FACIES ARCHITECTURE OF THE UPPER SEGO MEMBER OF THE MANCOS SHALE FORMATION, BOOK CLIFFS, UTAH

A Thesis

by

ERIC D. ROBINSON

Submitted to the Office of Graduate Studies of
Texas A&M University
in partial fulfillment of the requirements for the degree of

MASTER OF SCIENCE

December 2005

Major Subject: Geology

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Approved by:

Chair of Committee,	Brian J. Willis
Committee Members,	John Spang
	Beth Mullenbach
Head of Department,	Richard Carlson

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ABSTRACT

Facies Architecture of the Upper Sego Member of the Mancos

Shale Formation, Book Cliffs, Utah. (December 2005)

Eric D. Robinson, B.S., Morehead State University

Chair of Advisory Committee: Dr. Brian J. Willis

The Late Cretaceous upper Sego Member of the Mancos Shale exposed in the Book Cliffs of east-central Utah is a 30 m thick sandstone wedge that overlies the Anchor Mine Tongue of the Mancos Shale and underlies coastal plain deposits of the Neslen Formation. Although this sandstone has been interpreted to be comprised of transgressive valley fills, recent detailed facies architecture studies of the underlying lower Sego Sandstone suggest these deposits may instead be regressive deposits of tide-influenced deltas. This study maps facies associations, the geometry of lithic bodies, and key stratigraphic surfaces in order to define the architecture of a 12 km long cross section of the upper Sego Sandstone. This broadly depositional dip-oriented cross section exposes a vertical stack of three sandy intervals, truncated by a high-relief erosion surface, and capped by a shell rich lag. Sandy intervals are interpreted be an assemblage of forward stepping successions of tide-influenced delta lobes. Interval 1, dominantly highly marine bioturbated sandstones which thin landward over kilometers, is cut locally by an erosion surface overlain by tidal bed sets. It is capped by a localized transgressive shell lag and then a thin continuous marine shale. Intervals 2 and 3 are composed of stacked tidal bar deposits that successively coarsen upward and thicken basinward. Interval 2 is overlain by thin marine shales, whereas interval 3 is capped by a pronounced oyster shell lag ravinement surface. A high-relief erosion surface that extends from the top of the upper Sego sandstone down into the Anchor Mine Tongue Shale, is overlain by coarser-grained amalgamated fluvial channel deposits and is interpreted to be a incised valley fill. Erosion surfaces at the base of sandy intervals, thicken and decrease in marine bioturbation within successive intervals, and the valley cut into this succession reflects episodic forced regression of a deltaic shoreline.

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Last, but not least, thanks to my mother and father, Mike and Sheila Robinson, for their undying moral and financial support throughout very tough times. I would not be where I am today without such loving and caring parents. For this reason, I dedicate this thesis to them.

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INTRODUCTION

Facies models for tide-dominated deltas are less developed than those for wave and river dominated types due to the complexity of tide and river current interactions along coasts (Willis and Gabel 2001). It is difficult to distinguish tidal facies formed in estuary, delta, and shelf settings, because tidal influence is not restricted to the narrow littoral zone along the coastline (Johnson and Belderson, 1969). On river-dominated deltas, sediments can be assumed to fine systematically away from the coastline under a decelerating river mouth plume, and thus grain size trends within marine deposits can be inferred to reflect water depth (Bouma et. al, 1989). Tides complicate depositional patterns away from river distributaries by moving muds landward onto tidal flats and reworking sands into deeper waters below mean storm wave base (Willis et al., 1999). Depositional complexities caused by tidal reworking of river-deposited sediments make it difficult to recognize changes in relative water depth and to distinguish autocyclic from allostratigraphic facies trends and surfaces that define larger sequence stratigraphic divisions. High-relief erosion surfaces, which cut into coastal deposits, generally are related to either distributary channels or lowstand incised valleys (e.g., Van Wagoner, 1995). With the development of sequence stratigraphy, tide-dominated sandstones have been inferred to be transgressive estuarine deposits within incised valleys (Van Wagoner 1991), because they generally overly erosion surfaces. Although estuarine valley fills are clearly an important environment for preservation of tide-dominated deposits, evidence

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This thesis follows the style of the American Association of Petroleum Geologists.

of deep erosion may not always reflect lowstand fluvial incision (Ta et. al, 2002). The sandy tidal deposits formed in most depositional environments have basal erosion surfaces, and many recent oceanographic studies have documented significant submarine tidal erosion (Stride and others 1982). Although many studies of ancient outcropping marine sandstones report features which suggest tidal influence on deposition, few of these studies have documented the bedding geometry and the distribution of facies within a sequence stratigraphic framework (Van Wagoner, 1991; Kirchbaum and Hettinger, 1998; Willis, 2000; McLauren and Steel, 2000; Yoshida et al., 2001). Because sequence stratigraphers have generally inferred that all tide-influenced deposits are transgressive estuarine incised valleys fills, few outcropping tide-dominated deposits have been interpreted as deltaic.

Sandstone wedges in the foreland basin fill of the Sevier Orogeny commonly contain fluvial or wave-dominated coastal deposits (Kirchbaum and Hettinger 1998, Walker and Plint 1992). Tide-dominated deposits are less common, and have generally been inferred to be deposits of valleys or low relief embayments on delta top coastal plains formed as shorelines were transgressed (Willis and Bhattacharya 2001). The idea that strong tidal influence on deposition occurs only during transgressive embayment of shorelines lead Walker (1992) to propose that tide-dominated river deltas deposits do not exist. Because deltas had been previously defined to be areas of shoreline protrusion fed by rivers (Elliot, 1986), he suggested that tide-influenced deltas formed within larger-scale embayments would be better defined as sub-environments of estuaries. In contrast to Walker (1992), Dalrymple (1992) suggested that strong tidal influence would be expected during both retrogradational and progradational filling of embayments, and that tide-influenced prograding shorelines fed by a river should be referred to as deltaic whether they occurred within an embayed estuary or at a coastal protuberance. The idea that tide-influenced deltas occur only in embayments has been questioned in several recent studies of ancient deposits (Mellere & Steel, 1996; Bhattacharya and Willis, 2001; Willis and Gabel, 2003). Although Holocene sea level rise has lead to an abundance of modern embayed tide-influenced coastlines, in the past strongly tide-influenced coasts were expected to occur more commonly within other settings, including: areas adjacent to wide and shallow continental shelves, within shallow intracratonic seas, and along narrow seaways. More studies of tide-influenced strata are clearly needed to document the range of depositional setting where these types of deposits can form.

The Sego Sandstone exposed in the Book Cliffs of east-central, Utah, are classic outcrops of tide-influenced deposits. These strata show complex internal facies changes and multiple scales of heterogeneity. Because these tidal sandstones extend regionally over hundreds of miles, they are unlikely to have formed within isolated embayments along river- or wave-dominated coastlines. The Sego Sandstone is underlain by the transgressive marine Buck Tongue Member of the Mancos Shale Formation and is overlain by coast plain deposits of the Nelsen Formation (Young 1955, Van Wagoner, 1991; Kirchbaum and Hettinger, 1998; Willis, 2000; McLaurin and Steel, 2000; Yoshida et al., 2001). It is divided into two intervals separated by the Anchor Mine Tongue of the Mancos Shale. Previous detailed sedimentary architecture studies of these sandstones focused mostly on the lower Sego Sandstone (Van Wagoner, 1991; Willis and Gabel, 2001, 2003), which lead to different interpretations of depositional environments and sequence stratigraphic settings, ranging from a broadly transgressive successions of amalgamated valley fills to broadly regressive successions of top-eroded falling-stage tide-dominated delta deposits. Although facies within the upper Sego Sandstone are reportedly generally similar to those of the lower Sego Sandstone, details of facies variations and stratal architecture of the upper Sego Sandstone have not been documented.

This study presents a detailed mapping of facies architectural and stratigraphic surfaces across a continuous 12 km outcrop of the upper Sego Sandstone. This field investigation builds on previous work completed on the lower Sego Sandstone by tracing important stratigraphic surfaces (including high-relief erosion surfaces) and facies variations to interpret processes of deposition within the upper Sego Sandstone. Facies trends between allostratigraphic surfaces are used to define the sequence stratigraphic setting. Tide-influenced deposits comprise an important class of hydrocarbon reservoirs with complex internal heterogeneities that make it difficult to develop efficient production strategies. Analog deposits exposed in large outcrops show the scale and geometry of reservoir complexities. Understanding the internal structure of reservoir properties formed in different depositional environments and sequence stratigraphic settings, leads to predictive models of inter-well scale reservoir variability. Interpretation of this variability in terms of changes in depositional and sequence stratigraphic processes, can lead to more accurate models to predict properties of reservoirs that occur in analog deposits.

Tidal Influences on Deltaic Deposition

Deltaic deposits are regressive sediment bodies formed as fluvial sediment discharged into the sea (Coleman and Wright 1975). River-dominated deltas exhibit rapid unidirectional flows and faster deposition rates in shallower waters near distributary mouth bars, which decrease as flows disperse offshore (Dalrymple 1992). As mouth bars grow and fill local accommodation, associated distributary channels will avulse to another location. Following avulsion, waves can erode bar tops, or local flooding surfaces may develop as depositional grain sizes decrease to mostly mud carried along the coast from more distal distributaries. Repeated growth and

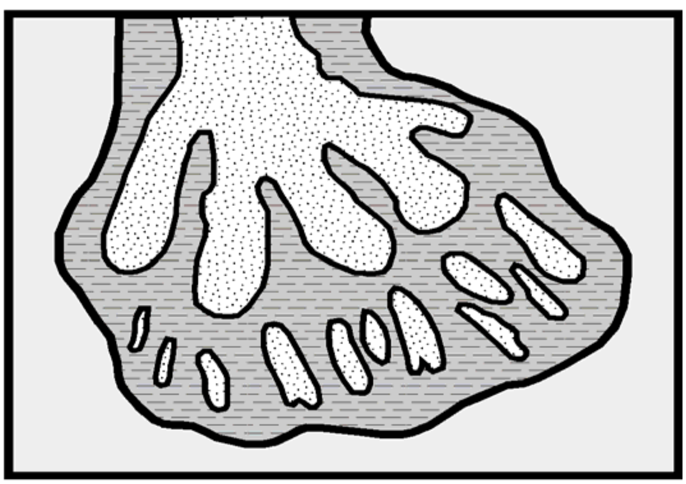


Figure 1. Net Distribution of Sand Bodies within Tide-Dominated Deltas. Illustrates the effect of low wave energy, low littoral drift, and high tide (Coleman and Wright, 1975).

avulsion of mouth bars and larger-scale delta lobes produce a lateral stacking of lobate sand bodies.

Tide-dominated deltas occur where wave energy is low and tidal range is high (Coleman and Wright, 1975). The reworking of river supplied sediments by tides tends to elongate coastal sand bodies perpendicular to the shoreline (Figure 1). This produces a more complex facies patterns from the coast offshore along the deltaic bathometric profile (Bhattacharya 1993). Most distinctions between tide-dominated and other types of deltaic facies have focused on recognizing evidence for frequent variations in paleocurrent speed and direction. Few studies have documented larger-scale differences in the distribution of facies, the geometry and arrangement of sediment bodies, and the expression of allostratigraphic surfaces that define sequence stratigraphic discontinuities.

Tide-dominated delta deposits (like those of river-dominated deltas) are fed by major distributary channels that deliver sediment to particular areas along the coastline (Dalrymple 1999). Tides tend to accelerate onshore into mouths of distributary and tidal channels and then decelerate onto interdistibutary tidal flats as these flows overbank channels during high tide. Muds can be carried offshore during infrequent major river floods or reworked onto proximal areas of the delta top by landward acceleration and then deceleration of flood tides during times of normal river discharge. This process leaves sandier deposits along the delta front and subtidal areas of the delta top, which can be reworked into shoreline elongate bars. Unlike river-dominated deltas, where sandy deposits along the coast tend to become covered by muds following upstream distributary channel avulsion, tidal reworking of the delta front may become more pronounced following distributary channel abandonment. This can produce erosion and a coarsening of deposits due to increased winnowing of fines following abandonment, rather than a fining of deposits.

Distributary channels on the tide-influenced Fly River delta were enlarged from 7 meters to nearly 30 meters deep in a period of 100 years following abandonment (Dalrymple et al., 2003 and Baker et. al, 1995). Bathymetric maps of other tide-influenced river deltas also reveal deep incisions cut into abandoned areas of deltas (Hori et al., 2002; Kuehl et al., 1989; 1997). The scale of enlarged distributaries incised into progradational delta lobes may resemble lowstand valleys, but internal facies trends should be distinct. Valley incision fills may have coarser-grained fluvial deposits at their base and are expected to have a retrograding estuarine fills deposited as seas progressively flooded the valley. Some areas within abandoned distributaries may initially fill with finer-grained sediments carried by tides along the coast (Nio and Yang, 1991). If a distributary channel has reoccurred, however, it may be filled by an upward-coarsening succession that records initial stages of renewed delta progradation. Deposits basinward of incised valleys should be lowstand delta deposits. Sands extending basinward of tidally enlarged distributaries are expected to be less extensive. That said, because the growth of tidal sand bars past the delta front edge may be suppressed in front of active distributaries by rapid deposition of prodelta mud, distributary erosion may initially be associated with tidal reworking of sands farther into the basin.

