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Fluvial–estuarine transitions in fluvial-dominated successions: examples from the Lower Pennsylvanian of the Central Appalachian Basin

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ABSTRACT

Early Pennsylvanian sedimentation in the Central Appalachian Basin was dominated by the successive development of south- to southwest-flowing, low-sinuosity streams in broad, longitudinal braidplains, which deposited a series of quartzarenites. Successive quartzarenite belts are locally separated by grey shales with brachiopods and other body fossils interpreted to represent marine- to brackish-water facies. Local features indicative of tidal sedimentation occur between fluvial facies and the marine- to brackish-water shales. Tidal features occur in transgressive successions between fluvial and overlying marine to brackish-water shales, and significant wave-generated features are absent, indicating that tide-dominated estuaries developed during transgressions. The boundary between fluvial facies and recognizable estuarine tidal facies represents a fluvial–estuarine transition. Tidal sedimentary features in the fluvial–estuarine transition, however, can be subtle, because upper estuarine channels may record only the most headward tidal effects in an otherwise fluvially dominated system.

Some of the possible tidal indicators noted in upper estuarine channel facies include local occurrences of opposing palaeoflow indicators, non-cyclic rhythmites, lenticular bedding, small reversing ripples on the crests of underlying current ripples, sigmoidal cross-strata, cross-strata with rising troughs, thick–thin laminae pairs and bundled laminae in ripple cross-lamination. None of these features is diagnostic for tidal sedimentation. Where multiple tidal indicators are found within otherwise fluvial facies, within a probable transgressive succession, interpretation of an upper estuarine channel facies becomes more tenable.

Recognition of fluvial–estuarine transitions is important in fluvial-dominated basins, especially in the upper reaches of longitudinal basins, because the transitions may be the only evidence of correlative down-dip marine flooding surfaces. Identification of the transition zone facilitates the distinction between lowstand and transgressive systems tracts. In turn, such sequence analyses can increase the potential for predicating lateral changes in fluvial channel continuity, and vertical changes in porosity and permeability characteristic of lithological changes from fluvial to estuarine facies, both of which are important in exploration for hydrocarbons.

INTRODUCTION

The location and character of the transition between fluvial and estuarine strata are somewhat dependent on the definition of estuary used. Estuaries can be defined based on geomorphological, hydrological, biological, and chemical attributes (summarized in Perillo, 1995). Estuaries can be defined as being partly enclosed by land, having an open connection to the sea and showing significant salinity dilution landward (e.g. Pritchard, 1967). A more geological definition describes estuaries as drowned river valleys that receive sediment from marine and fluvial sources, which are influenced by tide, wave and fluvial processes (Zaitlin & Shultz, 1990; Dalrymple *et al.*, 1992). Using the latter definition, the landward limit of the inner, river-dominated part of the estuary will extend farther inland than a definition based purely on salinity (Fig. 1) because tidal influences tend to occur farther headward than salinity influences (Dionne, 1963; Nichols & Briggs, 1985; Dalrymple *et al.*, 1992; Perillo, 1995).

Many tide-dominated estuaries can be divided into three parts based on salinity and the hydrological effects of wave, tidal and current processes, although form varies owing to differences in tidal range, and the relative effects of tides, waves, and rivers (Dionne, 1963; Dalrymple *et al.*, 1991, 1992; Allen, 1991). Those that have three parts consist of:

1 a marine-influenced lower estuary where waves and tides interact at the outer limit of estuarine sands, dominated by tidal sand bars and upper-flow-regime sand flats;

2 a zone of salt- and freshwater mixing called the middle estuary, which is dominated by tidal processes and is often muddy;

3 an upper estuary or fluvial-dominant zone marked by fresh water and the landward limit of tidally influenced sediments (Fig. 1; Dionne, 1963; Dalrymple *et al.*, 1991, 1992; Allen, 1991; Perillo, 1995).

The upper estuary encompasses the fluvial–estuarine transition, which is the focus of this study.

In the tide-dominant estuary model of Dalrymple *et al.* (1992), the fluvial–estuarine transition occurs within the upper estuarine channel, the boundary of which is marked by the

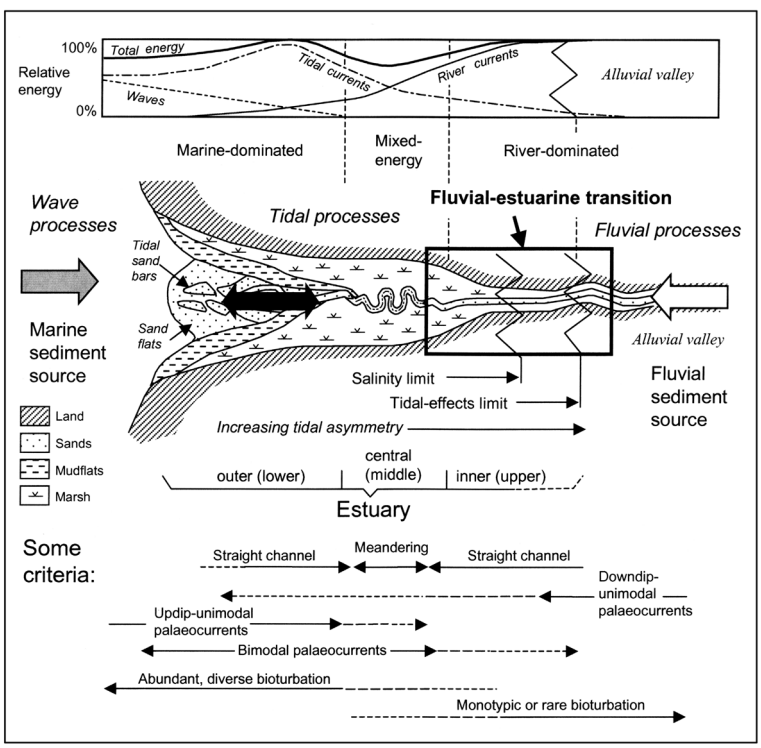

Fig. 1 Diagram showing typical features of a tide-dominated estuary (after Dalrymple *et al.*, 1992; Perillo, 1995). The fluvial–estuarine transition is highlighted in the black rectangle. No scale implied.

headward limit of tidal effects on sedimentation. Limited studies of modern fluvial–estuarine transitions in tide-dominated estuaries indicate relatively narrow transition zones between fluvial and tidal facies (e.g. Smith, 1988; Lanier & Tessier, 1998). A certain amount of temporal variability in the relative influence of tides, river discharge, waves, climates, rates of transgression, or net sediment supply could cause lateral migration of the transition based on the relative contribution of influencing factors. In addition, there are several modern tide-dominated estuaries that are associated with rivers with large discharge (Wells, 1995). In large, tidally influenced rivers tidal influences extend well up-river. For example, macrotidal conditions exist in the St Lawrence River 500 km inland from the Gulf of St Lawrence (Archer, 1995). In the Amazon River, flood tides reach 800 km inland and may influence sedimentation at least 200 km inland (Archer, this volume, pp. 17–39). If these rivers were transgressed, and estuaries developed within the drowned river system, the potential length of the upper estuary, if defined by the most landward occurrence of tidal sedimentation, could extend much farther inland than is typical for what is commonly considered an estuary.

Ancient estuarine strata have been interpreted in several basins, generally based on identification of a transgressive succession, and bimodal current indicators and/or trace-fossil suites suggestive of mixed or changing salinities (Rahmani, 1980; Diemer & Bridge, 1988; Smith, 1988; Shanley *et al.*, 1992; Archer *et al.*, 1994; Kvale & Barnhill, 1994; Greb & Chesnut, 1996; Els & Mayer, 1998; Lanier & Tessier, 1998). In ancient fluvial-dominated successions, sequence-stratigraphy analysis of fluvial successions is difficult where marine- and brackish-water indicators are absent because transgressive and maximum flooding surfaces cannot be defined (Emery & Myers, 1996). The purpose of this paper is to describe fluvial–estuarine transitions and their variability in fluvial-dominated Lower Pennsylvanian strata of the Central Appalachian Basin within a facies framework based on a geological definition of tide-dominated estuaries (*sensu* Dalrymple *et al.*, 1992). Recognition of the variability in this transitional facies may aid in the recognition of similar facies in fluvial-dominated successions in other basins.

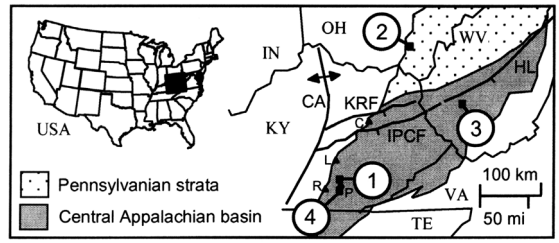

Fig. 2 Location of the Central Appalachian Basin showing major structures and study locations (numbered 1–4): CA, Cincinnati Arch; KRF, Kentucky River Fault Zone; IPCF, Irvine-Paint Creek Fault Zone; HL, hingeline; triangles, locations discussed in Greb & Chesnut (1996); C, Corbin Sandstone; L, Livingston palaeovalley; P, Pine Creek Sandstone; R, Rockcastle Sandstone; IN, Indiana; KY, Kentucky; OH, Ohio; TN, Tennessee; WV, West Virginia; VA, Virginia. *Pine Creek Sandstone is the informal name for the upper Bee Rock Sandstone in the study area.

STRATIGRAPHY AND PALAEOGEOGRAPHY

The Central Appalachian Basin is a foreland basin, which covers parts of Tennessee, Kentucky, West Virginia and Ohio. The basin is bordered on the west by the Cincinnati Arch (CA in Fig. 2) and is elongated in a NE–SW orientation (Fig. 2). The Kentucky River Fault System (KRF in Fig. 2) and Irvine–Paint Creek Fault System (IPCF in Fig. 2) in Kentucky, and a structural hingeline (HL in Fig. 2) in West Virginia, represent a structural hingeline on the northern margin of the basin. Strata thicken significantly south of this hingeline and continue to thicken to the south-east toward the preserved axis of the foreland basin in south-eastern Kentucky and western Virginia (Fig. 3A & B; Chesnut, 1992; Greb *et al.*, 2002).

The stratigraphical nomenclature of Lower Pennsylvanian strata is shown in Fig. 4.

Lower Pennsylvanian strata are dominated by thick quartzarenite tongues (Figs 3B & 4). In plan view, the quartzarenite tongues occur as elongate belts, which successively overlap the basin margin to the west and north (Fig. 3B). Successive quartzarenite belts both truncate and interfinger with coal-bearing strata to the south-east (Figs 3B & 4). Lower Pennsylvanian quartzarenites were previously interpreted as beach barriers (e.g. Ferm *et al.*, 1971), tidal strait deposits (e.g. Cecil & Englund, 1989) and fluvial facies (e.g. Potter &

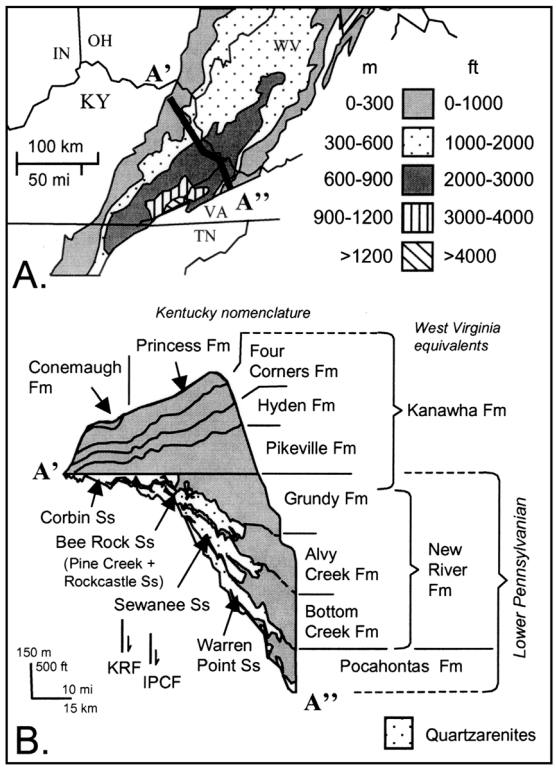

Fig. 3 Pennsylvanian strata in the Central Appalachian Basin. (A) Isopach showing south-eastward thickening and position of cross-section A′–A″. (B) Cross-section of Pennsylvanian strata showing the westward overlapping of quartzarenite belts on the western margin of the basin (modified from Chesnut, 1992). Datum is the Betsie Shale at the Lower–Middle Pennsylvanian boundary. KRF, Kentucky River Fault Zone; IPCF, Irvine–Paint Creek Fault Zone; IN, Indiana; KY, Kentucky; OH, Ohio; TN, Tennessee; WV, West Virginia; VA, Virginia.

