Sequence stratigraphy of the Dakota Formation (Cenomanian), southern Utah: interplay of eustasy and tectonics in a foreland basin

DAVID ULIČNÝ
Department of Geology, Charles University, Albertov 6, 128 43 Prague 2, Czech Republic (E-mail: ulicny@mail.natur.cuni.cz)

ABSTRACT
The Dakota Formation in southern Utah (Kaiparowits Plateau region) is a succession of fluvial through shallow-marine facies formed during the initial phase of filling of the Cretaceous foreland basin of the Sevier orogen. It records a number of relative sea-level fluctuations of different frequency and magnitude, controlled by both tectonic and eustatic processes during the Early to Late Cenomanian. The Dakota Formation is divided into eight units separated by regionally correlatable surfaces that formed in response to relative sea-level fluctuations. Units 1–6B represent, from bottom to top, valley-filling deposits of braided streams (unit 1), alluvial plain with anastomosed to meandering streams (2), tide-influenced fluvial and tide-dominated estuarine systems (3A and 3B), offshore to wave-dominated shoreface (4, 5 and 6A) and an estuarine incised valley fill (6A and 6B). The onset of flexural subsidence and deposition in the foredeep was preceded by eastward tilting of the basement strata, probably caused by forebulge migration during the Early Cretaceous, which resulted in the incision of a westward-deepening predepositional relief. The basal fluvial deposits of the Dakota Formation, filling the relief, reflect the onset of flexural subsidence and, possibly, a eustatic sea-level rise. Throughout the deposition of the Dakota Formation, flexure controlled the long-term, regional subsidence rate. Locally, reactivation of basement faults caused additional subsidence or minor uplift. Owing to a generally low subsidence rate, differential compaction locally influenced the degree of preservation of the Dakota units. Eustasy is believed to have been the main control on the high-frequency relative sea-level changes recorded in the Dakota. All surfaces that separate individual Dakota units are flooding surfaces, most of which are superimposed on sequence boundaries. Therefore, with the exception of unit 6B and, possibly, 3B, most of the Dakota units are interpreted as depositional sequences. Fluvial strata of units 1 and 2 are interpreted as low-frequency sequences; the coal zones at the base and within unit 2 may represent a response to higher frequency flooding events. Units 3A to 6B are interpreted as having formed in response to high-frequency relative sea-level fluctuations. Shallow-marine units 4, 5 and 6A, interpreted as parasequences by earlier authors, can be divided into facies-based systems tracts and show signs of subaerial exposure at their boundaries, which allows interpretation as high-frequency sequences. In general, the Dakota units are good examples of high-frequency sequences that can be misinterpreted as parasequences, especially in distal facies or in places where signs of subaerial erosion are missing or have been removed by subsequent transgressive erosion. Both low- and high-frequency sequences represented by the Dakota units are stacked in an overall retrogradational pattern, which reflects a long-term relative sea-level rise, punctuated by brief periods of relative sea-level fall. There is a relatively
major fall near the end of the *M. mosbyense* Zone, whereas the base of the Tropic shale is characterized by a major flooding event at the base of the *S. gracile* Zone. A similar record of Cenomanian relative sea-level change in other regions, e.g. Europe or northern Africa, suggests that both high- and low-frequency relative sea-level changes were governed by eustasy. The high-frequency relative sea-level fluctuations of ~100 kyr periodicity and ~10–20 m magnitude, similar to those recorded in other Cenomanian successions in North America and Central Europe, were probably related to Milankovitch-band, climate-driven eustasy. Either minor glacioeustatic fluctuations, superimposed on the overall greenhouse climate of the mid-Cretaceous, or mechanisms, such as the fluctuations in groundwater volume on continents or thermal expansion and contraction of sea water, could have controlled the high-frequency eustatic fluctuations.

**Keywords** Cenomanian, foreland basin, sea-level, sequence, Western Interior.

**INTRODUCTION**

The sedimentary infill of a foreland basin primarily represents a record of orogen tectonics, flexural subsidence of the foreland, rates of weathering and sedimentation and eustatic sea-level change. Field studies as well as numerical models of relative sea-level changes in foreland basins focus primarily on the relative roles of tectonic subsidence (or uplift) and eustasy in controlling the accommodation and formation of stratal packages and unconformities. Whereas long-term cycles of relative sea-level change (2–10 Myr) are generally thought to be controlled by accretionary wedge growth and long-term erosion rates in the orogen, short-term sea-level fluctuations (0.01–0.5 Myr) are generally attributed to high-frequency eustasy, thrust events or changes in intraplate stress levels (Peper, 1993). The driving mechanisms of the third- to fifth-order relative sea-level fluctuations (*sensu* Vail *et al.*, 1991) are a subject of ongoing controversy in sequence stratigraphy (cf. Christie-Blick & Driscoll, 1995).

The Dakota Formation in southern Utah is a well-exposed succession of fluvial and shallow-marine sediments formed during the initial phase of filling of the Cretaceous foreland basin of the Sevier orogen. This paper focuses on the sequence-stratigraphic aspects of the Dakota Formation in the Kaiparowits Plateau region (including the Paria River Amphitheatre, Fig. 1). Compared with the westernmost part of the foreland basin fill, adjacent to the orogenic front, this region was characterized by generally low rates of subsidence, and the depositional systems here were very sensitive to even small-scale relative sea-level changes (Shanley & McCabe, 1995).

The facies associations and their bounding surfaces are either physically traceable or correlatable throughout the region and offer an opportunity to study in detail the record of relative sea-level changes. The Dakota Formation in the Kaiparowits Plateau region has been studied most recently by Gustason (1989), am Ende *et al.* (1991) and Elder *et al.* (1994). This paper originated as part of a comparative study of the record of Middle and Late Cenomanian sea-level fluctuations in the US Western Interior and central Europe (preliminary results in Uličný, 1994).

**TECTONIC AND PALAEOGEOGRAPHIC SETTING**

During deposition of the Dakota Formation, the Kaiparowits region lay at the western margin of the Western Interior Seaway, 140–200 km east of the thrust front of the Sevier orogenic belt (Fig. 1a and b). The Western Interior foreland basin formed through flexural subsidence resulting from loading of the cratonic lithosphere of North America by low-angle thrust sheets of the Sevier orogen (e.g. Armstrong, 1968; Cross, 1986). Owing to the high flexural rigidity of the thermally mature western North American continental lithosphere, the foreland basin was relatively shallow and broad (Angevine & Heller, 1987; Bally, 1989). In addition to flexural subsidence, however, sedimentation in the basin was affected by normal faulting caused by load-induced reactivation of basement fault zones inherited from the Precambrian rifting and fragmentation of the North American continent (Picha, 1986; Schwans, 1995).
Clastic material of the Dakota Formation in southern Utah was shed generally south-eastwards from the NE- to SW-trending part of the Sevier belt (Peterson, 1969; Gustason, 1989; Fig. 1). Wolfe (1989) showed that, near the Arizona–New Mexico border, the sediment source during the mid-Cenomanian was from the Mogollon Highlands, a NW- to SE-trending uplift of possible rift-flank origin (Bilodeau, 1986; Gustason, 1989). The V-shaped intersection of the Sevier thrust belt and the Mogollon Highlands formed a topographic control of the ‘Grand Canyon Bight’ (Stokes & Heylum, 1963), an embayment formed during the mid-Cretaceous marine flooding of the Western Interior basin (Fig. 1; cf. Gustason, 1989; Elder & Kirkland, 1993). The Kaiparowits region was located at the north-western margin of this embayment, episodically flooded and subaerially exposed during the Late Cretaceous.

**STRATIGRAPHY AND NEW SUBDIVISION OF THE DAKOTA FORMATION**

The Dakota Formation in the Kaiparowits region rests with an angular unconformity on Jurassic
rocks of the Entrada and Morrison formations, which are gently inclined to the east (Gustason, 1989, his Fig. 2). Peterson (1969) divided the Dakota Formation into three informal members: the conglomeratic 'lower member' of fluvial origin; the 'middle member', also fluvial; and the 'upper member' formed by shallow-marine sandstones of the Dakota Formation interfingering with tongues of the offshore marine Tropic Shale. Details of the history of stratigraphic terminology and research are included in Gustason (1989) and am Ende et al. (1991). Correlation to ammonite stratigraphy is shown in Table 1.

The 'lower member' of the Dakota Formation (unit 1 in this paper) yielded no fossils that would allow reliable dating. Peterson (1969) assigned the lower member to the lower Cenomanian, whereas Gustason (1989) suggested an Albian age. Elder & Kirkland (1993) proposed early to middle Cenomanian age for the 'lower member'. The 'middle member' was dated by am Ende et al. (1991) as middle to late Cenomanian in age. The 'upper member' (units 4 to 6B here) belongs to the *Metoicoceras mosbyense* Zone. The top of the Dakota Formation in the Kaiparowits region is a major flooding surface, overlain by distal offshore

---

**Fig. 2.** Correlation of units and their bounding surfaces of the Dakota Formation across the study area (location of the measured sections shown in Fig. 1c). The sequence-stratigraphic interpretation includes only the main bounding surfaces; details of subdivisions of individual units are discussed in text. Tectonic elements interpreted as syndepositionally active normal faults are indicated by bars at the top of the cross-section; note the incremental westward increase in thickness across most of them. The shift in datum from the base of the main body of the Tropic Shale in the west to the base of unit 4 in the east is arbitrary and meant to prevent artificial westward plunge of bounding surfaces, which would result from the choice of one ravinement surface (base of the Tropic) as a datum for the whole region. Note the influence of differential compaction on the preservation of some units (especially above channels in units 3A and 3B). Section 5 is modified from Gustason (1989).