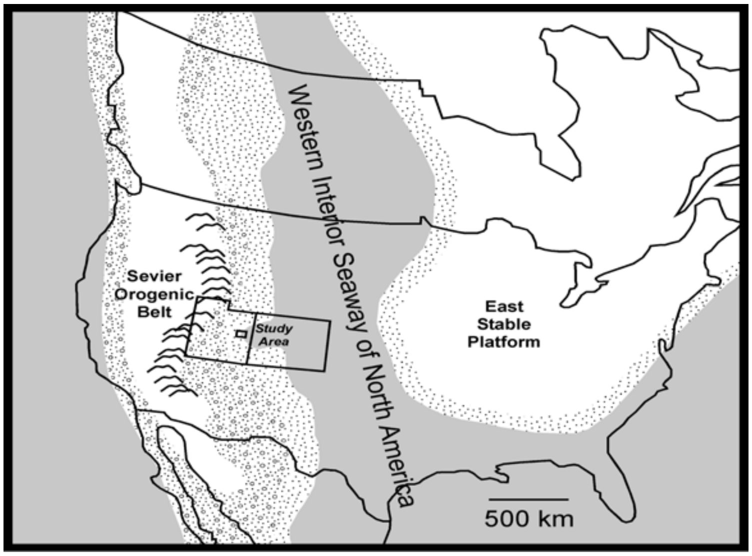
Regional regressions and transgressions of coasts are interpreted from deltaic deposits by recognizing systematic vertical changes between successive autocyclic progradation-abandonment successions (parasequences), which are predicted to show varying stacking trends in different systems tracts (e.g., highstand, falling stage, lowstand and transgression). During regional regression, delta front deposits within successive parasequences will shift basinward and may steepen and become more wave-dominated (e.g., Ta et al., 2002). Tidal erosion of abandoned distributaries may be more pronounced during falling stage sea level, because preserved deltaic deposits are likely to be relatively thin and the tidal prism may flood long distances across low gradient deltas. Elongate bar growth basinward of delta fronts may be promoted during sea level lowstands within intracratonic seaways, where wide and shallow seas enhance tide-influence. Delta top facies are more likely to be preserved during initial stages of sea level rise as delta front deposits vertically aggrade, preventing marine ravinement. With continued transgression, shoreline deposits may be reworked into shelf tidal bars (Snedden and Dalrymple, 1999). It may be difficult in tide-dominated deltaic deposits to distinguish facies variations caused by local delta progradation and then local abandonment from those produced by more regional regressions and transgressions. Outcrop analysis allows detailed examination of multiple scales of variations details of facies variations, the geometry and lateral continuity of lithologic trends, and characteristics of regional stratigraphic surfaces. An understanding these complexities can then be used to develop correlation strategies and methods of inter-well scale heterogeneity prediction within analog reservoirs.

GEOLOGIC SETTING

The upper Sego Sandstone was deposited on the margin of Cretaceous Western Interior Seaway, an elongate basin that extended from the present Gulf of Mexico to the Boreal Sea (Figure 2A). This foreland basin formed by crustal loading and flexural subsidence under thrusts in the Sevier Orogeny (Roberts and Kirchbaum 1995). Tidal deposits of the upper Sego Sandstone formed along the western margin of the Interior Seaway as a wedge of southeastward prograding siliciclastic strata shed from the Sevier fold-thrust belt located to the west (Van Wagoner, 1991). The Upper Cretaceous Sego Sandstone lies within a finer-grained interval of this foreland basin succession that separates coarse-grained fluvial wedges of the underlying Castlegate Member of the Mancos Shale Formation and the overlying Blue Castle Member (Figure 2B). A westward thinning marine shale, the Buck Tongue Member of the Mancos Shale Formation , overlies and onlaps a major sequence boundary near the top of the Castlegate Member (Van Wagoner, 1991; Kirchbaum and Hettinger, 1998). The Buck Tongue records major transgression across the relatively flat-topped Blackhawk-Castlegate prograding wedge. Overlying the Buck Tongue Member, the Sego Sandstone Member erosionally truncates and episodically progrades to the southeast, recording regression of shorelines obliquely into the basin (Figure 2B). The Neslen Formation, overlying the upper Sego Sandstone, records an aggradational coal-bearing coastal plain and fluvial successions located updip from vertically-stacked, wave-dominated shoreline deposits of the Corcoran, Cozzette and Rollins Sandstones located in western Colorado (Kirchbaum and Hettinger, 1998). The Sego Sandstone Member is divided into upper and lower members separated by the transgressive marine Anchor Mine Tongue Shale (Willis and Gabel, 2001). A biozone defined by ammonites from the upper Buck and Anchor Mine Tongues (Figure 2B) indicate the Sego Sandstone Member is Late Campanian (76.0 – 74.6 Ma) and that the lower Sego Sandstone was deposited over about 1.5 my (Van Wagoner, 1991).

Although there is broad agreement that depositional facies within the Sego Sandstone show significance influence of tides, previous studies have suggested widely different interpretations of depositional environments and sequence stratigraphic settings. Van Wagoner et. al (1990) suggested that the Sego Sandstone is composed of vertically amalgamated estuarine valley fills formed during high frequency variations in sea level. Nine regional sequences were defined by recognizing erosional transitions from wave-dominated strata to coarser-grained tide-dominated strata. Although he implied that the Sego Sandstone Member was a regionally regressive unit, he suggested regressive shoreline deposits were not preserved because rapid sea level falls caused abrupt shifts in shorelines basinward, and subsequent transgressive ravinement by tides reworked all falling stage deposits landward into valley fills. Yoshida et al. (2001) suggested that the Sego Sandstone records a long period of transgression above a major sequence boundary incised into open marine shales of the Buck Tongue. This was based on the observation that as the Buck Tongue thins landward, it becomes truncated at the erosional base of the Sego Sandstone Member. They suggested that this erosion surface was a regional angular unconformity of higher stratigraphic order than the

A)



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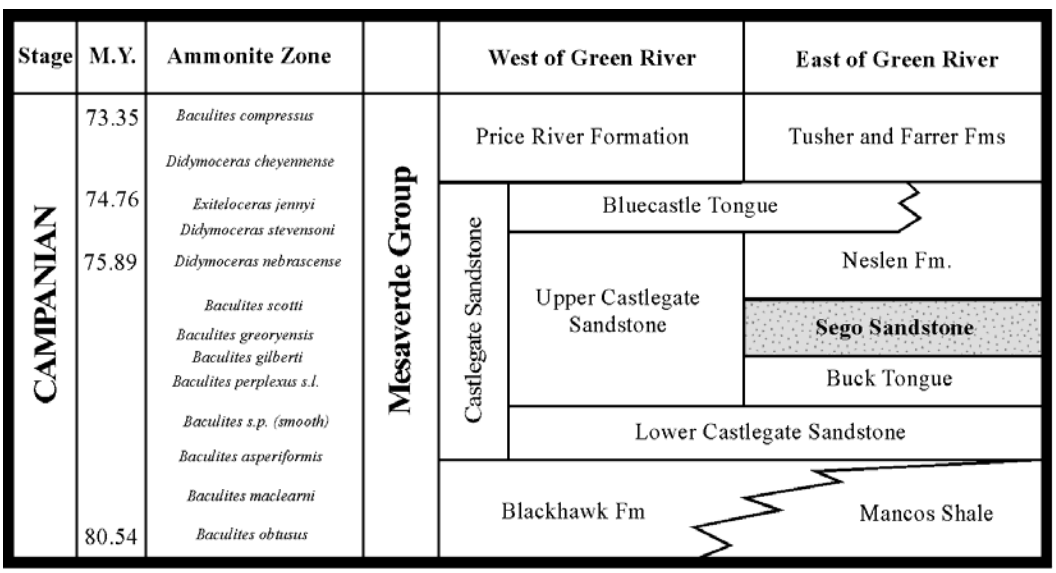
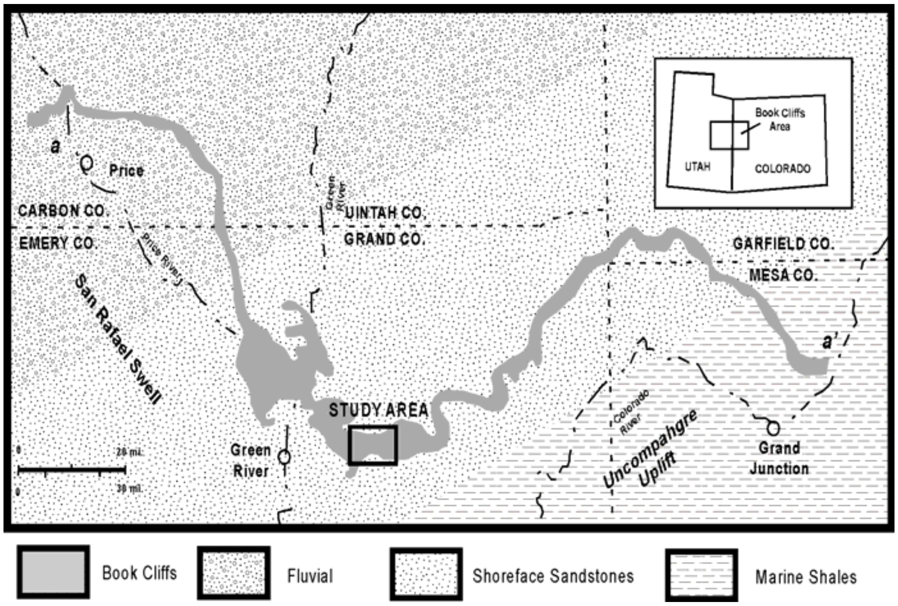


Figure 2. Paleogeography and Stratigraphic Nomenclature. A) Stratigraphic chart revealing biozonal control and overall stratigraphic positioning of the Sego Sandstone. B) Regional paleogeographic map showing location of the shoreline within the Western Interior Seaway of North America (after Roberts and Kirchbaum 1995). The sea embayment surrounding the the study area was postulated by Roberts and Kirchbaum (1995) to explain tide-dominated deposition within the Sego.

"high-frequency" sequence boundaries recognized by Van Wagoner et. al (1990). McLaurin and Steel (2000) disputed the existence of a high-order regional erosion surface at the base of the Sego Sandstone Member and suggested instead that there is a gradual facies transition landward from the Buck Tongue into tide-influenced layers within fluvial-dominated parts of this foreland basin fill.

Willis and Gabel (2001; 2003) recently reinterpreted the lower Sego Sandstone to be an overall progradational succession of tide-dominated deltaic deposits (Figure 3A and 3B), based on the interpretation of more detailed maps of facies trends and the geometry of key sequence stratigraphic surfaces. They recognized that the erosion surface at the base of the Sego Sandstone changed in stratigraphic position when traced along depositional strike, and suggested that many of the vertical transitions from wave- to tide-dominated strata within the Sego Sandstone defined by Van Wagoner et. al (1990) could reflect submarine tidal erosion of offshore marine areas basinward of prograding deltas. Their interpretations followed other recent papers of clastic wedges preserved in the Western Interior Seaway that recognized evidence of marine erosion in wave-dominated systems when falling sea level forced shorelines to regress and lowered wave base onto offshore areas of the basin floor (Plint, 1988; Posamentier et al, 1992; Plint and Nummedal, 2000; Posamentier and Morris, 2000). They suggested that the lower Sego Sandstone was dominantly a regressive succession of deltaic lobes that episodically prograded into the basin and were tidally eroded when local areas were abandoned. Although Kirchbaum and Hettinger (1998) provided insight that the most distal marine deposits of the upper Sego Sandstone were deposited during falling stage

A)



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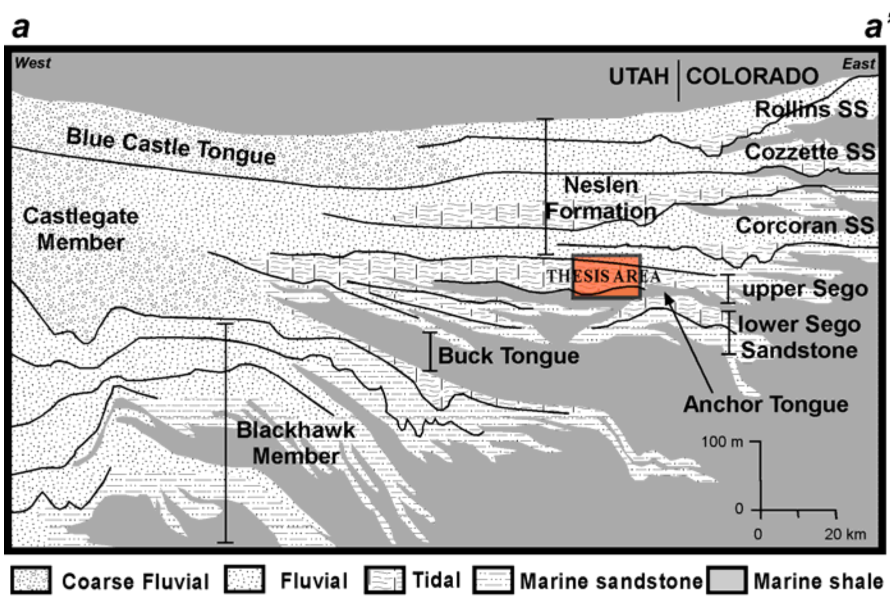

Figure 3. Book Cliff Basemap and Cross-section. A) Regional basemap illustrating paleo-depositional environments within exposures of the Book Cliff region. B) Regional cross-section of deposits exposed from Price, Utah to Grand Junction, Colorado (from Willis and Gabel 2001).

regression, there have not been comparable studies of more proximal exposures of this sandstone.

