The late Edwin McKee conducted an extensive stratigraphic study of the Cambrian of the Grand Canyon along its length (McKee, 1945). He discovered two important things. First, he recognized repetitive shale to nonskeletal sequences with sharp contacts—what we now would call cycles. Second, using fossil assemblages, he noted the stratigraphic units, he also found that distinctive concept features, such as the rusty-brown (trinitoidal) dolomite cliffs in the western Grand Canyon, were separating lithologic units of a different age than the rusty-brown dolomite cliffs in the eastern Grand Canyon (fig. 1). McKee demonstrated through paleontologic evidence that these individual sequences were correlatable for 100-200 km (62-120 mi) across depositional winkle but were diachronous—a time-transgressive sequence of the Middle Cambrian. These were important foundations.

Today we recognize that McKee's cycles, which are 10-70 m (30-230 ft) thick, are actually groupings of smaller-scale cycles (fig. 2; Wanless, 1973a,b). These smaller, 1-10-m thick (3.3-33-ft thick) cycles represent the fundamental depositional sequence or cycle of sedimentation. Scale content changes upward in each fundamental depositional sequence, and each has a sharp cap. Each of these fundamental depositional sequences appears to represent a shallow- or upward cycle (A → B), and many are capped by an exposure surface (C). Each fundamental depositional sequence can be correlated laterally, but the facies change across depositional strike (fig. 3). The upper parts of the cycles vary from fine silicic sands to glauconitic sands to skeletal and peloidal calcarenites to clean carbonate muds.

Fundamental depositional cycles commonly group into larger-scale sequences or cycles (grand cycles) of the Middle and Upper Cambrian of southern Canadian Rocky Mountains (Atkin, 1966) and lower Paleozoic of the central Appalachians, eastern United States (Rest, this volume).

Abstract  The design of useful models for predicting sequences and facies patterns of sedimentary cycles depends on an observational foundation that includes the recognition and adequate understanding of fundamental depositional sequences. Six facets must be met. The modeler must have (1) adequate documentation of sequence character; (2) adequate characterization of spatial and directional variability and continuity of facies; (3) an adequate database on the influence of primary and secondary controls on sedimentation; (4) an understanding of interactive or controls on sedimentation; (5) an understanding of the limits of physical, biological, and chemical controls; and (6) an understanding of diagenetic modifications to sequence nature and thickness. In addition, the modeler should work from the fundamental depositional sequence. The modeler should not caution on applying models designed for one scale of sedimentary sequence to another scale. Examples illustrate the necessity for meeting each facet.

Although McKee's aggregate cycles and the grand cycles of Atkin (1966) contain a number of smaller fundamental depositional sequences of decreasing scale upward, the deeper shaly facies (A) dominate in the lower part of the aggregate or grand cycle and the shallower nonskeletal facies (B) dominate in the upper part (fig. 2). Facies related to exposure (C) may cap the aggregate cycle. A random sampling through an aggregate or grand cycle would very likely miss the contained fundamental depositional sequences and would likely produce a confusing sequence of sedimentation. In fact, each fundamental depositional sequence in the Grand Canyon Cambrian strata is a shallowing-upward sequence, and many are capped by an exposure surface. The aggregate cycles are easily seen and traced from distant overviews of the Grand Canyon; the fundamental depositional sequences are seen only by detailed documentation of the complete sequence in steep, well-exposed outliers or core borings.

Here, a fundamental depositional sequence is considered the smallest subdivision of lithofacies assemblages that includes the main recurring lithofacies. As such, it records the basic sequence of processes and environments that constitute the larger sequence. A fundamental depositional sequence may or may not be cyclic. The fundamental depositional sequences of cycles of sedimentation are the fundamental lithologic building block or sequence unit of stratigraphic sequences. As such, they are the unit from which cyclic (or noncyclic) sedimentation, basin architecture. unit sequence stratigraphy must be modeled.

The term "parasequence" is a parallel arm but has been avoided here because it is formally tied to sea-level dynamics, as defined with respect to scale, and is commonly used in a seismic expression of the basic unit of sedimentation. Similarly, the term "punctuated gradational cycle" (PAC) (Anderton and Goodwin, 1978) is not used because it implies a genetic relation to sea-level change. Parasequence and PAC are specific interpretive types of fundamental depositional sequences. The building block for modeling must be based on...
Figure 1. East to west stratigraphic cross section of the Cambrian of the Grand Canyon, Arizona, from McKee (1945). By using traceable faunal and lithologic horizons, McKee showed that the sequence was time transgressive. Cycles within the sequence are equivalent to grand cycles and contain smaller fundamental depositional sequences. Thickness varies from 800 ft (240 m) in the east (right) to 1,500 ft (450 m) in the west.

Figure 2. The Cambrian of the Grand Canyon is organized into a series of grand cycles, basically equivalent to the sequences recognized by McKee (1945). Each grand cycle contains a series of 1-8-m-thick (~2.6-ft-thick) fundamental depositional sequences. These fundamental depositional sequences are cycles containing a sharp base, a lower sandy unit (a), an upper nonsandy unit (b), and a cycle cap (c). Sandy units dominate the fundamental depositional cycle in the lower part of a grand cycle. Nonsandy units dominate the upper portion of a grand cycle. See fig. 3 for facies types and internal variability of the fundamental depositional sequence.
Figure 3. Individual fundamental depositional cycles in the Middle Cambrian of the Grand Canyon, Arizona, can be traced and correlated laterally across depositional strike. Each cycle (increasing thickness from 1–3 m (3–9 ft) in the east (right) to 3–8 m (9–26 ft) in the west. The facies chart of the lower (shaly) part of each cycle has a gradual lateral gradational. The upper nonshaly part of the cycle contains more pronounced lateral facies variation from fine quartz and glauconite sands in the east to calcareous and algal-ball limestones and dolomites in the middle to increasingly clean limestones to the west. Cycle cap facies represent exposure conditions (Watkins, 1973b). Paleocurrents are shown for the lower shaly units (as ripple cross lamination) and the upper fine sandstone and glauconite sandstone units (im crossbed dip directions). These marine facies are in marked contrast with the paleocurrent flow direction from the rough crossbedded basal Tapeats Sandstone of fluvial origin.
lithologic sequences and should not be formally tied to any one depositional influence.

