ESTUARINE AND INCISED-VALLEY FACIES MODELS

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Abstract: Modern estuaries and incised valleys are important depositional settings that have widespread significance for human land use. The deposits of these environments are economically important for hydrocarbon exploration and production. Estuaries and incised valleys are a complex and possibly unique environmental grouping, inasmuch as they represent creation of depositional space by one process (mainly fluvial erosion) and fill of that space by a range of other processes (fluvial, estuarine, and marine deposition).

Early investigations of valleys began slowly in Greek and Roman times, but increased in the nineteenth century, when they were used to develop ideas on the age of the earth in uniformitarian debates. Gradual progress was made throughout the nineteenth and twentieth centuries with the introduction of ideas on river grade, fluvial equilibrium profiles, and base level, followed by the development of fluvial facies models in the 1960s. Studies on estuaries began in earnest much later than those on valleys, and major advances were not made until the mid-twentieth century, with development of the first comprehensive facies model in the 1990s.

Research on estuaries and incised valleys was energized in the 1980s by the concept of sequence stratigraphy, and work in the field has mushroomed since then. Indeed, the currently used facies models for estuaries and incised valleys were among the first to explicitly take into account the external control on the creation of accommodation and to be presented in a sequence-stratigraphic framework. In line with other sedimentary environments, the facies models for estuary and incised-valley environments have also proliferated, leading to the need for fundamental advances in how facies models are conceived.

Estuaries, as defined geologically here, are transgressive in nature. They receive sediment from both fluvial and marine sources, commonly occupy the seaward portion of a drowned valley, contain facies influenced by tide, wave, and fluvial processes, and are considered to extend from the landward limit of tidal facies at their heads to the seaward limit of coastal facies at their mouths. Estuaries can be divided, on the basis of the relative power of wave and tidal processes, into two main types, *wave-dominated estuaries* and *tide-dominated estuaries*. Estuarine facies models exhibit generally retrogradational stacking of facies and a tripartite *zonation* reflecting the interaction of marine and fluvial processes. All estuaries and incised valleys have a fluvial input by definition, but estuarine facies models reflect the balance between wave and tidal processes.

Valleys form because the transport capacity of a river exceeds its sediment supply. An incised-valley system is defined as a fluvially eroded, elongate topographic low that is characteristically larger than a single channel, and is marked by an abrupt seaward shift of depositional facies across a regionally mappable sequence boundary at its base. The fill typically begins to accumulate during the next base-level rise, and it may contain deposits of the following highstand and subsequent sea-level cycles if the accommodation is not filled during the first sea-level cycle. Incised valleys may be formed by either a piedmont or a coastal-plain river and can exhibit a simple or compound fill. The erosion that creates many incised valleys is thought to be linked to relative sea-level fall, although climatically produced changes in discharge and/or sediment supply may independently cause incision, even in areas far removed from the coast. In the case of valleys in coastal areas, fluvial deposition typically begins at the mouth of the incised-valley system when sea level is at its lowest point and expands progressively farther up the valley as the transgression proceeds, producing depositional onlap in the valley. Based on the longitudinal distribution of broad depositional environments, the length of an incised valley can be divided into three *segments*. Ideally, the fill of the seaward portion of the incised-valley (*segment 1*) is characterized by backstepping (lowstand to transgressive) fluvial and estuarine deposits, overlain by transgressive marine deposits. The middle reach of the incised valley (*segment 2*) consists of the drowned-valley estuarine complex that existed at the time of maximum transgression, overlying a lowstand to transgressive succession of fluvial and estuarine deposits similar to those present in segment 1. The innermost reach of the incised valley (*segment 3*) is developed headward of the transgressive estuarine–marine limit and extends to the point where relative sea-level changes no longer controlled fluvial style (i.e., to the landward limit of sea-level-controlled incision). This segment contains only fluvial deposits; however, the fluvial style changes systematically due to changes in the rate of change of base level. The effect of base-level change decreases inland until eventually climatic, tectonic, and sediment-supply factors become the dominant controls on the fluvial system. In valleys far removed from the sea, the fill consists entirely of terrestrial deposits, but shows changes in fluvial style that are similar to those in segment 3, even though the stacking patterns are controlled more by local tectonics and climate.

Recent and future development of estuarine and incised-valley facies models has emphasized the use of ichnology to recognize brackish-water deposits and the ability to subdivide compound valley fills on the basis of sediment composition. Imaging the valley and its fill has been greatly improved with 3D and 4D seismic techniques. Seabed mapping of modern estuaries has enabled detailed distributions of facies and morphology to be compiled, enhancing the ability to predict these features in ancient rocks. Our current set of facies models represents the early classification stage in the development of depositional models. The appropriate way forward appears to be a transformation from qualitative approaches to empirical and quantitative computer-based models with predictive capability, based on a thorough understanding of the dominant processes operating in each environment.

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INTRODUCTION

Estuaries and incised valleys and their preserved deposits (E&IVs; Figure 1) are important depositional settings that have widespread significance as the sites of human habitation, as harbors, as the host of significant hydrocarbon reserves, and as the repository of important information on lowstand to early transgressive sedimentation in ramp and shelf settings. Petroleum explorationists, in particular, have focused on E&IV deposits because of the economically significant quantities of hydrocarbons produced from reservoirs hosted by the fill of incised valleys (Table 1; e.g., Zaitlin and Shultz, 1984, 1990; Van Wagoner et al., 1990; Brown, 1993; Dolson et al., 1991). For example, Brown (1993) estimated that ~ 25% of worldwide off-structure conventional petroleum traps in clastic reservoirs are hosted within incised-valley systems, with the single largest petroleum reserves in the world (the Athabasca Tar Sands) being hosted by incised-valley deposits. Therefore, a clear understanding of the internal facies architecture, reservoir characteristics, and production behavior of incised-valley systems is of critical importance to the exploration for and exploitation of incised-valley reservoirs.

Estuaries and incised valleys are a complex and possibly unique grouping of sedimentary environments, inasmuch as their formation and development involve creation of depositional space mainly by one process (fluvial erosion) and the filling of that space by a range of other processes (fluvial, tidal, and wave), in the presence of water of variable salinity. It is the close association of incised valleys with estuarine fill that has resulted in these two environments being considered together here in a linked facies-model approach. The other main components of incised-valley fills (fluvial and marine sediments) are treated in separate sections of this publication.

Estuaries are also complex environments in that they contain the interrelated depositional products of wave, tide, and river processes within a relatively restricted area. This complexity caused the development of facies models for estuaries to lag behind those of most other adjacent environments such as rivers or beaches. For example, earlier editions of the Facies Models text (Walker 1979, 1984a, Walker and James 1992) did not have stand-alone consideration of either estuaries or incised valleys.

Because of the nature and complexity of E&IV facies models, this paper begins with a section on the development of both of these fields, to place the concepts in their historical framework. The next section details the authors' approach to facies models in general and the place of E&IV models in that approach. The remainder of the paper consists of outlining the current facies models for E&IVs, discussing how to use those models in practical applications, illustrated by reference to both ancient and modern examples and case studies. It concludes with a section on recent and future developments in the field.

HISTORICAL DEVELOPMENT OF IDEAS ON ESTUARINE AND INCISED-VALLEY FACIES MODELS

Incised Valleys

The following discussion represents a short historical summary of facies models for E&IVs. For more detail, the reader is referred to excellent reviews such as Dalrymple et al. (1994b),

TABLE 1.—Super-giant petroleum reservoirs hosted within incised-valley (IV) deposits (i.e., reservoirs with reserves > 50 MMBOE estimated ultimate recovery). Summarized from Dolson et al. (1991) and Pulham (1994).

Field/Trend	Basin	EUR (Mmboe)	Age	Environment
Athabasca Oil Sands Canada, Alberta	Western Canada Sedimentary Basin	665000	Cretaceous	Fluvial–Estuarine-IV
Messla-Faregh Libya	Sirte	1500	Cretaceous	Fluvial–Estuarine-IV
Burbank Oklahoma	Mcalester	500	Pennsylvanian	Fluvial-IV
Cutbank Montana	WCSB	199	Cretaceous	Fluvial-IV
Hilight Wyoming	Powder River	108	Cretaceous	Fluvial–Estuarine-IV
Churches Buttes Wyoming	Green River	77	Cretaceous	Fluvial-IV
South Glenrock Wyoming	Powder River	75	Cretaceous	Fluvial–Estuarine-IV
Clinton Oklahoma	Anadarko	67	Pennsylvanian	Fluvial-IV
Adena Colorado	Denver	60	Cretaceous	Estuarine-IV
Clareton Wyoming	Powder River	60	Cretaceous	Fluvial-IV
Stockholm–Arapahoe Kansas	Las Animas Arch	50	Pennsylvanian	Fluvial-IV
Cusiana Mirador Fm. Colombia	Llanos	> 100	Eocene	Fluvial–Estuarine-IV

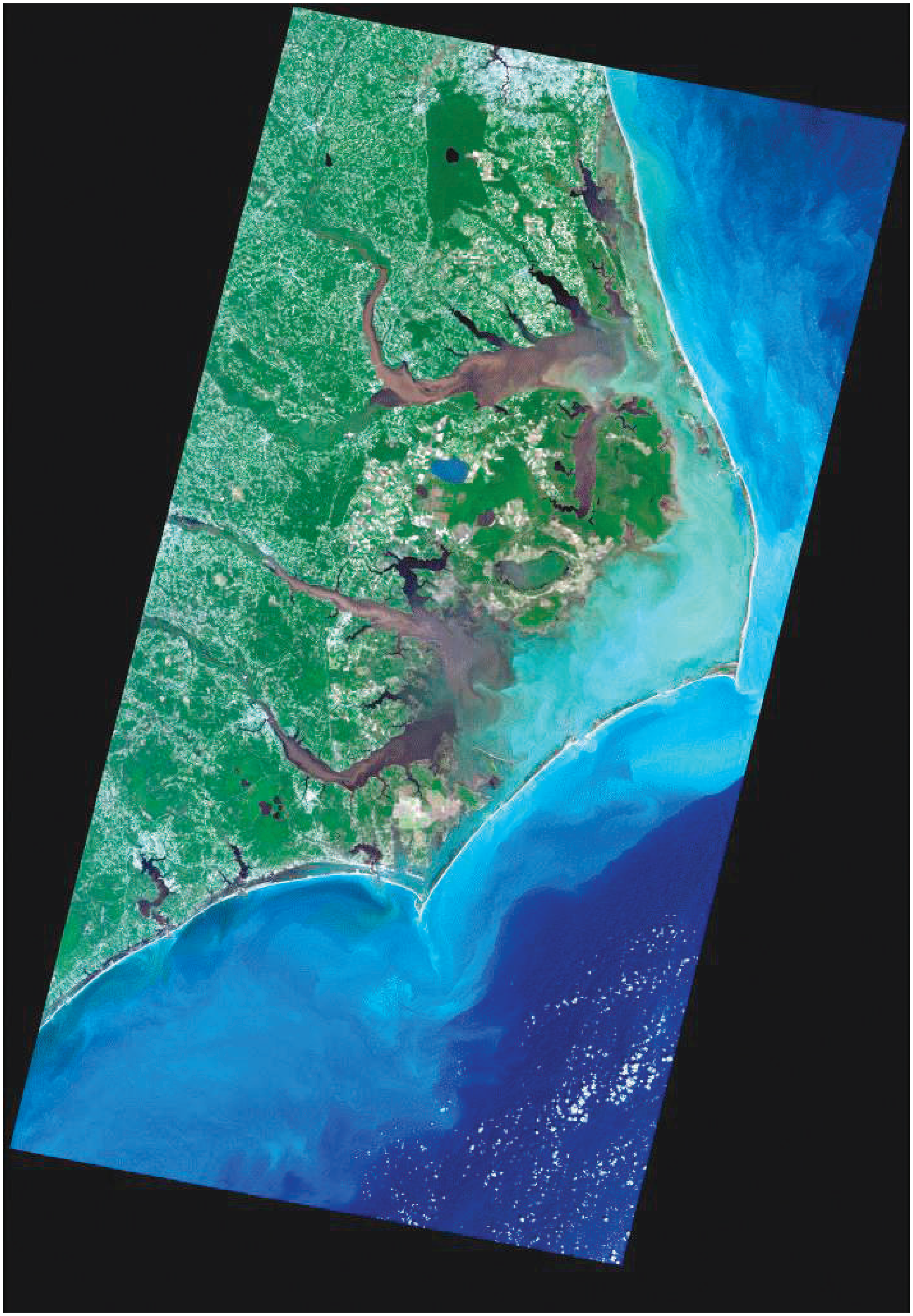


FIG. 1.—Landsat image of the Pamlico–Albermarle Sound area of North Carolina showing a range of incised valleys, estuaries, and lagoons. The valleys of the Pamlico, Neuse, and Roanoke rivers were incised during sea-level lowstand and have since been flooded by relative sea-level rise. This has transformed them into wave-dominated estuaries with extensive estuary-mouth-barrier and tidal-inlet systems. The regions between the valleys are also flooded and flanked seaward by barriers, tidal inlets, and tidal deltas, but are better described by the term lagoons.

Miall (1996), and Blum and Törnqvist (2000), from which parts of the following are derived.

The development of ideas on incised valleys is closely linked to the investigation of fluvial processes, and these have a long history of study back to Greek and Roman times. Plato and Homer both were aware of fluvial sedimentation processes. Herodotus around 450 B.C. realized the connection between the Nile River, its valley, and the deposits at its seaward end, to which he applied the term "delta". The term "valley" itself derives from Latin and Old French origins meaning "a long depression or hollow lying between hills or stretches of high ground and usually having a river or stream flowing along its bottom" (Oxford English Dictionary). It was not until the eighteenth century that more specialized study was devoted to river valleys, and this was mostly a result of the attempt to assign an age to the Earth. Hutton and his successor Playfair (1802) used the idea that river valleys were the product of long-term fluvial erosion to assign a much greater age to the earth than the opposing Neptunist concepts of recent catastrophism and origin from floods. These uniformitarian themes using fluvial processes were further developed in the classic work of Lyell (1830).

However, it was not until later in the nineteenth century that the concept of river grade and the fluvial equilibrium profile were developed, and their relationship to valley erosion and fluvial deposition was appreciated. Among the first to address these concepts were Powell (1875), Gilbert (1880), who developed the idea of base level, and Davis (1908), who illustrated the successive widening of a valley with age and the influence on the valley profile of strata of varying resistance. Around the same time, Penck and Brückner (1909) suggested a climatic control for the origin of valley terraces in southern Germany, thus initiating a continuing debate between climatic and fluvial control on valley development and fluvial deposition (e.g., Fisk 1944, 1947; Blum 1990, 1994; Blum et al. 1994; Blum and Törnqvist, 2000).

The next major development in our understanding of rivers and valleys came in the mid-twentieth century with the ideas of Lane (1935, 1955) and Mackin (1948), who took a more hydrodynamic approach and discussed the effects of equilibrium and the graded stream in terms of discharge, load, slope, and base-level parameters. Fisk's (1944) landmark work on the Mississippi River developed many of these concepts into a detailed approach to a single drainage system that considered its response to both internal sediment parameters and outside forcing by sea-level changes. Quantitative fluvial geomorphology themes were continued by Leopold and Wolman (1957), Leopold et al. (1964), and later Schumm and co-workers (e.g., Schumm 1972, Schumm and Khan 1972, Ethridge and Schumm 1978). At around the same time, incised-valley deposits were being recognized as hydrocarbon reservoirs for the first time. One of the earliest and best-described examples of a subsurface depositional system meeting the criteria of an incised valley was that of Harms (1966) in his description of stratigraphic traps within the extensive system associated with the Cretaceous "J" Sandstone in western Nebraska (Fig. 2).

Concurrently through the twentieth century, concepts of fluvial facies models were slowly being developed, beginning in the modern sense with the work of Melton (1936), Mackin (1937), and Happ et al. (1940), and also in Fisk's Mississippi studies (1944, 1947), culminating in the first major fluvial facies models developed by Allen (1963, 1964, 1965). Further developments in fluvial facies models were summarized in the first edition of Facies Models (Walker, 1979) drawing on many studies of the 1960s and 1970s integrated in papers such as Cant and Walker (1976, 1978), Miall (1977, 1978), and Rust (1978a, 1978b).

However, the majority of these advances did not deal with the longer-term evolution of river systems. Instead, they examined instantaneous fluvial geomorphology, sedimentary structures, bedforms, paleocurrents, and empirical relationships between parameters, finally integrating these features into static facies

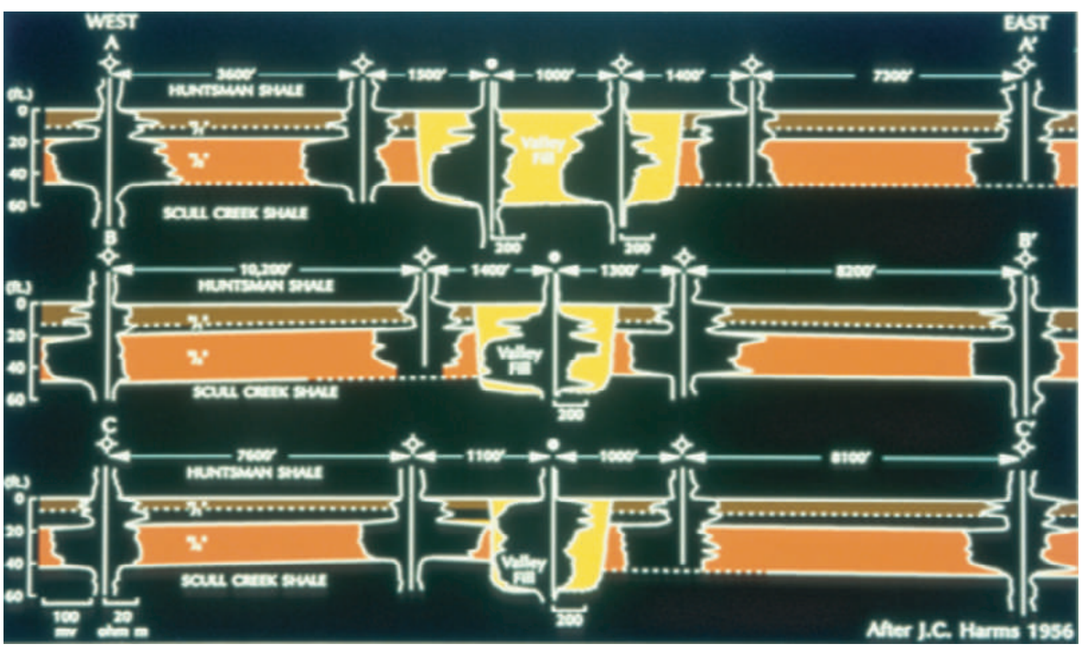


FIG. 2.—Wireline-log cross section from Harms (1966) showing one of the first and best-described examples of a subsurface depositional system meeting the criteria of an incised valley—the Cretaceous "J" Sandstone in western Nebraska. The figure shows the J Sandstone as "valley fill", incising the regional Skull Creek Shale and Huntsman Shale.

models for meandering and braided rivers. These studies were concentrated in terrestrial settings not linked to coastlines. They were also concerned mainly with the detailed nature of river deposits and not with a holistic approach to landscape evolution that encompassed river-valley incision and the large-scale stratigraphic organization of deposits within those valleys.

Developmental work on fluvial facies models concentrated more on the products of deposition than erosion and hence moved away from the early stratigraphic emphasis on unconformities (e.g., Blackwelder, 1909; Schuchert, 1927). A change back towards these larger-scale themes was precipitated by the development of seismic stratigraphy and later sequence stratigraphy in a series of papers by Vail and co-workers presented first in Payton (1977) and followed up in Posamentier and Vail (1988) and other related papers in SEPM Special Publication 42 (Wilgus et al., 1988). In the latter publications, incised valleys were seen as an integral component of a depositional sequence, formed during periods of decreasing and low accommodation (e.g., Van Wagoner et al., 1990). They were interpreted to form by fluvial incision at the exposed shelf break and to extend across the continental shelf and into the adjacent coastal plain. In this sense the application of the word incised, meaning "cut into" (Oxford English Dictionary) together with the word valley, was used to mean a valley that was eroded actively as a result of allocyclic factors (particularly falling relative sea level), to distinguish it from a valley resulting from other means (e.g., tectonic processes such as graben formation in a rift valley; Leeder and Gawthorpe, 1987). Thus, the criticism leveled by Blum and Törnqvist (2000) that all valleys are incised valleys is not valid when the term is used in a broad sequence-stratigraphic sense.

Further advances in the recognition of incised-valley deposits and documentation of fill styles were made in Van Wagoner et al. (1990), another book from the Exxon school of sequence stratigraphy. The rapid acceptance of sequence stratigraphy as the preferred method for stratigraphic analysis and hydrocarbon exploration placed a new significance on the recognition of incised-valley deposits and energized the decade of the 1990s to produce the first integrated facies models for these systems. As a result, facies models for E&IVs were the first to explicitly include a sequence-stratigraphic approach, and their usage increased rapidly (Fig. 3).

Advances in sequence stratigraphy and its emphasis on the evolution of depositional systems were instrumental in the development of these integrated dynamic models as compared to the more static or autocyclic focus of earlier facies models. A special session at the 1992 AAPG conference in Calgary was the source of many of the papers that made up SEPM Special Publication 51 on incised valleys (Dalrymple et al., 1994a). This publication presented the first integrated facies model for an incised-valley system (Zaitlin et al., 1994), together with summaries of the history of incised-valley research (Dalrymple et al., 1994b) and the origin, evolution, and morphology of fluvial valleys (Schumm and Ethridge, 1994). A further 19 papers described a range of incised-valley deposits. More recently, a 2003 SEPM research conference on incised valleys produced an updated collection of research papers in another SEPM Special Publication (Dalrymple et al., 2006).

Estuaries

Early work on applied and environmental aspects of estuaries is plentiful because of the widespread utilization of estuaries as

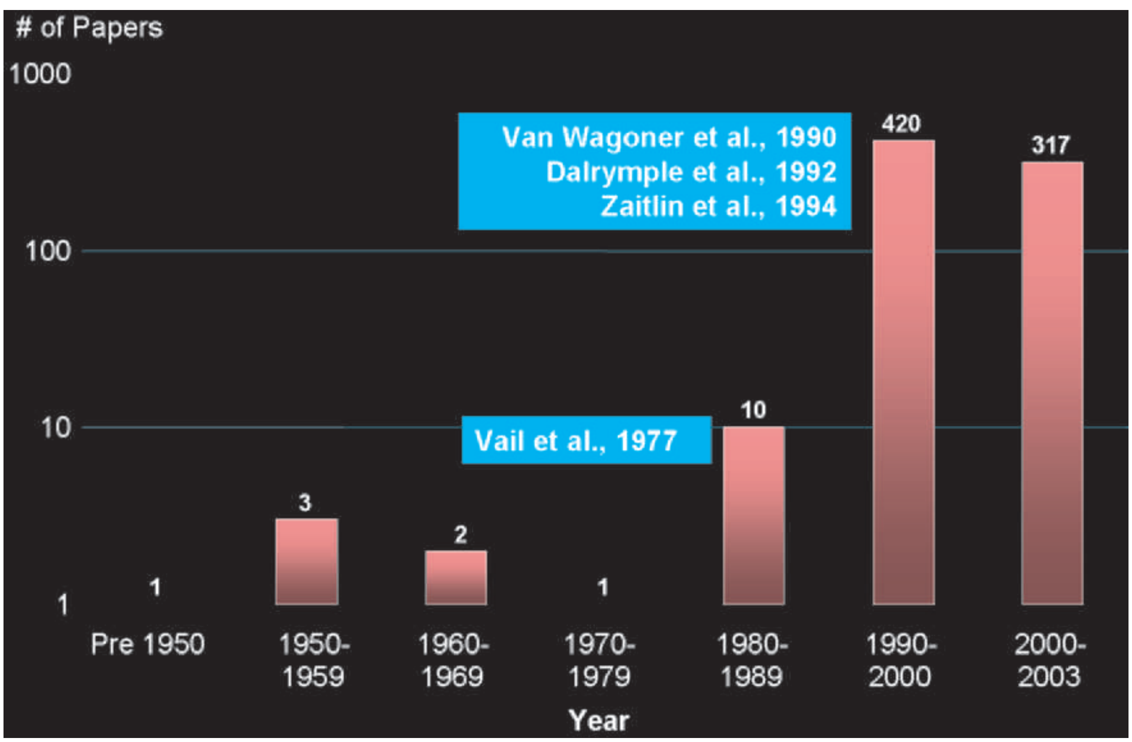


FIG. 3.—A search of the Georef data base (www.agiweb.org/georef) for the term "incised valley" shows a significant increase in usage during and after the 1980s, reflecting the widespread acceptance of the sequence-stratigraphy concept (e.g., Posamentier and Vail, 1988). Significant papers are shown in blue boxes.

harbors and ports (e.g., the Thames, the Hudson, and the Mersey; e.g., Preddy, 1954; Hughes, 1958) or living space (the Dutch lowlands; Oomkens and Terwindt, 1960). The presence of tidal influence is fundamental to the concept of an estuary, and the Oxford English Dictionary defines an estuary as "the tidal mouth of a great river, where the tide meets the current of fresh water" or more fully as a "semi-enclosed coastal body of water which has a free connection with the open sea and where fresh water, derived from land drainage, is mixed with sea water. Estuaries are often subject to tidal action..." (Allaby and Allaby, 1999). However, this is primarily an oceanographic definition derived from Pritchard (1967) and is difficult to apply to sediments and ancient rocks. It can also be ambiguous in a geological context because the active progradational distributaries of a delta such as the modern Mississippi can fulfill this definition of an "estuary" despite having profound stratigraphic differences from those drowned (i.e., transgressive) river mouths, which are also considered to be estuaries.

Early geological studies of the modern Severn estuary were conducted by Sollas (1883), who noted upstream sediment transport and determined a vertical stratigraphic succession. Other early studies of estuarine sediments were conducted in the Bay of Fundy (Kindle, 1917) and the Dutch estuaries and tidal flats (e.g., Oomkens and Terwindt, 1960; Terwindt, 1963; Van Straaten, 1952, 1954a, 1954b, 1961). Kindle also noticed upstream sediment transport, while Van Straaten (1952, 1954a) developed a model for tidal-channel migration (Fig. 4) well before Allen's (1963) fluvial version. Much early work seems to have differentiated between tidal flats and estuaries (e.g., Klein 1967); however, many of the tidal-flat studies were on sediments that were components of larger estuaries (e.g., the Bay of Fundy and the Dutch and German North Sea coasts). Summaries of estuaries were produced first in the early to mid-twentieth century (Twenhofel, 1932; Emery and Stevenson, 1957) and commonly dealt with the oceanographic and biological aspects, reinforced by detailed physical oceanographic studies such as Rochford's (1951) Australian work. Studies of ancient estuarine sediments were rare in the early twentieth century and included Arkell (1933), Pepper et al. (1954), and Allen and Tarlo (1963). A useful early comparison summary of modern and ancient estuarine and tidal-flat sediments was provided by Klein (1967).

However, while estuarine research concentrated on circulation studies and sediment dynamics, and models for fluvial systems, coasts, and deltas matured slowly, estuarine facies models did not make comparable advances. Schubel and Hirschberg as late as 1978 noted that "estuarine deposits rarely can be delimited unequivocally from other shallow marine deposits in the geological record because of their limited areal extent, their ephemeral character and their lack of distinctive features". However, great strides in understanding and recognizing estuarine sediments were made from the 1960s to the present, such that integrated models for estuaries were finally available by the 1990s (e.g., Dalrymple et al., 1992).

Some of the first major steps forward in understanding the geology of estuaries were the conferences held at Jekyll Island, Georgia (Lauff, 1967) and Myrtle Beach, South Carolina (Cronin 1975). In the published volumes from these two conferences, an oceanographic definition of an estuary was formulated (Pritchard, 1967) and later largely accepted. In addition, geomorphological elements of an estuary were defined (e.g., Russell, 1967; Steers, 1967, Jennings and Bird, 1967), rates of sediment transport and accumulation were determined (e.g., Postma, 1967; Rusnak, 1967), and studies that indicated the tripartite sedimentary subdivision of an estuary were presented (e.g., Kulm and Byrne, 1967; see Figure 5).

Numerous studies of the morphology and evolution of tidal inlets characterized work in the 1960s and 1970s (Hoyt and Henry, 1965; Vallianos, 1975; Oertel, 1975; Hine, 1975; Hubbard,

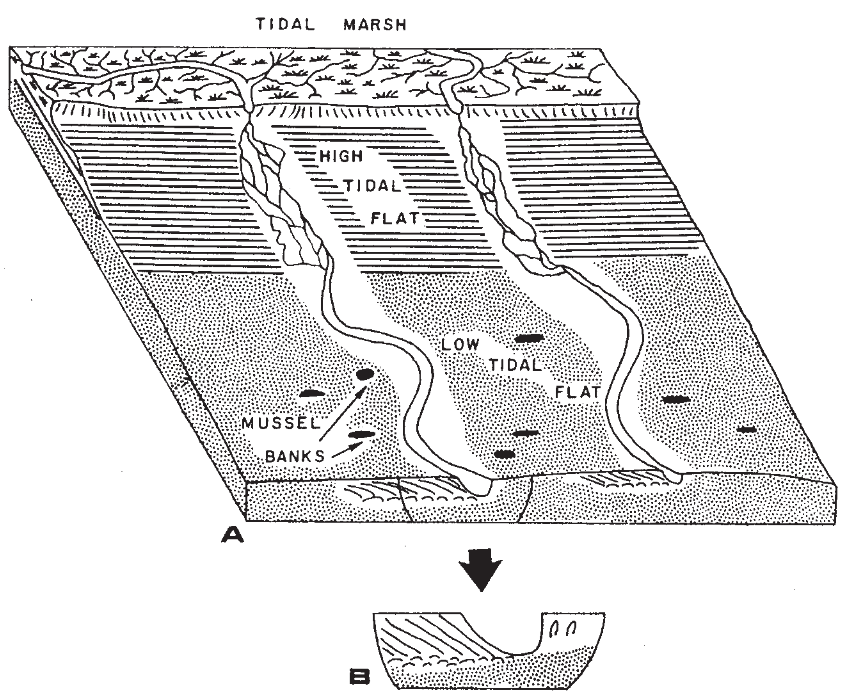


Fig. 4.—**A)** Block diagram and **B)** enlarged cross section of tidal-flat and tidal-channel sediments in the Dutch Wadden See (from Van Straaten, 1952, 1954, as modified by Klein, 1967.)

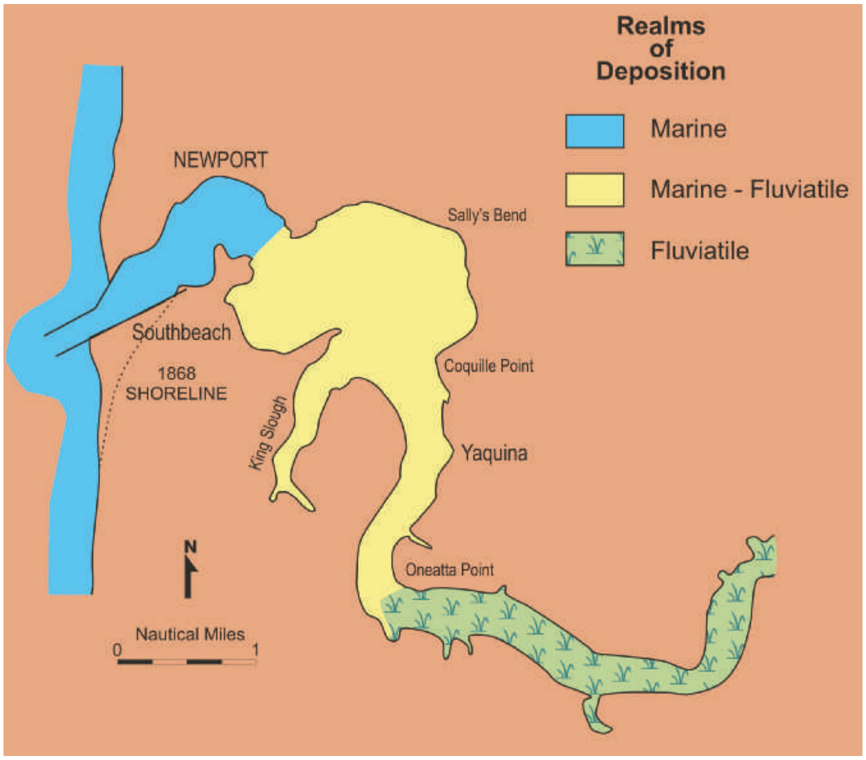


FIG. 5.—Early example of tripartite estuarine sedimentation zonation, Yaquina Bay, Oregon (original from Kulm and Byrne, 1967).

1975). Many of these studies were influenced by the estuary and tidal-inlet ideas of M.O. Hayes, who provided the first comprehensive sedimentary models for these settings in his classic 1969 and 1975 publications. Hayes (1975) also provided the basis for division of estuaries into microtidal, mesotidal, and macrotidal categories, following the tidal classification system of Davies (1964). These advances in modern systems began to be translated into detailed studies of ancient successions by authors such as Land (1972) in the Cretaceous of the Rocky Mountains, Bosence (1973) in the Eocene London Basin, and Horne and Ferm (1976) in the Carboniferous of the Appalachians. Beginning in 1985 and continuing through 2004, research symposia on clastic tidal sediments (e.g., de Boer et al., 1988; Smith et al., 1991; Bartholdy and Pedersen, 2004) have provided valuable studies of many modern and ancient tidal deposits, including documentation of the tidal sedimentary structures by which tidal deposits can be recognized. More recently, databases and volumes dealing with the distribution of estuaries across entire continents have been developed, such as those for Australia (www.ozestuaries.org) and South America (Perillo et al., 1999).

However, although extensive research continued on estuaries, no comprehensive model identifying and integrating the range of geomorphological and sedimentary elements was developed. Clifton's (1982) summary catalogued many estuarine sedimentary structures and proposed a tidal-channel succession. Roy (1984) summarized much research on Australian wave-dominated estuaries in a paper that identified a geomorphological evolution that is the basis of many later models. Nichols and Biggs (1985) provided an extensive review of estuaries, and, although summarizing processes and sediment dynamics comprehensively, noted that "it is still difficult to hindcast with certainty under what conditions and in what manner the sediment accumulated". Most of this earlier research tended to focus on wave-dominated rather than tide-dominated systems and on coastal segments that were not necessarily associated with river mouths.

Reinson's (1992) and Dalrymple's (1992) reviews in the third edition of Facies Models (Walker and James 1992) began to synthesize much of the earlier work on estuarine facies and facies successions and began to focus more on the role of tides. In this 1992 volume an early classification was developed (Fig. 6; Reinson, 1992), diagnostic sedimentary structures were identified, and summary vertical successions were provided. In addition, some integrated local studies had begun to assemble all of the basic elements required for later facies models in modern environments (e.g., Allen 1991; Dalrymple et al., 1990; Nichols et al., 1991), and in ancient rocks (Zaitlin and Schultz, 1984, 1990; Demarest and Kraft, 1987; Rahmani, 1988; Wood and Hopkins, 1989). By 1992, Dalrymple et al. had integrated many of these ideas into a conceptual facies model for estuarine systems that contained a geological definition of an estuary. This work has provided the main focus for research since then.

BACKGROUND TO FACIES MODELS AND THEIR APPLICATION TO ESTUARIES AND INCISED VALLEYS

Theoretical Basis of Facies Modeling

The facies-model concept as formulated by Walker (1984b, 1992) provides "a general summary of a depositional system written in terms that make the summary usable in at least (the following) four different ways": (1) As a norm for comparison, (2) As a framework and guide for future observations, (3) As a

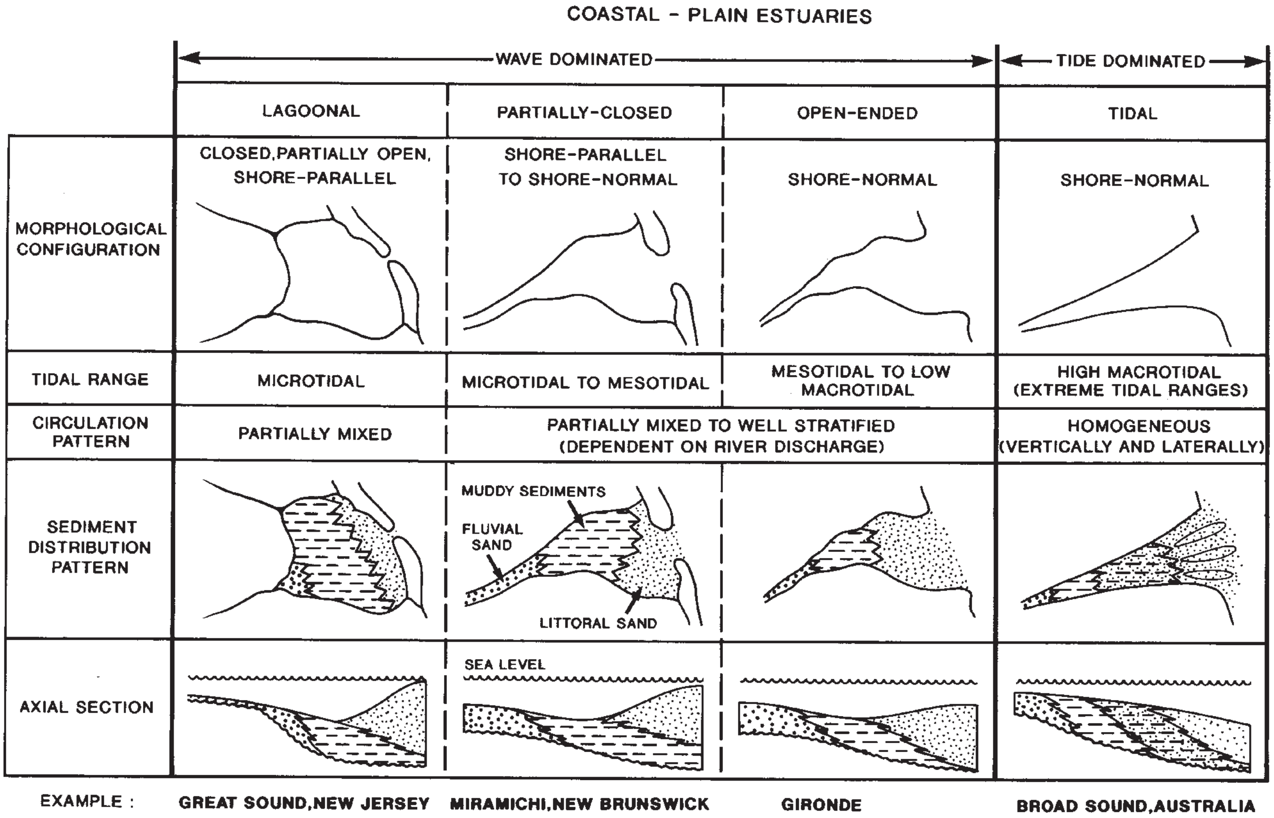


FIG. 6.—Early estuarine classification from Reinson (1992).

predictor in new geologic situations, and (4) As an integrated basis for interpretation for the system it represents. In practice, this often translated into an "idealistic" vertical succession of facies and/or a 3D block diagram of facies relationships that supposedly portrays the "essence" of the environment. Early facies models had only one or a limited number of vertical successions, and 3D block diagrams showed little internal information beyond the top and side panel(s) of the diagram.