METHODOLOGY

Sego Sandstone exposures extend 150 km from Green River, Utah, wedging out just west of Grand Junction, Colorado (Figure 4; Van Wagoner, 1991; Kirchbaum and Hettinger 1998). The study area is in the western (proximal) part of this exposure between Horse and Crescent Canyon in central-east Utah; a distance of about 12 km (Figure 4). There, the Sego Sandstone dips to the north at a few degrees, but is otherwise undeformed. Although exposures are nearly continuous across this interval, in places, finer-grained facies are poorly exposed. The upper Sego Sandstone interval examined is defined by the Anchor Mine Tongue at its base, and at its top by the first carbonaceous shale near the base of the Neslen Formation. These continuous, nearly stratigraphically horizontal layers were used as a datum for 20 sedimentological logs spaced every half kilometer along the outcrop. Sedimentologic logs record vertical variations in grain size, sedimentary structures, bioturbation, and the orientation of paleoflow indicators. Stratigraphic surfaces and facies variations were mapped between logs and projected into a plane oriented 120 degrees east of north (Figure 4). Photomossaics of well exposed cliff exposures along this cross section were compiled to show details of bedding and facies variations at specific locations (Plates 2 and 3). These facies variations and surfaces defined by erosional truncation, abrupt grain size changes, or distinctive thin beds were interpreted in terms of changing depositional environments and allostratigraphic discontinuities.

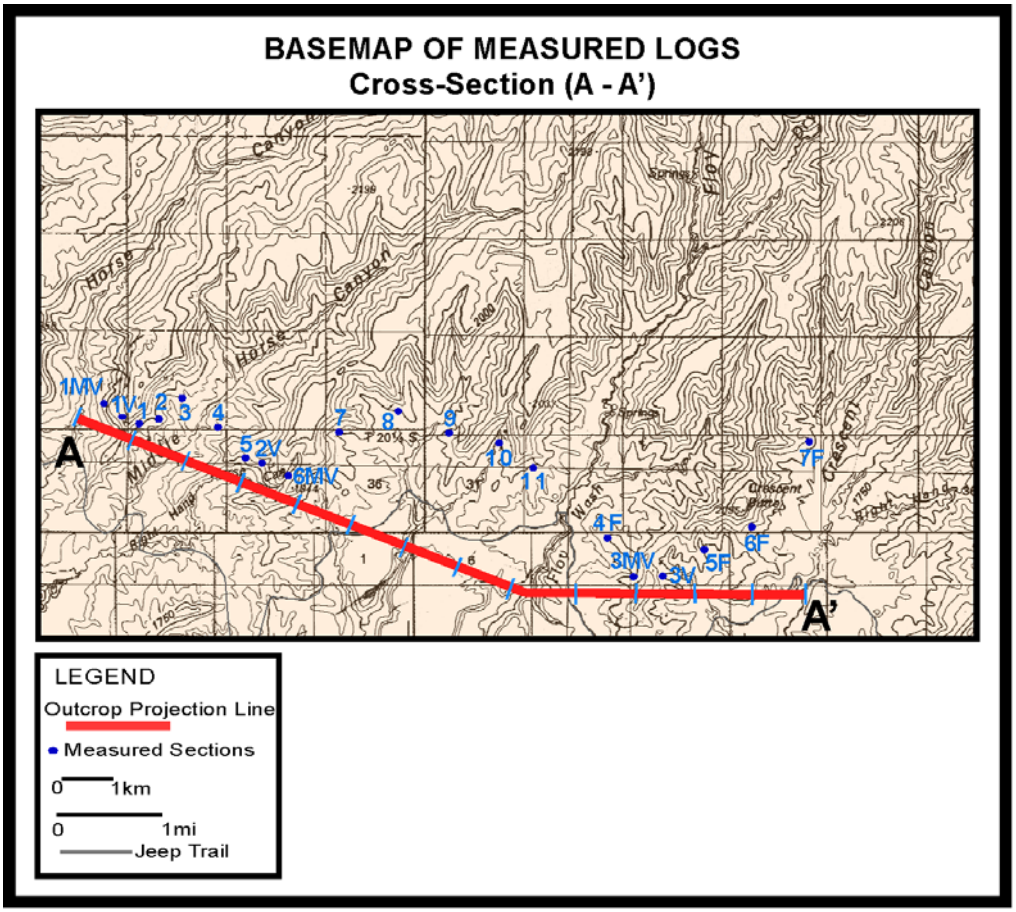

Figure 4. Topographic Basemap. Illustrates outcrop projection line marked by kilometer intervals and their associated points of measured section.

FACIES ASSOCIATIONS

The upper Sego Sandstone Member is subdivided into four facies associations: (1) Marine bioturbated shales with wavy bedded sandstones, (2) Marine bioturbated sandstone, (3) Tidal inclined-bed sets, and (4) Channel-form bed set. These facies are broadly similar facies were described by Van Wagoner (1991), Willis (2000), and Willis and Gabel (2001). Each is described in detail below.

Marine Bioturbated Shales with Wavy-Bedded Sandstones

Description

These 0.5 – 1 meter thick beds comprise medium-dark grey, bioturbated shales interbedded with lenses of symmetrical cross laminated, lenticular to wavy, very fine sandstones. Beds extend laterally across the 12 km interval studied (Figure 5A). Interval stratification is commonly obscured by intense bioturbation, which includes *Planolites* and *Chondrites* burrows. There is a gradual increase in sandstone, bioturbation, and a lightening of grey hues upward.

Interpretation

Shales accumulated from suspension offshore below fair-weather wave base. Lenticular sands with wave ripples record oscillatory flow during storms. Dark shales near the base of beds suggest sediment starvation and concentration of organic marine matter in deeper, oxygen-depleted areas offshore. More bioturbated, lighter-grey shales interbedded with wavy sandstones in the upper parts of beds indicate better open marine circulation above storm wave base (McCrimmon and Arnott 2002).

Marine Bioturbated Sandstones

Description

Beds are comprised of massive, pale to dark reddish brown sideritic, very fine to upper fine sandstone (Figure 5B). The facies occur in 2 – 6 meter thick beds that extend continuously across the 12 km area studied. Beds have erosional bases and in places are overlain by concentrations of sideritic nodules. Beds thicken towards the east. Although intense bioturbation by *Ophiomorpha*, *Chondrities*, and *Paleophycus* obscures most sedimentary structures, poorly preserved small-scale cross-strata, rare parallel laminations, and symmetrical ripples are observed locally.

Interpretation

Fine caliber sand, intense bioturbation by marine trace fossils and the lack of preserved sedimentary structures is characteristic of a low-energy marine distal shoreface setting below fair-weather wave base and above storm wave base (Howard and Reineck 1981, Pemberton and Frey 1992). The preservation of massive ichnofabrics reflects disruption of depositional stratification by low intensity currents during infrequent storms (Pemberton and Frey 1984). The thickening of beds to the east reflects increased accumulation below storm wave erosion basinward (Taylor and Gawthorpe, 1993).

Tidal Bed Sets

Description

Tidal bed sets range from 3 – 8 meters thick and vary gradually in thickness across

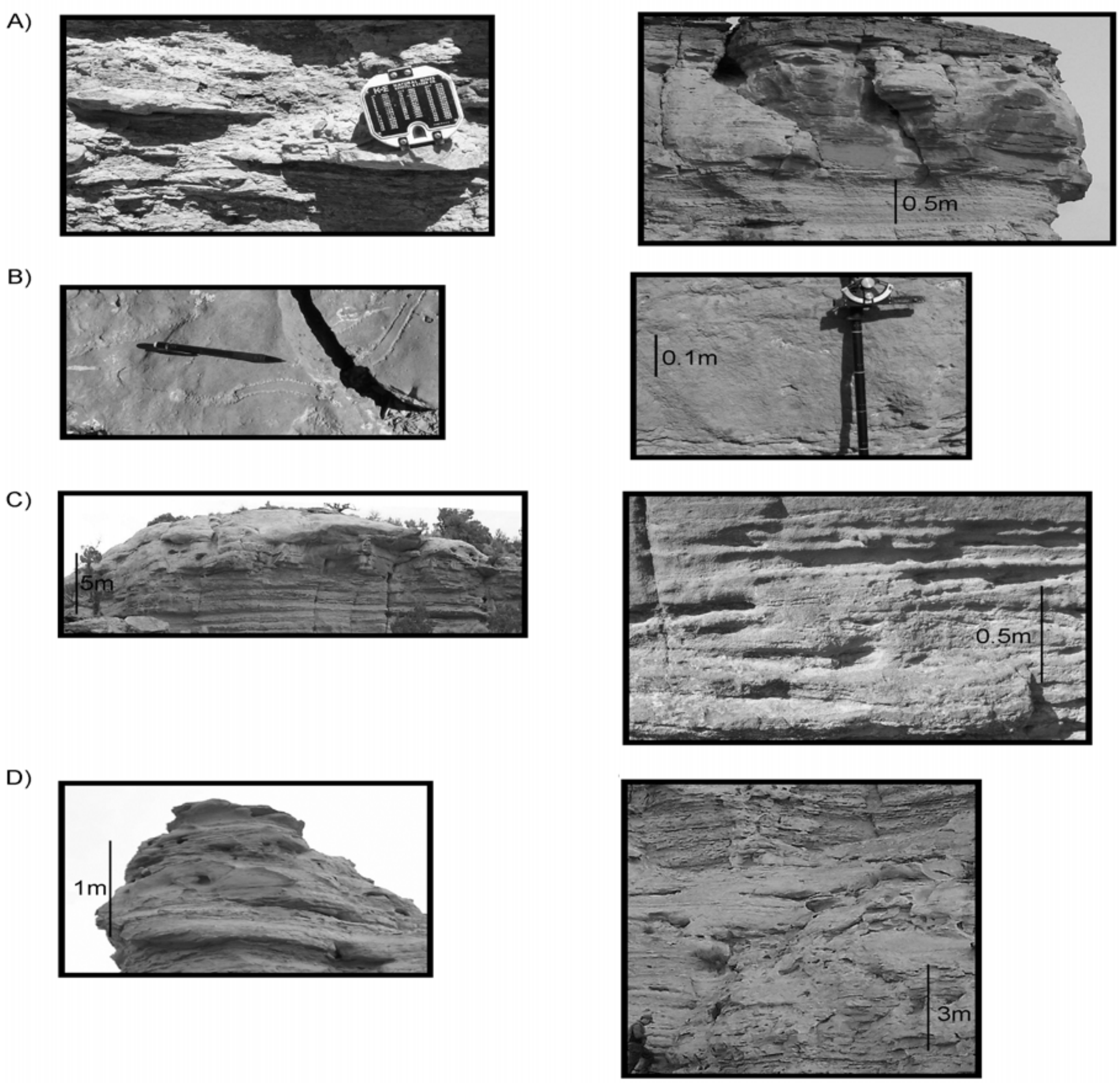

Figure 5. Facies Photographs. (A) Facies Association 1, laminated mudstone (B) Facies Association 2, *Ophiomorpha* sandstone (C) Facies 3 tidal inclined bed-sets, (D) Facies Association 4, Channel bed-sets.

the study area (Figure 5C). Bed sets have low-relief basinal erosion surfaces overlain by coarse intraclastic lags of mud clasts or cemented sandstone of underlying facies. The bed sets coarsen upward from a very fine to upper medium sandstone. Although internal beds locally dip at a few degrees to the southeast, in many locations dips are too low to easily observe in outcrop. Basal beds are thinner and more heterolithic; beds become thicker and more sand-rich upward. Bed sets are generally capped by extensively bioturbated sandstone. Thicker bed sets have a lower proportion of basal heterolithic beds than upper sandy beds. As bed sets thin they become more heterolithic on average. Vertical transitions between heterolithic and sandier beds are generally gradual, but can be locally abrupt. The degree of bioturbation (mostly *Ophiomorpha)* is highly variable, ranging from complete destruction of depositional stratification to beds with no evidence of borrowing activity.