It should be recognized that some sequences may be composites of more than one fundamental depositional sequence or cycle. These interfacing sequences may or may not be in phase or of the same duration. For example, constructive-destructive phases of lobes of a prograding delta produce one group of fundamental depositional sequences; sea-level rise and fall may produce another (different time scales, no phasing relation). Or, ice-rafted sediment related to a cycle of polar ice buildup may interleave with shelf or slope sedimentation influenced by sea-level rise and fall (same period of influence but not necessarily in phase).

In some sequences, event units (tempestites, turbidites, slumps) distinctly occur within fundamental depositional sequences (comprising as part of a shallowing-upward sequence) or contain vertical trends in the nature of event units (sequence of thickening upward tempestite units). Other sequences may not display clear fundamental depositional sequences but may be composed of numerous event units. Finer-scale event sedimentation units may not be conducive to predictive modeling, especially if the driving forces are episodic.

It is my goal in this article to offer some guidance for the recognition of the small-scale fundamental depositional sequences or cycles of sedimentation and to provide in awareness of the variety of depositional factors that can dramatically influence the nature of the preserved facies and fundamental sequence. It is hoped that this will lead to improved modeling of sequences of sedimentation.

Observational foundation

To get at the nature, variability, and meaning of the small-scale fundamental depositional sequence, it is necessary to look at three groups of observational foundations: descriptive characterization, understanding of influences, and incorporation of diagenetic modifications. Most fundamental to descriptive characterization is proper and adequate (1) characterization of the real-world sequences and (2) their lateral variability. To model a descriptive sequence so that its influences can be understood, one needs a good understanding of (3) the nature of controls on sedimentation, (4) the effects of their interactions with one another, and (5) the limits of a control's influence. Finally, one needs to have a good understanding of (6) diagenetic modifications to the nature of the sequence and its thickness. Rather than trying to give an overview with lists of hundreds of things that might be influential, a few specific examples are offered under each observational foundation topic. Most are within my experience. Some examples are a bit unusual but should serve to stimulate the reader's imagination for improved modeling efforts.

Sequence characterization

Sequence characterization must include (1) proper definition of the fundamental depositional sequence; (2) adequate statistical characterization of the vertical sequence; (3) proper definition of sequence boundaries; (4) definition of lateral facies continuity and boundaries; (5) clear differentiation of transgressive, regressive, and stratiform facies; and (6) differentiation of sheet deposition versus bank-mound ridge accumulation of facies. This characterization must be based on lithologic studies of outcrops and/or cores. To build sequence models entirely on interpreted seismic information is to build models from models—most likely a house of cards.

Proper definition of fundamental depositional sequence

Research on sequences should begin with proper documentation of the lithologic nature and variability of the fundamental depositional sequences, adequate evaluation of the depositional environments and processes recorded by those sequences.

Example

At a recent conference on Mississippiian sediments, Hunter (1989) led a portion of the field trip to the S. Genevieve Limestone in Indiana and offered strong evidence that significant portions of the shallowing-upward onlith cycles were regressive. This recognition revises the nature of the fundamental depositional sequence, completely changes the spatial relations of environments, changes conclusions about environments of onlith production, sediment budget, and transport processes, and thoroughly revises the possibilities for early diagenetic changes. It is thus important not only to recognize the fundamental depositional sequence but also to understand what it represents.

Adequate statistical characterization of the vertical sequence

It is extremely important to determine whether a cyclic sequence really exists.

Example

In the "golden rule" of cyclic sedimentation (Merritt, 1964), several vertical ("stratigraphic") number sequences were taken out of the Lawrence, Kansas, telephone book, and geologists unwittingly generated cyclic: from three number sequences and then correlated them (Zeller, 1964). Some sequences have only motifs or patterns (Walker and Iwans, 1971) but not cycles or trends.

Proper definition of sequence boundaries

In several cases sequences have been termed fining upward (because that is the type of sequence with which the geologist was familiar) when, in fact, the sequences are coarsening upward and the cycle boundary has been improperly placed. A supralatal facies zone can easily be called a cap to a shallowing-upward cycle, whereas it may actually represent either the transgressive beginning of a cycle or partly the cap of one cycle and partly the beginning of the next cycle of sedimentation.
Definition of lateral facies-continuity and boundaries: It is crucial to understand how sequences and facies within a sequence laterally interrelate and change.

Example: The dramatic Wauchopean-type eustatic mound builds of New Mexico are diverse features with a patchy distribution, high relief, and abrupt lateral boundaries with other facies (flanking grainstones) (Wilson, 1975). In contrast, the dolomitic crinoidal and algal bank facies in the cyclic Cambrian of the Grand Canyon are broad, low-relief features (figs. 1 and 3). They are 1-5 m (3.3 to 16.5 ft) thick, laterally continuous (perhaps for hundreds of kilometers) along depositional strike and 15-30 km (9.5 to 19 mi) wide perpendicular to depositional slope with gradual onlap boundaries with the landward glaciomarine sandstone facies, the seaward-beded limestone facies, and the underlying thin interbedded sandstone and shale facies. One must model the nature and influence of these two buildups differently because of their different sediment body geometry, form, continuity, scale, and lateral boundaries.

Clear differentiation of transgressive, regressive, and stillstand facies: Perhaps the most difficult problem is differentiation of transgressive stillstand, and regressive deposits.