---

© 1999 International Association of Sedimentologists, *Sedimentology, 46, 807–836*
shales of the Tropic Formation, belonging to the lower part of the *Sciponoceras gracile* ammonite zone (Gustason, 1989; Leithold, 1994). The strictly lithostratigraphic concept, which requires the offshore mudstones of units 4 and 5 within the ‘upper member’ to be termed ‘tongues of the Tropic Shale’, is not followed here: in the limited study area, these units are treated as subunits of the Dakota Formation.

The sequence-stratigraphic study of the Dakota Formation presented here allowed a more detailed subdivision of the Dakota rocks than that provided by Peterson (1969), Gustason (1989) and Elder et al. (1994). In this paper, the Dakota Formation is divided into six units (Table 1). The basis for the subdivision of the formation was the recognition of major, regionally extensive bounding surfaces (flooding surfaces, sequence boundaries), which reflect the response of depositional systems to base level changes. With the exception of the valley-filling unit 1, all units are present throughout the study region and can be either physically traced, or, where the exposure is discontinuous, correlated between the measured sections presented here.

**METHODS**

The Dakota Formation was studied in a number of vertical measured sections between the Paria Amphitheatre and the north-western margin of Lake Powell. The sedimentological observations included description of lithologies, sedimentary and biogenic structures, palaeocurrent measurements and fossils. Locally, the measured sections were supplemented by a study of architectures of particular sedimentary bodies. Attention was paid to the recognition of major, regionally extensive, bounding surfaces formed in response to base level/sea-level changes. The NW–SE correlation line (Fig. 2), connecting most of the
studied sections, provided a two-dimensional control of thickness and lithofacies variation. The sequence-stratigraphic terminology used in the paper follows the ‘Exxon’ concepts as presented by Van Wagoner et al. (1988, 1990), with the exception of the parasequence concept, as discussed in the text. The division of a foreland basin system into particular depozones follows the terminology of DeCelles & Giles (1996).

### DESCRIPTION AND INTERPRETATION OF UNITS 1–6

#### Unit 1

**Description**

Unit 1 infills the relief of the basal Cretaceous unconformity in southern Utah. Peterson (1969) found that the palaeovalleys underlying unit 1 are oriented uniformly towards the south-east. This was confirmed by a detailed palaeocurrent analysis (Gustason, 1989). In spite of the discontinuous occurrence of this unit, a systematic westward increase in thickness of the valley fills is observed in the study region and further west. The maximum thickness in the study region is 11 m in the westernmost sections (Sections 2 and 3 in Fig. 2). Gustason (1989) described a 50-m-thick, 8- to 10-km-wide valley fill in the south-western Markagunt Plateau region, 40 km west of the western margin of the study region.

Unit 1 comprises interbedded coarse-grained sandstone and conglomerate. The most common lithology is cross-bedded, moderately to poorly sorted, coarse-grained, pebbly sandstone. The cross-set thickness ranges between 0.1 and 3 m; trough cross-bedding generally prevails over planar cross-bedding. Conglomerates dominate in the lower part of unit 1 and range from poorly sorted, massively bedded, cobble conglomerates to trough and planar cross-bedded granule conglomerates. Thick sets of inclined stratification, where well-preserved, fine upward from conglomerate-dominated ‘toesets’ to medium-grained sandstone in the ‘topset’ strata (Fig. 3a). Clasts are occasionally up to 10 cm in diameter, especially in the lowermost parts of the unit. Grey to

---

### Table 1. Summary of the Dakota Formation units defined in this paper, with interpreted depositional environments, chronostratigraphic correlation and a comparison with earlier subdivisions of the Dakota Formation in southern Utah.

<table>
<thead>
<tr>
<th>Unit no.</th>
<th>Main depositional environment</th>
<th>Subdivisions of earlier authors</th>
<th>Biozones; Ar/Ar dates (Myr); chronostratigraphy</th>
</tr>
</thead>
<tbody>
<tr>
<td>6B</td>
<td>Estuarine–upper shoreface</td>
<td>Upper member (Peterson, 1969; Gustason, 1989) included in Parasequence 2 (Elder et al., 1994)</td>
<td>Metoicoceras mosbyense zone 93.9 ± 0.72&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>6A</td>
<td>Estuarine; offshore to shoreface</td>
<td>Upper member (Peterson, 1969; Gustason, 1989)</td>
<td>LATE (part)</td>
</tr>
<tr>
<td>5</td>
<td>Offshore to shoreface</td>
<td>Upper member (Peterson, 1969; Gustason, 1989) Parasequence 2 (Elder et al., 1994)</td>
<td>Equivalent to Calycoceras cantaitaurinum zone ~94.3&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>4</td>
<td>Offshore to shoreface</td>
<td>Upper member (Peterson, 1969; Gustason, 1989) Parasequence 1 (Elder et al., 1994)</td>
<td>~94.5&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>3B</td>
<td>Estuarine, tide-dominated</td>
<td>Middle member (Peterson, 1969; Gustason, 1989) (upper part)</td>
<td>~95.8&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>3A</td>
<td>Tide-influenced, fluvial</td>
<td>Middle member (Peterson, 1969; Gustason, 1989) (upper part)</td>
<td>?</td>
</tr>
</tbody>
</table>
dark-grey, carbonaceous and also greenish to yellowish mudstone interbeds occur rarely throughout unit 1 and are preserved as erosional relicts. Individual mudstone beds, up to 20 cm thick, have a limited lateral extent, generally less than 5 m. Mudstone intraclasts occur in the
conglomerates and, locally, root traces occur in mudstone and sandstone beds. Palaeocurrent indicators show a south-eastward mean flow direction, in accordance with previous observations (Peterson, 1969; Gustason, 1989).

The fossiliferous limestone and chert clasts in the conglomerates indicate derivation from Upper Palaeozoic rocks then exposed in Sevier thrusts in western Utah and Nevada (Gustason, 1989). Silicified wood clasts occur commonly as fragments several centimetres long but, exceptionally, logs several metres long and about 1 m in diameter have been found. They probably come from the weathering residue of the Jurassic Morrison Formation. Clasts of cross-bedded, fine-grained sandstone were reworked directly from the underlying Entrada Formation.

Unit 1 generally fines upwards. Thin beds of poorly sorted conglomerate form unit 1 outside the major valley fills. Gustason (1989) reported a widespread, poorly sorted, matrix-supported conglomerate or conglomeratic pavement occurring in the topmost parts of unit 1 and interpreted it as a product of pedimentation and formation of an unconformity at the top of unit 1. This lithology was recognized in some, although not all, sections in the present study (e.g. Sections 3 and 5 in Fig. 2; Fig. 3b). However, in Sections 6 and 7 (Fig. 2), up to 3.5 m of erosional relief occurs on top of unit 1. This relief is filled by whitish, carbonate-cemented, fine- to medium-grained sandstone.

Environmental interpretation

On the basis of the lithofacies and sedimentary geometries, unit 1 is interpreted as having been deposited by braided streams characterized by highly fluctuating discharge, leading to rapid shifting of channels. Gravel and sand bedforms migrated in the channels, some forming transverse bars (downstream accretion mesoforms) and linguoid or compound bars (lateral accretion mesoforms sensu Miall, 1985). Rapid channel migration permitted only limited accumulation of mudstone in abandoned channels or scour pits. This interpretation is in agreement with Gustason (1989), am Ende et al. (1991) and Kirschbaum & McCabe (1992). The braided streams of unit 1 transported sediment from the orogenic front in a south-eastwards direction. Peterson (1969) showed that the valley-fill trends coincide with the orientation of present-day structural features (Fig. 1c), indicating that basement fractures controlled the palaeovalley trends.

Unit 2

Description

Unit 2 is a succession dominated by mudstones, sandstones and coals. It corresponds to the lower and middle portions of the ‘middle member’ of the Dakota Formation (Table 1; the ‘lower coal zone’ and a part of the ‘middle alluvial interval’ of Gustason, 1989). It is either underlain by unit 1, commonly also separated by a coal zone (coal zone 1 of Kirschbaum & McCabe, 1992) or rests directly on Jurassic bedrock (Figs 3d and 4). The top of unit 2 is an erosional surface underlying tide-influenced deposits of units 3A and 3B. In the study region, unit 2 shows a distinct westward increase in thickness from =10 m in the east to =30 m in the west (Fig. 2).

Unit 2 is dominated by mudstones, mostly massive, pale grey to dark grey, locally carbonaceous. Pedogenic features such as rooted zones and slickensides occur commonly throughout unit 2. Sharp-based, thin, sheet-like sandstone beds occur encased in the mudstones. These sandstones commonly fine upwards into siltstones and normally show small-scale cross-bedding or ripple cross-lamination. Siderite concretions are found locally within the mudstones.

Isolated channel fills commonly display basal intraformational conglomerates, consisting mostly of mudstone and coal clasts and vertebrate bone fragments. The channel-fill lithology is dominated by well-sorted, fine-grained, trough cross-bedded sandstones, which may grade upwards into ripple cross-laminated sandstones with mudstone interbeds. Mudstone channel fills are relatively rare. Climbing ripples are common in the finer grained sandstone facies. Channels immediately overlying the basal coal zone in the easternmost sections (Fig. 3e) are filled with inclined heterolithic strata characterized by flaser–lenticular bedding. Commonly, channel fills in unit 2 show a high degree of synsedimentary deformation, which includes recumbent cross-beds and disharmonic folding (cf. Gustason, 1989; Kirschbaum & McCabe, 1992). Mean palaeocurrent direction within unit 2 is east-south-east (Gustason, 1989), but the dispersal of palaeocurrents is high.