Basal heterolithic intervals within bed sets vary from 0.5 to just over 1 meter thick. Centimeter- to decimeter-thick sandstone beds contain thin sets of very low-angle large-scale cross strata and fine upward to oppositely-dipping flaser cross lamination, asymmetrical ripple marks, rare hummocky cross-strata, and extensive mudstone drapes. Bioturbation is sparse to moderate. Decimeter- to meter-thick sandstone beds in upper parts of sets are generally erosionally amalgamated and contain near angle of repose, large-scale cross strata. These beds can be draped by discontinuous layers if cross lamination and scarcely preserved mudstone drapes. Cross strata sets contain abundant reactivation surfaces and rare oppositely oriented ripple marks and double mud drapes superimposed on cross strata. Cross strata dip directions are fairly consistent within bed sets, but vary significantly between different examples; commonly averaging either northwest or southeast. Extensively bioturbated beds capping sets can be distinctively coarser-grained. Although ichnofabrics are generally horizontal, *Ophiomorpha* with vertical shafts a half meter in length are observed.

Interpretation

Tidal inclined bed sets are deposits of subtidal coastal tidal bars. The thickness and continuity of bed sets suggest bars were several meters high, 3 to 8 km wide and several kilometers long. Low-relief basal erosion surfaces indicate a fairly regular depth of current scour between bars. Very low angle internal inclined beds indicate bars had subdued topography, rather than angle of repose front faces formed in a zone of lee flow separation. Basal heterolithic sands show greater evidence of tidal modification whereas thicker sandstone-rich beds higher in sets record stronger unidirectional currents. Intensely bioturbated sandstones capping bed sets indicate infaunal activity after the bar stabilized, perhaps after sand supply maintaining bar growth shifted to another location. Bed sets with internal cross strata that dip dominantly to the southeast grew by accretion of ebb transported sands, whereas those with cross strata that dip dominantly to the northwest record flood-oriented accretion. Similar upward-coarsening tidal bars have been documented in a range of modern tide-influenced depositional setting, including estuaries, deltas and on the shelf. Contrasts between tidal bars formed in these different marine environments are related to the extent of lee flow separation, dominance of offshore verses onshore bar migration directions, the elongation and orientation of bar axes relative to the coast, and the degree to which bed and suspended sediment loads have parted during transport. Although it is difficult to constrain many of these parameters from examination of a single cross section, some inferences can be advanced. The heterolithic nature of these deposits suggest deposition closer to shore rather than long distances out on the shelf. Uniformly low-angle accretion surfaces also suggest these are not the deposits of large marine dunes on the shelf. Dominance of ebb-oriented accretion bed dips (even in bars with dominantly flood-oriented internal cross strata) favors an offshore prograding deltaic rather estuarine tidal setting. Low relief basal erosion surfaces and open marine bioturbation of abandoned bars is more characteristic of bars that grew along the distal ends of a subtidal delta top and delta front than those within distributary channels incised into adjacent tidal flats.

Channel-Form Bed Sets

Description

Channel-form bed sets are 2 – 7 m thick, several hundred meters wide, and have upward concave basal erosion surfaces (Figure 5D and 6). These bed sets commonly occur in amalgamated stacks restricted to localized areas. Basal erosion surfaces are overlain by a concentration of mud chip intraclasts and coarse sand grains. Bed sets fine-upward from coarse to lower fine grained sandstone. Decimeter to meter thick beds within sets have erosional bases and can dip up to 5 - 10 degrees onto the basal erosion surface. Individual beds span horizontal distances less than 100 m. In one example, successive beds progressively decrease in dip along the set. Although decimeter-thick cross-strata dominate most beds, cross lamination caps some beds that fine upward.

Cross strata dips are typically southward, at a high angle to exposures of bed sets with steepest dipping internal beds. Trace fossils are generally absent, except in a few examples that directly underlie marine erosion surfaces that are extensively bioturbated by *Ophiomorpha.*

Interpretation

These are deposits of river or distributary channels that were up to at least 8 meters deep and on the order of 100 meters wide. Mud intraclasts at the base of bed sets accumulated as cut banks were eroded during channel migration. Internal inclined beds record episodic migration of point bars that did not have lee flow separation during floods. Distances spanned by individual inclined beds relative to the width of sets (> 1/2) indicate channels had low to moderate sinuosity. Cross strata within beds record migration of dunes across channel bars during major floods and cross laminations capping some fining upward beds record ripples that migrated during low stage flows. Bed dips highly oblique to dominate internal cross strata dip directions indicate lateral accretion. Dominate southward dips of cross strata are basinward. Little evidence for tidal modulation of flows was observed. Low levels of bioturbation is characteristic of fluvial channels and *Ophiomorpha* capping some bed sets occurred during marine transgression after deposition of the channel deposit.

STRATAL ARCHTECTURE

The upper Sego Sandstone is divided into three sandy intervals separated by shale layers that are nearly continuous across the area mapped. Facies variations and distinctive stratigraphic surfaces that define these intervals are shown in plate 1. Observed variations are described and interpreted in terms of changing depositional environments, before sequence stratigraphic interpretations are discussed.

Description

The upper Sego Sandstone is underlain by the regional Anchor Mine Tongue Shale. The base of the upper Sego Sandstone is defined by an abrupt coarsening to sandstone above a nearly horizontal erosion surface that is locally overlain by an intraclastic lag. The basal deposits are a laterally-continuous, 2-6 meter thick beds of marine bioturbated sandstone abruptly overlain by a laterally continuous, 0.5 to 1 meter thick bed of marine bioturbated shale. The marine bioturbated sandstone bed thins significantly to the west along the cross section documented. In one location, near the eastern end of the cross section, a channel-shaped erosion surface locally cuts down from the top of the marine bioturbated sandstone into the Anchor Mine Tongue. This nearly ten meter deep and a few hundred meter-wide incision is filled by two stacked tidal bed sets. Facies are heterolithic within thinner beds lower within sets and become highly bioturbated sandstones within thicker beds higher in sets. Basal beds within sets contain hummocky cross strata and symmetrical ripples marks, as well as tide-influenced cross strata more characteristic of this facies association.

The second sandy interval erosively overlies the shale bed that caps the first. The interval contains two stacked tidal bed sets and is capped by a meter thick bed of marine bioturbated sandstone abruptly overlain by a meter thick bed of marine shale. Although the interval has uniform thickness along the cross section, the lower tidal bed set thins significantly to the west as the upper tidal bed set thickens. Vertical upward-coarsening facies variations and maximum thickness (~4m) of these two bed sets are similar. The upper bed set is on average slightly coarser and somewhat more bioturbated. Cross strata within the lower bed set is dominantly to the southeast, whereas those in the upper bed set are dominantly toward the northwest. The capping marine bioturbated sandstone has distinctive dull red hue, contains randomly distributed mud clasts, and varies little in thickness along the cross section.

The third sandy interval, erosionally overlying the shale bed that caps the second, is 12 meters thick in the central area of the cross section and thins laterally in both directions to fewer than ten meters thick. The basal erosion surface is nearly stratigraphically horizontal. The interval consists of three upward-coarsening tidal bed sets. Successive vertically stacked bed sets are thicker, more sandstone-rich on average, and show more pronounced thickness variations in cross section. The basal bed set becomes as coarse grained as upper fine sandstone, whereas the upper most bed set can be as coarse as upper medium sandstone. The second bed set is separated vertically from the first by a bioturbated bed that is dull red and cemented. The base of the third bed set locally incises into the second. Bed sets tend to be sandier where they are thicker. Cross strata within the basal bed set is dominantly toward the northwest, whereas those within the upper two bed sets dominantly dip to the southeast.

Along the western side of the cross section, a high-relief erosion surface cuts from the top of the upper Sego Sandstone through the three sandy intervals just described. This erosion defines two distinct channel-form incisions (Figure 6). The larger one, several kilometers in width, cuts downward 25 meters into underlying deposits. At the point of greatest incision, this surface cuts through the Anchor Mine Tongue into the top of the lower Sego Sandstone. Where incision is greatest, basal deposits above the erosion surface are heterolithic. Most of the deposits that fill these incisions are sandy channel-form bed sets. Above the areas of deepest incision four channel-form bed sets are stacked vertically. In the other location of incision, the erosion surface is overlain by two vertically staked bed sets. Inclined beds that dip to the southeast within one of these bed sets are well exposed in a vertical cliff in the east side of Horse Canyon (Plate 2).

The upper Sego Sandstone is capped by a decimeters-thick cemented bed overlain locally by a distinct oyster shell lag. Although basal deposits of the Neslen Formation are uniformly poorly exposed along the cross section, a thin bed of coal defines a distinct dark layer across the weather slopes.

Interpretation

The Anchor Mine Tongue records slow, offshore accumulation of marine muds during regional transgression (Kirchbaum and Hettinger 1998). Increase in filling

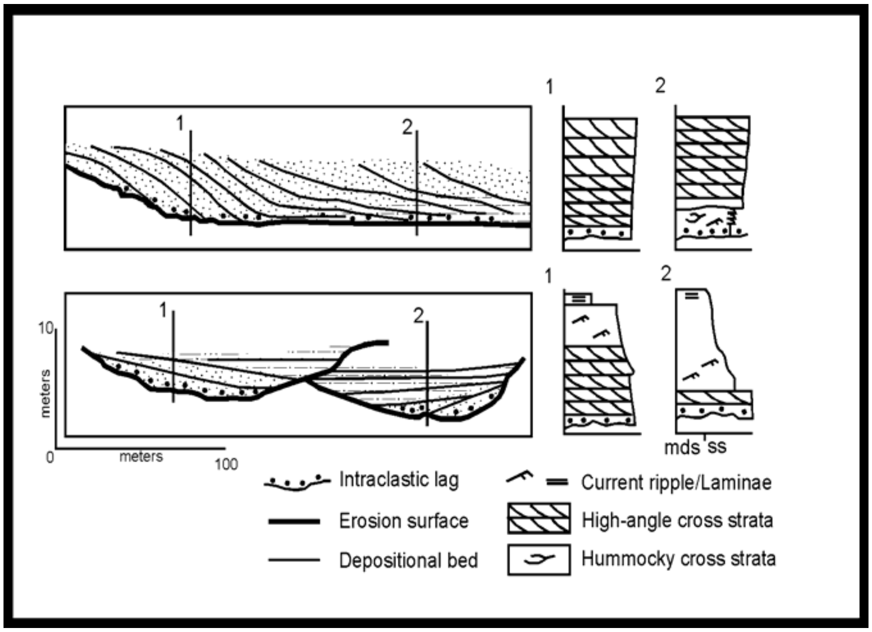

Figure 6. Tidal Bed Set and Channel-Bed Set Stratal Architecture. Stacking patterns and sharp facies variations between channel bed-set geometries, tidal inclined-bed sets, and high relief erosion surfaces. (from Willis 2000).

thin wave-rippled sandstone beds upward as shales become more bioturbated and lighter grey may record gradual shoaling before upper Sego Sandstone deposition. The three sandy intervals that comprise the upper Sego Sandstone, separated by marine shale beds, record episodic regression and then transgression of tide influenced shorelines (Figure 7). Increase in thickness and average grain size, and decrease in marine bioturbated sandstone of successive intervals, suggest preservation of progressively more shoreline proximal facies within successive intervals (i.e., a forward steeping progression).

Although the bioturbated basal interval is interpreted to be distal shoreline deposits, gradual lateral thinning of this sandstone along the cross section indicates lateral variation in sediment supply. Lateral thickness variations of the two tidal bar deposits within the second sandy interval suggest compensational deposition, with the upper bar in topography remaining after deposition on the underlying bar deposit. Southeastward dipping cross strata that dominates the lower bed set suggest this bar prograded offshore, whereas northwestward dipping cross strata that dominates the upper bed set suggests landward accretion. Although these differences in accretion direction may have formed during regression and transgression of shorelines, respectively, bars migrating simultaneously in ebb and flood tide directions are commonly observed along both estuarine (Harris, 1988; Harris et al., 1992) and deltaic (Allen and Chambers, 1998; Dalrymple et al, 1992) tide-influenced coastlines. It is a common observation in both settings that ebb and flood flows follow different paths on and off the coast due the oblique convergence of tidal waves onshore (Thomas and Anderson, 1989). Dominate paleocurrent directions within bed sets thus may reflect local (rather than regional stratigraphic) variations in sediment accretion. This interpretation is supported by variations within the third sandy interval, which has a basal flood accreting bar overlain by two basinward accreting bars. Lateral thickness changes of bed sets, more pronounced for thicker bed sets interpreted to reflect more proximal deposition, suggest bars had lobate to shoreline perpendicular elongate shapes.