Siever, 1956). Recent investigations have favoured fluvial deposition with most investigators interpreting the thick quartzarenites as braidplain deposits (Bement, 1976; Rice, 1984, 1985; Chesnut, 1992; Rice & Schwietering, 1988; Wizevich, 1991; Barnhill, 1994). Greb & Chesnut (1996) showed that although many of the quartzarenite belts are dominated by fluvial facies, tidal facies occur locally toward the top of each unit.

Figure 5 is a generalized view of Early Pennsylvanian palaeogeography in the Central Appalachian Basin and surrounding areas. Longitudinal, braided-stream trunk systems were developed on the western side of the basin. Quartz-pebble-bearing quartzarenites deposited within the braidplains had a north-eastern source (Donaldson *et al.*, 1985; Chesnut, 1992, 1994; Archer & Greb, 1995; Greb & Chesnut, 1996). Lithic arenites and sublitharenites were deposited on the eastern side of the basin within transverse drainages with a source to the south-east (Ferm, 1974; Englund, 1974; Houseknecht, 1980). These streams flowed toward the north-west, presumably draining into the longitudinal braidplains (Fig. 5). Greb & Chesnut (1996) noted that sedimentary structures typical of tidal sedimentation and local bioturbation were developed at the top of four different quartzarenites on the western side of the basin, which in turn were overlain by dark grey shales with marine fauna. These features were inferred to indicate Early Pennsylvanian transgressive successions. The identification of tidal sedimentary structures in these fluvial-dominated strata is important for the identification of transgressive successions as

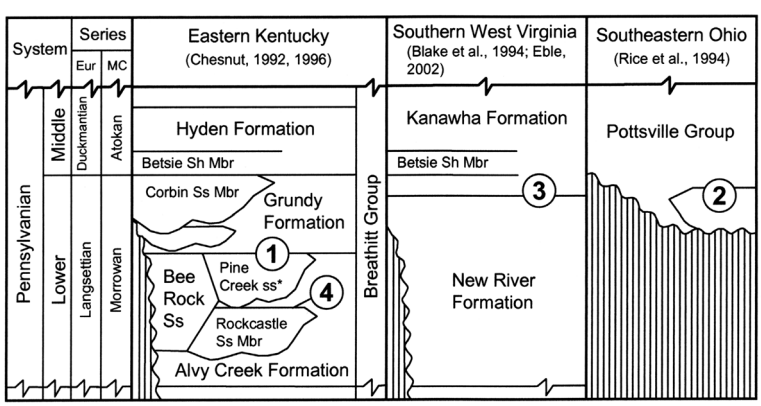

Fig. 4 Stratigraphical correlations of Lower and lower Middle Pennsylvanian strata in the Central Appalachian Basin. Numbers 1 to 4 refer to the stratigraphical positions of locations used in this study.

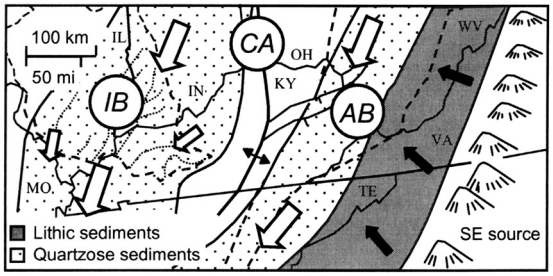

Fig. 5 Palaeogeographical map during the Early Pennsylvanian showing development of longitudinal and transverse drainages in the Central Appalachian Basin (AB). The Central Appalachian Basin was separated from the Illinois Basin (IB) by the Cincinnati Arch (CA). Quartzarenites occur in both basins and were sourced from the north-east (white arrows), whereas the source of lithic sediments in the Central Appalachian Basin was mostly to the south-east (black arrows) (modified from Chesnut, 1992; Greb & Chesnut, 1996). IN, Indiana; KY, Kentucky; OH, Ohio; TN, Tennessee; WV, West Virginia; VA, Virginia.

marine fauna are rare in Lower Pennsylvanian strata of the basin (e.g. Chesnut, 1991).

Early Pennsylvanian seas were interpreted to have transgressed from the south-west up the longitudinal drainage belts (Greb & Chesnut, 1996). In one example, the tidal facies succeeded fluvial facies within the Livingston Palaeovalley (L in Fig. 2), a bedrock-confined palaeovalley, indicating estuarine conditions during transgression. Wave-formed structures are rare to absent in these successions, so that if estuaries were developed as part of the succession, they were tide-dominated. In other parts of the basin, burrowed, compound cross-bedded sandstones in the Kanawha Formation of West Virginia below the Betsie Shale have been interpreted as tidal sand-bar facies (Martino, 1996; Hamrick, 1996). Such sand bar facies are typical of lower estuarine settings in modern macrotidal estuaries (Dalrymple *et al.*, 1992), which lends support to the application of tide-dominated estuary models.

Herein, facies are interpreted for Lower Pennsylvanian transgressive successions and examples are illustrated from different parts of the basin. Examples of the variety of vertical successions are shown in order to illustrate differences in fluvial–estuarine transitions in these fluvial-dominated sequences. Well-defined transitions are compared with more subtle transitions in order to determine the varied appearance of fluvial–estuarine transitions in individual vertical sections, which may aid in interpreting similar transitions in other basins, or in less well-exposed areas. Documentation of the variability in fluvial–estuarine transitions is important in fluvial-dominated successions because these may be the only facies that can be used to infer the up-dip equivalents of marine flooding surfaces, which are important for regional sequence analyses.

COMMON LITHOFACIES

Typical facies at the top of Lower Pennsylvanian quartzarenite belts in the Central Appalachian Basin are shown in Table 1 and described below.

Fluvial cross-bedded sandstone facies

Each of the Lower Pennsylvanian quartzarenite belts is dominated by coarse- to fine-grained, cross-bedded sandstones, arranged in a variety of downstream accreting-bar and channel macroforms (Wizevich, 1991, 1992, 1993; Barnhill, 1994). Scour-based, cross-bedded sandstones may fine upward into ripple-bedded sandstones. Palaeocurrent modes are unimodal to the south and south-west. In general, the cross-bedded sandstones have mostly been interpreted as fluvial deposits (Table 1). For the purpose of analysing fluvial–estuarine transitions, herein, cross-bedded sandstones with unimodal current indicators orientated in a down-dip direction within the quartzarenite belts are interpreted as fluvial cross-bedded sandstone facies. An attempt is not made to interpret individual macroforms or lithofacies within the all-inclusive fluvial sandstone facies for the purpose of defining fluvial–estuarine transitions herein.

Peat-mire coal facies

Fluvial cross-bedded sandstone facies may be immediately overlain or may fine upwards into thin rooted horizons overlain by coals. Coal beds occur above each of the major quartzarenite formations (Chesnut, 1992), and are widely accepted as the accumulations of topogenous to ombrogenous peat mires (Table 1).

Table 1 Facies used in this study.

Facies	Description	Physical structures	Palaeoflow	Interpretation
Fluvial cross-bedded sandstone channel	Scour-based, cross-bedded, fine to coarse-grained quartzose sandstones, locally conglomeratic with quartz pebbles. Generally multistorey deposits	Planar and trough cross-bedding common, local current-ripple lamination. Channel and down-stream-accreting, compound macroforms common. Occasional fossil plant debris	Unimodal down-dip	Fluvial bedload (braided) channels (Bement, 1976; Rice, 1984; Wizevitch, 1991, 1992, 1993; Greb & Chesnut, 1996)
Peat-mire coal	Thin (< 1.5 m) coal beds. Underlain by rooted palaeosols or seat earths	Layers of different texture and brightness. Underlain by root structures, seat earths/ palaeosols	No flow	Different kinds of peat mires (Donaldson *et al.*, 1985; Eble *et al.*, 1991; Eble, 1996)
Marine- to brackish-water shale	Dark grey to medium grey laminated, silty shales. Sharp-based and coarsening upward	Planar lamination. Lenticular, flaser, and wavy bedding where coarsening upward. Marine- to brackish-water body and trace fossils toward base	Few flow indicators	Marine- to brackish-water seaways (Chesnut, 1991, 1992; Martino, 1994, 1996)
Heterolithic tidal flat	Interlaminated silty shales, siltstones, and very fine- to fine sandstones. May fine upward. Only defined where bioturbated or tidal-bedding features are noted	Flaser to lenticular lamination and rhythmic lamination. Local climbing-ripple lamination. Biogenic structures and soft-sediment deformation rare to uncommon. Locally rooted at top	Unimodal down-dip to variable or bimodal	Intertidal flats (Greb & Chesnut, 1996), similar to modern flats (Klein, 1977; Terwindt, 1988; Nio & Yang, 1991; Dalrymple *et al.*, 1991)
Upper estuarine cross-bedded sandstone channel	Scour-based, cross-bedded, fine to medium-grained quartzose sandstones, locally interbedded with ripple-bedded sandstone. Only defined where tidal bedding features are noted	Bedding similar to fluvial cross-bedded facies, but cross-beds exhibit local sigmoidal bedding, bundled foresets, shale-draped foresets, thick–thin foreset couplets or rhythmic lamination. Generally more shale inter-lamination than fluvial channel facies. Bioturbation absent to rare	Unimodal down-dip to variable with rare to uncommon up-dip modes preserved locally	Fluvial-dominant, straight channels that become more tidal seaward characterize the upper estuary (Allen, 1991; Dalrymple *et al.*, 1992). Shale-draped bedding more common than in fluvial channel (Rahmani, 1988; Smith, 1988)
Middle estuarine heterolithic channel	Scour-based, interbedded silty shales, siltstones, and very fine- to fine sandstones as inclined stratification with lateral accretion. May be interstratified with cross-bedded sandstones	Parallel laminated to rhythmically laminated, local flaser to lenticular bedding, soft-sediment deformation, and cross-bedding as in the upper estuarine channel facies. Bioturbation rare to abundant. Top may be rooted	Unimodal down-dip, unimodal up-dip, bimodal, or variable	Heterolithic meandering channels common in the middle estuary of tide-dominant estuaries (Allen, 1991; Dalrymple *et al.*, 1992) and tidal channels (Thomas *et al.*, 1987)

Marine- to brackish-water grey-shale facies

Fluvial facies at the top of the quartzarenites, or the coals atop the quartzarenites, are locally succeeded by dark grey shales. These shales may contain local marine- or brackish-water fauna towards their base. Shales coarsen upward into laminated siltstones and sandstones. Similar Middle Pennsylvanian shales are widely interpreted as marine zones deposited in seaways during transgressions (Table 1). The bases of Middle Pennsylvanian marine zones have been interpreted as marine flooding surfaces and have been used to define third-order sequences (Aitken & Flint, 1995). At least one Lower Pennsylvanian shale (above the Pine Creek Sandstone, Fig. 4) has a fossiliferous siderite layer at its base, also interpreted as a marine flooding surface (P in Fig. 2, Greb & Chesnut, 1996). The fact that marine fauna are much lower in diversity and abundance in Lower Pennsylvanian transgressive grey shales than lithologically similar Middle Pennsylvanian transgressive grey shales does not preclude the Lower Pennsylvanian grey shales from representing transgressive deposits, but indicates that Early Pennsylvanian transgressions were more diluted or less extensive than their Middle Pennsylvanian counterparts. The coarsening upward part of the facies is interpreted to represent progradation of coastal–deltaic facies during highstands, similar to interpretations for the Middle Pennsylvanian.

Middle estuarine heterolithic channel facies

Greb & Chesnut (1996) identified scour-based, bioturbated, inclined heterolithic strata (IHS of Thomas *et al.*, 1987) at the top of the Rockcastle Sandstone (R in Fig. 2), one of the Lower Pennsylvanian quartzarenites (Figs 3B & 4). Inclined heterolithic strata form in meandering mixed-load channels in fluvial and tidal environments (Thomas *et al.*, 1987). Where non-bioturbated IHS occurs in vertical succession above the fluvial cross-bedded sandstone facies, the IHS could represent a change in channel morphology from bedload fluvial-braided streams to mixed-load meandering streams. Such a change could occur as rising base level decreased stream gradient. In cases where the IHS are bioturbated with brackish- to marine-trace fossils, a tidal channel origin is probable (Table 1). Likewise, shaly IHS, with shales draped across the macroform, may be more common in tidal IHS than fluvial IHS. Shale layers extending from the upper to lower margins of IHS have been used to infer tidal influences in other deposits (Rahmani, 1988). Deposition of mud on fluvial point bars occurs where suspended loads are high and system energy is relatively low, but generally is restricted to the upper portion of the point bar (Collinson, 1996). Such mud layers become more common and more extensive where even minor tidal influence occurs (Smith, 1988; Allen, 1991).

In the depositional model developed by Dalrymple *et al.* (1992) for tide-dominated estuaries, the alluvial channel enters the inner estuary as a tidally influenced, but fluvial-dominated straight channel. There is a net seaward transport in the channel owing to the prevalence of fluvial currents. Mud-rich meandering channels are more likely to develop in the central, low-energy zone of tide-dominated estuaries (Smith, 1988; Allen, 1991; Dalrymple *et al.*, 1991, 1992). Such facies are a distinctive component of the tide-dominated estuarine model, and are not found in wave-dominated estuaries.

