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HIERARCHY OF SEDIMENTARY DISCONTINUITY SURFACES AND CONDENSED BEDS FROM THE MIDDLE PALEOZOIC OF EASTERN NORTH AMERICA: IMPLICATIONS FOR CRATONIC SEQUENCE STRATIGRAPHY

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ABSTRACT

Sedimentological analyses of middle Paleozoic epeiric sea successions in North America suggest a hierarchy of discontinuity surfaces and condensed beds of increasing complexity. Simple firmgrounds and hardgrounds, which are comparatively ephemeral features, form the base of the hierarchy. Composite hardgrounds, reworked concretions, authigenic mineral crusts and monomictic intraformational conglomerates indicate more complex histories. Polymictic intraformational conglomerates, ironstones and phosphorites form the most complex discontinuity surfaces and condensed beds. Complexity of discontinuities is closely linked to depositional environments duration of sediment starvation and degree of reworking which in turn show a relationship to stratigraphic cyclicity.

A model of cratonic sequence stratigraphy is generated by combining data on the complexity and lateral distribution of discontinuities in the context of facies successions. Lowstand, early transgressive and late transgressive systems tracts are representative of sea-level rise. Early and late transgressive systems tracts are separated by the maximum starvation surface (typically a polymictic intraformational conglomerate or condensed phosphorite), deposited during the peak rate of sea-level rise. Conversely the maximum flooding surface, representing the highest stand of sea level, is marked by little to no break in sedimentation. The highstand and falling stage systems tracts are deposited during relative sea-level fall. They are separated by the forced-regression surface, a thin discontinuity surface or condensed bed developed during the most rapid rate of sea-level fall. The lowest stand of sea level is marked by the sequence boundary. In subaerially exposed areas it is occasionally modified as a rockground or composite hardground.

RESUME

Les analyses sédimentologiques de successions de mers épicontinentales du Paléozoïque moyen en Amérique du Nord suggèrent une hiérarchie de surfaces de discontinuité et de couches condensées de plus en plus complexes. Des surfaces fermes (firmgrounds) et des surfaces durcies (hardgrounds), qui sont des caractéristiques assez éphémères, forment le fond de l'hiérarchie. Des hardgrounds composées, des concrétions modifiées, des croûtes de minéraux authigènes et des conglomérats intraformationnels monomictiques indiquent des histoires plus complexes. Des conglomérats intraformationnels polymictiques, des cailloux ferrigineux et des phosphorites forment les surfaces de discontinuité et les couches condensées les plus complexes. La complexité de discontinuités est liée prochement à un environnement de dépôt, surtout quant à la durée de privation sédimentaire et au degré de modification. Ceux-ci indiquent eux-mêmes une relation aux cycles stratigraphiques.

On obtient un modèle de stratigraphie de séquence cratonique en combinant des données sur la complexité et la distribution latérale de discontinuités quant aux successions de faciès. Des prismes de bas niveau marin, de première transgression et de transgression tardive représentent une montée du niveau marin. Les prismes de première transgression et de transgression tardive sont séparés par la surface de privation maximale (typiquement un conglomérat polymictique intraformationnel ou une phosphorite condensée) déposée pendant la vitesse maximale de la montée du niveau marin. D'autre part, la surface d'inondation maximale, qui représente la plus grande élévation du niveau marin, est marquée par peu ou pas d'interromption de sédimentation. Le prisme de haut niveau et celui de la chute sont déposés pendant la chute relative du niveau marin. Ils sont séparés par la surface de régression forcée, une surface mince de discontinuité ou une couche condensée développée pendant la chute la plus rapide du niveau marin. Le niveau marin le plus bas est marqué par la limite de la séquence. Dans des régions exposées à l'atmosphère, il est parfois modifié et devient une surface rocheuse (rockground) ou une surface durcie (hardground) mixte.

INTRODUCTION

The continued development of sequence stratigraphy over the last few decades has encouraged a re-evaluation of numerous sedimentary successions. However, a general sequence stratigraphic model specifically applicable to epicontinental successions has not been formally proposed outside of that designed for the siliciclastic margin of foreland basins. More commonly, sequence stratigraphic models have been developedfor specific time intervals and/or geographic regions. Sequence stratigraphy is aptly termed a stratigraphy of surfaces. Thus we present here first a discussion of the various discontinuity surfaces and condensed beds in cratonic successions. We then combine these data with that of the enclosing facies successions to generate a general sequence stratigraphic model applicable deposits of epeiric seas, especially of foreland basins.

Assessing the relative temporal significance of discontinuities and diastems is critical to generating accurate paleoenvironmental reconstructions (Loutit et al., 1988; Brett, 1995). Unfortunately, the resolution of most geochronologic techniques is generally too low for such work, especially in Paleozoic and older rocks. However, taphonomy, sedimentary geochemistry and cyclostratigraphy provide techniques for assessing the relative 'time-richness' of discontinuities (etc. Fürsich, 1978; Kidwell and Bosence, 1991). Such studies, as discussed below, supply evidence that a spectrum of time-averaging is present in many cratonic successions. Hardgrounds, concretions, ironstones and phosphorites are among the most readily identified indicators of condensation and/or erosion on the seafloor. Examples provided here are primarily taken from analysis of middle Paleozoic strata in eastern North America, a period of calcite seas (etc., Stanley and Hardie, 1998; Wilson and Palmer, 2004) during an overall greenhouse interval (Frakes et al., 1992). Throughout this region Upper Ordovician, Lower Silurian and Middle Devonian strata can be traced from the craton margin, where they are composed mostly of siliciclastic-dominated foreland basin sediments, to the cratonic interior, where they are almost entirely carbonate (Fig. 1). This framework permits analysis of individual discontinuity surfaces across a range of sedimentary regimes. In particular, the high frequency and diversity of Upper Ordovician discontinuity surfaces gives a unique view into the short-duration end of the discontinuity hierarchy that is much more difficult to detect in other Paleozoic successions.

HIERARCHY OF DISCONTINUITIES AND CONDENSED BEDS

Analysis of a variety of discontinuity surfaces and condensed beds from the middle Paleozoic of eastern North America and additional examples from the literature forms the groundwork for the following hierarchy (from shortest duration to longest; Table 1).

FIRMGROUNDS

Firmgrounds are not fully indurated sediment layers on the seafloor with irregular surfaces produced by sharply defined burrows; they lack borings and encrusters (Fig. 2; Bromley, 1990). Simple firmgrounds include upper surfaces of fine-grained limestone beds with low densities of burrows (<2–3 cm deep) such as the *Trichophycus* ('turkey track') horizons common in mid-ramp storm beds throughout the Upper Ordovician of eastern North America (Osgood, 1970). More complex firmgrounds are typified by *Thalassinoides* burrow networks that may penetrate several successive beds. Complex firmgrounds are often closely spaced stratigraphically near major facies offsets and tend to be more laterally continuous than simple firmgrounds. They likely represent extensive periods of slowed sedimentation. Complex firmgrounds often occur as predecessors to composite discontinuity surfaces (discussed below). Both simple and complex firmgrounds have been described from throughout the Phanerozoic (etc., Ghibaudo et al., 1996; Bertling, 1999).

SIMPLE HARDGROUNDS

Hardgrounds are surfaces of synsedimentarily cemented carbonate layers that have been exposed on the seafloor (Wilson and Palmer, 1992). Simple hardgrounds are typified by: (1) minimal modification of surface topography (etc., no extensive boring or thick encrustation); (2) preservation of encrusters (suggesting abrupt burial); and (3) little to no evidence for multiple generations of encrusters (Fig. 3). Surfaces encrusted with *Sphenothallus* sp. (Bodenbender et al., 1989; Neal and Hannibal, 2000) are among the most common types of simple hardgrounds found in the Upper Ordovician (Maysvillian–Richmondian), though they are often overlooked because of their subtle character. Cemented shell pavements form a distinctive type of simple hardground, exhibiting thin crusts of bryozoans, crinoid holdfasts, edrioasteroids, sphenothallids, cornulitids etc. (Fig. 3). For example, a *Rafinesquina* shell pavement described by Meyer (1990) from the Upper Ordovician (Fig. 3) shows a simple hardground with exceptional preservation of edrioasteroids, which, as delicate, multi-plated organisms typically would have disarticulated quickly after death, as is characteristic of asteroids and crinoids. Thus the exposure time of this hardground was relatively short once it was colonized. The encrusted *Rafinesquina* vary from whole, articulated specimens to broken and slightly abraded valves. A few of the edrioasteroids occur directly on the matrix between shells indicating that it too was stabilized and partially cemented. Meyer (1990) noted a bimodal distribution of edrioasteroids and suggested that the shell pavement remained exposed

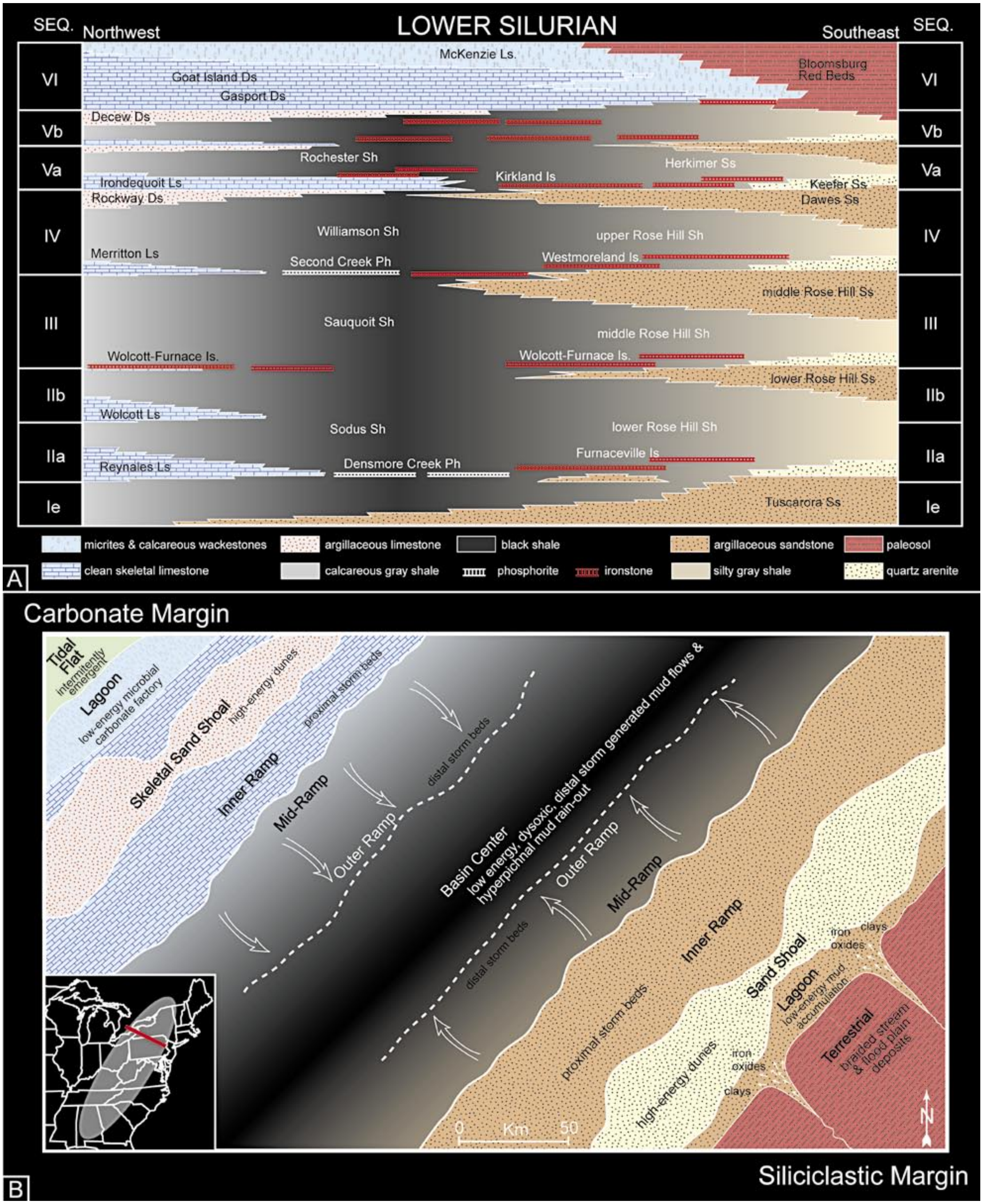

Figure 1. (A) Schematic cross-section from central Pennsylvania (southeast) to western New York (northeast) showing the Lower Silurian Taconic and Salinic foreland basins with facies belts roughly divided into a carbonate-dominated margin to the northwest, a mud-dominated basin center, and a (coarse) siliciclastic-dominated margin to the southeast. Ls=limestone; Ds=dolostone; Sh=shale; Ph=phosphorite; Is= ironstone; Ss=sandstone. (B) General distribution of depositional environments in the Taconic, Salinic and Acadian foreland basins (grey-shaded area of inset map marks position of strata from these foreland basins in eastern North America; cross-section line marked in red). Modified from Cotter (1988).

Discontinuity surfaces and condensed beds	Associated lithologies	Taphonomic features	Lateral Distribution	Commonly associated discontinuities/ Condensed beds
Simple firmgrounds	Form in fine-grained grainstone, calcisiltite or siltstone	Sharp contrast between burrow fill and surrounding sediment	1–10s km^2	None
Simple hardgrounds	Form on fine-grained grainstone and packstone blanketed by shale	Well preserved encrusters, low diversity and density	1–10s km^2	None
Concretions	Composed of cemented shale, calcisiltite, and/or skeletal stringers	Lack of encrusters, orientation parallel to bedding	10s–1000s km^2	Composite hardgrounds, authigenic crusts
Reworked concretions	Typically enclosed in shale or grainstone	Encrusters, borings, and/ or orientation oblique to bedding	10s–1000s km^2	Composite hardgrounds, authigenic crusts
Hiatus concretions	Typically enclosed in shale or grainstone	Multiple layers of encrustation and concretionary cements	10s–1000s km^2	Composite hardgrounds, authigenic crusts
Rockgrounds	Most common on fine-grained grainstone and micrite	karsting and/or diagenetic mottling cut by borings or encrusted, highly irregular (scalloped) erosion surface	10s–100s km^2	Authigenic mineral crusts, composite hardgrounds
Authigenic mineral crusts	Phosphate, glauconite, hematite, pyrite and/or ferric layer(s) on fine- to medium-grained grainstone	Occasionally show degraded holdfasts and/or interlayering of multiple generations of crusts and encrusters, borings rare–common	10s–1000s km^2	Composite hardgrounds, reworked concretions, mono- and polymictic intraformational conglomerates
Composite hardgrounds	Fine- to medium-grained grainstone	Encrusters showing a range of degredation and/ or mineralization, locally densely spaced borings	10s–1000s km^2	Authigenic mineral crusts, reworked concretions monomictic conglomerates
Modified firmgrounds	Form in/on fine- to medium-grained grainstone, micrite	Borings and/or encrustation of firmground	10s–1000s km^2	Authigenic crusts, mono- or polymictic conglomerates
Monomictic intraformational conglomerates	Composed of fine- to medium-grained grainstone or calcisiltite	Large angular or small rounded clasts of a single lithology showing encrustation and/or borings	10s–1000s km^2	Authigenic mineral crusts, composite hardgrounds
Polymictic intraformational conglomerates	Composed of fine- to medium-grained grainstones, calcisiltite, shale and concretions; typically cap skeletal grainstone	Small rounded clasts of multiple lithologies showing encrustation and/ or borings	1000s–10,000s km^2	Authigenic mineral crusts, composite hardgrounds, reworked concretions
Condensed phosphorites	Composed of phosphatic grains and cements; Enclosed within black shales or cap skeletal grainstone	Phosphatic ooids, phosphate and pyrite steinkerns, phosphatized and pyritized skeletal grains, teeth, bone, wood, and burrow tubes, chert, quartz sand-pebbles, limestone clasts	1000s–10,000s km^2	Grade into polymictic intraformational conglomerates and/or ironstones; associated with reworked and/or hiatus concretions

Table 1. Characteristics and distribution of discontinuity surfaces and event beds.