Because tides can rework sands into bars that extend a variety of distances from the shoreline, it is difficult to estimate both depositional water depth and how relative sea level changed during the accumulation and transgression of the sandy intervals. Key to interpretation of these intervals is the origin of basal erosion surfaces. Prograding shorelines and delta lobes are expected to generate gradually-upward-coarsening deposits, reflecting deposition under progressively stronger currents as waters shoal. It is increasingly recognized that shoreline regression, associated with sea level falls ("forced

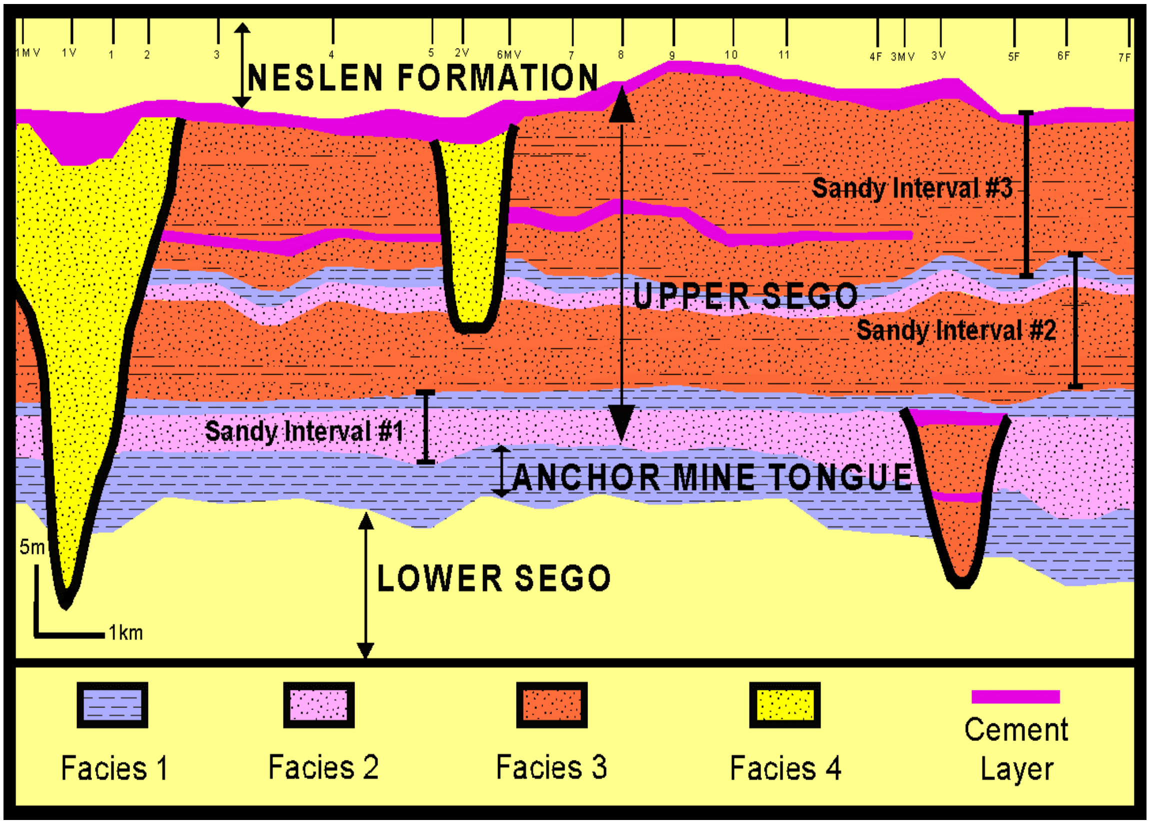

Figure 7. Compressed Cross-Section. Illustrates facies associations and their applicable sequence stratigraphic designations. Bedding diagram (true scale) illustrating facies association distribution within the upper Sego.

regression" of Posamentier et al., 1992), can lower wave base level to the sea floor farther offshore. Shoreface deposits formed in this setting can abruptly coarsen vertically from offshore shales to proximal shoreface sands deposited above wave base (McCrimmon 1996). Presumably, this same process can occur along tide-influenced shorelines, even though tidal current energy is less well correlated with water depth than it is within wave-dominated settings. Alternatively, marine wave or tide current energies may have changed over time with changes in climate or basin shape induced tidal resonance, but these mechanisms cannot be assessed by examination of only this local

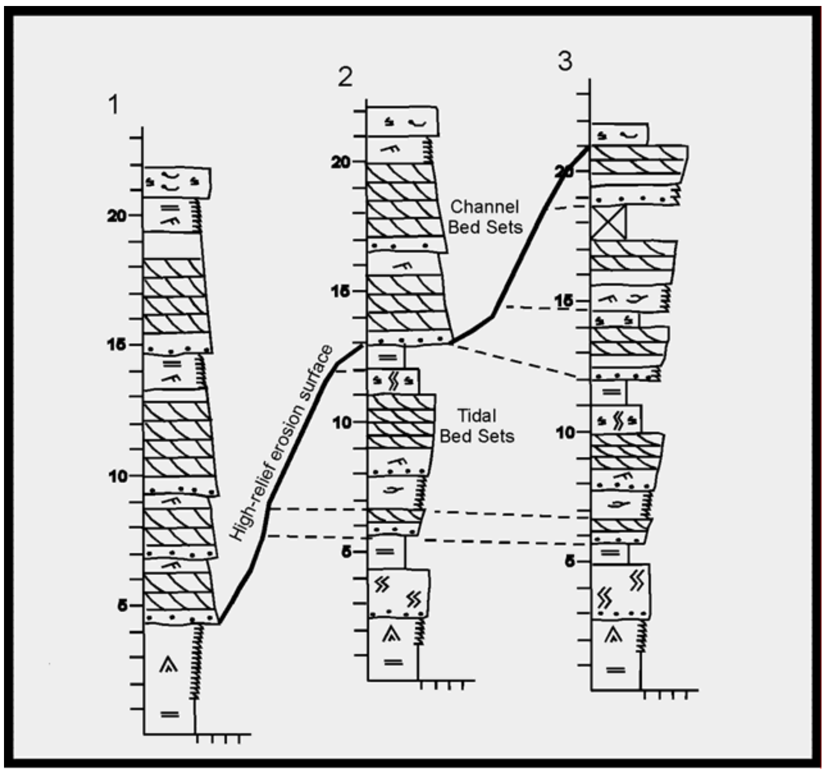

Figure 8. Diagram of a High-Relief Erosion Surface. Eastern margin of valley #1 illustrating sharp lateral facies variations.

cross section. Given the clear regressive trend upwards through these three intervals, it seems unlikely that each layer records lowstand incision and estuary filling (Van Wagoner et. al, 1990). Although tens of meters of transgressive erosion have been documented at the top of some sandstones within Cretaceous Interior Seaway (Posamentier and Vail 1988, Bhattacharya and Willis 2001) deep enough to rework entire shoreface deposits landward, there is no evidence of coarse-grained lags of extrabasinal clasts that would suggest this magnitude of erosion along basal erosion surfaces of intervals in the upper Sego Sandstone. Increase in ebb-oriented paleocurrents within successively thicker sandy intervals suggest progradation, rather than accumulation of sands during ravinement.

The 10 meter-deep incision into the first sandy layer, a layer interpreted to record the most shoreline distal facies, may have formed by localized fluvial incision and then subsequently filled by estuarine tidal bars during transgression (plate 3) This interpretation seems unlikely, however, because it suggests profound short-term changes in sea level. An alternative is that this feature records local incision into the sea bed by tides in deeper waters offshore. These alternatives are discussed further in the section on sequence stratigraphy below.

The high relief erosion surface that cuts from the top through the base of the upper Sego Sandstone is a lowstand incised valley (Figure 8) . At 25 meter deep and only a few kilometers wide, this valley is narrower than most ancient subsurface analogs (e.g., Van Wagoner et. al, 1990; Allen and Posamentier 1992; Zaitlin et al., 1994; Morton and Suter 1996). The depth of this incision indicates that adjacent areas were subaerially exposed during maximum sea-level lowstand. Although heterolithic deposits above the most deeply incised area along this incision may record estuarine intrusion into this valley, the dominance of sandy channel-form bed sets within the incision fill indicates that the river carried enough sediment to keep pace with sea level rise during valley filling. The size of the river that cut and filled this valley (indicated by the thickness of channel-form bed sets and the distance spanned by individual inclined beds within these sets) was on the same order of magnitude as the largest channels interpreted from deposits of the lower Sego Sandstone and within basal deposits of the overlying Neslen Formation (Kirchbaum and Hettinger 1998; Willis and Gabel 2001). The second location of incision, cuts down from the same horizon as the first, potentially reflecting the down cutting of a tributary that joined the main truck stream basinward. The size of the bed sets within this incision appear to be similar to those in the larger incision, however, this may alternatively indicate that the river avulsed during incision in areas upstream that remained unincised.

A soil is expected to have covered interfluves during lowstand valley incision, and thus the absence of a paleosol along the top of the upper Sego Sandstone suggests significant erosion under the cemented shell lag that occurs at this stratigraphic level. Such transgressive ravinement surfaces commonly cap prograding shoreline deposits. The depth of erosion along the base of the valley fill thus provides a minimum depth of river down cutting during sea level fall. Facies of the Nelsen Formation record deposition above this discontinuity, which may record a significant time gap.

SEQUENCE STRATIGRAPHY

Sequence stratigraphic interpretations of facies variations and allostratigraphic surfaces observed in this study predict larger-scale variations of depositional systems during formation of the upper Sego Sandstone. These models relate shifts in the position of deposition along a basin to changes in rates of sediment supply to accommodation development (Mitchum et. al, 1977). It is assumed that sediment will accumulate at the first location with available accommodation along a depositional profile. Deposition will thus shift landward as accommodation increases due to sea level rise, increases in basin subsidence rate, or the slowing of sediment input. Deposition will shift basinward as accommodation declines due to falls in sea level, or the slowing of sedimentation to subsidence rates. Key to such interpretations is recognition of allostratigraphic surfaces that record either times when sedimentation was sequestered landward and the study area was sediment starved, or times when accommodation in the study area was full and sediment was bypassed deeper into the basin.

Possible allostratigraphic (sequence stratigraphic) surfaces include: 1) maximum flooding surfaces formed when this area was transgressed and sediment starved; 2) regressive surfaces of erosion formed when falling sea level caused offshore marine areas to shoal enough that they could be reworked by marine currents; 3) lowstand surfaces of erosion formed as rivers incised to bypass sediments basinward of this area; and 4) initial flooding or ravinement surfaces formed as sea level rise lead shorelines to be transgressed (Figure 7). These surfaces define the base of the highstand, falling stage, lowstand, and transgressive system tracts (respectively). Sequence stratigraphic nomenclature is complicated because different investigators choose dissimilar surfaces to define sequences.

Standard “Exxon” sequence stratigraphic nomenclature define the lowstand surface of erosion to be the sequence boundary, because they were particularly interested in the record of sediment bypass that would produce additional reservoirs farther basinward. In ramp basin settings, like the Cretaceous Interior Seaway, rivers are less prone to incise deeply, bypassing sediments to a distinct lowstand wedge during sea level falls, because there is no distinct shelf break. All systems tracts can grade laterally within a continuous clastic wedge. Although Exxon stratigraphic nomenclature (Figure 9) requires differentiation of erosion surfaces formed during different stages of sea level change, it is difficult to distinguish falling stage, lowstand, and transgressive erosion in the Sego Sandstone. Sequences of a given order can develop bounding lowstand surfaces of erosion during lowstand parts of lower order sequences but not during highstand parts of the same sequences. As defined by the Exxon nomenclature, the base of the upper Sego Sandstone sequence would be somewhere within the lower Sego Sandstone, at a location defined only by interpretation (or very detailed regional 3D mapping), rather than outcrop observation. Furthermore, it becomes nearly impossible to differentiate lowstand and transgressive erosion where distal ends of deltaic deposits are significantly reworked by tidal currents during the start of transgression (Figure 9). Because different types of erosion surfaces associated with sandy intervals can be difficult to differentiate, the Sego Sandstone is more efficiently separated into stratigraphic units defined by marine maximum flooding surfaces, following Galloway (1989).

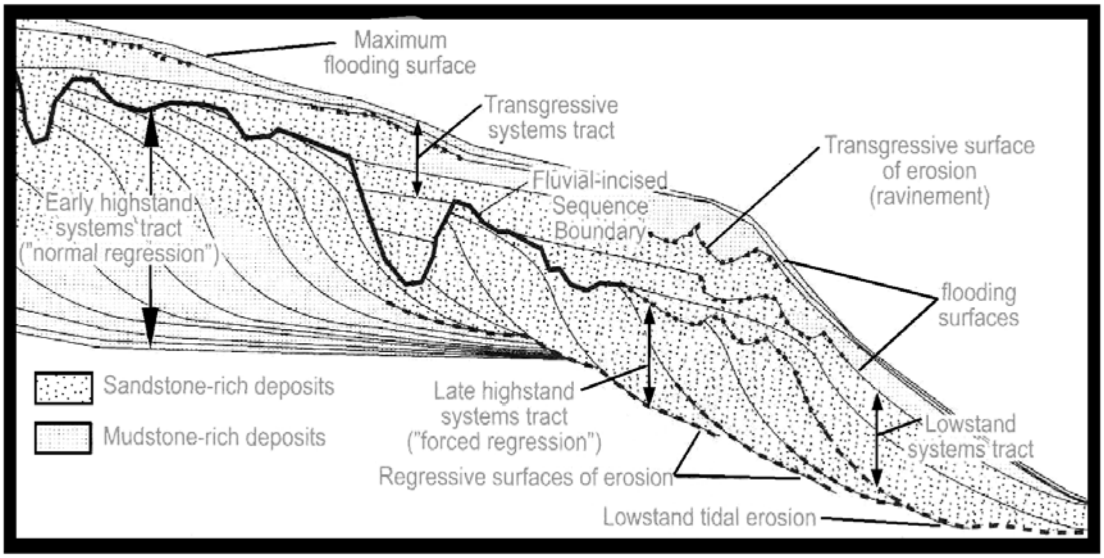

Figure 9. Sequence Stratigraphic Schematic Diagram. Exhibits highstand and lowstand deposition in combination with key stratal surfaces which define sequence designation (lowstand incisions), transgressive and regressive erosion, and marine flooding surfaces (Willis and Gabel 2001).