Estuarine tidal-flat facies

Greb & Chesnut (1996) also noted local occurrences of bioturbated, parallel-laminated to ripple-laminated heterolithic strata at the top of one of the Lower Pennsylvanian quartzarenites (L in Fig. 2). Some of these facies contain rhythmic lamination typical of tidal facies, although non-cyclic rhythmites (*as defined by* Greb & Archer, 1998) dominate. Laminated heterolithic strata may grade upward into rooted strata capped by coals. In a transgressive succession, this type of bedding could be interpreted as tidal-flat deposits (Table 1). In modern, tide-dominated estuaries, tidal flats occur from the lower to upper estuary, and are often bounded landward by marshes (e.g. Dalrymple *et al.*, 1992). In general, tidal flats become sandier, exhibit increasing sedimentation rates, and become less bioturbated headward in the estuary (Dalrymple *et al.*, 1991, 1992). Rhythmites recording daily to neap–spring cyclicity are perhaps best recorded on tidal flats in the

inner or upper estuary where there is little bioturbation and increased tidal amplification owing to funnelling effects of the estuary (Dalrymple *et al.*, 1991, 1992; Tessier, 1993, 1998). Decreasing preservation potential of sedimentation within rhythmic lamination is noted higher on flats, which can lead to the deposition of non-cyclic, rather than cyclic tidal rhythmites (Archer, 1998; Tessier, 1998).

Upper estuarine sandstone channel facies

Upper estuarine channels near the transition zone with fluvial channels would be expected to be river-dominated straight channels, exhibiting some evidence of tidal sedimentation (Fig. 1 and Table 1). In Lower Pennsylvanian transgressive successions, down-dip-orientated planar cross-beds typical of the fluvial cross-bedded sandstone facies are locally interbedded with flaser- to lenticular-bedded sandstones that contain rare brackish-water trace fossils, non-cyclic rhythmites, and/or sedimentary structures with up-dip-orientated current indicators. Cross-beds may show increased foreset concavity, crude thick–thin foreset alternations, shale-draped foresets, sigmoidal-shaped foresets or rising trough levels. If bioturbation, rhythmic lamination or bedding indicates tidal effects, in an otherwise fluvial-dominated channel, the deposits are reinterpreted as upper estuarine sandstone channel facies. The headward limit of the upper estuary is the headward limit of tidal influence (Dalrymple *et al.*, 1992), so the headward limits of upper estuarine channels might exhibit only rare tidal signatures and be difficult to interpret.

TRANSGRESSIVE SUCCESSIONS

Examples of Lower Pennsylvanian transgressive successions at the tops of the quartzarenites are documented around the basin to show their varied appearance. In some cases, there is abundant evidence of tidal sedimentary features, which can be used to interpret an upper estuarine sandstone channel facies. In other cases, delineation of the facies is based upon more speculative evidence, and the facies is difficult to differentiate from underlying fluvial deposits.

Example 1

Greb & Chesnut (1996) previously used a vertical section at the top of the Pine Creek sandstone along Kentucky Highway 80 in Pulaski County to illustrate a gradual transition from fluvial to tidal bedding within previously inferred fluvial facies (location 1, Figs 2 & 6). Herein, strata within the transgressive succession are used as an example of a gradual change from fluvial to tidal deposition within a tidally dominated estuary framework.

Bedding within the lower two-thirds of the sandstone is dominated by unimodal, southwest-orientated cross-bedding (Fig. 7A) of the fluvial cross-bedded sandstone facies (Table 1). The upper part of the sandstone is shown in Fig. 6. At this location, there is a vertical shift from unimodal, down-dip-orientated palaeocurrents to variable and then up-dip-orientated palaeocurrents (Fig. 6), assigned to upper estuarine sandstone channel facies, middle estuarine heterolithic channel facies, and estuarine tidal-flat facies.

Upper estuarine channel facies at this location occur above two scours at the top of the Pine Creek sandstone. The lower scour fill is sheet form, fines upward (Fig. 7B), and exhibits a low angle of dip to the south-west. Sheets contain isolated trough cross-beds down-dip and are dominated by flaser bedding up-dip. Some ripples exhibit crude bundling of foreset laminae (alternating thicker and thinner sets) separated by shale drapes (Fig. 7C). Some ripples in the upper channel exhibit possible smaller scale ripples orientated in opposing directions on their crests or in troughs between ripples (Fig. 7D). Rare vertical burrows also occur but are not very distinct (Fig. 7D). Some down-dip-orientated foresets in isolated cross-beds (Fig. 7E) exhibit possible thick–thin alternations (Fig. 7F). A single up-dip-orientated cross-bed occurs at the top of the sandstone (Fig. 6).

The upper channel is incised sharply into the underlying sheet-form sandstones (Fig. 6). The upper channel is also dominated by ripple bedding and cross-bedding, but contains more parallel lamination and shale clasts than the underlying sheet sandstone. Asymmetric cross-beds have mostly up-dip orientations (Figs 6 & 8A–C). Some cross-beds exhibit crude bundling of

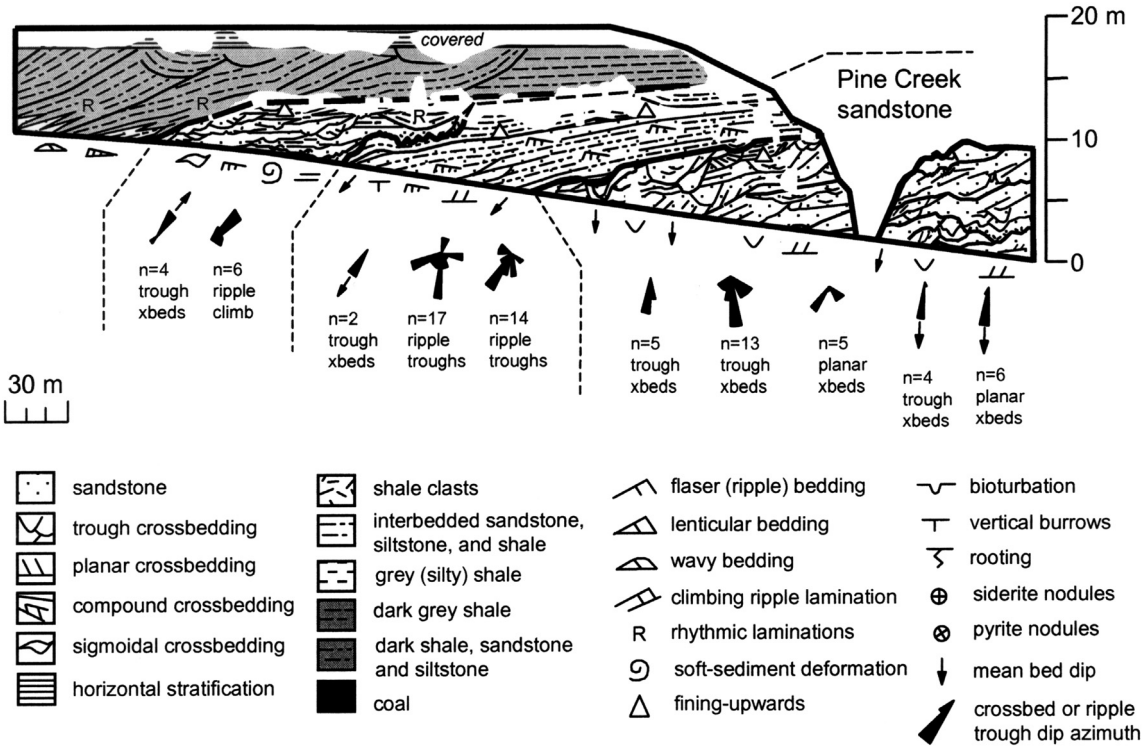

Fig. 6 Line drawing of upper Pine Creek sandstone at location 1, Kentucky Highway 80, east of the Rockcastle River, Laurel County, Kentucky.

thicker, sandy foresets, and thinner, shale-draped foresets (black arrows in Fig. 8A), thick–thin foreset alternations (Fig. 8B), laterally rising trough levels and sigmoidal foresets (Fig. 8C). Non-cyclic, rhythmic lamination occurs above a small scour at the top of the sandstone (Figs 6 & 8D). Crude thick–thin bundling occurs within climbing ripples within this scour (arrows in Fig. 8E).

The sandstone is sharply overlain by dark grey, shaly heterolithic strata correlated to the Dave Branch Shale (Figs 6 & 8F). At other locations marine fauna have been found in this shale (Chesnut, 1991; Greb & Chesnut, 1996). The lower part of the heterolithic facies here contains bundles of sandier and shalier strata (Fig. 8F & G). Sandier bundles consist of rhythmic lamination and lenticular bedding. Individual ripple laminae are separated by packets of 5–8-mm-scale rhythmic laminations (Fig. 8G). Some bundles show thickening of ripples toward the centre of bundles, and possible reversing crest ripples (black arrow in Fig. 8H).

Interpretation

The base of the upper estuarine sandstone channel facies is placed along the scour at the base of the sheet-form sandstones because this marks a major bedding change. Dalrymple *et al.* (1992) noted that facies contacts within estuarine successions were likely to coincide with erosional channel bases. Ripple cross-laminae azimuths are more varied than in the underlying fluvial facies, but reversing current indicators are rare (Fig. 6). Likewise, cross-beds are still mostly orientated down-dip, but show thick–thin foreset alternations. Thick–thin foreset alternation is a common feature of diurnal inequality in tides formed in semi-diurnal settings (Rahmani, 1988; Smith, 1988; DeBoer *et al.*, 1989; Kvale *et al.*, 1989; Nio & Yang, 1991). Visser (1980) considered persistent

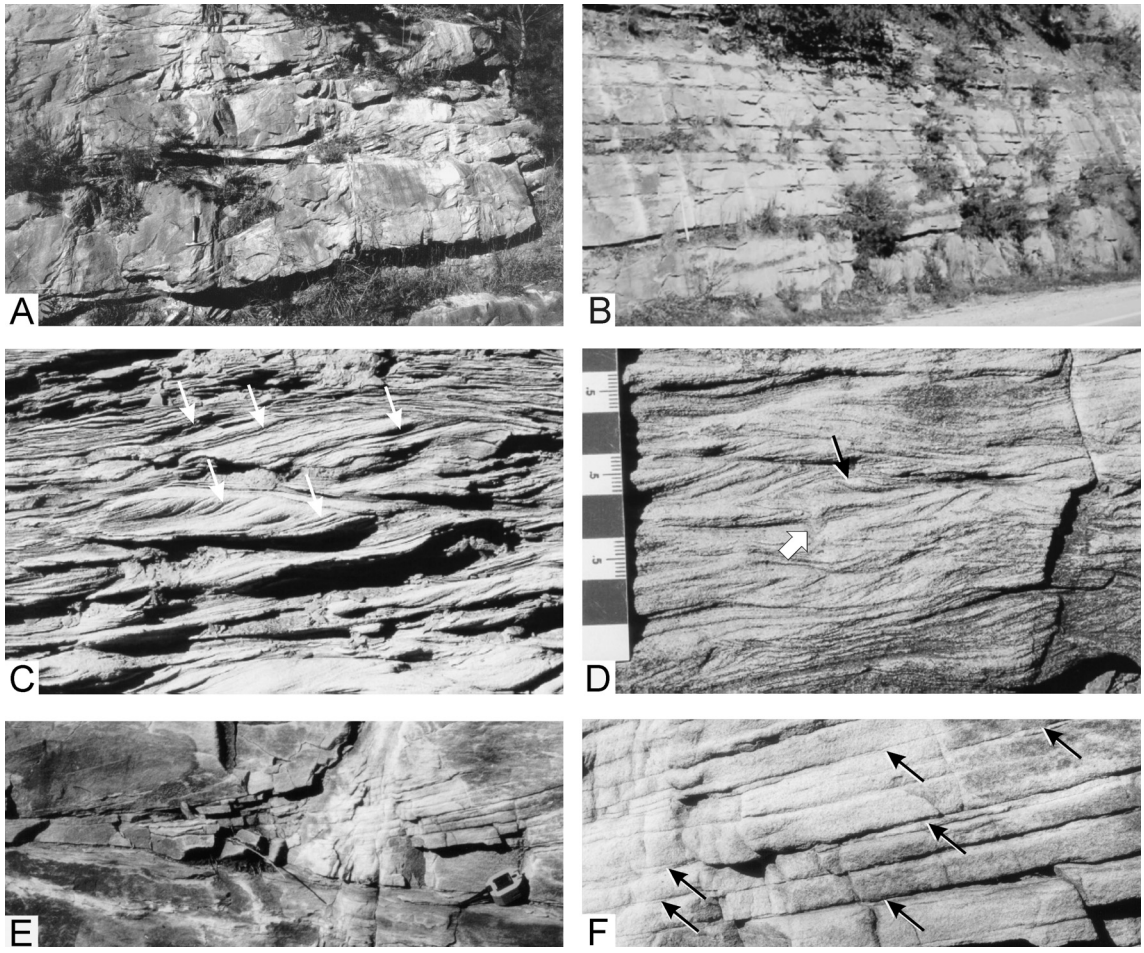

Fig. 7 Pine Creek exposures, location 1. (A) Planar and trough cross-beds in fluvial facies. Hammer scale = 30 cm. (B) Sheet sandstones at the top of the upper main sandstone. Yard stick scale = 0.9 m. (C) Ripples in sheet sandstones showing alternation between thicker, sandier bundles of foresets, and thinner, shalier bundles of foresets (white arrows). Large ripples are *c.* 2 cm thick. (D) Vertical burrow (white arrow). Note possible reversing crest ripple on overlying ripple (black arrow). Scale in millimetres. (E) Isolated down-dip-orientated cross-bed in ripple-bedded sandstone sheets. Brunton scale. (F) Alternating thick–thin laminae foresets (arrows point to thin parts of each pair) in cross-bed. Thick laminae are *c.* 5 to 6 mm thick.

thick–thin laminae alternations in cross-beds diagnostic of estuarine facies. The thick–thin laminations here are not persistent, but may be the first headward indicator of tidal influence in this succession. In the headward part of upper estuarine channels, where fluvial conditions dominate, persistent tidal features would not be expected.