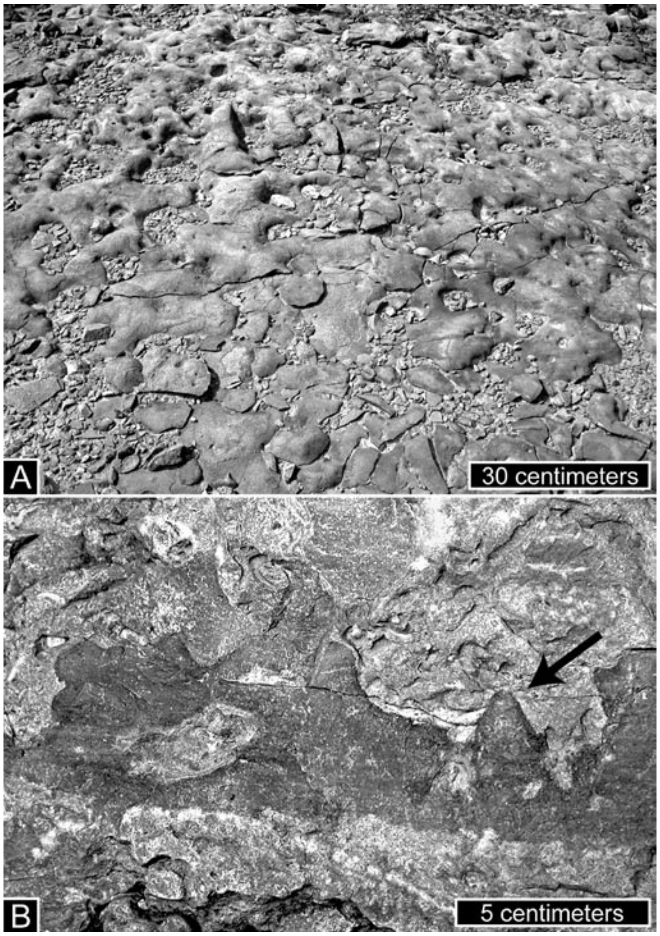

Figure 2. Simple firmgrounds. (A) Bedding plane of heavily bioturbated grainstone (Upper Ordovician Point Pleasant Member of the Lexington Limestone; Peaks Mill, Kentucky). (B) Cross-section of sharp-sided burrows (arrowed) in grainstone (Upper Ordovician Perryville Member of the Lexington Limestone; Salvisa, Kentucky).

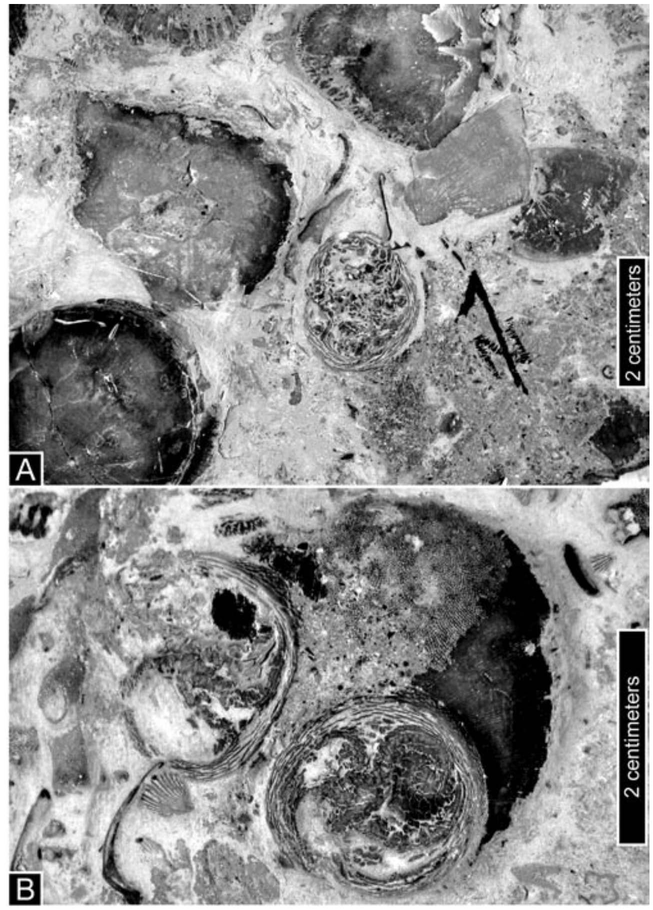

Figure 3. Simple hardground (Upper Ordovician Corryville Formation; Florence, Kentucky). (A) Edrioasteroid (left of north-arrow) attached to cemented sediment between brachiopod valves comprising a shell pavement. The remnant mudstone partially obscuring the hardground surface represents catastrophic burial of this layer. (B) Two edrioasteroids and a small bryozoan colony encrusting a brachiopod shell. The encrusters come close to abutting but do not overlap, suggesting they represent cohorts rather than multiple generations of encrusters. The short duration of exposure of this hardground is supported by the scarcity of borings.

for only the duration of two spat falls (up to tens of years). Commonly simple hardgrounds are aerially restricted, often not traceable beyond a single exposure, though in rare cases (as in the edrioasteroid example above) traceable for a few tens of kilometres. They are most commonly found at limestone–shale contacts, but may also occur within amalgamated limestone beds.

CONCRETION BEDS

Beds of simple ellipsoidal concretions of carbonate-cemented mudstone or siltstone are common in some mixed carbonate–siliciclastic successions. Concretions typically occur in grey to brown shales containing *Chondrites* trace fossil assemblages and thin, scattered laminae of skeletal hash (distal storm layers), suggestive of outer-ramp to deep subtidal deposits (Fig. 4). Concretionary horizons die out up-ramp into more proximal sections and down-ramp into black shales of the basin centre. Concretions are thought by some to form during periods of reduced sedimentation and accrete at a rate of a few centimetres per thousand years (etc., Canfield and Raiswell, 1991; Raiswell and Fisher, 2004).

In the Upper Ordovician of the Cincinnati region concretion horizons are most abundant within shale-dominated units (Fig. 4A; Brett et al., 2003a). However, not all mixed carbonate–siliciclastic successions grade down-ramp into shales containing concretions in this region; often concretions are absent even when very similar biofacies are present. The controlling factor appears to be lithologic: where shales are greater than 10 cm thick, concretions are much more common, whereas, concretions are almost absent within the Lexington

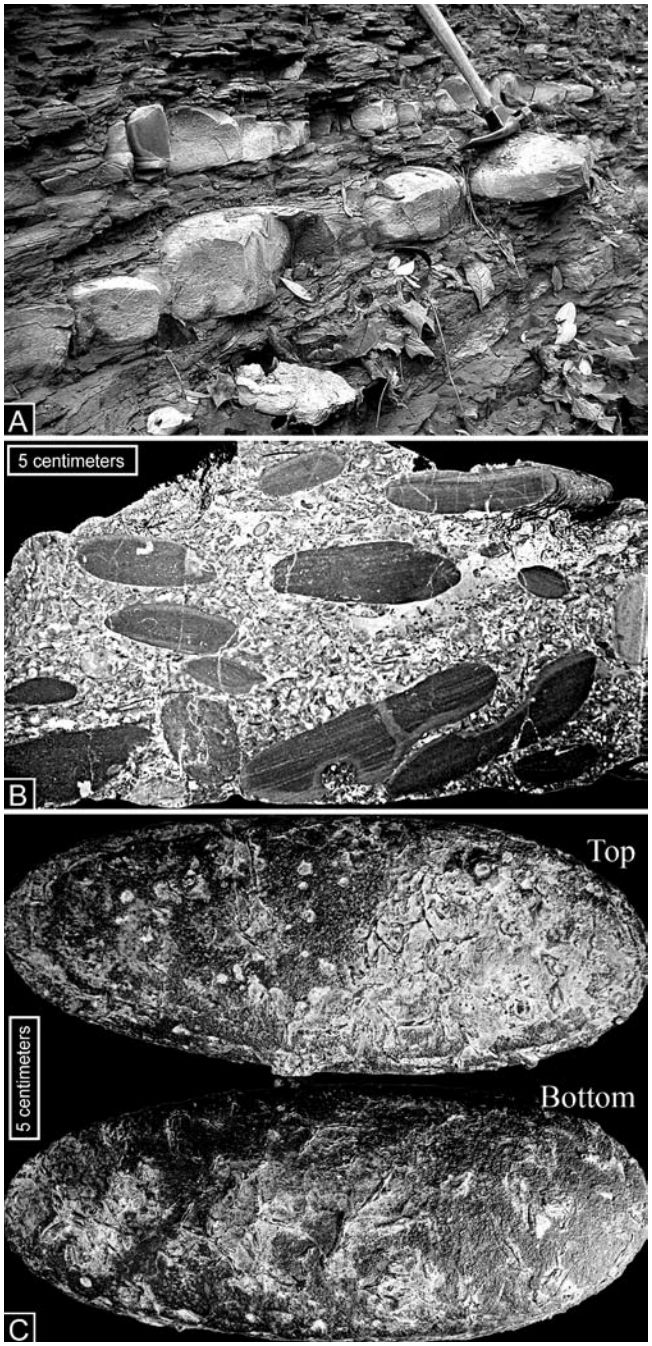

Figure 4. Concretions and reworked concretions. (A) Concretion-bearing horizons forming a nearly continuous bed in organic-rich shales (Upper Ordovician Kope Formation; Brent, Kentucky); hammer is 30 cm long. (B) Polished section showing imbricated reworked concretions in crinoidal grainstone (Upper Ordovician Kope Formation; Monterey, Kentucky). Many of these concretions are encrusted by bryozoans and crinoids on both their upper and lower surfaces. (C) Reworked, large ellipsoidal concretion encrusted by blackened alloporid corals and crinoid holdfasts and covered by a thin black phosphatic crust on both its upper and lower surface (Middle Devonian Chenango Member of the Skaniateles Formation; Moonshine Falls, New York).

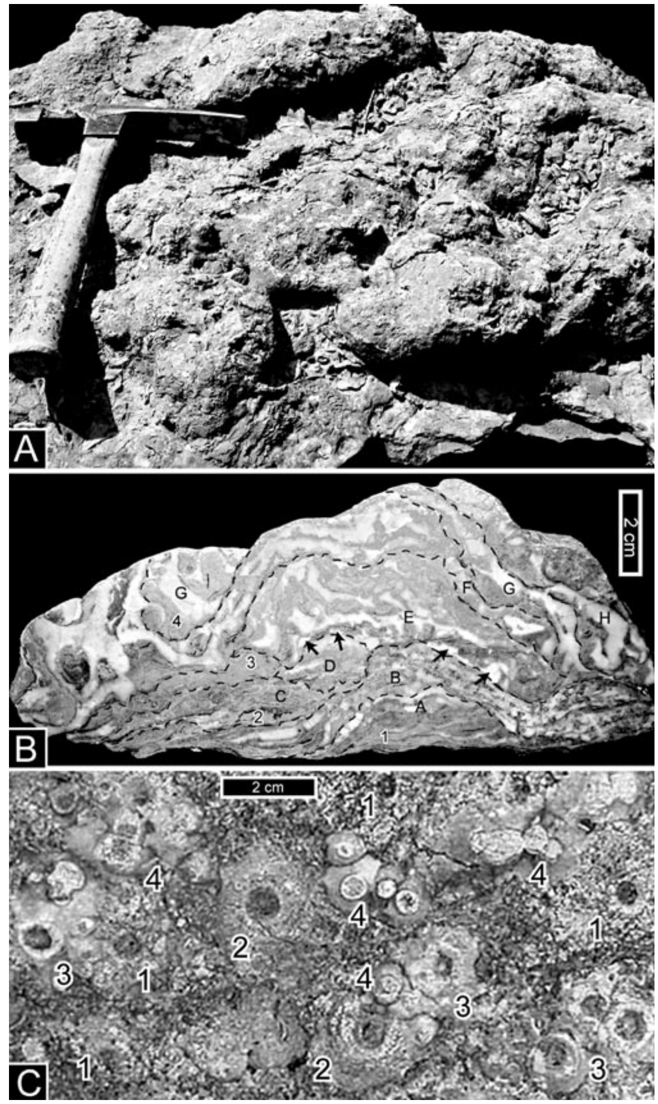

Figure 5. Composite hardgrounds. (A) Bedding plane of mounded bryozoan- and crinoid-encrusted hardground (Upper Ordovician Point Pleasant Member of the Lexington Limestone; Georgetown, Kentucky); hammer is 30 cm long. (B) Polished section through the encrusting layers of the hardground shown in A. This small mound records eight encrustation episodes (A–H) some of which are separated by mud burial events (1–4). (C) A crinoid-encrusted hardground at the same stratigraphic horizon 55 km to the northeast (Wollcott, Kentucky). Holdfasts are overlapping and variably preserved (grades 1–4), suggesting four encrustation events likely corresponding to those recorded in the mounded lateral equivalent.

Limestone, which typically only contains thin shales (<10 cm thick) within outer-ramp shale–marl successions. Concretions are much more common in the offshore facies of the overlying Kope Formation where limestones are commonly separated by tens of centimetres of shale. This may seem a paradox as it suggests that successions with relatively high siliciclastic sedimentation rates also contain some of the most easily recognizable evidence of sediment starvation in the form of concretions. The absence of concretions and abundance of marls in outer ramp facies of the Lexington Limestone provides support for the interpretation that marls form in a manner similar to concretions in settings of relatively low siliciclastic sedimentation. Therefore, marls may also be considered indicative of periods of slow sedimentation.

Numerous examples of the stratigraphic distribution of concretions in epicontinental seas, from the Middle Devonian (Baird, 1976), the Jurassic (Fürsich et al., 1992) and the Upper Cretaceous (Kauffman, 1977), seem to exhibit similar stratigraphic relationships. In the Miocene of Egypt concretion beds are paired with hardgrounds that formed within a few tens of centimetres above the concretion levels (Malpas et al., 2004). This pairing of concretions and hardgrounds is also seen in the Kope Formation (Brett et al., 2003a). The co-occurrence of concretions and hardgrounds in these sections may indicate concurrent responses to a period of sediment starvation; one response at the seafloor and the other within the zone of sulfate reduction just below the sediment–water interface.