An important extension of this approach is the display of a range of vertical successions in different parts of the model (e.g., for deltas; Coleman and Prior, 1980; Galloway and Hobday, 1996) or a spectrum of vertical successions that illustrate the variability that is possible, as illustrated by the multiple models for braided-fluvial deposits by Rust (1978a), Rust (1978b), and Miall (1978), summarized into the 16 "models" provided for fluvial systems by Miall (1996). However, when this approach is extended to its logical conclusion, the number of "models" can proliferate, and hence lose the ability to provide a relatively simple environmental summary. In this sense the proliferation of models brings into question the provision of a "norm" (use #1 above, as discussed by Anderton, 1985). Part of the problem here is the degree to which each researcher utilizes the technique of "distillation" (Walker, 1984b), in which local variability is removed and replaced with a simplified model that is based on a summary of the representative geomorphology and facies (i.e., an idealized view of what should occur at a specific place on the earth's surface). Hence we have problems in appreciating (for example) what the ideal view of a delta is when we have to confront the contrasts between a temperate-climate river-dominated mid-latitude delta and a frozen arctic delta or a tide-dominated tropical delta.

In our search for facies and geomorphological simplicity we may have neglected the fundamental basis for our development of models, which lies in the characteristic processes that control sedimentation in any one depositional setting. Hence we should not expect a single model for the deltas listed above, but we should expect that all of them follow similar physical laws such as the dispersion of suspended sediment, the response of bed material to wave motion and the action of biological agents in the presence of a salinity gradient. Thus, the key to understanding depositional environments is to identify the processes that operate in each one and to determine their sedimentary response or combination of responses. For example, the combination of waves, longshore-directed currents, and offshore-directed rip currents in the surf zone makes for a unique process environment. If we can identify the corresponding sedimentary responses and deposits for this combination of processes we will have generated a model that summarizes those deposits and their formational processes. It may not be the **only** model for nearshore marine settings, but it should be the only one that experiences that specific process combination. We then need to examine the physical, chemical, and biological processes of an environment, as well as the properties of the sediment supplied to it, to determine the **range** of possible outcomes for that environment. Secondly we need to determine the **probability** of occurrence of those process combinations and sediment types. Our ideal facies model then becomes one that covers the environmental range but recognizes the most probable combination of processes and sediments (this is the real distillation process of Walker, 1984b). Many situations are possible in the real world, but only a small number are common. Environments with many variables that do not display clustering of common processes and sediment types will not produce a single representative, useful facies model. On the other hand, the best facies models will result from environments with few variables that exhibit frequent repetition of the same process combinations. Our approach to building an ideal facies model should then be a quantitative approach that models the processes and sediments and is capable of creating the full range of process–sediment interactions in an environment. Examples of this approach are Syvitski and Daughney (1992) as applied to deltas, or Cowell et al. (1992, 1995) as applied to transgressive continental shelves. Our observations derived from experiments and field work provide the experience that identifies the processes and geomorphological components, and the probability of encountering the individual examples throughout the range of possibilities.

The response to the process combination in each part of the environment will be a 3D sediment body of a particular shape that contains a number of characteristic properties. Sediment bodies of this sort have been termed architectural elements (e.g., Miall, 1985) and equated with facies successions by Walker (1992). The frequent association of processes results in architectural elements occurring in common relationships with other adjacent or linked elements. An example is the frequent association of river flood plains with levees and channels because of the linked processes of channel hydraulics and flooding. Because of the direct link between processes and facies models, the critical laboratory for constructing models is the modern environment, where the interplay between sedimentary process and product can be observed and recorded in a wide range of settings. Modern environments are also becoming better suited to the documentation of sedimentary architecture with the advent of high-resolution seismic surveys (particularly 3D surveys), ground-penetrating radar, multibeam bathymetric surveys, and other remote-sensing techniques such as resistivity surveying. Ancient examples are not as useful because of the possibility of ambiguity in interpretation of the contemporaneous processes (the Shannon Sandstone is a celebrated although extreme example; see for example, Suter and Clifton, 1999) and the inability to observe those processes directly. Nevertheless, once processes have been documented and understood, observation of their depositional products in ancient rocks can be used to: (1) provide good information on the 3D geometry of the deposits, (2) extend the range of variability and scale for examples (such as ice-house versus hothouse climates and a wide range of tectonic basin settings), (3) document paleogeographic development and preservation potential through time, as well as (4) provide the only information on non-uniformitarian situations such as the pre-Silurian terrestrial processes prior to the advent of land plants and the widespread presence of microbial mats prior to the advent of metazoan grazers in the latest Precambrian (MacNaughton et al., 1997).

The process characterization of an environment takes place at two scales, the local, **autocyclic scale** and the regional to global **allocyclic scale**. In the first case the controlling variables are things like fluid shear, salinity, density, and sediment size. The response is the production of distinctive sedimentary bodies that reflect the genetic process—these bodies are facies, facies successions, and architectural elements, and the sedimentological features that they contain, such as bedding structures, bioturbation, and their geometry. In the second (allocyclic) case, the controlling variables are accommodation (the space made available for sedimentation, *sensu* Jervey, 1988) and the amount and textural character of the sediment flux as determined by tectonic, climatic, and sea-level behavior. The sediment responses here are the production of distinctive bounding surfaces, and the generation, preservation, and juxtaposition of stratigraphic units, including their stacking patterns. A facies model for any one environment should take into account both the autocyclic products to provide the building blocks and the allocyclic products that describe the geometric arrangement of those building blocks into the finished end product.

These principles can be explicitly applied to E&IV facies models. In this case, the processes involved are primarily a combination of fluvial, wave, and tidal processes supplemented locally by other processes such as organic production (e.g., peat or shell), wind, and density stratification. Wave and tidal processes provide a range of possibilities in estuarine systems, generating a spectrum between macrotidal, tide-dominated settings and microtidal, wave-dominated settings. The combination of all E&IV processes produces a range of characteristic morphological elements including river channels and flood plains, bayhead deltas, estuarine central basins, barriers and tidal inlets, tidal deltas, and tidal sand flats and ridges. The allocyclic variables produce fluvial incision during decreases in sediment input, increases in water flux or lowered relative sea level, and fluvial deposition followed by estuarine deposition during increases in relative sea level and the landward migration of fluvial, estuarine, and marine lithofacies. Estuarine facies models are amongst the most complex due to the occurrence of multiple dominant processes (river, wave, and tide) and specific varied responses to a range of relative sea-level and sediment-flux parameters. This complexity contributed to the slow development of facies models for E&IV systems.

FORMATION AND FILL OF INCISED VALLEYS

Incised valleys are containers. They are significant stratigraphic entities because they create a localized space in which sediment can accumulate, often in areas where space may be uncommon otherwise (such as the coastal plains of low-accommodation basins). Incised valleys should be regarded as a system in which there are two components, the valley and its fill. These components may or may not be related in time or formational process. To understand the incision of a valley by fluvial processes (the only mechanism we will address here, neglecting valleys of structural or tectonic origin) we must consider the sediment continuity equation , which can be written in its simplest one-dimensional form as

$$dz/dt + dq_s/dx = 0$$

where z = bed elevation, t = time, q_s = width-averaged sediment transport rate, and x = distance along the channel. Blum and Törnqvist (2000) show how this equation can be used to identify channel incision (an increase in z) as the result of the sediment transport capacity exceeding the sediment supply. Steeper slopes and coarser grain sizes increase the magnitude and rate of incision. Incision can result from a change in climate, tectonics, or sea level, with climate and tectonics becoming more important landward from the shoreline (Shanley and McCabe, 1994).

Much of the modern significance associated with incised valleys derives from their association with sequence-stratigraphic concepts (e.g., Posamentier and Vail, 1988; Van Wagoner et al., 1988; Van Wagoner, 1990; Van Wagoner et al., 1991) and economic importance (e.g., Brown, 1993; Dolson et al., 1991). In areas on the margin of a marine basin, incised valleys are considered to have formed primarily in response to a fall in relative sea level and a resulting decrease in accommodation, and are associated with a regional unconformity. Such a response requires a specific coastal-plain and continental-shelf geometry to satisfy the sediment continuity equation. In particular, for the sedimenttransporting capacity of the stream to increase during sea-level fall, the river must encounter a significant increase in gradient (a knickpoint) somewhere seaward of the highstand shoreline (e.g., Summerfield, 1985; Schumm, 1993). In other words, the fluvial equilibrium profile lies below the level of the land surface (Summerfield, 1985). The incision initiated at this location then propagates headward to create the valley. In areas with a relatively low-gradient shelf and a distinct, exposed shelf–slope break, a knickpoint generally coincides with the shelf edge. However, in cases without a distinct shelf break, or where the shelf edge lies below the lowstand elevation, incision may not extend to the shelf edge; instead, recent studies of several such shelves have shown that incision begins at the break in slope associated with an earlier lowstand shoreline (e.g., Woolfe et al., 1998; Posamentier, 2001; Fielding et al., 2003; Wellner and Bartek, 2003) and/or with the immediately preceding highstand coastline.

The lateral extent along the stream channel that can be affected this way is highly debated (see Blum and Törnqvist, 2000), but Quaternary examples suggest that incision across the entire exposed continental shelf is possible if sea level falls below the shelf edge (e.g., Suter and Berryhill, 1985), and that incision upstream of the highstand shoreline (e.g., Ethridge et al., 1998) is possible for some tens to hundreds of kilometers: Blum and Törnqvist (2000) suggest a range of from 40 to 400 km for the upstream limit of coastal onlap. Examples of ancient incised valleys can reach hundreds of kilometers in length if the sea-level fall is of sufficient duration and magnitude. The Mississippian Morrow Formation along the Sorrento–Mt. Pearl–Siaana and Stateline trends is such an example of a well-documented valley form that is mappable over hundreds of kilometers (e.g., Krystinik and Blakeney-DeJarnett, 1994; Krystinik, 1989; Bowen and Weimer, 1997). Another documented subsurface example of a long incised valley is provided by the Lower Cretaceous Basal Quartz and its time-equivalent units (e.g., Hayes et al., 1994; Zaitlin et al., 2002; Leckie et al., 2005). The several valleys forming this compound valley fill can be traced for over 800 km south to north in the Western Canadian Sedimentary Basin. Other examples of throughgoing valley systems include the Pennsylvanian of the Illinois Basin (Howard and Whitaker, 1988), the Permian of west-central Texas (Bloomer, 1977), the Lower Cretaceous Glauconitic Formation of Alberta (Sherwin, 1994), the Lower Cretaceous Viking–Muddy equivalents in western U.S.A. (Harms, 1966; Weimer, 1984; Reinson et al., 1988; Martinsen et al., 1994; Porter and Sonnenberg, 1994), and the Upper Cretaceous Dunvegan Formation, Alberta (Plint, 2002; Plint and Wadsworth, 2003). More localized incision is also possible: in cases where sea level does not fall very far, incision may occur only in the vicinity of the immediately preceding highstand shoreline as a result of the relatively steep slope of the highstand shoreface. Distinguishing such localized incisions from tidal inlets may be difficult. In areas far removed from the sea, incision can be induced by increases in slope caused by tectonic activity or by an increase in the ratio of water discharge to sediment discharge: determining the cause(s) of incision in an ancient example can be very difficult.

Incised-valley filling is also highly dependent on the relationship between accommodation and sediment flux, with filling beginning when the fluvial equilibrium profile rises above the level of the valley base. Clearly, because valleys are incised by fluvial processes, one can expect fluvial sediments to be deposited at the base of the valley, even if these deposits are only one meander-belt or channel-bar height thick. In the case of valleys cut into coastal plains, these fluvial deposits have a marine influence for some distance landward of the lowstand shoreline but lack marine influence farther inland. The facies boundary between tidal–fluvial and purely fluvial deposits migrates landward as base level rises. Landward of the marine limit of inundation during relative sea-level highstand, valley-fill deposits consist entirely of fluvial, lacustrine, and organic facies (e.g., Shanley and McCabe, 1994). If there is sufficient terrestrial sediment supplied during valley filling, the valley may be both cut and filled by fluvial processes. If the valley remains at least partially unfilled after sea-level lowstand, then the downdip end experiences estuarine sedimentation during the subsequent transgression. Seaward of the highstand shoreline, if the valley is still underfilled after transgression, some of the valley fill is marine and includes shelf sand and mud facies. In valleys far removed from coastal areas, all of the valley fill is fluvial in nature. Terrestrial and marine sediments are covered in detail elsewhere in this volume and will not be considered further here. Instead we will concentrate on identifying the character of estuarine sediments that are a common component of valley fills in coastal areas and developing an appropriate facies model for them. Later we will return to see how estuarine sediments fit into an overall facies model for incised-valley systems.

COASTAL CLASSIFICATION

Defining precisely what is or is not an estuary, and providing a useful geological classification scheme for estuaries, as a necessary basis for creating a facies model, has been a long-standing problem in coastal studies. In order to solve this problem, it is first necessary to present some basic ideas on coastal classification to see what estuaries are and how they fit in (see Boyd et al., 1992, Perillo, 1995, and Bird, 2000, for a more detailed treatment of this material).

Firstly, we divide coasts into either transgressive or regressive categories (Figs. 7, 8). Secondly, we divide coasts into those that are significantly influenced by rivers and those that are not. On regressive coasts, the interaction between river sediment input and the ability of marine processes to redistribute that input determines if the coast will be an elongate or lobate protuberance (i.e., deltaic) or linear (i.e., strandplain or shoreface or tidal flat; Boyd et al., 1992). When the rate of relative sea-level rise exceeds the rate of sediment supply (area above the diagonal line in Figure 7), transgression with deposition (blue color in Figure 7) results in the generation of estuaries and lagoons on embayed coasts and the landward migration of the shoreline and continental shelf on all linear (tidal-flat and headland) coasts. Coastal cliffs fall into this latter category and form where the terrestrial gradient is relatively steep and there is net erosion. It is implicit in this arrangement that estuaries and lagoons form in areas of low terrestrial gradient, and only during regional or local trangression. They should not form or persist through a shoreline regression, and they should occupy only an ephemeral position at sea-level highstand until infilled (a critical point to appreciate for management of present-day highstand shorelines). However, estuaries are commonly reestablished in the same location during subsequent sea-level cycles, leading to multiple cut-and-fill events in the sedimentary record. Confirmation of the formation of estuaries during transgressions and their disappearance during regressions is provided by the history of the 3 m sea-level oscillation of the Caspian Sea over 65 years (Kroonenberg et al., 2000).

Another way of describing the influence of the major coastal processes is to employ a ternary diagram identifying their relative power (Fig. 9). Here the three main process agents are considered to be river currents, waves, and tidal currents. When the ternary diagram is constructed such that the vertical axis for

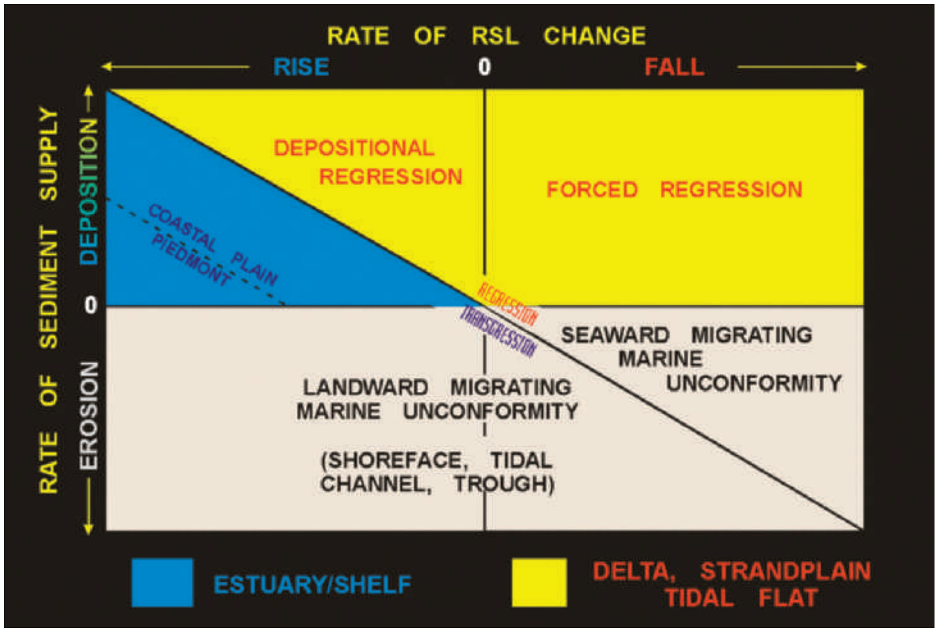


FIG. 7.—Shoreline response (transgression versus regression) to change in sea level and sediment supply (modified from Boyd et al., 1992).

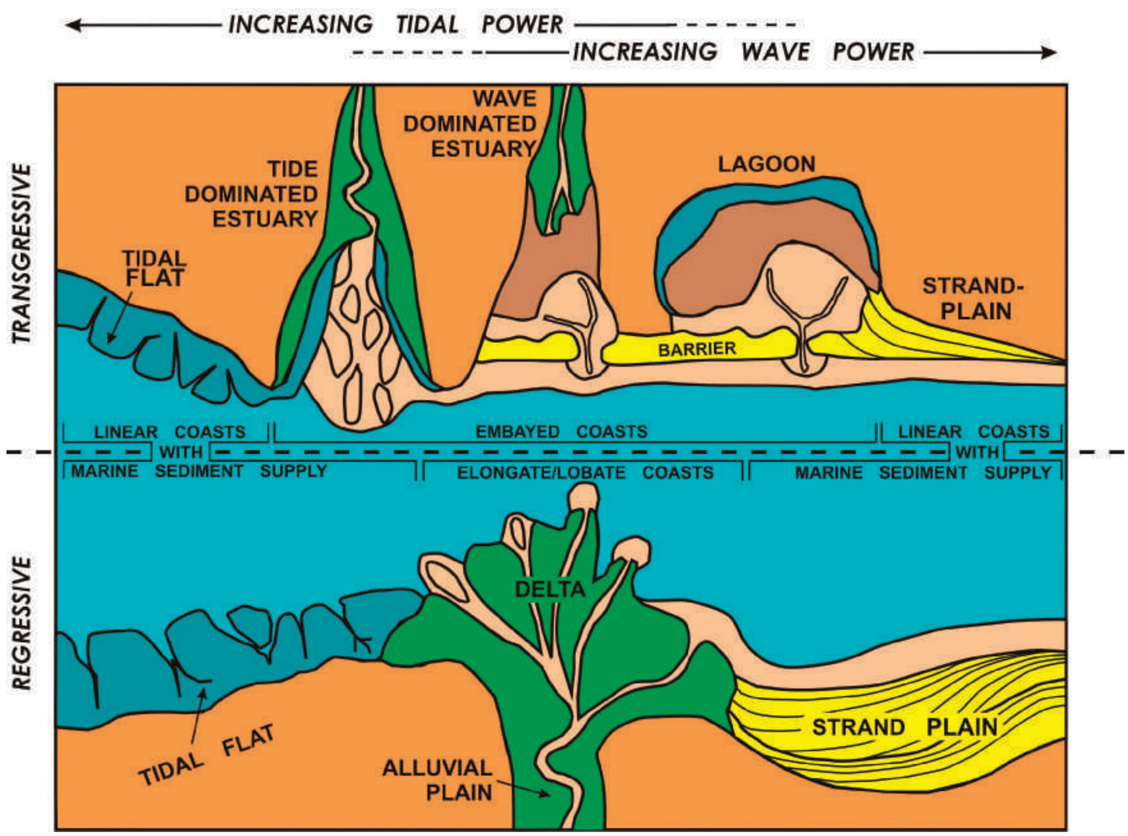


FIG. 8.—Classification from Boyd et al. (1992), illustrating organization of all of the major clastic coastal depositional environments based on shoreline translation direction (i.e., progradation or transgression) and relative power of waves, tidal currents, and river currents. The upper coastline is transgressive, and the lower coastline is regressive. The influence of tides relative to wave power increases from right to left.

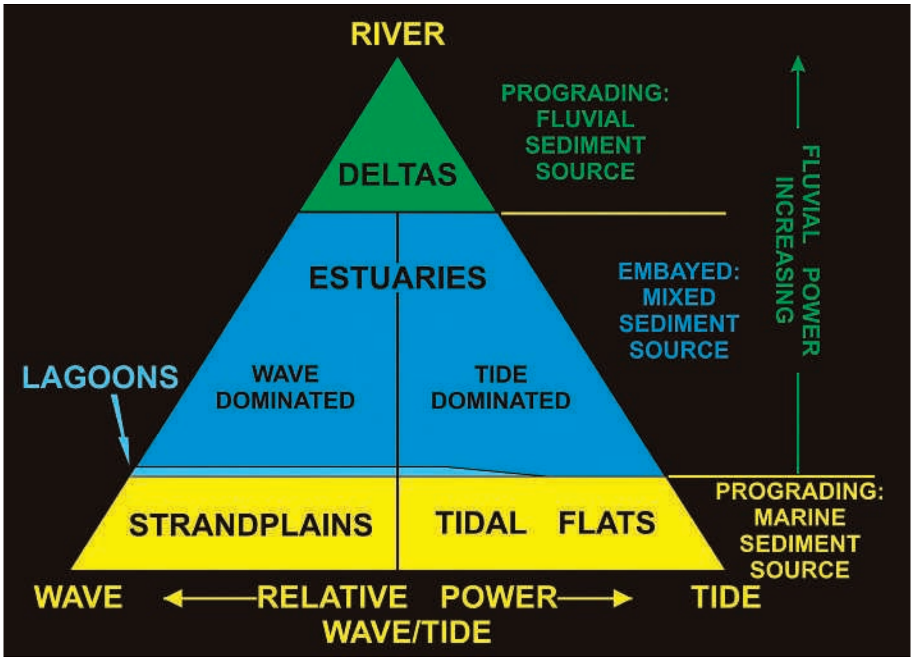


FIG. 9.—Triangular coastal classification using the three parameters of river, wave, and tidal processes, together with direction of sediment supply.

fluvial power is combined with a factor that discriminates prograding coasts from embayed transgressive coasts, and a second factor that discriminates direct sediment supply from a river from sediment that is supplied to the coast by marine processes (termed a marine sediment supply in Figures 8 and 9), then a clear definition of the major coastal sedimentary environments can be achieved. Estuaries occupy the center of the ternary diagram, where the coast is embayed and receives sediment from both marine and fluvial sources.

Estuaries can be distinguished by their mixed sediment source and association with a river input, whereas lagoons have no strong river-valley association and have only a marine sediment source (Figures 10, 11; Boyd et al., 1992;). In this scheme, estuaries and lagoons are intergradational, with lagoons representing the end-member situation where the river influence is negligible. By contrast, prograding deltas (the top triangle of Figure 9) derive sediment directly and only from a fluvial source, whereas prograding linear coasts (strandplains and tidal flats as shown at the base of Figure 9) are supplied only by marine processes (waves and/or tides), although that sediment must ultimately be derived mostly from a river source. It should be noted that virtually all coastal embayments have some form of fresh-water drainage into them, making the recognition of the gradational boundary between estuaries and lagoons difficult. It is suggested here that the term lagoon be used when there is no significant bedload supplied to the system by fluvial processes, as shown, for example, by the absence of a bayhead delta.

ESTUARINE FACIES MODEL

Once we have identified the dominant coastal processes and the relationship of relative sea level to sediment flux, we can develop a practical definition of an estuary. Perillo (1995) provides an extensive discussion of estuarine definitions and classifications, identifying a range of oceanographic, biologic, and geomorphologic–geologic approaches. For facies-models usage, a geological definition is most useful because it can be applied to ancient estuarine successions as well as modern estuaries. An estuary in geological terms receives sediment from both fluvial and marine sources, commonly occupies the seaward portion of a drowned valley, contains facies influenced by tide, wave, and fluvial processes, and is considered to extend from the landward limit of tidal facies at its head to the seaward limit of coastal facies at its mouth (cf. Dalrymple et al., 1992). This definition overcomes the limitations of the widely used oceanographic definition of Pritchard (1967) based on salinity, because Pritchard's definition applies to both regressive and transgressive settings in addition to being difficult to use in ancient estuarine deposits. Estuaries as defined here are present at the mouths of valleys that are being transgressed, and Dalrymple et al. (1992) restricted the use of "estuary" to such settings. However, we now recognize that transgressive embayments that do not contain a river-carved valley (e.g., the "abandoned" portion of a delta) may also contain environments that fulfill the criteria for an estuary provided above. Therefore, we extend the term "estuary" to such transgressive settings.

Most estuaries contain brackish water, but brackish water can occur in other settings (e.g., progradational deltas and even some shelves); hence, the identification of a trace-fossil assemblage indicating reduced salinity in an ancient succession does not necessarily mean that the deposits are estuarine (*sensu* Dalrymple et al., 1992). Salt-water intrusion up rivers is never as extensive as tidal action, so an estuary as defined above extends farther inland than if a salinity-based definition is used (e.g., Buatois et al., 1997): the tidal limit on many modern rivers lies tens of kilometers (in microtidal and steep-gradient settings) to hundreds of kilometers (in low-gradient, macrotidal settings) landward of the coast.

Because of the profound influence that waves and tides have on their basic morphology, estuaries can be divided into two

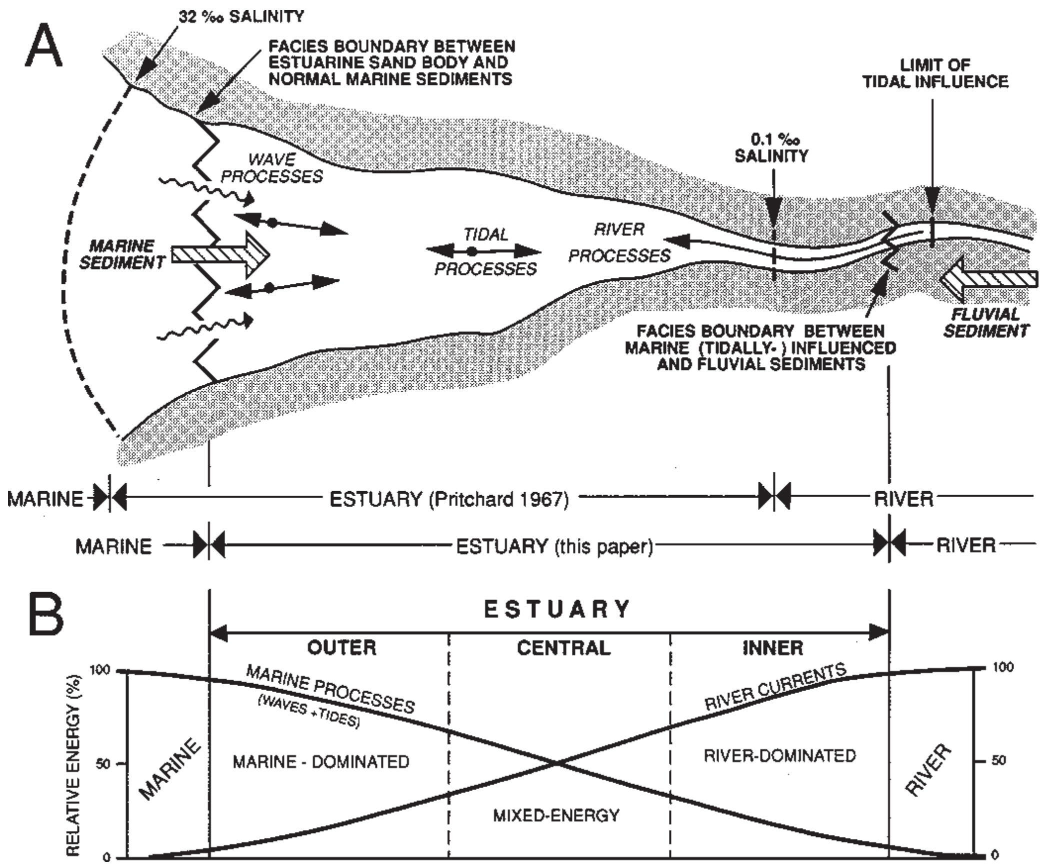


FIG. 10.—**A)** Schematic representation of the definition of an estuary according to Pritchard (1967) and Dalrymple et al. (1992). **B)** Schematic distribution of the physical processes operating within estuaries, and the resulting tripartite facies zonation.

main types, wave-dominated estuaries and tide-dominated estuaries, based on the relative power of waves and tidal processes (Figs. 8 , 9). This distinction determines the range of the resulting facies model. Fluvial processes primarily control the upstream sediment flux during estuary evolution and do not alter the fundamental morphology of the system. This point will be discussed further in the section on criticisms, misuses, and refinements of the E&IV model.

We believe that the interaction between river and marine processes provides the basis for a generalized estuarine facies model. Fluvial energy, as given by the energy flux per unit cross-sectional area or other suitable measure, typically decreases down an estuary (Fig. 10), because the hydraulic gradient decreases and the valley and its associated marine water bodies widen as the river approaches the sea. Marine energy, by contrast, generally decreases headward, because oceanic wave energy is dissipated by a wave-built barrier or tidal sand-bar complex and/or because tidal-current speeds eventually decrease up the estuary as a result of friction. Ideally, therefore, both wave- and tide-dominated estuaries can be divided into three zones (Fig. 10): (1) an outer zone dominated by marine processes (waves and/or tidal currents); (2) a relatively low-energy central zone, where marine energy (generally tidal currents) and river currents are approximately equal in strength **in the long term** (i.e., averaged over many years); and (3) an inner, river-dominated zone. (*Note that this estuarine* ***zonation*** *must be distinguished from the three-part* ***segmentation*** *of valley fills to be discussed below, because the two schemes have no relationship to each other*).

The tripartite estuarine zonation (Figs. 5, 10, 11) also corresponds with the general patterns of net bedload transport. Long-term (averaged over several years) transport of bedload is seaward in the river-dominated zone, whereas coarse sediment moves up estuary in the marine-dominated zone as a result of waves and/or flood-tidal currents (Guilcher, 1967; Kulm and Byrne, 1967; Roy et al., 1980; Dalrymple and Zaitlin, 1989). Thus, the central zone is an area of net bedload convergence and typically contains the finest-grained bedload sediment in the estuary, regardless of whether the estuary is wave- or tide-dominated. Once the process-based tripartite division of wave- and tide-dominated estuaries has been established, we can then examine each of these estuary types to see the major depositional elements developed and the facies successions they produce.

Elements of a Wave-Dominated Estuary

The profile of "total energy" (i.e., the sum of energy from all sources) for an ideal wave-dominated estuary shows two maxima, one at the mouth caused by wave energy and one at the head produced by river currents, separated by a pronounced energy minimum (Fig. 11). This distribution of total energy produces a clearly defined, "tripartite" distribution of lithofacies (coarse–fine–coarse) within most wave-dominated estuaries (e.g., Figs. 5, 11, 12; Kulm and Byrne, 1967; Roy et al., 1980; Zaitlin and Shultz, 1984, 1990; Rahmani, 1988; Nichol, 1991; Nichols et al., 1991). As the estuary fills, the central energy minimum becomes less pronounced.

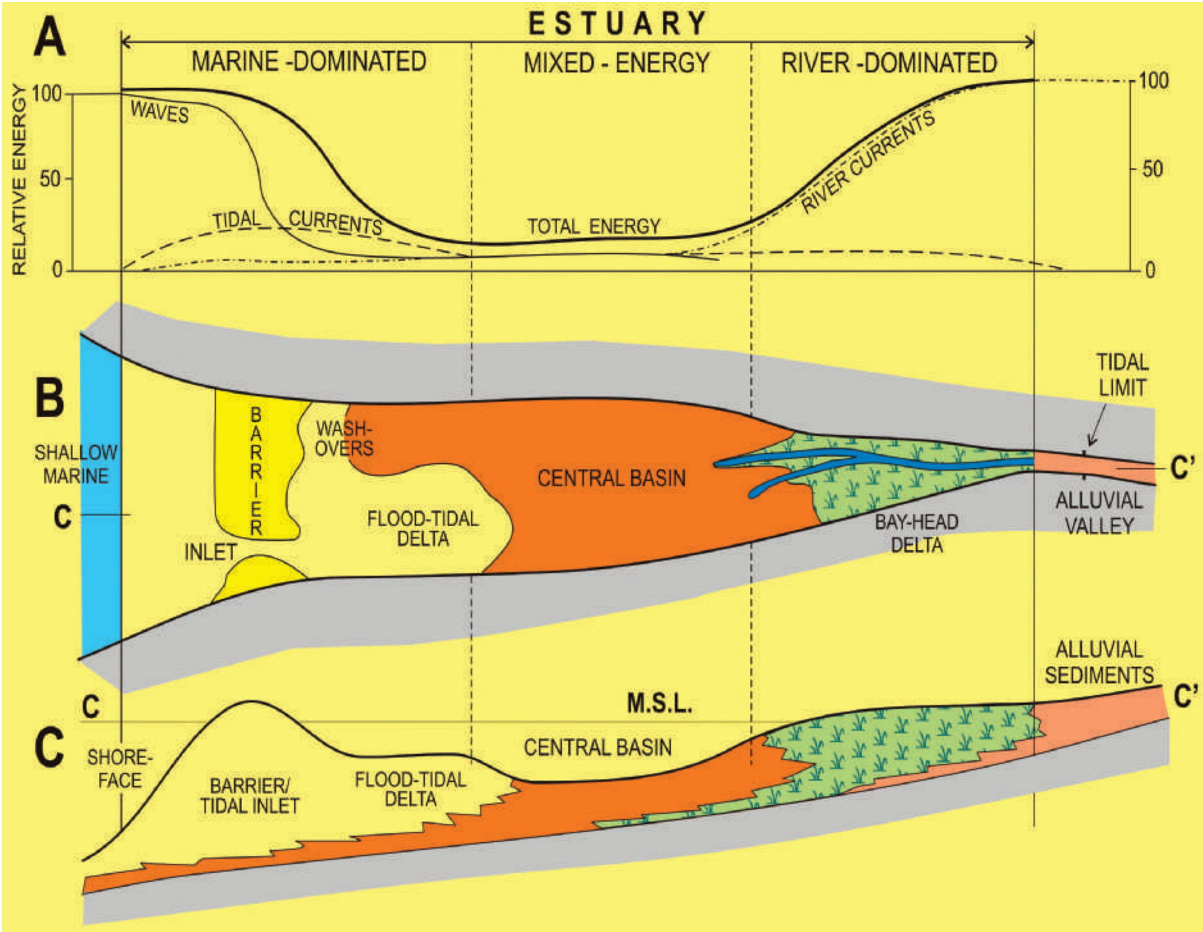


FIG. 11.—Distribution of **A)** energy types, **B)** morphological components in plan view, and **C)** sedimentary facies in longitudinal section within an idealized wave-dominated estuary. MSL = mean sea level (from Dalrymple et al., 1992). Note that for simplicity the complete transgressive succession that would be formed by landward migration of the estuary is not shown.

A marine sand body accumulates in the area of high wave energy at the mouth (Figs. 11, 12). It consists of a barrier, cut by one or more tidal inlets that terminate in ebb and flood tidal deltas. A shoreface, which typically experiences net erosion, lies seaward of the barrier. The limit of this shoreface or the distal ebb-tidal delta is the marine limit of the estuary *(sensu lato)*, and typically occurs in water depths less than 20 m. A subsurface example of such a barrier deposit located at the mouth of a wave-dominated estuary is provided by the Lower Cretaceous Lloydmister Formation Senlac Pool (Zaitlin and Shultz, 1984, 1990), which is described below in the Incised Valley Segment 2 portion of this review. Sand and/or gravel are also deposited at the head of the estuary by the river, forming a bayhead delta. This bayhead accumulation has a typical deltaic character with subaerial delta plain and a subaqueous mouth bar, prodelta, and delta front. The morphology is typically river-dominated because of the low-energy nature of the central basin, but wave- and tide-dominant varieties can occur if the local processes allow. A subsurface example of such a bayhead-delta deposit is the Lower Cretaceous Glauconitic Formation Lake Newell Pool of Broger et al. (1997), described in the Incised Valley Segment 1 portion of this review. The low-energy central part of the estuary (the "central basin") acts as the prodelta region of both the bay-head delta and the flood-tidal delta, and fine-grained, organic-rich and normally bioturbated muds accumulate there (Biggs, 1967; Donaldson et al., 1970). The margins of wave-dominated estuaries typically contain salt marshes, and/or mangroves cut by tidal channels, and sandy or muddy tidal flats. A comparison of central-basin deposits between the Glauconitic and Viking Formations was presented by Leroux et al. (2001) and MacEachern (1999). Beaches may occur along the margins of large central basins with fetch sufficient for the local generation of waves.