Up-dip-orientated cross-beds in the upper channel are more easily inferred to be tidally influenced. This channel is still dominated by ripple-bedding and cross-bedding so is still inferred to represent an upper estuarine channel, although it contains some features more typical of the inner estuary. The increase in up-dip palaeocurrent indicators (Fig. 6), parallel lamination, shale clasts and cross-beds with tidal indicators suggest a position more seaward than the underlying sheet sandstones. Bundled foresets are common in tidal-estuarine cross-beds (Visser, 1980; Boersma & Terwindt, 1981; Nio & Yang, 1991). Thinner,

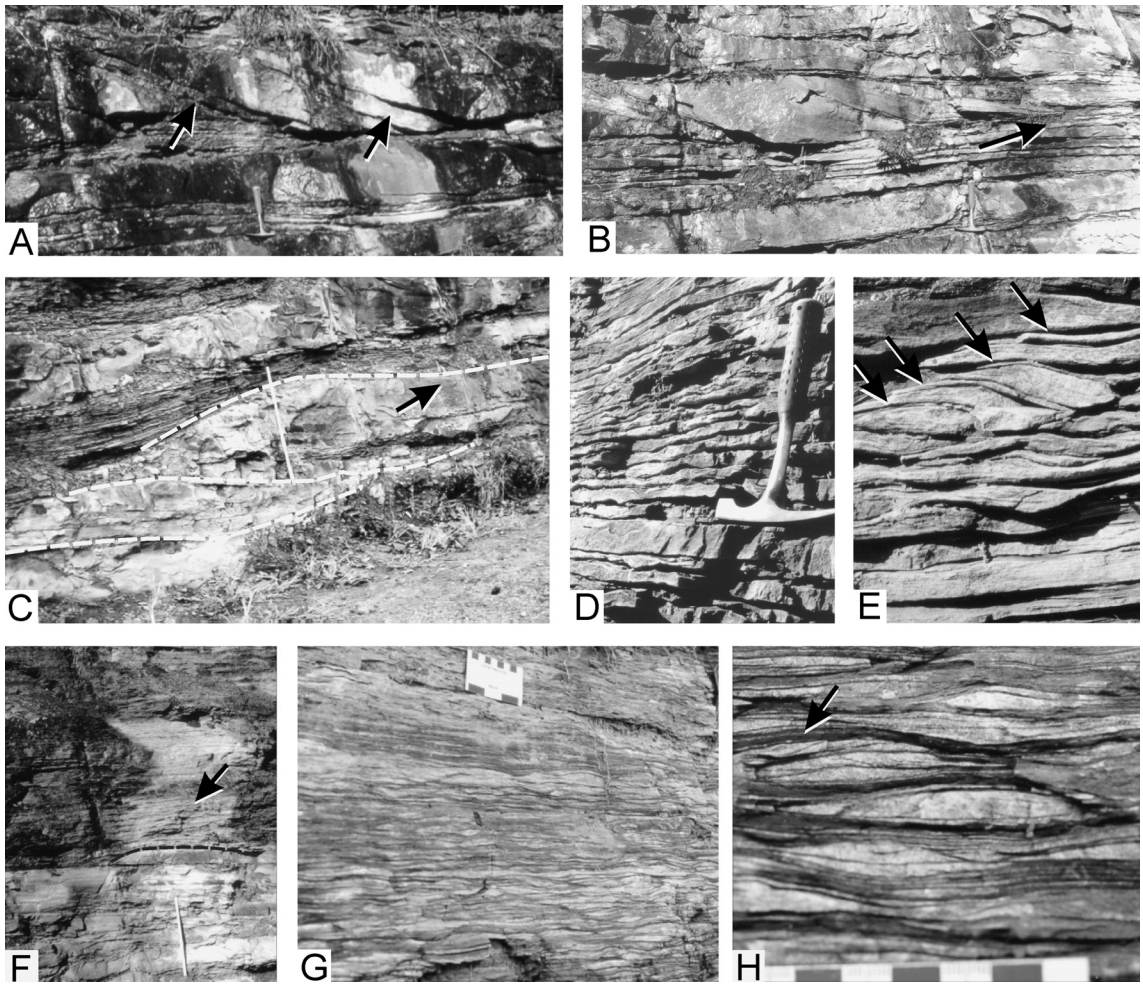

Fig. 8 Pine Creek exposures, location 1. (A) Up-dip-orientated asymmetric cross-bed showing crude bundling of thinner, shale-draped foresets (black arrow) with thicker bundles of sandier foresets. Hammer scale = 30 cm. (B) Small, up-dip-orientated asymmetric cross-bed (black arrow) showing crude thick–thin foreset alternation and soft-sediment deformation (white arrow). (C) Sigmodial foresets (dashed lines) in up-dip-orientated asymmetric cross-bed. Yard stick scale = 0.9 m. (D) Rhythmic lamination in small scour fill. Hammer scale = 30 cm. (E) Climbing ripples in same scour fill showing thick–thin foreset alternations along sigmoidal reactivation surfaces (black arrows). (F) Base of shaly, inclined heterolithic strata (dashed line) with bundles of rhythmically laminated strata (black arrow). Yard stick scale = 0.9 m. (G) Bundles of laminae at the base of the shaly heterolithic strata. Note alternating bundles of sandier, ripple-bedded strata with shalier flaser- to lenticular-bedded sandstone. (H) Detail of bundle showing ripple cross-lamination alternating with millimetre-scale rhythmic lamination. Some ripples have possible reversing crests (black arrow). Scale in centimetres.

shalier foresets bundles represent neap-tide migration, and thicker, sandier foreset bundles represent spring-tide migration. Likewise, rising trough cross-beds with increasing foreset concavity can be formed by rapid upgrowth of troughs during neap periods (Boersma & Terwindt, 1981). Sigmodial bedding (Kreisa & Moiola, 1986; Nio & Yang, 1991) and thick–thin foreset lamination are also common in tidal cross-bedding.

The unusual occurrence of bundled climbing ripples in the small scour at the top of the upper channel suggests periodic alternations in deposi-

tion across a relatively short time span, which also would be more typical of tidal influences than fluvial influences. Climbing ripples with neap–spring–neap changes in shale concentration and climb angle related to neap–spring–neap changes in energy have been documented in the Mont-St-Michel estuary (Lanier & Tessier, 1998). Ebb-dominant climbing ripples are only common in chute channels and point bars of the meandering middle estuary in that analogue. The example herein is ebb-orientated (Fig. 6) and occurs in a small scour at the top of the sandstone, which could represent a chute channel at the inner–middle estuary transition. Overlying shaly heterolithic strata above the sandstone are inclined and represent middle estuarine heterolithic channel facies, although better examples of this facies occur in examples 2 and 3.

Within the middle estuarine heterolithic channel facies, bundles of five to eight laminae couplets overlain or truncated by ripple cross-lamination (as in Fig. 8H) could be interpreted as individual neap–spring deposits. Bundles of neap–spring deposits are arranged in alternating bundles of shalier, rhythmic-lamination-dominated and sandier, ripple cross-bed-dominated bundles, interpreted as annual deposits, similar to bundling described in tidal channels and tidal flats elsewhere (Greb & Archer, 1998; Tessier, 1998).

Example 2

In some transgressive successions, the succession is not as gradational but a middle estuarine heterolithic channel facies can be readily identified and used to re-examine previously interpreted fluvial facies as possibly parts of upper estuarine channels. The top of a Pottsville Sandstone, a Lower Pennsylvanian quartzarenite (Location 2 in Fig. 2), is exposed along US Highway 35 near Jackson, Ohio (Fig. 9A), on the northern margin of the basin (Fig. 2). The fluvial cross-bedded sandstone facies is exposed at outcrops D1 and D2 (Figs 9A & 10). It exhibits mean palaeoflow azimuths to the west, which are typical of fluvial cross-bedded sandstone facies in this area (e.g. Dominic, 1992). Palaeocurrent measurements for vertically adjacent storeys varied by 45° or less, but flow divergence between trough and planar cross-beds within storeys was as much as 90°. A single cross-bed climbs reactivation surfaces in the opposite direction of underlying planar cross-beds (Dominic, 1992; Fig. 11).

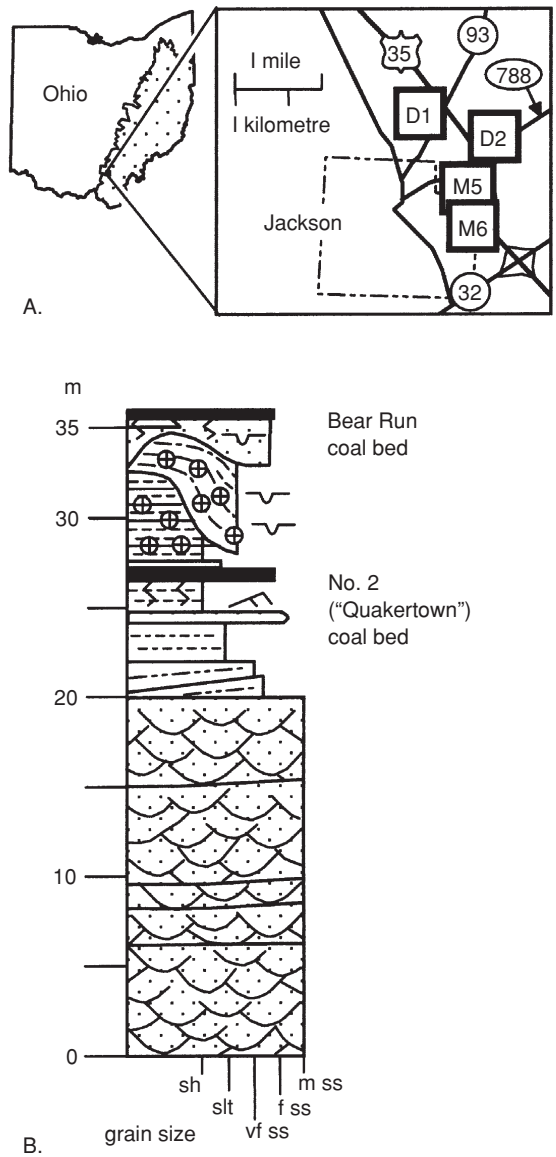

Fig. 9 A lower Pottsville Sandstone at location 2, near Jackson, Ohio. (A) Map of sample outcrops discussed in text (from Martino *et al.*, 1992, Fig. 2.7, p. 13). (B) Stratigraphical section through lower Pottsville strata at exposures shown on the map (after Dominic, 1992, p. 22). See Fig. 6 for legend.