Architecturally, the channel-fill bodies include lateral accretion, vertical and concentric accretion, and are documented in detail by Kirschbaum & McCabe (1992). A part of the channel fills described by these authors from the Henrieville area is, however, now included in unit 3A and was found to fill a topography incised into unit 2.

© 1999 International Association of Sedimentologists, Sedimentology, 46, 807–836
Gustason (1989) described an apparent change in fluvial architecture across the region, from isolated channel-fill bodies encased in mudstones in the west, towards wide, laterally persistent, lateral-accretion lithosomes in the east. Re-examination of relevant sections by the author showed that most of the laterally extensive channel systems in the 'middle Dakota' are, in fact, incised channels of units 3A and 3B and cannot be correlated to unit 2 in the western part of the study region.

A conspicuous component of unit 2 is coal, occurring in distinct assemblages of seams (coal zones) ranging in thickness from 5 cm to 2.5 m. Unit 2 comprises coal zones 1–3 of Kirschbaum & McCabe (1992). Coal zones 1 and 2 correspond to the Bald Knoll coal of Doelling & Graham (1972). Coal zone 1 is the least extensive, whereas coal zone 2 is a major group of thick coals that are regionally correlatable. Dispersed, thin coals of limited lateral extent within unit 2 correspond to coal zone 3 of Kirschbaum & McCabe (1992). The coal zones 4 and 5 of the 'middle member of the Dakota' are genetically related to units 3A, 3B and, possibly, 4.

Environmental interpretation

Unit 2 is interpreted as representing an anastomosed river system, characterized by isolated channels encased in floodplain and swamp deposits (cf., e.g. Smith, 1986). The sheet-like sandstones interbedded with the mudstones and coals represent crevasse-splay deposits. This interpretation of the depositional system is in agreement with that of Gustason (1989) and Kirschbaum & McCabe (1992), but it applies only to the lower portion of their 'middle Dakota'. The wide palaeocurrent dispersal reported by Gustason (1989) is partly caused by the sinuosity of the anabranches, and partly because the measurements of Gustason (1989) also included units 3A and 3B, commonly characterized by intense meandering.

Kirschbaum & McCabe (1992, their Fig. 13) pointed out the apparent similarities between Figs 4 and 5. Gustason (1989) described an apparent change in fluvial architecture across the region, from isolated channel-fill bodies encased in mudstones in the west, towards wide, laterally persistent, lateral-accretion lithosomes in the east. Re-examination of relevant sections by the author showed that most of the laterally extensive channel systems in the 'middle Dakota' are, in fact, incised channels of units 3A and 3B and cannot be correlated to unit 2 in the western part of the study region.

A conspicuous component of unit 2 is coal, occurring in distinct assemblages of seams (coal zones) ranging in thickness from 5 cm to 2.5 m. Unit 2 comprises coal zones 1–3 of Kirschbaum & McCabe (1992). Coal zones 1 and 2 correspond to the Bald Knoll coal of Doelling & Graham (1972). Coal zone 1 is the least extensive, whereas coal zone 2 is a major group of thick coals that are regionally correlatable. Dispersed, thin coals of limited lateral extent within unit 2 correspond to coal zone 3 of Kirschbaum & McCabe (1992). The coal zones 4 and 5 of the 'middle member of the Dakota' are genetically related to units 3A, 3B and, possibly, 4.
the Dakota anastomosed rivers and the Okavango fluvial system in Botswana and interpreted the Dakota system as ‘some form of large inland delta introducing clastics into a widespread mire’. However, the occurrence of inclined heterolithic stratification, locally rhythmic and containing flaser to lenticular bedding, close to the base of unit 2 in the eastern part of the study region is interpreted here as a record of tidal influence on fluvial deposition (Fig. 3e; cf. Thomas et al., 1987; Shanley et al., 1992). Therefore, it is suggested here that the fluvial system of unit 2 was not a ‘large inland delta’, as hypothesized by Kirschbaum & McCabe (1992), but may have been connected to an estuarine system east of the Kaiparowits region.

**Units 3A and 3B**

Units 3A and 3B share similar facies and architectural characteristics: both are markedly heterolithic, show signs of tidal influence in the depositional environment and occupy incised channel systems. The units are separated from one another by a palaeosol capped with a distinctive coal bed containing a thin bentonite interbed, belonging to coal zone 4 of Kirschbaum & McCabe (1992). However, channels at the base of unit 3B commonly incise into unit 3A, removing the bounding palaeosol and coal and thus making it difficult to distinguish between the two units. For this reason, Uličný (1996) treated them as a single unit, and their similarity is reflected in their designation as units 3A and 3B.

**Description**

Unit 3A is 0–10 m thick and occupies channels incised into the underlying unit 2 (Fig. 2). It comprises trough cross-bedded sandstone and heterolithic sandstone–mudstone laminites and grades upwards into silty mudstones, commonly carbonaceous, capped by a palaeosol and a coal bed. The channel fills show lateral accretion architecture (Figs 4 and 6a), commonly with

---

Fig. 5. Measured sections and palaeocurrent data for units 2 to 3B (Sections 4a–d, locations shown in Fig. 4a). The locations of features illustrated in Figs 6 and 7 are shown. Symbols as in Fig. 2.

© 1999 International Association of Sedimentologists, *Sedimentology, 46*, 807–836
transitions between inclined heterolithic stratification (sensu Thomas et al., 1987) and cross-bedded sandstones forming the upper and lower portions of the channel fills. Sandstone-dominated channel fills, with abundant lags of mud and coal clasts, are characterized by trough cross-sets with common mud drapes and, locally, mud couplets on foresets (Fig. 6). The drapes are commonly formed by carbonaceous debris instead of mud. Bottomsets show flaser lamination. Synsedimentary deformation is common locally. Heterolithic laminites (Fig. 6c) show a typical continuum of flaser to lenticular, ‘tidal’ bedding, locally rhythmic. A channel fill in Section 2 (Fig. 6a) contained corbulid bivalves; elsewhere, bivalve shell moulds occur in sandstones. Palaeocurrent indicators, where measured, show eastward sediment transport (Fig. 5).

Unit 3B is 0-8 to more than 9 m thick and is a highly heterolithic facies succession. It is separated from unit 3A by a coal bed (Fig. 4; coal zone 4 of Kirschbaum & McCabe, 1992), except for cases where the coal has been removed by subsequent erosion. Unit 3B was identified in more sections than the underlying unit 3A. Although unit 3B does not occupy such markedly incised channels as unit 3A, regional correlation shows that it changes thickness across relatively short distances (Fig. 2), and local incision into underlying units is common (Fig. 7).

Unit 3B is most commonly characterized by a fining-upward trend, from fine-grained, well-sorted, planar and trough cross-bedded sandstones at the base, through heterolithic laminites to black, carbonaceous mudstones at the top, rooted and capped with a coal bed (Fig. 7b and c).
Large-scale depositional architecture, where sufficiently exposed (over tens to hundreds of metres), is dominated by lateral accretion, in both the cross-bedded sandstones and the laminites (Fig. 4). The sandstone foresets are draped by mud and carbonaceous debris, commonly grouped to form tidal bundles (Fig. 7; cf. Visser, 1980; Nio & Yang, 1991). Some cross-sets display sigmoidal-shaped foresets and, locally, herringbone cross-bedding. Palaeocurrents are directed predominantly west to north-west (Fig. 5 and 7). The interlaminated sandstones and mudstones show various types of ‘tidal bedding’, locally markedly rhythmic (Fig. 7a). In places, the unit contains extremely abundant corbulid bivalves. Arenicolites and Ophiomorpha occur in the sandstones. In places where both units 3A and B are absent or very thin, a thick palaeosol is developed under unit 4 (Sections 5 and 6 in Fig. 2).

Environmental interpretation

The occurrence of coupled mud drapes on foresets, sigmoidal-shaped foresets, herringbone cross-bedding and flaser- to lenticular-bedded laminites indicates clearly that units 3A and 3B formed in tide-influenced settings (Figs 5, 6 and 7; cf. Nio & Yang, 1991). In Section 4, sampled for palynological analysis, marine microplankton was found in the coals capping both unit 3A and unit 3B, as well as in the mudstones within unit 3B (M. Svobodová, pers. comm., 1996). Also, the occurrence of corbulid bivalves attests to brackish water influence in both units.

The two units differ primarily in the abundance of tidal indicators, which are more common in unit 3B than in unit 3A, and in the lateral extent of their deposits. Although both units are characterized by an order of magnitude variation in thickness across the study area, which results from their occurrence as channelized sediment bodies, the channel systems of unit 3A are much more localized than those of unit 3B (Fig. 2).

Unit 3A is interpreted as deposited in a tide-influenced fluvial environment, based on the criteria listed above and on comparisons with other tide-influenced fluvial systems (e.g. Shuley et al., 1992). The fluvial system occupied channels and shallow valleys incised into unit 2.
Individual channels meandered within the incised valleys. The influence of brackish water, documented by corbulids, as well as the probable neap–spring tide rhythmicity recorded in the laminites (Fig. 6c) suggest that the environment of deposition was close to the junction of the fluvial system and the upper estuary (Rahmani, 1988; Shanley et al., 1992).