COMPOSITE HARDGROUNDS

Composite hardgrounds show evidence of multiple encrustation events (Fig. 5), but typically contain only a single biofacies. Bioerosion, corrosion and abrasion on the seafloor all work to remove the record of encrusting organisms; the longer the exposure time on the seafloor the greater the degree of degradation and number of preservation states of encrusting organisms. Therefore, hardgrounds that show a variety of preservational states among adjacent encrusters suggest multiple episodes of encrustation, in some instances even indicating intermittent burial of the hardground surface.

Many composite hardgrounds occur within the Upper Ordovician (upper Mohawkian) of central Kentucky and southwestern Ohio (McLaughlin and Brett, 2007). These surfaces commonly display multiple overlapping layers of crinoid holdfasts and encrusting bryozoans developed upon a highly abraded and occasionally heavily bored and iron-mineralized fine- to medium-grained skeletal limestone (Figs. 5A–C). They display irregular upper surfaces, having relief of up to 3 cm, where developed upon firmgrounds. Encrusters show multiple grades of preservation, often from highly abraded to pristine in the uppermost layer (Fig. 5C). Colour alteration is also a common feature; for example, the best preserved crinoid holdfasts are pink whereas more heavily degraded encrusters are blackened. Consecutive layers of encrusting bryozoans and crinoids interspersed with thin mud drapes form mounds up to 8 cm high. In some cases, cessation of hardground formation resulted from rapid mud burial as indicated by the presence of large, well preserved edrioasteroids that cap the mounds. In places, cemented layers were undercut by erosion leading to the formation of composite hardgrounds along edges or bases of beds. Bryozoans and crinoid and holdfasts, borings and even edrioasteroids have been observed on such undercut, cavernous hardgrounds. Encrustation of such cavities, on an otherwise flat seafloor, is also described from the Jurassic (Palmer and Fürsich, 1974; Fürsich et al., 1992). It is not uncommon to find composite hardgrounds in which only elevated portions of the surface are heavily encrusted, suggesting a barrier to colonization of the lower areas, likely sediment cover (Palmer and Palmer, 1977; Brett and Liddell, 1978). Raised areas on upper Mohawkian hardgrounds in Iowa and Minnesota have been interpreted as formed by buckling in response to cement growth (Palmer, 1978), but alternatively may record deformation by earthquakes (see below).

Composite hardgrounds occasionally show evidence of firmground precursors (Fig. 6). In the upper Mohawkian, closely spaced *Thalassinoides* firmground networks occur in fine-grained skeletal limestone. These burrows are distinct, as they are lined by thin, discontinuous iron oxide and phosphatic crusts (see Fürsich et al., 1992). Encrusting bryozoans are present, though typically not in high numbers. Slightly ferruginous, dolomitic and phosphatic skeletal carbonate sand infills the burrow networks, providing further evidence that the passageways remained open for an extended period (Pope and Read, 1997). Similar composite hardgrounds with firmground precursors are described from age equivalent strata in Iowa (Palmer, 1978), the Middle Jurassic of Israel (Wilson et al., 2005), Upper Jurassic of western India (Fürsich et al., 1992), and Cretaceous of southwestern England (Garrison et al., 1987) and Texas (Fürsich et al., 1981). These examples vary in the degree of physical modification of the firmground, but all were formed in carbonates adjacent to the interface between shallow-water facies (within which the firmgrounds developed) and overlying deep-water facies.

Composite hardgrounds typically occur at sharp facies offsets and are traceable over broad areas. Within Upper Ordovician and Lower Silurian, mixed carbonate-siliciclastic successions of the Cincinnati Arch, composite hardgrounds typically cap packstone–grainstone beds forming sharp contacts with overlying grey shales and argillaceous limestones. In these successions composite hardgrounds form distinctive marker horizons over hundreds to thousands of square kilometres (Gordon and Ettensohn, 1984; Norrish, 1991; McLaughlin and Brett, 2007). Levorson and Gerk (1972) reported a similar distribution of Upper Ordovician hardgrounds in Iowa. In the Lexington Limestone composite hardgrounds commonly show a gradient across depositional dip from planar with abraded encrusters in proximal areas, to irregular with a high diversity of encrusters that range from abraded to pristine in the mid-ramp, to irregular with a low diversity of encrusters showing little abrasion and high degree of mineralization in the deeper areas. Similarly,

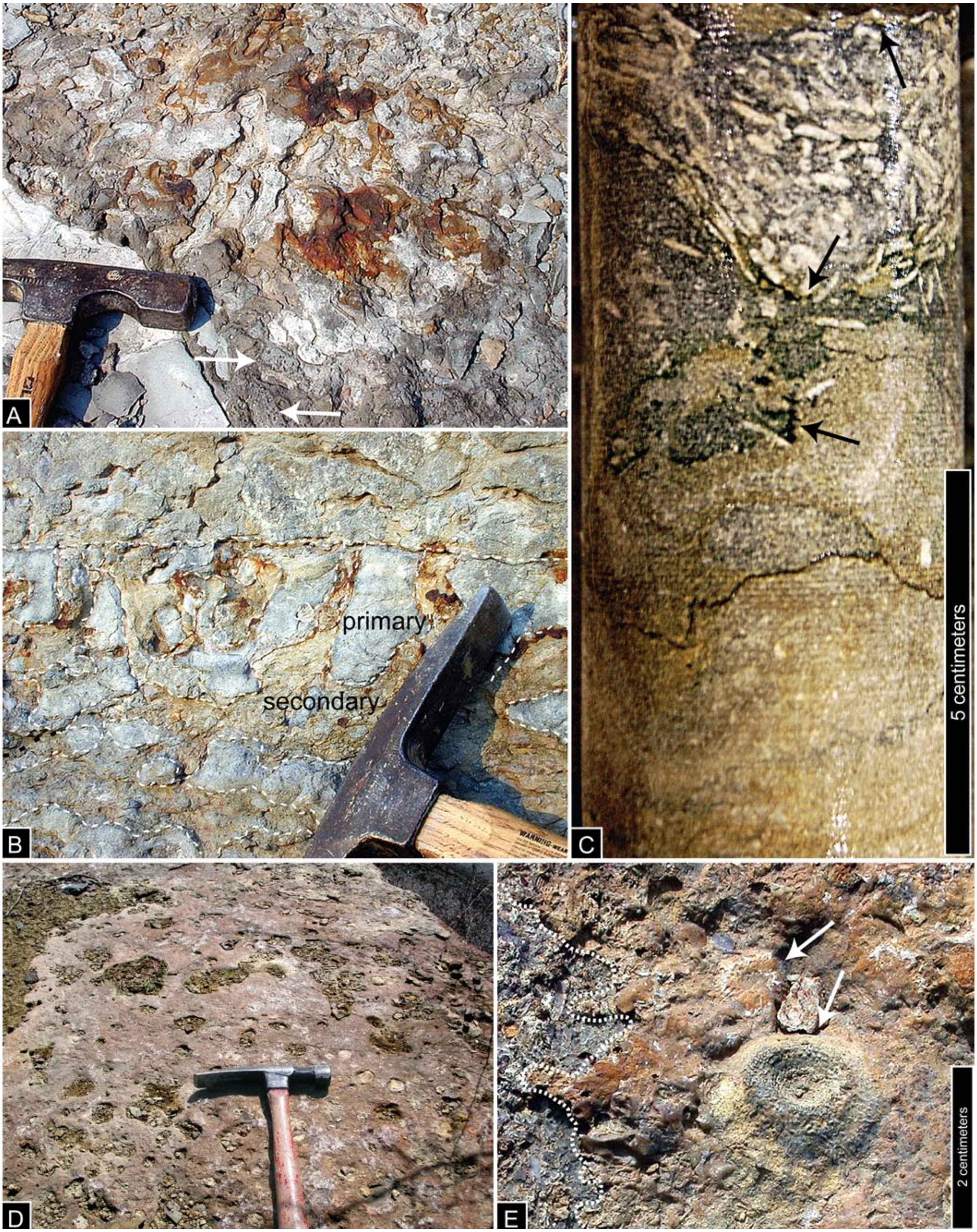
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Figure 6 (left page). Composite hardgrounds with firmground precursors and authigenic mineral crusts. (A) Bedding plane of composite hardground (arrows point to bored areas); Upper Ordovician Curdsville Member of the Lexington Limestone). (B) Cross-section through composite hardground showing phosphatic and dolomitic burrow filling. Limonitic staining in both A and B represent oxidation of pyrite concentrated in burrow filling and along burrow walls. (C) Core showing irregular upper surface (suggesting primary firmgound) coated by glauconitic mineralization (lower arrow) and overlain by glauconitic and phosphatic skeletal grainstone containing at least two additional mineralized surfaces (middle and upper arrows; Lower Silurian Brassfield Formation; Somerset, Kentucky). (D) Iron oxide crust developed on a skeletal grainstone with a bevelled upper surface displaying partially exhumed large burrows (large pits; Upper Ordovician Point Pleasant Member of the Lexington Limestone); hammer is 30 cm long. (E) Close-up of the iron oxide crust showing degraded crinoid holdfastss (arrows) on the upper surface. Dotted line marks a margin where the upper layer of crust has broken away.

Malpas et al. (2004) were able to use continuous exposures of Miocene strata in Egypt to trace hardground surfaces down-slope, recording little change in encruster composition but a significant change in density of encrustation.

REWORKED AND HIATUS CONCRETIONS

Both reworked and hiatus concretions occur in mud-rich, down-ramp sections where storm processes exhumed concretions (or concretionized skeletal elements; Fig. 4B), which were later reworked. They are interpreted as the counterpart of composite hardgrounds in the more distal and siliciclastic-dominated settings. Reworked and hiatus concretions exhibit borings, encrustation and more rarely mineralized coatings; both indicate removal of surficial sediment and thus a change from sediment starvation to submarine erosion.

Reworked concretions occur at various levels within the Upper Ordovician (lower Cincinnatian) in central Kentucky and southwestern Ohio (Wilson, 1985; Algeo and Brett, 1999). As noted above, this shale-dominated (~60–80%) interval contains multiple concretion-bearing horizons, but most are not reworked. Reworked concretions and concretionized skeletal grains are concentrated near the base of the Kope Formation where they are often found reworked into skeletal grainstone beds (Fig. 4B). Similar to the composite hardgrounds described above, these concretions are encrusted both by sheet-like bryozoans and crinoid holdfasts that show varying states of preservation. In many cases, both sides of the concretion are encrusted (Fig. 4C).

Similar occurrences of reworked concretions are described from the Devonian of New York (Brett, 1974) and the Jurassic of India (Fürsich et al., 1992). At a few localities reworked concretions are only surrounded by shale and thus mark subtle shale-on-shale discontinuities. Such contacts are also described from the Devonian of New York (Baird, 1981;

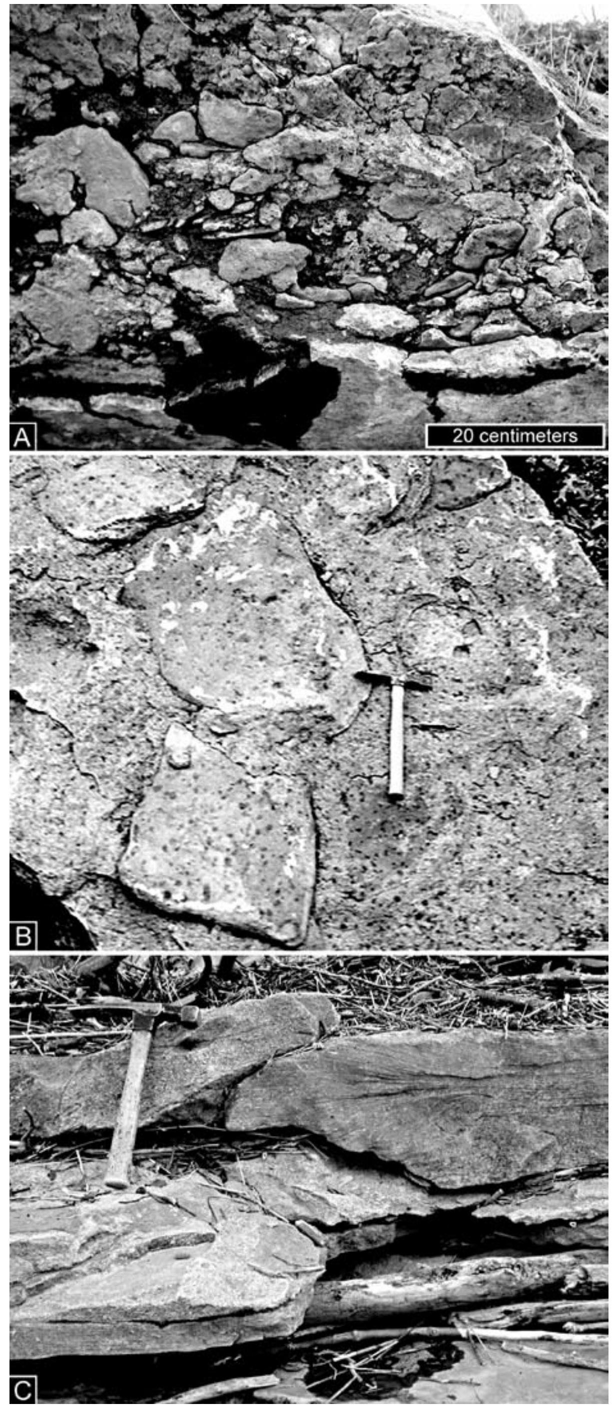

Figure 7. Monomictic intraformational conglomerates (Upper Ordovician Point Pleasant Member of the Lexington Limestone). (A) Underside of a thick (~40 cm) amalgamated skeletal grainstone bed. The middle of this complex bed is composed of densely packed and imbricated, reworked, encrusted and bored fine-grained grainstone clasts. (B) Large (some > 30 cm long) angular slabs of fine-grained grainstone in the lower part of a thick bed of skeletal grainstone/rudstone; hammer is 30 cm long. (C) Broken and over-thrust skeletal grainstone bed disruption of cemented layers initiating formation of intraformational conglomerate; surrounding beds are laterally continuous. Laminae bifurcate toward fault.