Elements of a Tide-Dominated Estuary

Most tide-dominated estuaries are macrotidal, but tidal dominance can also occur at much smaller tidal ranges if wave action is limited and/or the tidal prism is large. Tidal-current energy exceeds wave energy at the mouths of tide-dominated estuaries, and elongate sand bars are typically developed there (Figures 13, 14; Hayes, 1975; Dalrymple et al., 1990). These bars dissipate the wave energy that does exist, causing it to decrease with distance up the estuary. On the other hand, the incoming flood tide is progressively compressed into a smaller cross-sectional area because of the funnel-shaped geometry that characterizes

A

B

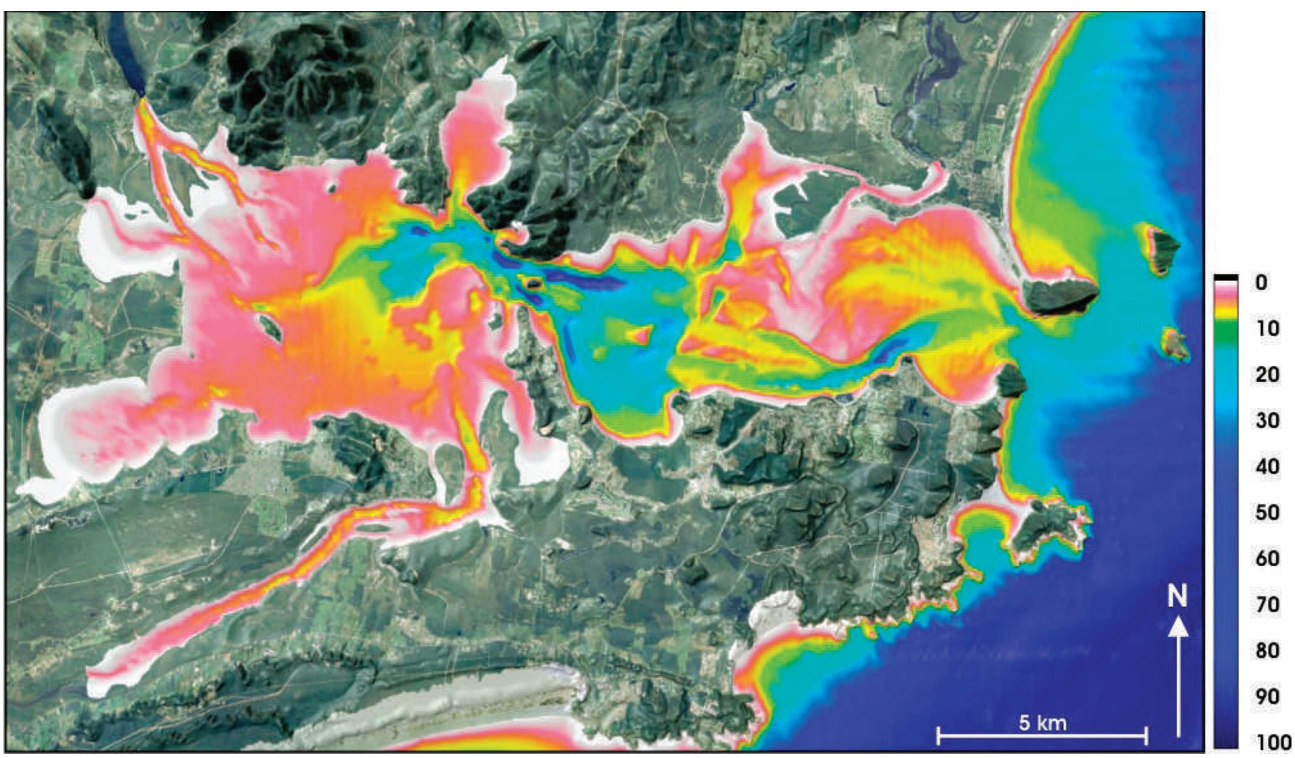


FIG. 12.—Examples of wave-dominated estuaries: **A)** Tuggerah Lake, NSW, on the southeast coast of Australia. Wyong Creek (right) and Ourimbah Creek (center) are building prograding bay-head deltas into the muddy central basin of Tuggerah "Lake", while The Entrance tidal inlet (foreground) is building a marine sand body landward into the estuary. **B)** Port Stephens, NSW, showing a merged landscape and seascape DEM illustrating the division of wave-dominated estuaries into an outer flood tidal delta and barriers (right), a deeper central basin (middle), and an inner river valley and bay-head delta (Karuah River upper left). Depth color bar at right is in meters below sea level.

these estuaries (Langbein, *in* Myrick and Leopold, 1963; Wright et al., 1973). Flood tidal currents increase in speed landward until frictional dissipation exceeds the effects of amplification produced by convergence, causing the tidal energy to decrease, eventually reaching zero at the tidal limit. Fluvial energy decreases seaward as in wave-dominated systems. The location where flood-tidal and fluvial energy are equal lies landward of the tidal-energy maximum (i.e., the location where the tidal current speeds are greatest; Fig. 13A). As in wave-dominated systems, this bedload convergence is the location of a minimum in the total-energy curve, but this minimum is not as pronounced as it is in most wave-dominated estuaries because the flow is channelized along the entire length of the estuary. Subsurface examples of such tide-dominated estuarine systems have been proposed from the Lower Cretaceous McMurray Formation (e.g., Flach and Mossop, 1985; Ranger and Pemberton, 1988) and in outcrop from the Proterozoic of Utah (Ehlers and Chan, 1999) and the Eocene of Spitsbergen (Plink-Björklund, 2005).

In high-tidal-range end-member cases such as the Severn and Cobequid Bay–Salmon River estuaries, the marine sand body consists of two strongly contrasting facies. The best known is the elongate tidal sand-bar zone (Harris, 1988; Dalrymple and Zaitlin, 1989; Dalrymple et al., 1991), which is characterized by cross-bedded medium to coarse sand (Fig. 14). These bars lie seaward of the tidal-energy maximum. The second facies, which coincides with the tidal-energy maximum, consists of upper-flow-regime (UFR) sand flats which display a braided channel pattern where the estuary is broad and shallow (Fig. 15A), but these become confined to a single channel farther headward as the estuarine funnel narrows (Figs. 13, 15B; Hamilton, 1979; Lambiase, 1980; Dalrymple et al., 1990). This facies, which may not be present in tide-dominated estuaries that are deeper and/or have smaller tidal ranges, consists of parallel-laminated fine sand.

The tripartite facies distribution is not as obvious in tide-dominated estuaries because the energy minimum is not as pronounced within these channelized systems, and sands occur in the tidal–fluvial channels that run along the length of the estuary (Woodroffe et al., 1989; Dalrymple et al., 1990). Nevertheless, the energy minimum is the site of the finest channel sands. In the central, low-energy zone of systems in which the main channel is unconfined by older material, this channel consistently displays a regular progression of sinuosities (e.g., Ashley and Renwick, 1983; Dalrymple and Zaitlin, 1989; Woodroffe et al., 1989) that is termed "straight–meandering–straight" (Figs. 13, 15). The outer straight reach in these estuaries is tidally dominated and the net sediment transport and barform asymmetry are headward due to strong flood-tidal currents (e.g., Dalrymple et al., 1990). The channel contains alternate, bank-attached bars (Fig. 15B) and some mid-channel bars. The inner straight reach contains similar bar types, but here the net sediment transport and barform asymmetry are downstream due to the long-term dominance of river flow over tidal currents.

The region between the two straight reaches contains tight meanders (Figs. 13, 15B) that commonly exhibit symmetrical point bars (Dalrymple and Zaitlin, 1989). A subsurface example of such a symmetrical tidal point-bar deposit, also from the Lower Cretaceous Glauconitic Formation, is provided by the Lathom "A" Pool described by Zaitlin et al. (1998). This meandering zone is the lowest-energy portion of the system and is the position of net bedload convergence. Grain sizes in the channel become finer toward this area from both directions (Dalrymple and Zaitlin, 1989). Muddy sediments accumulate primarily in tidal flats, marshes, and flood plains along the sides of the estuary. Subtle

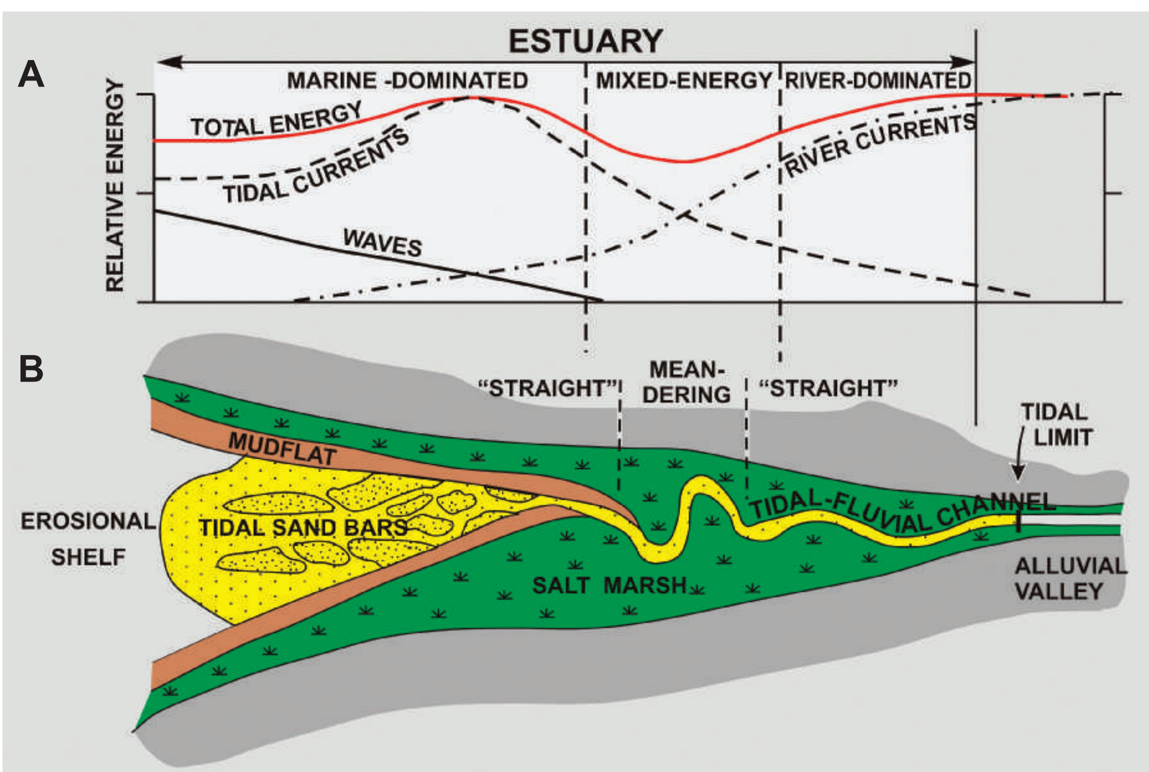


FIG. 13.—Distribution of **A)** energy types and **B)** morphological elements in plan view within an idealized tide-dominated estuary.

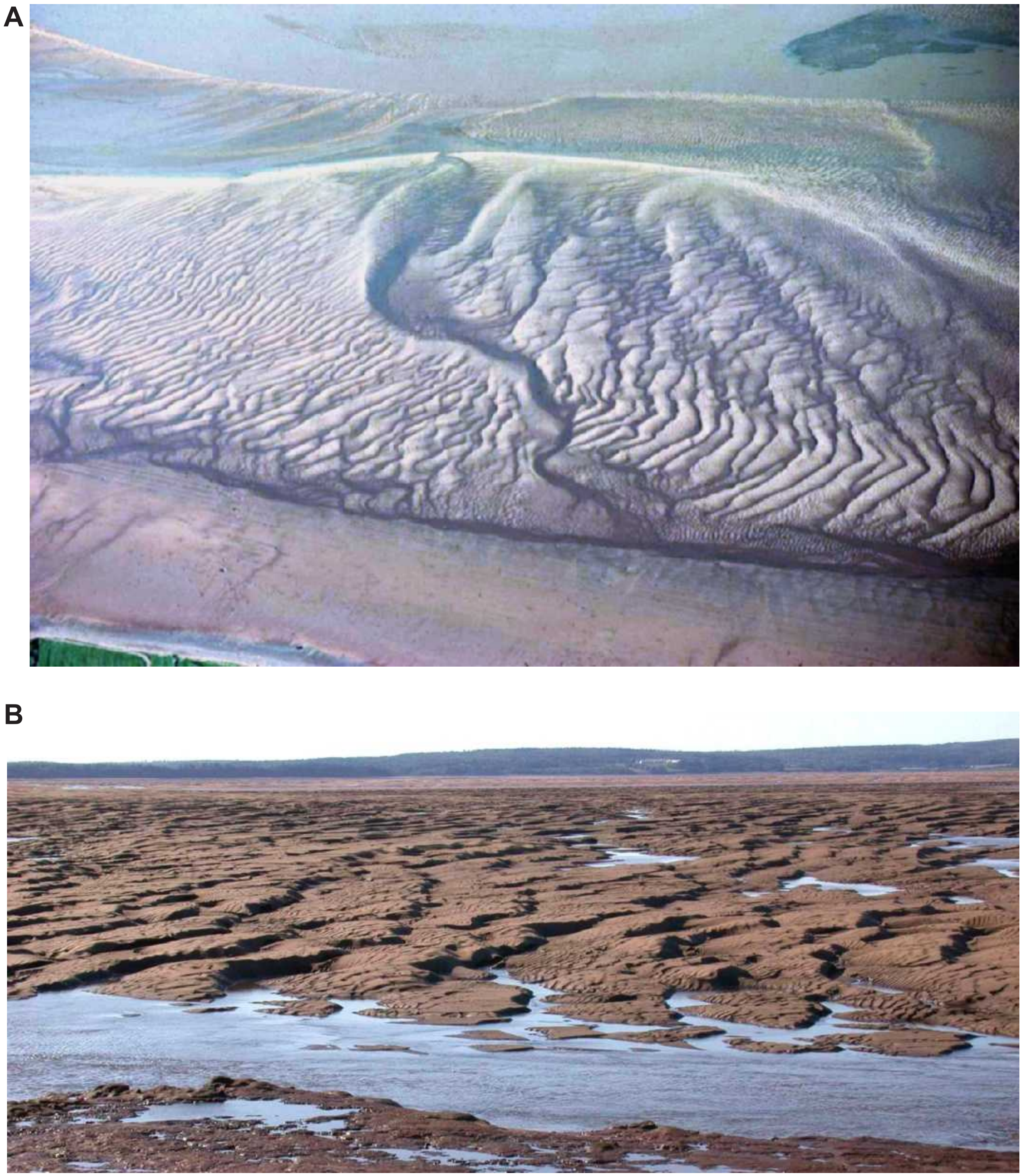


FIG. 14.—**A)** Overview of elongate sand bars developed in the outer (marine dominated) part of the Cobequid Bay–Salmon River Estuary, Bay of Fundy, Canada. **B)** Close up of one elongate sand bar from Part A showing the scale of the bar (approximately 500 m across) and the superimposed dunes on the bar at several different length scales (Both photos by R. Dalrymple).

A

B

FIG. 15.—**A)** The inner part of the upper-flow-regime sand flats of Zone 2, where marine energy is at a maximum (see Fig. 13). **B)** the straight–meandering–straight transition in the mixed energy, upper part (Zone 3) of the Cobequid Bay–Salmon River Estuary, Canada. This photo is taken from approximately the same position as Figure 15A but is looking in the opposite direction. The straight channel with bank-attached bars is in the foreground, the meandering channel is in the middle distance above the bridge, and the upper straight channel is in the upper center near the town of Truro. Both photos courtesy of John Suter.

levees flank the channel, but crevasse-splay deposits become progressively less abundant in a seaward direction through the tidal–fluvial reach because the intensity of river floods is damped by tidal action. A discrete bayhead delta is **not** present in the river-dominated portion of tide-dominated estuaries because there is no open-water body into which the sediment can be dumped. Instead the tidal–fluvial channel passes directly into the river above the tidal limit.

Organization of Estuary Elements into a Facies Model

The allocyclic components of estuarine sedimentation are fixed, in that relative sea-level rise over the long term exceeds the sediment input from both marine and fluvial sources, resulting in transgression, a necessary condition for the formation of estuaries (as defined geologically; Dalrymple et al., 1992). Estuaries are typically initiated with the beginning of the transgression and continue accumulating sediment throughout the transgression, up to the time of maximum flooding, when the shoreline reaches its most landward position, before finally filling at the beginning of the subsequent highstand. If the highstand is of short duration, sea level may fall before the estuary is completely filled; however, if the highstand is long and / or the rate of sediment input is high, then the estuary fills completely in the transition to highstand progradation. As a result, an assemblage of estuarine facies, termed here an **estuarine lithosome**, stretches along a substantial portion of the valley or the length of the embayment, from near the lowstand mouth of the river to the landward extent of marine influence at the time of maximum transgression (Figs. 16, 17). In this lithosome, facies are stacked retrogradationally such that the most landward terrestrial facies is overlain by central estuarine facies and lastly by the most marine facies.

The contact between the fluvial and overlying estuarine sediments is termed the initial flooding surface (FS; Figs. 16, 17), or, alternatively, the transgressive surface. As the estuary continues to translate landward, the upper portion of the transgressive succession is generally removed by shoreface and / or tidal-channel erosion (generating wave and tidal ravinement surfaces, respectively), depending on whether wave or tidal processes dominate. The amount of section removed varies between examples, depending on the relationship between the rates of sea-level rise and transgression, the rate of sediment input, the depth of the shoreface and tidal-channel thalweg, and the depth of the paleovalley (cf. Davis and Clifton, 1987; Demarest and Kraft, 1987). Partial transgressive successions, in which the basal fluvial and fluvial–estuarine facies have the highest preservation potential, should occur along the transgressed portion of a paleovalley, seaward of the highstand shoreline (Figs. 16, 17). Fluvial deposits should occupy the deepest portions of the valley, except near the lowstand river mouth, where tidal–fluvial sediments may occur. Along the flanks of the valley, estuarine deposits lie directly on older deposits and the sequence boundary, without intervening fluvial sediments. In settings where estuaries occupy embayments that are not paleoriver valleys, the estuarine deposits overlie either earlier deposits such as deltas and are separated by a

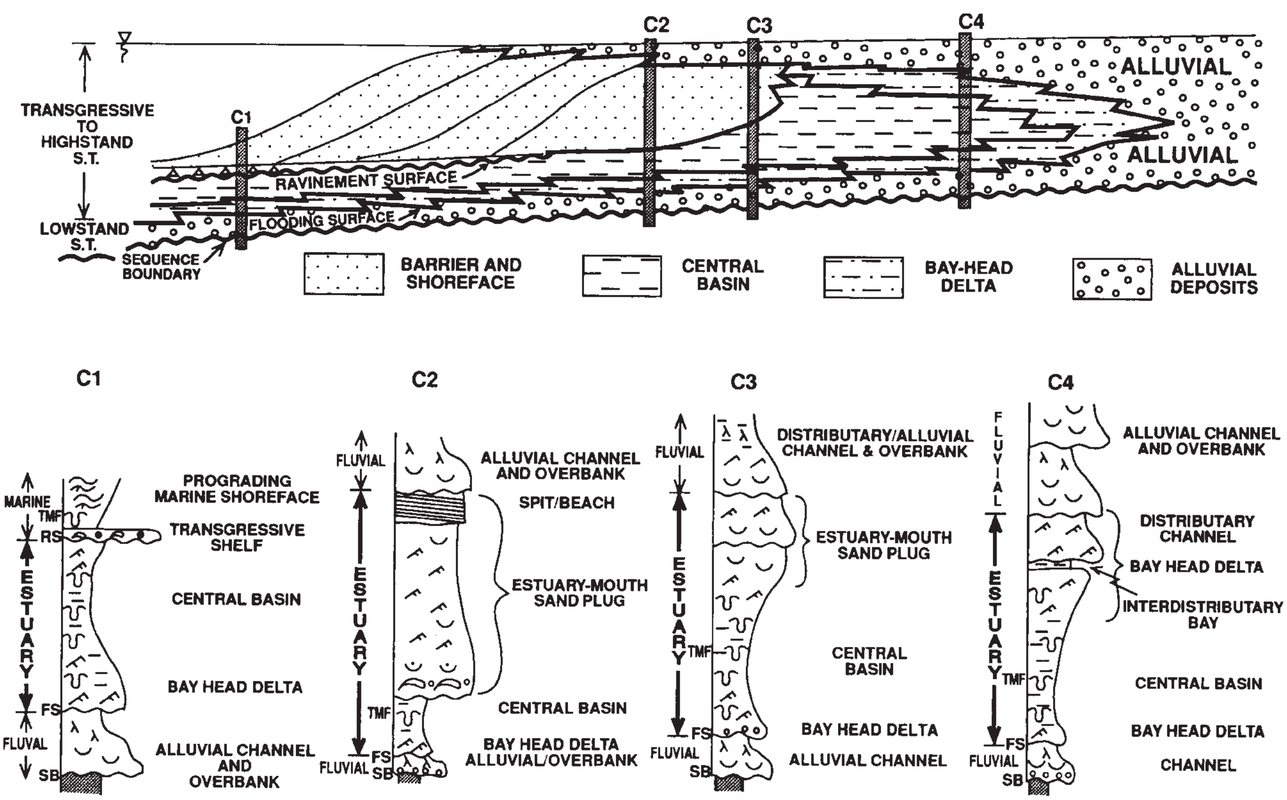


FIG. 16.—Schematic section along the axis of a wave-dominated estuary, showing the distribution of lithofacies resulting from transgression of the estuary, followed by estuary infilling and shoreface progradation at the time of sea-level highstand. The completeness and thickness of the preserved transgressive succession depends on the relative rates of sea-level rise and the headward translation of the shoreface. See Figure 17 for legend (from Zaitlin et al., 1994). "Flooding surface (FS)" refers to the initial flooding surface at the beginning of transgression.

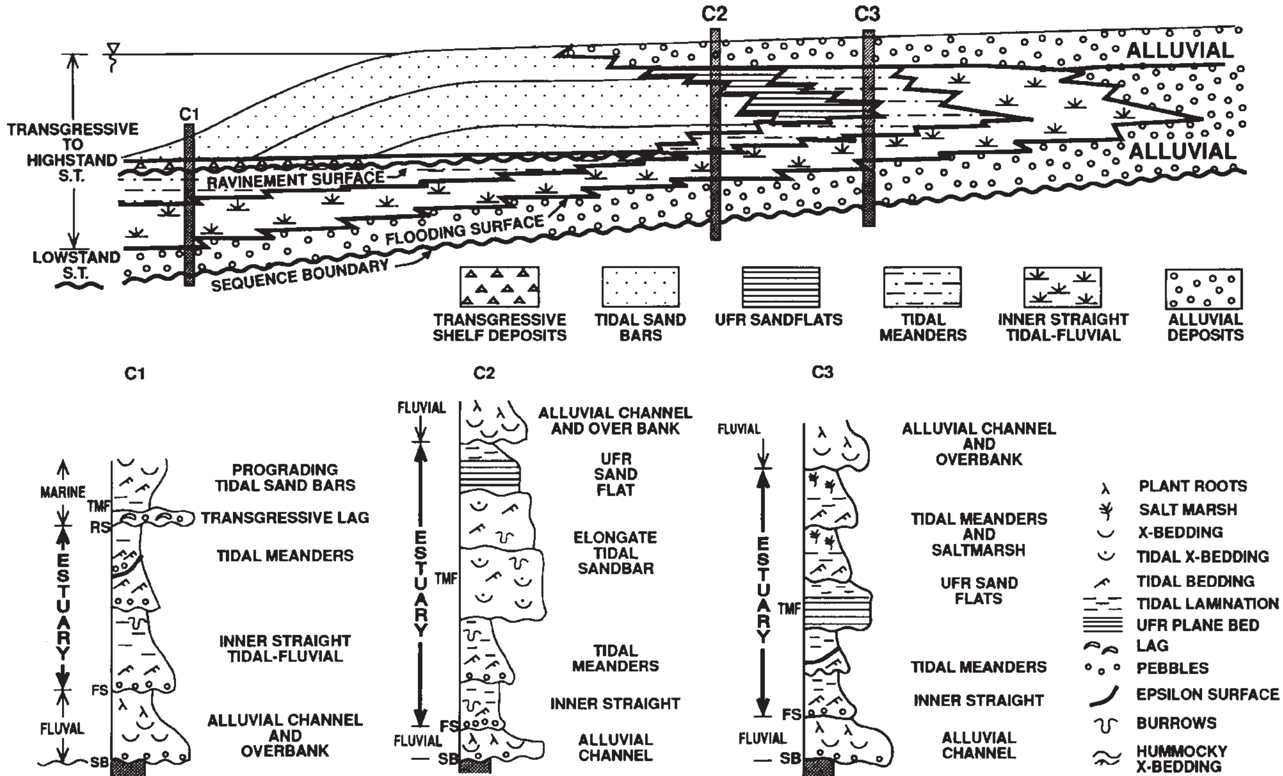


FIG. 17.—Schematic section along the axis of a tide-dominated estuary, showing the distribution of lithofacies resulting from transgression of the estuary, followed by estuary infilling and progradation of sand bars or tidal flats. The completeness and thickness of the preserved transgressive succession depends on the relative rates of sea-level rise and the headward translation of the thalweg of tidal channels (from Zaitlin et al., 1994). "Flooding surface (FS)" refers to the initial flooding surface at the beginning of transgression.

flooding surface, or older unrelated units located below an unconformity.

Wave-Dominated Estuarine Model

The marine sand body in these estuaries is a composite feature that may contain several discrete facies. In transgressive successions, some or all of the barrier complex is likely to be eroded during shoreface retreat and overlain by a wave ravinement surface (Fig. 16, C1). If any part of the barrier remains, it consists of the deeper and/or more landward facies including erosionally based tidal-inlet deposits and the landward-directed cross bedding of washovers and flood-tidal deltas that may interfinger with the underlying central-basin muds (e.g., Roy et al., 1980; Roy, 1984; Zaitlin and Schulz, 1984, 1990; Boyd and Honig, 1992). In vertical profile, fine-grained central-basin sediments ideally exhibit a symmetrical grain-size trend (Fig. 16, C4). The basal upward-fining portion represents the passage from transgressive, fluvial, and bayhead-delta deposits through progressively more distal prodelta sediments. More commonly, the base of the central-basin muddy facies is an abrupt flooding surface that might display some evidence of erosion (i.e., a "bay ravinement surface") that occurred as the low-energy central-basin shoreline transgressed. The finest sediments represent the center of the central basin and are frequently the mostly intensely bioturbated (although often with an impoverished, brackish-water trace-fossil assemblage). Organic facies, including peat, coal, and oyster buildups, may also be present at this stratigraphic level. The finest sediments are overlain in turn by an upward-coarsening succession passing into either flood-tidal delta and washover sediments (Fig. 16, C1, C2, C3) along most of the length of the estuarine lithosome, or into bayhead-delta deposits (Fig. 16, C3) at locations where there is episodic bayhead-delta progradation. Tidal-channel migration during transgression generates a tidal ravinement surface landward and ahead of the wave ravinement surface, providing at least two possible erosion surfaces within the wave-dominated estuarine succession.

The bayhead delta deposits are distinguished from true fluvial sediments by the presence of tidal structures and/or a brackish-water fauna as well as a deltaic geometry and stratigraphy. Bayhead-delta sediments are likely to be common in the lower part of transgressive valley-fill successions, and will occur at the up-dip end of the estuarine lithosome where they will exhibit an upward-coarsening succession resulting from progradation either during stillstands or during estuary filling at highstand (Fig. 16, C4; Rahmani, 1988; Reinson et al., 1988; Allen, 1991; Allen and Posamentier, 1993; Broger et al., 1997).

Meandering tidal channels containing inclined heterolithic strata (Flach and Mossop, 1985; Thomas et al., 1987; Pemberton and Wightman, 1992) are likely to be most abundant in the late stage of estuary filling, when the prograding bayhead delta merges with the flood-tidal delta (Smith, 1987; Nichol, 1991). Such channels may erode some or all of the underlying central-basin succession and might scour down to the basal unconformity. An additional stratigraphic surface, termed the bayhead-delta diastem, may be generated by erosion at the base of laterally migrating bayhead-delta distributaries (e.g., Nichol, 1991). Ancient wave-dominated estuarine systems such as the Lower Cretaceous Lloydminster Member and the Albian Paddy Member (Leckie and Singh, 1990; Leckie et al., 1990) will be discussed in the later section dealing with incised valleys, segment 2.

Tide-Dominated Estuary Model

During transgression, the elongate tidal sand bars that constitute the outer part of the marine sand body in tide-dominated estuaries are likely to be erosionally truncated or completely removed (Fig. 17, C1) by the headward migration of the erosional zone that coincides with the "bedload parting" that lies seaward of the estuary mouth (Dalrymple, 1992; Dalrymple et al., 1992) and/or the headward and lateral migration of tidal channels that separate the sand bars. The amalgamation of these scours produces a tidal ravinement surface. Erosion by the channels during transgression also causes the cross-bedded sand bars, or the parallel-laminated, UFR sand-flat deposits, to overlie or abut erosionally against mudflat and salt-marsh sediments along the margins of the estuary and/or on more headward facies in the axis of the valley (Fig. 17, C2). If the transgressive succession contains both sandy facies (i.e., cross-bedded medium to coarse sand and parallel-laminated fine to very fine sand), they produce an overall upward-coarsening trend. The contact may be either erosional or gradual.

The central, mixed-energy (tidal–fluvial meanders) and inner, river-dominated portions of the estuary are characterized by tidal-channel deposits that are flanked by vertically accreted, salt-water, brackish-water, and fresh-water marsh sediments. If sufficient accommodation is generated, the point-bar sediments of the meandering zone are overlain and underlain by the deposits of straighter channels (Fig. 17) that display opposite paleocurrent directions; if there is low accommodation, the last channel to cross the area incises into the older tidal-channel deposits. Upper-flow-regime parallel lamination predominates in the shallower parts of the outer (tide-dominated) straight reach (Fig. 15A), while dunes may occur in the deeper channels. Ripples and/or dunes are likely to be more abundant in the meandering and inner straight reaches. The channel sediments are finest, and the mixing of fluvially and tidally supplied sediment is most pronounced, in the zone of tight meandering. The contacts between facies zones coincide with erosional channel bases. The channel-bank sediments consist of tidally bedded sands and muds that occur either as erosionally bounded wedges of flat-lying strata (Dalrymple et al., 1991) or as inclined heterolithic strata (IHS); see Flack and Mossop (1985). IHS is most prevalent in the meandering reach. A well-documented example of an ancient tide-dominated estuary is the Cretaceous Lower Greensand in the Leighton Buzzard area of England (Johnson and Levell, 1995).

INCISED-VALLEY FACIES MODEL

To develop an appropriate facies model for an entire incised valley, compared to an estuary, we need to address the wider concept of an incised-valley system (IVS). An incised-valley system (e.g., Fig. 18) must incorporate elements of the erosional valley itself, the strata that it eroded into, and the entire fill consisting of fluvial, estuarine, and marine facies (Fig. 19). In this context, an incised-valley system is defined as "a fluvially eroded, elongate topographic low that is typically larger than a single channel form, and is characterized by an abrupt seaward shift of depositional facies across a regionally mappable sequence boundary at its base. The fill typically begins to accumulate during the next base-level rise, and may contain deposits of the following highstand and subsequent sea-level cycles" (Zaitlin et al., 1994).

Types of Incised Valleys

There are two major physiographic types of incised valley. Incised-valley systems that have their headwaters in a (mountainous) hinterland, and that cross a "fall line" where there is a significant reduction in gradient, are here considered to be ***piedmont incised-valley systems***. There are many ancient examples from the North American Western Interior Seaway that can be interpreted as piedmont incised-valley systems including the Lower Cretaceous Cutbank, Taber, and Basal Quartz of northern Montana–Alberta (e.g., Hayes, 1986; Dolson and Piombino, 1994; Ardies et al., 2002; Lukie et al., 2002; Zaitlin et al., 2002), Glauconite Formation (Rosenthal, 1988; Sherwin, 1994) and Muddy Sandstone and its Canadian equivalents the Viking and Bow Island Formations (Gustason et al., 1986; Dolson et al., 1991; Pattison, 1991; Pattison and Walker, 1994, 1998; MacEachern and Pemberton, 1992, 1994). Incised-valley systems that are confined to low-gradient coastal plains and that do not cross a "fall line" are termed ***coastal-plain incised-valley systems***. Subsurface examples of coastal-plain estuaries include parts of the Cretaceous Viking Formation (e.g., Pattison, 1991; MacEachern and Pemberton, 1994) at Sundance, Edson, and CynPem, and the southern portions of the Paddy–Cadotte (e.g., Leckie and Singh, 1991).

Piedmont incised-valley systems are characterized by a longer fluvial reach than coastal-plain systems and are commonly associated spatially with underlying structural features in the hinterland, e.g., the Upper Cretaceous Dunvegan System (Plint, 2002, and the Mississippian Morrow System (Bowen and Wiemer, 1997, 2003). As a result, these river systems may be longer lived than coastal-plain systems. Also, piedmont systems more commonly contain coarse-grained, less-mature, fluvially supplied sediment, whereas coastal-plain systems are usually filled by finer-grained and more mature deposits recycled from coastal-plain sediments. Piedmont systems may have overall higher rates of sediment supply because they have larger catchment areas. In both piedmont and coastal-plain systems, marine-derived sediment is preserved in the estuarine portion of the valley fill (see below). Coastal-plain and piedmont incised-valley systems occur adjacent to each other in modern coastal areas (e.g., Hayes and Sexton, 1989).

Simple and Compound Incised-Valley Fills

The fill of any incised-valley system can be classed as either ***simple or compound*** depending on the absence or presence, respectively, of multiple, internal, high-frequency sequence boundaries. If the valley is filled completely during one cycle such that the depositional surface rises above the level of the original interfluves, the fill is termed a "simple fill". An ancient example of a simple fill has been described by Zaitlin and Schultz (1984, 1990; see more below). A "compound fill" records multiple cycles of incision and deposition resulting from fluctuations in base level and is therefore punctuated by one or more sequence boundaries in addition to the main, lower-order sequence boundary at the base of the incised valley (e.g., the Mississippian Morrow Formation; Krystinik and Blakeney-DeJarnett, 1994; Krystinik, 1989; Bowen and Weimer, 1997, 2002); and the Lower Cretaceous Basal Quartz Formation (Ardies et al., 2003; Zaitlin et al., 2002; Leckie et al., 2005), the Lower

FIG. 18.—Example of incised valley with incised tributaries, Red Deer River south of East Coulee, Alberta, Canada. An incised-valley system consists of the erosional form seen here, plus the sediments that will ultimately fill this container.

Cretaceous Glauconitic Formation (e.g., Wood and Hopkins, 1989, 1992; Broger et al., 1997), and the Viking/J/Muddy Formation (Gustason et al., 1986, Gustafson et al., 1988, Reinson et al., 1988). Due to the presence of structural control on their location, piedmont river systems may exist through more than one sequence of sea-level fall and rise; thus, their incised valleys may contain a compound fill, although higher rates of sediment supply may counteract this tendency (e.g., Gustason et al., 1988; Dolson et al., 1991; Ardies et al., 2002; Zaitlin et al., 2002). Coastal-plain systems are more likely to exist through only one regression–transgression cycle and therefore have a simple fill, unless the rate of sediment supply is too low to fill the valley during a single cycle.

Recognition of Estuarine and Incised-Valley Systems

E&IV deposits are among the hardest to recognize because of their low width:depth ratio, limited lateral extent and ribbon geometry, and the complex association of fluvial, tidal, wave, and marine facies within them (Figs. 19–21). The following is a list of criteria for recognizing E&IV systems:

(1) **The valley is a negative (i.e., erosional) paleotopographic feature, the base of which truncates underlying strata, including any regional markers** (such as bentonites, coals, flooding surfaces, or seismic markers) that may be present (Fig. 22, green arrow). The valley container has a characteristic size, shape, and regional extent. The valley should be larger than a single channel (e.g., Figs. 18, 23) and commonly has an erosional relief (from the valley base to the original floodplain level) of 10 m or more. However, there is a complete gradation from non-incised channels, through shallowly incised systems, to very deeply entrenched valleys (Fig. 23). Studies of both modern and ancient valleys show that the depth of incision is not constant along their length (Schumm and Ethridge, 1994). Deeper-than-average incision occurs at the location where tributaries join the trunk river (scour depths at these locations may be up to five times the depth of adjacent parts of the valley; Best and Ashworth, 1997), at the location of flow constrictions where the river cuts across a more resistant underlying unit, and at the outsides of bends. Ardies et al. (2002) show a well-documented ancient example of all three types of channel-bottom irregularity (Fig. 24). The valley width may also be quite variable; it increases with time (e.g., Schumm and Ethridge, 1994) and is wider where the river cuts into less resistant lithologies (e.g., Ardies et al., 2002). However, typical dimensions are in the range of several hundreds of meters to several tens of kilometers, with most valleys in the range of 1–10 km wide.

(2) **The base and walls of the incised-valley system represent a sequence boundary** (Fig. 22, red line) that correlates to an erosional (or hiatal) surface outside the valley (i.e., on the interfluve areas). This erosional surface may be modified by later transgression, forming an E/T (erosive–transgressive) surface; Plint et al., 1992), or a combined flooding surface and

sequence boundary (an FS/SB surface; Van Wagoner et al., 1990). The sequence boundary may be mantled by a pebble lag and/or characterized by burrows belonging to the *Glossifungites* ichnofacies (MacEachern et al., 1992; MacEachern and Pemberton, 1994). On the interfluves the exposure surface may be characterized by a particularly well-developed soil or rooted horizon (Leckie and Singh, 1991; McCarthy and Plint, 1998). Such paleosols may show evidence of lower groundwater tables and more prominently developed soil horizons than paleosols formed syndepositionally within the TST and/or the HST.

(3) **Because the river erodes below the level of the interfluves when it creates the valley, water drains downward into the valley; as a result, the trunk river may be fed by smaller incised tributary valleys that are themselves incised** (e.g., Figs. 18, 24; Posamentier, 2001; Ardies et al., 2002). These tributary valleys aid in distinguishing incised valleys from unincised channels and augment criterion 2 above. On the regional scale, the planform geometry of tributary networks should mimic the river system(s) that became entrenched. As a result, the various river patterns identified by geomorphologists (e.g., Howard, 1967) may be recognized in valley systems. For example, dendritic patterns predominate in areas with uniform slopes and substrate erodibility, whereas rectilinear patterns occur in jointed bedrock or in areas with a crosscutting network of subtle faults. Recent work suggests that faults that are active during incision may have a strong influence on the location and planform pattern of valleys (e.g., Ardies et al., 2002).