Example: In many models of clastic shelf-plain sedimentation, current thinking is that sea level rises and then sedimentation (aggradation and progradation) occurs during a stillstand (Vail et al., 1977). In fact, Brown's rule suggests that a significant portion of inner shelf sedimentation should occur during a rise in sea level (as sand is transferred from the shoreface to a deepening inner shelf) (Brown, 1962; Schwartz, 1967; Swift, 1988; Dominguez, 1987; Dominguez et al., 1987) and that shoreface progradation should occur during stillstand (longshore sediment supply) or falling sea level (sand transfer from the seaward inner shelf to the shore) (Dominguez, 1987; Dominguez et al., 1987). Although erosion of the shoreface occurs with sea level rise, the deepening inner shelf provides accommodation for sediment accumulation. This deposit of transgressive shelf sand may become an important part of the sedimentary sequence (fig. 4A). Dominguez (1987) and Dominguez et al. (1987) have shown this to be the case for the Holocene shelf and sublittoral sequence in Baja. Transgressive and regressive shelf sensu deposit can become an amalgamated subtidal sand unit that would be easy to misinterpret as entirely regressive, especially if the shelf was shallow. Calculated rates of sediment supply and the nature of facies deposition are different if deposition occurs during transgression.

Example: It is also important to understand that there are expected facies successions that occur with time within transgressive, stillstand, and regressive deposits. For example, the Pleistocene and Holocene of the western portion of Cuba and platform (southeast Bahamas) contains three reef to oolitic grainstone sequences (fig. 5) (Wallace et al., 1989).

The two Pleistocene reef to oolitic sequences are capped by calcite exposure surfaces. Each sequence is a single cycle of sedimentation in which initial reef growth is eventually overwhelmed and smothered by oolitic sand. I have observed this reef to ooloid sequence from the Lower Cambrian 

archaoscopic banks to the Holocene. Some researchers have suggested the choice is reefs or ooids (Perkins, 1986) rather than reefs, then ooids. Models will differ greatly depending on the path chosen.

Example: Heckel (1978, 1983) has attempted to differentiate shallow carbonates formed by high stands from carbonates formed during the following transgression. He has used stratigraphic relationships, succession of conodont biofacies, and diagenetic characteristics to differentiate transgressive and regressive portions of the cycles in the Pennsylvanian Captain Creek Limestone Member of southeastern Kansas. Transgressive facies include basin stromatolites, oolite, and phytoplankton mound facies and an absence of marine or eustatic diagenesis. Regressive carbonates record marine and meteoric diagenesis.

Diferentiation of shelf deposition versus bank-ruled ridge accumulation of facies: It is important to understand whether a deposit has the form of a sheet or a mound.

Example: The dolomitic crinoidal and algal-bank facies of the Grand Canyon Cambrian occur between the seaward-beded limestones and the landward glaciomarine sandstones (fig. 3). The crinoidal and algal-bank facies appear to have formed as a broad, shallow bank. This dolomitic crinoidal bank appears partway through a depositional cycle (fig. 3) and served largely to block the seaward transport of fine-grained sands and muds and to isolate a broad innerlagoon of glaciomarine and quartz sands from the outer platform of carbonate deposition. Fine siliciclastic sand and glaciomarine extend seaward into the dolomitic crinoidal facies and fade out. Crossbedded glaciomarine and quartz sands in the inner lagoon are oriented in a north-south longshore direction, indicating energy for clastic sediment transport but seaward restrictions to flow. The algal-bank facies formed as a synchronous or subsequent prograding seaward flank to the crinoidal bank. Recognition that a facies forms as a shelf bank rather than as a simple subtidal sheet greatly modifies the influence one must give to it in a sedimentary model. Exposure features may cap the inner coarse and bank parts of the cycle, and style of the subsequent cycle is spread across the entire sequence, marking the synchronous initiation of a new cycle of sedimentation.

Example: This problem becomes particularly evident when evaluating why the carbonate stones of the Great Bahama Bank form a broad subtidal shelf deposit, whereas those of Florida Bay and Biscayne Bay are organized into a variety of mound or bank complexes (fig. 6) (Wallace and Tagen, 1989; Wallace et al., 1989). The answer appears to lie in the influence of preexisting topography on circulation and
Figure 4. (A) Application of Brazeau’s rule to rising (top) and falling (bottom) sea level (from Dominguez 1987). Landward-migrating barrier islands and shell accretion characterize rising sea level as sand is transferred from the shore face to the inner shelf to restore an equilibrium profile; strandplain progradation and shell erosion characterize lowering of sea level as sand is transferred downdrift from the inner shelf in an effort to maintain a profile of equilibrium. (B) Sedimentary environments on Caravelas beach-ridge plains, southeastern coast of Brazil, a strand plain not directly associated with a river mouth (from Dominguez et al. 1987). (C) Sedimentary environments on São Francisco beach-ridge plains, eastern coast of Brazil. Progradation is associated with times of lowering relative sea level.
sedimentation and the rates of flooding of the platforms. The Great Bahama Bank was inundated 5,360–7,000 years ago when the rate of relative sea-level rise was greater than 5 m (15 ft) per 1,000 years. During this time of flooding, topo-

graphic irregularities had such a rapidly changing influence on sedimentation that they had little overall influence on it. In contrast, Florida Bay and Biscayne Bay were inundated during the past 4,500 years when relative sea level was rising at less than 2.5 m (8 ft) per 1,000 years. Much of the growth of the south Florida mudbanks has been during the past 3,600 years, in association with a sea-level rise of less than 0.5 m (1.6 ft) per 1,000 years. These slower rates of sea-level rise have resulted in relatively subdued topographic irregulari-
ties with a more prolonged, persistent influence on patterns of sedimentation. This appears to have been the primary cause for initiation and growth of south Florida's carbonate banks (Wanless and Taggart, 1989).