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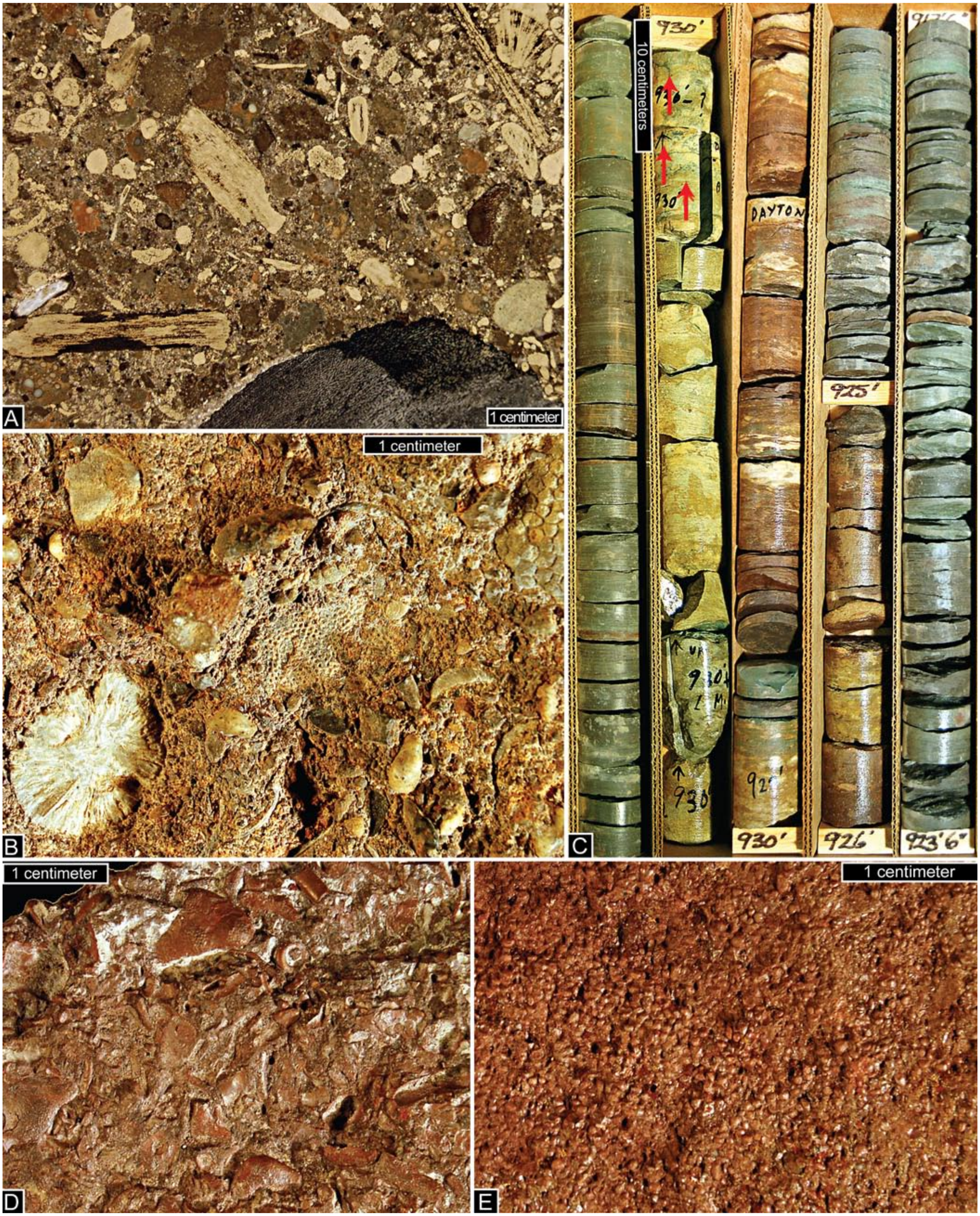
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1 centimeter

Figure 8 (previous left page). Polymictic intraformational conglomerates (all examples from the contact of a grainstone horizon with an overlying shale-dominated interval). (A) Pyrite- and phosphate-mineralized interval containing at least three different types of crinoid-encrusted, blackened limestone clasts capping a 2 m-thick skeletal grainstone succession (this is the down-ramp equivalent of the composite hardground shown in Figure 6B; Upper Ordovician Perryville Member of the Lexington Limestone; Swallowfield, Kentucky); hammer is 30 cm long. (B) Iron- and phosphate- mineralized upper surface of a 5 m thick grainstone-dominated succession displaying least four different lithologies of limestone clast in a variety of orientations and sizes (Upper Ordovician Point Pleasant Member of the Lexington Limestone; Swallowfield, Kentucky). (C) Bedding plane of a skeletal grainstone showing a variety of limestone clasts with glauconite and phosphate mineralization as well as an abundance of small phosphate pebbles (Lower Silurian Merriton Limestone; Hamilton Ontario). (D) Core showing a 3–5 cm thick bed composed of phosphate granules, phosphate- and pyrite-mineralized clasts, bored limestone clasts, chert pebbles and pyrite steinkerns overlain by shale (Upper Ordovician Curdsville and Logana members of the Lexington Limestone; Jackson County, Ohio). (E) Core showing phosphate encrusted bioclastic grainstone overlain by phosphate encrusted and bored grainstone clasts, phosphate pebbles and skeletal grains in a limonitic matrix (Lower Silurian Bischer Formation; Pike County, Ohio).

Figure 9 (previous right page). Condensed phosphorites and ironstones. (A) Polished bedding-parallel section through a thin (~5 cm) lenticular phosphorite bed containing phosphate nodules and steinkerns, phosphatized skeletal grains, placoderm bone, fish teeth, pyritized burrow tubes and skeletal grains, quartz pebbles and granules, coalified and calcified wood in a matrix of conodont elements (Middle/Upper Devonian boundary interval, Blocher Formation; Clay City, Kentucky). This bed spans at least three conodont zones. (B) Bedding plane of a thin (~3 cm) phosphorite containing phosphate and quartz granules and pebbles, glauconitic and hematitic sand and ooids, limestone clasts, rugose and tabulate corals, and crinoid columnals (Lower Silurian Second Creek Phosphorite; Alton, New York). (C) Core showing olive-coloured shale (at left) sharply overlain by bioclastic grainstone (phosphatic hardgrounds with red arrows) that passes upward into bioclastic and oolitic ironstone which is, in turn, overlain by glauconitic shales (Lower Silurian Lulbegrud Shale–Dayton Limestone–Estill Shale succession; Pike County, Ohio). (D) Bedding plane of ironstone showing skeletal grains coated and surrounded by hematite (Lower Silurian Kirkland Ironstone; Clinton, New York). (E) Bedding plane of oolitic ironstone showing small hematite ooids cemented by hematite (lateral equivalent of the phosphorite shown in B; Westmoreland Ironstone; Kirkland, New York).

Mayer et al., 1994), Jurassic of England (Hallam, 1969; Hesselbo and Palmer, 1992) and the Cretaceous of Israel (Soudry and Lewy, 1990).

Hiatus concretions form when reworked concretions become buried and additional layers of concretionary carbonate are added (Kennedy and Klinger, 1974). Hiatus concretions have not been reported from the Ordovician in North America. However, they are known from the Devonian of New York (Baird, 1978), the Jurassic of India (Fürsich et al., 1992), England (Hesselbo and Palmer, 1992) and Switzerland (Wetzel and Allia, 2000), and the Cretaceous of South Africa (Kennedy and Klinger, 1974).

AUTHIGENIC MINERAL CRUSTS

Thin (1–5 mm) authigenic mineral crusts rich in iron (typically oxidized to limonite), glauconite and phosphate. These crusts are common to abundant in the Upper Ordovician (upper Mohawkian) and Lower Silurian (Llandovery–Wenlock) successions across eastern North America, especially along the carbonate margin of the Taconic and Salinic foreland basins and cratonic interior (Sardeson, 1898; Palmer, 1978; Brett et al., 1990; Norrish, 1991; Pope and Read, 1997; McLaughlin et al., 2004). The presence of crinoid and bryozoan holdfasts observed directly cemented to iron oxide crusts (Fig. 6A–E) suggests that they formed early and demonstrates that they were exposed on the seafloor. *Trypanites* borings are occasionally found penetrating the crusts, showing that they were hard substrates. Similar crusts are described from the Devonian of New York (Baird, 1979), the Jurassic of England (Palmer and Wilson, 1990) and India (Fürsich et al., 1992) and the Cretaceous of Israel (Soudry and Lewy, 1990) and are interpreted to represent structures formed by bacterial, fungal or algal biofilms.

By contrast, Palmer (1978) suggested that iron mineralization in some Upper Ordovician hardgrounds in Iowa was, in part, the early diagenetic replacement of calcite by pyrite, stressing that the mineralization did not simply represent a surface crust developed through modern weathering, but rather, was embedded within the limestone. In such cases, even primary pyrite replacement of calcite could indicate stratigraphic condensation.

Authigenic mineral crusts have a lateral and stratigraphic distribution similar to composite hardgrounds. Iron oxide, glauconite, and phosphate crusts are most common within carbonate successions and typically become more closely spaced near facies offsets (i.e. deep-water facies juxtaposed on shallow-water facies). Many can be traced laterally for hundreds to thousands of kilometres.

MONOMICTIC INTRAFORMATIONAL CONGLOMERATES

Monomictic intraformational conglomerates commonly occur as medium- to coarse-grained grainstone–rudstone (skeletal and/or ooid) containing incorporated limestone clasts and blocks, representing some of the most complex discontinuities found in carbonate successions. These beds contain clasts of a single limestone lithology, differentiated from concretions by their skeletal/ooid composition.

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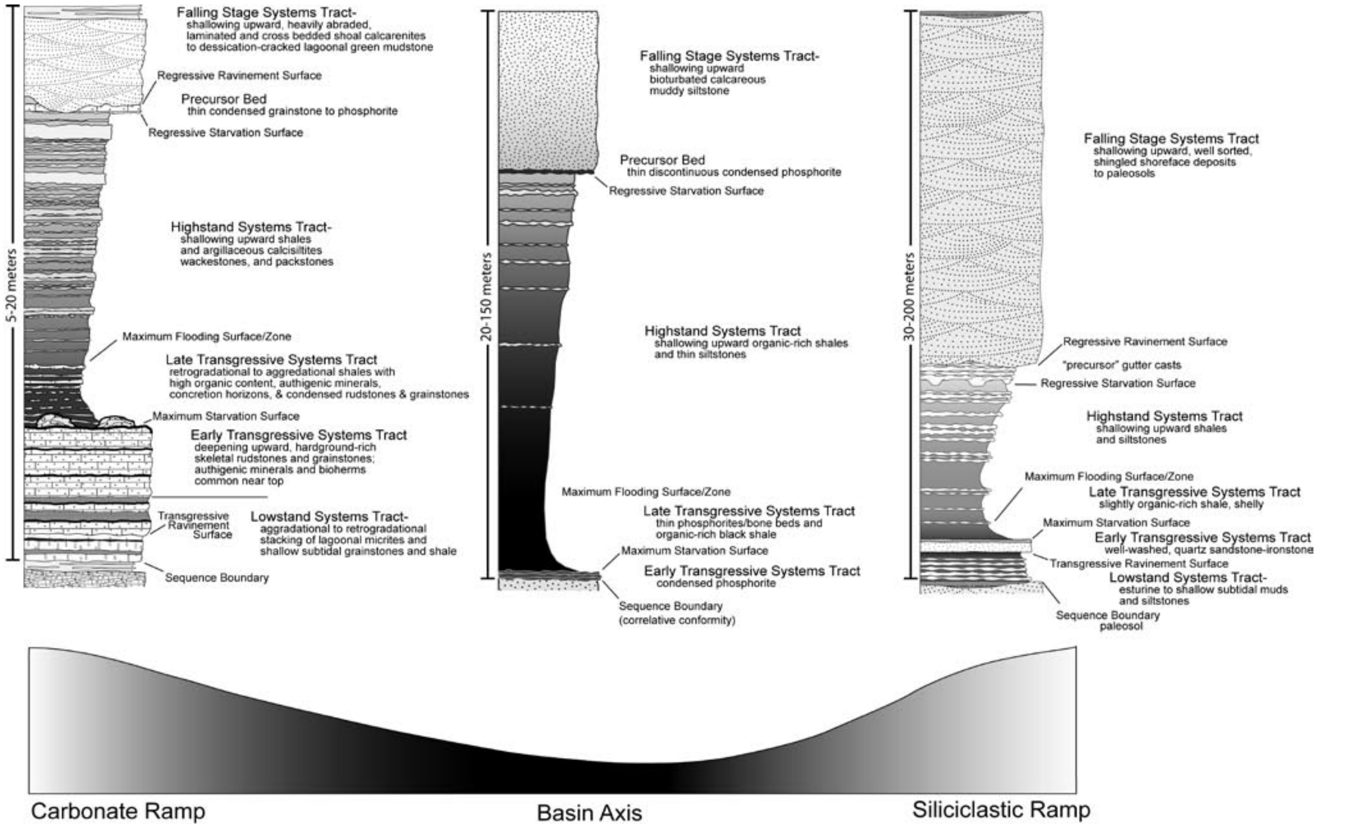

Figure 10. Variation in motif of generalized depositional sequences and types of component discontinuity surfaces and condensed beds across a schematic foreland basin profile.

Monomictic intraformational conglomerates occur at multiple levels within the Upper Ordovician (upper Mohawkian) strata of Kentucky and Ohio (Fig. 7). The clasts are commonly thin (~1–2 cm thick), elongate in plan view, 1 cm to as much as 30 cm in maximum diameter, with rounded edges. Some clasts are morphologically similar to portions of *Thalassinoides* burrows (etc., showing sinuosity and branching) and may represent infilling, lithification, exhumation and reworking of burrow fillings. Similar to reworked concretions, both the top and the bottom of the clasts are frequently encrusted and/or bored showing little variation in the density of borings and degree of mineralization from clast to clast.

Similar Upper Ordovician examples of monomictic intraformational conglomerates elsewhere include crinoid-encrusted clasts from Iowa and Minnesota (Sardeson, 1908) and coral-encrusted clasts from southern Ohio (Foerste, 1917). They are also common in the Lower Ordovician of Korea (Lee and Chankim, 1992; Kim and Lee, 1996), upper Mohawkian strata in New York (Brett and Baird, 2002) and slightly younger (basal Cincinnatian) strata in Ontario (Brett and Brookfield, 1984). Examples are also known from the Lower Silurian of Ohio and Kentucky (Foerste, 1895; Norrish, 1991), New York (LoDuca and Brett, 1994) and Pennsylvania (Sumrall et al., in press). Mesozoic examples are described from the Jurassic of India (Fürsich et al., 1992) and the Cretaceous of southwestern England (Garrison et al., 1987).

By contrast, relatively large angular blocks occasionally form monomictic intraformational conglomerates as well (Fig. 7B). Unlike the smaller, rounded clasts described above, these blocks are rectangular to diamond shape, 2–10 cm thick, 10–40 cm broad and coarse-grained. Furthermore, their distribution tends to be more localized. Foerste (1917) described similar blocks from the upper Cincinnatian in southern Ohio as did Fürsich et al. (1992) from a Jurassic oolite in India. In the latter case, the blocks were interpreted to have spalled off a nearby submarine escarpment. Instead, the Upper Ordovician examples in Kentucky developed upon a gently dipping cratonic ramp lacking evidence of submarine cliffs. In these cases, widespread horizons of soft-sediment deformation (>6000 km^2; McLaughlin and Brett, 2004) are present just a few metres below, providing evidence that seismic activity was affecting the area during that time period. Thus, these angular blocks may be a manifestation of earthquake activity that fractured the early cemented seafloor. Similar features and interpretations are described by Suuroja et al. (1999) from the Middle Ordovician of Estonia, and Pratt (2002a,b) from the Cambrian of Montana and Alberta, respectively.