(4) A fundamental aspect of incised valleys is their formation at times of erosion and falling base level; in cases where the area lies close to the shoreline, coastal regression accompanies incision. Hence, the base of the incised-valley fill (Figs. 19–21) exhibits an erosional juxtaposition of more proximal

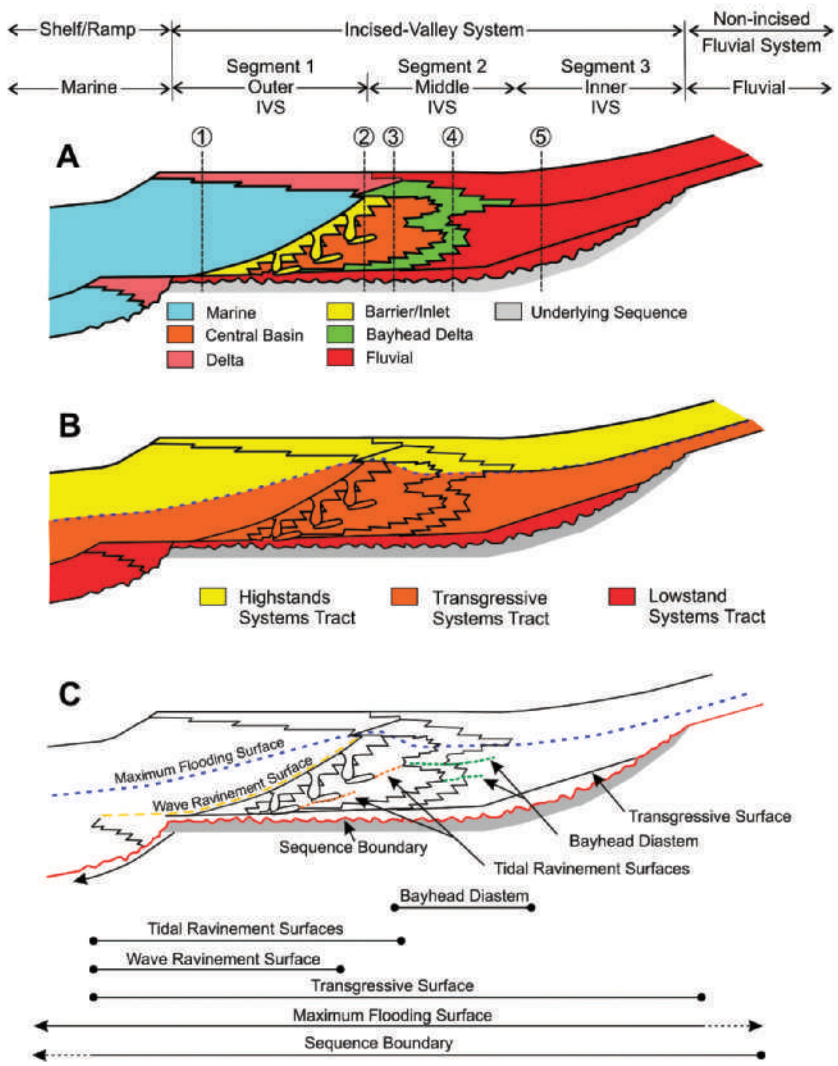


FIG. 19.—Idealized longitudinal section of a simple incised-valley system showing the distribution of **A)** depositional environments, **B)** systems tracts, and **C)** key stratigraphic surfaces. A wave-dominated estuary has been used in this model. Segments 1 and 3 are typically much longer than segment 2, and are compressed here for presentation purposes. Also shown are the locations of the schematic profiles illustrated in Figure 20. Modified from Zaitlin et al. (1994).

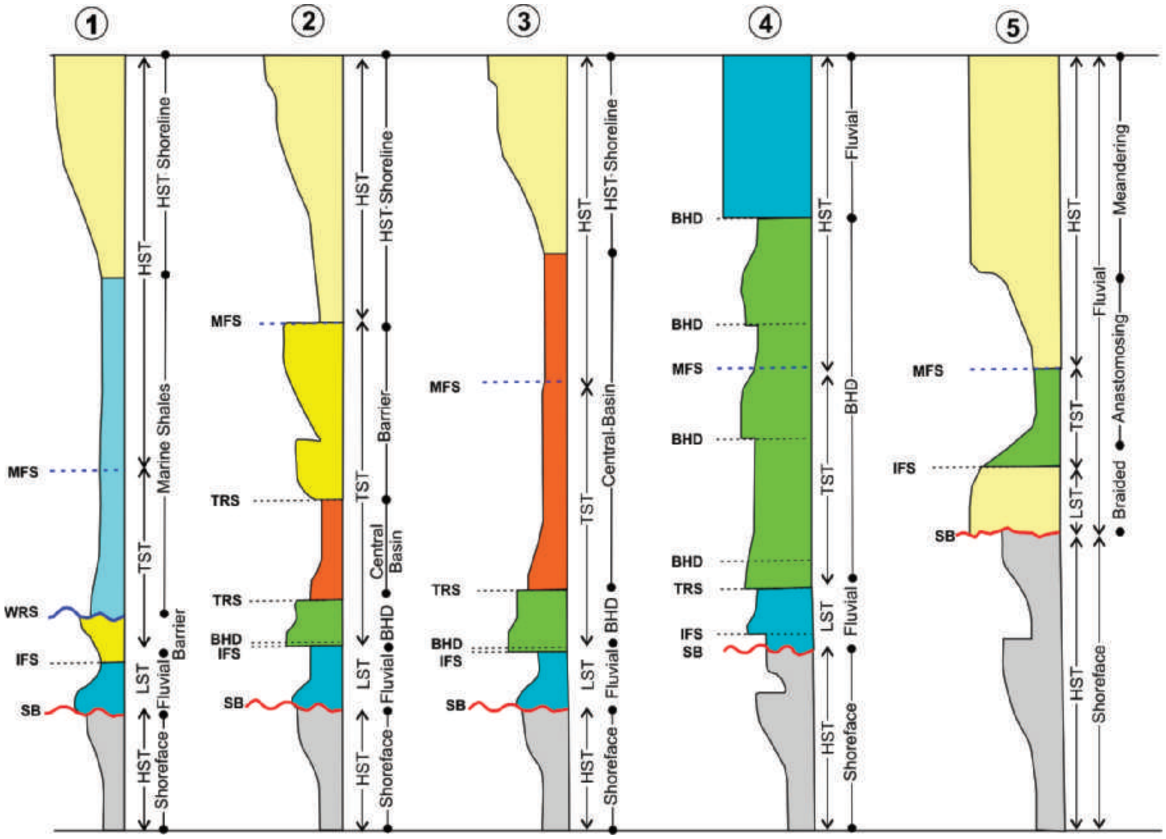


FIG. 20.—Five representative vertical sections of facies and sequence-stratigraphic surfaces in an idealized incised-valley system, based on an estuarine system that is wave dominated. WRS = wave ravinement surface, MFS = maximum flooding surface, IFS = initial flooding or transgressive surface, SB = sequence boundary, TRS = tidal ravinement surface, BHD = bay-head delta. Numbers in circles identify location of sections shown in Figure 19. Modified from Zaitlin et al. (1994).

(landward) facies over more distal (seaward) deposits (i.e., **a "basinward shift in facies" *sensu* Van Wagoner et al., 1990**), across a regional hiatus (unconformity)—vertical white arrow in Figure 22. The subsequent filling of the valley occurs partially or wholly during rising base level and is accompanied by transgression in near-coast situations. The latter typically results in more downdip facies (marine, estuarine) being deposited on top of more updip facies (terrestrial). In the case of valley fills consisting solely of fluvial facies, those facies reflect the change from a low-accommodation to a higher-accommodation style, for example by changing the channel stacking patterns, the relative preservation of overbank deposits, or the amount of organic facies (Fig. 21), and / or by a change in any paleosols from well-drained and more mature to poorly drained and immature as accommodation increases.

(5) As a result of filling in response to rising base level, depositional markers within the **deposits of the incised-valley fill onlap the valley base and walls** but do not occur outside the valley (smaller white horizontal arrow in Figure 22), except where they can be traced in a seaward direction into equivalent marine deposits.

(6) In terms of **sequence-stratigraphic surfaces** (Figs. 19–21), the formation of a valley generates a sequence boundary at the base, and a transgressive surface within the fill of a simple valley, or of each sequence constituting a compound valley fill. A maximum flooding surface lies above the valley fill in segment 1, within the estuarine deposits in segment 2, and likely low in the fluvial deposits in segment 3. Wave and tidal ravinement surfaces are commonly present between the sequence boundary and the maximum flooding surface in the areas transgressed by the shoreline. Additional flooding surfaces, bay ravinement surfaces, and erosional surfaces of more local extent, including bayhead and fluvial diastems, are likely to be formed during backstepping of fluvial and estuarine subenvironments.

(7) **Channels contained within the valley should be substantially smaller than the valley itself** (e.g., Figs. 18, 21, 23). However, it is recognized that channels that experienced only a short period of incision may be incised only slightly, with insufficient widening to form a pronounced valley. In addition, as discussed above, individual scours within a channel may be much deeper than the average channel depth, for example at tributary junctions (e.g., Best and Ashworth, 1997; Ardies et al., 2002). In these cases, the deeper scour could be mistaken for a valley but is of local extent only (Fig. 24), whereas a valley exhibits an elongate erosion surface of more regional extent. Where the valley and channel boundaries can be observed together, floodplain or terrace surfaces attached to channels within the valley can occur at lower stratigraphic elevations than the adjacent valley walls (M. Boyles, personal communi-

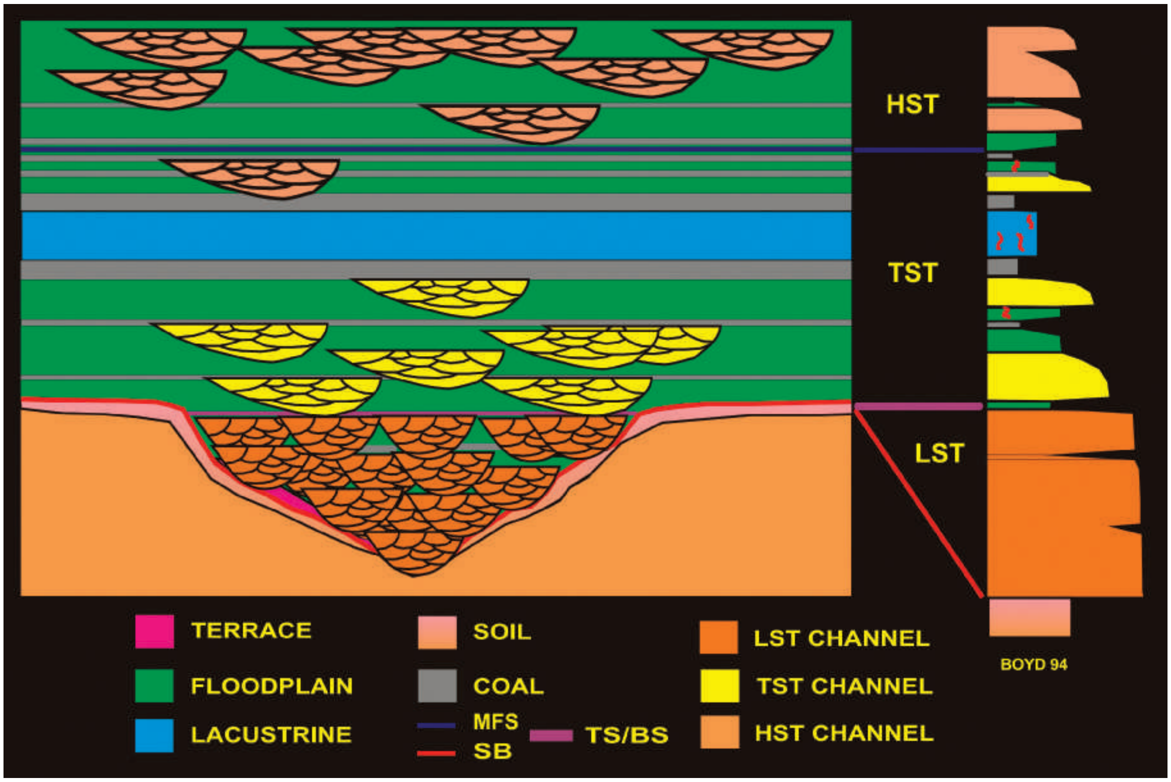


FIG. 21.—Nonmarine sequence-stratigraphic model showing the change in channel stacking patterns and organic facies responding to a cycle of accommodation change such as may be seen in segment 3 of an incised-valley fill. From Boyd and Diessel (1994).

cation, 2002) and / or interfluves outside the valley. This recognition feature augments those listed in 4 and 5 above.

(8) **Estuaries as defined above, following Dalrymple et al. (1992), are transgressive, tidally influenced environments that constitute an important and distinctive component of incised valleys in their seaward parts.** Because estuaries tend to enhance tidal action because of flow constriction, tidal indicators and distinctively tidal deposits may be especially abundant within the fill of incised valleys. The most distinctive of these (Fig. 25) are flat-lying tidal rhythmites and tidal bundles in cross beds, both of which record the neap–spring tidal cyclicity. In more general terms, single and paired (i.e., double) mud drapes, which give the deposits a heterolithic nature, are indicative of tidal sedimentation. In addition, other classic features such as reactivation surfaces, bidirectional paleocurrent patterns, herringbone cross stratification, flaser to lenticular bedding, and the large scale of cross-beds are distinctive (Dalrymple, 1992). In relatively low-accommodation settings and in basins with small tidal ranges, incised-valley fills may be the only place where tidal deposits are preserved. In such cases, the presence of tidal deposits can be used to suggest the existence of an incised valley.

(9) **The mixing of fresh and salty water is a fundamental characteristic of estuaries.** This stressed environment produces a characteristic ichnological suite and faunal composition (Pemberton et al., 1992, and early articles in Lauff, 1967) that are characterized by a low ichnospecies diversity, with populations consisting of small individuals (smaller then their open-marine counterparts) that exhibit opportunistic behavior (cf. Howard and Frey, 1973, 1975; Howard et al., 1975; MacEachern and Pemberton, 1994; Buatois et al., 1997; Buatois et al., 2005; Gingras et al., 1999; Pemberton et al., 2001). The degree of bioturbation (i.e., the bioturbation index; Droser and Bottjer, 1986, 1989) is commonly highly variable, with essentially unbioturbated beds interbedded with extensively bioturbated deposits that may contain a monospecific assemblage of traces. The unconformity at the base of the valley can display a *Glossifungites* ichnofacies (MacEachern et al., 1992), and individual forms such as *Gyrolithes* are distinctive of the estuarine environment (see the Brackish Ichnology section below). Brackish-water microfauna and macrofauna also display distinctive diversity and occurrence trends that are useful for the recognition of estuarine deposits, such as marsh foraminifera (*Ammonia, Haplophragmoides, Trachammina* sp.) that occur primarily in the intertidal zone in combination with *Spartina* sp. flora. Bivalves such as the modern *Rangia cuneata* that are overwhelmingly found in estuarine settings and oysters such as *Crassostrea* sp. are also useful environmental indicators. **However, it is important to note that many of these brackish-water features may occur in settings other than estuaries and should not be used on their own to interpret the presence of an estuary or an incised valley.**

(10) **Estuaries receive sediment input from both the marine and terrestrial ends of the system** (Figs. 9–12), creating the potential for the mixing of sediment with two different compositions. The sediment supplied directly by the river reflects the bedrock composition of the fluvial drainage basin, while the sediment provided by the marine source reflects shelf litholo-

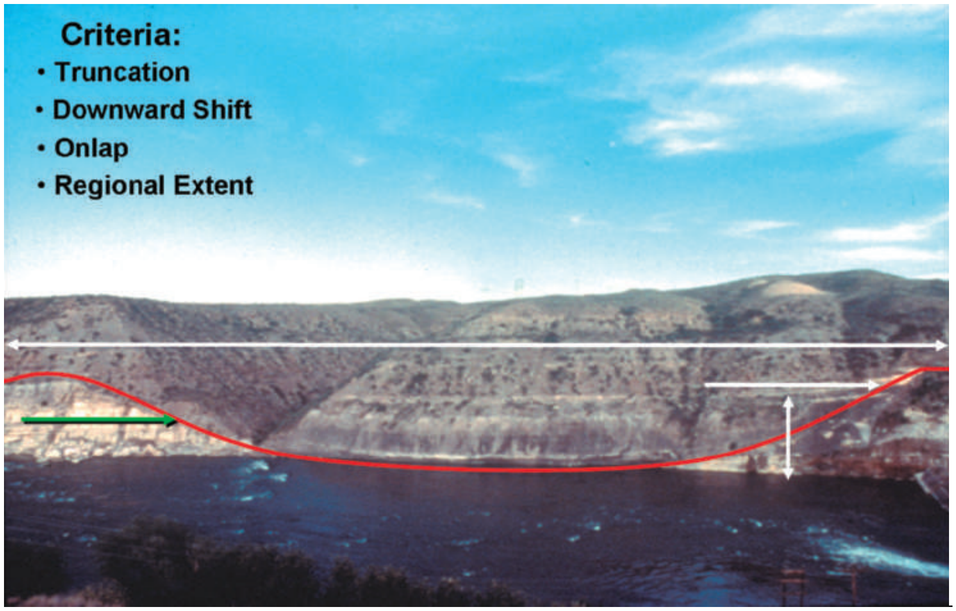


FIG. 22.—The four main criteria for recognizing an incised-valley system illustrated using a Lower Cretaceous (Basal Quartz equivalent), muddy incised-valley fill cutting into shoaling-upward shelf–shoreface parasequences along the Missouri River in northern Montana, U.S.A. Photo courtesy of P. Putnam, Petrel Robertson Research.

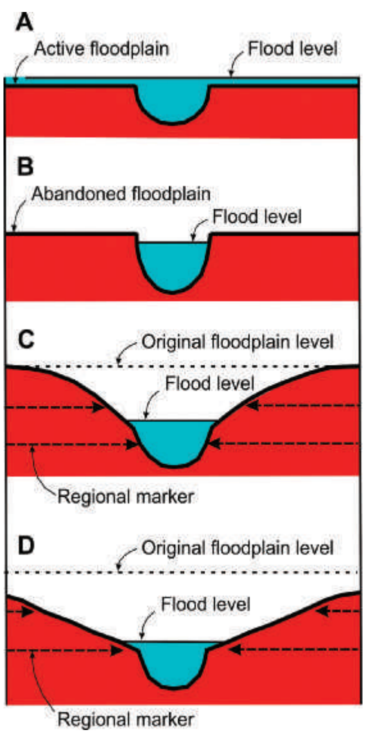


FIG. 23.—Incised-valley formation and entrenchment. If the floodplain is periodically inundated by large floods, the river is not incised, regardless of the relief between the low-stage water level in the river and the floodplain. The situation shown in Part B is the minimum incision required to qualify the river as incised in the modern, but such situations may be difficult to distinguish from non-incised channels in the ancient; the degree of development of floodplain paleosols, if preserved, would be the key distinguishing factor.

gies and / or source regions updrift in the longshore transport system. Because the marine-sourced material has been reworked from older deposits, it is often more mature than the terrestrial sediment (e.g., Roy, 1977). If source regions change during the fill of a compound valley, it may be possible to distinguish individual sequences within the compound valley fill by their compositional differences (e.g., Zaitlin et al., 2002).

(11) **E&IVs contain a characteristic mix of sedimentary facies. These include terrestrial, estuarine, and marine facies and range from fluvial, to tidal–fluvial channel, bayhead delta, central basin, barrier, and tidal sand ridge**. When found in combination, and especially when such facies are not present in the surrounding regional deposits, this set of facies identifies an estuarine setting, provided that they display a transgressive stacking arrangement (Figs. 16, 17) and may also point to the presence of an incised valley if a suitable container is present. Note that the presence of fluvial facies at the base of the estuarine or valley-fill succession is helpful but not essential for identification. Transgression subsequent to fluvial deposition can result in reworking and removal of fluvial facies by tidal and wave processes, especially by means of erosion at the bases of migrating tidal inlets. In other situations, the fluvial sediments may not be widespread and may occur only in a geographically restricted zone along the valley axis. Near the seaward end of segment 1 of the incised-valley system, all channel facies are likely to be tidal–fluvial in nature and hence display tidal features.

(12) **The central zone of incised-valley estuaries is occupied by a low-energy region** (Figs. 11, 13) representing either the finer-grained central basin of wave-dominated estuaries or the fine-grained meandering reach of tide-dominated estuaries.

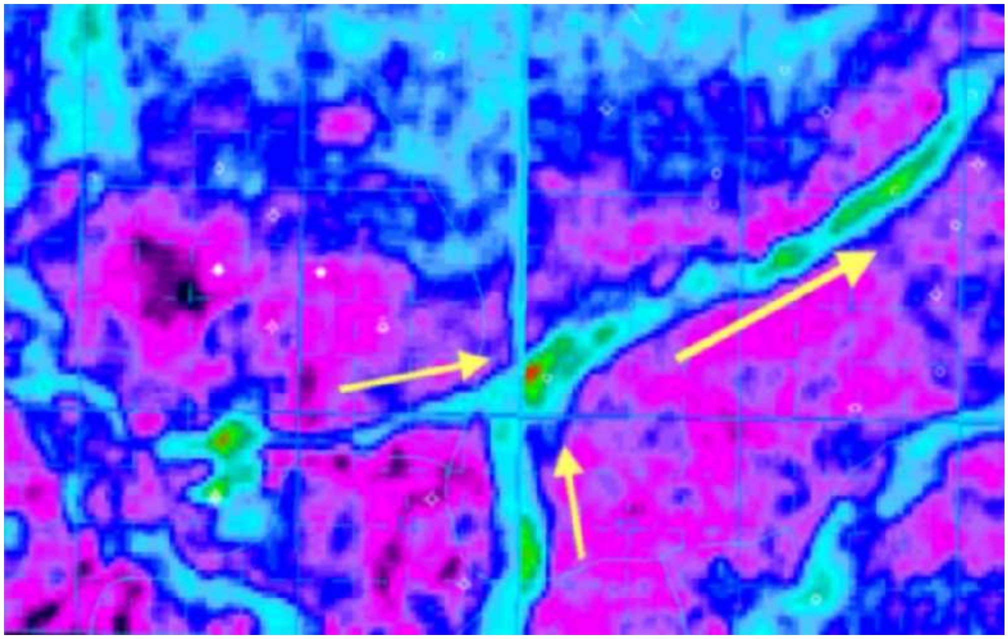


FIG. 24.—A 3-D amplitude anomaly map of a part of the Basal Quartz Formation (Lower Cretaceous) of southern Alberta, showing a tributary-junction scour (TJS) (cf. Ardies et al., 2001; 3-D image courtesy PanCanadian Energy (now EnCana) Corporation).

(13) **In the case of valley incision during regression and relative sea-level fall due to steepening of the fluvial profile as a result of seaward extension of the river, the regional marine gradient** is **greater than the terrestrial gradient of the river valley**.

(14) **E&IV deposits occupy fluvial drainage corridors**, **and their locations are often determined by underlying paleotopographic and structural trends**, with valleys occurring especially in areas of subtle downward flexure and/or parallel to fault traces (cf. Ardies et al., 2002; Plint and Wadsworth, 2003). By contrast, valleys tend to avoid areas of subtle upwarping.

An early example of an interpreted subsurface incised-valley system that subsequently met many of the above criteria for an incised valley was that of Harms (1966) in his description of the Cretaceous "J" Sandstone in western Nebraska. Harms' correlation (Fig. 2), based on a detailed electric-log cross section, demonstrated the truncation of regionally mappable, coarsening-upward marine parasequences by blocky to fining-upward fluvial valley-fill deposits, thus fulfilling recognition criteria (1), (2), (4), and (5) above. Other examples include those of the Mississippian Morrow Formation (e.g., Krystinik, 1989; Krystinik and Blakeney, 1990; Krystinik and Blakeney-DeJarnett, 1994; Bowen and Weimer, 1997, 2003) and parts of the Lower Cretaceous Glauconitic Formation (Wood and Hopkins 1989, 1992; Broger et al., 1997). An example of a more recent study that illustrates criterion (3) above is that of Ardies et al. (2002), who, in their study of the Basal Quartz unit, recognize tributaries and tributary junction scours, both in seismic and by detailed wireline well correlation.

It is critical when identifying the extent of the incised-valley system to document the geometry of the sequence boundary, both inside and outside of the incised valley. The paleotopography of the incised-valley network (e.g., tributary orientation or valley width/depth) may allow one to determine the paleodrainage direction as an aid in paleogeographic reconstruction. In addition, paleovalley networks are proving to be powerful tools for the identification of subtle structural warping and/or fault movement (e.g., Ardies et al., 2002), because rivers seek out the lowest part of the eroding landscape. A variety of techniques have been employed to identify and map paleovalleys, including: (1) geological structure mapping of the erosional surface from 2D–3D seismic (e.g., Broger et al., 1997) or from wireline logs (e.g., Krystinik, 1989; Van Wagoner et al., 1990; Krystinik and Blakeney-DeJarnett, 1994; Bowen and Weimer, 1997); (2) third- or higher-order residual mapping of the erosional surface in areas affected by postdepositional structuring (e.g., Zaitlin and Shultz, 1984, 1990); (3) detailed isopach mapping of the interpreted fill, or of an interval between the unconformity and an overlying horizontal marker that extends over the interfluves, to locate anomalously thick sections confined to the paleotopographic lows (e.g., Siever, 1951; Van Wagoner et al., 1990, Ardies et al., 2002). Other techniques include petrographic and chemostratigraphic typing of sediment composition, gravity techniques, resistivity mapping, and mapping of hydrocarbon production trends.

Model for a Simple Incised-Valley Fill

For simplicity, here we present a model for a simple incised-valley fill, based primarily on Zaitlin et al. (1994). We will consider the case of a piedmont incised-valley system, which is cut and filled in a single cycle of base-level change and which is connected to a marine shoreline; valleys that are located far inland with no marine link are considered later. We will also assume that fluvial sediment supply and the rate of transgression are constant, that waves are more significant than tides in the coastal zone, and that any estuaries that develop are wave-dominated (*sensu* Dalrymple et al., 1992) , because this is the situation most commonly documented in ancient incised-valley systems. For the sake of completeness, we have explicitly included the succeeding highstand systems tract, assuming that sediment supply is sufficient, relative to the length of the sea-level highstand, to allow shoreline progradation following the transgression. At times of high-frequency, high-amplitude sea-level changes, such as have occurred during the Pleistocene, this assumption may not be fulfilled, in which case sea level falls and the river reincises before the estuary is completely filled.

STRATIGRAPHIC ORGANIZATION OVERVIEW

Models for incised valleys that are connected to a marine shoreline are based on an ability to subdivide the valley fill longitudinally (Fig. 19) into three *segments* (as distinct from the tripartite estuarine facies *zonation* discussed above). This three-fold subdivision reflects the unique depositional and stratigraphic organization of the fill, which results primarily from lowstand erosion, followed by transgressive deposition, and finally highstand progradation.

As relative sea level falls, the entire length of the incised valley is characterized by (net) fluvial erosion, which creates the basal sequence boundary and may also leave intermediate, falling-stage terraces within the valley. When relative sea level starts to rise after reaching its lowest level, fluvial deposition begins at the mouth of the incised-valley system and extends progressively farther up the valley (i.e., the deposits onlap) as the transgression proceeds. Ideally, the fill of the seaward portion of the incised valley (segment 1, Fig. 19) is characterized by backstepping (lowstand to transgressive) fluvial and estuarine deposits, overlain by transgressive marine sands and/or shelf muds. The middle reach of the incised valley (segment 2, Fig. 19) consists of the drowned-valley estuarine complex that existed at the time of maximum transgression, overlying a lowstand to transgressive succession of fluvial and estuarine deposits like those in segment 1. The innermost reach of the incised valley (segment 3, Fig. 19) is developed headward of the transgressive estuarine limit and extends to the point where relative sea-level changes no longer control fluvial style. This segment is characterized by fluvial deposits throughout its depositional history; however, the fluvial style changes due to systematic variations in the rate of change of base level. The effect of base-level change decreases inland until eventually climatic, tectonic, and sediment-supply factors become the dominant controls on the fluvial system.

In the following sections we present additional detail on the characteristics of each incised-valley segment and then provide a range of representative outcrop and subsurface studies.

FIG. 25.—Examples of diagnostic tidal sedimentary structures; **A)** tidal rhythmites, **B)** tidal mud drapes in a cross bed that separates the cross bed into tidal bundles (from MacEachern and Pemberton, 1994).

Segment 1—Outer Incised Valley

The outer incised valley (segment 1) extends from the most seaward extent of valley incision, near the lowstand mouth of the incised valley, to the point where the shoreline stabilizes at the beginning of highstand progradation (Fig. 19). As in the other segments, this reach of the valley initially undergoes fluvial incision with the lowering of base level. Sediment is bypassed to the mouth of the valley, where it is deposited as a lowstand delta and/or prograding shoreline. This period is represented by the sequence boundary, which may be overlain by lowstand fluvial to tidal–fluvial deposits (Fig. 20, profile 1). As sea level begins to rise and the lower reaches of the system are transgressed, the lower reaches of the incised valley change from being a conduit for fluvially eroded sediment to the site of fluvial and (subsequently) estuarine deposition. Fluvial deposition, although initiated during the late lowstand, continues during the early stages of transgression, with the locus of deposition shifting landward as relative sea level rises and the shoreline transgresses (Wright and Marriott, 1993; Wescott, 1993). The transition from erosion and fluvial bypass to fluvial deposition migrates landward as the transgression proceeds. Thus, the boundary between the lowstand and transgressive systems tracts (i.e., the transgressive surface) may lie within the fluvial deposits rather than at their top and is diachronous if it is picked at a facies boundary. For this reason, the lowstand systems tract (LST—i.e., those deposits that accumulated before the shoreline begins to migrate landward) within the valley may effectively pinch out landward (Figs. 19, 20), although there should be at least a thin layer (ca. one channel depth thick) of lowstand-age fluvial deposits along the length of the valley, unless they have been removed by later channel erosion. Near the mouth of the valley, most of the fill may be deposited during lowstand-systems-tract time, but farther up the valley the greater part of the fill accumulates during transgressive-systems-tract time.

Within the fluvial succession near the river mouth, the early deposits accumulate when the rate of creation of accommodation is low (i.e., near maximum lowstand time), hence channel amalgamation is common, leading to the formation of a coarse-grained succession in which muddy overbank deposits are scarce. As base level begins to rise ever more rapidly during the TST, the fluvial channels become progressively less amalgamated and fine-grained deposits are preserved more commonly (Fig. 21; e.g., Boyd and Diessel, 1994). The fluvial style (i.e., braided, meandering, anastomosed, or straight) within the incised valley is dependent on a variety of factors, including sediment supply, grain size, discharge, valley gradient, and rate of transgression (Schumm, 1977, 1993; Schumm and Ethridge, 1994). These variables likely change during the rise in sea level associated with the marine transgression (Gibling, 1991; Wright and Marriott, 1993; Törnqvist, 1993). Thus, in the simplest case where all other factors remain constant, the character of the lowstand to transgressive fluvial sediments should change vertically as the depositional gradient and capacity of the fluvial system decreases as the shoreline approaches. This change would most likely result in successively younger channels having finer-grained sands than preceding channels, in part because of the seaward decrease in grain size within the river but also because of deposition of the coarser portions of the sediment load in more inland areas. This overall upward decrease in the grain size of subsequent channels accompanied by and upward decrease in channel amalgamation, with a change from higher-energy (such as sandy braided) to lower-energy, (such as mixed-load meandering) fluvial deposits. An excellent example of this is provided by the Quaternary sediments in the Rhine–Meuse valley (Törnqvist, 1993), and in conceptual form in Figure 21. Note, however, that the amalgamated channel deposits at the base of the valley fill cannot be assumed to be braided, simply because of the absence of overbank deposits; they could equally well be meandering-river deposits with negligible preservation of muddy overbank deposits because of the low accommodation.

The thickness of the fluvial succession, and the extent to which the predicted changes in fluvial style are developed, may be variable along the length of segment 1. The ultimate thickness is controlled by the accommodation developed during the rise in sea level (Jervey, 1988), with the major factor being the ratio of the rate of fluvial-sediment input to the rate of sea-level rise. In the situation where sea-level rise greatly outpaces fluvial input, transgression is rapid and the thickness of the fluvial deposits is less than in the case where abundant sediment input occurs during a slow rise in sea level. In the special case where sediment input matches sea-level rise, the fluvial deposits aggrade vertically and the shoreline does not transgress. In all cases, the preserved thickness of the fluvial succession may be affected by subsequent erosion associated with transgression. While this fluvial stacking is best preserved in Segment 3 (discussed below), and documented in Arnott et al. (2000, 2002) and Lukie et al. (2002) from the Basal Quartz Formation, examples of preserved fluvial stacking controlled by accommodation in Segment 1 are found in the Upper Cretaceous of the Kaiparowits Plateau, Utah (Shanley and McCabe, 1991, 1994) and the Mesaverde Group (Olsen et al., 1995).

As the transgression proceeds, the estuarine conditions that are established in the seaward end of the valley migrate landward. In a wave-dominated estuarine setting, the first estuarine deposits over the fluvial sediments are tidally influenced fluvial and bayhead-delta (distributary channel, levee, and interdistributary bay) deposits (Fig. 20, profile 1). As transgression continues, central-basin deposits then overlie the bayhead delta across a flooding surface that may correlate updip to a change in fluvial style. The central-basin deposits in turn are overlain by the estuarine flood-tidal-delta and other barrier deposits (cf. Boyd et al., 1992; Dalrymple et al., 1992). This contact may be gradational if it corresponds to the prodelta deposits of the flood tidal delta, but it is equally likely to coincide with the erosional base of a tidal channel (Boyd and Honig, 1992), with the deepest incision occurring at the location of the tidal inlet. The erosion surface at the base of such channels is referred to as a tidal ravinement surface (Allen and Posamentier, 1993).

As transgression proceeds, the shoreface passes the former location of the estuary. Wave erosion associated with shoreface retreat produces a wave ravinement surface that may truncate the underlying estuarine deposits (Fig. 20, profile 1; e.g., Ashley and Sheridan, 1994; Belknap et al., 1994; Kindinger et al., 1994; Thomas and Anderson, 1994). The depth of erosion depends on a variety of factors, the more important of which are:

1. The depth of the base of the shoreface: a more intense wave climate leads to deeper erosion.

2. The resistance to erosion of the interfluves: lithified bedrock is more resistant to erosion than unconsolidated material, and may cause the shoreface to ride up and over the valley fill.

3. The depth of the valley: shallow valleys may be completely removed, whereas more of the fill of deeper valleys escapes truncation.

4. The rate of relative sea-level rise: more rapid rates of sea-level rise promote more rapid transgression, which reduces the potential for deep truncation.

In many cases, all but the deepest and most landward parts of the estuary-mouth sand body are removed. Flood-tidal deltas and the bases of the deepest tidal channels, including the tidal inlet, have the highest preservation potential (e.g., Belknap et al., 1994, Thomas and Anderson, 1994). The wave ravinement surface may then be overlain by transgressive shoreface to nearshore sands, which may vary in thickness from almost nothing to many meters in shelf sand banks and ridges (Snedden and Dalrymple, 1998) that were created by shelf processes. Finally, the valley is capped by open-marine mudstones associated with the succeeding highstand. The landward limit of these mudstones is an indicator of the inner end of segment 1.

Segment 2—Middle Incised Valley

Segment 2 lies between the inner end of segment 1 (i.e., the initial highstand shoreline) and the estuarine limit (i.e., the landward limit of recorded tidal influence) at the time of maximum flooding (Fig. 19). It therefore corresponds to the area occupied by the drowned-valley estuary at the end of the transgression. In this segment the sequence boundary is overlain by lowstand to early-transgressive fluvial deposits similar to those in segment 1. These are in turn overlain by transgressive estuarine facies, but in this segment the nature of the overlying estuarine succession varies along the length of the segment (Figs. 16, 17; cf. Dalrymple et al., 1992) because the estuarine facies are (ideally) preserved with the spatial distribution they had in the contemporaneous estuary.

Near its seaward end, (i.e., beneath the preserved barrier that forms the landward margin of any subsequent highstand strandplain, assuming as we have throughout this discussion that the coastline is wave dominated) the succession is similar to that in segment 1, with fluvial and bayhead-delta sediments overlain by central-basin deposits , which are, in turn, capped by estuary-mouth-barrier sands. Because open-marine conditions do not transgress into this segment, the barrier sediments are overlain by highstand fluvial deposits (Fig. 20, profile 2), unless sea level falls before the estuary fills completely, in which case the estuarine deposits are capped by the next sequence boundary. In the middle portion of segment 2, barrier sands are absent, and central-basin deposits coarsen upwards above the maximum flooding surface into progradational, bayhead-delta and fluvial sediments of the succeeding highstand deposits (Fig. 20, profile 3) that fill the estuary if the highstand is of sufficient duration. At the headward end of segment 2, central-basin sediments are absent, and the bayhead delta is overlain directly by highstand fluvial deposits (Fig. 20, profile 4). The most landward limit of the detectable marine influence (i.e., tidal features in fluvial deposits) is taken as the inner end of segment 2. This point corresponds with the inner end of the estuary as defined by Dalrymple et al. (1992), and is also approximately equivalent to the "bayline" of Posamentier et al. (1988) and Allen and Posamentier (1993).

Barrier islands are rarely preserved in incised valleys because typically they are removed by shoreface ravinement during transgression. However, preservation may be possible at the highstand shoreline as the barrier stabilizes and then evolves into a strandplain, as is just beginning on Galveston Island, Texas (e.g., McCubbin, 1982). A potential subsurface example of such a preserved barrier sand body is provided by the Lower Cretaceous Lloydminster Member (Mannville Group) Senlac heavy-oil pool in southwestern Saskatchewan (Zaitlin and Shultz, 1984, 1990).