A short distance away, at outcrops M5 and M6 (Fig. 9A), the middle estuarine heterolithic

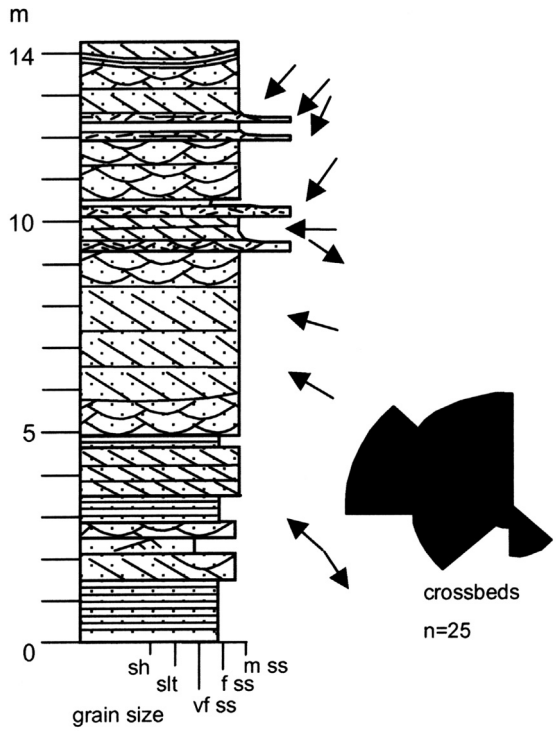

Fig. 10 Measured section of a lower Pottsville Sandstone at location 2, outcrop D2 in Fig. 9A (after Dominic, 1992, p. 16). See Fig. 6 for legend.

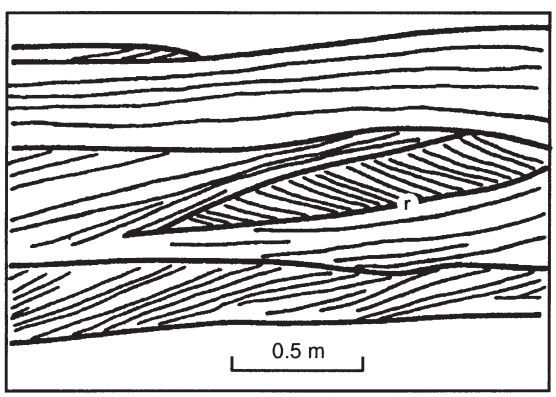

Fig. 11 Line drawing of planar cross-beds with opposing cross-beds above reactivation surface (r), location 2, outcrops D1 and D2 in Fig. 9A (from Dominic, 1992, Fig. 2.7, p. 19).

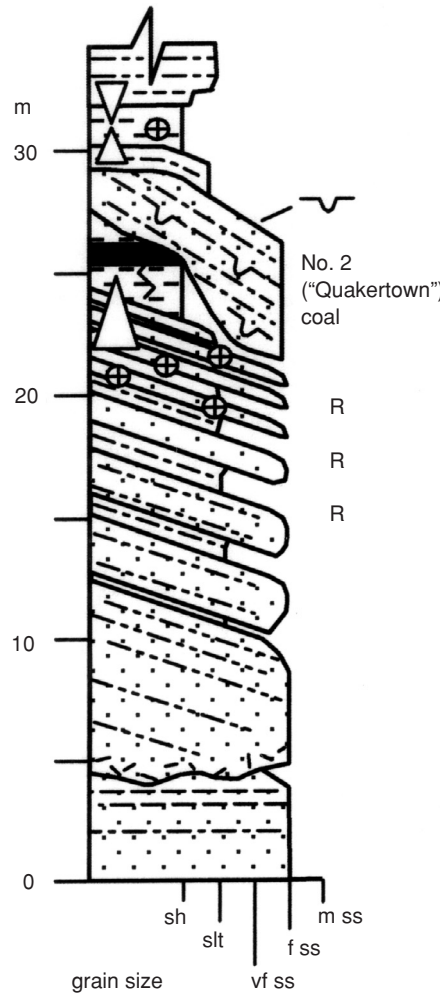

Fig. 12 Measured section of inclined heterolithic strata at location 2, outcrop M6 in Fig. 9A (after Martino *et al.*, 1992). See Fig. 6 for legend.

channel facies occurs between the Quakertown coal and the fluvial cross-bedded sandstone facies. The channel containing the heterolithic facies is 10.5 m deep and contains 7.25-m-thick, large-scale accretion surfaces dipping toward the north-west (320°) at angles of as much as 13° (Figs 12 & 13A). Ripple cross-lamination within inclined sandstone-dominated sets indicates flow toward the south-west (230°). Accretion surfaces are defined by very fine to fine sandstone-dominated intervals, as much as 30 cm thick, which alternate with dark grey, interlaminated very fine sandstone to shale. A crude cyclicity is evident, with sand-dominated intervals alternating with widely spaced shale layers, and more heterolithic intervals alternating with thinner sand layers

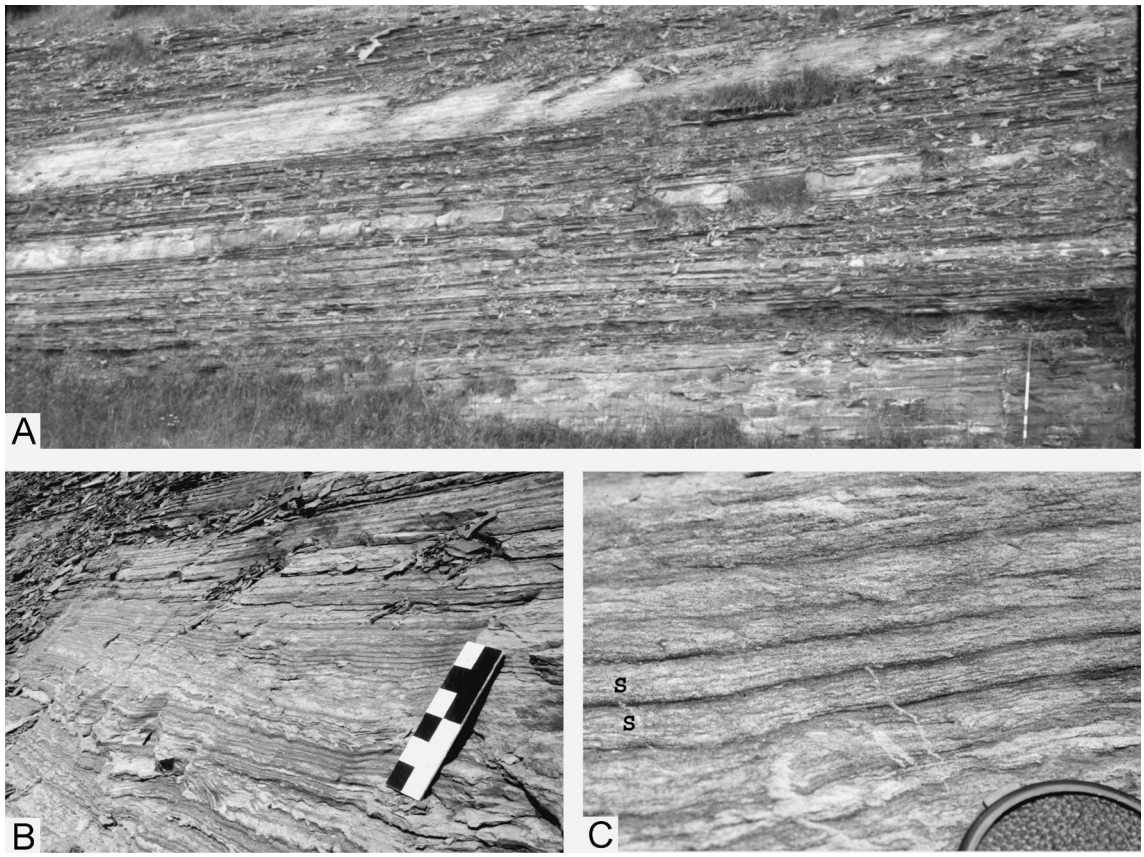

Fig. 13 Inclined heterolithic strata (HIS) at location 2, outcrop M6 in Fig. 9A. (A) Low-angle dipping heterolithic strata. Jacobs staff at lower right 1.5 m. (B) Rhythmic laminations within IHS at M6. Note bundled laminations and alternating larger scale bundles of sandier and shalier strata. Bar scale = 30 cm. (C) Sand–shale couplets in sandstone bedding within IHS. Note sandier parts of bundles (S) are composed of millimetre-scale laminae couplets. These are interpreted as spring tidal deposits that alternate with dark shale intervals formed during neap tides when currents were too weak to entrain sand. Lens cap for scale.

(Fig. 13B). Sand/silt–shale laminae couplets occur in bundles within the heterolithic intervals, which are separated by a persistent shale laminae. Five to eight laminae couplets occur in each laminae bundle (Fig. 13C). Thicker sandstone laminae occur toward the middle of many bundles, although there is local truncation of laminae within bundles. Some of the thicker sandstone units truncate underlying strata. Bioturbation is sparse to abundant, although specific ichnogenera could not be identified. The inclined sandstone–shale interval grades vertically and laterally into thin-bedded dark grey shale and sideritic siltstone. The channel-fill is capped by root-traced seatrock of the overlying Quakertown coal, which is overlain by burrowed sandstones, and shales and limestones with marine invertebrates (Fig. 12).

Interpretation

In this series of outcrops, typical fluvial cross-bedded sandstone facies and middle estuarine heterolithic channel facies can be documented. Both are capped by the peat-mire coal facies and the marine- to brackish-water grey-shale facies, so that a transgressive succession is identified. The middle estuarine heterolithic channel facies here contains decimetre-scale mud couplets. Smith (1988) inferred that there was a transition between fluvial and tidal creek point bars, largely reflected

in increased development and continuity of decimetre-scale sand–mud couplets and frequency in tidal creeks and from micro- to mesotidal conditions within an estuary. The increased frequency of mud drapes in tidally influenced point bars is related to influences of turbidity maximum in the middle estuary (e.g. Thomas *et al.*, 1987). Within the decimetre-scale couplets, shale-draped bundles of laminae on accretion surfaces are similar to non-cyclic rhythmites interpreted as tidal rhythmites in other parts of the basin (Martino & Sanderson, 1993; Martino, 1996; Adkins & Eriksson, 1998; Greb & Archer, 1998). The preservation of five to eight laminae couplets within each bundle does not match a complete tidal cycle duration, but could represent incomplete preservation of neap–spring cycles.

At this particular series of outcrops, the middle estuary heterolithic channel facies succeeds the fluvial cross-bedded sandstone facies, and may actually be laterally equivalent to that facies, because both are overlain by the Quakertown coal. Therefore, in vertical section, a fluvial–estuarine transition can at least be defined across the scour at the base of the middle estuary heterolithic channel facies. A lateral transition may also occur between the two outcrops. Interestingly, the upper part of the fluvial cross-bedded sandstone facies contains a cross-bed orientated opposite to the dominant fluvial mode. Rare opposing palaeocurrent modes can occur in fluvial channels from reversing eddies in unidirectional flow (Allen, 1982a). This example, however, does not appear to occur on the slope or in front of a larger bedform. Additionally, it occurs in an interval of increased overall current variability, relative to more uniform, unimodal palaeoflow indicators lower in the fluvial facies. Hence, it is possible that the feature represents preservation of tidal flood currents in an otherwise fluvial-dominant channel. In that case, the sandstone is part of an upper estuarine sandstone channel facies, and the fluvial–estuarine transition would occur across that facies, at least to the base of the channel scour in which the reversing cross-bed is noted.

Example 3

In some cases, heterolithic strata indicative of tidal flat origins supercede fluvial facies and can be used to interpret the fluvial–estuarine transition. The Kanawha Formation at Low Gap in Boone County, West Virginia (location 3, Fig. 2), is exposed in a series of channels 40 m below the base of the Betsie Shale Member, which defines the Lower–Middle Pennsylvanian boundary (Blake *et al.*, 1994). The lower channel fill is 10 m thick and contains three different types of bedding (Fig. 14). The lower part of the channel fill is typical of the fluvial cross-bedded sandstone facies in the area. It consists of 1.75 m of cross-stratified, medium-grained sandstone, with compound cross-bedding orientated to the south-west. The

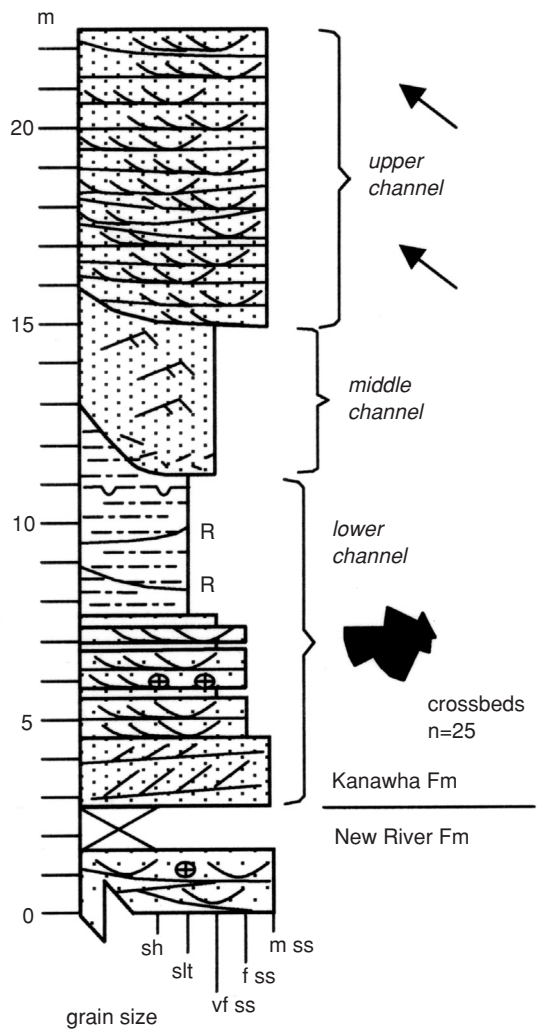

Fig. 14 Measured section through lower Kanawha Formation at location 3, Low Gap, West Virginia. See Fig. 6 for legend.

middle part of the lower channel fill consists of 3.35 m of trough cross-stratified medium- to very-fine sandstone with unimodal palaeocurrents to the south-west similar to the base of the channel fill. Shale-bounded isolated trough sets occur upwards within the middle part of the channel.