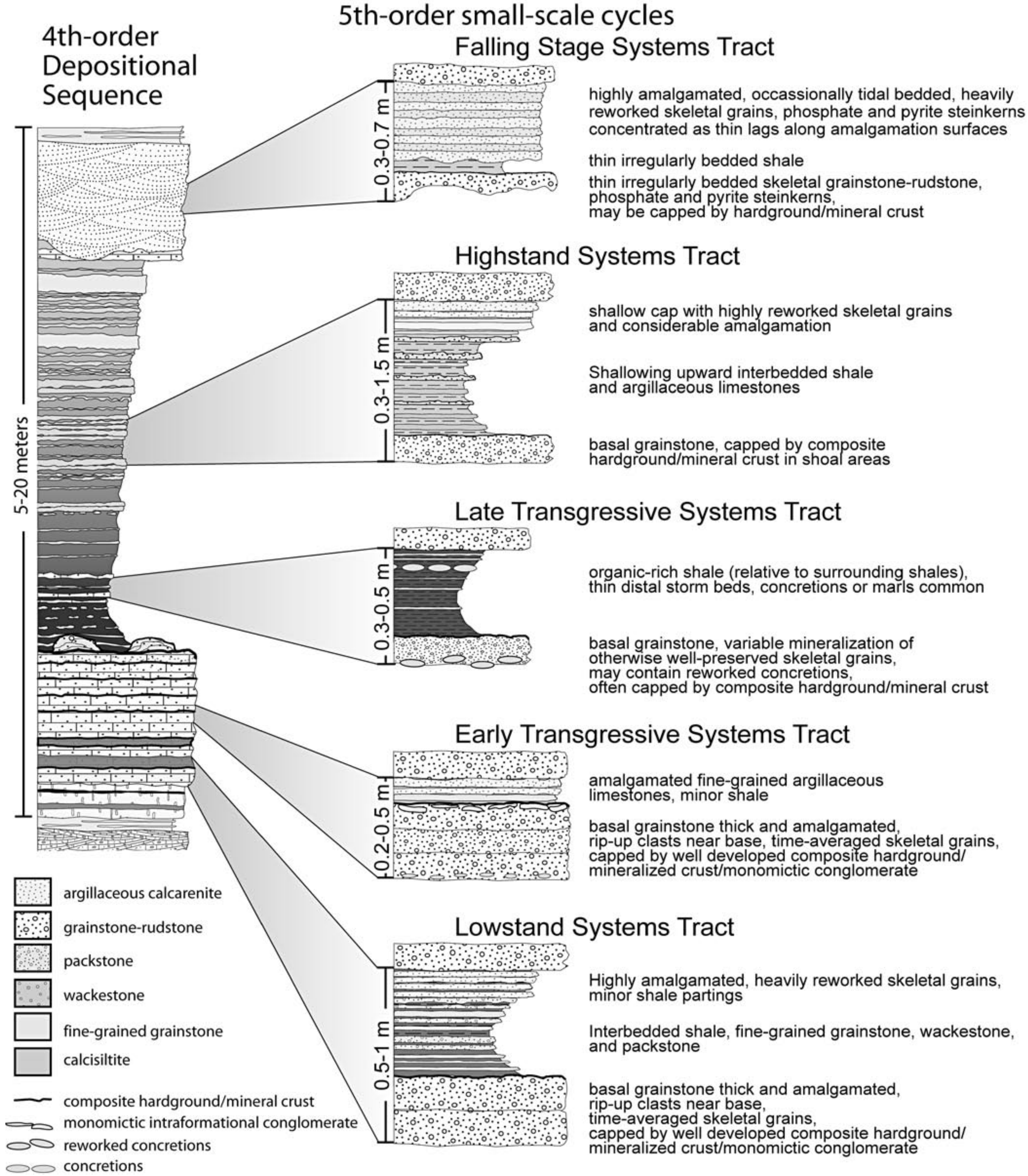

Figure 11. Hierarchy of cyclicity observed on the carbonate margin of middle Paleozoic foreland basins, depicted by generalized stratigraphic sections. Left: depositional sequences (4[th] order) comprise systems tracts that are in turn composed of small-scale cycles. Right: motifs of small-scale cycles organized on the basis of the systems tract in which they occur.

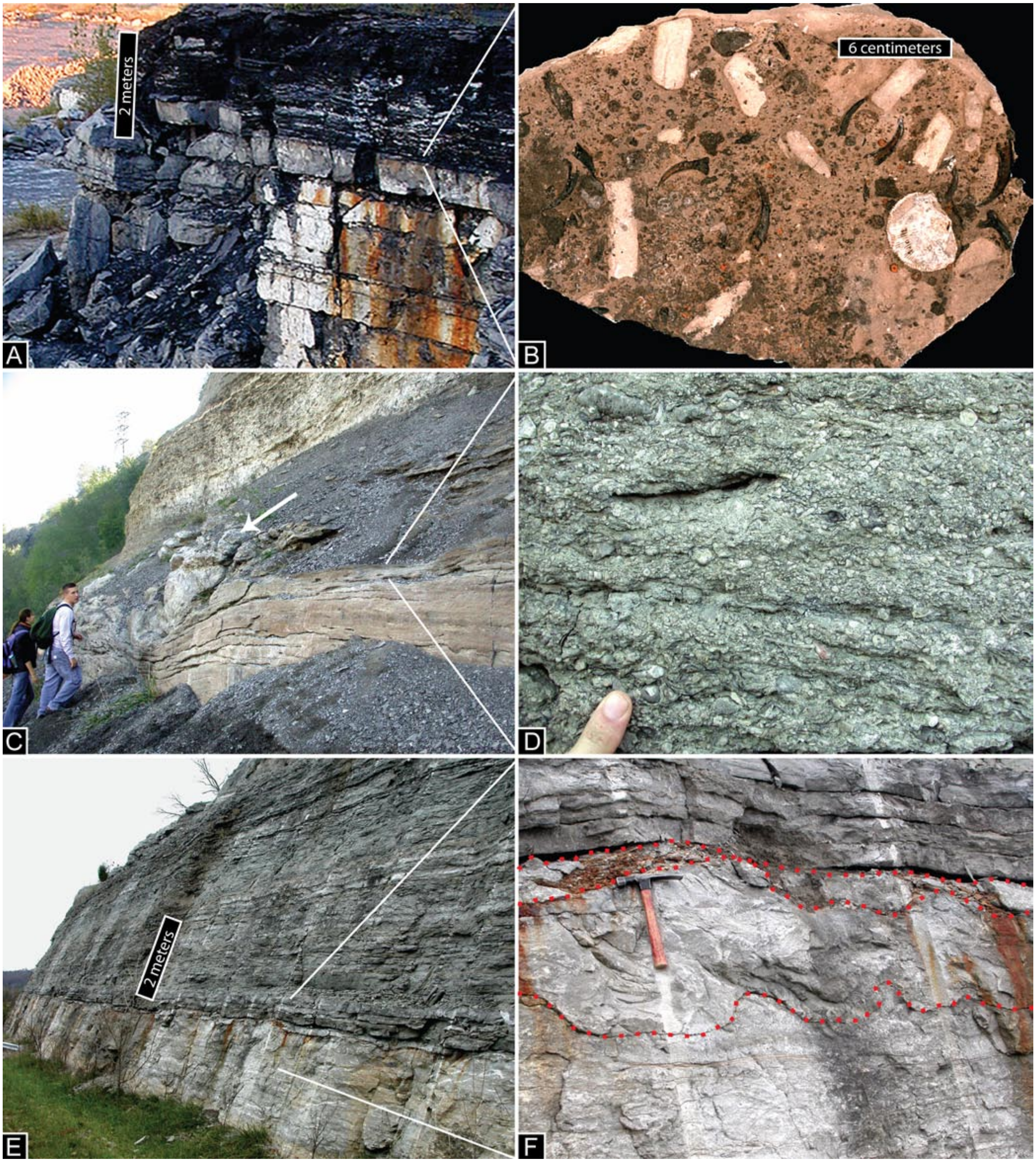

Figure 12. Outcrops showing the stratigraphic context of maximum starvation surfaces. (A and B) Sharp change from wackestones and packstones into organic-rich shale is marked by lenticular condensed phosphorite (Middle Devonian Onondaga Limestone/Marcellus Shale; Jamestown, New York). (C and D) Sharp facies offset marked by a glauconitic crinoid grainstone overlain by a 2 m thick bioherm that is draped by organic-rich shale (Lower Silurian Irondequoit Limestone/Rochester Shale; Lewiston, New York). (E and F) Massive skeletal grainstone/rudstone containing multiple closely spaced iron oxide crusts which become clustered near the contact with the overlying argillaceous wackestone- and packstone-bearing shale (Upper Ordovician Sulphur Well/Stamping Ground member contact within the Lexington Limestone; Peaks Mill, Kentucky). Small (10 to 20 cm thick) algal–bryozoan bioherms occur the contact (outlined with dotted red lines in F.

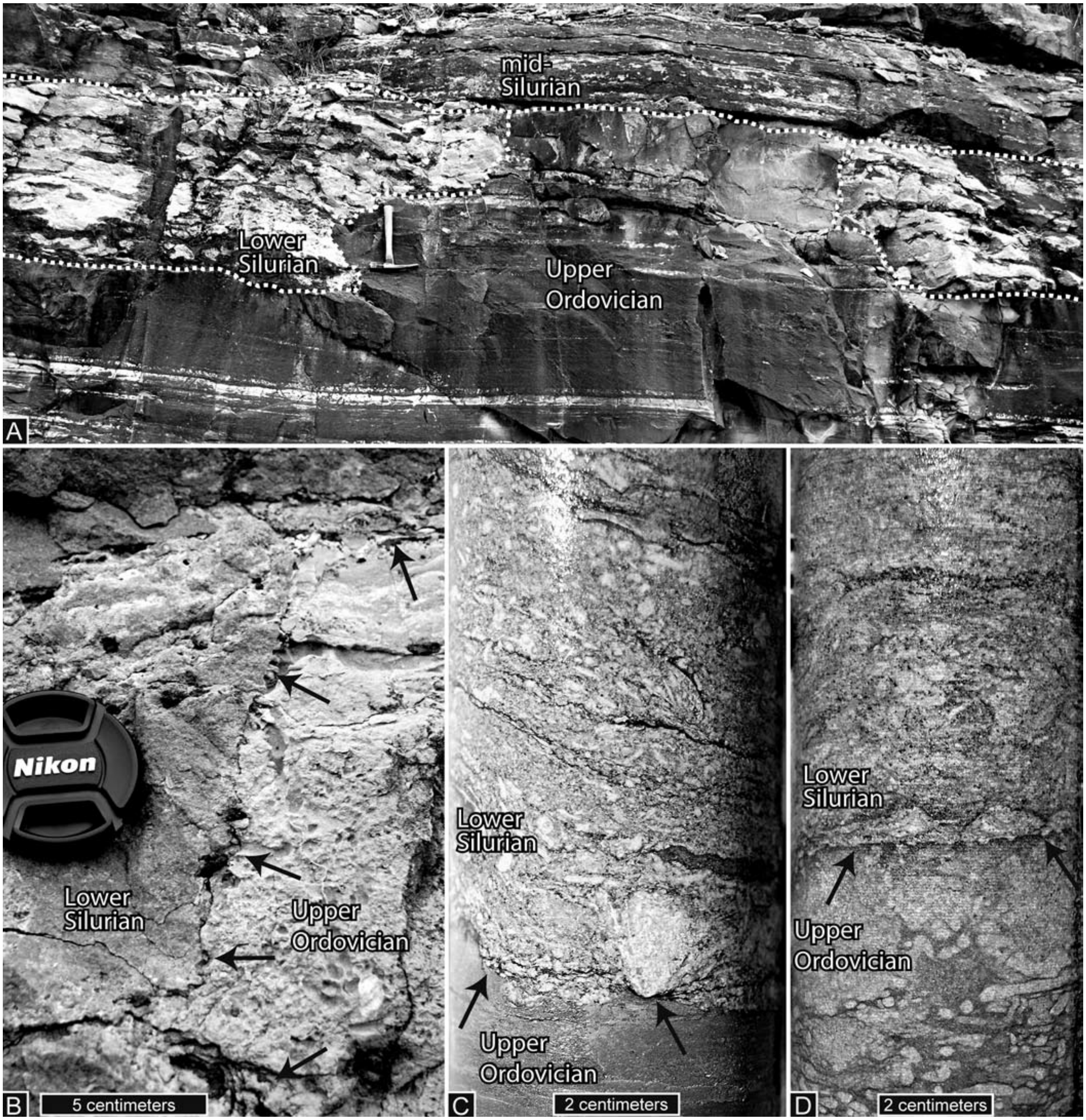

Figure 13. Discontinuity surfaces at sequence boundaries. (A) Outcrop showing irregular surface forming the Upper Ordovician/Lower Silurian boundary (Saluda and Brassfield formations; Crestwood, Kentucky) and planar surface marking the Lower Silurian (lowermost Llandovery)/mid-Silurian (uppermost Llandovery Dayton Limestone–Osgood Shale) contact. Both of these unconformities mark erosion of tens of meters of strata and long periods of subaerial exposure yet show relatively little downcutting. (B) Cross-section of Upper Ordovician/Silurian contact in nearby road cut showing a vertical surface with numerous borings and minor pyritic mineralization (arrows) suggesting composite hardground development. (C) and (D) Cores of the Upper Ordovician/Lower Silurian boundary (Clinton and Greene counties, Ohio). Both show minor authigenic mineralization and encrustation (rugose coral in C, right arrow) on an undulating contact, overlain by coarse skeletal grainstone containing abundant of glauconite as disseminated granules and concentrated in stringers along foreset beds (black specks).

Taphonomic properties reveal that monomictic intraformational conglomerate beds have a complexity similar to that of composite hardgrounds and thus may form over a similar time interval. Extended periods were required to complete the processes of lithification, erosion, undercutting, rounding and re-sedimentation. The rounded outlines of these clasts may be a reflection of concretionary cementation, possibly tracing older burrow networks. Aggregation of many clasts of similar lithology and limited fragmentation suggests early cementation close to the sediment–water interface followed by reworking and little subsequent transport.

POLYMICTIC INTRAFORMATIONAL CONGLOMERATES

Polymictic intraformational conglomerates resemble monomictic conglomerates, but contain mixtures of limestone clasts of variable lithologies and sizes. Clasts typically vary from pebbles to cobbles; large limestone blocks are rare. Their texture ranges from fine- to coarse-grained, and their lithology grainstone to mudstone. The degree of biologic modification of the clasts ranges from pristine to heavily encrusted, mineralized and bored.