Segment 3—Inner Incised Valley

The innermost segment (segment 3) of the incised-valley system lies landward of the transgressive marine–estuarine limit, but it is still influenced by changes in base level associated with relative sea-level change (Fig. 19). This segment may extend for tens to hundreds of kilometers above the limit of marine/estuarine influence (Shanley et al., 1992; Schumm, 1993; Levy and Christie-Blick, 1994). The fill of this segment is entirely fluvial, with no evidence of tidal action or brackish water. Channels may be braided, meandering, anastomosed, and/or straight, depending on a variety of factors such as sediment supply, gradient, discharge, and sediment size. However, relative sea-level changes associated with the lowstand–transgression–highstand cycle produce predictable variations in the rate of creation of accommodation through time and may also produce a predictable vertical succession of fluvial styles (Fig. 20, profile 5; Fig. 21; Gibling, 1991; Wright and Marriott, 1993).

Lowstand fluvial deposits are expected to be relatively thin, because the fluvial system in these inland locations would have been erosional or would have acted mainly as a transport conduit (a bypass zone) at that time. Late lowstand to early transgressive deposits at the base of the fill may be characterized by relatively coarse-grained, amalgamated channel deposits (Fig. 21). As transgression proceeds, an overall upward-fining succession of channels should be developed as the gradient and stream capacity decrease as the backwater zone landward of the estuary migrates up the valley. The deposits that accumulated during times of rising base level should contain more isolated, channel-sandstone bodies, interbedded with a higher percentage of overbank deposits (e.g., Törnqvist, 1993, Shanley and McCabe, 1994). Freshwater organic facies (e.g., peat or lacustrine carbonates) might be abundant and the soils less mature and wetter than those associated with the lowstand (Cross, 1988; Boyd and Diessel, 1994; Wadsworth et al., 2002). The overlying highstand deposits may be expected to coarsen upward overall, due to progradation in response to decreasing rates of base-level rise and accommodation creation (Schumm, 1993).

In terms of relative length, the three incised-valley segments identified above may be quite variable. If the transgression has been extensive, however, segment 1 is likely to be long and may extend for most of the width of the formerly exposed continental shelf. The length of segments 2 and 3 is related to the depth of valley incision and the gradient above the highstand shoreline. For example, on many old, wide passive margins such as the Gulf of Mexico, segment 1 is much longer than segments 2 and 3 (Thomas and Anderson, 1994; Blum and Törnqvist, 2000). Overall, segment 2 is likely to be the shortest of the three because it corresponds to the length of the estuary at one point in the sea-level cycle.

ANCIENT CASE STUDIES OF INCISED-VALLEY DEPOSITS

The model for an incised-valley fill described in the preceding paragraphs has been applied to a large number of ancient examples. Here we review several of these to illustrate typical examples and to highlight controls on the nature of incised-valley deposits that are not discussed elsewhere in this chapter, such as the influence of the overall accommodation regime on the character and stratigraphic organization of such deposits. The petroleum-industry applications are highlighted in several of these examples.

Case Study 1: The Mississippian Morrow Formation: Fluvial to Fluvial–Estuarine Deposits of Segment 1

A number of well-documented subsurface examples of segment 1 incised-valley deposits exist from the Western Interior Seaway of North America. One such example is the Carboniferous Morrow Formation of the Anadarko Basin, as described by Krystinik and Blakeney-DeJarnett (1994), Krystinik (1989), and Bowen and Weimer (1997) (Figs. 26, 27).

The Morrow Formation has been the target of extensive exploration over the last forty years and is characterized by several well-documented productive trends. The Morrow Formation is distributed on the north flank of the Anadarko Basin, in what was a broad, low-relief shelf subject to glacio-eustatic exposure and inundation. During glacio-eustatic lowstands (Fig. 27B), the shelf was largely exposed and subject to fluvial erosion by drainage networks that fed deltas along the rim of the Anadarko Basin. During interglacial highs (Fig. 27A) the shelf was inundated, with the deposition of mudstone and carbonate. The shoreline position is thought to have moved in excess of 145–200 km (90–125 miles) per cycle.

The Morrow incised valleys are characterized by multiple exposure surfaces and fluvial incision interpreted to have been cut by repeated high-frequency glacio-eustatic sea-level drops, and backfilled with fluvial and fluvial–estuarine deposits during transgression. The Sorrento–Mt. Pearl–Siaana and Stateline trend is an example of a well-documented Mississippian Morrow structurally controlled valley that is mappable over hundreds of kilometers (Fig. 26). Repeated transgressive–regressive events developed a compound valley and terrace geometry (Leighton, 1997), similar to that observed in the modern Colorado River in Texas (Blum, 1990, 1994). Individual incised-valley systems are between 50 and 80 feet (15–25 m) thick and 0.5 and 2 miles (0.8–3.2 km) wide (Krystinik and Blakeney-DeJarnett, 1994; Krystinik, 1989). The incised-valley systems are cut into marine mudstones and limestone of the preceding highstand and are blanketed by similar deposits of the succeeding highstand. Multiple unconformity and exposure surfaces merge onto the interfluves.

An individual cycle of fill from segment 1 of the Morrow incised valleys consists, from base to top, of:

(1) basal (braided) fluvial deposits composed of coarse–medium-grained cross-bedded sandstones (core porosity to 25%; core permeability 0.1–4 darcys), grading upward into

(2) meandering fluvial (core porosity 20–25%; core permeability 0.1–300 md) and floodplain mudstones and green-waxy paleosols that are overlain by

(3) estuarine (bayhead delta) sandstones (core porosity 3–12%; core permeability 0.1–2 md) and mudstones displaying tidal influence and restricted bioturbation, topped by

(4) glauconitic sandstone and transgressive marine mudstones that rest on a shell-rich pebble lag (i.e., the wave ravinement surface) that indicates the Segment-1 character of this example. The "hour glass" shape of the well logs through the fill (i.e., a basal blocky to fining-upward fluvial to estuarine succession, overlain by coarsening–upward central-basin to estuary-mouth deposits) appears to be characteristic in most

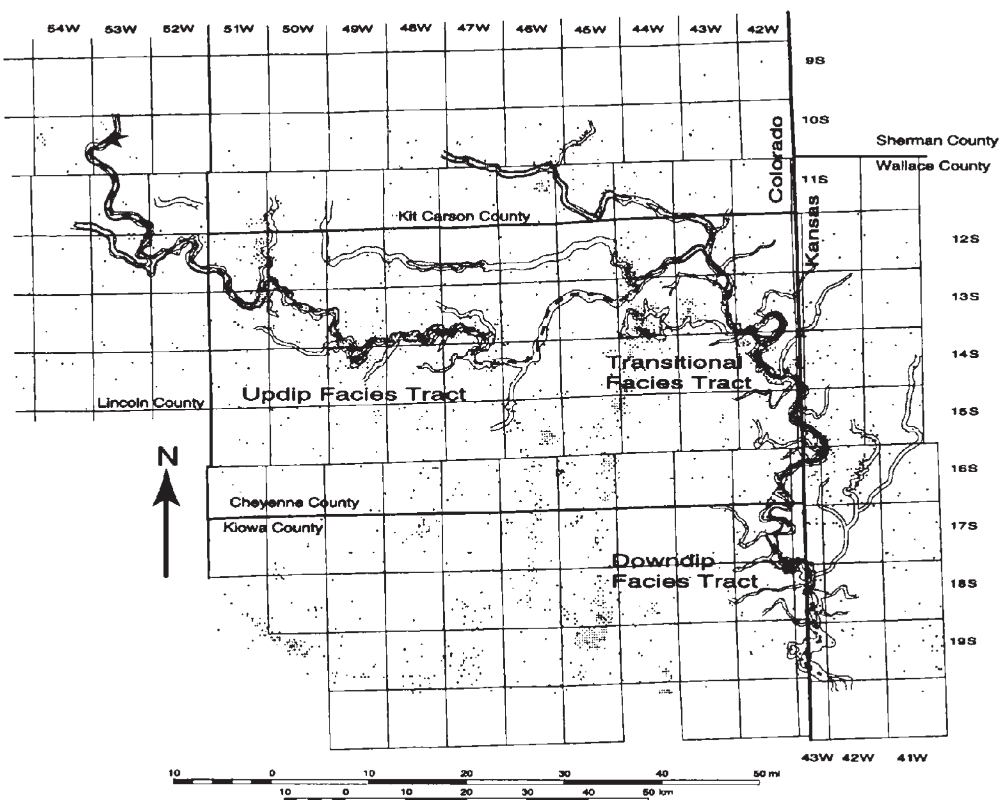


FIG. 26.—Map showing the distribution of the Mississippian Morrow Formation incised-valley fills (from Bowen and Wiemer, 2003).

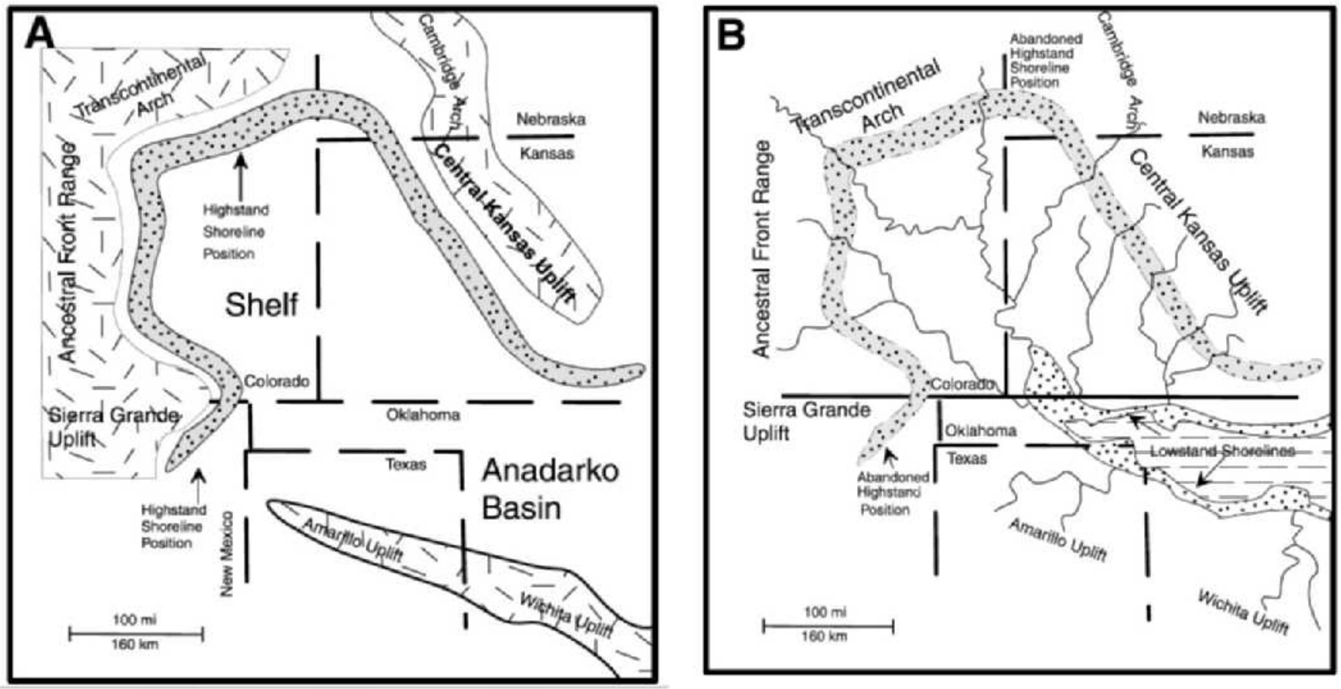


FIG. 27.—Schematic diagram showing the distribution of depositional systems during deposition of the Morrow Formation. **A)** During relative highstands of sea level, shorelines rimmed the basin and black muds were deposited on a broad, shallow-marine shelf. **B)** During relative lowstands, an extensive series of river valleys were developed in eastern Colorado and western Kansas that flowed into the Anadarko Basin (from Bowen and Wiemer, 2003).

of these incised-valley systems. Outside of the incised-valley networks, the interfluve areas are characterized by extensive paleosol surfaces.

Case Study 2: Lower Cretaceous Glauconitic Sandstone: Fluvial to Fluvial–Estuarine Deposits of Segment 1

The Lower Cretaceous Glauconitic Formation of the Western Canada Sedimentary Basin is characterized by a network of northwestward-trending, compound, piedmont incised-valley systems that are interpreted to feed lowstand to early transgressive east–west-trending shorelines to the north. The Glauconitic incised-valley system is mappable for at least 535 km south–north, from Montana into central Alberta (Wood and Hopkins, 1989, Sherwin, 1994, Broger et al., 1997, Peijs-van Hilten et al., 1998). The fills of the 1–5 km wide valleys exhibit a progressive northward change in character. In the south, the fill consists of lowstand to early transgressive, fluvial to fluvial–estuarine deposits characterized by multiple erosive events (i.e., they are compound valley fills) as a result of low accommodation. In the north, the accommodation was greater and the individual valleys are separated by coarsening-upward highstand shoreface parasequences, resulting in full preservation of individual, simple valley fills.

The Countess–Alderson trend is a 56 mile (90 km) reach of one such Glauconitic IVS that extends over 300 miles (480 km) from northern Montana into central Alberta, Canada. Along this reach there are 122 hydrocarbon pools (e.g., Countess YY and Lathom A pools) that have produced over 100 MMBBL of oil and 300 BCF of gas since the 1950s. Recent optimization of many pools using a multidisciplinary approach has led to a better understanding of the nature of this incised-valley system. The majority of pools produce from backstepping (LST to TST), transgressed fluvial and estuarine bayhead-delta and central-basin deposits.

The Countess YY pool (Fig. 28), one of several reservoirs located beneath Lake Newell and adjacent areas in southern Alberta, Canada, is interpreted by Broger et al. (1997), Peijs-van Hilten et al. (1998), and Zaitlin et al. (1998) to lie within segment 1 of a wave-dominated incised-valley system. A low-permeability Middle Glauconitic channel (Fig. 29) incises into the producing channel and locally forms an updip seal to trap hydrocarbons in the Lower Glauconitic channel.

Both the Countess YY and Lathom "A" pools contain a number of characteristic depositional facies that are stacked in a manner that is consistent with the vertical succession proposed for a segment 1 incised-valley system. The base of the valley is overlain by *fluvial facies* that consist of litharenitic, coarse- to medium-grained, large-scale trough and planar-tabular cross-bedded sandstone that overlies erosional surfaces that are usually covered by a pebble lag (Figs. 30, 31). This facies has excellent reservoir quality (Fig. 29), ranges in thickness from 1 m to more than 10 m, and is encountered at the base of the incised-valley system. The gamma-ray log signature shows a blocky or fining-upward profile. The sediments are interpreted to be deposited by a highly connected braided to coarse-grained meandering fluvial channel system.

The bay-head delta facies has moderate to poor reservoir quality and either gradationally overlies the fluvial facies or immediately overlies the basal sequence boundary in areas off the axis of the valley. Thickness ranges from 3 m to more than 11 m. The gamma-ray log signature shows an overall coarsening-upward trend, indicating a progradational environment. Individual blocky to fining-upward units 3–7 m thick are interpreted to represent bayhead-delta distributary channels. Single and

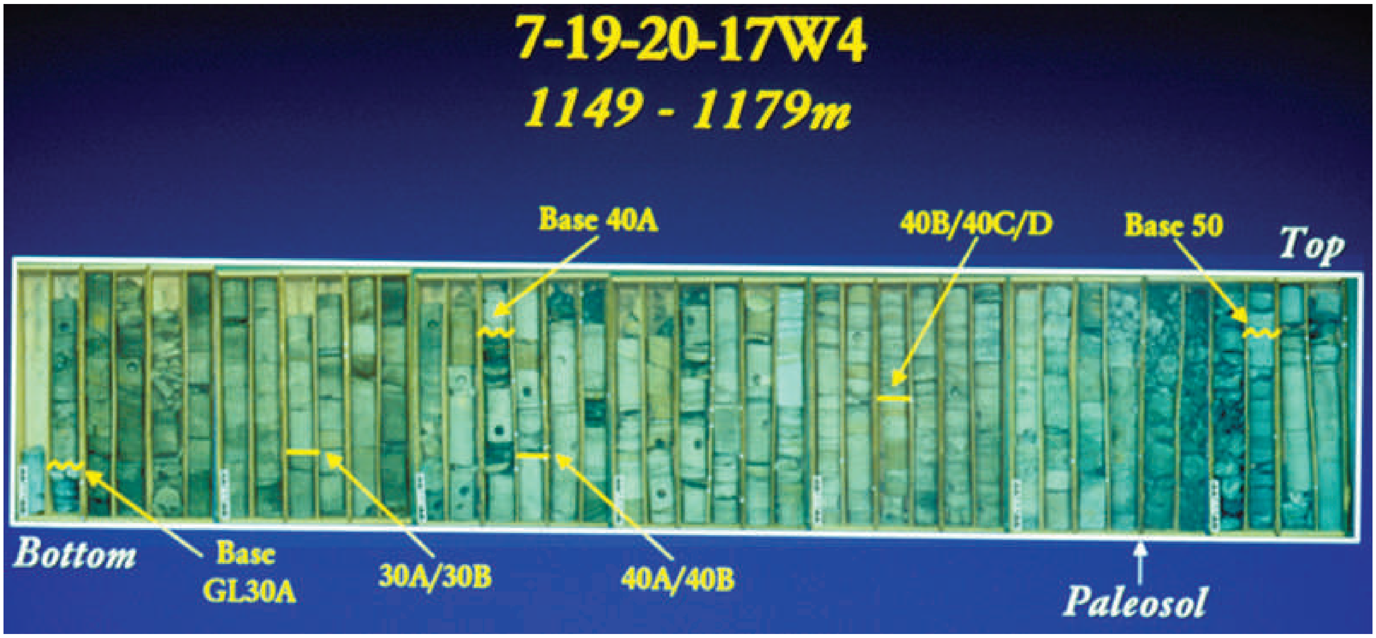


FIG. 28.—An example of a compound incised-valley fill from the type Lower Cretaceous Glauconitic Formation well from the Latham "A" field in southern Alberta, Canada. (Zaitlin et al., 1998). Bottom of core to the lower left; top to the top right. Glauconitic Sandstone Member; 30, 40, and 50 represent informal units in the Glauconitic; A = fluvial facies, B = tidal–fluvial, C/D = tidal–fluvial bayhead-delta to central-basin facies.

stacked channel units show a fining-upward and an obvious shaling-upward trend in core and on wireline logs, with tidal mud drapes (Fig. 32) being more abundant in the upper parts of the units, indicating either an increase in tidal influence or a decrease in energy during deposition. These *bayhead-delta distributary-channel deposits* are composed of medium- to coarse-grained, planar-tabular cross-bedded, flaser-bedded, and tidally bedded sandstones (Fig. 32). Massive to repetitive fining-upward successions are characterized by basal scour surfaces marked by shale rip-up clasts and channel lags. From seismic-amplitude maps, a northwest (downvalley) bifurcation of the channel facies is observed, indicating a distributary-channel pattern (Fig. 33). Evidence of tidal activity is indicated by the presence of mud drapes and couplets, as well as by a typical estuarine ichnofossil assemblage in the associated central-basin facies (Fig. 34).

In some cases, these channel deposits display a more heterolithic character and are interpreted to consist of an *inclined heterolithic tidal point-bar facies* that is characterized by a sharp to erosional basal contact with a fining-upward trend. These units consist of fine, massive to tidally bedded, flaser-bedded sandstones, alternating with 2-cm-thick continuous mudstones. All of the strata display a consistent dip of 5–19° and can be considered to be inclined heterolithic stratification (IHS; cf. Thomas et al., 1987) of point-bar origin. Locally, a restricted trace-fossil assemblage may be present. This facies has moderate to good reservoir quality and overlies the bayhead-delta distributary-channel facies. Thickness ranges from 5 to 7 m. The gamma-ray log signature shows an irregular but clearly fining-upward profile. The abundance of inclined shale intervals increases toward the top of the succession. Sandstone intervals containing tidal mud drapes also are more abundant in the top part of this facies.

The *delta-front turbidite facies* lies adjacent to the distributary-channel deposits and consists of a regular interbedding of planar to wavy parallel-laminated sandstones and weakly burrowed, dark-gray mudstones (Broger et al., 1997; Peijs-van Hilten et al., 1998). Locally an abundance of wave-generated physical sedimentary structures are present, such as current-ripple lamination. Fine mud laminae are present in some intervals. The mudstone beds are locally highly carbonaceous, and typically much thinner than the intervening sandstone beds (1–5 cm thick). They commonly contain convolute lamination, syneresis cracks, and small-scale gravity faults. Bioturbation is rare in the sandier portions of the facies but increases in the mudstone interbeds. The trace-fossil assemblage is restricted in diversity *(Planolites, Teichichnus, Cylindrichnus, Skolithos, and Tigillites* sp.), indicating the presence of a stress, most likely because of salinity fluctuations or water turbidity. The heterolithic character indicates repeated fluctuations in the energy regime, and the sedimentary structures indicate that the sand beds were emplaced by density flows that were caused by wave, storm, and/or river-flood processes. The deformation features indicate a depositional slope, and failure of the heterolithic succession. These deposits have thicknesses ranging from 8 to 24 m. The gamma-ray log clearly shows an irregular alternation of clean sandstone and shale intervals. Reservoir properties and thicknesses of the sandstone intervals increase upward, suggesting a sanding-upward and coarsening-upward trend that indicates progradation. This facies is interpreted to be deposited in a bayhead delta-front turbidite environment, and is inferred to have a lobate geometry. In some wells, the successions show overall lower porosity and permeability values than elsewhere, indicating an areal variation in grain size or sand proportion.

The *central-basin facies* consists of fine-grained, rippled, flaser-bedded and tidally bedded sandstones displaying abundant shale laminae and double mud drapes with a low-diversity ichnofossil assemblage (Fig. 34). This facies has poor reservoir quality and occurs in intervals with a thickness of 1 to 5 m at various stratigraphic positions, most commonly on top of bayhead-delta sandstone facies and below capping marine shales or the crosscutting Middle Glauconitic channel sediments. The gamma-

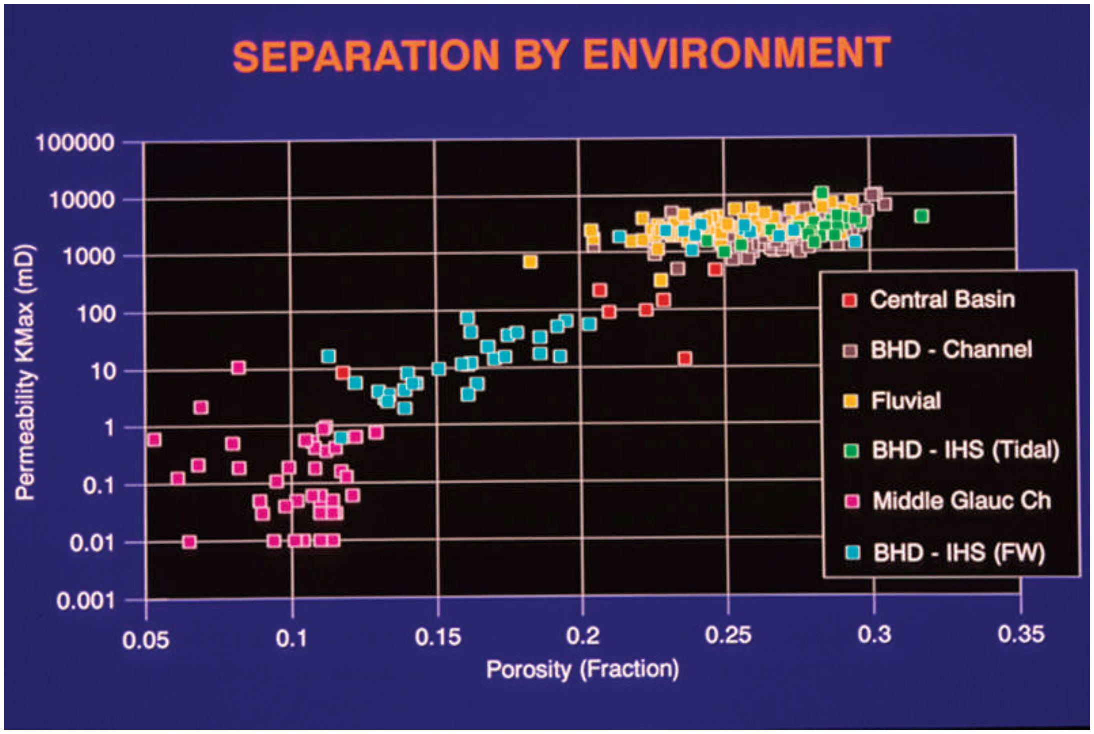


FIG. 29.—Porosity (%) versus permeability (millidarcys) cross plot for the various incised-valley facies in the Lower Cretaceous Glauconitic Formation, southern Alberta, Canada. The data points cluster about four major categories: (1) best reservoir produceability occurs in fluvial, bayhead-delta channel, and bayhead-delta tidal point-bar (IHS) facies; (2) moderate-reservoir deposits consisting of sandy, central-basin bay-fill deposits; (3) moderate- to poor-reservoir deposits consisting of muddy, bayhead-delta fresh-water and central-basin deposits; and (4) Middle Glauconitic lithic-channel facies that locally forms a lateral seal to the reservoirs because of extensive diagenetic alteration (Broger et al., 1997). BHD = bayhead delta; IHS = inclined heterolithic stratification; CH = channel; FW = fresh water.

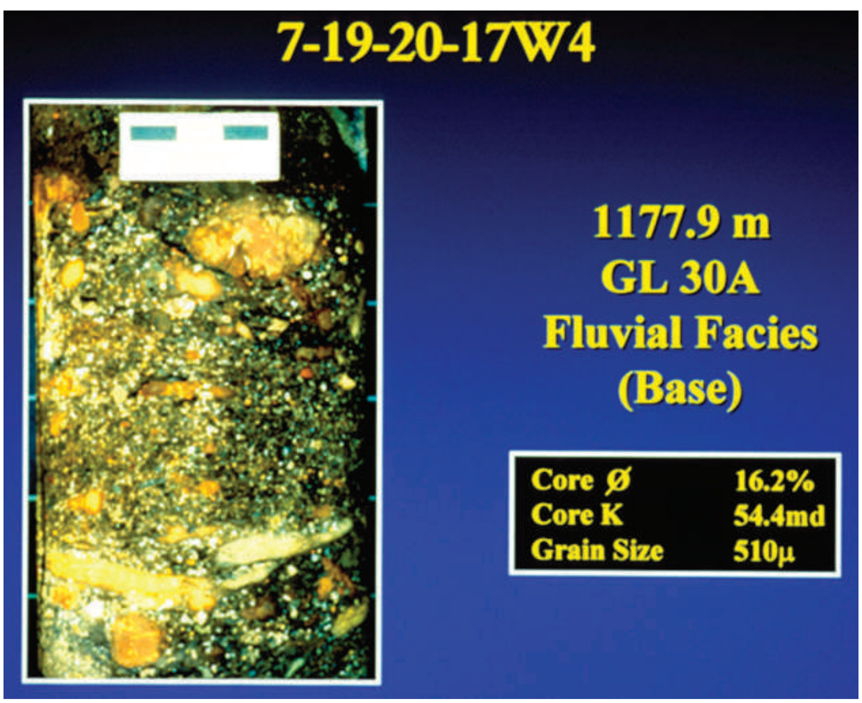


FIG. 30.—A typical basal pebbly fluvial-lag facies from the Lathom "A" 7-19-20-17W4 well, with its associated grain size, porosity (Ø), and permeability (K) values (from Zaitlin et al., 1998).

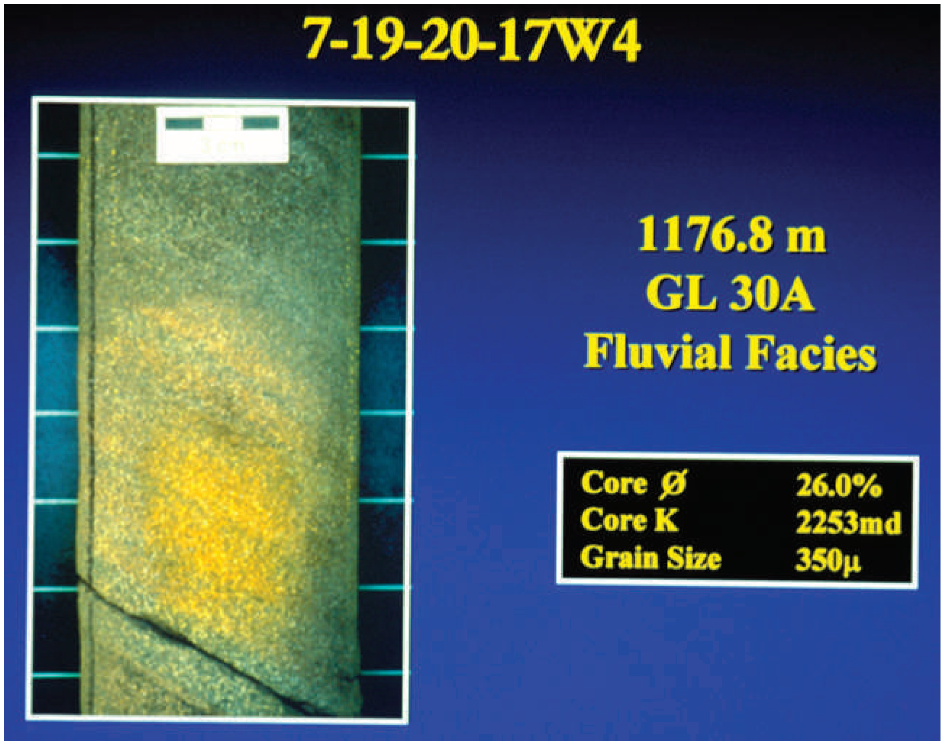


FIG. 31.—Sandy cross-bedded fluvial facies from the Lathom "A" 7-19-20-17W4 well, with its associated grain size, porosity (Ø), and permeability (K) values (Zaitlin et al., 1998).

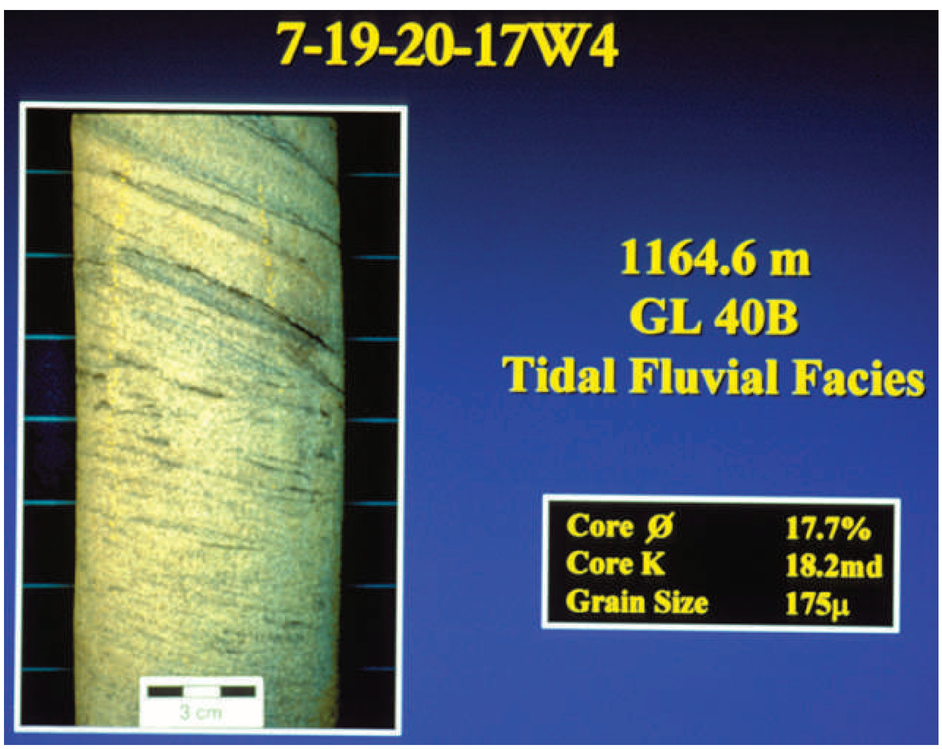


FIG. 32.—Typical tidal–fluvial facies with tidal couplets and mud drapes from the Lathom "A" 7-19-20-17W4 cored well, with its associated grain size, porosity (Ø), and permeability (K) values (from Zaitlin et al., 1998). The mud layers reduce permeability and make this facies a poorer reservoir than the fluvial facies (Figs. 30, 31) that underlie these deposits (see Fig. 29).

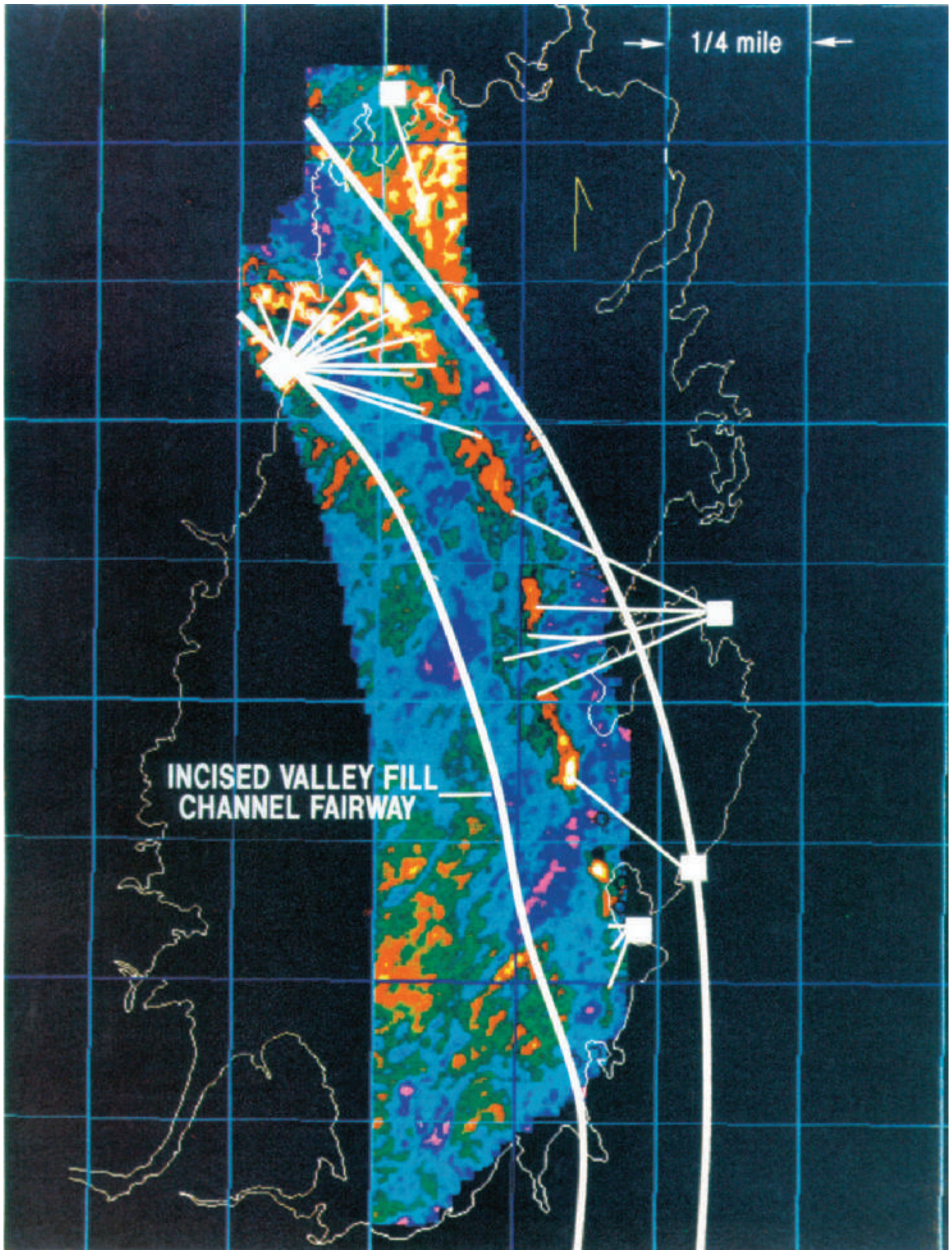


FIG. 33.—Map of the Lake Newell area, southern Alberta, Canada, showing the distribution of seismic amplitudes along the trend of the lower Glauconitic incised valley (margins indicated by heavy white lines). Warm colors are interpreted to be either porous sand (i.e., potential hydrocarbon reservoirs) or undisturbed Ostracod shale, whereas cooler colors indicate nonporous shales. Note the northwestward bifurcation of the inferred sandstones in the northern part of Lake Newell. This pattern is interpreted to represent a bayhead delta. The locations of the drillsites are also shown. (Broger et al., 1997.)

ray log signature is irregular. This facies is interpreted to have been deposited in the central-basin environment. As is common in segment 1, no preserved barrier exists in the study area because it was removed by ravinement. This is particularly true of low-accommodation settings; in areas with higher accommodation, portions of the barrier (dominantly the tidal inlet and flood-tidal channels that cut down into the central-basin deposit) may escape removal—e.g., Cretaceous Viking Formation in the Crystal Field (Reinson et al., 1988; Pattison, 1991 or in outcrops of the Paddy–Cadotte interval (Leckie and Singh, 1991), both in Alberta.