The deposition of unit 3B indicates the establishment of an estuarine, tide-dominated depositional environment. The more widespread nature of this unit relative to unit 3A probably results both from a larger area of deposition and possibly from erosion of unit 3A by scour in subtidal channels (cf. Oertel et al., 1991; Uličný & Špičáková, 1996). Herringbone cross-bedding indicates a strongly asymmetrical tidal ellipse caused by confinement of flow in channels; this is also supported by the abundance of mud drapes (cf. Yang & Nio, 1989). The lateral accretion architecture of both the cross-bedded sands and heterolithic strata reflects lateral migration of gently dipping sand bodies, interpreted as tidal point bars, within estuarine channels (Thomas et al., 1987; Ainsworth & Walker, 1994). Thin successions dominated by tidal laminites, commonly fining upwards, represent tidal flat deposits outside the main channel courses. Palaeocurrent data show consistently a SE–NW trend of the channels, similar to the drainage pattern of underlying fluvial units. Local accumulations of corbulid bivalves in areas where the unit is thin seem to reflect low local rates of sedimentation (e.g. Section 2 in Fig. 2). Locally, the laminites show rhythmic thinning and thickening of the two lithologies, reflecting neap–spring tide rhythmicity (Fig. 7; cf. Kvale & Archer, 1990; Nio & Yang, 1991).

A trend of increasing tidal influence in an eastward direction, supported by the abundance of sand-dominated channels filled by subtidal dunes in the Big Water area, indicates that the western part of the study area represents the middle estuary, with heterolithic estuarine point bars and tidal flats, whereas the eastern part was close to the estuary mouth (Fig. 7d and e). The position of the estuary-open marine transition is unclear, but a thin, coarsening-upward, shoreface succession that occurs in the easternmost part of the study region (Fig. 2), tentatively interpreted as coeval to tidal deposits of unit 3B, suggests that this transition may be preserved in the southeastern tip of the Kaiparowits region. Precise correlation is difficult because of discontinuous exposure.

**Units 4 to 6B**

Units 4 to 6B represent the ‘upper member’ of the Dakota Formation in the study area (Table 1). A feature common to units 4, 5 and 6A is an upward-coarsening facies succession interpreted as offshore to shoreface deposits. Specific to individual units are occurrences of other lithofacies, including cross-bedded sandstones, heterolithic laminites, coals and others. Their relevance to sequence-stratigraphic interpretation necessitated separate descriptions of these facies associations for individual units. Units 6A and 6B are locally difficult to distinguish because of very similar facies associations.

Gustason (1989) and Elder et al. (1994) divided the ‘upper member’ in the study area into two shallowing-upward parasequences and correlated them to limestone–shale couplets in the distal facies of the Western Interior. It will be demonstrated below that the sequence stratigraphy of the ‘upper member’ is significantly more complicated.

**Description**

Unit 4 is 0–6 m thick and comprises an upward-coarsening succession of shallow-marine mudstones and sandstones (Fig. 8a and c). It thins both downdip and updip from the zone of maximum thickness between Sections 2 and 9. Downdip, it pinches out near Section 14 (Fig. 2).

The basal surface of unit 4 is sharp, commonly overlain by a fossiliferous sandstone, 0.1–1 m thick, rich in bivalve shells, mostly small oysters (Fig. 8). This deposit, commonly cemented by siderite and interbedded with coal debris, locally overlies a thin (up to 1 m thick) succession of mudstones rich in corbulid bivalves or a thin, lenticular-bedded, heterolithic succession.

The main body of unit 4 consists of grey marine mudstones, which alternate with fine-grained, well-sorted, subarkosic to arkosic sandstones from the middle part of the unit upwards. The sandstone beds thicken upwards. From bottom to top, a succession of wave-ripple lamination, undulating parallel lamination and hummocky to swaley cross-stratification occurs in the sandstones. Synsedimentary deformation is common, especially in sections at the western edge of the Kaiparowits Plateau (Sections 8 and 9 in Fig. 2), and is dominated by convolute bedding and ball-and-pillow structures (Fig. 8c).

Unit 5, up to 12 m thick, is an upward-coarsening, shallow-marine succession similar to that of unit 4. The boundary between units 4
and 5 is either sharp, overlain by an oyster lag, commonly sideritic, or it is marked by shallow scours and, locally, channels up to 4 m deep. The channels are filled with cross-bedded sandstone with coal debris and very abundant oysters (Fig. 8d and e). The top of unit 5 is an erosional surface, locally penetrated by roots, overlain by various facies assemblages of units 6A and 6B.

In the uppermost parts of unit 5, trough cross-bedded sandstones are thicker than analogous facies of unit 4, but are not continuously preserved. To the east of Section 9, gutter casts begin to occur in the middle portion of the unit. Further eastwards, the thickness of unit 5 is reduced to a few metres because of erosion at the base of overlying unit 6A.

Unit 6A, up to 8 m thick, comprises two different facies associations, here called proximal and distal. The distal part of unit 6A is an upward-coarsening succession, analogous to the underlying units 4 and 5, of dark, silty mudstones, wave-ripped sandstones and thin HCS beds. It occurs east of Section 12 and, in the Coyote Creek area (Section 11), it passes updip into the proximal assemblage. The top of unit 5 is overlain by a distinctive carbonaceous layer, up to 40 cm thick, formed largely by redeposited carbonaceous debris with silty and sandy admixture and locally abundant thin-shelled bivalves and gastropods (Fig. 9f and g). The topmost parts of unit 6A pass upwards into heterolithic laminates and carbonaceous mudstones of the proximal facies assemblage.

The proximal part of unit 6A is exposed west of Section 11 (Fig. 2), where it rests on unit 5. Its lower bounding surface is locally burrowed and penetrated by roots and shows a pronounced erosional relief (Fig. 9b and h; Section 8 in Fig. 2). The proximal facies association includes planar and trough cross-bedded, fine- to medium-grained sandstones; massive, intensely bioturbated, fine-grained sandstones; heterolithic laminites; carbonaceous mudstones and coals. Cross-bedded sandstones occur as channel fills, commonly several metres thick, interbedded with heterolithic laminites. Common trace fossils include Ophiomorpha, Thalassinoides and Arenicolites. The massive sandstone facies contains abundant trace fossils, which show that intense bioturbation caused the loss of original physical structures. The heterolithic laminites show a continuum of flaser, wavy and lenticular bedding and commonly fine upwar ds, capped by coals or carbonaceous mudstones (Fig. 9b). Distinct cracks occur in the heterolithics and coals of the proximal part of unit 6A in sections near the East Kaibab monocline, filled by overlying sands and showing compactional folding (Fig. 9b).

Unit 6B overlies an erosional surface on top of the proximal facies of unit 6A. The boundary is marked by tree stumps locally preserved under unit 6B or by coals at the top of unit 6A, with erosional features such as gullies and channels (Fig. 9a–e). In the Rimrocks area (Figs 2 and 9a), the channels incised into the coals are filled with dark-coloured, bioturbated sandstone and siltstone with extremely abundant carbonaceous debris. Other facies include cross-bedded sandstones with bidirectional (WNW–ESE) palaeocurrent patterns, mudstone drapes on foresets, sigmoidal foresets, locally well-developed mudstone couplets forming tidal bundles and bundle sequences (Fig. 9i–l). Unit 6B is easily identified to the east of Section 5, where its most conspicuous facies, sandstone with accumulations of thick-shelled Exogyra oysters, mostly preserved in situ as ‘oyster banks’ occurs below the basal surface of the Tropic Shale (Fig. 9d and e). This facies is commonly underlain by fine-grained, bioturbated sandstone with scattered bivalves.

Although it is difficult to correlate unit 6B physically to the west of Section 5 because of

---

**Fig. 8.** Units 4 and 5. (a) Typical appearance of units 4 and 5 as simple shallowing-upward ‘parasequences’ in Section 10; see text for discussion and (e) for a measured section. (b) Lower part of unit 4, interpreted as a backbarrier transgressive systems tract, overlain by a wave ravinement surface marked by a shell lag (arrow); Section 12. (c) A body of convolute-bedded sandstone (arrow), interpreted as a submarine slump, thickening towards the right from 0.7 to 2.5 m, occurs in the uppermost part of unit 4; Section 8. (d) Cross-bedded sandstone with oyster coquinas on foresets, filling a channel incised into unit 4 and interpreted as a tidal inlet fill, part of a transgressive systems tract of unit 5; Section 2. (e) Selected measured sections illustrating the internal complexity of units 4 and 5, interpreted as high-frequency sequences; note that where the transgressive systems tracts (TSTs) are not preserved (Section 10), both units have the appearance of para-sequences. The systems tracts defined are facies based: RST, ‘regressive systems tract’ (regressive deposits in which the boundary between the highstand (HST) and falling stage (FSST) systems tracts cannot be identified); SB, sequence boundary; FS, flooding surface; RSME?, tentatively interpreted regressive surface of marine erosion, defined by the sharp base of shoreface sandstones passing downdip into a gradual upward-coarsening succession. See Discussion for comments.
Sequence stratigraphy of Dakota Formation