Polymictic intraformational conglomerates the Upper Ordovician (upper Mohawkian) typically show firmground precursors and are usually associated with thick iron oxide and phosphatic crusts in mid- to outer-ramp facies (Fig. 8). The surfaces they rest upon are occasionally modified by abrasion and corrosion and irregular. Their distribution is largely restricted to major facies offsets where they cap deepening-upward skeletal grainstone successions. Thus, polymictic intraformational conglomerates are rare relative to monomictic varieties. They also occur in the Upper Ordovician (McLaughlin and Brett, 2007) and Lower Silurian (Llandovery–Wenlock) strata of the Cincinnati Arch (Foerste, 1895) and Upper Silurian of the Welsh Borderland (Cherns, 1980). Polymictic intraformational conglomerates are also present in Upper Ordovician (Maysvillian–Richmondian) black shales of Iowa and Minnesota

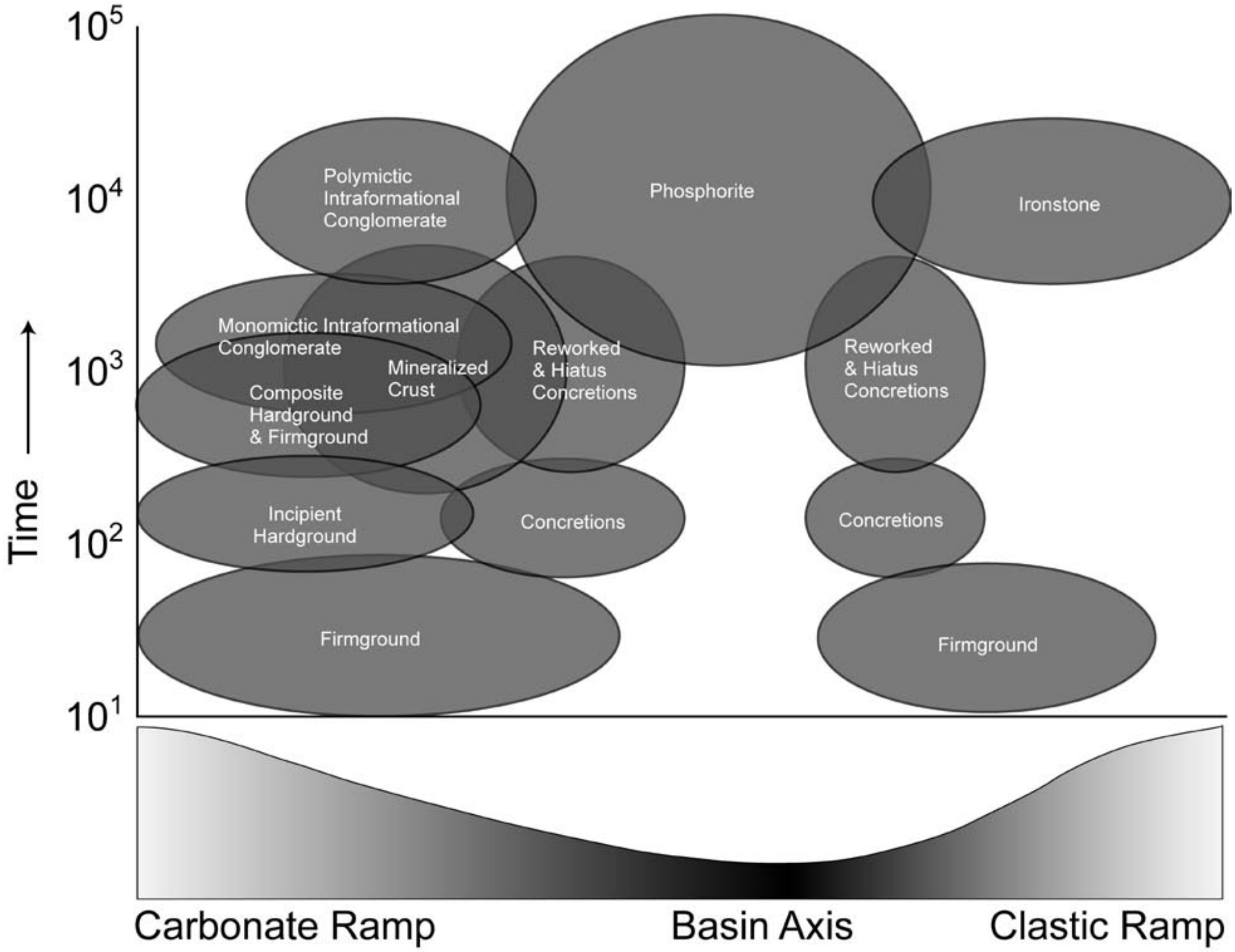

Figure 14. Hierarchy of discontinuity surfaces and condensed beds across a schematic foreland basin profile plotted against time in years. The duration assigned to individual discontinuities and condensed beds is estimated based on increasing complexity and sedimentology. The carbonate side of the basin preserves abundant sediment starvation indicators.

associated with condensed phosphorite beds (Raatz and Ludvigson, 1996). Undoubtedly, other occurrences exist, description may be inadequate to differentiate of mono- from polymictic intraformational conglomerates in the literature is rare, thus the accurate reporting of their full distribution in epicontinental seas is not possible at this time.

CONDENSED PHOSPHORITES

Also referred to as "bone beds", condensed phosphorites (Föllmi, 1996) generally form thin (<10 cm; most commonly 1–2 cm) discontinuous beds or nodule layers within organic-rich shales or cap limestone intervals at their contact with an overlying shale succession. Condensed phosphorites are generally composed of phosphatized and pyritized skeletal grains, phosphate and pyrite steinkerns, conodonts, bone fragments and other refractory materials such as chert nodules (Fig. 9). The stable fluorapatite composition of phosphatic particles ensures that these materials remained little changed on the seafloor for up to millions of years (e.g., Donovan 1991; Lucas and Prévôt, 1991), and may act as nucleation points for further phosphate precipitation (Föllmi, 1996). Formation of coated phosphate grains occurs just below the sediment–water interface in the suboxic zone and requires low sedimentation rates (Pufahl and Grimm, 2003). It is typically contended that phosphorites form below highly productive surface waters, although fluvially derived phosphorous is suggested by epicontinental phosphate deposits such as the Cretaceous of Egypt (Glen and Arthur, 1990). Macquaker et al. (1996) argued that apatite authigenesis pre-dates sulfate reduction, implying precipitation of phosphate just above the oxic–suboxic/sulphidic interface. Concentrations of pyrite with morphologies of burrow tubes, steinkerns and internal molds are often a dominant component within these beds, in addition to chert and silicified skeletal particles (e.g. Baird and Brett, 1986, 1991; Brett et al., 2003b). The presence of concentrations of pyrite grains suggests dysaerobic to anaerobic depositional environments and moderate to intense winnowing which liberated the pyrite from the enclosing muds. Taphonomic studies of condensed phosphorites reveal a high degree of time-averaging (e.g. Baird and Brett, 1986 and references therein; Brett et al., 2003b).

Condensed phosphorites in the Upper Ordovician, Lower Silurian and Middle Devonian of the Appalachian Basin are associated with thick successions of shale representing deposition near the deep the centres of foreland basins (Fig. 9B; Baird and Brett, 1986; Brett et al., 1998). Similarly, Pennsylvanian and Permian condensed black shales, bordering on phosphorites, are common deep-water components of mid-continent cyclothems in North America and are likewise interpreted to represent extended periods of sediment starvation (Heckel, 2002; Algeo et al., 2004). Alternatively, condensed phosphorites within Upper Devonian strata of the Cincinnati Arch appear to have been deposited at relatively shallow depths within a restricted, but also sediment starved, basin (Fig. 9A; Conkin, 1986; Brett et al., 2003b; Schieber and Riciputi, 2004).

IRONSTONES

Ironstones are beds of hematite, goethite, and siderite-cemented sediments that may contain abundant hematite and phosphate-coated particles as well as glauconite, skeletal grains and quartz sand and pebbles grains (Figs. 9C–E). Primary mineralization within ironstones is early-diagenetic, forming within a few centimetres below the sediment–water interface (Taylor et al., 2002). Petrographic data suggest that chamosite and bertherine iron-rich clays are precursors of hematite or goethite which commonly make up the present composition of many ironstones (Cotter and Link, 1993). Because bertherine and siderite pre-date pyrite in ironstones and thus must have formed in the sub-oxic zone before sulfate reduction, in this way they are similar to condensed phosphorites (Macquaker et al., 1996). Ironstone formation is governed primarily by availability of reduced iron and low sedimentation rates and extensive physical reworking (Macquaker et al., 1996; Taylor et al., 2002). Ironstones commonly grade laterally into other authigenic mineral deposits such as glauconitic sandstones, limestones and phosphorites whose genesis demands similar environmental conditions. However, ironstones are largely restricted to the clastic side of foreland basins and passive margins where there is run-off from iron-rich basement rocks. High levels of bioturbation are indicative of formation in relatively well oxygenated waters (Brett et al., 1998; Taylor et al., 2002). Furthermore, ironstones commonly exhibit cross-stratification, ripples, sharp bases, stringers of coarse quartz sand and pebbles, and sparry calcite cement, suggestive of deposition within high-energy environments near fair-weather wave-base. They are typically widespread and range from only a few centimetres to over a metre in thickness, but are commonly ~10 cm thick which remains uniform across strike. Phanerozoic ironstones are most abundant during regional tectonic quiescence between tectonic pulses in greenhouse periods (Ordovician–Devonian and Jurassic–Paleogene; Van Houten, 1985, 1990; Van Houten and Arthur, 1989; Young, 1992; Brett et al., 1998).

Ironstones are prominent components of Lower Silurian successions in the Appalachian Basin. On the siliciclastic margin of the Taconic Foreland Basin widespread ironstones cap shallowing-upward sandstone successions at the contacts with overlying shales, while more localized lenticular ironstones may be scattered throughout sandstone intervals. Lenticular ironstones also occur at the lateral interfaces of inner shoal margins and lagoons where skeletal limestones laterally grade into siliciclastic mudstones (Brett et al., 1998). Widespread ironstones in the Appalachian basin are commonly highly fossiliferous; component skeletal grains commonly show a high degree of within-habitat time-averaging. Some of the thickest and most widespread ironstones show a mixing of shallow- and deep-water faunas and are thus considered ecologically time-averaged (see also Kidwell and Bosence, 1991). Oolitic ironstones are relatively rare on the carbonate margin of the Taconic Foreland Basin, though fossil ironstones and hematitic skeletal grainstones are common (Fig. 9C). These hematitic deposits are most common in mid-ramp facies (commonly 1–2 m thick) and grade up-ramp into hardground-rich glauconitic skeletal grainstones and wackestones (commonly 2–5 m thick). They often show a similar vertical gradient from hardground-rich skeletal grainstone to hematitic grainstone and ironstone from bottom to top, interpreted as deepening-upward (Fig. 9C).

CONTRIBUTIONS TO A MODIFIED SEQUENCE-STRATIGRAPHIC MODEL

Stratigraphic units in foreland basins can be traced across the basin from areas of high sedimentation rates—the siliciclastic margin—to areas of low sedimentation rates characteristic of the cratonic interior. Because discontinuity surfaces and condensed beds show variable levels of complexity, paralleled by verging lateral extent and stratigraphic distribution, they provide information about the temporal affinities of the enclosing strata. Widespread discontinuity surfaces serve to separate the rock record into discrete, genetically related facies packages. These are the building blocks of sequence stratigraphy and are now well studied in a range of tectonic and sedimentary settings. While refined models for carbonate platforms on passive margins (e.g. Sarg et al., 1988) and the siliciclastic margins of foreland basins has been developed (e.g. Van Wagoner et al., 1990; Van Wagoner and Bertram, 1995), simply combining these two models does not generate a hybrid model that explains the patterns observed in the strata of carbonate margins of foreland basins, and the sedimentary and tectonic dynamics of these basins as a complete system.

Foreland basins display an asymmetric sediment fill indicative of differential subsidence. Loading of the craton margin, as in peripheral and retro-arc basins, results in high levels of subsidence adjacent to the rapidly wasting orogenic highlands, resulting in thick siliciclastic-dominated sedimentary successions. Crossing the foreland basin, subsidence rates drop toward the hinge line where the effective subsidence rate is zero. Still further cratonward, uplift as a peripheral bulge may be the dominant tectonic control. In these areas siliciclastic sedimentation rates are relatively low. On flooded cratons the carbonate margins of foreland basins are commonly thousands of kilometres from the seashore and cratonic source areas for terrigenous, sediment, meaning that fine-grained siliciclastics are delivered across the foreland basin from the much closer siliciclastic margin. Judging by the overall thickness of foreland basin fill, carbonate production rates on the carbonate margin were relatively low as well. Taking into account these tectonic and sedimentological regimes, it follows that stratigraphic facies changes on the stable carbonate margin may more closely record eustatic fluctuations than those in the basin centre or on the siliciclastic margin.

The complex discontinuity surfaces and condensed beds occurring at facies offsets help define packages of strata representing depositional sequences and component systems tracts. In the middle Paleozoic of eastern North America five genetically related packages of strata (systems tracts) can commonly be recognized within 4^{th}-order depositional sequences: lowstand (LST), early transgressive (ETST), late transgressive (LTST), highstand (HST), and falling stage (FSST). Condensed beds within this interval are primarily found within the ETST. Discontinuity surfaces mark the sequence boundary (SB) separating the FSST and LST, the maximum starvation surface (MSS) separating the ETST and LTST, and the forced regression surface (FRS) separating the HST and the FSST (Fig. 10). These surfaces are relatively isochronous by comparison with the transgressive ravinement and regressive ravinement surfaces (which do not separate systems tracts). Systems tracts are composed of 5^{th}-order parasequences (on the siliciclastic margin) and laterally equivalent small-scale cycles (on the carbonate margin; Fig. 11).

The LST is typically composed of alternations of clean skeletal limestone and interbedded argillaceous limestones and shales on the carbonate margin, silty mudstones with carbonate stringers and laminae in the basin centre, and a mixture of silty mudstones, siltstones and sandstones on the siliciclastic margin. These sediments are deposited during the initial phase of sea-level rise when sedimentation rates generally exceed the rate of accommodation created in nearshore areas on the siliciclastic margin of the foreland basin. Thus the LST represents a period of progradation to aggradation on the siliciclastic margin and into the basin centre; condensed beds are therefore generally lacking in these areas during this phase. However, on the carbonate margin this interval of sea-level rise is marked by aggradation to retrogradation. In the mid-ramp, relatively condensed skeletal grainstone–packstone beds are commonly capped by composite hardgrounds, which are in turn overlain by argillaceous limestones and shales (Fig. 11). These 5^{th}-order cycles record fluctuations in the amount of clay transported across the foreland basin. The composite hardgrounds record the cessation of carbonate sedimentation prior to this clay influx. The contact between the LST and ETST is typically not marked by a condensed bed but, rather, by a sharp change to nearly pure carbonate sedimentation.

The ETST is deposited during rapid sea-level rise when the rate of relative sea-level rise (eustasy plus subsidence) outpaces sedimentation on the siliciclastic margin of the foreland basin, resulting in siliciclastic sediment starvation of the entire basin. On the siliciclastic margin the ETST is represented by thin condensed beds composed of shelly mudstones and reworked sands, commonly with authigenic mineralization. When present, ironstones form the ETST on the siliciclastic margin. These condensed beds can be traced down-ramp into condensed phosphorites near the basin centre. Condensed phosphorites begin to split apart into closely stacked polymictic intraformational conglomerates alternating with skeletal packstone–grainstone (commonly phosphatic, glauconitic and/or hematitic) up the carbonate ramp. Polymictic intraformational conglomerates give way to monomictic conglomerates, iron oxide and phosphatic crusts, and composite hardgrounds intercalated with skeletal grainstones in the mid- to outer-carbonate ramp (Fig. 11). The ETST is commonly thickest in the upper mid-ramp resulting from moderate levels of seafloor energy, increasing accommodation space, and increased skeletal carbonate production rates in response to siliciclastic sediment starvation. Authigenic mineralization commonly marks increasing condensation toward the top of the ETST (Fig. 9C). Farther up-ramp, in condensation winnowing plays the dominant role rather than carbonate factory cessation through drowning. Skeletal grainstones of the ETST in shoal areas commonly contain highly reworked skeletal grains, amalgamation, an abundance of disseminated phosphate steinkerns, and simple planar hardgrounds indicative of condensation.