Case Study 3: Lower Cretaceous Senlac (Lloydminster Formation) Sandstone: An Example of an Estuary-Mouth Barrier Sandbody of Segment 2

The Senlac heavy-oil pool, located in Townships 38–39, Range 26–27W3, of Saskatchewan (Fig. 35), was discovered in 1980. It has been estimated to contain 1.3×10^7 m^3 (84.3×10^6 barrels) of 13–15 degree API oil in a barrier and tidal-inlet complex at the mouth of a paleovalley system (Fig. 36). The existence of an intact barrier

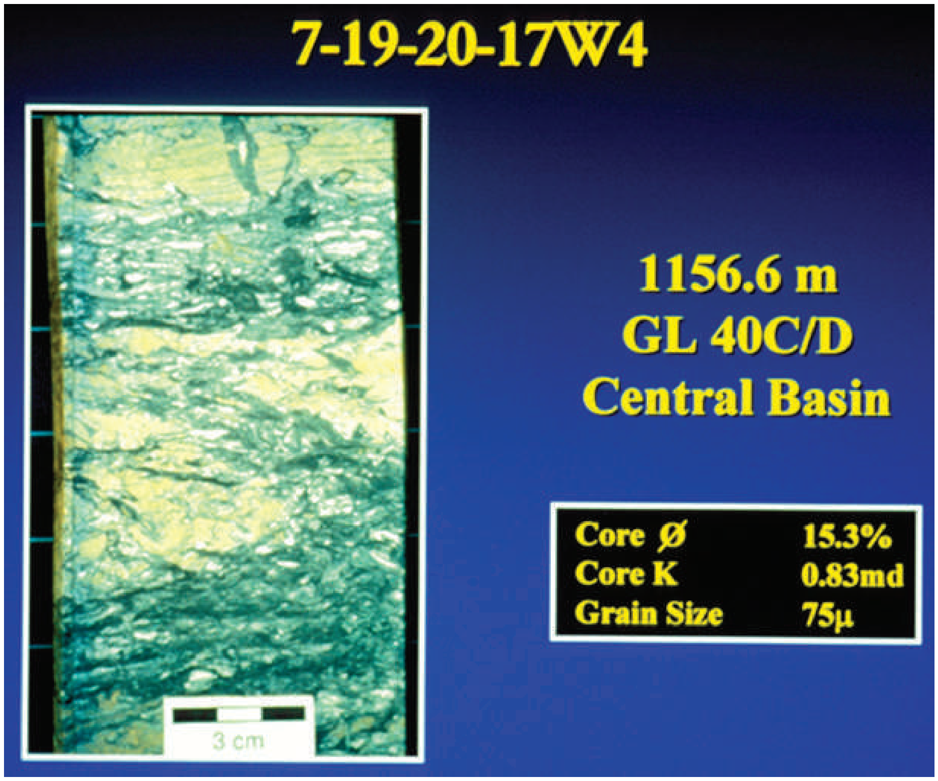


FIG. 34.—An example of bioturbated central-basin facies from the Lathom "A" 7-19-20-17W4 cored well, with its associated grain size, porosity (Ø), and permeability (K) values (from Zaitlin et al., 1998).

complex with its associated estuary is the reason for assigning this deposit to segment 2 of the incised-valley model.

Four main environments can be identified beneath and within the barrier at the mouth of the paleovalley (Figs. 36, 37):

(1) A basal fluvial sandstone to siltstone ~ 5 m thick, organized into repetitive fining-upward cycles of massive to cross-bedded to rippled sandstone with local rootlets and a restricted trace-fossil assemblage consisting of *Paleophycus herberti*, *Conichnus* sp., *Lokeia* sp., and small *Thalassinoides* sp. The sands display excellent reservoir quality but are wet, whereas the siltstone has an effective permeability of < 0.01 md and porosity of < 5%.

(2) An ~ 4-m-thick coal and carbonaceous shale that accumulated in marsh environments.

(3) A bioturbated central-basin to fringing tidal-flat mudstone.

(4) A complex sandbody that consists of upward-coarsening shoreface deposits (effective permeability 2.5 darcys and porosity 27–31%) that are cut by blocky to fining-upward tidal-inlet channels (permeability ~ 3 darcys and porosity 25–30%), with back-barrier flood-tidal deltas on its south side (permeability ~ 2.7 darcys and porosity 29–31%).

The position and preservation of the barrier imply a wave-dominated shoreline deposit at the transgressive limit of the shoreline.

There is marked variation in the production history (Fig. 38) between the subfacies of the barrier because of internal heterogeneity, variation in lateral continuity, and porosity–permeability differences associated with original textural characteristics. The flood-tidal delta, with increased bioturbation and a higher proportion of introduced mud, has the poorest production characteristics, whereas the tidal inlets have the coarsest grain size and the highest initial porosity and permeability, which leads to the most rapid production. Another example of a preserved segment 2 barrier has been documented in outcrop sections from the Paddy Member of the Albian Peace River Formation (Leckie et al., 1990).

Case Study 4: Lower Cretaceous Basal Quartz Sandstone: A Low-Accommodation Compound Incised-Valley Deposit

One of the most complex successions of incised-valley deposits yet described in detail is provided by the Lower Cretaceous Basal Quartz Formation and its equivalents (i.e., the Coverley, Lakota, Cutbank, and Sunburst units) in southern Alberta and northwestern Montana (e.g., Way et al., 1998; Dolson and Piombino, 1994; Lukie et al., 2002; Zaitlin et al., 2002; Leckie et al., 2005). The Basal Quartz (BQ) is a relatively thin unit (typically < 100 m) that was deposited in an accommodation-limited setting and is characterized by multiple, closely spaced unconformities that define a set of more than ten complexly nested incised-valley fills. The BQ was deposited as part of an elongated NNW–SSE trending foreland trough in which there is pronounced isopach thickening toward the northwest. The trough contains three major north-south paleodrainage systems (the Spirit River, Edmonton Channel, and McMurray valleys; Fig. 39). The older and more southerly occurrences provide

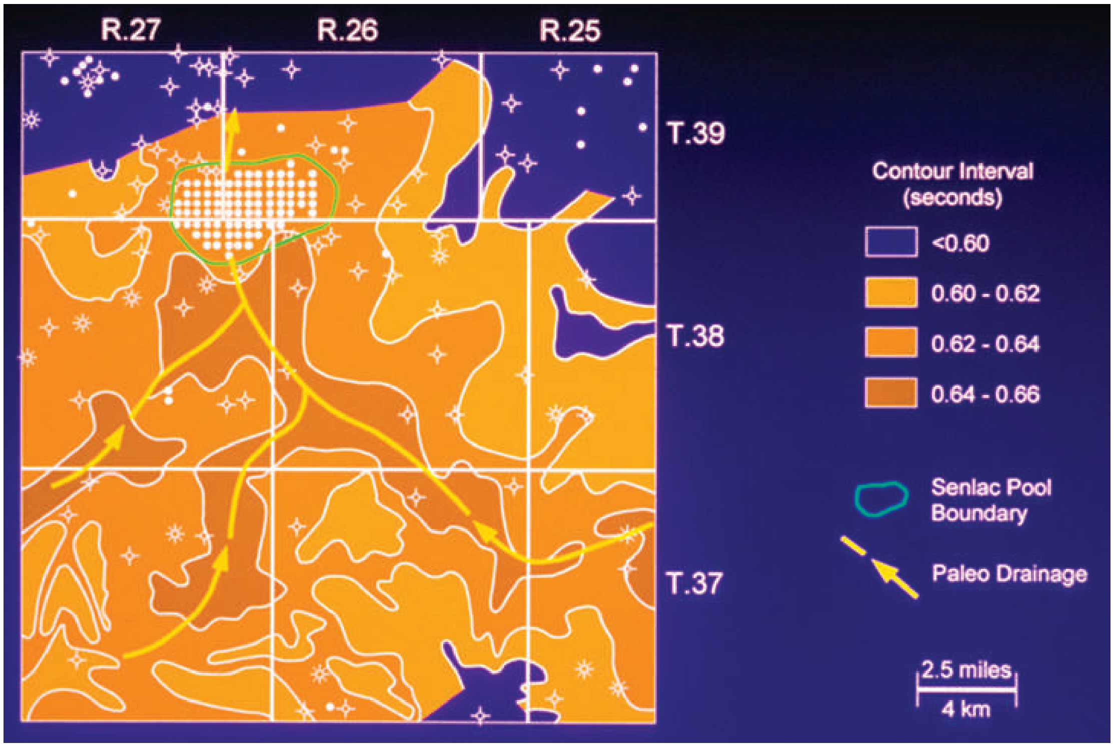


FIG. 35.—Two-way travel time in seconds to the sub-Cretaceous unconformity in the Senlac area of southwestern Saskatchewan, Canada. Darker colors (longer times) indicate areas where the unconformity is deeper. The pattern is interpreted to represent an incised-valley network. Arrows indicate interpreted paleovalley trends and inferred paleodrainage directions. (From Zaitlin and Shultz, 1990.)

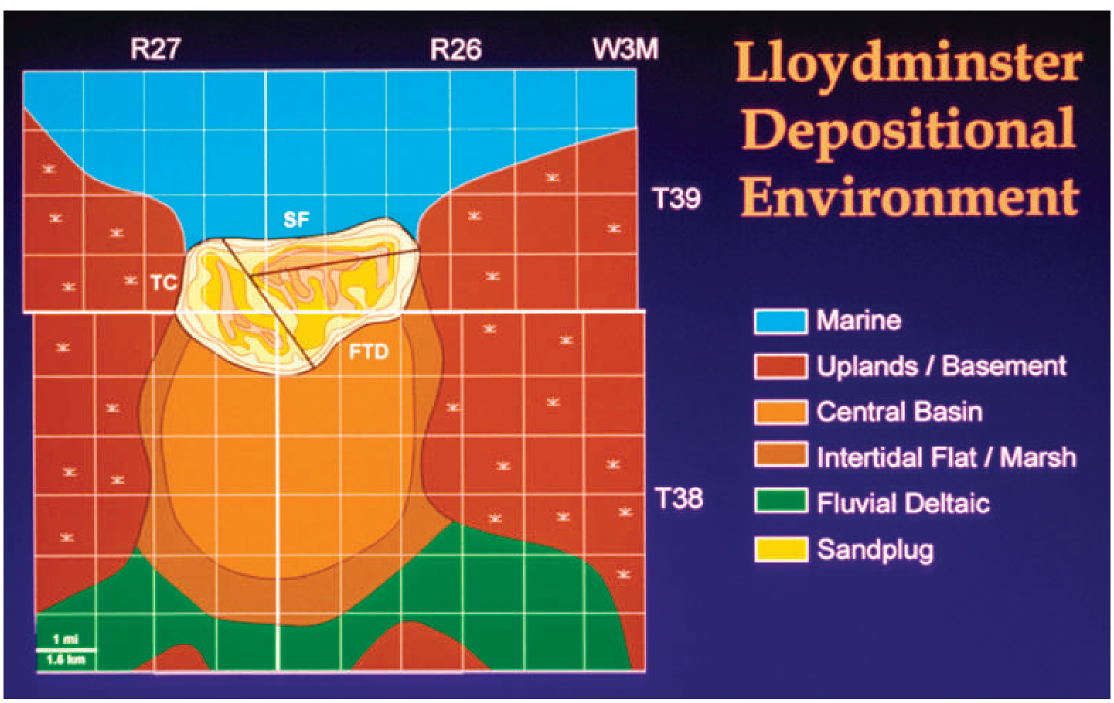


FIG. 36.—Distribution of inferred depositional environments during Lloydminster Formation time in the Senlac incised valley. SF. = shoreface deposits; TC. = tidal-channel deposits; FTD = flood-tidal-delta deposits. Heavy black lines separate depositional environments within the estuary-mouth sand plug. (From Zaitlin and Shultz, 1990.)

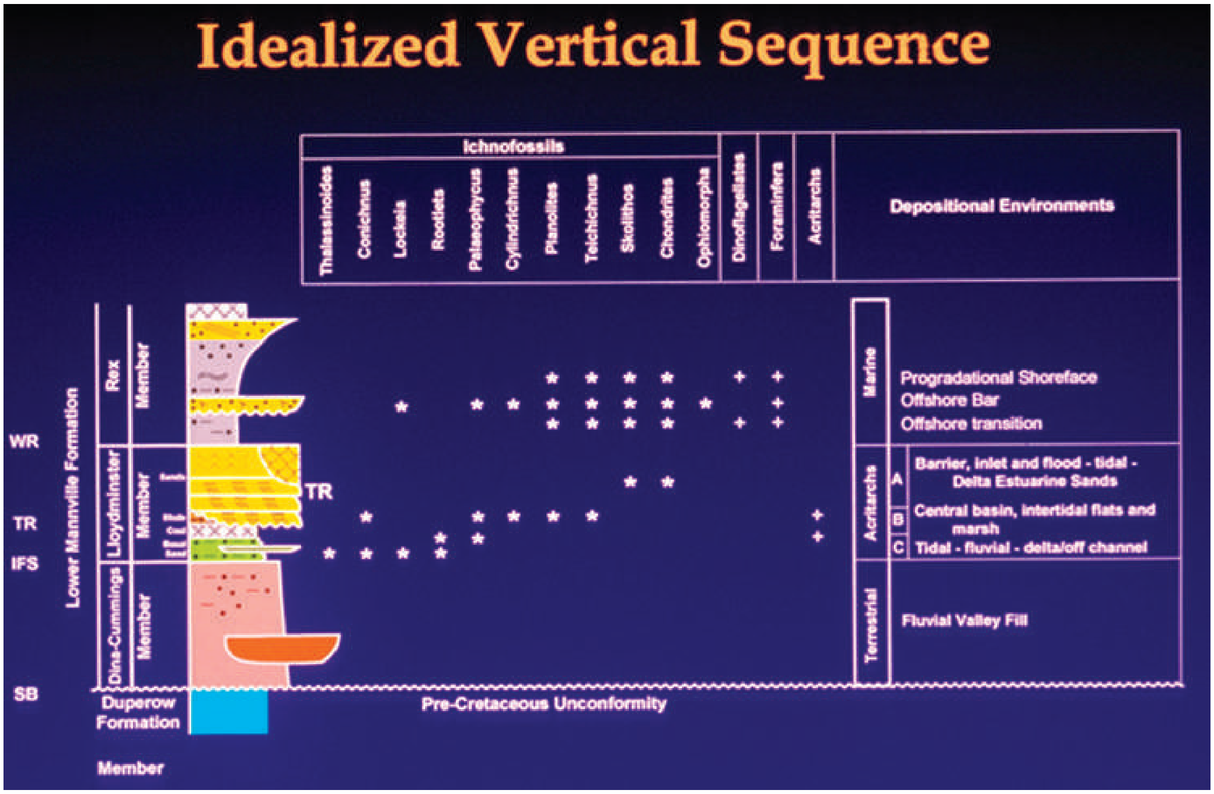


FIG. 37.—Idealized vertical sequence of the Lower Mannville Group in the Senlac area. Ichnofossils identified by Dr. G. Pemberton (University of Alberta); micropaleontological data provided by Robertson Research and Dr. C. Vervoloet. A, B, and C refer to zones in Figure 36: SB = sequence boundary; IFS = initial flooding surface; TR = tidal ravinement surface; WR = wave ravinement surface. (From Zaitlin and Shultz, 1990.)

well-documented examples of segment 3 fluvial deposits that pass northward into segment 1 fluvial, estuarine, and marine deposits.

Within the study area of Zaitlin et al. (2002) (Fig. 39), accommodation ranges between the following two end members:

(1) An area of extremely low accommodation in the southeast corner of Alberta, where isopach values range between 0 and 40 m and net sedimentation rates are less than 2.2 m/My. This area was dominated by long periods of erosion and exposure, the development of paleosols, and polycyclic incision of valley systems characterized by thin, sheet-like, braided to coarse-grained meandering-fluvial deposits.

(2) An area of low–intermediate accommodation in the northwest where thicknesses range between 40 m and more than 200 m and net sedimentation rates ranged between 1.3 and 11.1 m/My, and valley systems are less amalgamated and more easily mappable, with sheet-like fluvial to coarse-grained meandering deposits, paleosols, and thin coals at their bases, changing upward into finer-grained meandering-fluvial to fluvial–estuarine systems.

The transition between these two areas corresponds closely to a geophysically defined ENE-trending structural zone termed the Vulcan Aeromagnetic Low (Ross et al., 1997).

The BQ has an extensive data base of wireline logs, cores, cuttings, and producing pools that allow the succession to be divided into four informal mappable units (A Sandstone, Horsefly, BAT, Ellerslie), each of which can be further subdivided (Zaitlin et al., 2002) (Fig. 40). In particular, the A Sandstone has been divided into the Regional A (oldest), Carmangay, Mesa IV, and Valley and Terrace units. This informal stratigraphic breakdown was later substantiated by chemostratigraphic analysis of the succession (Figs. 42, 43; Ratcliffe et al., 2004). There are two cycles of increasing-upward mineralogical and textural maturity, the first associated with the A Sandstone and the second associated with the Horsefly–BAT–Ellerslie succession. The subdivision of the BQ into discrete valley systems allows recognition of how the paleodrainage changed through time. There is both a progressive spatial and stratigraphic change in valley organization, from thin and wide valley forms in the south and at the base of the maturity cycles, to thicker, narrower, and more deeply cut systems toward the northwest and top of the cycles (Figs. 40, 41). There is also a spatial and temporal change in the development of tributary systems for the Horsefly–BAT–Ellerslie (upper) cycle. The Horsefly Sandstone has few well-developed tributaries, whereas the BAT is characterized by narrow and thin tributaries south of the Vulcan Low, deeply cut complex tributary patterns within the Vulcan Low, and linear deep tributaries north of the Vulcan Low (Ardies et al., 2002; Zaitlin et al., 2002).

The style of depositional fill also changes stratigraphically and spatially, from braided and coarse-grained meandering-fluvial sheet deposits in the Regional A Sandstone, Carmangay, and Horsefly units south of the Vulcan Low and in the low-

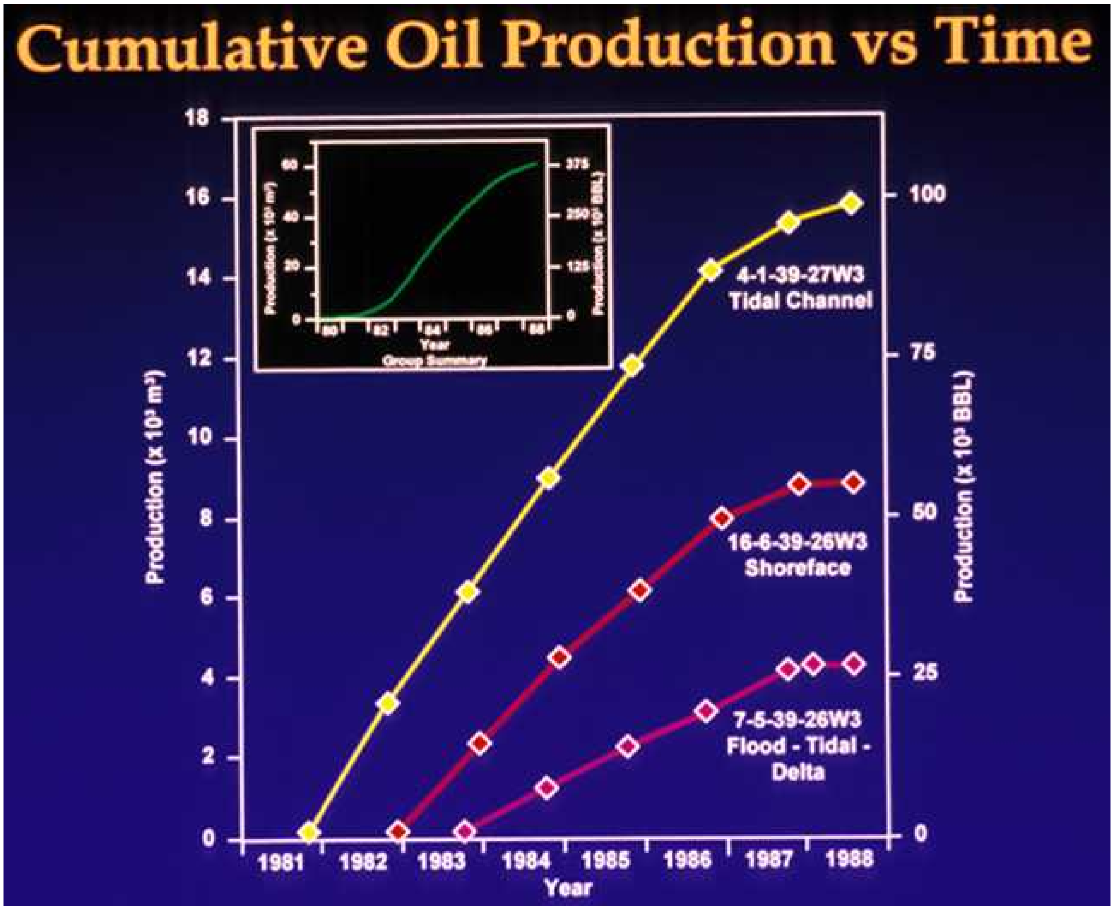


FIG. 38.—Plot of cumulative oil production vs. time for three typical wells completed in the tidal-channel, shoreface and tidal-delta lithofacies. Inset: total cumulative oil production vs. time for the Senlac Pool. (From Zaitlin and Shultz, 1990.)

accommodation portions of the Valley and Terrace and BAT units north of the Vulcan Low, to meandering-fluvial deposits associated with somewhat higher-accommodation Mesa IV, Valley and Terrace, Horsefly, and BAT units, and then to fluvial–estuarine deposits in the portions of the Valley and Terrace, BAT, and Ellerslie units, which accumulated in the highest-accommodation settings north of the Vulcan Low (Figs. 40, 41).

The Carmangay unit (Figs. 40, 41) forms a thin sheet-like sandbody, up to 20 m thick, in the southwest corner of the study area and is interpreted to have accumulated entirely in segment 3. It consists of multiple cycles of erosionally based, fining-upward channel deposits, 1–5 m thick, of medium- to coarse-grained, pebbly, cross-bedded sandstones, fining upward into fine- to medium-grained cross-bedded to rippled, well sorted sandstones. Where preserved, the cycles are locally capped by thin variegated to green waxy paleosols. During Carmangay time, braided-fluvial to coarse-grained meandering-fluvial systems migrated across the depositional surface. The lateral migration of the channels effectively removed most fine-grained overbank deposits and left multiple basal scour surfaces. Reservoir parameters range from < 0.01 md to 4 darcy permeability, with 20% porosity, that yield excellent reservoir quality.

The Mesa IV unit (Figs. 40, 41) also lies entirely in segment 3 and consists of multiple cycles of erosionally based, fining-upward medium- to coarse-grained pebbly cross-bedded quartz and rusty-chert sandstones fining upward into fine- to medium-grained cross-bedded to rippled, well sorted sandstones. These sandstones may be capped by thin, variegated to green waxy paleosols that formed during long periods of exposure. Partial pedogenic clay plugging is pervasive and typically degrades the porosity and permeability of the Mesa IV deposits. The Mesa IV valleys contain narrow, sinuous, ribbon-like channel deposits, less than 15 m thick and 1.6 km wide. Individual channels are difficult to map in the absence of core. The Mesa IV deposits are interpreted to have been formed by coarse-grained meandering-fluvial systems. Locally, the Mesa IV sandstones constitute fair to excellent gas reservoirs with 12–25% porosity and < 0.1 md to 0.8 darcy permeability.

The Valley and Terrace deposits (Figs. 40, 41; cf. Hamilton et al., 2001) consist of braided and coarse-grained meandering-fluvial deposits, grading upward into fluvial and interbedded floodplain and paleosols. Toward the north the Valley and Terrace deposits contain tidal–fluvial channel and estuarine central-basin deposits of segments 1 and 2. As the name indicates, the Valley and Terrace unit consists of a series of nested terraces that formed during repeated periods of base-level fall and subsequent backfilling at a time of overall falling base level, resulting in an architecture that is similar in style to the Quaternary Colorado River (Blum, 1990, 1994; Blum and Valastro, 1994; Blum et al., 1994). Southward-directed transgression of the northern Boreal Seaway during Valley and Terrace time resulted in the backstepping of estuarine deposits over fluvial deposits. Reservoir param-

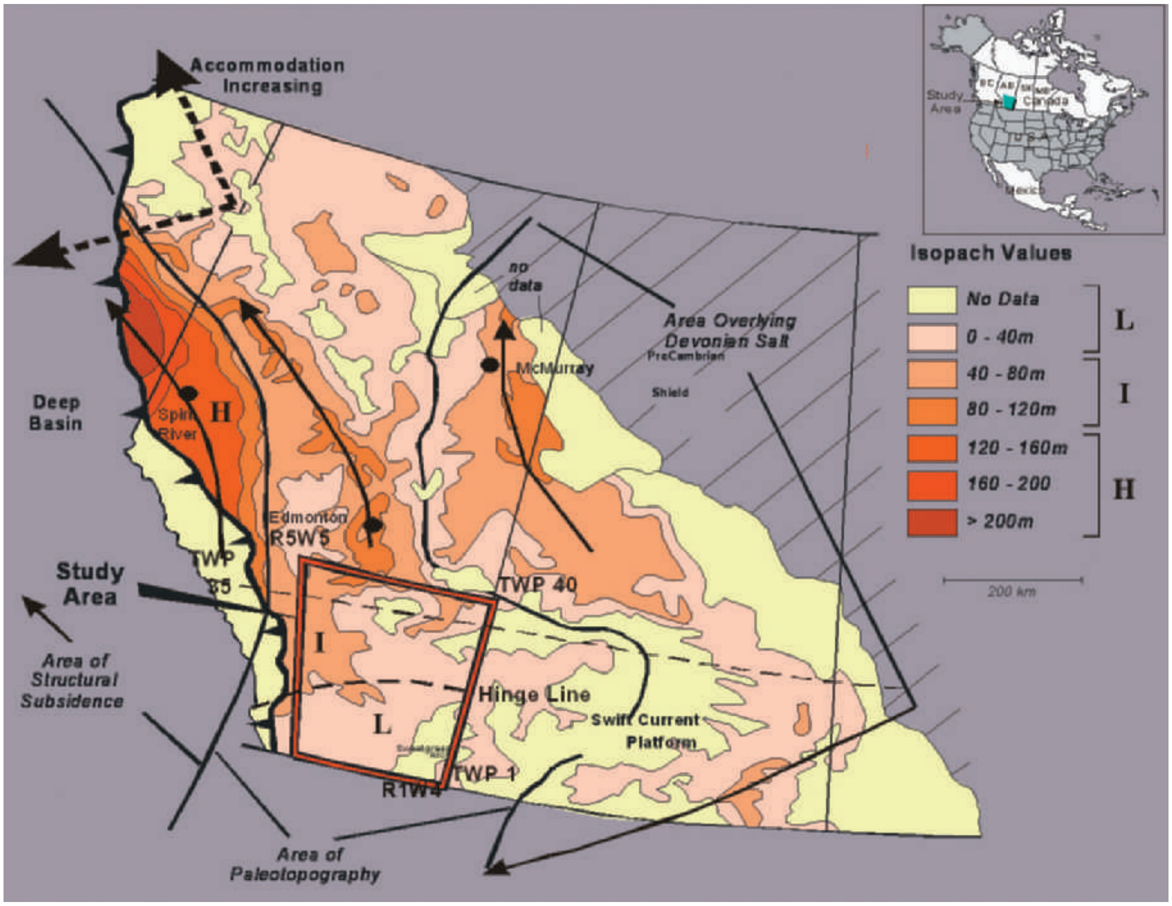


FIG. 39.—Isopach map of the Lower Mannville Group in the Western Canada Sedimentary Basin. The northwestward increase in thickness indicates that accommodation increased in that direction during deposition of the Basal Quartz. L = low-accommodation area; I = intermediate-accommodation area; H = high-accommodation area. Arrows indicate paleodrainages of the McMurray, Edmonton, and Spirit River valley systems. BC = British Columbia; AB = Alberta; SK = Saskatchewan; MB = Manitoba; MT = Montana. (From Zaitlin et al., 2002.)

eters range from 5–28% porosity, and 0.06 md to 1.2 darcy permeability.

The Horsefly unit is confined to two major compound incised-valley systems termed the Whitlash Valley (Hayes, 1986; Hayes et al., 1994) and Taber–Cutbank Valley (Lukie, 1999; Lukie et al., 2002; Arnott et al., 2000, 2002), both of which extend southward into northern Montana (e.g., Dolson and Piombino, 1994), where the Horsefly is termed the Cutbank Sandstone. The Horsefly succession is up to 25 m thick, and the Taber–Cutbank Valley is approximately 50 km wide (Fig. 41). The valley fill consists of repeated fining-upward successions of braided-fluvial to coarse-grained meander sandstones overlain by thick successions of muddy paleosols. The basal strata consist of poorly sorted, matrix-supported conglomerate with a medium- to coarse-grained sandstone matrix. Clasts are subrounded and several decimeters in diameter and are composed of sandstone and silty mudstone. The basal unit is overlain by cross-stratified upper medium- to coarse-grained sandstone, gradationally overlain by massive to small-scale cross-stratified fine-grained sandstone, in turn overlain by siltstone and silty mudstone. The overlying paleosol deposits are composed of variegated red, green, and gray siltstones and mudstones that locally reach 30 m in thickness (Lukie, 1999; Lukie et al., 2002; Arnott et al., 2000, 2002; Zaitlin et al., 2002).

The channel deposits in the Horsefly Sandstone exhibit a classic upward change from amalgamated to isolated (cf. Fig. 21). The cycle begins with a regionally mappable erosional surface that is overlain by amalgamated braided-fluvial sandstones. Any contemporaneous overbank mudstones were completely eroded. These sandstones are then overlain by mudstone-dominated overbank deposits that encase "ribbon" channel and sheet-like crevasse-splay deposits (Arnott et al., 2000, 2002; Lukie et al., 2002; Zaitlin et al., 2002). Two such successions are present within the Horsefly. Each of these sequences accumulated under conditions of continuously increasing accommodation. Tectonic movements, perhaps in response to episodic thrust loading, are thought to have been the major control on accommodation; eustatic fluctuations were probably not important because the study area lay far inland at the time of deposition, landward of the landward limit of estuarine conditions (i.e., in segment 3). Reservoir parameters of the Horsefly unit range from 3–24% porosity and < 0.01 md to > 1.2 darcy permeability.

The BAT can be divided spatially, on the basis of depositional style, into two sub-units (Fig. 41; Zaitlin et al., 2002). The first is a low-accommodation BAT in areas where the total BAT unit isopach is less than 30 m. South of Township 20 along the Taber–Cutbank valley system the width is of the order of 1–5

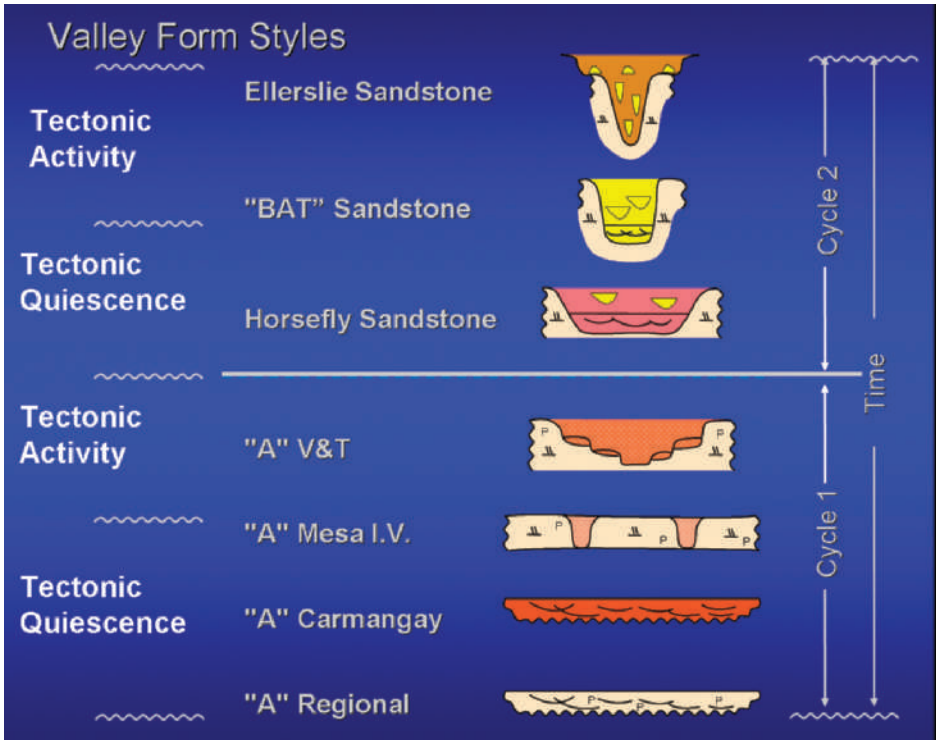


FIG. 40.—Variations in BQ valley form and width:depth ratios. Cycles 1 and 2 relate to tectonic stages of the adjacent Cordillera during accumulation of the Basal Quartz (as defined in Zaitlin et al., 2002). Note how in each cycle the first valleys are broad and relatively shallow, whereas younger valleys have greater depth-to-width ratios.

km. The second is a high-accommodation BAT along the Carseland–Crossfield–Penhold trend, and the Provost trend (where the term Dina is used), where total isopach values can reach up to 100 m and the valley width is approximately 6 to 10 km and the valley-filling deposits are characterized by fluvial–estuarine deposits of segment 1. In low-accommodation areas, the BAT consists of stacked, erosionally based, fining-upward sheet-like sandstones (Ardies, 1999; Ardies et al., 2002; Arnott et al., 2000, 2002; Zaitlin et al., 2002). Each succession grades upward from coarse to medium sandstone, to lower medium–upper fine sandstone. All of these sandstones are pervasively cross stratified and are interpreted to have accumulated in braided to coarse-grained meandering rivers. Very rarely does the low-accommodation BAT display any form of marine bioturbation. The BAT sandstones display excellent reservoir quality and are a prime exploration target. In low-accommodation BAT reservoirs in the southern and eastern portions of the study area, reservoir parameters range from 3–28% porosity and < 0.01 md to 5 darcy permeability.

CRITICISMS, MISUSES, AND REFINEMENTS OF THE E&IV MODEL

The E&IV facies model detailed above (Dalrymple et al., 1992; Zaitlin et al., 1994) has gained widespread usage (Fig. 3) and acceptance over the past decade and could now be regarded as a mature and established model. However, like all facies models that are necessarily based on a "distillation" of natural variability (Walker, 1984b), it does represent a simplification of natural complexity and cannot be expected to match every specific example, whether modern or ancient. As a result, it is perhaps natural that there have been suggestions that the model fails to take into account important variables and thus does not accurately reflect certain aspects of estuarine and incised-valley deposits. There have also been attempts to develop refinements and/or elaborations of the model, in the same way that the models for a meandering-river point bar have multiplied from the single vertical succession proposed by Allen (1963) to the 16 successions shown by Miall (1996). In addition, there has been inadvertent misuse of the model by some workers. Here we examine some of the issues raised by these developments, because they illustrate useful information about incised-valley estuarine systems or about the nature of facies models in general.

Estuary Versus Estuary: The Implications of Applying a Name

One of the most fundamental problems with estuarine facies models has been the ongoing confusion between the oceanographic, salinity-based definition of estuaries (Pritchard, 1967) and the modified geologic definition of Dalrymple et al. (1992) used here. This, in turn, has led to the potential for inaccurate interpretations of ancient successions and/or to suggestions that one or other of the definitions is inappropriate. At the outset, it must be recognized that both definitions are "valid" in their own right. The problem arises through failure to carefully articulate which definition is being used and/or to implicitly switch between definitions without saying so.

The most common expression of this problem is the growing tendency to deduce that certain ancient deposits accumulated in an area of brackish water, on the basis of the nature of the trace-fossil assemblage as described below. From this, the authors state

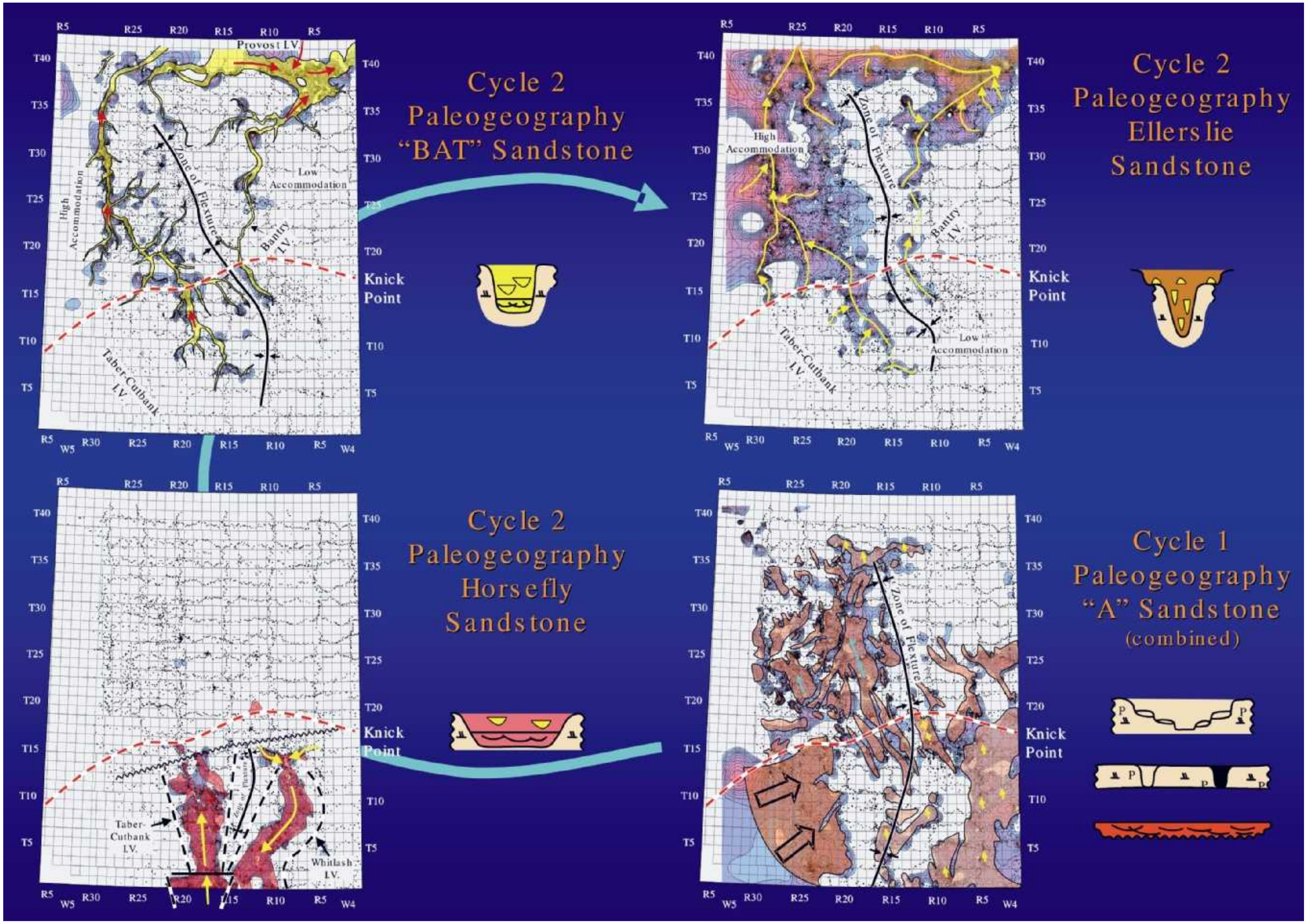


FIG. 41.—Cycle 1 (lower right): Composite isopach and paleogeographic map of the A Sandstone. Contour interval = 5 m, values range from 0 to 30 m. Cycle 2 (lower left) Isopach and paleogeographic map for the Horsefly unit. Cycle 2 (upper left): Isopach and paleogeographic map of the BAT unit. Cycle 2 (upper right): Isopach and paleogeographic map of the Ellerslie unit. In all panels, arrows represent inferred paleodrainage direction. Where no contours are present, no deposits are present, due either to nondeposition or to subsequent erosion. Original from Zaitlin et al. (2002).