In addition, Florida Bay contains several different bank types. In central Florida Bay the banks are rather narrow (a few hundred meters wide) and are dynamic, actively migrating features. In central Florida Bay display southward- and westward-dipping stratigraphic units, indicating that the banks are migrating (fig. 6, cross section A–A'). In contrast, the mudbanks in western Florida Bay are as much as several kilometers across, and in stratigraphic cross section (Wanless and Taggart, 1989) it can be seen that bank size is due to the coalescence of several smaller core banks of layered mudstone resulting from infilling of the shallow interior basin (fig. 6, cross section B–B'). Bank coalescence results from excess sediment input from the shelf to the west (James, 1984) and more intense sediment production associated with more open marine waters (Wanless and Taggart, 1989). These individual and coalesced banks would be extremely difficult to recognize in outcrop or core bits of ancient rock sequences because of the scale (1–2 m (3–6 ft of relief over 0.5–3 m (0.3–10 ft that the recognition of such bank sequences faces the nature of sedimentary processes, produces a spatial variability of diagenetic potential, and defines spatial continuity and variability in relation to eko-

nomic resource distribution. Variability includes intrablue and intercycle variations in facies type, morphology, and growth habit with orientation of platform margin, coastline, or sediment body; intercycle variations in lateral facies shifts with variations in orientation; variations in early diaginesis with orientation and variations in emergent topography with orientation.

EXTRACTION OF VARIABILITY IN TIDAL PLATFORMS

An excellent example of how facies vary with position and orientation on a platform is seen in contrasting the tidal flat deposits of Andros Island, Great Bahama Bank; Caicos platform, British West Indies; and Club Cay, Great Bahama Bank. On Andros Island, the tidal flats face the platform interior to the west and northwest. Northwest wind pulses behind winter cold fronts provide repetitive bursts of anhore flooding of sediment-laden water onto the flats (Hartley, 1977). As a result, there are well-developed shore and channel-margin levees in which millimeter-thick laminas dominate (Hartley, 1977; Wanless, 1967). These levees and laminae are the resultant frequent small sedimentation events (1–20 per year) that create a high-precision profile of algal-bound millimeter-

thick laminae. This influence dies out inland across the flats, and the inner and outer islands marsh receive sedimentation events only during major storms. Sediments of the inner marsh are similar to those described from Caicos platform. The tidal flats of Caicos platform face the platform interior but, in contrast to Andros Island, face the south (Wanless, Tedesco, and Tyrell, 1988; Wanless et al., 1989). Northwesten winds, following the passage of winter storms, blow off the Caicos flats, and winter storms produce no depositional events. The tidal flats over most of Caicos receive sedimentation events only during hurricanes (average of 1 every 5.7 years; Neumann et al., 1979). Hurricanes deposit centimeter-

thick layers of carbonate grainstone or packstone and generate broad, low-relief bedforms (Wanless et al., 1988). The longer time between events permits a more mature algal-

cyanobacterial marsh community and a thicker organic mat over tidal flats across nearly the entire flat. Orientation of a platform interior thus defines bedform character and organic content of the sediment.
Figure 6. Zones of different dynamic growth history of carbonate muddocks and coastal sand dunes caused by differences in sediment supply, energy, and sea-level history in Florida Bay. Stratigraphic cross section A-A', across Russell Bank in central Florida Bay, is a migrating bank dominated by layered carbonate mudstone. Stratigraphic cross section B-B', across Dildo Key mudflats in western Florida Bay, shows that this broad bank initiated as four core banks of layered mudstone. Subsequent sedimentation filled the shallow bays in between, creating a broad coalesced bank. Map and cross sections are from Winter and Taggart (1989).
a carbonate-tidal flat. Similarly, the southwestern portion of Andros Island faces southwest, is not a focus of frequent sedimentation events by winter storms, and has a broad, continuous shore levee built by less frequent hurricane flooding events.

On Chub Cay a tidal flat occurring as a south-facing shore embayment, but one facing a narrow, sediment-starved platform margin. Here, there is an insufficient source of fine-grained carbonate sediment adjacent to the flat and the sediments on the flats is entirely red-marsh organics (Vanless, 1974).

**Example of Variability Resulting from Process Intensity** The spatial variability can result from one process, where the intensity of that process varies spatially. For example, the coarse-compensate deposits on the shelf of the Holocene (fig. 77) (Aigner and Rinkle, 1982) display dramatic changes in texture, bed thickness, structure, and abundance, from proximal overwash to sand bars to distal scattered sand layers in a shelf-basin.

**Influences on Sedimentation**

**Primary controls on sedimentation** include sea-level dynamics, sediment supply, topography, and physiographic setting; relative importance of prevailing energy, water storms, hurricanes, ocean swells, and tides; tectonic forces, palaeo-environments, and climate, especially rainfall and temperature. These primary controls define a variety of secondary controls, including erosion and deposition rates, biologic sediment production, and topographic processes. The primary controls and secondary controls act in concert to produce the modern carbonate platform sedimentation patterns.

**Secondary controls** on the carbonate platform include the processes of sedimentation and sedimentation processes. It is much more difficult to translate to mass accumulation than accumulation.

**Example of Sea-Level Influence** Wright and Coleman (1975) offered a model, which has become widely used, to explain deltaic form and the varying nature of delta sequences. In this model they show the fluvial sediment discharge as a series of waves (fig. 77) (Coleman and Wright, 1975) to determine the delta morphology and the sediment type. A primary delta from the birdfoot-delta of the Mississippi River as the muddy end member to the stacked sand-plain complex of the Sao Francisco River in Brazil as the coarse end member. They did not consider the role of wave level.

**Example of a Coastal Plain** Wright and Coleman (1975) have subsequently shown that the Sao Francisco delta is one of the most important coastal-plain wedges on the Brazilian coast and that the strand plains reflect sea-level history more than interaction of river discharge and ocean processes (fig. 4c).