The upper part of the lower channel fill consists of 5 m of interlaminated very fine sandstone, siltstone, and shale with well-developed rhythmic lamination (R, Fig. 14). Horizontal, sand-filled burrows, ripple cross-lamination, and broad scour surfaces occur locally. Analysis of sand lamination thickness shows abundant sand/silt–shale laminae couplets arranged in a hierarchy of bundled thickening and thinning cycles (Fig. 15A; Martino, 1996). Minor cycles contain five to eight layers; intermediate-scale cycles contain 15 to 18 layers; Major cycles contain 50 to 62 layers (Fig. 16; modified from Martino, 1996). In addition, thick–thin couplets are also common (Figs 15B & 16). The rhythmically laminated part of the channel fill is interpreted as the estuarine tidal-flat facies.

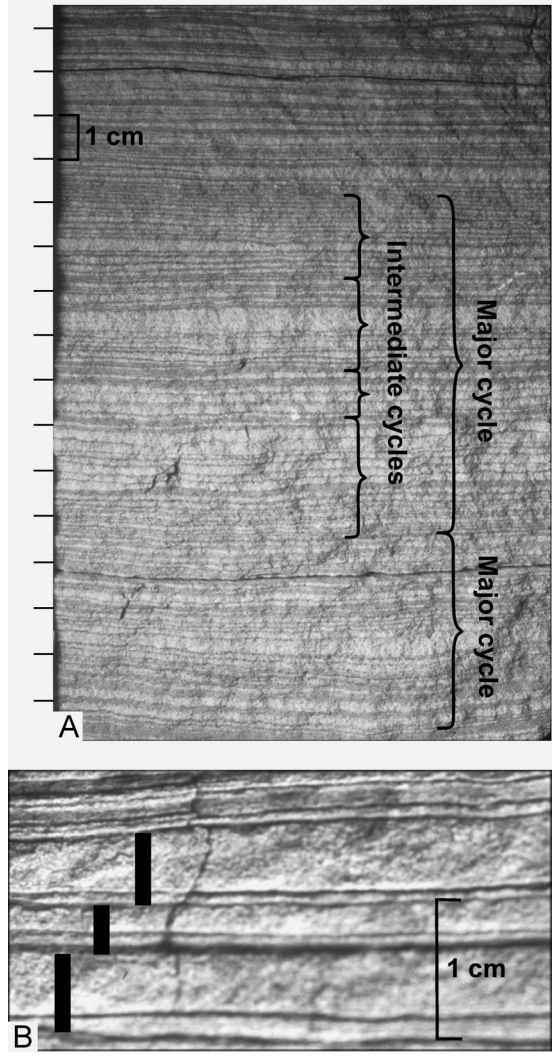

Fig. 15 Rhythmic laminations at Low Gap. (A) Sandstone–shale laminae couplets arranged in rhythmic successions from a block that was collected and analysed; see Fig. 16 for analysis. Vertical thickness = 16 cm. (B) Thick–thin laminae pairs and alternating thick–thin pairs in couplets (black bars).

Interpretation

Rhythmites in intertidal flats are most commonly preserved in the upper middle to inner (upper) estuary (Dalrymple *et al.*, 1991; Tessier, 1993, 1998). Furthermore, thick–thin alternation of laminae couplets would be uncommon in purely fluvial environments, and suggest diurnal inequality in tides formed in semi-diurnal settings (e.g. Kvale *et al.*, 1989). The rhythmites at Low Gap preserve distinct orders or scales of cyclicity but do not preserve complete tidal sedimentation records. In an intertidal setting, tidal currents may be too weak to entrain sediments during the neap portion of spring–neap cycles (Fig. 17), in the higher portions of intertidal flats, or headward within the fluvial–estuarine transition, resulting in increasingly non-cyclic (incomplete) rhythmites (Dalrymple *et al.*, 1991, 1992; Tessier, 1993, 1998; Archer, 1998). The partial preservation or amalgamation of tidal laminae in rhythmites at Low Gap could have resulted in spring–neap cycles that contain only five to eight mud-draped sand layers (minor cycles in Fig. 16). Likewise, lunar perigee–apogee cycles might contain less than 30 layers owing to the weakness of neap tides at this position in the palaeo-estuary and be represented by the intermediate cycles of 15 to 18 layers (Figs 15A & 16). The major cycles of 50 to 62 layers (Figs 15A & 16) may represent attenuated 6-month (solstice–equinox) cycles. Seasonal tidal cycles with maximum tidal ranges occurring during June and December have been interpreted for

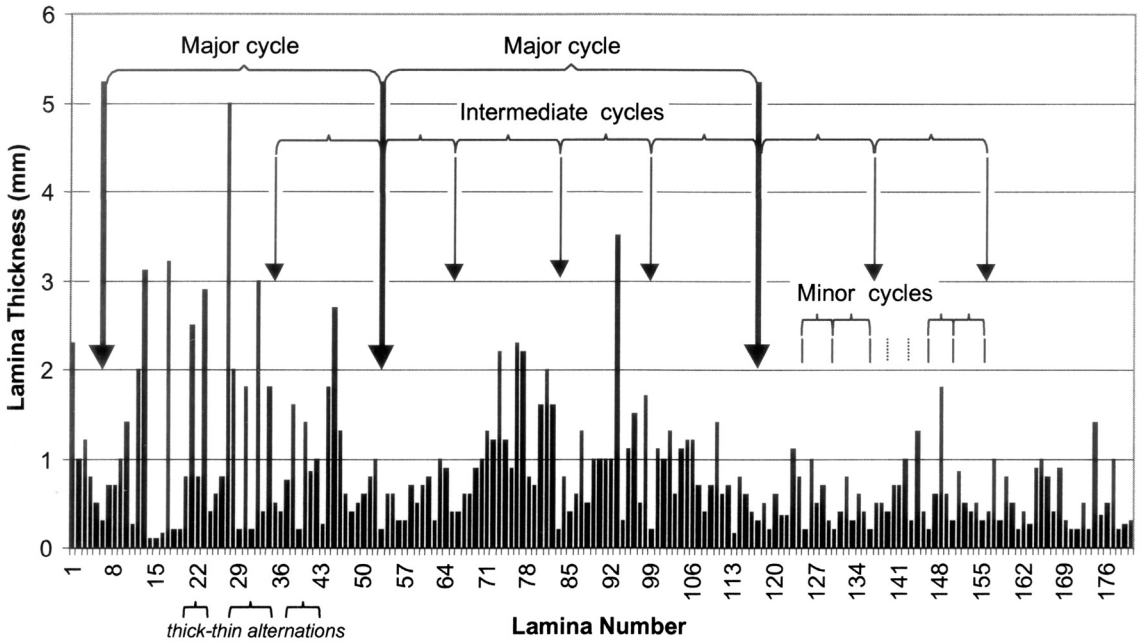

Fig. 16 Plot of laminae thickness measured from collected block, a portion of which is illustrated in Fig. 15A. Coarse layers (very fine sandstone, siltstone) in each couplet are plotted.

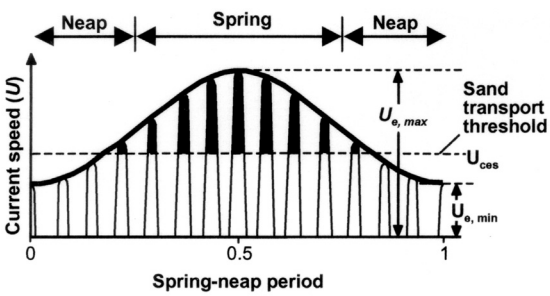

Fig. 17 Plot of current speed variation during a hypothetical spring–neap cycle. Sandstone laminae may only be preserved above the sand transport threshold (after Allen, 1982b).

other Carboniferous rhythmites (Kvale *et al.*, 1989; Archer, 1991; Martino & Sanderson, 1993).

The tidal rhythmites occur above a channel with otherwise fluviatile characteristics. The lower part of the channel fill contains cross-bedding orientated in the inferred direction of fluvial flow. Hence, the fluvial–estuarine transition occurs at least across the scour in the upper sandstone overlain by the estuarine tidal-flat facies. The trough cross-beds within the middle interval of the channel fill are draped by continuous shale laminae. These could represent seasonal fluctuation in river discharge within the fluvial facies, but in a transgressive succession they could indicate rapid sedimentation from suspension caused by the interaction of flood tidal currents in the fluvial–estuarine transition. If so, the fluvial–estuarine transition would occur across this part of the channel fill, rather than at the scour at the top of the trough cross-bedded interval. There is no supporting evidence within the trough cross-bedded interval to substantiate tidal influences, but such evidence might not be preserved in the most headward part of an upper estuarine channel.

Example 4

In some cases, diagnostic tidal features are not found, but possible tidal features are noted within transgressive successions at stratigraphical horizons where estuarine facies are noted elsewhere. Greb & Chesnut (1996) identified a transgressive succession at the top of the Rockcastle Sandstone on the western margin of the basin. Bioturbated IHS (herein referred to as the middle estuarine

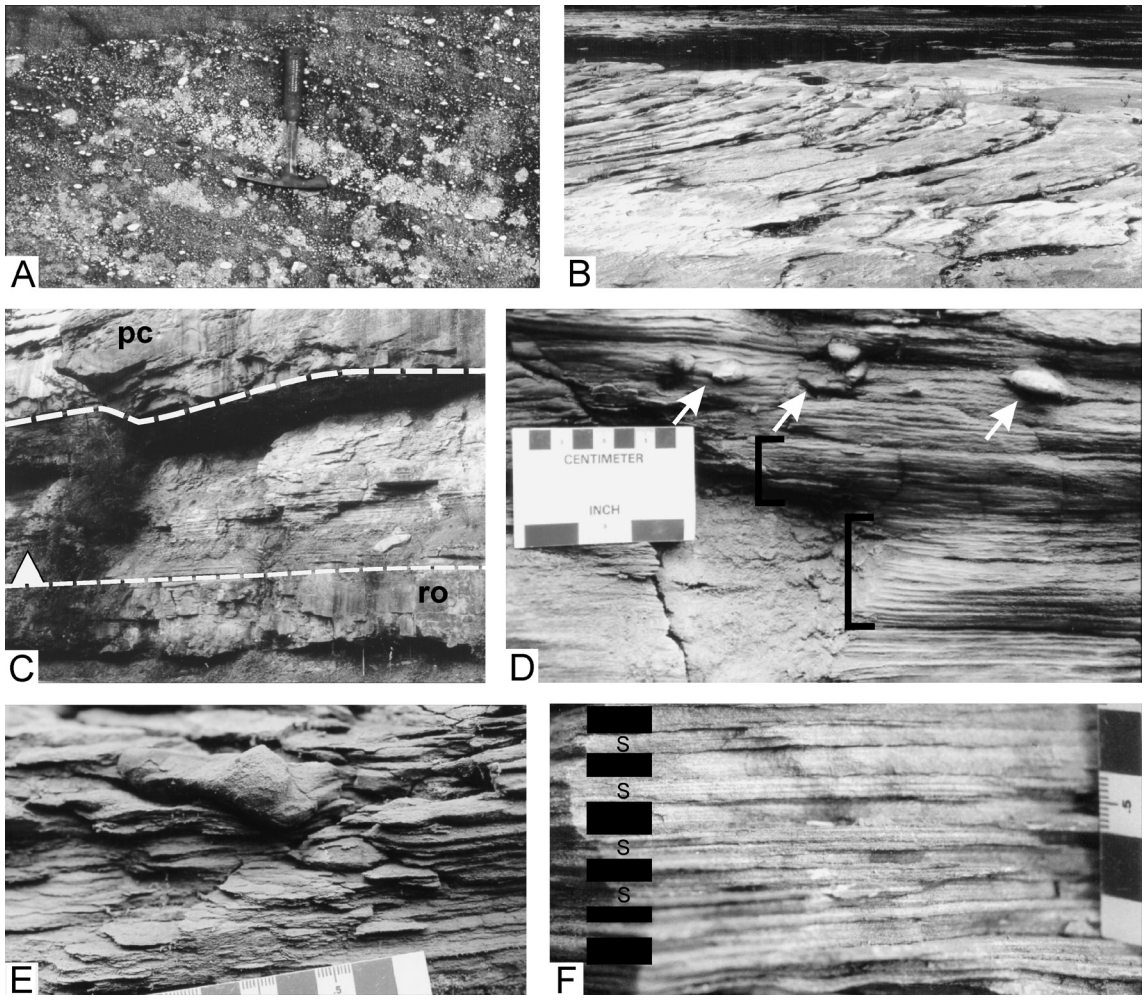

Fig. 18 Rockcastle Sandstone at Cumberland Falls, location 4. (A) Conglomeratic cross-bed facies typical of fluvial facies in the Rockcastle Sandstone. Hammer scale = 30 cm. (B) Bedding-plane exposures of south-easterly orientated, broadly arcuate, planar-tabular cross-beds in fluvial facies. (C) Channel exposed at the top of the Rockcastle Sandstone (ro) north of the entrance to Cumberland Falls State Park: bf, Barren Fork coal; pc, Pine Creek Sandstone. (D) Siderite nodules (white arrows) in non-cyclic rhythmites. Rhythmites consist of bundled laminae couplets (brackets). (E) Siderite nodule with *Rosselia* shape. Scale in centimetres. (F) Detail of sandy part of rhythmite bundle showing alternating groups (s) of sandier couplets and shalier (black boxes) couplets. Scale in centimetres.

heterolithic channel facies) was noted in a transgressive succession. Examination of other localities within 15 km of the IHS locality shows that there is considerable variability in facies beneath the coal and grey shale. An example is described here (location 4, Fig. 2).