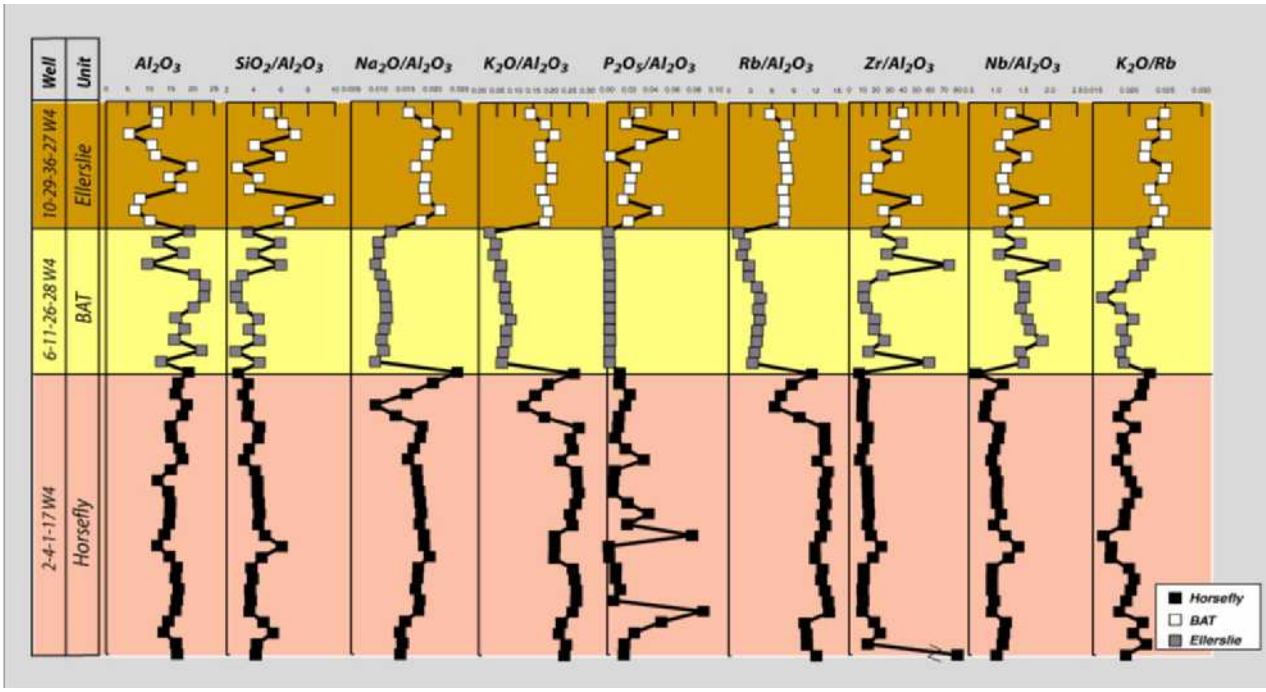


FIG. 42.—Vertical changes in the geochemistry of silty claystone in the Horsefly, BAT, and Ellerslie units of the Basal Quartz Formation (Zaitlin et al., 2002; Ratcliffe et al., 2004). The data come from several cores, with samples placed in their correct, relative stratigraphic position. Al_2O_3 and SiO_2 values demonstrate that there are only minor differences in the silt and clay content of the various units. However, to minimize the influences of subtle changes in silt content, the values for the other elements have been normalized against Al_2O_3.

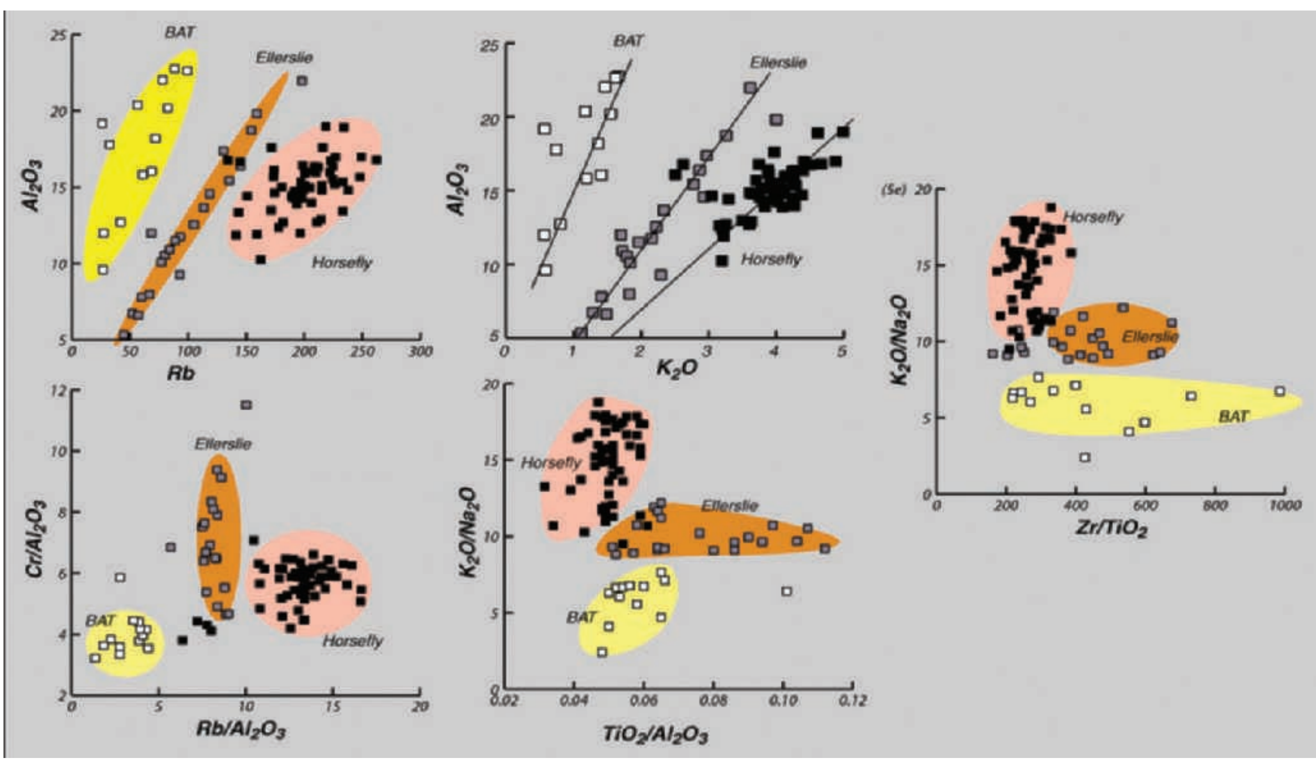


FIG. 43.—Cross plots of normalized elemental ratios to illustrate differentiation of the Horsefly, BAT, and Ellerslie units (Basal Quartz Formation) using geochemical data.

that the deposits are "estuarine", implicitly utilizing the salinity-based definition of Pritchard (1967). Then, on the basis of this estuarine interpretation, the deposits are said to be transgressive and/or to demonstrate the existence of an incised valley, which implicitly represents a switch to the Dalrymple et al. (1992) geological definition. Alternatively, the authors might demonstrate that the succession is, in fact, regressive and go on to suggest that progradational estuaries exist, in contravention of the Dalrymple et al. (1992) definition.

Such switching between the two definitions of estuary is inappropriate because, as stated above, the salinity-based definition includes a much broader range of environments than the geological definition: although the two definitions overlap in their application, they are not equivalent. It is certainly the case that estuaries (*sensu* Dalrymple et al., 1992) may have a phase of progradational filling at the end of the transgression, when the coastal zone switches from transgression to regression. However, this progradational phase must overlie a transgressive succession. Furthermore, the application of the term "estuary" *sensu* Dalrymple et al. (1992), in combination with the (now modified) idea that such estuaries are restricted to incised valleys, also implies that there was a relative sea-level lowstand and the development of a sequence boundary at the base of the valley (cf. Hein and Langenberg, 2003). This, in turn, has important implications for our understanding of the geological history of the area and for the prediction of petroleum-reservoir play types (e.g., lowstand deltas). However, brackish-water trace-fossil assemblages can occur in progradational deltaic settings and even in some shelf environments. It may be, therefore, that the inappropriate switching between the two definitions of estuary has led to the misidentification of deltaic distributaries as estuaries and the incorrect sequence-stratigraphic interpretation of some successions (cf. Reinson and Meloche, 2002; Zaitlin, 2003; Krystinik and Leckie, 2005).

It should be noted that the definition of estuary presented in this paper is a modification of that presented in Dalrymple et al. (1992). Since the original definition was constructed it has become apparent that there are numerous settings such as abandoned deltas and structural embayments that possess the characteristics of estuaries but are not necessarily associated with paleovalleys. While the origin and classification of these types of settings are usually apparent in modern environments, it is much more difficult to discern them in ancient sediments. Hence, while it might be preferable to identify an abandoned delta as such in an ancient deposit, it might not always be possible to do so, in which case the use of the term estuary would be justified if it met the criteria identified in this paper.

The types of problems that may result from switching between the two definitions of estuary represent inappropriate use of the models rather than deficiencies in the definitions. Both definitions of estuary have their use, but they should not be confused. We suggest that, if the Dalrymple et al. (1992) definition is to be used at any point in a study, the salinity-based definition be avoided. Instead, we recommend the use of the term "brackish-water" as the more acceptable term (in place of "estuarine") for deposits believed to have accumulated in an area of reduced salinity. Conversely, if the decision is to use the salinity-based definition, then the Dalrymple et al. (1992) definition should be avoided and the term "transgressive" should be used for retrogradationally stacked facies successions.

Classification of Estuaries

Several authors, beginning with Cooper (1988), have suggested that a third type of estuary (fluvially dominated) should be added to the two-fold wave- and tide-dominated subdivision proposed by Dalrymple et al. (1992). Such a proposal would seem reasonable by analogy with the three-fold subdivision of delta facies models (Coleman and Wright, 1975; Galloway, 1975). It is certainly the case that there is a wide range in the size of rivers feeding estuaries. However, this proposal for a river-dominated class of estuary has weaknesses for three reasons:

(1) In the specific instance described by Cooper (1988), the short-term and long-term behavior of the system was not adequately taken into consideration. Because of the extreme variability of discharge in that situation, the river-mouth area alternated between two conditions: at the time of the infrequent but very large river floods, sand was exported beyond the mouth of the river to the marine environment, whereas, during the much longer, intervening periods, the river-mouth area was refilled by sand carried to the area by river and flood-tidal processes. During the times when sediment was being imported, the system was a wave-dominated estuary (*sensu* Dalrymple et al., 1992) with a barrier, flood-tidal delta, and central basin). In the longer term, however, sediment was being supplied by the river to a beach and shoreface system. As a result, in the longer term the system described by Cooper (1988) is not an estuary but is a river feeding an incipient strandplain.

(2) In the more general sense, one of the most important, even defining, characteristics of estuaries (*sensu* Dalrymple et al., 1992) is the existence of two sediment sources: fluvial and marine. In the limiting cases where one or other of these two sediment sources goes to zero, it is legitimate to argue that the systems are no longer estuaries in the original sense. Therefore, systems with only a marine sediment source and no river influence might legitimately be considered barrier–lagoon systems that are gradational with estuaries (cf. Boyd et al., 1992). Such systems, in our opinion, form exclusively in transgressive situations. Systems with negligible marine sediment input (i.e., they are "river-dominated") are, by contrast, almost certainly regressive, at least locally at the river mouth, at the time of consideration. Therefore, they fail to fulfill one of the fundamental criteria of "estuary" (*sensu* Dalrymple et al., 1992). There is no good, existing term for a semi-enclosed coastal area with no marine input that might otherwise be called river-dominated. One possibility would be to call such systems "embayments", as is commonly done in the coastal geomorphological literature (e.g., an open-mouthed bay with no bay-mouth barrier or other marine-sourced sediment body, but with river input at its head). Therefore, given the essential character of estuaries as proposed by Dalrymple et al. (1992) a prograding river-dominated system cannot be an "estuary".

(3) A careful review of modern river-mouth areas (cf. Dalrymple et al., 1992) indicates that the size of the river does not fundamentally change the geomorphic character of the estuarine system. Therefore, valley mouths that have unfilled accommodation (i.e., they are estuaries *sensu* Dalrymple et al., 1992) have similar morphologies regardless of whether the river is small or large. For example, the Severn River (England) and Salmon River (Cobequid Bay, Bay of Fundy) tide-dominated estuaries have essentially identical morphological and facies zonations despite the fact that the water and sediment discharges of the Severn River are several orders of magnitude larger than those of the Salmon River. Similarly, the fundamental morphology of the large Mobile Bay estuary (Kindinger et al., 1994) is identical to that of the small

Narrawallee and Wapengo estuaries of southern New South Wales (with coast-parallel barrier, low-energy and muddy central basin and bayhead delta; Nichol, 1991).

Thus, the creation of a river-dominated class of estuaries would appear, at least at this time, to be unnecessary. People working in the ancient rock record who have adopted this concept may have fallen victim to the inadvertent mixing of estuary definitions discussed in the preceding section.

Systems-Tract Assignment of Valley Fills

Some confusion exists regarding the assignment of incised-valley fills to individual systems tracts. The original work on incised-valley deposits by the Exxon group (e.g., Van Wagoner et al., 1988, Posamentier and Vail 1988) considered all of the deposits within an incised valley to belong to the LST. In this context, this was reasonable because they were dealing with relatively low-resolution seismic data and large-scale stratigraphic sequences of second or third order. In this context, the fill of the valley could not be subdivided in detail and the valley-fill succession represented a very small volume at the base of the much larger sequence. By contrast, detailed examination of both modern and ancient valley-fill successions (e.g., Roy, 1984; Reinson, 1992; Boyd and Honig, 1992; MacEachern and Pemberton, 1994; Demarest and Kraft, 1987) show clearly that a significant fraction of the valley-filling deposits in many systems was deposited during base-level rise, commonly at a time when the shoreline had migrated substantial distances landward of its lowstand location. Incised-valley estuaries along modern coastlines illustrate this point: valley filling continues at a relative highstand of sea level. As a result, many, but not all, workers have tended to recognize both LST and TST deposits within incised valleys, with TST deposits predominating throughout most of the length of the valley.

Although this situation is perhaps the "norm" (*sensu* Walker, 1992), valleys, or portions of valleys, that are filled entirely during the lowstand are a possibility. In particular, this may occur for a distance landward of the lowstand shoreline, with the valley fill consisting of fluvial deposits that accumulated during the fluvial aggradation that accompanied sea-level rise during the late LST and earliest TST. High rates of sediment supply at lowstand would favor valley filling at this time. However, the inland extent of this lowstand fluvial aggradation would be limited to the area where the "backwater" effect exists (a few kilometers to several tens of kilometers at most; e.g., Blum and Törnqvist, 2000, and references therein) during the late lowstand. At the same time, areas farther landward in the valley would be bypass zones with little or no net deposition. As the lowstand shoreline experienced initial transgression at the onset of the RSL rise, the transgressive surface would be formed. This surface, where it is possible to recognize it, would onlap into the valley. Landward of the point of onlap of the transgressive surface, the valley fill would consist of a thin LST (possibly only one channel depth thick in many cases) consisting of relatively coarse-grained fluvial deposits, overlain by finer-grained fluvial sediments of the TST.

Both systems-tract assignments of valley-fill deposits are probably valid, but at very different scales of consideration. The early Exxonian view that all valley-fill deposits are LST should be used only at very large spatial and temporal scales, whereas a more refined subdivision into LST and TST is more likely to be correct in high-resolution studies. In our opinion, an example of what can happen by an inappropriate use of the Exxonian view in a high-resolution study is provided by Bowen and Weimer (1997, 2003). In these papers, the authors use the Exxonian approach without clearly explaining why. They then proceed to document the nature of the valley fill in detail and show tens of kilometers of backstepping of facies, which clearly lie within the TST as defined by most workers, but which they say forms part of the LST. Such inconsistent use of terminology is confusing at best and deviates from the original intent of systems tracts.

Relative Abundance of Facies and Systems Tracts within Incised Valleys

The original model for incised valleys (Figs. 19, 20; Zaitlin et al., 1994) shows fluvial deposits as constituting a very small proportion of the entire valley fill, which was dominated by estuarine facies. As a result, the TST was volumetrically predominant, with minimal LST. While these authors explicitly said that the relative proportion of fluvial (and LST) deposits was subject to considerable variability, some subsequent workers have criticized the model, suggesting that this is not a universal aspect of incised-valley successions.

Such criticisms may have some validity, but they fail to recognize the nature and role of facies models. As already stated, facies models represent a distillation of existing knowledge and are not intended to illustrate the only possible stratigraphic expression. Variability is to be expected, and deviations from the model can be used to deduce important information about the situation under study. For example, the complete absence of fluvial deposits and the presence of tidally influenced deposits right to the base of the valley may indicate either (1) that the erosional feature is not a valley but instead represents a tidally scoured depression that may not correlate to a sequence boundary or (2) that deposition within a valley took place near the lowstand river mouth in a tidal–fluvial environment. On the other hand, a valley filled entirely with fluvial deposits indicates that the rate of fluvial sediment supply was high relative to the rate of creation of accommodation by sea-level rise, or that the location in question lay sufficiently far inland that estuarine conditions never reached there (i.e., the valley lies within segment 3). In retrospect, the original Zaitlin et al. (1994) representation with minimal fluvial and LST deposits may have been unduly influenced by the then predominance of systems in which there was a relatively small fluvial sediment input and of modern systems in which the rate of RSL rise was so rapid that minimal fluvial–LST deposition occurred, especially in the inner part of segment 1 and in segment 2 (e.g., incised-valley systems along the US east coast such as described in Ashley and Sheridan, 1994). A better "distillation" might well have included more fluvial sediment as the "norm".

Additional Critiques of Estuarine and Incised-Valley Models

Other discussions of the E&IV models have been published by Washington and Chisick (1994) and Blum and Törnqvist (2000). Washington and Chisick (1994) suggested that several factors were missing from the estuary model of Dalrymple et al. (1992). They identified the *in situ* production and accumulation of biogenic material (peat and carbonate), the rate of sea-level rise relative to the rate of marine sediment input, and climate (temperate versus tropical) as factors that should have been included. In response we note that no generalized model can include all factors that are present in a depositional sedimentary environment. The full range of boundary conditions and processes in an environment determines the spectrum of deposits that may be produced by that environment; however, only the commonly occurring combinations will be useful for a widely applicable model. Hence, while the three factors identified by Washington and Chisick (1994) may be important in local examples, the lack of explicit inclusion of them in the Dalrymple et al. (1992) model illustrates the distillation process identified by Walker (1992), by which variability is removed and generalized facies models are produced. In the case of the three factors above, they are not included in the general model because: (1) they do not control the basic geomorphic organization of estuarine facies; (2) their influence is less pervasive or less intense than that of the fundamental interaction of fluvial and marine processes; and / or (3) the nature or distribution of their influence is controlled by the fluvial–marine interaction in an estuary (i.e., the latter factor is more fundamental; cf. Dalrymple et al., 1994a). As estuarine facies models become progressively more refined, however, future workers might well wish to create a "new" facies model (i.e., a variant on the models proposed by Dalrymple et al., 1992) to explicitly incorporate the distribution of carbonate facies in tropical estuaries with low fluvial influence.

More recently, Blum and Törnqvist (2000) have criticized how some workers have used the incised-valley concept because it implies a "vacuum cleaner" approach to fluvial sediment transport rather than a "conveyor belt" approach. Blum and Törnqvist (2000) have disputed the influence of relative sea-level fall as the initiator of incision, accompanied by "an upstream-propagating wave of stream rejuvenation, which produces sediments that entirely bypass the coastal plain and newly emergent shelf to provide a critical volume of sediment for systems tracts further basinward" (the vacuum-cleaner model that results in an incised valley). This is contrasted with the conveyor-belt model, "where sediments are continuously delivered to the basin margin from a large inland drainage". Instead, they suggest that it is the climatically produced changes in discharge that drive incision. While this may be true in many cases and is not explicitly considered in many discussions of valley formation, it is hard to neglect the role of relative sea level (RSL) fall as a trigger for valley formation, because a fall in RSL may cause the river to encounter new areas of steep gradient on the continental shelf that promote incision. Although the impact of these new gradients is not felt throughout the drainage basin, incision is present on many rivers 40–400 km upstream of the present shoreline (data of Blum and Törnqvist, 2000). Indeed, the very abundance of Holocene incised valleys containing estuaries on many coastal streams near the highstand shoreline points to the strong influence of RSL change on their development. So, although much sediment is transported through alluvial valleys in response to climatic forcing during times of sea-level fall and steeper shelf gradients, some sediment is also removed from the coastal plain, generating a container (the valley) for later filling. This line of argument highlights one of the new features to emerge since the development of E&IV models, namely the recognition that valley incision may take place only at localized changes in gradient where knickpoints can be created. Hence, although full cross-shelf incision may occur when the shoreline drops below the shelf break (e.g., Suter et al., 1987; Porębski and Steel, 2003), a more common situation results from sea-level change that exposes a local gradient increase at an old shoreface. This may cause incision at several localized sites while the greater part of the exposed continental shelf and the upstream alluvial channel remain unincised (e.g., Woolfe et al., 1998; Posamentier, 2001; Fielding et al., 2003; Wellner and Bartek, 2003).

AN EXAMPLE OF FACIES-MODEL USAGE: THE TIDE-DOMINATED DELTA VERSUS ESTUARY CONTROVERSY

At a recent SEPM research conference (Dalrymple, 2003) it was suggested by some participants that any brackish-water, tidally influenced facies should be considered an estuarine deposit, a suggestion that implicitly follows the Pritchard (1967) definition of an estuary. It was also suggested that many deltaic deposits had been incorrectly identified as estuaries because of the recent popularity of E&IV models (e.g., Reinson and Meloche, 2002; Leckie and Krystinik, 2005). This illustrates the need for practical and accurate facies models, because brackish-water tidal facies actually occur in several distinct environments, and because deltas should not be confused with estuaries. By developing clear facies models based on distinctive combinations of sedimentary processes it is possible to correctly identify and differentiate these environments.

While it was noted in an earlier section of this paper that each facies model necessarily is a simplification of a wide spectrum of similar environments, there should be fundamental differences in facies models from different depositional environments. So facies models from estuaries, tide-dominated deltas, lagoons, and tidally influenced shelves should not be the same. For example, the differentiation of tide-dominated deltas from tide-dominated estuaries provides a convincing argument for the clear establishment of facies models for each setting and their appropriate use, and it represents an important example of the value of the facies-model concept.

The delta-versus-estuary problem was formally raised by Walker (1992), who suggested that the triangular classification of deltas (e.g., Galloway, 1975) was inappropriate and that it should be modified or abandoned. Walker's (1992) emphasis on sea-level change and the presence of a coastal protuberance (i.e., a bulge) as a distinguishing feature of deltas led him to believe that tide-dominated deltas, which commonly occur at the heads of embayments, were not related to other deltas and were better considered as tidal estuaries. However, the problem results from a fundamental confusion of the factors that make up the essence of facies models for deltas and estuaries.

Many of the detailed features of tide-dominated deltas and tide-dominated estuaries are certainly similar. For example, both of them contain brackish water and hence restricted faunal and ichnological assemblages. They contain very similar physical sedimentary structures (e.g., tidal bundles, inclined heterolithic stratification, and all other tidal indicators listed in recognition criteria 8 and 9 above), as well as similar depositional sub-environments and facies (e.g., tidal–fluvial channels and elongate tidal sand bars). However, that is as far as the similarity goes. There are fundamental differences that distinguish the two depositional environments and their facies models.

(1) Estuaries are commonly associated with incised valleys while deltas are not. However, early highstand progradation of some deltas may be restricted to incised-valley settings, while some abandoned deltas that are not incised take on an estuarine character during transgression.

(2) Estuaries display a tributary pattern (see above) while deltas display a distributary pattern.

(3) Deltas have only one sediment source and hence single composition, while estuaries have two.

(4) Deltaic sands fine unidirectionally seaward while estuaries show a grain-size peak at either end of the system, reflecting the two sediment sources.

(5) Deltas are fundamentally regressive systems while estuaries are transgressive.

(6) Because deltas are regressive in nature, their stratigraphy differs fundamentally from transgressive estuaries. In deltas, marine sand bars are underlain by prodelta and marine sediments. In estuaries, marine sand bars are underlain by a tidal ravinement surface and more landward estuarine and fluvial facies (Figs. 16, 17, 19, 20).

(7) The prodelta environment is missing in estuaries.

(8) The estuary typically lies on a regional unconformity or on fluvial deposits , which in turn lies on an unconformity. It has a maximum flooding surface located within or above the estuarine fill. A highstand delta typically lies above a maximum flooding surface and has a sequence boundary developed above it (Fig. 19).

(9) In sequence-stratigraphic terms, estuaries more commonly occupy the transgressive systems tract while deltas more commonly occupy the highstand systems tract (although it is recognized that these depositional systems can occur in a range of systems tracts, especially when considering lower-order sequences).

So, while many aspects of tide-dominated estuaries and deltas look superficially similar, they should not share the same facies model. When the correct identification of estuarine and deltaic deposits in their appropriate stratigraphic context is made, it is clear from the nine issues listed above that there are fundamental differences in the two facies models. Our conclusion is that the "offending corner of the delta triangle" that was removed by Walker (1992, his Figure 7) should be firmly reaffixed. In addition, it should be placed correctly in the triangular coastal classification of Boyd et al. (1992) and separated as shown in Figure 9 from tide-dominated estuaries. The reasoning behind this return to the triangular classification is the contrasting processes that distinguish deltas (e.g., Wright, 1985) from estuaries. Chief among these is the balance between sediment flux and relative sea-level rise. In deltas, over a longer term, the sediment flux outstrips any change in relative sea level, while in estuaries the reverse is true. In deltas, the fluvial processes delivering sediment to the coastline overwhelm the marine processes because there is no available onland accommodation, and they result in a unidirectional seaward flux of sediment. In estuaries, because there is unfilled accommodation within the drowned coastal zone, wave and tidal processes produce a landward sediment flux from the marine end of the system that supplements that from the fluvial end. In addition, the geometry of a delta tends to favor ebb-tidal dominance while that of an estuary tends to favor flood-tidal dominance (cf. Friedrichs and Aubrey, 1988).

RECENT AND FUTURE DEVELOPMENT OF ESTUARINE AND INCISED-VALLEY FACIES MODELS

In this section we first look at the general concept of scientific models, to identify the current state of evolution of facies models. We then examine some specific advances in the field of E&IV models and look forward to the approach of the future.

Development of Scientific Models

Goodwin (1999) provides an insight into the evolutionary stages in the development of a scientific field such as sedimentology (Fig. 44). He identifies an early observation stage that is then followed by a need for classification of the observations. Both of these stages occur early in the development of a discipline. As the field advances, however, classification gives way to the development of empirically based laws and finally to theoretical understanding (Hempel, 1965).

The field of facies models is still a relatively young field with a history of less than forty years. Hence, we are in the early stages of its development, in which we have made a large number of observations in the form of surveys and process measurements in modern environments, outcrop studies, wireline-log, core and borehole studies, and remote-sensing studies (e.g., seismic, radar). These observations have been incorporated into depositional facies models since the middle of the twentieth century in what is essentially a form of classification. An approach of this kind describes the delta and coastal classification triangles presented earlier.

Inherent in this approach is an organization of the processes that control deposition and hence involves some understanding of the relationships between the controlling parameters (in these cases, for example, waves, tides, and rivers). Therefore, our scientific field is at the point of transition to the next stage , which involves empirical approaches and finally theoretical approaches to understanding.

In the E&IV field, empirical laws have been developed and applied, for example, to paleohydraulics (Miall, 1996), simulation of alluvial stratigraphy (Bridge and Leeder, 1979), the influence of relative sea level on river incision (Wood et al., 1993), the continent-wide quantitative classification of coastal systems based on physical processes (Harris et al., 2002) and the preservation of estuarine strata after shoreface erosion (Cowell et al., 1999). All of these examples and many others have begun to take a quantitative approach to sedimentation problems with the ultimate aim of achieving a theoretical understanding. We believe that the quantitative approach to sediment modeling is the best way to advance our field. Our current stage of development is the formulation and application of facies models, with a resulting proliferation of these models. The way to avoid becoming bogged down in this classification stage, as also occurred, for example, in the study of cyclothems (e.g., Wanless and Weller, 1932) or geosynclines (e.g., Kay, 1951) in the earlier twentieth century, is to employ a quantitative approach to determine the predictive relationships governing the sedimentary processes. An approach of this kind represents a way forward (most likely through the techniques of computer modeling; see section below) that will provide a better ability to predict facies relationships.

Brackish Ichnology

Because estuaries, like other river-mouth coastal environments, are characterized by brackish-water conditions, the development of techniques to identify brackish-water deposits using trace fossils has greatly assisted the recognition of estuarine deposits. In many deposits, distinctive body fossils are either lacking or poorly preserved, whereas trace fossils are abundant and preserved *in situ*. The development of brackish-water ichnology is a relatively recent field, with early work in the 1980s (e.g., Wightman et al., 1987) and first-generation summaries published in the 1990s (MacEachern and Pemberton, 1994; MacEachern in Zaitlin et al., 1995). More recent reviews are provided by Pemberton et al. (2004) and Buatois et al. (2005).

Recent research (e.g., MacEachern and Pemberton, 1994) has shown a distinctive assemblage of trace fossils for brackish-water settings that contrasts strongly with surrounding terrestrial or fully marine trace-fossil suites (Fig. 45). Additional work by

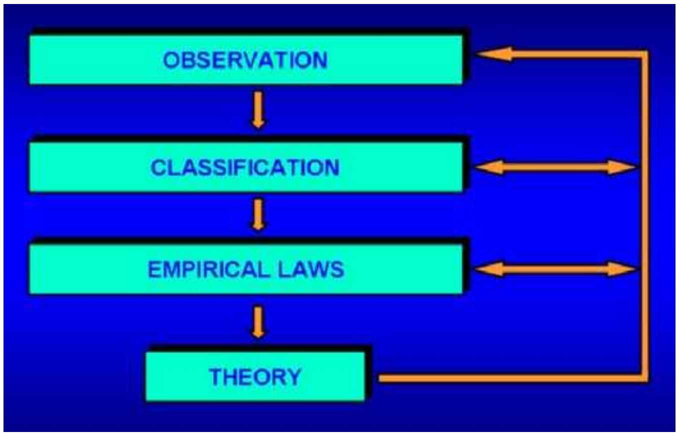


FIG. 44.—Evolutionary stages in the development of a scientific field (after Goodwin, 1999).

Buatois et al. (1997) has shown that terrestrial trace-fossil assemblages in tidal rhythmites can be used to locate the innermost tidally influenced freshwater zone of an estuary (see Fig. 10) The trace-fossil suite of brackish-water environments is characterized (MacEachern and Pemberton, 1994) by "a variable and sporadic distribution of burrowing, variability in ichnogenera distribution, and dominance by simple structures of trophic generalists. The suite is dominated by opportunistic suites characteristic of stressed environments, particularly those subjected to fluctuations in salinity, episodic deposition, variable aggradation rates, and variability in substrate consistency."

Recognition of these ichnological characteristics in combination with the other criteria for distinguishing E&IV systems given above provides a strong basis for identifying E&IV systems, even where they exhibit a mud-on-mud or a sand-on-sand contact with the deposits of other environments. In addition, careful documentation of ichnofacies assemblages may enable an internal subdivision of estuarine depositional settings into bayhead delta, central basin, and barrier components (Fig. 46) on the basis of a longitudinal gradient of salinity (from nearly freshwater at the head to nearly marine salinity near the mouth; cf. MacEachern et al., 1992). A key to the ichnological identification of incised-valley deposits is the presence of a firmground *Glossifungites* ichnofacies (Fig. 47) that frequently occurs on the sequence-bounding unconformity at the base of the valley (MacEachern et al., 1992; Pemberton et al., 1992). It must be remembered, however, that all of these ichnological characteristics may occur in any brackish-water setting and not just estuaries.

Subdivision of Compound Incised-Valley Fills

A recent advance has been to use detailed compositional data to subdivide complex, compound valley fills into their constituent sequences. Early approaches to incised valleys regarded the fill as an undifferentiated entity, and while later work identified individual components such as bayhead deltas and muddy cen-

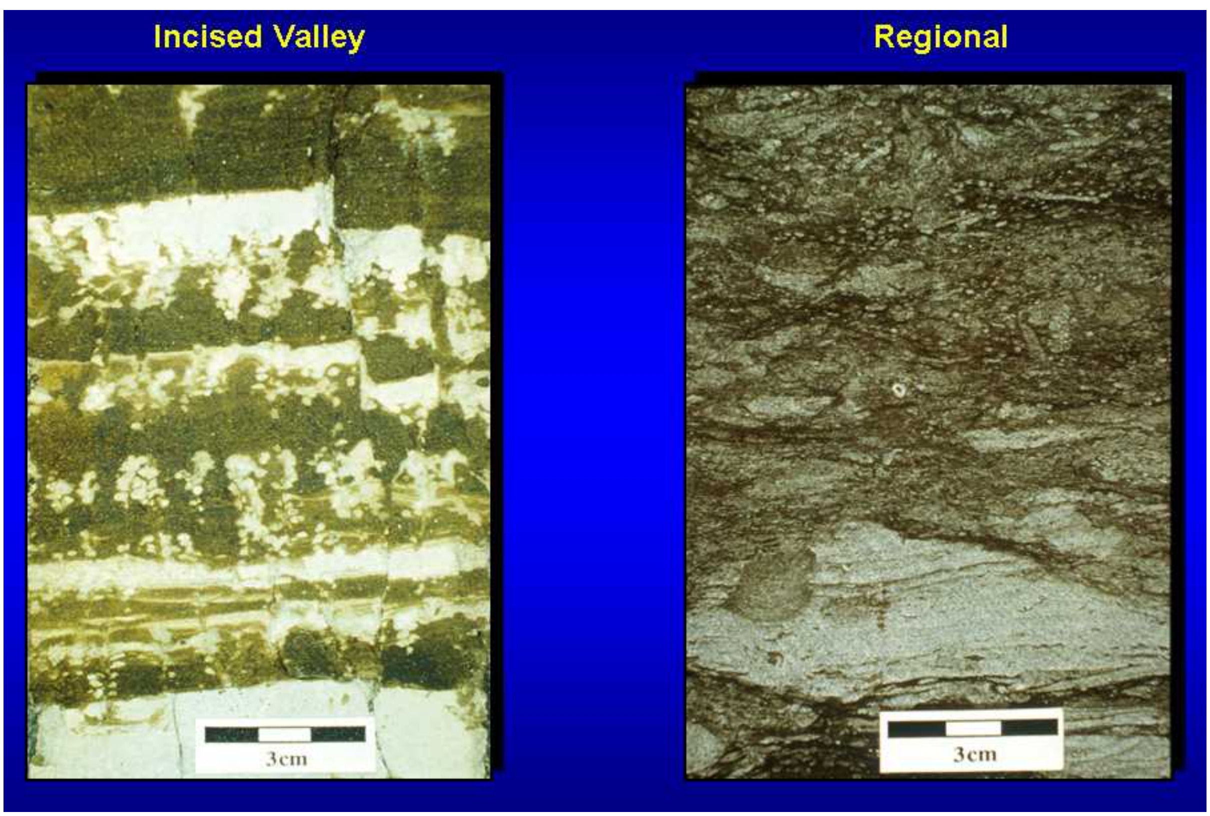


FIG. 45.—Comparison of ichnological traces from brackish sediments (left: monospecific *Gyrolithes traces*) and marine sediments (right: high-species-diversity traces with *Helminthopsis* and *Chondrites* dominant) in the Viking Formation, Western Canada Sedimentary Basin. (Figure courtesy of James MacEachern.)