More important, these strand plains occur along a coastal zone that does not have especially high energy, some strand plains are not directly associated with river discharge areas (fig. 4b) (Dominguez et al., 1987, p. 121), and the sand building the strand plains is largely derived from cannelization of the inner shelf during the last 6,000 years as a result of the relative sea level lowering along the coast of Brazil (reverse of Brauns's rule during rising sea levels) (Dominguez, 1987; Dominguez et al., 1987). The Sao Francisco delta is basically a prograding strand-plain deposit that formed under conditions of lowering sea level.

If you miss an influential primary control on sedimentation, your interpretation (and model) can be completely misleading.

**Example of Proximate Topographic Influence** Another example on influences on sedimentation comes from Florida Bay. The pattern of modern carbonate mudflats correlate extremely well with irregularities in bedrock topography (Conwell, 1989). D'Alvi (1980) and Taggart (1989) have documented subtle topography on the surface of the underlying Pleistocene limestone--bread along the northeastern margin of Florida Bay. In coastal areas, a coastal storm levee deposit formed in various positions of sea level (Conwell, 1989). Taggart (1989) and Vanless and Taggart (1989) have shown that, during the last Holocene transgression, mangrove and freshwater peats filled the lagoon troughs, and coastal levees formed a retreating succession of reingtune--sea build-ups along the northeastern margin of Florida Bay. Continued transgression the intertidal marsh deposits were overgrown by marine waters, leaving the more resistant elongate peat deposits and storm levee marks as emergent peninsulas extending into Florida Bay. Many of these peat peninsulas and coastal buildups have been dissected by coastal erosion. The emergent cap to most peninsulas, coastal buildups, and islands has migrated southward and westward in response to winter storms (Conwell, 1989).

These dissected and partly eroded peat and coastal deposits have also served to define and initiate most of the subtidal carbonate mudflats within Florida Bay (Conwell, 1989; Taggart, 1989). Vanless and Taggart, 1989). This is an example of the dramatic control on sedimentation by even subtle preexisting topography. This influence is one that is not easily recognized in modern environments and would be extremely difficult in ancient rocks. Yet without the underlying topographic influence, the mudflats may not have formed.

**Example of Topographic and Physiographic Setting** The oolitic sediment bodies of the Bahamas provide a different influence of preexisting topography. There, in the east coast, the oolitic sediment bodies formed because of protection and current flow defined by preexisting topography (fig. 8) (Vanless et al., 1989). Two oolitic
grainstone banks have formed down-current from emergent islands on the windward side of Caicos platform (behind South Caicos and Ambergris Cay fig. 8). Cross-platform current flow has streamlined these oolitic sands into elongate drumlins (fig. 8, area 1 and 3). The southerly bank (Ambergris shoal: area 1 in fig. 8) extends across the entire platform, shedding sediment off the leeward platform margin (area 2 in fig. 8). The position of an emergent feature on the eastern side of Caicos platform defines not only the position of sediment bodies on the platform but also the focus of off-bank sediment transport on the opposite side of the platform. The northerly bank (nib platform shoal: area 3 in fig. 8) separates a protected peloidal packstone-wackestone lagoon to the north (area 4 in fig. 8) from the grainstones of the more open platform.

Near the northern and western margins of Caicos platform, emergent ridges of oolitic sand formed during the middle Pleistocene. Extensions of these ridges have been gradually closing off the leeward (western) margin during the late Pleistocene and Holocene (fig. 9) (Wanless et al., 1989). These ridges have permitted extensive barrier reef growth on the reward margin during high sea-level stands by blocking the east to west transport of sediment across the bank, have initiated new sediment facies on the platform interior (e.g., the windward platform interior progradation of oolitic sand plains by production of ooids in the shore zone (B. Floyd et al., 1987), and have completely modified sedimentation in the adjacent basin by blocking and diverting cross-bank and off-bank transport (Wanless et al., 1989).

Example of importance of ecologic communities Afinal example is the influence of organic deposits of mangrove peat that have formed because of the favorable influence of primary controls of sedimentation. In south Florida mangrove communities form a coastal swamp that ranges from a few hundred meters to tens of kilometers in width (Davis, 1980; Wannebo, 1974; Wanless et al., 1989). The red mangrove community has produced extensive mangrove peats beneath these swamps. These peats form a major sediment body as much as 8 m (26 ft) thick and 0.1-1.5 km (0.06-0.9 mi) wide. This mangrove peat deposit is a significant sediment body, separating the marine and freshwater depo- sitions and environments and receiving the bottom of hurricane energies. It produces acidic tannic waters that corrode and dissolve both associated and underlying Holocene carbonate sediments and Pleistocene limestones and has deep-penetrating acidic root systems that can generate breccias in the Pleistocene limestone 2.5 to 7.16 ft beneath the sediment surface (Wanless et al., 1989). Most important, through bacterial oxidation and/or sulfate reduction, the mangrove peat deposits will most probably be obliterated and will become part of the sedimentary record (Wanless et al., 1989). We are used to this concept with evaporites. It is just as important for this swamp deposit.

Interaction of controls on sedimentation Interaction of primary controls on sedimentation commonly generates facies characteristics that would not be formed independently. Incorporation of two independent variables into a model may be insufficient; their interaction may have to be defined. For example, sediment accumulation or loss rates and sediment texture are defined by the interaction of (1) rate of sediment dispersal by high-energy events and platform
width, (2) potential sediment circulation by energy events and openness of platform margins, (3) rate of sediment production or supply and shallow water bioerosion (loss of coarse particles) or dissolution (loss of fines), (4) energy potential for physical sediment transport and stabilization by vegetation or cementation, and (5) surficial sediment supply and subsurface excavation of the sequence by deep burrowers and or subaerial in-fillings of channel networks with sediment.