The first sequence stratigraphic interpretations of the Sego Sandstone inferred that all abrupt vertical transitions from offshore marine deposits to coarser-grained tide-influenced deposits occur across lowstand surfaces of erosion (Exxon "sequence boundaries"). The tidal deposits were thus interpreted to be late lowstand deposits reworked into incised valleys during sea level rise. Using this idea, the three sandy intervals observed in this study area (each with a basal erosion surface that records abrupt vertical transition from offshore marine shale to coarser-grained tidal sandstones) would each be interpreted to be lowstand surfaces of erosion overlain by late lowstand to transgressive system tract deposits. The fluvial deposits above the high-relief erosion surface that cuts through the upper Sego Sandstone would constitute a fourth sequence.

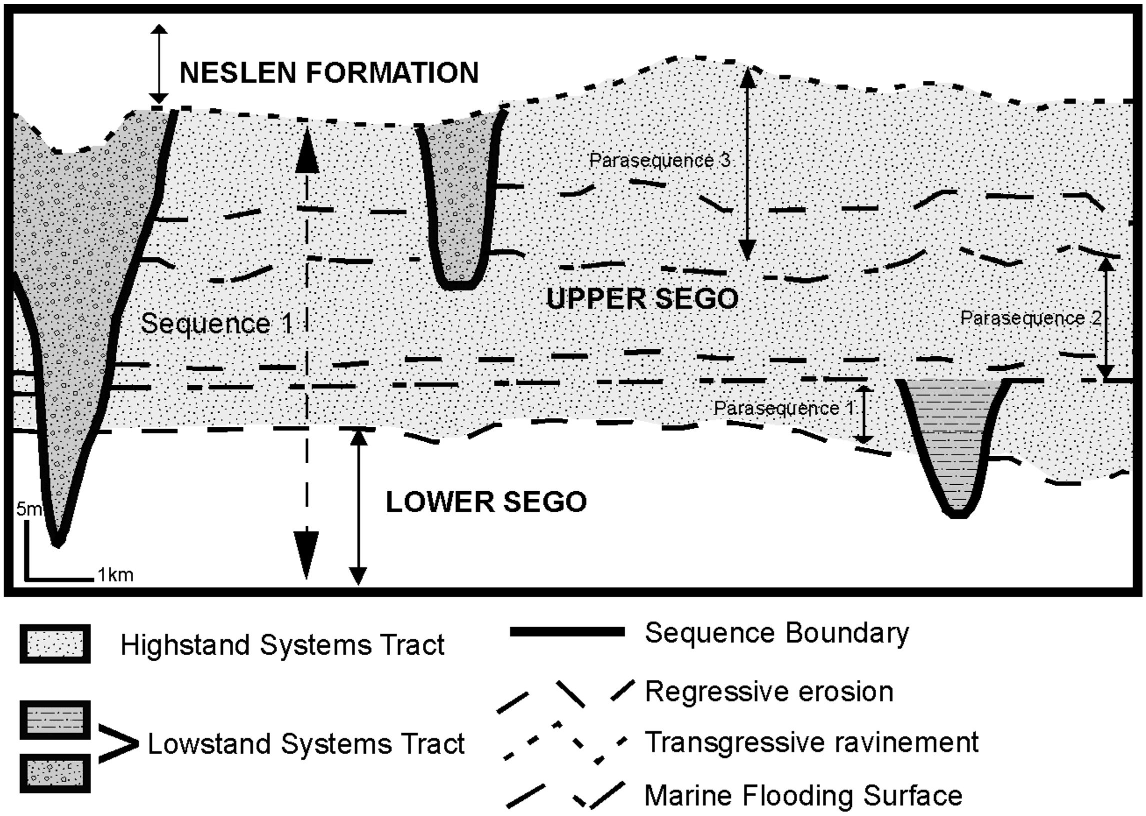

Figure 10. Exxon Sequence Stratigraphic Application. Regional cross-section exhibiting key stratal surfaces identified within the upper Sego.

A potential depositional model applicable to the Exxon sequence stratigraphic association is includes the shelf sand ridge facies (Figure 10). Shelf sand ridges are existent within the modern shelves due to the overall sea level rise prevalent within the Holocene. This produced a predominance of continental margins, which were covered by sediment left behind as shorelines transgressed. Transgressive conditions are more favorable for shelf sand ridges to occur since regressive areas lack the existence of loose sands, currents capable of moving that sand, and a preexisting irregular substrate (Sneeden and Dalrymple 1999). Shelf sand ridge morphology can show considerable variability due to the irregularity of coastal and/or shelf processes and the complex

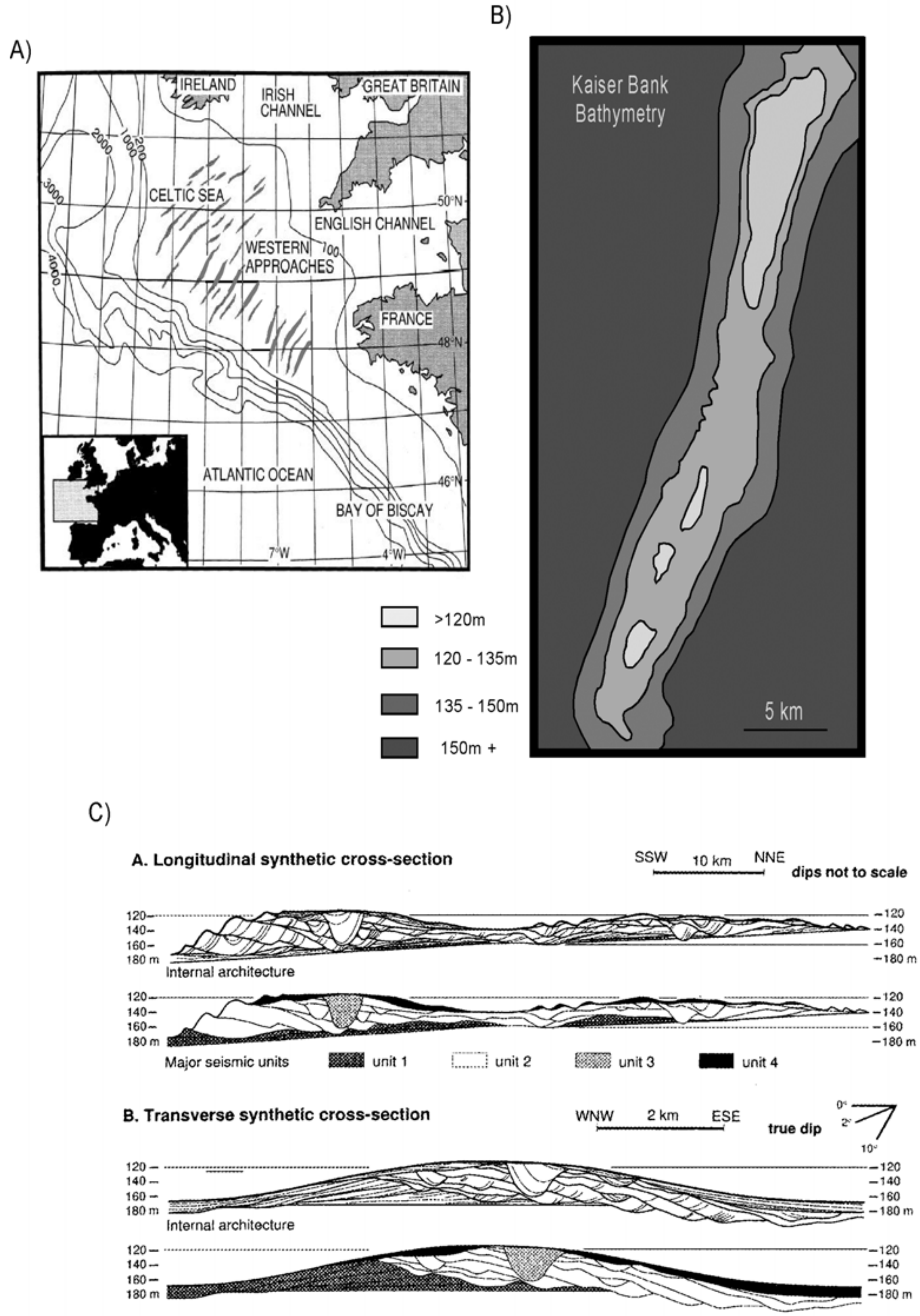

Figure 11. Architecture of the Deep Sand Banks of the Celtic Sea (From Reynaud et al., 1999a). (A) Basemap illustrating the elongate and perpendicular to shoreline sand geometries of the deep sand banks. (B) Bathymetric map of the Kaiser Bank (C) Schematic diagram illustrating the stratal geometries of interpreted within the tidal sand ridge of Kaiser Bank. Longitudinal and transverse cross sections are interpreted. Note the difference in scale.

evolution of the sand ridge in result to continued interaction with shelf and nearshore currents.

An analog for this interpretation of the shelf sand ridge models may lie within Holocene deposits on broad shelves formed during sea level lowstands. For example, shelf sand banks under the Celtic Sea analyzed by Reynaud and others (1999b) using high-resolution seismic data have erosional bases and relatively flat tops similar to the sandy intervals in the upper Sego Sandstone (Figure 11A). These sand banks are elongate perpendicular to lowstand paleoshorelines and were interpreted to have formed by the offshore tidal reworking of lowstand shoreline sediments when shallow seas enhanced tidal resonance (Figure 11B). Within these sand ridges 20 meter high bedsets were interpreted to be the deposits of marine dunes formed during relative sea level rise under 60 meter deep waters of the shelf (Figure 11C). High-relief erosion surfaces observed within ridge deposits where inferred to form as tidal currents cut and filled channels through the elongate ridges (Bridges, 1982). Flat tops of sands were interpreted to reflect an increased influence of waves on deposition as waters deepened and the basin departed from tidal resonance conditions.

Although the transgressive tidal deposits in the Celtic Sea have some parallels with those in the upper Sego Sandstone, there are several weaknesses when applying these deposits as analogs. There is no evidence that remnants of lowstand fluvial deposits are preserved in basal parts of sandy intervals within the upper Sego Sandstone, and basal erosion surfaces of these intervals are very flat, unlike what would be expected for fluvial lowstand incision. Bedsets within these intervals have very low-angle inclined internal bedding, much shallower than fronts of shelf marine dunes. Shelf ridge sands are expected to accrete laterally (normal) to paleocurrents, rather than dominantly parallel as interpreted for bed sets within the upper Sego Sandstone. As indicated in the interpretation of tidal bedsets, above, sands accumulated on offshore ridges are expected to be fairly clean because extensive reworking of these deposits should winnow muds from sands. In addition, shelf sandstones are generally reported to be highly marine bioturbated, rather then having relatively undisrupted depositional stratifications observed in upper Sego Sandstone tide-influenced facies. In contrast to Reynaud and others (1999b), Berne et. al (1998) suggested that Holocene ridges in the Celtic Sea might be dominantly erosional features formed by transgressive tidal erosion, and that bed sets within these deposits formed on shallow marine tidal bars on lowstand deltas or within estuaries during early stages of transgression. The interpretation that the erosion surface with the greatest relief, cut into the top of the upper Sego Sandstone, is filled with fluvial channel deposits, would suggest a very and dramatic short-term drop in sea level, far greater than simply the 30m of observed incision. It is difficult to define a mechanism for this scale of sea level fall, particularly during a “green house” period like the Cretaceous. Therefore, for these several reasons this suite of shelf sand analog application to the upper Sego is less applicable (Figure 11).

An alternative depositional model proposes that sandy intervals within the upper Sego Sandstone record autocyclic avulsion of delta lobes during regional shoreline progradation in a forced regressive (falling stage) setting. This specific depositional setting refers to the process of basinward migration of paleoshoreline in direct response to a combination of sediment influx and/or relative sea level fall. Forced regressive deposition is a more fit interpretation for the upper Sego through several parallels (observations) made in the field including: (1) the absence of deltaic top deposition, (2) the occurrence of basinward-dipping upper bounding surface at the top of the regressive succession (ravinment surface), (3) the presence of increasing grain size up section, and (4) and the occurrence of long distance regression (Posamentier and Morris 2000). These factors, common within the upper Sego, fit the depositional mold that Posamentier and others (1992) present supporting the stratal architecture of forced regressive deposits.