© 1999 International Association of Sedimentologists, Sedimentology, 46, 807–836
Fig. 9. Units 6A and 6B. (a) A channel at the base of unit 6B (arrow) incised into the coals of the proximal part of unit 6A and filled with siltstones containing a high proportion of carbonaceous debris; the lenticular shape of the channel fill is caused by high compaction of coal underlying the channel flanks; Section 9. (b) A crack in coal and underlying heterolithic strata of unit 6A, filled with overlying bioturbated sandstone and showing compactional folding; Section 9; note also slumped beds overlying the channelized 5/6A boundary. (c) The 6A/6B bounding surface showing erosional scour in coal, filled by bioturbated sandstone of unit 6B; Section 9. (d) Base of the main body of the Tropic shale, marked by a lag deposit dominated by *Pycnodonte newberryi* oysters, overlying the *Exogyra* sandstone of unit 6B; Section 9. (e) Casts of tree roots and stumps preserved beneath the 6A/6B boundary; note the *Exogyra* sandstone directly overlying the scoured surface; arrows show branching roots; Section 7. (f) The 5/6A boundary marked by accumulated carbonaceous debris (arrow) overlying an erosional surface at the base of the distal facies of unit 6A; Section 13. (g) Close-up view of the base of unit 6A shown in (f).
Fig. 9h–l. (h) Units 6A and 6B filling a channel incised into unit 5; note the lateral accretion surfaces in the sandstones (arrow); Section 2. (i) Neap–spring tidal cyclicity preserved in tidal bundles of subtidal channel of unit 6B; Section 2, see measured section in (j). (j) Lithofacies and palaeocurrents of units 6A and 6B; Section 2. (k) Sandstone-filled channel incised into the distal facies of unit 6A and interpreted as a tidal inlet fill related to the 6B transgression; Section 12; see Depositional history for details. (l) Measured section of the channel fill shown in (k).
discontinuous exposure, several channel fills consisting of cross-bedded sandstone with tidal bundles and heterolithic laminite interbeds at the western margin of the study area are interpreted as belonging to unit 6B (Fig. 9i and j). This correlation, however, is tentative, based on sequence-stratigraphic reasoning explained further below.

Unit 6B is overlain by a thin, bioturbated, fossiliferous sandstone rich in *Pycnodonte newberryi* oysters, commonly forming laterally extensive coquinas, which is the basal deposit of the main body of the Tropic Shale (Fig. 9d).

**Environmental interpretation**

The upward-coarsening successions of units 4 and 5 and the distal part of unit 6A are typical successions formed in offshore to wave-dominated lower to upper shoreface environments by the progradation of a wave-dominated strandplain (for example, cf. Walker & Plint, 1992 and references therein). Orientation of wave-ripple crests is generally north–south (Fig. 8e), which corresponds to the shoreline strike during the deposition of these units. Gutter casts found in units 4 and 5 were oriented essentially perpendicular to the shoreline. Mudstones with abundant corbulid bivalves, found locally in the basal parts of units 4 and 5, as well as heterolithic laminites, are interpreted as preserved relics of back-barrier, lagoon/estuarine deposits, and channel-filling sandstones found locally at the base of unit 5 are interpreted as relics of tidal inlet fills (Fig. 8e).

The proximal part of unit 6A is interpreted as deposits of a tide-dominated estuarine setting. Gustason (1989) also identified the tidal depositional regime in this part of the Dakota Formation, but he interpreted the tide-influenced deposits as lagoon deposits formed behind a prograding barrier island system. Although the fauna of unit 6 is similar to that described by Fürsich & Kirkland (1986) in a late Cenomanian brackish lagoon deposit from the Black Mesa, Arizona, muddy facies that could be interpreted as back-barrier lagoon-fill sediments are not common in unit 6. Strong tidal currents are recorded in the abundant cross-bedded sandstones filling channels and representing subtidal sand bars, as well as in tidal laminites, formed on tidal flats. Carbonaceous mudstones and coals were deposited in supratidal marshes. This facies assemblage is very similar to cases described by Devine (1991), Leckie & Singh (1991) and Ainsworth & Walker (1994) from tide-dominated, transgressive estuarine deposits filling channels incised into wave-dominated shoreface deposits.

Unit 6B was deposited in environments similar to the proximal part of unit 6A. Whereas the cross-bedded sandstones and heterolithic laminites are interpreted as tidal channel and tidal flat facies, the sandstones with *Exogyra* accumulations correspond to a high wave energy environment and are interpreted as foreshore deposits. A special case among various tidal channel fills of unit 6B is a large channel complex in the Wahweap Creek area. The channel, up to 300 m wide, =10 m deep and dominated by bidirectionally cross-bedded sandstones is incised into the offshore–shoreface deposits of the distal part of unit 6A (Fig. 9k and l). It may represent either a tidal inlet fill, as argued by Gustason (1989), or a transgressive estuarine channel occupying inherited erosional topography of a sequence boundary. A combined interpretation is also possible, because it is common for tidal inlets within transgressive barrier systems to form in pre-existing channels and further enhance their relief (e.g. Oertel et al., 1991). In any case, the channel is considered a relic of a transgressive barrier system, not a prograding one, as argued by Gustason (1989) and followed...

DEPOSITIONAL HISTORY

Unit 1

The depositional history of unit 1, complicated by the lack of reliable age data, is closely related to the preceding events responsible for the incision of the topography underlying unit 1. This erosional episode probably coincided with the eastward tilting of the Jurassic bedrock, which occurred during the Early Cretaceous (the ‘Neocomian’ regional uplift; Peterson, 1969). The eastward tilting and resulting westward-deepening erosion of the Jurassic strata are explained here as a consequence of the rise of a forebulge formed in response to initial thrusting to the west of the mid-Cretaceous position of the Sevier thrust front (Fig. 10). A similar explanation has been proposed recently by Currie (1998) for the palaeovalley systems underlying the Buckhorn Conglomerate member of the Cedar Mountain Formation in north-eastern Utah. Other, alternative interpretations, such as a thermal uplift (cf. Heller & Paola, 1989) or fluctuations in compressional intraplate stress (Heller et al., 1993; Peper, 1993), are not preferred here (for detailed discussion, see Currie, 1998). The forebulge uplift would be consistent with the hypothesis of an Early Cretaceous foredeep developed in front of the Pavan and Canyon Range thrust sheets in western Utah (Royse, 1993; Schwans, 1995).

The data of Currie (1997, 1998) indicate that, during this time, eastward migration of the forebulge shifted the NE Utah region from the ‘back-bulge depozone’ of the foreland basin system (as defined by DeCeilles & Giles, 1996) to the zone of forebulge uplift. In the Kaiparowits region, situated further east of the Sevier orogenic front, this shift may have occurred later. The subsequent eastward propagation of the Sevier thrust front during the Cenomanian caused the onset of the typical flexural regime of the foredeep depozone: the backtilting of the Jurassic basement in southern Utah, possibly combined with a eustatic rise, led to deposition of the unit 1 valley fills. Although detailed research is clearly necessary to test the hypothesis of a forebulge-related incision, the depth of erosional relief preserved (=50 m maximum) and the width of the area affected by erosion (=150 km along section in Fig. 1d, normal to the thrust front) are within the range of values estimated for the height and wavelength of a forebulge, given the flexural rigidity expected of the Western Interior basin (cf. DeCeilles & Giles, 1996 and references therein).

The Ar/Ar dates from bentonites in the lower part of unit 2 suggest that the youngest possible age of unit 1 is Early or early Middle Cenomanian (based on Bohor et al., 1991; see below). On the basis of lithologic similarity between unit 1 and the Lower Cretaceous conglomerates found in the Western Interior (e.g. the Buckhorn Conglomerate), Gustason (1989) interpreted the age of unit 1 as Early Cretaceous (Albian). However, the Lower Cretaceous conglomerates found in the foreland of the Sevier belt and interpreted formerly as synorogenic conglomerates recording active thrusting are now considered to be older than the flexural subsidence phase in the Cretaceous foreland basin (cf. Heller & Paola, 1989; Currie, 1998). The incision of the valleys in the Kaiparowits region most probably provided clastics for Albian gravels, studied by Heller & Paola (1989), further eastward in the Sevier foreland. The incision might have been enhanced by the effects of Early Cretaceous eustatic fluctuations, as interpreted by Aubrey (1992) for the sub-Dakota surface in New Mexico, but this possibility has to be tested by improved stratigraphic correlation.

The evidence of erosion and pedogenesis at the top of unit 1 suggests a gradient change caused either by uplift (favoured by Gustason, 1989) or by a sea-level fall. Although signs of subaerial erosion were recognized on bounding surfaces of many of the units of the Dakota Formation, the lack of physical correlation to coeval paralic strata, as well as the lack of reliable chronostratigraphic data for unit 1, preclude an unequivocal explanation of the cause of this erosional event.

Unit 2

The contrast in fluvial style between units 1 and 2 (Fig. 3d), as well as the internal organization of unit 2, suggest a significant change in stream profiles and a major decrease in regional sediment input rate in the Dakota fluvial system. Such a change in depositional style, emphasized by the occurrence of coal on top of unit 1 over most of the area, can be attributed either to an increased rate of flexural subsidence in the foreland basin leading to accelerated backtilting of the stream profiles and trapping of coarse sediment close to the orogenic front, or to a relative sea-level rise causing transgression on the coast east of the

© 1999 International Association of Sedimentologists, Sedimentology, 46, 807–836
Kaiparowits area and resultant flattening of the stream profiles. The latter explanation is favoured here because of the occurrence of probably tide-influenced strata immediately above the base of unit 2 in the eastern part of the study area.