**Example of vegetative modification of sediment transport.** The decrease or increase in energy that permits the establishment of a sea grass (or eelgrass) community is typically sufficiently mild so that it would not, in and of itself, cause a significant change in bottom sedimentation or in the benthic environment. Once established on the bottom, however, sea grasses will function to significantly modify wave and current energy at the bottom (Scofield, 1970), the stability of the bottom (Pattigquin, 1975; Wanless, 1981), Wanless et al., 1989), the texture of the bottom sediments (Pattigquin, 1975; Wanless, 1981; Wanless et al., 1989); the epifaunal and infaunal benthic community (Wanless, 1981); and the vertical sequences generated (Wanless, 1991; Wanless et al., 1989).

**Example of sediment dispersal and platform width.** Within a setting of uniform energy, platform (or shelf) width may have a dramatic influence on sediment texture. Observations by Bagnold (1966), McCave (1971), Passorga (1984), and Southard and Bogucki (1990) suggest that we consider four dynamic populations of grain size:

- 630 μm: tends to move as a true bedload
- 175–650 μm: mixed bedload and saltation load
- 40–175 μm: short-term or graded suspension load (drops out as current energy stops or detaches from bottom)
- <40 μm: long-term suspension (remains in suspension for some time after currents stop or detach from bottom because of ambient turbulence in the water column)

On narrow shelf and platform margins energy events may be sufficient to move all these textural populations of sediment off the platform before final deposition. With increasing platform width, the coarsest textural group will be partly and then completely re-accumulated on the platform under the same...
intensity and frequency of energy events. As platform width further increases, the finer-grained populations will be partly and then completely retained on the shelf. Thus, on the interior of extremely wide platforms, such as the Cambrian cratonic margins, the fine silt and clay fractions may be nearly completely retained in the shallow-marine environment simply because the platform width is too great for effective sediment dispersal. Similar variation also could be achieved with a constant platform width and varying frequency and intensity of reworking, rate of sediment input, and presence of a seaward barrier to sediment dispersal. These factors interact to define the rate of accumulation and the texture of the deposits.

**EXAMPLE OF EVOLUTION.** The carbonate reefs of Biscayne Bay provide another example of the interacting influences of controls on sedimentation (fig. 10). Burton (1984) and my observations. Marine organisms in Biscayne Bay produce aragonite, high-magnesium calcite, and minor low-magnesium calcite. The sediments on the mudbanks forming the seaward boundary of Biscayne Bay contain predominantly high-magnesium calcite and aragonite. These components gradually decrease along a landward transect across the bay until only low-magnesium calcite is present in the landward 1–3 km (0.6–5 mi) (fig. 10). Although there is both biogenic production and storm dispersal of aragonite and high-magnesium calcite throughout this open bay, the inshore waters are seasonally of low salinity and tannin rich. These waters, together with decay processes in the upper part of the sediment sequence, promote selective dissolution of the less stable carbonate components. This dissolution is visible in the partly dissolved and crumby skeletal macrofauta. Thus, minor variation in the waters across Biscayne Bay is producing sediments with strikingly contrasting mineralogies, different rates of sedimentation, and different diagenetic potentials.

**Limits of depositional controls.** Physical, biogenic, and chemical controls have discrete but subtle temporal and spatial boundaries to their influence. For example, winter storms, hurricanes, and other physical controls occur only over certain geographic ranges, and this range changes with geologic time. Subtle changes in water chemistry, temperature, and residence time define limits for various types of carbonate sedimentation and diagenesis (e.g., ooid production, bioturbation, intragranular and intergranular cementation).

**EXAMPLE OF SPATIAL LIMITS.** Hurricanes and winter storms are dramatic influences on sedimentation, but both have
defined spatial and temporal limits. Shallow-marine Holocene sedimentation in the Caribbean, Bahamas, or the Atlantic and Gulf coasts of the United States is strongly influenced by hurricanes (Neumann et al., 1978). The influence is present throughout these areas, although the statistical frequency and intensity spectrum vary from place to place. Shallow-marine Holocene sedimentation along the eastern coast of Brazil, however, is not influenced by hurricanes or winter cold fronts (Lee, 1982). There are large areas on the globe today where hurricane-type storms do not occur. The character of a reef sequence in Brazil (Lee, 1982) will be different from one in Bermuda, Bahamas, south Florida, and the Caribbean (e.g., Ball et al. [1967]).

EXAMPLE OF TEMPORAL LIMITS. Barron (1989) suggests that there were times in the Earth's history, such as the Eocene, when hurricanes were totally absent because the oceanic temperatures were not sufficient to maintain intense cyclonic circulation and perhaps other times when warm sea surface temperatures promoted more frequent and intense hurricanes than those occurring today. Barron also suggests that an absence of landmasses in polar latitudes may have limited the development of winter storm systems during certain periods (e.g., the Cretaceous). Winter storms are today a major influence on some carbonate depositional environments. The physical dynamics and sedimentary structures that characterize the channelled tidal flats on Andros Island, Bahamas, and the subtidal mudbanks in Florida Bay are largely a product of winter storms sedimentation (Ginsburg, 1956; Hardie, 1977; Eno and Perkins, 1979; Wanless and Taglerrt, 1989). What would be the nature of these depositional environments in the absence of winter storm and hurricane influences?

It is important to consider other high-energy-event influences in addition to or instead of local storms. Long-period
swells moving across the ocean from distant storms may cause a sedimentation event on shelves, reefs, and shorelines equal to or greater than locally generated, shorter-period storm waves (Wanless et al., 1989). Similarly, tectonic activity generates intense local surges and far-reaching tsunamis.