The upper 3 m of the Rockcastle Sandstone is dominated by the fluvial cross-bedded sandstone facies at most outcrops (Fig. 18A & B; Wizevitch, 1991; Greb & Chesnut, 1996). Locally, however, a fining-upward, heterolithic interval occurs between the fluvial facies and Barren Fork coal bed (Figs 18C & 19A & B). Also, at many localities, the marine- to brackish-water grey shale facies that overlies the coal in other areas (Greb & Chesnut, 1996) is replaced by scour-based fluvial cross-bedded sandstone facies of the overlying sandstone (Figs 18C & 19A & B). The heterolithic

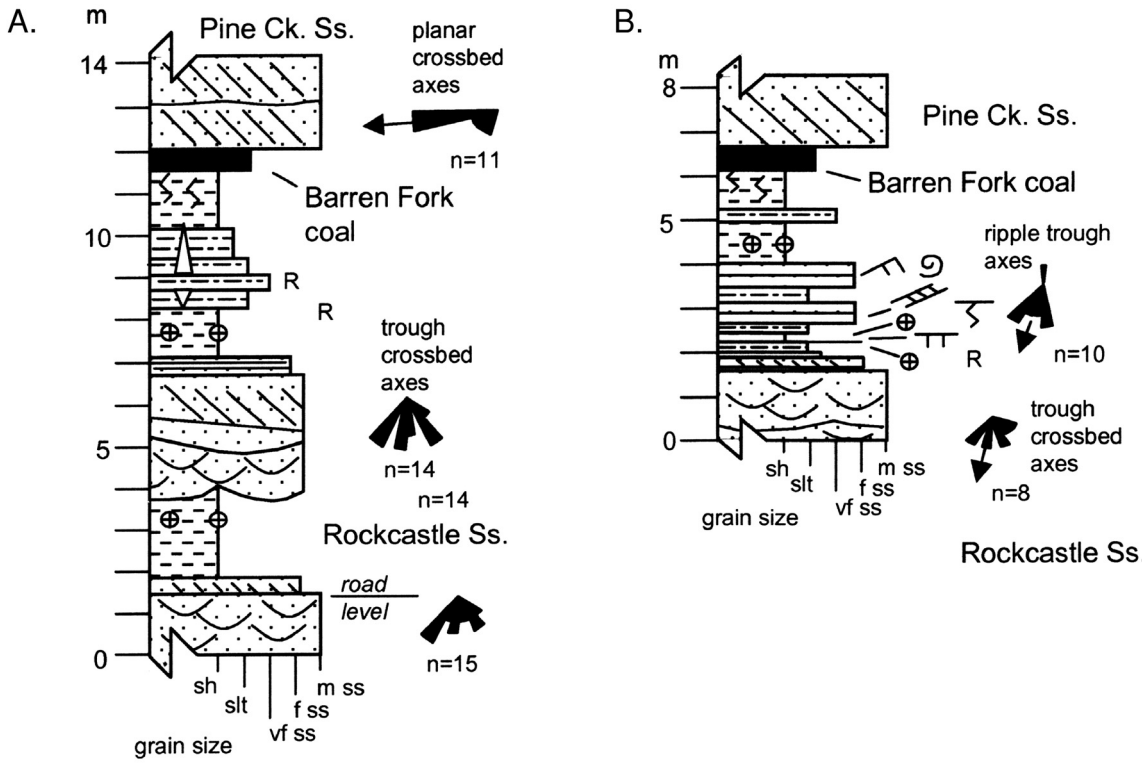

Fig. 19 Vertical sections from the upper Rockcastle Sandstone, location 4. (A) North side of Cumberland River near the entrance to Cumberland Falls car park at Cumberland Falls State Park. (C) South side of the Cumberland River near the entrance to Cumberland Falls car park at Cumberland Falls State Park. See Fig. 6 for legend.

strata at the top of the Rockcastle Sandstone exhibit ripple-bedded sandstones with climbing ripples, crude rhythmic laminations and contorted bedding. Large *Stigmaria* roots penetrate the sandstone and heterolithic layers (Fig. 19A & B). Within the heterolithic layers, siderite nodules occur along distinct bedding planes. In one layer, sideritized vertical, cylindrical structures occur above an exposed *Stigmaria* root. Some nodules are tubular and parallel to bedding, such that they are similar to *Planolites*. Some are more conical or irregular in shape (Fig. 18D), with one having the shape of *Rosselia* (Fig. 18E). Rhythmic laminations are ordered in bundles of laminae 3 to 4 cm thick, consisting of a sandier and a shalier half (brackets in Fig. 18D). These bundles are composed of smaller scale bundles, each 0.2 to 0.7 cm thick (Fig. 18F). Sandy bundles (s in Fig. 18F) alternate with shalier bundles (black boxes in Fig. 18F). The small-scale sandy bundles consist of three to five laminae couplets, each couplet consisting of a very fine sandstone or siltstone and shale laminae. Some bundles show vertical thickening and thinning of laminae couplets, although discontinuities are common.

Interpretation

Local, rooted, fining upward, heterolithic channel fills could be interpreted as floodplain deposits of the typical fluvial facies reported for Lower Pennsylvanian sandstones. In fluvial floodplain deposits, however, fining upward laminations from suspension deposition or coarsening upward laminations from bedform progradation might be more typical than the sand–shale laminae couplets arranged in bundles preserved here. As a transgressive succession was interpreted at this stratigraphical horizon nearby (Greb & Chesnut, 1996), the non-cyclic rhythmites in this example

are interpreted as incomplete records of tidal sedimentation, similar to the non-cyclic rhythmites at the other locations previously discussed. If the laminae couplets represent daily deposition, then the alternating sandy and shaly small-scale bundles might represent incomplete neap–spring cycles. The occurrence of five to six neap–spring cycles in the sandy part of each larger scale bundle represents a higher energy five- to six-month seasonal influence. Larger scale bundles with sandy and shaly halves could represent annual deposits consisting of a high- and low-energy seasonal deposit, similar to annual rhythmites described from Carboniferous tidal channels at other locations (Greb & Archer, 1998). The non-cyclic rhythmites are crudely ordered, indicating that the fining upward heterolithic interval should be placed within the estuarine tidal-flat facies. Rooting structures in these rhythmites may indicate a relatively high position on tidal flats, which would preclude preservation of the high tidal signature, and lead to incomplete tidal records of tidal deposition (Dalrymple *et al.*, 1991; Tessier, 1993, 1998).

Where climbing ripples have been found in tidal deposits, they are restricted to the fluvial–estuarine transition (Lanier & Tessier, 1998). Interestingly, in the Mont-St-Michel estuary, flood-dominant climbing ripples (as occur here) are restricted to areas along the margins of inner straight channels of the fluvial–estuarine transition, where they are also associated with rhythmites, soft-sediment deformation and rooting (Lanier & Tessier, 1998). Rooting is common in fluvial–estuarine transition zones because of the common development of marshes in the inner estuary (Lanier & Tessier, 1998; Tessier, 1998). Most of the siderite nodules at this location are rootlets, but some could represent invertebrate bioturbation. Sideritization would have destroyed the internal structure of any structures that might have been burrows. At least one of the siderite nodules is similar in shape to *Rosselia*, which is common in Carboniferous marine and tidal facies in the basin (Martino, 1989, 1992, 1994; Greb & Chesnut, 1992, 1994). In this case, the nodule may represent a rootlet, but the point of similarity is brought up because differentiating infaunal invertebrate burrows from nodular root traces can be difficult in the fluvial–estuarine transition.

DISCUSSION

Controls on fluvial–estuarine transitions

There are numerous controls on the development of palaeofluvial–estuarine transitions in tide-dominated estuaries. As estuaries are coastal features, the relative position of the palaeocoastline and orientation of valleys and lowlands along that coast are critical to the development of estuaries. If river valleys are transgressed, the overall shape and slope of the valleys will influence the development of tidal prisms and resultant tidal sedimentation. Estuaries become longer as coastal gradient decreases and/or tidal range increases (Dalrymple *et al.*, 1992). Upstream changes in valley shape could cause changes in the character of tidal sedimentation as the seas transgressed across those changes. Likewise, the amount of sediment flux, seasonality and other variables controlling fluvial discharge will influence whether a delta or estuary forms, and in tide-dominated estuaries, the preservation potential of tidal features in the most landward areas of tidal influences (Dalrymple *et al.*, 1992; Wells, 1995). Moreover, the position and extent of fluvial–estuarine transitions in the rock record will be dependent on the rate and amount of relative sea-level rise, which will be a function of eustasy, sediment flux and tectonic accommodation.

During transgression and into highstand, the overall size of drainage basins decreases so that fluvial input into up-dip-advancing estuaries is decreased if climate is unchanged. Overall river grade also decreases. As base level rises, increasing numbers of tributaries can be converted to estuaries, because of the upstream-branching in pre-existing fluvial systems (Archer & Greb, 1995). The increase in the number of transgressed tributaries causes an increase in the number of locales in which there can be fluvial–estuarine transitions and the number of sediment catchments in which fluvial–estuarine transitions can be preserved. The preservation potential of tidal-estuarine facies in these catchments will be dependent on the geometry of the valley as it is drowned, and the amount of fluvial input during transgression. The relative amount of fluvial and tidal energy will determine the length of the fluvial- and tidal-dominant parts of the estuary (Dalrymple *et al.*,

1992), as well as the nature of the fluvial–estuarine transition. Some tributaries (secondary and tertiary drainages) will have little fluvial input so that tidal influences might leave a greater imprint on the sedimentology of the preserved fill, as long as the tributaries are large enough or shaped in such a way that they can generate a tidal prism. First-order drainages with active discharge will be more likely to be fluvially dominated, even if influenced by flood-tidal currents. Studies of modern estuaries with fluvial inflow have shown that during floods:

1 more sediment may be supplied to the estuary in a matter of days than is supplied by months or years of average sedimentation;

2 there is lateral translation of salt/fresh-water interface;

3 there is translation of zones of suspended sediment accumulation (Nichols and Briggs, 1985).

All of these factors could lead to the net dominance of fluvial sedimentation at the expense of tidal sedimentation across the fluvial–estuarine transition, even if fluvial dominance was temporally subordinate. Additionally, as individual estuaries translate landward, the upper portion of the transgressive succession can be removed by tidal channel erosion (Davis & Clifton, 1987; Dalrymple *et al.*, 1992), further complicating identification of transgressive successions and fluvial–estuarine transition facies. Considering all of these factors, variability in fluvial–estuarine transitions might be expected both spatially and temporally if transgressions flooded broad alluvial braidplains, such as those envisioned for the Early Pennsylvanian of the Central Appalachian Basin.

Early Pennsylvanian estuaries

During the Early Pennsylvanian a series of longitudinal braidplains was developed in the Central Appalachian Basin (Chesnut, 1992, 1994; Archer & Greb, 1995; Greb & Chesnut, 1996). Each of the braidplains was succeeded by marine- to brackish-water grey-shale facies, indicating that the braidplains were transgressed. The development of estuarine facies between the typical fluvial facies of the braidplains and the overlying transgressive shale indicates local estuarine development during transgression (Greb & Chesnut, 1996). Some inferences can be made about the development of Early Pennsylvanian estuaries, based upon comparison to Pleistocene analogues. During Pleistocene trangsressions, the slope of the flooded coastline dramatically affected the development of estuaries. In general, estuaries developed within valleys where there was an increase in the slope of the transgressed surface, whereas bays and lagoons developed on flat, or low, continuous slopes (Emery, 1967).