Further considerations of the depositional history of unit 2 depend on the accuracy of its age estimates, further complicated by the lack of biostratigraphic data. Bohor et al. (1991) dated altered volcanic ashes from the Cenomanian of the Kaiparowits area, including the coal zones 2 and 3; although the absolute ages are ≈3 Myr younger than comparable Cenomanian dates of Obradovich (1993), the internal consistency of the Bohor et al. (1991) data set allows an estimate of the time span represented by the strata between the two zones as ≈1-4 Myr. The corrected age for the coal zone 2 is close to 95-8 Myr, the Early-Middle Cenomanian boundary (Table 1; based on Obradovich, 1993).

Coal seams in fluvial strata deposited on coastal plains have been interpreted recently as time-equivalents of flooding surfaces at coeval coastlines, reflecting a rise in water table on the coastal plain, as well as the decrease in rate of sediment input resulting from base-level rise (Flint et al., 1995 and references therein). Based on the characteristics of the coal zones and the clastic deposits between them, unit 2 is interpreted as recording a long-term cycle of base-level change. The discontinuous coal zone 1 is interpreted as reflecting initial flooding, whereas the laterally continuous, thick coal zone 2 may correlate to a maximum flooding interval. The clastic deposits between these coal zones are dominated by mudstones and are tentatively interpreted here as a transgressive systems tract. The deposits overlying the coal zone 2 are characterized by an upward increase in the abundance of channels and by thin, discontinuous coal seams (coal zone 3 of Kirschbaum & McCabe, 1992), which may represent a relative base-level highstand (cf. Wright & Marriott, 1993; Olsen et al., 1995). The erosional surface underlying unit 3A is interpreted as a sequence boundary (see below). It must be kept in mind, however, that numerous palaeosols and coals within unit 2 may reflect a number of higher frequency base-level fluctuations superimposed on the long-term trend.

**Units 3A and 3B**

The occurrence of the tide-influenced unit 3A above the fluvial deposits of unit 2 clearly indicates a relative sea-level rise. However, the erosional relief underlying unit 3A and thick palaeosols developed where channel-fill sandstones of unit 3A are absent (e.g. Section 6 in Fig. 2) are regarded as evidence for a preceding base-level fall, and thus the base of this unit is interpreted as a sequence boundary coinciding with a flooding surface. The fining-upward pattern within unit 3A is a response to filling of accommodation, terminated by deposition of mudstones with pedogenic features and coals at the top of unit 3A.

Erosional topography also exists at the base of unit 3B, locally forming channels up to 8 m deep (Fig. 7), which formed either by subaerial erosion as a result of a base-level fall or by autogenic channel incision by tidal currents in the estuarine environment. Such ‘tidal ravinement’ (Allen, 1991) is common in estuarine settings and may cause some flooding surfaces to resemble sequence boundaries superficially (Uličný & Špičáková 1996). Erosion by tidal currents seems likely because the deepest incision is observed in the eastern part of the study area, where evidence exists for vigorous current action in subtidal channels of the ‘lower estuary’ (Fig. 7d and e). On the other hand, in the western part of the area, coal zone 4 on top of unit 3A is overlain by sandstones and heterolithics of the ‘middle estuary’ portion of unit 3B, without significant erosion. It is also possible that the coal bed underlying the sandstones and heterolithic deposits of unit 3B in the western part of the study area reflects mire formation on top of unit 3A, in response to early sea-level rise.

The depositional history of units 3A and 3B is summarized as follows: (i) after a relative sea-level fall, incision of shallow channel systems into unit 2 occurred, accompanied by pedogenesis in the interfluves; (ii) a relative sea-level rise caused filling of the incised topography by tide-influenced fluvial deposits of unit 3A; (iii) after a stillstand or a minor relative sea-level fall, deposition of unit 3B occurred in response to significant flooding of the area and onset of deposition in a tide-dominated, estuarine system, which covered a much larger area than the localized channels of unit 3A. The carbonaceous mudrocks and coals in the topmost parts of unit 3B formed in a supratidal environment in response to filling of the accommodation. The possibility of a significant relative sea-level fall terminating deposition of unit 3B can be neither ruled out nor confirmed by existing data, but the marine strata of unit 4 clearly overlie a subaerially exposed surface.
Units 4 and 5

The base of unit 4 is a record of the first marine flooding of the Kaiparowits area, at the beginning of the *M. mosbyense* Zone (cf. Elder *et al.*, 1994). Locally preserved mudstones with brackish water bivalves and heterolithic laminites of intertidal origin in the lower part of unit 4 (e.g. Sections 6, 8 and 12; Figs 2 and 8b and e) probably represent remnants of back-barrier deposits formed in a transgressive barrier–lagoon depositional system. Most of the deposits of this system were destroyed by subsequent wave ravinement, which resulted in the deposition of an oyster lag either on top of the underlying unit 3B or on top of the relicts of back-barrier deposits. This transgressive lag is overlain by the shallowing-upward offshore–shoreface succession, which forms the bulk of unit 4. No direct signs of subaerial erosion were found at the top of unit 4.

The base of unit 5 is another prominent marine flooding surface, in many ways analogous to that of unit 4. The back-barrier and tidal-inlet facies found in the westernmost part of the area represent relict transgressive deposits of a barrier system similar to that of unit 4. The shell lags at the base of the offshore muds of unit 5 overlie a wave ravinement surface, which marks the maximum flooding during deposition of the unit. The overall thickness of unit 5, compared with that of unit 4, indicates a relatively larger increase in accommodation. The gradual filling of accommodation, represented by the bulk of unit 5, was terminated by subaerial exposure and valley incision. Based on the data available, it is difficult to speculate about the position of the boundary between the highstand and falling stage deposits (*sensu* Plint, 1996) within the shallowing-upward succession of unit 5.

Both unit 4 and unit 5 superficially resemble offshore–shoreface parasequences, as defined by Van Wagoner *et al.* (1990), and were interpreted as such by Elder *et al.* (1994). The internal structure of these units, however, allows interpretation of both transgressive and regressive portions (systems tracts) of depositional sequences within both units and is discussed further below.

Units 6A and 6B

A significant basinward shift in facies occurs at the top of unit 5, an erosional surface with up to 10 m relief, traceable throughout the study region and locally penetrated by plant roots, which justifies the interpretation of a sequence boundary underlying unit 6A. Additional supporting evidence is the separation of the distal part of unit 6A from the underlying deposits by a bed containing reworked coal material. This bed is interpreted as a lag deposit overlying a wave ravinement surface, i.e. a surface that was initially exposed subaerially and covered by a mire. This interpretation means that the marine lowstand deposits overlying the sequence boundary were deposited east of the Kaiparowits area. A tidal channel below the base of unit 6A near Section 15 (Fig. 2) may correspond to this lowstand, but the correlation is tentative because of discontinuous exposure. Apart from the major sequence boundary separating units 5 and 6A, the evidence provided by incised channels in coals at the top of unit 6A and tree stumps overlain by the *Exogyra* sandstones of unit 6B suggest an episode of subaerial exposure before the deposition of unit 6B.

The 5/6A sequence boundary correlates chronostratigraphically with a more deeply incised sequence boundary, of up to 20 m relief, in the upper part of the Twowells Sandstone in north-western New Mexico (Mellere, 1994; Uličný, 1994). The evidence for a relative sea-level fall contradicts earlier interpretations by Gustason (1989) and Elder & Kirkland (1994) that the uppermost parts of the Dakota formed during a short-term sea-level stillstand.

The deposition of unit 6 took place in several stages: (i) initial transgression, causing the establishment of a mire on the sequence boundary; (ii) marine transgression leading to destruction of the mire and deposition of tide-dominated, estuarine facies in the western part of the area (unit 6A, proximal) and, later, a highstand or falling stage leading to the deposition of wave-dominated, strandplain deposits (unit 6A, distal); (iii) minor relative sea-level fall, which caused local subaerial erosion of unit 6A; and (iv) flooding leading to the deposition of unit 6B.

The formation of the wave ravinement surface at the base of the offshore Tropic Shale marked the major flooding event of the early *S. gracile* Zone (cf. Gustason, 1989; am Ende *et al.*, 1991). The low-diversity molluscan assemblage formed during conditions of low sediment supply and possibly high influx of nutrients, indicating abrupt deepening.
DISCUSSION

Sequences vs. parasequences

Units 1–6 of the Dakota Formation have been defined on the basis of the interpretation of their bounding surfaces: all unit boundaries were shown to represent flooding (and correlative) surfaces. Most of them formed initially during base-level falls, i.e. as sequence boundaries, *sensu* Van Wagoner et al. (1988). The interpretation of the individual units as sequences, not parasequences, as thought earlier by Elder et al. (1994) and Uličný (1994), stems from the evidence for relative sea-level falls discussed in previous sections. The ambiguity of interpretation of the basal surfaces of units 3B, 4 and 5 emphasizes the problems concerning the distinction between parasequences and high-frequency sequences. Especially where evidence of subaerial erosion is equivocal, or where it disappears in a basinward direction, it is difficult to distinguish between the two categories (Figs 8 and 11).