**Example of need for empirical verification:** Modeling of the nature and distribution of storm influences needs empirical verification. King (1990) recently suggested (through modeling) that the Miocene Mississippi Waukaraian carbonate mud mounds formed in a time and area where storms were absent—low latitude on the margin of the Laramian shelf or seaway. Previously, Wright (1986) had documented (through field studies) that these ramped shelf sequences (in southeast Wales) were dominated by storm processes but that the Waukaraian mud mounds were positioned on the deeper outer shelf below storm wave base. It is fundamental to empirically test and calibrate models in the field for at least use the information that is provided in the literature.

**Early- to late-stage diagenetic modifications:** Diagenetic modifications include syndepositional biogenic modifications through repetitive burrow excavation and infillings causing transformation of an existing deposit by subsurface generation of facies with new sediment composition, texture, permeability, and diagenetic potential; loss of transgressive facies (Wanless and Taggett, 1986, 1989; Taggett, 1989); amalgamation of sedimentary cycles (Wanless and Tedesco, 1987); and reworking, transformation, or obliteration of the sediment facies that initiated or constructed a sediment body and later stage compaction and pressure dissolution that reduces depositional thickness and modifies the fabric, mineralogy, and porosity and permeability of sequences.

The profound influence of diagenetic modifications on the nature of the sedimentary sequence is illustrated by several examples.

**Example of transformation of existing deposit:** Excavating burrowers provide a dramatic example of the influence of early diagenetic modifications on both the nature of the preserved sequence and the subsequent diagenetic potential. Deep burrowers, such as Calianna resorts, are not mixed with the sediment; they are excavators of extensive open cavities that extend one to several meters beneath the sediment surface and may comprise as much as 10% of the sequence. Storms may produce catastrophic infilling of these open networks (fig. 11) (Wanless, Tedesco, and Tyrell, 1988). Repetitive excavation and infilling of the structure can produce a sequence or portion of a sequence that is a new deposit produced in the subsurface by repetitive tubular transgressive infillings, completely replacing the precursor sequence, and has a depositional sediment fabric formed in the subsurface with only indirect relation to the syndepositional or precursor surficial depositional texture (Wanless, Tedesco, and Tyrell, 1988; Wanless et al., 1989; Tedesco and Wanless, 1991). Failure to recognize this will cause misinterpretation of the depositional environments and processes and misrep
are immense. In the Glen Rose Limestone (Lower Creta-
cene), for example, fillings of *Thallusinosidae* are important
unlined burrow excavation structure extending back to the
Cambrian (Groszer and Brojér, 1988; Kjær, 1981), have
generated or transformed significant portions of the sequence
(Walless and Ts-dos, 1987). The Glen Rose contains nu-
merous meter-scale shallowing-upward sequences. In many
of these, excvation and infilling have transformed the basal
transgressive portion of the cycle. If the regressive (peritidal)
cycle cap is not hardened by subaerial exposure or marine,
cementation, burrowers of the next cycle of sedimentation
party to completely obliterate the cycle boundary, causing
local amalgamation of cycles.

**Example of Obliteration of Original Facies**

In the Holocene of Florida Bay, peritidal coastal mangrove peat
and storm levee marls provide the initiation for many of the
subtidal mudbanks (see last example [preexisting topogra-
phy] on p. 47) (Walless and Taggert, 1989). As sediment
bodies, many of the mudbanks are simply transgressed coastal
leaves and mangrove peat deposits (Connell, 1989). In the lower
parts of Florida Bay this transgression is commonly well
preserved in bank stratigraphy because fluctuating salinity
conditions inhibit a deep burrowing fauna. In central and
outer Florida Bay, however, deep-exca
dinating burrowers are
present and have transformed these transgressed coastal peat
and marl deposits into marine mudbanks (Taggert, 1989) (fig.
17). That is, some mudbanks are sediment bodies generated
as coastal peat levee deposits but were subsequently trans-
formed into marine mudbanks! Where completely trans-
formed, the only evidence of precursor facies is the coarse
skeletal concentrates of freshwater, paralic, and brackish
d fauna in lower burrow infills (Taggert, 1989; Walless and
Taggert, 1989). The origin of significant sediment bodies can
be obscure.

Similarly, many of the modern carbonate mudbanks build and
expand by event deposition—layered mounds in the banks of
the conical lagoons and back to the grainstone tempestite units in the more exposed banks. As the banks build to sea level, the rate of vertical accretion of the bank
interior (bank core) slows, and the underlying sequence can
be partly to completely transformed by burrow excavation
and storm infilling. Layered mounds, packstones, and
grainstones are commonly transformed into bioturbated ske-
letal packstones by repetitive burrow excavation and infill
(Fig. 14) (Trabucco and Walless, 1991; Walless and Taggert,
1989). Such transformations also determine the sediment
texture, fabric, and composition during the gradual vertical
accretion of the cove facies of modern (and ancient) mudbanks.

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**Figure 12.** Core sequences taken bankward from the eolian
south-facing tidal-flat coastline of North Calico (see star in fig. 8
above 'A' for location). Core sequences record progressive trans-
formation of the transgressed tidal-flat deposit into a marine deposit
by repetitive excavation and infilling of Callionymus burrow tubes
and chambers. Coarse tidal-flat fauna, not expelled by excavation,
gradually worked toward the bottom of the sequence during this
transformation (from Walless et al., 1989).

**Figure 13.** Sequence of eolian cross sections of a coastal levee
at the northern margin of Florida Bay as it becomes inundated by
rising sea level and is gradually transformed into a marine mudbank
by repetitive excavation and infilling of burrow complexes. As in
fig. 12, the only evidence that this sediment body originated as a
coastal levee buildup might be an anomalous coarse fauna concen-
trated at the base of the sequence.