Most vertical sections through the Lower Pennsylvanian quartzarenite belts of the Central Appalachian Basin show a succession from the fluvial cross-bedded sandstone facies to peat-mire coal facies to marine- to brackish-water grey-shale facies, suggesting that broad floodplains and abandoned braidplains within the quartzarenite belts were paludified and capped by peats (especially in topographic depressions and coastal lowlands), which accumulated as water tables rose. Peats were transgressed across broad areas where underlain by floodplains and abandoned parts of the braidplain. Facies containing tidal sedimentary structures occur only locally, mostly in channel facies. This indicates that Early Pennsylvanian transgressions initially followed the paths of abandoned or active channels within valleys. Fluvial channels within valleys were locally converted to upper estuarine channels.

Upper estuarine channels, by definition, are fluvially dominant with some tidal influences (Allen, 1991; Dalrymple *et al.*, 1992). The upper estuarine sandstone channel facies defined herein is difficult to interpret at any outcrop if it contains only a few, possible, tidal indicators. In cases of more limited outcrop, isolated evidence for opposing currents might only represent reversing eddies in a fluvial environment. Where a transgressive succession can be defined at a specific stratigraphical interval and there are at least middle estuarine heterolithic channel facies or estuarine tidal-flat facies at the same horizon, then interpreting fluvially dominant channels with subtle tidal indicators as upper estuarine sandstone channel facies is tenable.

Recognition of tidal-estuarine indicators

In analyses of Cretaceous fluvial–estuarine transitions, Shanley *et al.* (1992) noted that the preponderance of evidence was sometimes needed

to infer tidal influences in the fluvial–estuarine transition; such is the case in the Pennsylvanian examples studied herein. In the examples studied, tidal indicators were sometimes subtle, and often individually inconclusive. The greatest diversity and abundance of possible tidal structures not surprisingly occurred in the upper part of the preserved succession, either in middle estuarine heterolithic channel facies or estuarine tidal-flat facies. With decreasing frequency and variety of possible tidal features down-section, the possible lower (headward) limit of the fluvial–estuarine transition within the underlying fluvial deposit is difficult to delineate precisely.

In a complete transgressive succession, a vertical section through a palaeoestuary might preserve upper, middle and lower estuary facies. In the headward regions of estuaries, however, the succession might not contain lower estuarine facies. Dalrymple *et al.* (1992) illustrated hypothetical vertical sections through estuarine fills produced by transgression followed by progradation. In the headward examples of these fills, the maximum flooding surface occurred within the estuary. A vertical section through a similar succession in which the middle estuarine heterolithic channel facies was the point of maximum transgression is shown in Fig. 20A. This is similar to the succession noted at location 1 (Fig. 2), where the Dave Branch Shale may contain (in part) the middle estuary heterolithic channel facies (Fig. 6). In most cases, however, the marine- to brackish-water grey shale facies caps estuarine facies and a wide variety of vertical successions is noted.

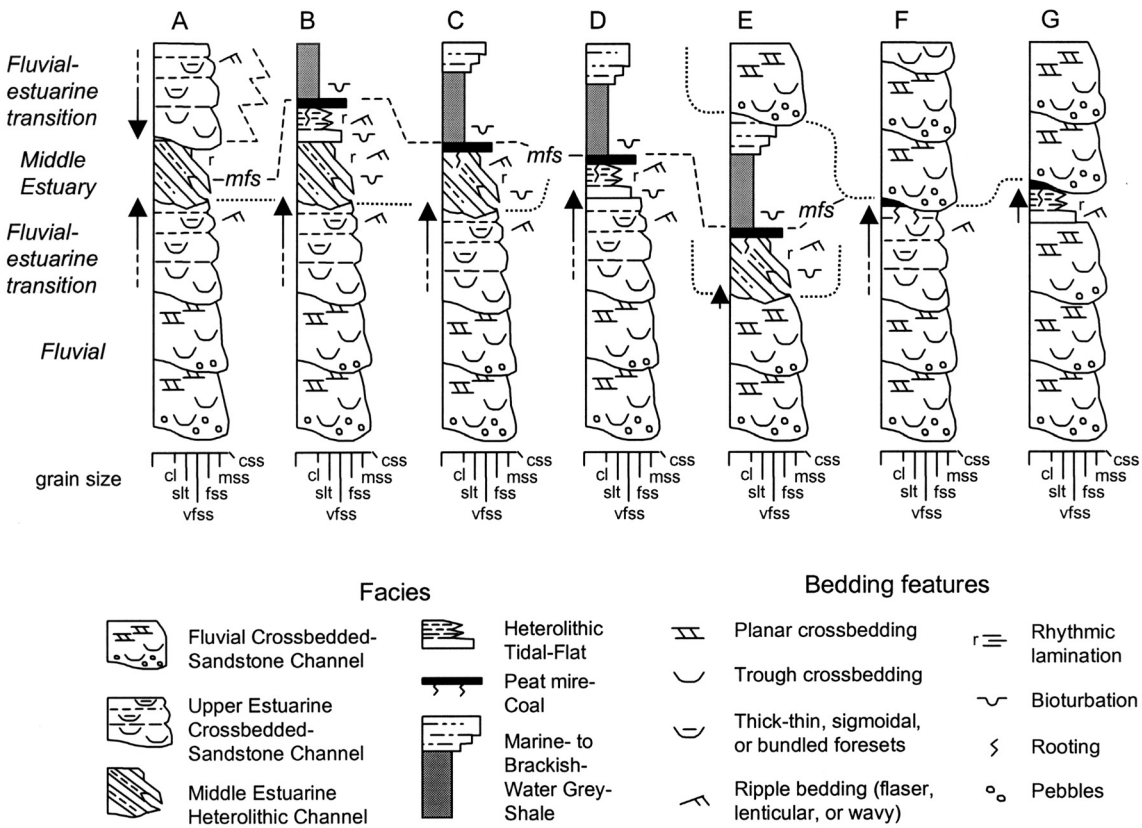

Fig. 20 Schematic vertical sections (A–G) showing variability in transgressive successions in fluvial-dominated strata. The fluvial–estuarine transition is easier to infer when middle estuarine heterolithic channel facies can be defined in a transgressive succession (A, B, C & E). Where that facies is missing (D, F & G) or the overlying marine facies are missing (F & G), the fluvial–estuarine transition is more difficult to interpret.

If each of the middle and upper estuarine facies described among the various outcrops studied (Table 1) were stacked in vertical succession, they might appear in the succession illustrated in Fig. 20B. In some cases, middle estuarine heterolithic channel facies overlie upper estuarine channel facies (Fig. 20C), as in Example 2, or estuarine tidal-flat facies overlie upper estuarine channel facies (Fig. 20D), as in Example 3. In some cases, the middle estuarine heterolithic channel facies might directly overlie fluvial facies without an intervening upper estuarine channel transition (Fig. 20E). This is what could be interpreted for Example 2 if the up-dip cross-bed in underlying 'fluvial' facies was not noted, or not considered a tidal indicator. Where middle estuarine heterolithic channel facies or estuarine tidal-flat facies can be identified (Fig. 20B, C & E), detailed examination of underlying strata can be undertaken to determine if there is any evidence for tidal sedimentation that could lead to reinterpretation of fluvial cross-bedded sandstone facies as upper estuarine cross-bedded sandstone channel facies. It is more complicated to interpret estuarine facies where the transgressive shale or middle estuarine facies is missing (Fig. 20F & G). If marine facies are truncated by overlying sequences, then the transgressive succession may not be apparent, and subtle tidal indicators of the upper estuarine cross-bedded sandstone channel facies might be overlooked.

Bioturbation, one of the most commonly used indicators of salinity change, is not pervasive in the fluvial–estuarine transition of modern environments (Howard & Frey, 1973, 1975; Allen, 1991; Dalrymple *et al.*, 1991, 1992; Tessier, 1993, 1998), and is uncommon in the upper estuarine facies interpreted herein. Unless the study area is at a location where the transgression continued past the fluvial–estuarine transition into the middle or lower estuary, bioturbation may not be common. Additionally, in quartzarenites, there may not be sufficient variations in grain colour or texture to distinguish burrows in outcrop. Weathering accentuated the burrow shown in Fig. 7D. If bioturbation did occur lower in the section, where interlaminated shales are less common, it might not be noticed. Another complication to recognition of invertebrate bioturbation in the fluvial–estuarine transition is the common occurrence of rooting, as occurs in Example 4. Many Lower Pennsylvanian fluvial facies are capped by coal beds such that paludification and rooting of immediately underlying facies might destroy important tidal evidence. Root disturbance of tidal flats in the fluvial–estuarine transition is common in modern estuaries (Dalrymple *et al.*, 1992; Tessier, 1993, 1998). Geochemical indicators of salinity changes, such as carbon isotope ratios in siderite (Weber *et al.*, 1979), or pyritization and carbon/sulphur ratios of shales (Berner & Raiswell, 1984), may be needed to help support a fluvial–estuarine transition where bioturbation is lacking or inconclusive.

Bipolar cross-bedding is probably the most commonly cited criterion used to interpret tidal influences in fluvial–estuarine transitions (e.g. Thomas *et al.*, 1987; Rahmani, 1988). Features formed by opposing tidal modes in sediments within the fluvial–estuarine transition may be very subtle. Rather than truly bipolar palaeocurrent distributions, preserved flow indicators of fluvial–estuarine transitions are generally unimodal with only a few opposing orientations. It is difficult to conclusively interpret these as tidally modulated without supporting tidal evidence; however, their limited occurrence in a transgressive succession might indicate the most headward limits of tidal influences in the palaeoestuary.

Several studies of Cretaceous fluvial–estuarine transitions have indicated that sigmoidal reactivation surfaces within cross-strata where they bound thickening and thinning foreset bundles are a tidal indicator (Kreisa & Moiola, 1986; Shanley *et al.*, 1992). In several of the Pennsylvanian outcrops studied here, sigmoidal bedding and asymmetrically-filled cross-beds with rising troughs were noted stratigraphically below the lowermost occurrence of bidirectional dips or bioturbation. Bundled cross-beds in these facies, however, were uncommon. If these features record tidal influences, then the fluvial cross-bedded sandstone facies at the top of quartzarenite belts in some areas may actually belong to the upper estuarine cross-bedded sandstone channel facies.

Another type of feature reported for other fluvial–estuarine transitions is so-called tidal lamination, which can include bundled laminae

bounded by shale laminae, thick–thin laminae alternations and rhythmic lamination (e.g. Rahmani, 1988). Part of the evidence for showing cyclicity in tidal rhythmites is counting continuous repetitive thickening and thinning laminae couplets related to known tidal cyclicities as was done for the rhythmites at location 3 (Fig. 16). The inner or upper estuary is a zone in which tidal amplification can be favourable for the deposition of cyclic rhythmites on intertidal flats. Yet even in these areas, the preservation potential of the rhythmites is dependent upon local accommodation, sediment flux, tidal asymmetry and position of deposition relative to the level of the highest tides. In general, cyclic rhythmites will become increasingly non-cyclic (incomplete) higher on tidal flats or channel margins where weaker tides do not reach (Dalrymple *et al.*, 1991; Archer, 1998; Greb & Archer, 1998; Tessier, 1998). Most of the rhythmites noted in the Lower Pennsylvanian transgressive successions were non-cyclic. By themselves, non-cyclic rhythmites are not diagnostic tidal indicators (Greb & Archer, 1995). Likewise, thick–thin alternations in foreset laminae were noted in many of the examples studied, but none were persistent throughout several trough cross-beds, such that they also are not diagnostic tidal indicators.

Subtle, relatively isolated tidal features may be the only indication of tidal influences in many fluvial–estuarine transitions. If marine flooding surfaces had not been defined previously for each of the study intervals, subtle tidal features might not have been looked for in underlying strata and facies would have been interpreted as wholly fluvial. This is particularly true in areas where incision of the following fluvial sandstone truncates the marine- to brackish-water grey-shale facies or peat-mire coal facies, as happens at location 4. In areas of low accommodation, or in the up-dip parts of basins where fluvial sequences dominate, subtle tidal features in upper estuarine facies may be the only record of transgressions and down-dip marine flooding surfaces for sequence-stratigraphy implications. Recognition of these transitions can enable more accurate delineation of lowstand and transgressive systems tracts within fluvial-dominated strata, and will improve the ability to anticipate vertical and lateral changes in fluvial style, and channel-fill geometry, lithology and continuity. Recognition of fluvial–estuarine transitions may also help in interpreting or predicting lateral and vertical changes in porosity and permeability owing to lithological changes from fluvial to estuarine facies in hydrocarbon reservoirs.

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