Individual sandy intervals within the upper Sego Sandstone would then be interpreted to form forward of a distributary channel. Abrupt fining and increase in marine bioturbation at sandy interval tops would then form after distributary avulsion. Erosion surfaces at the base of these intervals would reflect marine erosion of the prodelta. Many recent studies have proposed that similar marine erosion occurs forward of prograding shorelines when regional sea levels fall (Plint and Nummedal, 2000). In this scenario, erosion surfaces at the base of individual sandy intervals would not reflect allocyclic changes in sea level, but rather would simply record temporal variation in sediment supply along a regionally prograding shoreline. All three sandy intervals would thus be within the same falling stage (late highland) systems tract, capped by a single lowstand surface of erosion (Exxon sequence boundary), incised into the top of this progradational stack.

Distinguishing deposits of individual progradational lobes ("parasequences") from high frequency sequences can be difficult in this setting because individual distributaries can incise into associated mouth bars even as channel segments in areas landward remain unincised and continue to avulse. For example, modern analogs of tide-influenced, falling stage deltaic systems do not occur because of the early Holocene sea level rise. Quaternary examples of falling stage deltaic deposits along the Gulf of Mexico shelf may have some similarities, despite occurring along wave- rather than tide-dominated shorelines (Sydow et al. 1994; Edwards, 1994; Figure 12). In this system, delta lobes stack laterally across the shelf, rather than vertically as along the early highstand edge of the system. Earlier deposited lobes are incised by distributary channels that fed similar lobes basinward. Because all these deposits formed during falling sea level, distributaries may incise locally into delta fronts exposed by continuing sea level fall. Given this pattern of distributary incision and continued avulsion, Edwards (1994) suggested that lowstand surfaces of erosion (Exxon sequence boundaries) would be discontinuous across the shelf, making them a poor choice for the definition of mappable allostratigraphic units. Regressive surfaces of erosion would similarly be local, because some lobes may prograde into water deep enough to prodelta muds, while they prograde into other areas less deep than wave base.

This model and analog better explains the vertical thickening, coarsening, and increase in proximal shoreline facies within successive sandy intervals, specifically within the upper Sego Sandstone. The idea that all intervals formed on the same prograding shoreline also better explains the depth of valley incision, which is expected to be much greater than the knickpoint created by the delta front. Also, the fluvial fill of the valley would not require a sea level fall much greater than the observed depth of incisions (unlike the previous model) which would suggest less radial high-frequency sea level changes. This interpretation, furthermore, suggests that most deposits formed during shoreline regression and that deposits preserved during transgression were restricted to a late lowstand valley fill. Delta top deposits are seldom observed in falling stage deltaic deposits; they are thin initially because delta top accommodations are kept low by the basinward fall in the shoreline and are prone to subsequent transgressive marine erosion. Thick marine lags, like the accumulation of cemented shells capping the upper Sego Sandstone, are the record of transgressive marine erosion.

The two sequence stratigraphic models of transgressive shelf sand ridges and forced regressive wedges outlined above have different implications for facies likely to occur landward and seaward of this exposed cross section. Van Wagoner's approach, assuming sandy intervals are broad transgressive-filled valleys, implies a significant amount of bypass occurred during episodic regressions. Therefore, seaward of the study area a thick aggradational succession of lowstand deltaic sands should have been deposited during terminal regression and early stages of transgression. Landward of the study area would be sandy fluvial deposits, recording multiple periods of negative accommodation that should have favored preservation of channel deposits relative to overbank deposits and increased depositional slopes. Valley fills formed during multiple eposodic

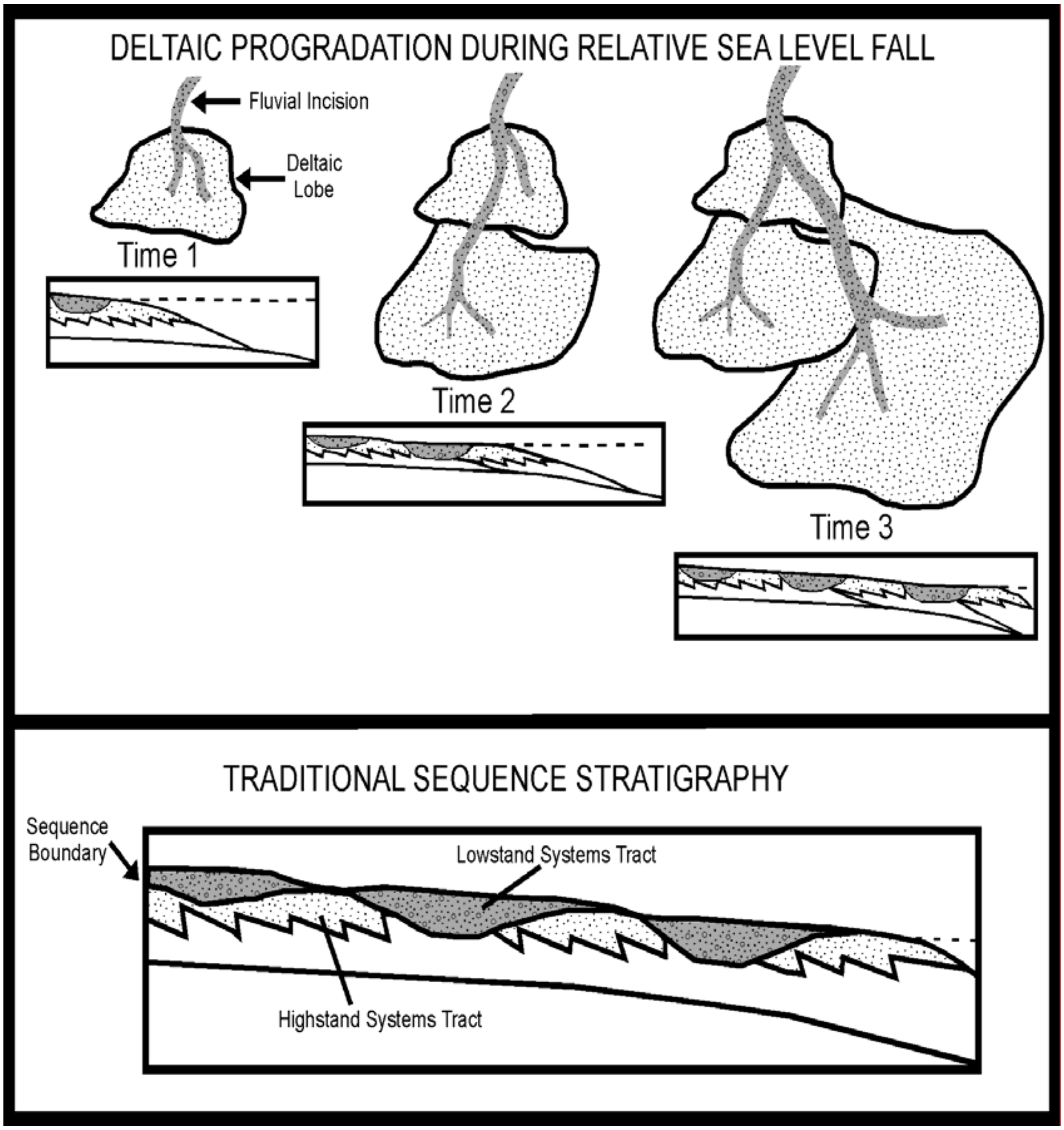

Figure 12. Schematic Diagram of Lateral Stacking of Deltaic Lobes. Note upstream avulsion of distributary channels is not taken in account for the traditional sequence stratigraphic model. (from Edwards 1994).

transgressions should be relatively common and in many cases may be compound (Zaitlin et al., 1994).

Alternatively, if the upper Sego Sandstone is a succession of falling stage deltaic lobes (forced regressive sands), each sandstone interval should gradually thin.and become more marine influenced over a few kilometers basinward. Seaward, lowstand deposition can be detached or attached dependant on the extent of a bypass zone seaward. In either case, bypass of sediment basinward would occur along certain portions within the basin, defined by locations of incised valleys. Landward, fluvial facies formed during initial stages of progradation may be muddier, and these deposits may gradually coarsen upward, reflecting the gradual transition to falling stage incision. Valley fills within these deposits are likely to be narrow and few in number.

CONCLUSIONS

The upper Sego Sandstone is a tide-influenced clastic wedge shed from the Sevier Orogeny obliquely into the Cretaceous Western Interior Seaway. It is underlain by the Anchor Mine Tongue, interpreted to be a maximum marine flooding surface, and overlain by coastal plain deposits of the Neslen Formation. Three sandy layers, increasing in thickness and grain size up-section, record an overall forward stepping deltaic progradation. Sandy layers are commonly capped by cemented ravinement surfaces. Shales formed during episodic transgressions separate the three progradational sandy intervals. Facies abruptly coarsen to channel deposits across a high-relief erosion surface interpreted to be a lowstand valley.

Two depositional models have been proposed for the Sego Sandstone including: (1) Sands supplied during episodic regressions that were completely reworked tidal bars into very broad lowstand incised valleys, and (2) falling stage, sharp-based prograding deltaic lobes that were cut by major valleys. The latter interpretation is favored because it better explains 1) vertical thickening and coarsening of successive sandy layers, 2) ebb-dominated paleo-flow indicators, 3) the nature of the erosion surfaces at the base of the sandy intervals, and 4) upward decrease in open marine styles of bioturbation. Like many other forced regressive wedges, delta top and prodeltaic facies are missing due to regressive erosion and subsequent transgressive ravinement. Interpretation that the upper Sego Sandstone records regional progradation contradicts the conventional assumption that tidal deposition is restricted to transgressive periods of siliciclastic wedge development.

REFERENCES CITED

Allen, G.P., and Chambers, J.L.C., 1998, Sedimentation in the modern and Miocene Mahakam Delta: Indonesian Petroleum Association, Jakarta, Indonesia, 236 p.

Allen, G.P., and H.W. Posamentier, 1992, On the origin of subsequent transgression modifications of incised valleys: American Association of Petroleum Geologists Annual Meeting Expanded Abstracts, v. 76, p. 2.

Baker, E.K., P.T. Harris, J.B. Keene, and S.A. Short, 1995, Patterns of sedimentation in the macrotidal Fly River delta, Papua New Guinea, *in* B.W. Flemming, and A. Bartholoma, eds., Tidal Signatures in Modern and Ancient Sediments: International Association of Sedimentologists, Special Publication 24, p. 193–211.

Berne', S., Gilles, L., Marsset, T., Bourillet, J.F., and M. de Batist, 1998, Erosional offshore sand ridges and lowstand shorefaces: examples from tide and wave-dominated environments of France: Journal of Sedimentary Research, v. 68, p.540-555.

Bhattacharya, J.P., 1993, The expression and interpretation of marine flooding surfaces and erosional surfaces in core: examples from the Upper Cretaceous Dunvegan Formation, Alberta foreland basin, *in* H.W. Posamentier, C.P. Summerhayes, B.U. Haq, and G.P.Allen, eds., Sequence Stratigraphy and Facies Associations: International Association of Sedimentologists, Special Publication 18, p. 125–160.

Bhattacharya, J.P., and B.J. Willis, 2001, Lowstand deltas in the Frontier Formation, Wyoming: implications for sequence stratigraphic models: American Association of Petroleum Geologists, Bulletin, v. 85, p. 261–294.

Bouma, A. H., J. M. Coleman, C. E. Stelting, and B. Kohl, 1989, Influence of relative sea level changes on construction of the Mississippi fan: Geo-Marine Letters, v. 9, p. 161–163.

Bridges, P.H., 1982, Ancient offshore tidal deposits, *in* A.H. Stride, ed., Offshore Tidal Sands: Processes and Deposits, Chapman & Hall, London, p. 172-192.

Coleman, J.M. , and L.D.Wright, 1975, Modern river deltas: variability of processes and sand bodies, *in* M.L. Broussard, ed., Deltas, models for exploration: Houston, Texas, Houston Geological Society, p.99–149.

Dalrymple, R.W., 1992, Tide depositional systems, *in* R.G. Walker, and N.P. James, eds., Facies models: response to sea-level change: Geological Association of Canada, St. John's, Newfoundland, Canada, p. 195-218.

Dalrymple, R.W., 1999, Tide-dominated deltas; do they exist or are they all estuaries. Official Program Abstracts. AAPG, 8, A29.

Dalrymple, R.W., E.K Baker, P.T Harris, and M.G. Hughes, 2003, Sedimentology and stratigraphy of a tide-dominated, foreland-basin delta (Fly River, Papua New Guinea): SEPM Special Publication, No. 76, p. 147-173.

Dalrymple, R.W., B.A. Zaitlin, and R. Boyd, 1992, Estuarine facies models: conceptual basis and stratigraphic implications: Journal of Sedimentary Petrology, v. 62, p. 1130-1146.

Edwards, M.B., 1994, Enhancing sandstone reservoir prediction by mapping erosion surfaces, lower Miocene deltas, southwest Louisiana, Gulf Coast Basin: Transactions of the Gulf Coast Association of Geological Societies, v. 10004, p. 207-212.