**Figure 14.** Biogenic transformation across a prograding carbon-
ate mudbank in Florida Bay. Bank growth is primarily by deposition
of layered mudstone units (some with an eolian base and a
coarser lower part and/or stabilization by seawater before deposition
of the next layer of mudstone). The sequence undergoes progressive
biogenic transformation, forming skeletal wackestone or packstone
by repetitive excavation and infilling of deep, open burrow com-
plexes and resulting in the complex facies patterns.
EXAMPLE OF BIVALVE EXTREMES - The New Market Lime- stone (Ordovician) of western Maryland contains striking carbonate cycles of sedimentation (Reinhardt and Hardie, 1976; Wanless, 1986). Figure 15 shows one measureable cycle of sedimentation recorded from bottom to top as underlying supratidal dolomitic deposits (p); a deepening peritidal to shallow subtidal sequence of high-relief stromatolitic sheets with planar millimeter-thick laminae to high-relief postural stromatolites (thrombolites) to microthickened of my corals (Lichenaria, Fletcheria, and Prinomya); Neuman, 1951) in a thinly interbedded limestone and dolomite with burrow structures, gastropods, and other skeletal grains. A (1) to B (2), a “deep” sequence of interbedded limestone and dolomite that contains no macrofauna and no trace fossils (D); a shallower subtidal sequence of interbedded limestone and dolomite in which gastropods, other skeletal grains, and burrowers reappear (E); a shallow shelf to peritidal sequence of thinly interbedded limestone and dolomite with microthickened corals to low-relief postural stromatolites (thrombolites) to low-relief stromatolite sheets with millimeter thick laminae to decalcification cracks in low-relief stroma- tolic sheets with millimeter thick laminae (F→G→H), and increasingly dolomitic supratidal deposits (l and J). Note that the form of the stromatolites is high relief during the deepening (transgressive A→B) phase and low relief during the shallowing (regressive G→H) phase (Wanless, 1986).

The described sequence forms an essentially symmetric cycle with the abiotic interbedded limestone and dolomite zone (D), representing the deepest portion of the cycle. This unit presumably represents a time of aoxic bottom waters because of the lack of macrofauna and lack of burrowers. The gastropods in the overlying units reflect the reappearance of algae as a bottom food source, and the burrowers reflect oxygenated bottom waters.

The vertical distance from the deepest portion of the cycle (D) to the overlying stromatolites with millimeter-thick laminae (base of H) is 1.8 m (5.9 ft); to the first decalcification cracks (middle of H) is 2.1 m (6.9 ft). If this represents the valid thickness of sediment accumulation from the deepest subtidal anoxic sedimentation to the supratidal, a fluctuation of sea level would seem necessary to generate this cyclic sequence. Anoxic bottom environments are not at all common in modern carbonate environments ≈ 6 depths of 1.8–2.1 m (5.9–6.9 ft). There are, however, two critical flaws with this reasoning.

First, in evaluating cyclic sequences, it is critical to determine the increase in thickness that has occurred because of later diagenesis, especially pressure dissolution. In the described cycle, a bed-by-bed examination was made to determine the minimum pressure dissolution that has occurred by saturated sea bottom dissolution, nonsaturated sea bottom dissolution (sulfate dissolutions), and pervasive dissolution and dolomitization (Wanless, 1979). Thinning associated with larger surface detritus were defined by height in outcrop. thinning because of smaller seams, nonsaturated seams, and pervasive dissolution thinning was assessed in thin section and by outcrop measurements of bed thinning. The result is that the sequence has thinned by at least 50% because of pressure dissolution (see Fig. 15). Thus the sequence from “depths” to peritidal was at least 3.4–4.3 m (11–14 ft). Anoxic conditions occur in a number of modern coastal lagoons, bays, and estuaries at these water depths (Reynolds, 1984; Wanless et al., 1984). Thus, by correcting for thickness reduction by subsequent pressure dissolution, other mechanisms for cycle generation become available for consideration. Sea-level fluctuations are not required to produce the observed cycles from the point of view of thickness of cycles.

Second, during and just after times of transgression and rising sea level, large volumes of nutrient can be expected to be released from coastal freshwater, paralic, and marine swamps, marshes, tidal flats, and upland soils (Hall and Schlager, 1986). This may provide a phase of high nutrient stress to the coastal lagoons, bays, and estuaries. Thus, during and just after transgression, anoxia might form in shallower bays than would be the case during times of sea-level stability and regression. Study of modern environments, however, would not pick up this type of anoxic bottom because sea level has been stable to only gradually rising for the last 3,000 years (Wanless, 1981) and the protected coastal and adjacent freshwater environments mostly have not been transgressive (Wanless et al., 1989). Nutrient excess, even if not producing anoxia, is well recognized on carbonate production and even on the survival of carbonate platforms (Hall and Schlager, 1986).

Scallop

One must be careful in applying influence, patterns, and rates from one scale of sedimentation to another. Many fundamental controls on sedimentation have different effects on different scales of sedimentary cycles. The time scales of sedimentation may not be conducive to predictive modeling, especially if the driving forces are episodic.

Holocene analogues represent sedimentary cycles invarious stages of completeness. Integration with Pleistocene sequences are a first step toward assessing the duration necessary for generating complete fundamental depositional sequences and assessing usefulness of the Quaternary as analogues to times of dampened eustatic driving forces.

Summary

Most modeling efforts attempt to incorporate those influences that the modeler recognizes as important. It is hoped that this article has increased the modeler’s awareness of influences on sedimentation, shown the importance of defini-
Figure 15. Characteristics of one typical fundamental depositional sequence in the Ordovician New Market Limestone, western Maryland (in pasture 0.5 km southwest of St. Paul's Church, US-40, Washington County). The left-hand side is measured section and distribution of petrofacies characteristics. Units A to J are described in the text. The right-hand side is the uncremented sequence obtained by expanding the section for the minimum observed thinning by pressure dissolution. Minimum water depths are indicated by the solid and dashed lines adjacent to the measured and uncremented sections (figure expanded from Wilkins [1986]).