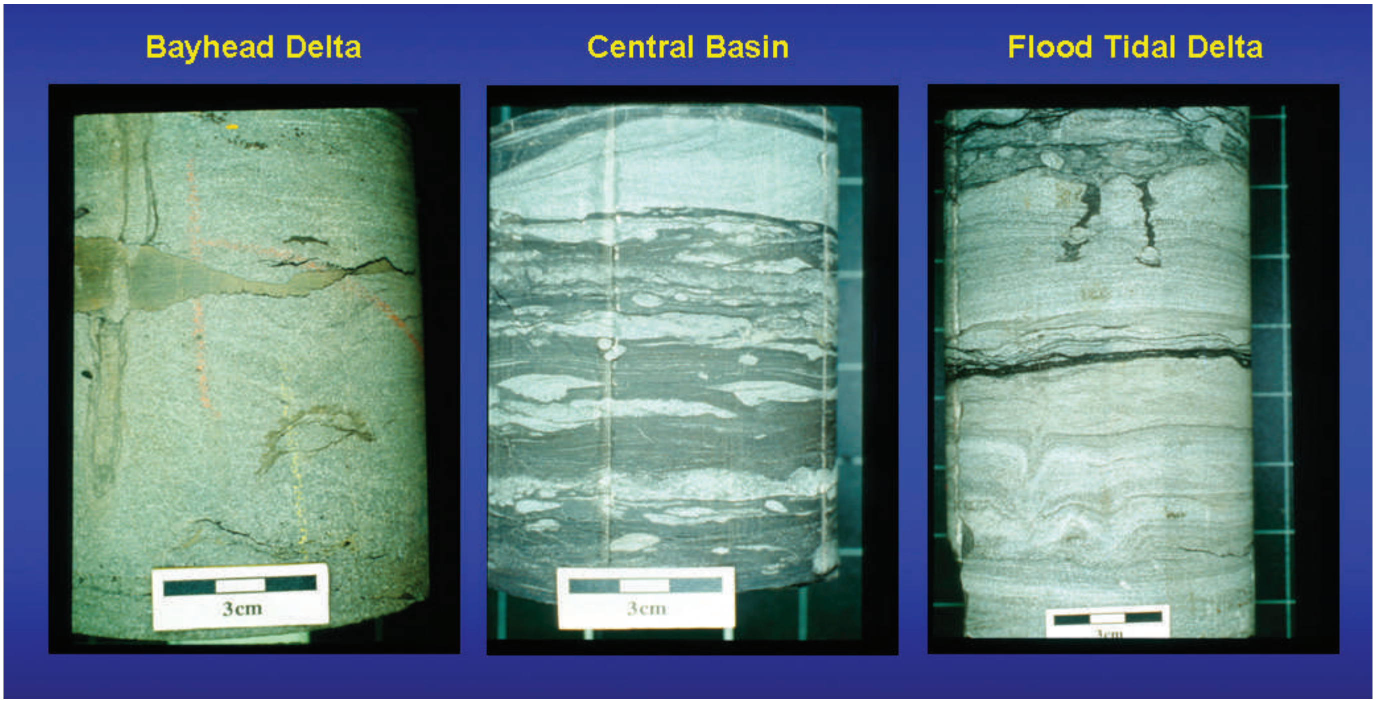


FIG. 46.—Examples of distinctive ichnofacies from the inner (bay-head delta), middle (central basin), and outer (flood-tidal delta) regions of an estuary. (After MacEachern and Pemberton, 1994.)

tral basins, the work of Zaitlin et al. (1994) highlighted the complex nature of many incised-valley deposits as a result of cut and fill over several sea-level cycles. However, these cycles remain difficult to subdivide, especially in low-accommodation settings such as described in Case Study 4 above, despite being of prime importance in establishing petroleum reservoir and seal relationships. Work by Zaitlin et al. (2002) has illustrated how the use of a small number of diagnostic petrological components can be used to differentiate two cycles and five units of cut and fill in a single formation (Figs. 40–43, 48). Other similar opportunities exist to use complementary parameters such as chemostratigraphy, heavy minerals, reservoir properties such as pressure and flow, and remotely sensed electrical properties to identify and subdivide compound valley fills, as well as to determine their provenance. Chemostratigraphy, for example, involves the characterization and correlation of strata using major-element and trace-element geochemistry and has been used effectively in the North Sea (e.g., Preston et al., 1998) and the Western Canada Sedimentary Basin (e.g., Ratcliffe et al., 2004).

3-D Seismic

Earlier 2D seismic-reflection technology was not effective at imaging E&IV systems in the subsurface. This was because the frequencies generated by conventional seismic sources were in the range of 20–100 Hz, which is generally not sufficient to resolve incised valleys with only a few meters to several tens of meters of relief. In addition, 2D seismic collected in single lines could not provide a regional map of incised-valley distribution, which typically exhibits a complex regional pattern (e.g., Figs. 26, 35, 41, 49).

The advent of 3D seismic changed this scenario in several important ways. Firstly, because 3D-seismic acquisition works with an array of receivers for each shot location, there are multiple return paths from each location in the subsurface, providing continuous coverage and a resulting 3D cube of seismic data rather than a 2D slice. Secondly, the 3D method of generating seismic data enables the 3D cube to be sliced horizontally as well as vertically. It also allows the 3D cube to be imaged along individual reflection horizons, which in turn allows visualization of complex paleogeomorphological features. A range of seismic attributes can be used to highlight aspects of the 3D data (e.g., Figs. 24, 33, 49). These include peak-amplitude maps of the depositional surface, and classification of the waveforms being reflected from that surface. These techniques have greatly enhanced our ability to image E&IV settings (e.g., Zeng et al., 1996; Posamentier, 2001; Miall, 2002; Reuter and Watts, 2004) because the fill of incised valleys frequently differs seismically from the surrounding regional sediments. The acoustic-impedance contrast at the base of the valley aids further in imaging the valley container. Finally, the fragmentary coverage of 2D seismic that was ineffective at detecting E&IV facies has been replaced by horizontal maps of seismic attributes that are particularly effective in connecting together the linked reflections that result from long, linear coherent features such as channels and valleys (e.g., the tributary valleys seen in Figure 49). Increased future use of 3D seismic processing and enhancement algorithms will be especially powerful for delineating valley networks and longitudinal changes in the nature of the valley-filling deposits.

Numerical Modeling

As discussed above regarding scientific models, forward progress in the field of facies models will require the development of quantitative techniques to predict the response of E&IV systems to the dominant processes, and to assess the balance between sediment flux and relative sea-level changes. Some important steps have already been taken in this direction, and

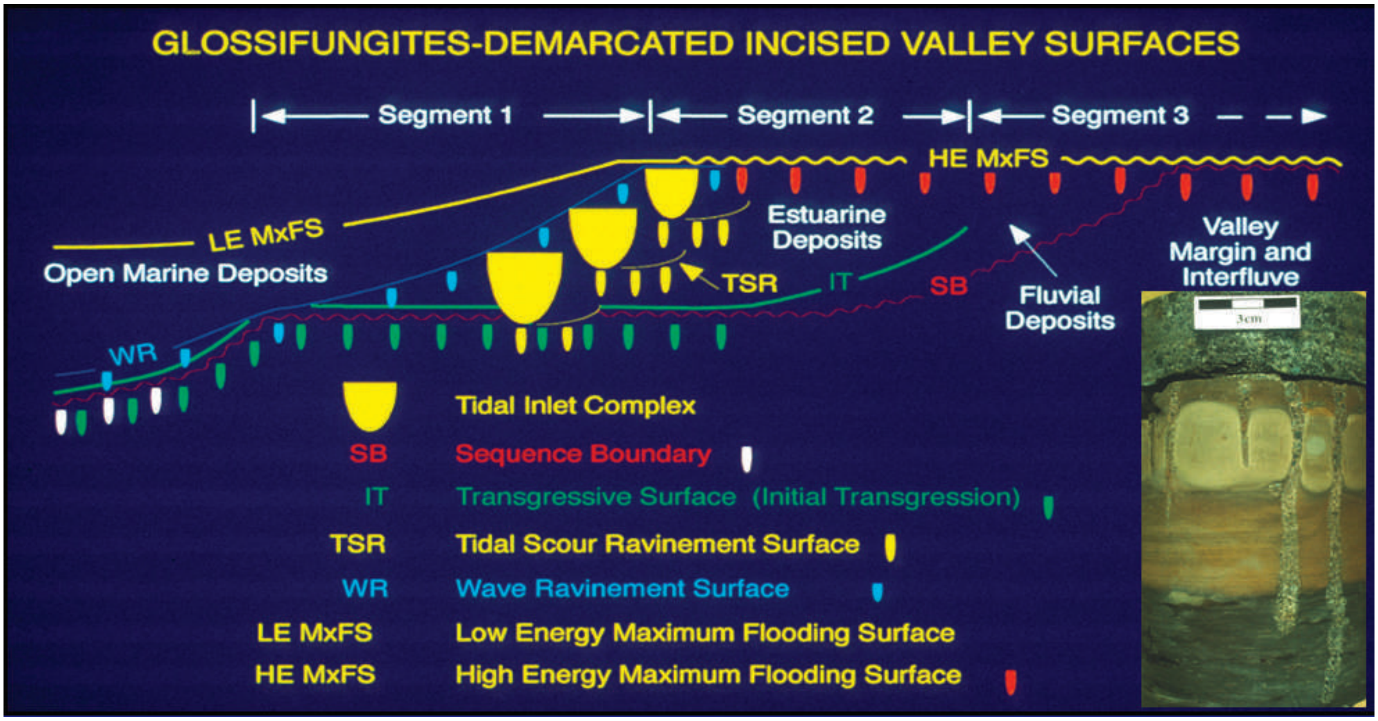


FIG. 47.—Demarcation of incised-valley surfaces by the *Glossifungites* ichnofacies (from McEachern and Pemberton, 1994). Note that this ichnofacies is not unique to sequence boundaries.

preliminary results are available from a number of approaches. There is insufficient space to review this field here, but some of the more interesting approaches are as follows:

(1) The generation of valleys and their fill has been modeled from the perspective of landform evolution models (e.g., Willgoose et al., 2003; Whipple and Tucker, 2002), fluid mechanics models (e.g., Thorne 1994), and alluvial-simulation stratigraphic models (Bridge and Leeder, 1979; Bridge and Mackey, 1993);

(2) The question of E&IV preservation has been modeled with a shoreface-erosion approach (Figs. 50–51) by Cowell et al. (1995), Cowell et al. (1999), and Cowell et al. (2003);

(3) Extensive numerical modelling of estuarine circulation (e.g., the NOAA model for Chesapeake Bay; NOAA, 2003; Fig. 52) and sediment transport has been conducted; and

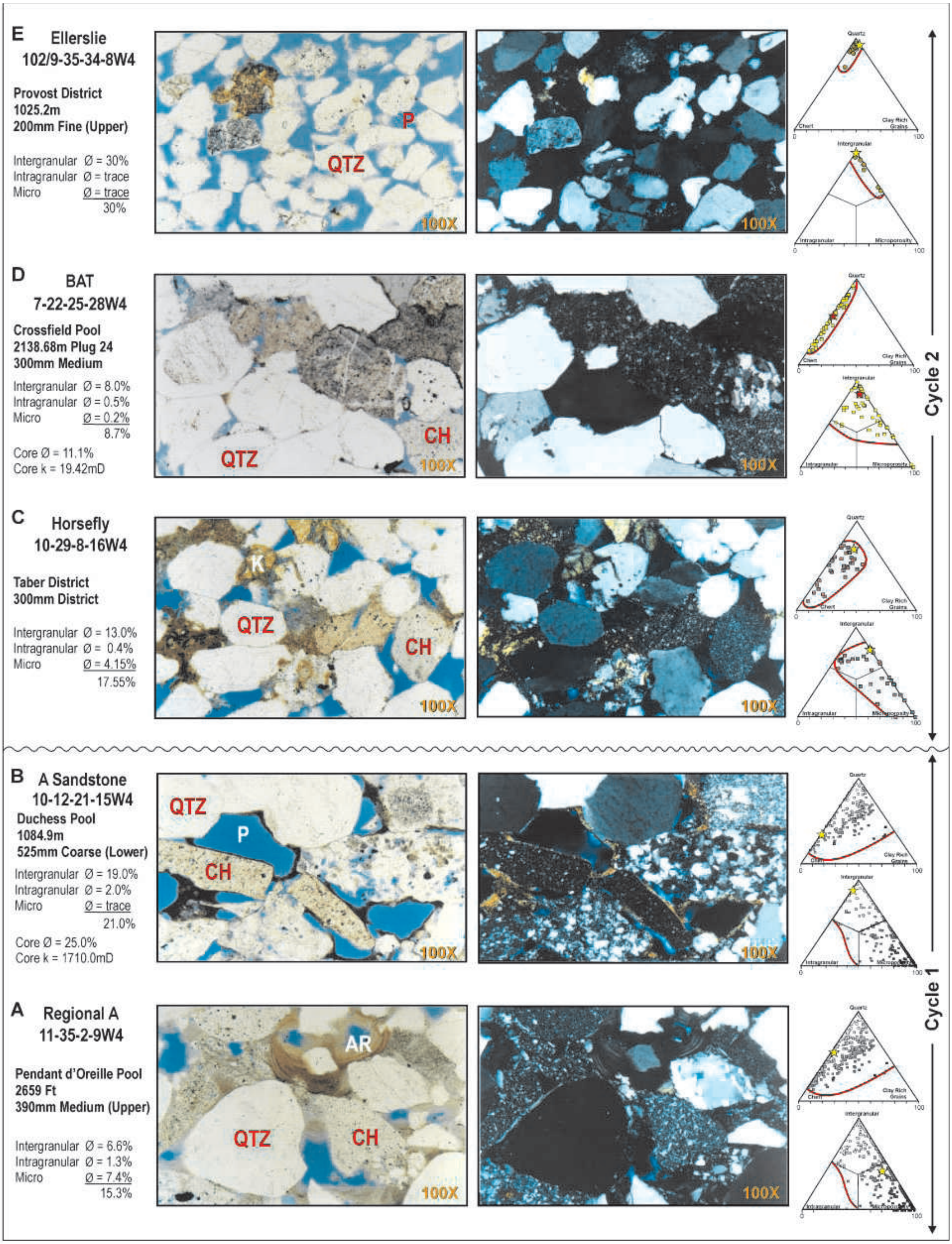
(4) Quantitative relationships have been developed for the balance between river, wave, and tidal power (Fig. 53) and used to test the Boyd et al. (1992) coastal classification through the analysis of all major Australian estuaries (Harris et al., 2002).

However, these quantitative approaches only address individual components of the larger system; a full simulation of E&IV stratigraphy has not yet been attempted.

Seabed Imagery

Improved technology for imaging the modern seabed offers important new insight into marine sedimentary environments. Earlier views of the seabed were derived primarily from individual soundings, followed more recently by continuous 2D seismic and/or echosounder profiles. These earlier acoustic techniques relied on wide-angle single-beam methods with limited spatial coverage. Results were frequently contoured to give a final representation of the current marine depositional surface. However, the detailed character of the seafloor remained elusive, and the ability to image details of the marine depositional surface lagged behind equivalent land-based approaches such as aerial photography and satellite imagery. The development of multibeam sounders, wide-swath side-scan sonars, and the first seabed returns from 3D seismic surveys, combined with accurate satellite position fixing, have fundamentally changed our view of the seabed over the past twenty years, but particularly over the past five to ten years (e.g., Fig. 54). All three of these depth-measuring methods rely on the propagation of sound waves through the ocean and their reflection from the seabed, providing a marine acoustic image equivalent of aerial photographs, Landsat images, and digital-elevation models for the terrestrial environment. This provides us with our first real view of what is on the ocean floor at the same degree of resolution as that available on land. Detailed understanding of the modern depositional surface in estuaries enables us to interpret better the vertical stacking of depositional and erosional surfaces that are imaged below the seabed in 3D seismic data.

These new views of estuaries have shown us tidal bedforms in great detail, from the centimeter to the tens-of-meters scale (Fig. 54). They have provided details of separation of flood and ebb tidal currents, maps of the distribution of the marine flora and fauna, the nature of deep scour holes, and the release of biogenic and thermogenic gas from pockmarked estuary floors. Derivation of acoustic backscatter values from side-scan and multibeam data has enabled correlations to be made with sediment grain size and hence has provided the promise of remotely mapping the detailed distribution of sediment texture on the floor of estuarine and adjacent shallow-marine areas.


E Ellerslie
102/9-35-34-8W4
Provost District
1025.2m
200mm Fine (Upper)
Intergranular Ø = 30%
Intragranular Ø = trace
Micro Ø = trace
30%
P
QTZ
100X
100X
D BAT
7-22-25-28W4
Crossfield Pool
2138.68m Plug 24
300mm Medium
Intergranular Ø = 8.0%
Intragranular Ø = 0.5%
Micro Ø = 0.2%
8.7%
Core Ø = 11.1%
Core k = 19.42mD
QTZ
CH
100X
100X
C Horsefly
10-29-8-16W4
Taber District
300mm District
Intergranular Ø = 13.0%
Intragranular Ø = 0.4%
Micro Ø = 4.15%
17.55%
K
QTZ
CH
100X
100X
B A Sandstone
10-12-21-15W4
Duchess Pool
1084.9m
525mm Coarse (Lower)
Intergranular Ø = 19.0%
Intragranular Ø = 2.0%
Micro Ø = trace
21.0%
Core Ø = 25.0%
Core k = 1710.0mD
QTZ
P
CH
100X
100X
A Regional A
11-35-2-9W4
Pendant d'Oreille Pool
2659 Ft
390mm Medium (Upper)
Intergranular Ø = 6.6%
Intragranular Ø = 1.3%
Micro Ø = 7.4%
15.3%
AR
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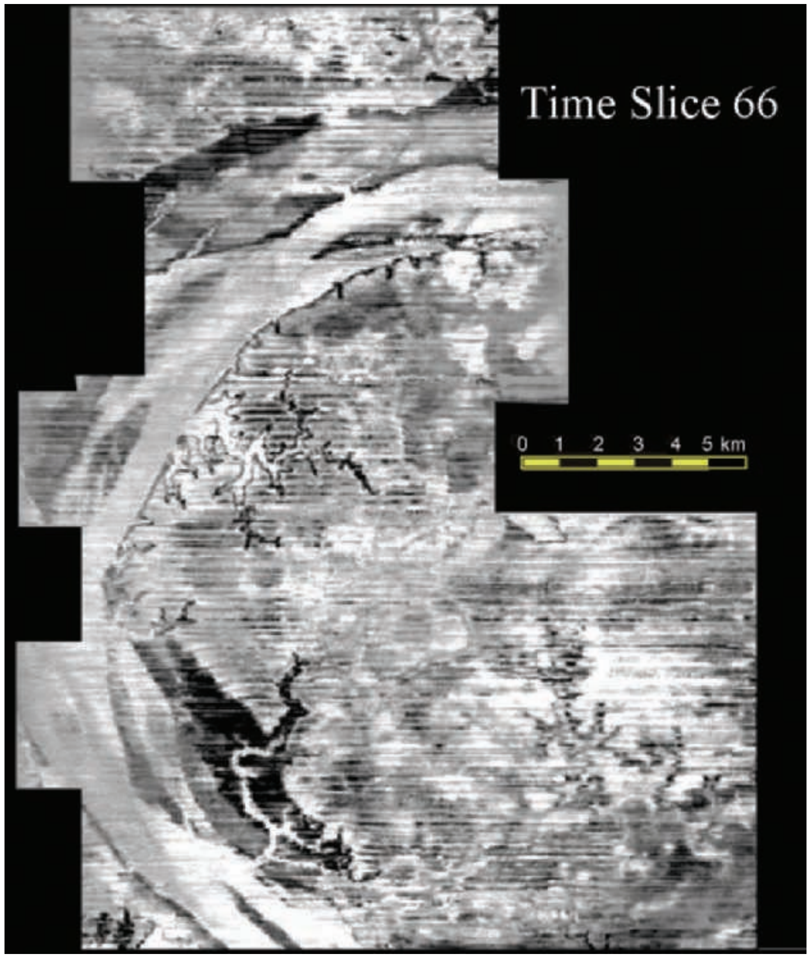

FIG. 49.—3D seismic time slice of a Late Pleistocene incised valley from the Java Sea shelf, offshore Indonesia. Note valleys tributary to the main valley. (From Posamentier, 2001.) Compare with Figure 18.

A LOOK FORWARD—ELEMENTS OF AN E&IV FACIES MODEL FOR THE TWENTY-FIRST CENTURY

The recent advances in E&IV models documented above, and progress in the field of facies models in general, enable us to delineate an ideal facies model of the future. Such a model would: (1) produce a range of realistic E&IV stratigraphy and facies from a given set of input parameters (see example list below), (2) identify the preservation potential of the stratigraphy produced in that model, (3) hindcast the input parameters for a given field example, and (4) predict the rest of the model or example from elements of the component data set. Following the approach of developing more realistic and quantitative facies models outlined above, the following elements represent important components of an E&IV facies model for the twenty-first century:

(1) A precise definition of the E&IV system and its morphological elements.

(2) A quantitative (digital) database of the geometry and facies of entire systems and their component elements from many global examples, both ancient and modern. This should cover the spectrum of systems and be able to

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FIG. 48 (opposite page).—Representative thin sections of the major Basal Quartz units with associated point-count data. Two sets of ternary diagrams are used to illustrate variations in textural and mineralogical maturity. The upper ternary diagram of each pair has quartz, chert, and clay-rich grains at the apices and is effective in partitioning the petrographic data into distinctive populations of mineralogical maturity. The lower ternary diagram of each pair has intergranular, intragranular, and microporosity pore types at the apices and is used to illustrate porosity fabric and reservoir quality. The representative thin sections are organized into two cycles (see Fig. 40). Star and triangles represent locations of point-counted samples in the ternary diagrams. Left photomicrograph in each pair taken in plane light, right photomicrograph of each pair in crossed polars. Magnification 100x. QTZ = quartz; CH = chert; AR = argillans; P = porosity.

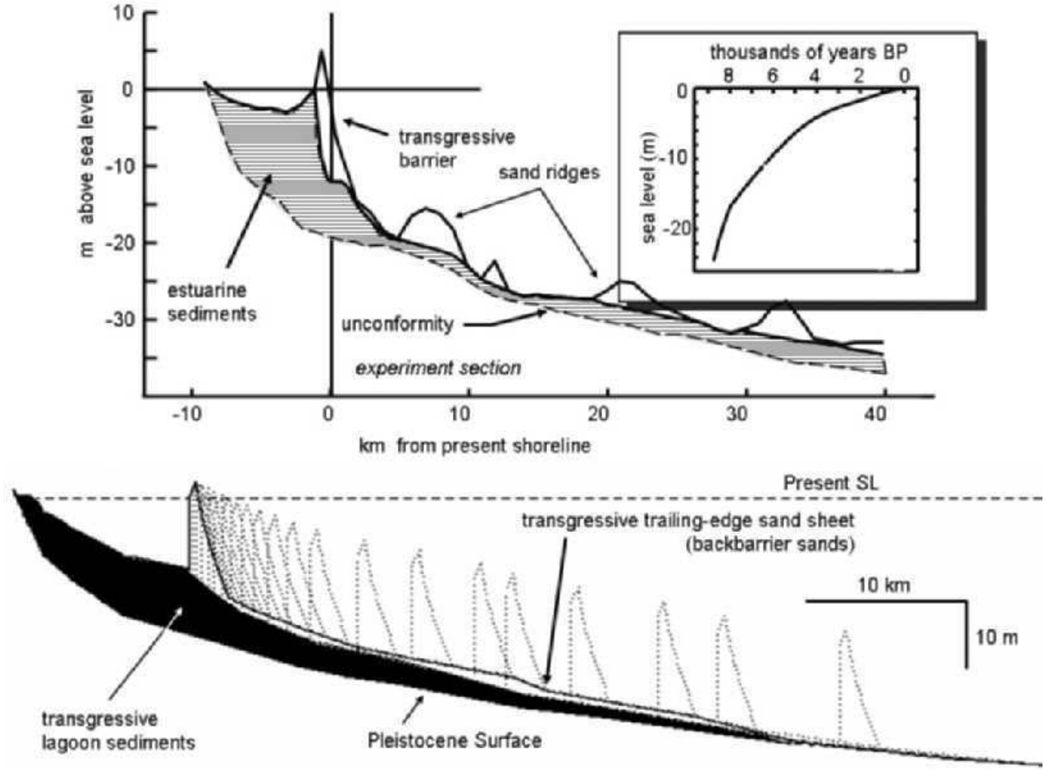

FIG. 50.—Transgressive shoreface simulation of Duck, North Carolina, U.S.A., showing retention of a thin estuarine valley fill (horizontal stripes) after shoreface translation during the last 9 ky (sea-level curve in upper right). From Cowell et al. (1999) and Cowell et al. (2003).

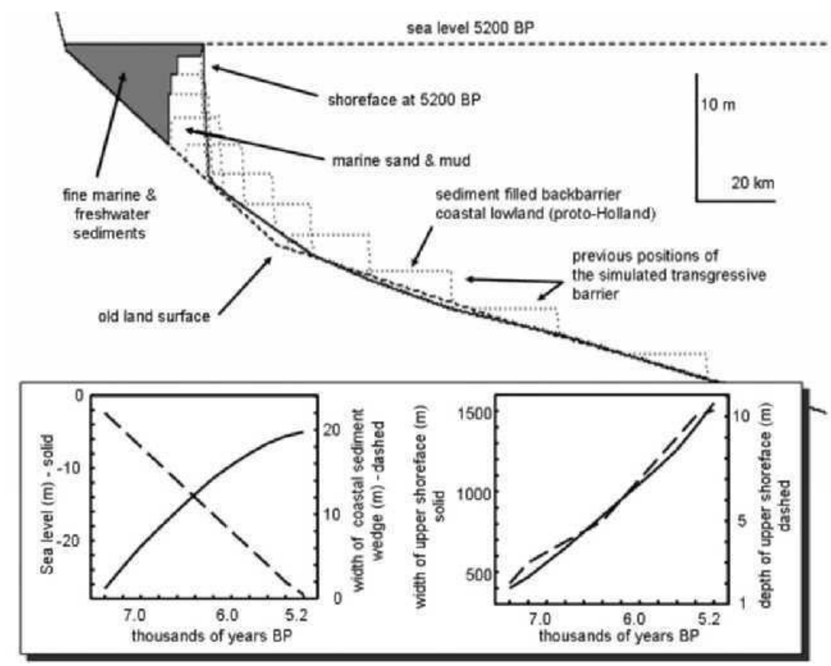

FIG. 51.—Transgressive shoreface simulation of Haarlem, The Netherlands, showing reworking of shelf deposits into the backbarrier during transgression. In contrast to the situation shown in Figure 50, almost all of the estuarine sediments (gray color) have been removed from the shelf but have been preserved behind the aggrading barrier at the present-day shoreline (which marks the position of maximum transgression and the landward limit of the ravinement surface). From Cowell et al. (1999) and Cowell et al. (2003).

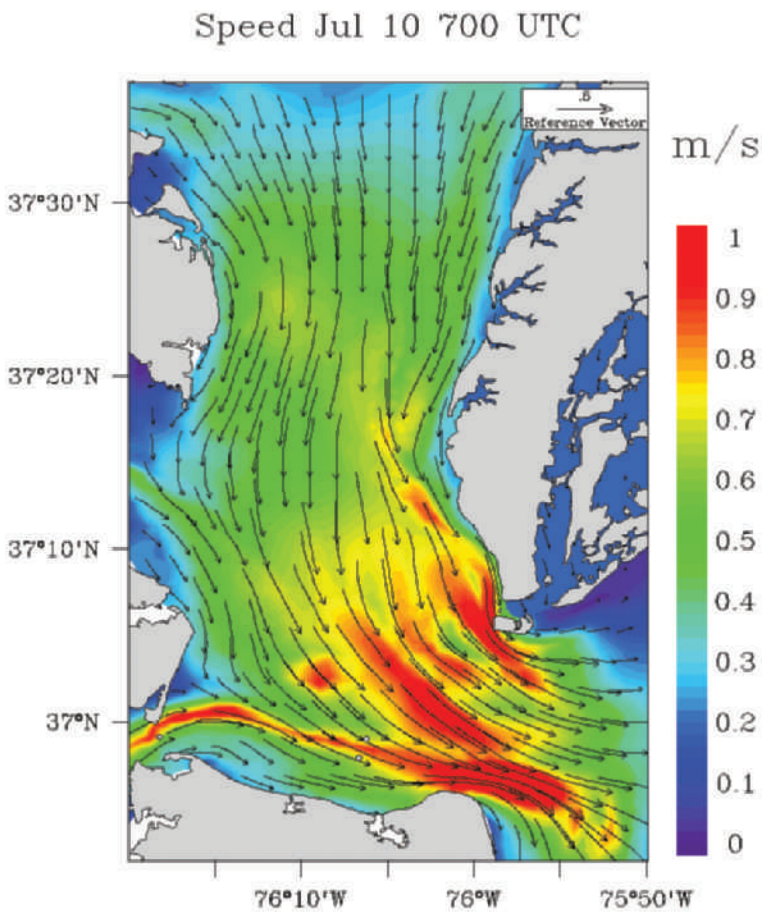

FIG.52.—Animation of tidal circulation in Chesapeake Bay. (From http://ccmp.chesapeake.org/C3POANIM/). Color bar on right shows surface tidal current speed in m/s at 0700 on July 10, 2006. Arrows on figure show direction of water transport.

identify "average" or most frequently occurring geometries and the common internal facies characteristics of each system element. The database should be managed as an open structure able to be accessed by all researchers via grid or web-based computing and have a template for common data entry.

(3) A list of the major processes operating in the E&IV system and a description of their dynamic characteristics. Examples of these processes include, but are not limited to, plane jet flow in bayhead and tidal deltas, channelized flow in inlets and tidal–fluvial channels, wave motion at the seaward margin, and relative sea-level changes throughout the system.

(4) A sediment-input component providing sediment volume, direction, texture, and composition. These inputs could be empirical values or derived in turn from models such as climate simulations, wave, and tide predictions.

(5) Computer-modeling software developed to simulate the processes identified in #3 above using inputs of geometry and facies from #2 and sediment input from #4. For complex systems such as E&IVs, the software models would require a number of linked modules to incorporate the range of processes present. Early models could utilize a smaller subset of the processes to describe wave-dominated estuaries, for example, or a fluvially eroded valley, while more complex models would be required to describe the response of fluvial and estuarine systems to sea-level change or incised-valley evolution over a complete sea-level cycle, and to predict the range of subsequent preservation outcomes. Ideally, computer-modeling software would also have an open architecture and be available on line so that users could simulate parts of the overall system or link several modules together following the lead of other geoscience modeling networks such as www.geoframework.org for internal earth processes.

(6) Output models would exhibit a spectrum of 3D examples spanning the range of natural E&IV variability, together with a set of "average" models that would describe the most frequently occurring combinations of natural parameters (e.g., most common values of wave height, tidal range, valley size, rate of sediment supply, and rate of sea-level variation) and sediment characteristics. Model output would be evaluated on how well it reproduced type field examples.

The field of facies models in general has had a rapid rise in knowledge and application over the past forty years, with estuary and incised-valley models exhibiting a similar rise in popularity over the past thirteen years. As new strides are

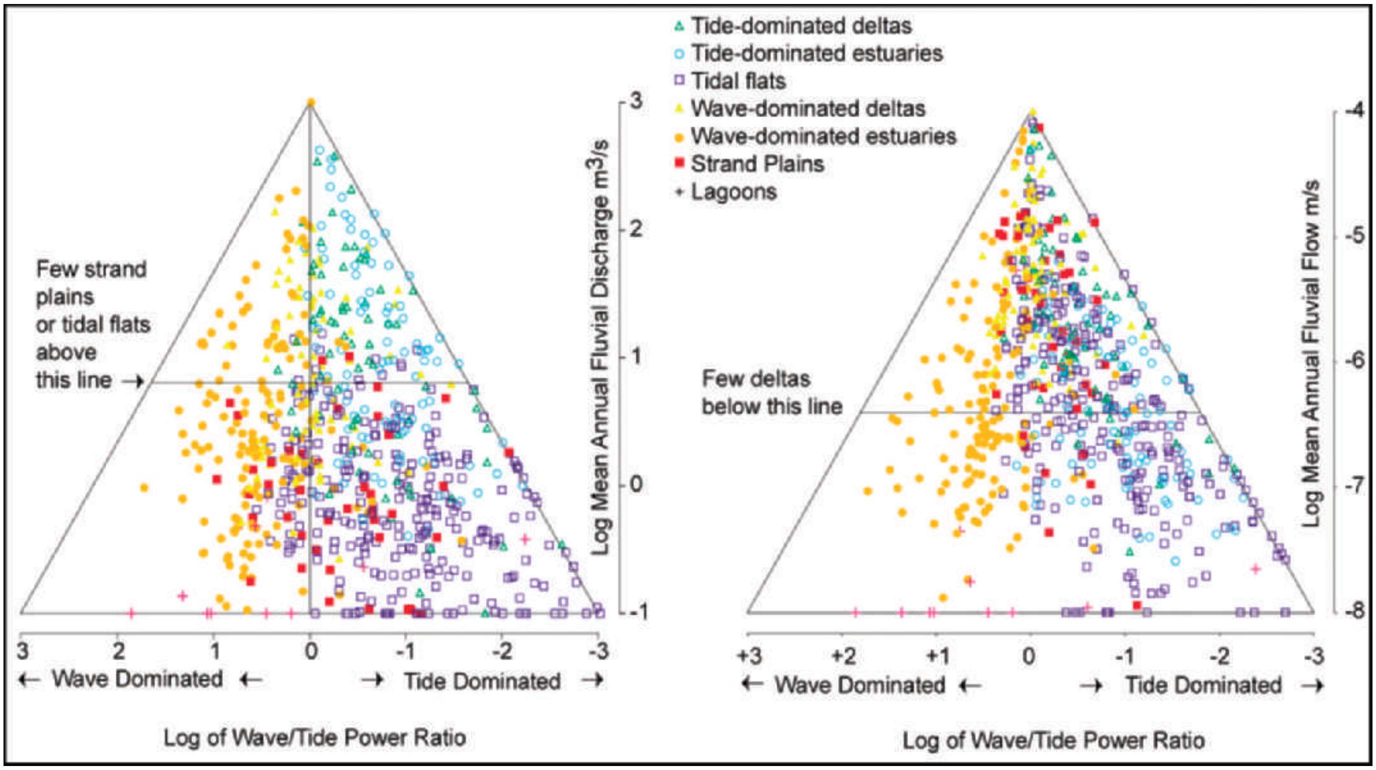

FIG. 53.—Coastal classification and empirical model testing from Harris et al. (2002). In this approach, the parameters of log [mean annual fluvial flow] (vertical axis, right) and fluvial discharge (left) have been plotted on a ternary diagram against log [ratio of tidal power to wave power]. This provides a quantitative test for the Boyd et al. (1992) coastal-classification scheme using all the river mouths on the Australian coast. Note segregation of major coastal depositional settings.

made to transform the current spectrum of classification models into empirical and theoretical models, simulated on computers and tested in the field, further advances to a new level of understanding sedimentary depositional systems can be anticipated.

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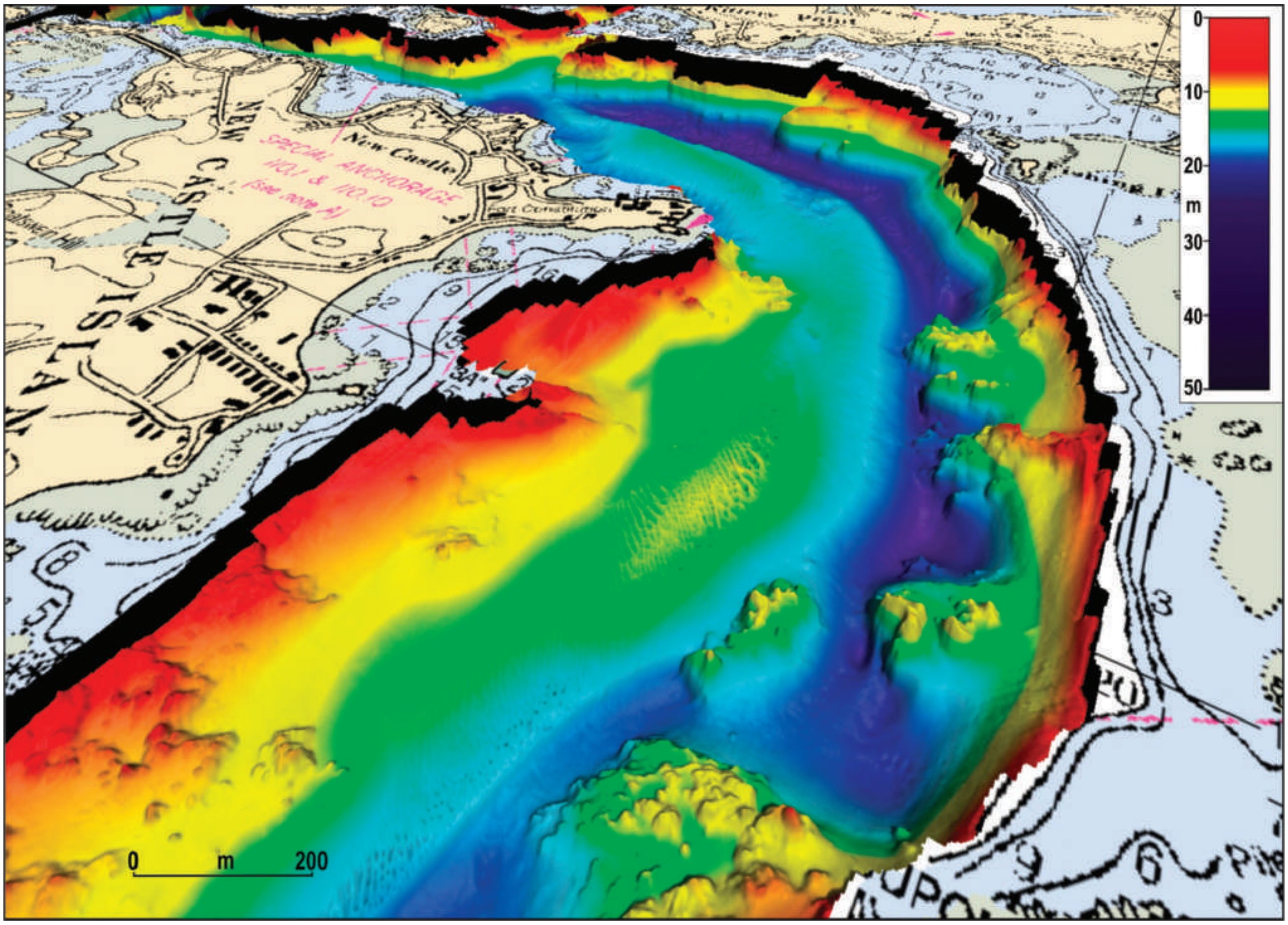

FIG. 54.—Reson 8101 multibeam data from Portsmouth Harbor, New Hampshire, U.S.A., showing a high-resolution image of estuarine geomorphology including the channel thalweg (dark blue), an extensive tidal dune field (center) and localized bedrock outcrops (e.g., right-hand side of channel). Data collected by NOAA as part of the Shallow Survey 2001 Common Data Set (Mayer and Baldwin, 2001) and processed by the Center for Coastal and Ocean Mapping, University of New Hampshire. 3-D visualization created using the Fledermaus software suite. Color bar, top right, shows depths in meters below sea level.

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