Elliot, T., 1986, Deltas and siliciclastic shorelines, *in* H.G. Reading ed., Sedimentary Environments and Facies: Oxford, England, Blackwell Scientific Publications, p. 113–188.

Galloway, W.E., 1989, Genetic stratigraphic sequences in basin analysis I: architecture and genesis of flooding-surface bounded depositional units: American Association of Petroleum Geologists Bulletin, v.73, p. 125-142.

Harris, P.T., 1988, Large-scale bed forms as indicators of mutually evasive sand transport and sequential infilling of wide-mouthed estuaries: Sedimentary Geology, v.57, p. 273-298.

Harris, P.T., C.B. Pattiaratchi, A.R. Cole, and J.B. Keene, 1992, Evolution of subtidal sandbanks in Moreton Bay, eastern Australia: Marine Geology, v. 103, p. 225-247.

Hori, K., Y. Aito, Z. Quanhong, and P. Wang, 2002, Architecture and evolution of the tide-dominated Changjiang (Yangtzee) River delta, China: Sedimentary Geology, v. 146, p. 249-264.

Howard, J.D., and H.E. Reineck, 1981, Depositional facies of high energy beach-to-offshore sequence, comparison with low energy sequence, American Association of Petroleum Geologists, Bulletin, v. 65, p. 807-823.

Johnson, M.A., and Belderson, R.H., 1969, The tidal origin of some vertical sedimentary changes in epicontinental seas: Journal of Geology, v. 77, p. 355–357.

Kirchbaum, M.A., and R.D. Hettinger, 1998, Stratigraphy and depositional environments of the Late Cretaceous coal-bearing Neslen/Mount Garfield Formations, eastern Book Cliffs, Utah and Colorado: U.S. Geological Survey, OF 98-43, p. 98-43.

Kuehl, S.A., T.M. Harris, and W.S. Moore, 1989, Shelf sedimentation off the Ganges-Brahmaputra river system: evidence for sediment bypassing to the Bengal fan: Geology, v. 17, p. 1132-1133.

Kuehl, S.A., B.M Levy, W.S Moore, and M.A. Allison, 1997, Subaqueous delta of the Ganges–Brahmaputra river system: Marine Geology, v. 144, p. 81–96.

McCrimmon, G.G. 1996. Sedimentology and sequence stratigraphy of the Lower Cretaceous Clearwater Formation, Cold Lake, Alberta. Unpublished M.Sc. thesis, University of Ottawa, Ottawa, Ontario, 281 p.

McCrimmon, G.G., and R.W.C. Arnott, 2002, The Clearwater Formation, Cold Lake, Alberta: a worldclass hydrocarbon reservoir hosted in a complex succession of tide-dominated deposits: Bulletin of Canadian Petroleum Geology, v. 50, p. 370-392.

McLauren, B.T., and R.J Steel, 2000, Fourth-order nonmarine to marine sequences, middle Castlegate Formation, Book Cliffs, Utah: Geology, v. 28, p. 359-362.

Mellere, D., and R.J. Steel, 1996, Tidal sedimentation in Inner Hebrides half grabens, Scotland: the Mid-Jurassic Bearreraig Sandstone Formation, *in* M. De Batist, and P. Jacobs, eds., Geology of Siliciclastic Shelf Seas: Geological Society of London, Special Publication 117, p. 59–79.

Mitchum, R. M., P. R. Vail, and S. Thompson, 1977, Seismic stratigraphy and global changes of sea level, part 2: the depositional sequence as a basic unit for stratigraphic analysis, *in* C. E. Payton, ed., Seismic stratigraphy—applications to hydrocarbon exploration: AAPG Memoir 26, p. 53–62.

Morton, R.A., and J.R. Suter, 1996, Sequence stratigraphy and composition of Late Quaternary shelf-margin deltas, northern Gulf of Mexico: American Association of Petroleum Geologists, Bulletin, v. 80, p. 510-530.

Nio, S.D. and C.S. Yang, 1991, Diagnostic attributes of clastic tidal deposits: a review. *in* D.G. Smith, G.E. Reinson, B.A. Zaitlin and R.A. Rahmani, eds., Classic Tidal Sedimentology. Canadian Society of Petroleum Geologists, Memoir 16, p. 15-27.

Pemberton, S.G., and R.W. Frey, 1984, Ichnology of storm-influenced shallow marine sequence: Cardium Formation (Upper Cretaceous) at Seebe, Alberta, *in* D.F. Stott and D.J. Glass, eds., The Mesozoic of Middle North America. Canadian Society of Petroleum Geologists, Memoir 9, p. 281-304.

Pemberton, S.G., J.A. MacEachern, and R.W. Frey 1992, Trace fossil facies models: environmental and allostratigraphic significance, *in* R.G. Walker and N.P. James, eds., Facies Models, Response to Sea Level Change. Geological Association of Canada, p. 47-52.

Plint, A.G., 1988, Sharp-based shoreface sequences and "offshore bars" in the Cardium Formation of Alberta: their relationship to relative changes in sea level, *in* C.K. Wilgus, B.S. Hastings, C.G. Kendall, H.W. Posamentier, C.A. Ross, and J.C.Van Wagoner, eds., Sea-level changes: an integrated approach: Society for Sedimentary Geology Special Publication 42, p. 357–370.

Plint, A.G., and D. Nummedal, 2000, The falling stage systems tract: recognition and importance in sequence stratigraphic analysis, *in* D. Hunt, and R.L. Gawthorpe, eds., Sedimentary Responses to Forced Regressions, Geological Society Special Paper 172, p. 1-18.

Posamentier, H.W., and W.R. Morris, 2000, Aspects of the stratal architecture of forced regression deposits, *in* D. Hunt, and R.L. Gawthorpe, eds., Sedimentary Responses to Forced Regressions, Geological Society of London, Special Paper 172, p. 19–46.

Posamentier, H.W., G.P. Allen, D.P. James, and M. Tesson, 1992, Forced regressions in a sequence stratigraphic framework: concepts, examples, and exploration significance: American Association of Petroleum Geologists Bulletin, v. 76, p. 1689-1694.

Posamentier, H. W., and P. R. Vail, 1988, Eustatic controls on clastic deposition sequence and systems tract models, *in* C. K. Wilgus, B. S. Hastings, C. G. St. C. Kendall, H. W., Posamentier, C. A. Ross, and J. C. Van Wagoner, eds., Sea-level changes: an integrated approach: SEPM Special Publication 42, p. 125–154.

Reynaud, J.Y., B. Tessier, J.N. Proust, R. Dalrymple, J.F. Bourillet, M. de Batist, G. Lericolais, S. Berne', and T. Marsset, 1999a, Architecture and sequence stratigraphy of a late Neogene incised valley at the shelf margin, southern Celtic Sea: Journal of Sedimentary Research, v. 69, p. 351-364.

Reynaud, J.Y., B. Tessier, J.N. Proust, R. Dalrymple, J.F. Bourillet, M. de Batist, G. Lericolais, S. Berne', and T. Marsset, 1999b, Eustatic and hydrodynamic controls on the architecture of a deep sand bank (Celtic Sea): Sedimentology, v. 46, p. 703-715.

Roberts, L.N.R., and M.A. Kirschbaum, 1995, Paleogeography of the late Cretaceous of the Western Interior of Middle North America—Coal Distribution and Sediment Accumulation: U.S. Geological Survey, Professional Paper 1561, 115p.

Snedden, J.W. , and R.W. Dalrymple, 1999, Modern shelf sand ridges: historical review of modern examples and a unified model of ridge origin and evolution, *in* K.M. Bergman and J.W. Snedden, eds., Isolated shallow marine sand bodies: sequence stratigraphic analysis and sedimentologic interpretation: SEPM Special Publication 64, p.13–28.

Stride, A.H., R.H. Belderson, N.H. Kenyon, and, M.A. Johnson, 1982, Offshore tidal deposits: sand sheet and sand bank facies, *in* A.H., Stride, ed., Offshore Tidal Sands: Processes and Deposits: London, U.K., Chapman & Hall, p. 95–126.

Ta, T.K.O., V.L. Nguyen, M. Tateishi, I. Kobayashi, Y. Saito, and T. Nakamura, 2002, Sediment facies and Late Holocene progradation of the Mekong River Delta in Bentre Province, southern Vietnam: an example of evolution from tide-dominated to tide- and wave-dominated delta: Sedimentary Geology, v. 152, p. 313-325.

Taylor, A.M., and R.L. Gawthorpe, 1993, Application of sequence stratigraphy and trace fossil analysis to reservoir description: examples from the Jurassic of the North Sea, *in* J.R. Parker, ed., Petroleum Geology of Northwest Europe, Proceedings of the 4th Conference: The Geological Society, London, v. 1, p. 317–336.

Thomas, M. A., and J. B. Anderson, 1989, Glacial eustatic controls on seismic sequences and parasequences of the Trinity/Sabine incised valley, Texas continental shelf: Transactions of the Gulf Coast Association of Geological Societies, v. 39, p. 563–570.

Van Wagoner, J.C., 1991, High frequency sequence stratigraphy and facies architecture of the Sego Sanstone in the Book Cliffs of western Colorado and eastern Utah, *in* J.C. Van Wagoner, D. Nummedal, C.R. Jones, D.R. Taylor, D.C. Jeanette, and G.W. Riley, eds., Sequence Stratigraphy Applications to Shelf Sandstone Reservoirs: Outcrop to Subsurface Examples: Tulsa, Oklahoma, American Association of Petroleum Geologists, Field Conference, p. 1-22.

Van Wagoner, J.C., 1995, Sequence stratigraphy and marine to nonmarine facies architecture of foreland basin strata, Book Cliffs, Utah, U.S.A., *in* J.C. Van Wagoner, and G.T. Bertram, eds., Sequence Stratigraphy of Foreland Basins: Outcrop and Subsurface Examples from the Cretaceous of North America: American Association of Petroleum Geologists, Memoir 64, p. 137–223.

Van Wagoner, J.C., R.M. Mitchum, K.M. Campion, and V.D. Rahmanian, 1990, Siliciclastic sequence stratigraphy in well logs, cores and outcrops: Concepts for high-resolution correlation of time and facies: American Association of Petroleum Geologists, Methods in Exploration Series 7, 55 p.

Walker, R.G., 1992, Perspective, facies modelling and sequence stratigraphy: Journal of Sedimentary Petrology, v.60, p.777–786.

Walker, R.G. and A.G. Plint, 1992, Wave- and storm-dominated shallow marine systems, *in* R.G. Walker, and N.P. James, eds., Facies Models: Response to Sea Level Change: St. Johns, Newfoundland, Geological Association of Canada, p. 219–238.

Willis, A., 2000, Tectonic control of nested sequence architecture in the Sego Sandstone, Neslen Formation and Upper Castlegate Sandstone (Upper Cretaceous), Sevier Foreland Basin, Utah, USA: Sedimentary Geology, v. 136, 277-317.

Willis B.J., J.P. Bhattacharya, S.L. Gabel, and C.D. White, 1999, Architecture of a tide-influenced delta in the Frontier Formation of central Wyoming: Sedimentology, v. 46, p. 667-688.

Willis B.J., and S.L. Gabel, 2001, Sharp-based, tide-dominated deltas of the Sego Sandstone, Book Cliffs, Utah, USA: Sedimentology, v. 48, p. 479-506.

Willis B.J., and S.L. Gabel, 2003, Formation of deep incisions into tide-dominated river deltas: Implications for the stratigraphy of the Sego Sandstone, Book Cliffs, Utah, U.S.A.: Journal of Sedimentary Research, v. 73, p. 246-263.

Yoshida, S., A. Willis, and A.D. Miall, 2001, Fourth-order nonmarine to marine sequences, middle Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Utah: Journal of Sedimentary Research, v.66, p. 737-748.

Young, R.G., 1955, Sedimentary facies and intertonguing in the Book Cliffs, Utah–Colorado: Geological Society of America, Bulletin, v. 55, p. 177–202.

Zaitlin, B.A., R.W. Dalrymple, and R. Boyd, 1994, The stratigraphic organization of incised- valley systems associated with relative sea-level change, *in* R.W. Dalyrymple, R. Boyd, and B.A. Zaitlin, eds., Incised-Valley Systems: Origin and Sedimentary Sequences, SEPM, Special Publication 51, p. 45-62.

APPENDIX

(Plates are in separate pdf. files)

Plate 1 – Bedding diagram overlain by 20 measured sections within the Horse – Crescent Canyon thesis area.

Plate 2 – Photomosaic of westernmost valley incision comprised of amalgamated lowstand fluvial fill.

Plate 3 – Photomosaic of easternmost valley incision.

VITA

Eric D. Robinson received his Bachelor of Science degree in geology from Morehead State University, eastern Kentucky, in May of 2003. He began his graduate studies in geology at Texas A&M University in September of 2003, and received his Master of Science degree in December of 2005. His research interests include field application of sequence stratigraphy to regional clastic wedges in the Book Cliff Mountains of Utah in order to improve subsurface hydrocarbon reservoir predictability.

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