The MSS (see Baum and Vail, 1988) forms during the most rapid rate of sea-level rise and is marked by the most complex discontinuity surface or condensed bed within the depositional sequence at a facies offset that can be traced across nearly the entire foreland basin. On the siliciclastic margin the MSS marks the period of greatest siliciclastic sediment starvation and may be marked by an iron oxide or phosphate crust, thin condensed phosphorite or ironstone. In the basin centre the MSS is marked by continued formation of phosphorite. On the carbonate margin the MSS is marked by an iron oxide or phosphate crust, thin condensed phosphorite, polymictic intraformational conglomerate, or some combination of these features capping the underlying skeletal grainstones of the ETST at the contact with the overlying argillaceous carbonates of the LTST (Fig. 12). The MSS on the carbonate margin forms in response to drowning of the carbonate factory. Common in the middle Paleozoic are bioherms—from small mounds a few centimetres thick to pinnacle reefs up to >20 m tall—directly underlying the MSS, recording the onset of rapid sea-level rise.

The LTST forms during the slowing phase of sea-level rise. On the siliciclastic margin this period is associated with renewed progradation of muds and silts as the rate of relative sea-level rise is outpaced by sedimentation rate. On the outer-ramp of both the carbonate and siliciclastic margins concretion horizons mark the early part of the LTST. In the basin centre this interval is typically marked by deposition of organic-rich shales, which may contain K-bentonite horizons. In the mid- to upper carbonate ramp the argillaceous carbonates of the LTST share many similarities with the LST, though with evidence of increasing-upward rather than decreasing-upward input of siliciclastics (Fig. 11). This interval typically displays alternations between condensed skeletal grainstone beds (occasionally containing reworked concretions) and shaly intervals containing a mixture of concretions (indicative of sediment starvation) and obrution deposits (indicative of event sedimentation). Faunally, the FSST indicates deepening-upward, but lithologically is suggestive of increased siliciclastic sedimentation. The contact with the overlying HST may be lithologically subtle, marked only by the deepest water biofacies (the maximum flooding surface/zone).

The HST forms during initial sea-level fall. Relatively low seafloor energy associated with high sea level and renewed influx of siliciclastics (primarily clay) into the basin generally results in high net siliciclastic sedimentation rates, producing a general lack of discontinuities and condensed beds across the foreland basin. The HST on the siliciclastic margin is marked by renewed growth of deltas resulting in deposition of mud, silt and sand onto the siliciclastic margin. The HST in the basin centre is commonly composed of a thick, monotonous shale succession. The HST on the carbonate margin is marked by a progradational succession of shales and argillaceous carbonates. Only in the high-energy inner-ramp facies of the carbonate margin, where winnowing reduces net sedimentation rates, are simple hardgrounds and authigenic mineral crusts occasionally found in the HST.

The FRS forms during the most rapid sea-level fall and is marked by a discontinuity surface and/or condensed bed at a facies offset across much of the foreland basin (Fig. 9). On the siliciclastic margin, the FRS corresponds with the RRS on the shallow-ramp where it is marked by a planar, erosive discontinuity surface typically lacking indications of sediment starvation. However, down-ramp the FRS occurs below the RRS and is commonly marked by a quartz arenite or ironstone. Toward the basin centre the FRS may be marked by a thin discontinuous condensed phosphorite ("precursor bed" of Brett and Baird, 1996). Up-ramp on the carbonate margin where the rapid drop in base level results in sediment starvation via winnowing and bypass, the FRS is typically marked by a thin, condensed skeletal grainstone lag bed and in some cases a composite hardground.

The FSST forms as sea-level fall slows toward its lowest point. On the siliciclastic margin of the foreland basin the FSST is dominated by amalgamated delta and prodelta sands which mark rapid regression and corresponding formation of large delta systems (see Plint and Nummedal, 2000). Toward the basin centre the FSST is dominated by interbedded silty mudstones and siltstone storm beds. As a result of high siliciclastic sedimentation rates, discontinuity surfaces and condensed beds are largely lacking on the siliciclastic margin and in the basin centre during deposition of the FSST. On the carbonate margin the FSST is dominated by amalgamated argillaceous calcarenites composed of abraded and fragmented skeletal grains. Though this interval is condensed through winnowing, hardgrounds and other discontinuity surfaces are uncommon. Apparently continuous reworking during formation of the FSST largely precludes early cementation on the carbonate margin or consistently removes the record of encrusters and/or mineralization.

The SB marks the lowest point of the sea-level cycle and bounds the depositional sequence. However, only on the shallow margins of foreland basins are SBs major erosion surfaces. In up-ramp parts of the siliciclastic margin the SB may be marked by incised valleys, but more commonly evidence of erosional discontinuity is subtle. For example, sequence boundaries are often marked simply by a paleosol or the base of an estuarine succession. Farther offshore the SB corresponds to the turn around of the RRS into the TRS at the most basinward position of upper shoreface (see Catuneanu, 2002). Basinward the SB is largely correlative and is typically marked by a switch from sand-dominated facies and evidence for rapid progradation to more heterolithic mixed mud-, silt- and sandstone facies with a decrease in the degree of amalgamation indicating accommodation increase. Like the MFS, the SB in the outer siliciclastic ramp and basin centre is not a discontinuity surface but is simply marked by the shallowest water facies within the sequence. On the mid- to outer-carbonate ramp the SB is also largely in continuity and is marked by the shallowest water facies and a switch from highly amalgamated to more heterolithic bedding. As on the shallow siliciclastic margin, the SB in the up-dip portion of the carbonate ramp corresponds with the turn around point of the RRS–TRS where it becomes an erosive discontinuity surface. Shoreward the SB is a discontinuity surface associated with subaerial exposure and in some cases may be marked by a rockground developed on the karst surface during exposure and/or a hardground developed during initial submersion; both may show evidence of boring, iron oxide precipitation, and encrustation. However, discontinuity surfaces developed on exposure surfaces, even those at mega-sequence boundaries (Fig. 13), rarely show the complexity characteristic of the MSS. In fact, they typically share much in common with closely overlying composite hardgrounds and authigenic crusts developed within the LST and ETST.

In summary, discontinuity development and stratigraphic facies change represent a common response to variations in siliciclastic influx and carbonate production associated with oscillations of relative sea level in epicontinental seas. The most widespread discontinuity surfaces and condensed beds form during the most rapid rate of sea-level rise and fall. Development of cemented layers and their exposure and modification as hard substrates requires very low net sedimentation rates and/or enhanced seafloor erosion. In terms of sequence stratigraphy, this is the critical connection in the present context: best developed hardgrounds and other discontinuities mark prolonged reduction of sediment accumulation and thus are especially characteristic of transgressions.

CONCLUSIONS

The stratigraphic record of epicontinental seas is full of complex surfaces and beds that represent of periods of slowed or no net deposition. Comparative analysis of these discontinuities in sedimentation reveals a general hierarchy of increasing complexity. The relationships of these features to one another can be represented by plotting their observed lateral distribution across a foreland basin against estimates on the duration of their genesis from experimental data in the literature and taphonomic observations (Fig. 14). Good agreement is found between interpreted duration of cycles of 4th- and 5th-order and estimated duration of discontinuities and condensed beds characteristic of those cycles. In order of increasing complexity the discontinuity surfaces and condensed beds found in epicontinental seas include:

(1) Simple firmgrounds and hardgrounds—heavily burrowed beds and surfaces with limited encrustation;
(2) Composite hardgrounds—beds and surfaces showing a complex history of encrustation and bioerosion;
(3) Authigenic mineral crusts—thin glauconite, phosphate and iron oxide layers capping beds and commonly encrusted or penetrated by borings;
(4) Monomictic intraformational conglomerates—reworked clasts of a single lithology often showing encrustation, bioerosion or mineralization;
(5) Polymictic intraformational conglomerates—thin beds or surfaces containing reworked limestone clasts of varying composition often showing authigenic mineralization, encrustation and/or bioerosion;
(6) Condensed phosphorites—thin phosphate-rich beds and lenses containing a variety of reworked and/or mineralized clasts, skeletal grains and refractory materials such as teeth;
(7) Ironstones—thin to thick beds and lenses cemented by hematite, occurring as bioclastic or oolitic varieties; along with reworked clasts, quartz grains, and refractory materials.

These surfaces have specific relationships to the sequence stratigraphy of epeiric foreland basins:

(1) Simple firmgrounds and hardgrounds may occur in any part of a depositional sequence, as their limited lateral distribution and relatively common occurrence suggests they were ephemeral features that record local rather than regional or global events;
(2) Composite hardgrounds, authigenic mineral crusts and monomictic intraformational conglomerates typically mark the maximum starvation surface of 5th-order small-scale cycles; and separating the early from late transgressive systems tract;
(3) Polymictic intraformational conglomerates, condensed phosphorites and ironstones are the most complex discontinuities and typically mark the MSS of 4th-order depositional sequences.

Lateral tracing of discontinuity surfaces and condensed beds reveals a compositional gradient along depositional dip. The compositional gradient is related both to changes in water depth which affect environmental energy, water temperature, salinity, light penetration and so on, and proximity to the siliciclastic margin which influences the rate of relative sea-level fluctuation, sedimentation rate, turbidity and water chemistry.

A sense of the general characteristics of discontinuities in cratonic successions aids in the construction of a sequence stratigraphic framework at a resolution in many cases beyond that which biostratigraphy alone can provide. While the sequence stratigraphic model presented here focuses primarily on foreland basin successions, the distribution and types of discontinuity surfaces characteristic of the carbonate margin share much in common with the purer carbonate strata of other epeiric seas. We propose that this model, with some modification, will hold for those strata as well, even though the greater uniformity of lithology increases the difficulty of recognizing facies change.

ACKNOWLEDGEMENTS

The authors are indebted to many colleagues for fruitful discussions that greatly helped refine our ideas. In particular, we would like to thank Steve Donovan and Paul Taylor for organizing the special session "Hard Substrates Through Time" at the 2004 Geological Society of America annual meeting, which gave impetus for compiling this study. Reviews by Brian Pratt, Andreas Wetzel and Duncan McIlroy greatly helped to improve this paper. Acknowledgement is also made to the donors of the Petroleum Research Fund of the American Chemical Society and National Science Foundation Grant # 0518511 for partial support of this research.

REFERENCES

Algeo, T.J., Schwark, L. and Hower, J.C., 2004, High-resolution geochemistry and sequence stratigraphy of the Hushpuckney Shale (Swope Formaiton, eastern Kansas): implications for climato-environmental dynamics of the Late Pennsylvanian Midcontinent Seaway: Chemical Geology, v. 206, p. 259–288.

Algeo, T.J. and Brett, C.E., 1999, eds., Sequence, Cycle, and Event Stratigraphy of Upper Ordovician and Silurian strata of the Cincinnati Arch region: Field Trip Guidebook for the 1999 Field Conference, Great Lakes Section of SEPM and Kentucky Society of Professional Geologists, Kentucky Geological Survey, 152 p.

Baird, G.C., 1976, Coral encrusted concretions: a key to recognition of a 'shale on shale' erosion surface: Lethaia, v. 9, p. 293–302.

Baird, G.C., 1978, Pebbly phosphorites in shale: a key to recognition of a widespread submarine discontinuity in the Middle Devonian of New York: Journal of Sedimentary Petrology, v. 48, p. 545–555.

Baird, G.C., 1979, Sedimentary Relationships of Portland Point and Associated Middle Devonian Rocks in Western New York: New York State Bulletin 433, 20 p.

Baird, G.C., 1981, Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian: Lethaia, v. 14, p. 105–122.

Baird, G.C. and Brett, C.E., 1986, Erosion on an anaerobic seafloor: significance of reworked pyrite deposits from the Devonian of New York State: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 57, p. 157–193.

Baird, G.C. and Brett, C.E., 1991, Submarine erosion on the anoxic seafloor: stratinomic, palaeoenvironmental and temporal significance of reworked pyrite-bone deposits, *in* Pearson, T.H., ed., Modern and Ancient Continental Shelf Anoxia: Geological Society Special Publication 58, p. 233–257.

Baum, G.R. and Vail, P.R., 1988, Sequence stratigraphic concepts applied to Paleogene outcrops, Gulf and Atlantic Basins, *in* Wilgus, C.K., Hastings, B.S., Kendall, C.G.St. C, Posamentier, H., Ross, C.A., Van Wagoner, J., eds., Sea-level Changes: an Integrated Approach: Society of Economic Paleontologists and Mineralogists, Special Publication 42, p. 309–328.

Bertling, M., 1999, Taphonomy of trace fossils at omission surfaces (Middle Triassic, East Germany): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 149, p. 27–40.

Bodenbender, B.E., Wilson, M.A. and Palmer, T.J., 1989, Paleoecology of *Sphenothallus* on an Upper Ordovician hardground: Lethaia, v. 22, p. 217–225.

Brett, C.E., 1974, Contacts of the Windom Member (Moscow Formation) in Erie County, New York, *in* Peterson, D.N., ed., Guidebook, Geology of Western New York State: State University of Western New York– Fredonia, p. C1–C22.

Brett, C.E., 1995, Sequence stratigraphy, biostratigraphy, and taphonomy in shallow marine environments: Palaios, v. 10, p. 597–616.

Brett, C.E. and Baird, G.C., 1996, Middle Devonian sedimentary cycles and sequences in the northern Appalachian Basin, *in* Witzke, B.J., Ludvigsen, G.A. and Day, J.E., eds., Paleozoic Sequence Stratigraphy: Views from the North American Craton: Geological Society of America, Special Paper 306, p. 213–241.

Brett, C.E. and Baird, G.C., 2002, Revised stratigraphy of the Trenton Group in the type area, central New York State: Sedimentology and tectonics of a Middle Ordovician shelf-to-basin succession: Physics and Chemistry of the Earth, v. 27, p. 231–263.

Brett, C.E. and Brookfield, M.E., 1984, Morphology, faunas and genesis of Ordovician hardgrounds from southern Ontario, Canada: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 46, p. 233–290.

Brett, C.E. and Liddell, W.D., 1978, Preservation and paleoecology of a Middle Ordovician hardground community: Paleobiology, v. 4, p. 329–348.

Brett, C.E., Goodman, W.M. and LoDuca, S.T., 1990, Sequences, cycles, and basin dynamics in the Silurian of the Appalachian Foreland Basin: Sedimentary Geology, v. 69, 191–244.

Brett, C.E., Baarli, B.G., Chowns, T., Cotter, E., Driese, S., Goodman, W. and Johnson, M.E., 1998, Early Silurian condensed intervals, ironstones, and sequence stratigraphic in the Appalachian foreland basin, *in* Landing, E., ed., Silurian Cycles: Linkages of Dynamic Stratigraphy with Atmospheric, Oceanic, and Tectonic Changes: New York State Museum Bulletin 491, p. 89–143.

Brett, C.E., Algeo, T.A. and McLaughlin, P.I., 2003a, Use of event beds and sedimentary cycles in high-resolution stratigraphic correlation of lithologically repetitive successions: the Upper Ordovician Kope Formation of northern Kentucky and southern Ohio, *in* Harries, P. and Geary, D. eds., High-Resolution Approaches in Stratigraphic Paleobiology: Topics in Geobiology 21, Kluwer Academic, Derdrecht, p. 315–350.