A parasequence (Van Wagoner et al., 1990) is a ‘relatively conformable succession of genetically related beds or bedsets bounded by marine flooding surfaces or their correlative surfaces’, in which the vertical facies succession reflects a gradual decrease in water depth. Kamola & Van Wagoner (1995, p. 31) state explicitly that ‘there are no flooding surfaces nor unconformities within a parasequence’. The shallow-marine units 4, 5 and 6A, however, show characteristics contradicting the parasequence definition. Units 4 and 5 show a more complicated internal organization of facies assemblages than a simple shallowing upward associated with parasequence progradation. The locally preserved transgressive deposits fill a pronounced relief of the underlying sequence boundary; the regressive portion of the sequence is almost detached from the transgressive systems tract because the high-frequency relative sea-level rise is superimposed on a low-frequency fall (exemplified by unit 6A, inset), resulting in a long-term progradational stacking. Symbols in insets show positions of underlying sequence boundaries and position of the time slices (b) and (c) in the relative sea-level curve. The systems tracts defined are facies based. TST, transgressive systems tract; RST, ‘regressive systems tract’ (regressive deposits in which the boundary between the highstand and falling stage systems tracts cannot be identified); SB, sequence boundary; TS, transgressive surface; LST, lowstand systems tract. No scale implied.

© 1999 International Association of Sedimentologists, *Sedimentology, 46*, 807–836
back-barrier and inlet deposits are interpreted as a transgressive systems tract, whereas the offshore-shoreface successions form the highstand to falling stage portion of a sequence (Figs 8e and 11), which is called here informally the regressive systems tract because of difficulties in field recognition of the boundary between the highstand and falling stage deposits. The systems tracts are defined on the basis of facies and key surfaces, not on the basis of internal geometric arrangement (stacking pattern) of depositional units, which is typical of high-frequency sequences at a scale similar to that of the Dakota units (cf. Swift et al., 1991a,b; Plint, 1996). Thus, units 4 and 5 can be termed high-frequency sequences, although some of their bounding surfaces cannot be shown unequivocally to represent sequence boundaries (for instance, at the top of unit 4, any evidence of subaerial erosion may have been destroyed by subsequent transgressive erosion). On the contrary, unit 6A is bracketed by relatively well-constrained sequence boundaries. Its proximal facies assemblage is interpreted as estuarine deposits of the transgressive systems tract, whereas most of the distal portion of this unit is interpreted as the ‘regressive’ systems tract, based on the spatial distribution of facies. The division of the sequences into the facies-based, transgressive and regressive portions is conceptually very close to the depositional couplets of Devine (1991).

Although a more general discussion of the sequence vs. parasequence definitions is beyond the scope of this paper, the examples from the ‘upper Dakota’ illustrate that the distinction between high-frequency sequences and parasequences commonly depends primarily on the availability of evidence and the amount of data (Christie-Blick & Driscoll, 1995; Swift et al., 1991b). At the scale of the high-frequency sequences observed in the Dakota Formation, it is considered more practical to use the term ‘parasequence’ for a special case of a depositional sequence (cf. Swift et al., 1991b), rather than to amend the Van Wagoner et al. (1988, 1990) parasequence definition by including transgressive deposits (Arnott, 1995).

**Periodicity and magnitude of relative sea-level changes**

Good geochronological and biostratigraphical constraints are available for units 4–6B (Table 1), but the accuracy of dating decreases downsection, which affects the estimates of relative sea-level change periodicity. In the case of unit 1, for which reliable information on age is absent, no relationship between the deposition and some hypothetical period of base-level change can be established.

The long-term (up to 1.4 Myr) cycle of base-level change, recorded by the deposition of unit 2, is an order of magnitude longer than the cycles recorded in the overlying units (Table 1). The fact that the thickness of unit 2 is commonly close to that of units 4–6B combined (total time of deposition = 400 kyr; Table 1 and below) illustrates the importance of compaction in modifying the preserved thickness. Considering compaction ratios and peat growth rates, Kirschbaum & McCabe (1992) concluded that coals and carbonaceous muds in strata equivalent to unit 2 may represent up to 90% of the time of deposition. A number of relative base-level changes of higher frequency may have influenced the deposition of unit 2, but their understanding will require a detailed correlation of the numerous coals and palaeosols within the unit.

The *M. mosbyense* biozone, which comprises units 4–6B, represents 0.4 Myr or less, based on Obradovich (1993). The simplistic assumption that units 4–6B each represent equal increments of time gives a maximum average frequency of relative sea-level change of 100 kyr per unit. Similar periodicities, suggesting the involvement of Milankovitch forcing in relative sea-level change, were found by Plint (1996) in the Early to Middle Cenomanian Dunvegan Formation of Western Canada (= 150 kyr) and by Uličný & Špičáková (1996) in the Middle to Late Cenomanian of Bohemia (100–140 kyr). Based on the Ar/Ar ages from coal zone 3 (= 94.5 Ma), units 3A and 3B represent a similar frequency of relative sea-level changes (Table 1).

The magnitudes of the relative sea-level fluctuations are difficult to determine exactly, but the comparison of palaeobathymetric interpretations of facies juxtaposed across the bounding surfaces of units 3A to 6B allows an estimate of 10–20 m, possibly up to 30 m at the 5/6B sequence boundary.

**Regional correlation of Dakota units**

Elder et al. (1994) established regional correlation between carbonate-shale couplets in the pelagic rhythmites of the central part of the Western Interior and the nearshore ‘parasequences’
in southern Utah (including the ‘upper Dakota’ of the Kaiparowits area). Such correlation seems to support the above hypothesis of Milankovitch forcing of relative sea-level changes. However, part of the correlation by Elder et al. (1994) is based on recognition of only two ‘parasequences’ in the ‘upper Dakota’ in the Kaiparowits region: units 6A and 6B are considered parts of their ‘parasequence 2’ and, in the SE Kaiparowits area, marine mudstones of unit 5 are treated as a downdip continuation of unit 4 (their ‘parasequence 1’). Also, Elder & Kirkland (1993) do not recognize a relative sea-level fall around the M. mosbyense/S. gracile boundary.

Two linked problems arise when comparing the ‘parasequences’ of Elder et al. (1994) and the units 4–6B presented in this study. First, this study shows that more than two units separated by flooding surfaces occur in the ‘upper Dakota’ in the southern Kaiparowits area, and thus the correlation of individual surfaces to concretion and limestone beds across some 200 km between the Kaiparowits area and the nearest offshore section in Elder et al. (1994), as well as the number of cycles correlated, should be revised. Secondly, the evidence for sequence boundaries, especially 5/6A and 6A/6B, presented here implies that the theoretical concept of correlating the nearshore ‘parasequence’ flooding surfaces to distal basinal rhythms should be modified to include also the response to relative sea-level fall and lowstand deposition in the distal part of the basin (e.g. considering the role of skeletal limestones; cf. Sageman, 1996).

Regional and local tectonic controls on depositional style

The main control on the depositional style of the Dakota Formation was regional flexural subsidence within the foreland basin. The overall north-westward increase in total thickness of the Dakota Formation reflects the position of the study region within the foredeep depozone of the foreland basin system (sensu DeCelles & Giles, 1996) during the Cenomanian, where it remained until the Santonian (cf. Eaton & Nations, 1991; Leithold, 1994). In the region west of the Paunsaugunt Fault, the total thickness of strata coeval to units 1–6B reaches more than 300 m. Gustason (1989) reported an incremental westward increase in thickness of the Dakota across prominent fault zones, such as the Paunsaugunt Fault, and concluded that pre-existing basement fractures, approximately parallel to the orogenic front, were reactivated as normal faults and added a local component of subsidence to the regional flexural pattern. Schwans (1995) termed the portion of the foreland basin characterized by this subsidence pattern a ‘proximal zone’ of the foreland basin.

The study region, however, is located east of the Paunsaugunt Fault and belongs to the ‘distal zone’ of the foredeep depozone sensu Schwans (1995), where similar rapid lateral changes in thickness are far less pronounced. In the Henrieville area, abrupt changes in thickness of most units and local erosion of most of unit 4 were recorded in sections spaced 700 m apart across the NNE-striking Dry Creek valley, which separates Sections 3 and 4. This valley probably follows a basement fault roughly parallel to the Paunsaugunt and other major faults. The erosion of unit 4 in Section 4 (Fig. 2) is interpreted as the result of a coincidence of transgressive erosion and minor local uplift along a reactivated basement fault. Elsewhere, similar local changes in thickness of individual units are associated with differential compaction, namely of units 2, 3A and 3B, where channel-fill sandstones laterally interfinger with mudrocks and coals. On short time scales, compaction rates may have been more important in creating accommodation than the relatively slow tectonic subsidence.

Abundant deformational structures in unit 2 have been interpreted by Kirschbaum & McCabe (1992) as sediment liquefaction resulting from seismic activity. However, these phenomena occur throughout unit 2, and it is difficult to link them to particular faults; a more likely explanation of liquefaction is high sedimentation rates in the fluvial channels. Along the East Kaibab Monocline, sandstone bodies generated by slumping on the shoreface of unit 4 are interpreted as recording local seismic activity (Figs 2 and 8c). Also, the clastic dykes in unit 6A suggest involvement of seismic shocks in this area (Fig. 9b); they are downward injected, do not form polygonal patterns and penetrate various lithologies, which implies that coal compaction alone was probably not sufficient to cause their formation (cf. discussion in Tanner, 1998, and references therein). The precursor of today’s East Kaibab Monocline was apparently a fault parallel to and with a similar history to the Paunsaugunt, Sevier and Hurricane Faults in southern Utah (cf. Picha, 1986).
Causes of high- and low-frequency sea-level changes

Based on the periodicity estimates above, the Dakota units fall into two distinct groups: (i) units 1 and 2, representing depositional sequences formed in fluvial environments and recording presumably long-term (1.5 Myr for unit 2, unknown for unit 1) base-level changes; and (ii) units 3A to 6B, interpreted as high-frequency sequences formed in response to >100 kyr relative sea-level fluctuations. The overall stacking of the Dakota units results from a combination of low- and high-frequency relative sea-level fluctuations (Fig. 12). However, no attempt is made here to establish hierarchies of sequence-stratigraphic subdivisions such as sequence and parasequence sets of various orders sensu Mitchum & Van Wagoner (1991).

For unit 1, the onset of flexural subsidence was the most important causal mechanism, although a possible involvement of eustasy in the base-level rise causing the filling of valleys cannot be ruled out. As long as the duration of deposition remains unknown, the 1/2 sequence boundary may be explained as a response either to a eustatic fall or to crustal relaxation after a temporary cessation of thrusting activity. The marked base-level rise recorded at the base of unit 2 was probably linked to a eustatic rise, but it may have been assisted by accelerated back-taulting of the coastal plain, which helped to trap the coarser clastics closer to the source during the beginning of deposition. If the eustatic explanation is adopted, unit 2 would correspond to a long-term cycle analogous to the ‘third-order’ cycles of Vail et al. (1991). Individual widespread coal zones and palaeosols within unit 2 may represent a higher frequency signal of base-level changes.

For units 3A to 6B, eustasy is believed to have been the cause of the relative sea-level fluctuations of >10±20 m magnitude. All units occur consistently across the study area, >65±70 km along depositional dip, although their thicknesses vary in response to regional changes in subsidence rate, local syndepositional faulting, local erosion during sea-level falls and differential compaction. The estimated periodicity of the relative sea-level changes, >100 kyr, falls within the range of the Milankovitch-band orbital cycles, the eccentricity cycle being the closest match, which favours climate-driven eustasy as the mechanism responsible for the relative sea-level changes.

Some authors interpret flooding events of similar frequency as reflecting periods of punctuated thrusting (Kamola & Huntoon, 1995). Peper (1993) also includes changes in intraplate stress level in the group of mechanisms potentially causing high-frequency sea-level changes. However, the fact that the thicknesses of all

Fig. 12. Summary of the depositional history and sequence stratigraphy of the Dakota Formation in the Kaiparowits region, southern Utah. Comments in text.

© 1999 International Association of Sedimentologists, Sedimentology, 46, 807–836
high-frequency units are of similar orders of magnitude suggests that the mechanism driving the sea-level changes was cyclic, which supports the eustatic forcing, rather than thrust events or changes in compressional stress.

Firm evidence for the above interpretation of a eustatic control on relative sea-level changes would, of course, be inter-regional or global correlation. Correlation of high-frequency sea-level changes in the Cenomanian is still very complicated but, for certain intervals of the Late Cenomanian, the eustatic nature of low-frequency cyclicity is well documented by biostratigraphic correlation. The 5/6 sequence boundary in the Kaiparowits region correlates to a marked sea-level fall at the end of the M. mosbyense biozone (*Calycoceras naviculare* in Europe) documented, for example, from the Twowells Sandstone, New Mexico (Mellere, 1994), British Chalk (Owen, 1996), Tunisia (Robaszyński et al., 1993) or Central Europe (the ‘sub-Plenus’ sea-level fall; Voigt et al., 1992; Uličný et al., 1997). Although stacking patterns in siliciclastic successions are sensitive to local rates of sediment input, the overall retrogradational stacking of high-frequency sequences between unit 2 and the base of the Tropic Shale shows a marked similarity to the record of late Middle–early Late Cenomanian, long-term sea-level rise in central Europe (Uličný & Spičáková, 1996), punctuated by short-term relative falls in sea level.

It is not the purpose of this paper to discuss the possible driving mechanisms of high-frequency eustatic sea-level changes, but it is worth noting that the global mechanisms causing eustatic fluctuations of such a short periodicity during the Middle–Late Cenomanian, a time interval close to the mid-Cretaceous greenhouse climatic peak (Jenkyns et al., 1994), are still controversial. Although Cenomanian glacioeustasy producing 10–30 m of sea-level change is still difficult to accept for many researchers (cf. discussion in Christie-Blick & Driscoll, 1995), some recent computer models produce seasonal or even permanent continental ice close to the South Pole and thus raise the possibility of limited glacioeustasy even during the Cenomanian. Alternative mechanisms could be changes in the volume of groundwater stored on continents, which could be responsible for sea-level changes of up to 20 m over periods of 10–100 kyr, or ocean steric (thermohaline) volume changes (Revelle, 1990).

**CONCLUSIONS**

1 The tectonic controls on the sedimentation of the Dakota Formation included:

(a) eastward migration of a forebulge causing the tilt of pre-Cretaceous basement strata and the incision of a westward-deepening valley system into them;

(b) flexure-dominated subsidence regime of the foredeep depozone, which probably started during the early Cenomanian and controlled the long-term subsidence rate and its dip-parallel accommodation profile;

(c) reactivation of basement faults, which led to an abrupt westward increase in subsidence rate across some faults, as well as minor uplifts, which caused local erosion and submarine slumping.

2 Eustatic sea-level changes had the following effects:

(a) formation of regionally extensive flooding surfaces, mostly coinciding with previously formed sequence boundaries, which separate the Dakota units;

(b) the preserved thicknesses of individual Dakota units vary because of regional and local subsidence rate variations as well as because of local incision on sequence boundaries, but all units occur consistently across the study region;

(c) the fluvial strata of units 1 and 2 are capped by erosional surfaces, interpreted as updip equivalents of sequence boundaries; the coal zones at the base and within unit 2 may represent a response to higher frequency flooding events;

(d) the shallow-marine units, interpreted as parasequences by earlier authors, can be divided into facies-based systems tracts and show signs of subaerial exposure on their boundaries, which allows them to be interpreted as high-frequency sequences;

(e) the overall stacking pattern of the high-frequency sequences is retrogradational, reflecting a long-term relative sea-level rise, punctuated by brief periods of relative sea-level falls, including a relatively major fall near the end of the *M. mosbyense* Zone; this is similar to the record of Cenomanian relative sea-level change in other regions, e.g. central Europe or northern Africa, suggesting that eustasy also played an important role in governing the low-frequency relative sea-level changes.

3 The driving mechanisms of the eustatic changes are controversial. The high-frequency relative
sea-level fluctuations of 100 kyr periodicity and 10–20 m magnitude were probably related to Milankovitch-band, climate-driven eustasy. Either minor glacioeustatic fluctuations, superimposed on the overall greenhouse climate of the mid-Cretaceous, or mechanisms such as the fluctuations in groundwater volume on continents, or thermal expansion and contraction of seawater, may have controlled the high-frequency eustatic fluctuations.

ACKNOWLEDGEMENTS

Special thanks are due to Jeff Eaton for having introduced me to the field area, for generous logistical support that made the project possible, for fruitful discussions and camaraderie during the fieldwork periods. Criticisms and suggestions by Gary Nichols helped to improve the early draft of the paper, and Gus Gustason is thanked for an informal discussion of different views on Dakota sequence stratigraphy. I am grateful to the reviewers, Peter Schwans and Mark Kirschbaum, and to Sedimentology editor Guy Plint for incisive reviews and suggestions that improved the paper significantly. M. Kirschbaum is thanked for providing unpublished data on bentonite Ar/Ar ages. Marcela Svobodova provided unpublished palynological data. Rita Ulicña was the most patient and enthusiastic field assistant. The fieldwork was supported financially by GSA Foundation grant no. 5071-92 and by donors to the Czechoslovak Charta 77 Foundation. This study was initiated during part of the author’s doctoral study at the University of Illinois, Champaign-Urbana.

REFERENCES


Elder, W.P. and Kirkland, J.I. (1994) Cretaceous Paleo-

geography of the southern Western Interior region. In: Mesozoic Systems of the Rocky Mountain Region, USA (Ed. by M.V. Caputo, J.A. Peterson and K.J. Franczyk), SEPM Rocky Mountain Section, 415–440.


Kamola, D. and Van Wagoner, J.C. (1995) Stratigraphy and facies architecture of parasequences with examples from the Spring Canyon Member, Blackhawk Formation, Utah. In: Sequence Stratigraphy of Fore-

Kirschbaum, M.A. and McCabe, P.J. (1992) Controls on the accumulation of coal and on the development of anastomosed fluvial systems in the Cretaceous Da- 

kota Formation of southern Utah. Sedimentology, 39, 581–598.
Mellere, D. (1994) Sequential development of an estu- 

Miall, A.D. (1985) Architectural-element analysis: a new method of facies analysis applied to fluvial de- 
Mitchum, R.M. and Van Wagoner, J.C. (1991) High-

frequency sequences and their stacking patterns: 

sequence-stratigraphic evidence of high-frequency eustatic cycles. Sedim. Geol., 70, 131–160.


Owen, D. (1996) Interbasinal correlation of the Cen-


Record in Foreland Basins. PhD Thesis, Vrije Univer-

siteit, Amsterdam.
Peterson, F. (1969) Cretaceous sedimentation and tec-


© 1999 International Association of Sedimentologists, Sedimentology, 46, 807–836


Wolfe, D.G. (1989) The Stratigraphy and Palaeoenvironments of Middle Cretaceous Strata along the
Central Arizona–New Mexico Border. MSc Thesis, University of Colorado, Boulder.


*Manuscript received 11 November 1996; revision accepted 13 January 1999.*