Brett, C.E., Turner, A.H., McLaughlin, P.I., Over, J. and Storrs, G., 2003b, Middle–Upper Devonian (Givetian–Famennian) bone/conodont beds from central Kentucky: reworking and event condensation in the distal Acadian foreland basin: Courier Forschungsinstitut Senckenberg, v. 242, p. 125–139.

Bromley, R.G., 1990, Trace Fossils: Biology, Taphonomy and Applications: Chapman & Hall, London, 361 p.

Canfield, D.E. and Raiswell, R. 1991, Carbonate precipitation and dissolution: its relevance to fossil preservation, *in* Allison, P.A. and Briggs, D.E.G., eds., Taphonomy: Releasing the Data Locked in the Fossil Record: Plenum Press, New York, p. 411–463.

Catuneanu, O., 2002, Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls: Journal of African Earth Sciences, v. 35, p. 1–43.

Cherns, L., 1980, Hardgrounds in the Lower Leintwardine Beds (Silurian) of the Welsh Borderland: Geological Magazine, v. 117, p. 311–326.

Conkin, J.E., 1986, Late Devonian New Albany–Ohio–Chattanooga shales and their interbasinal correlations in Indiana, Kentucky, and Tennessee: University of Louisville Studies in Paleontology and Stratigraphy, v. C, p. 217–259.

Cotter, E., 1988, Hierarchy of sea-level cycles in the medial Silurian siliciclastic succession of Pennsylvania: Geology, v. 16, p. 242–245.

Cotter, E. and Link, J.E., 1993, Deposition and diagenesis of Clinton ironstones (Silurian) in the Appalachian Foreland Basin of Pennsylvania: Geological Society of America Bulletin, v. 105, p. 911–922.

Donovan, S.K., ed., 1991, The Processes of Fossilization: Colombia University Press, Columbia, New York, 303 p.

Foerste, A.F., 1895, On Clinton conglomerates and wave marks in Ohio and Kentucky: Journal of Geology, v. 3, p. 50–60, 169–197.

Foerste, A.F., 1917, Intraformational pebbles in the Richmond Group, at Winchester, Ohio: Journal of Geology, v. 257, p. 289–306.

Föllmi, K.B., 1996, The phosphorus cycle, phosphogenesis and marine phosphate-rich deposits: Earth-Science Reviews, v. 40, p. 55–124.

Frakes, L.A., Francis, J.E. and Syktus, J.I., 1992, Climate Modes of the Phanerozoic; the History of the Earth's Climate Over the Past 600 Million Years: Cambridge University Press, Cambridge, 274 p.

Fürsich, F.T., 1978, The influence of faunal condensation and mixing on the preservation of fossil benthic communities: Lethaia, v. 11, p. 243–250.

Fürsich, F.T., Kennedy, W.J. and Palmer, T.J., 1981, Trace fossils at a regional discontinuity surface: the Austin/Taylor (Upper Cretaceous) contact in central Texas: Journal of Paleontology, v. 55, p. 537–551.

Fürsich, F.T., Oschmann, W., Singh, B. and Jaitly, A.K., 1992, Hardgrounds, reworked concretion levels and condensed horizons in the Jurassic of western India: their significance for basin analysis: Journal of the Geological Society, v. 149, p. 313–331.

Garrison, R.E., Kennedy, W.J. and Palmer, T.J., 1987, Early lithification and hardgrounds in Upper Albian and Cenomanian calcarenites, southwest England: Cretaceous Research, v. 8, p. 103–140.

Ghibaudo, G., Grandesso, P., Massari, F. and Uchman, A., 1996, Use of trace fossils in delineating sequence stratigraphic surfaces (Tertiary Venetian Basin, northeastern Italy): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 120, p. 261–279.

Glen, C.R. and Arthur, M.A., 1990, Anatomy and origin of a Cretaceous phosphorite–greensand giant, Egypt: Sedimentology, v. 37, p. 123–154.

Gordon, L.A. and Ettensohn, F.R., 1984, Stratigraphy, depositional environments, and regional dolomitization of the Brassfield Formation (Llandoverian) in east-central Kentucky: Southeastern Geology, v. 25, p. 101–115.

Hallam, A., 1969, A pyritized limestone hardground in the Lower Jurassic of Dorset (England): Sedimentology, v. 12, p. 231–240.

Heckel, P.H., 2002, Genetic stratigraphy and conodont biostratigraphy of upper Desmoinian–Missourian (Pennsylvanian) cyclothem successions in Midcontinent North America, *in* Bamber, W.E. and Henderson, C.M., eds., Carboniferous and Permian of the World: Canadian Society of Petroleum Geologists, Memoir, 19, p. 99–119.

Hesselbo, S.P. and Palmer, T., 1992, Reworked early diagenetic concretions and the bioerosional origin of a regional discontinuity within British Jurassic marine mudstones: Sedimentology, v. 39, p. 1045–1065.

Kauffman, E., 1977, Upper Cretaceous cyclothems, biotas and environments: Mountain Geologist, v. 14, p. 129–152.

Kennedy, W.J. and Klinger, H.C., 1974, Hiatus concretions and hardground horizons in the Cretaceous of Zululand (South Africa): Palaeontology, v. 15, p. 539–549.

Kidwell, S.M. and Bosence, D.W.J., 1991, Taphonomy and time-averaging of marine shelly faunas: *in* Allison, P.A., and Briggs, D.E.G., eds., Taphonomy: Releasing the Data Locked in the Fossil Record: Plenum Press, New York, p. 116–209.

Kim, J.C. and Lee, Y.I., 1996, Marine diagenesis of Lower Ordovician carbonate sediments (Dumugol Formation), Korea: cementation in a calcite sea: Sedimentary Geology, v. 105, p. 241–257.

Lee, Y.I. and Kim, J.C., 1992, Storm-influenced siliciclastic and carbonate ramp deposits, the Lower Ordovician Dumugol Formation, South Korea: Sedimentology, v. 39, p. 951–969.

Levorson, C.O. and Gerk, A.J., 1972, Revision of Galena Stratigraphy: Geological Society of Iowa, Field Trip Guidebook, 10 p.

LoDuca, S.T. and Brett, C.E., 1994, Revised stratigraphic and facies relationships of the lower part of the Clinton Group (middle Llandoverian) of western New York State, *in* Landing, E., ed., Studies in Stratigraphy and Paleontology in Honor of Donald W. Fisher: New York State Museum Bulletin, 481, p. 161–182.

Loutit, T.S., Hardenbol, J., Vail, P.R. and Baum, G.R., 1988, Condensed sections: The key to age determination and correlation of continental margin sequences, *in* Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C, Posamentier, H., Ross, C.A., Van Wagoner, J., eds., Sea-level Changes: an Integrated Approach: Society of Economic Paleontologists and Mineralogists, Special Publication 42, p. 183–217.

Lucas, J. and Prévôt, L.E., 1991, Phosphates and fossil preservation, *in* Briggs, D.E.G. and Allison, P.A., eds., Taphonomy: Releasing the Data Locked in the Fossil Record: Plenum Press, New York, p. 389–409.

Macquaker, J.H.S, Taylor, K.G., Young, T.P. and Curtis, C.D., 1996, Sedimentological and geochemical controls on ooidal ironstone and 'bone-bed' formation and some comments on their sequence-stratigraphical significance, *in* Hesselbo, S.P., ed., Sequence Stratigraphy in British Geology: Geological Society Special Publication 103, p. 97–107.

Malpas, J.A., Gawthorpe, R.L., Pollard, J.E. and Sharp, I.R., 2004, Ichnofabric analysis of the shallow marine Nukhul Formation (Miocene), Suez Rift, Egypt: implications for depositional processes and sequence stratigraphic evolution: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 215, p. 239–264.

Mayer, S.M., 1994, Field studies of the Middle Devonian Ludlowville–Moscow sequence in the Genesee and Seneca valleys, New York State, *in* Brett, C.E. and Scatterday, J., eds., New York Geological Association Field Trip Guidebook, 66, p. 491–504.

McLaughlin, P.I. and Brett, C.E., 2004, Eustatic and tectonic control on the distribution of marine seismites: examples from the Upper Ordovician of Kentucky, USA: Sedimentary Geology, v. 168, p. 165–192.

McLaughlin, P.I. and Brett, C.E., 2007, Signatures of sea-level rise in mixed carbonate–siliciclastic foreland basin successions: Palaios, v. 22, p. 245–267.

McLaughlin, P.I., Brett, C.E., Taha McLaughlin, S.L. and Cornell, S.R., 2004, High-resolution sequence stratigraphy of a mixed carbonate–siliciclastic, cratonic ramp (Upper Ordovician; Kentucky–Ohio, USA): insights into the relative influence of eustasy and tectonics through analysis of facies gradients: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 210, p. 267–294.

Meyer, D.L., 1990, Population paleoecology and comparative taphonomy of two edrioasteroid (Echinodermata) pavements: Upper Ordovician of Kentucky and Ohio: Historical Biology, v. 4, p. 155–178.

Neal, M.L. and Hannibal, J., 2000, Paleoecologic and taxonomic implications of *Sphenothallus* and *Sphenothallus*-like specimens from Ohio and areas adjacent to Ohio: Journal of Paleontology, v. 74, p. 369–380.

Norrish, W.A., 1991, The Lower Silurian Brassfield Formation: a condensed sequence of punctuated deposition?: Unpublished PhD Thesis, University of Cincinnati, Cincinnati, 316 p.

Osgood, R.G., 1970, Trace fossils of the Cincinnati area: Paleontographica Americana, v. 6 (41), p. 281–444.

Palmer, T.J. and Fürsich, F.T., 1974, The ecology of a Middle Jurassic hardground and crevice fauna: Palaeontology, v. 17, p. 507–524.

Palmer, T.J., 1978, Burrows at certain omission surfaces in the Middle Ordovician of the Upper Mississippi Valley: Journal of Paleontology, v. 52, p. 109–117.

Palmer, T.J. and Palmer, C.D., 1977, Faunal distribution and colonization strategy in a Middle Ordovician hardground community: Lethaia, v. 10, p. 179–199.

Palmer, T.J. and Wilson, M.A., 1990, Growth of ferruginous oncoliths in the Bajocian (Middle Jurassic) of Europe: Terra Nova, v. 2, p. 142–147.

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

Pope, M.C. and Read, J.F., 1997, High-Resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian–Cincinnatian) foreland basin rocks, Kentucky and Virginia: American Association of Petroleum Geologists Bulletin, v. 81, p. 1866–1893.

Pratt, B.R., 2002a, Storms versus tsunamis: dynamic interplay of sedimentary, diagenetic, and tectonic processes in the Cambrian of Montana: Geology, v. 30, p. 423–426.

Pratt, B.R., 2002b, Tepees in peritidal carbonates; origin via earthquake-induced deformation, with example from the Middle Cambrian of western Canada: Sedimentary Geology, v. 153, p. 57–64.

Pufahl, P.K. and Grimm, K.A., 2003, Coated phosphate grains: proxy for physical, chemical, and ecological changes in seawater: Geology, v. 31, p. 801–804.

Raatz, W.D. and Ludvigson, G., 1996, Depositional environments and sequence stratigraphy of Upper Ordovician epicontinental deep water deposits, eastern Iowa and southern Minnesota, *in* Witzke, B.J., Ludvigson, G.A. and Day, J.E., eds., Paleozoic Sequence Stratigraphy: Views from the North American Craton: Geological Society of America, Special Paper 306, p. 143–159.

Raiswell, R. and Fisher, Q.J., 2004, Rates of carbonate cementation associated with sulphate reduction in DSDP/ODP sediments: implications for the formation of concretions: Chemical Geology, v. 211, p. 71–85.

Sardeson, F.W., 1898, Intraformational conglomerates in the Galena Series: American Geologist, v. 22, p. 315–323.

Sardeson, F.W., 1908, Discoid crinoidal roots and *Camarocrinus*: Journal of Geology, v. 16, p. 239–254.

Schieber, J. and Riciputi, L.R., 2004, Pyrite ooids in Devonian black shales record intermittent sea-level drop and shallow-water conditions: Geology, v. 32, p. 305–308.

Soudry, D. and Lewy, Z., 1990, Omission-surface incipient phosphate crusts on early diagenetic calcareous concretions and their possible origin, Upper Campanian, southern Israel: Sedimentary Geology, v. 66, p. 151–163.

Stanley, S.M. and Hardie, L.A., 1998, Secular oscillations in the carbonate mineralogy of reef-building and sediment-producing organisms driven by tectonically forced shifts in seawater chemistry: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 144, p. 3–19.

Suuroja, K., Saadre, T. and Kask, J., 1999, Geology of Osmussaar Island: Estonia Maritima, v. 4, p. 39–63.

Sumrall, C., Brett, C.E., and Klock, G., in press, A new edrioasteroid from encrusted platters of the Lower Silurian McKenzie Formation of Pennsylvania: Journal of Paleontology.

Taylor, K.G., Simo, J.A., Yocum, D. and Leckie, D.A., 2002, Stratigraphic significance of ooidal ironstones from the Cretaceous Western Interior Seaway: The Peace River Formation, Alberta, Canada, and the Castlegate Sandstone, Utah, U.S.A.: Journal of Sedimentary Research, v. 72, p. 316–327.

Van Houten, F.B., 1985, Oolitic ironstones and contrasting Ordovician and Jurassic paleogeography: Geology, v. 13, p. 722–724.

Van Houten, F.B., 1990, Stratigraphic and palaeogeographic distribution of Paleozoic oolitic ironstones: *in* Scotese, C.R. and McKerrow, W.S., eds., Palaeozoic Palaeogeography and Biogeography: Geological Society Memoir 12, p. 87–93.

Van Houten, F.B. and Arthur, M.A., 1989, Temporal patterns among Phanerozoic oolitic ironstones and oceanic anoxia, *in* Taylor, W.E.G., ed., Phanerozoic Ironstones: Geological Society Special Publication 46, p. 33–49.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M. and Rathmanian, V.D., 1990, Siliciclastic Sequence Stratigraphy in Well Logs, Core, and Outcrops: Concepts for High-Resolution Correlation of Time and Facies: American Association of Petroleum Geologists, Methods of Exploration Series, 7, 55 p.

Van Wagoner, J.C. and Bertram, G.T., 1995, Sequence stratigraphy of foreland basin deposits; outcrop and subsurface examples from the Cretaceous of North America: American Association of Petroleum Geologists, Memoir 64, 487 p.

Wetzel, A. and Allia, V., 2000, The significance of hiatus beds in shallow-water mudstones; an example from the Middle Jurassic of Switzerland: Journal of Sedimentary Research, v. 70, p. 170–180.

Wilson, M.A., 1985, Disturbance and ecological succession in an Upper Ordovician cobble-dwelling hardground fauna: Science, v. 228, p. 575–577.

Wilson, M.A. and Palmer, T.J., 1992, Hardgrounds and Hardground Faunas: University of Wales, Aberystwyth, Institute of Earth Studies Publications, 9, 131 p.

Wilson, M.A. and Palmer, T.J., 2004, Calcite precipitation and dissolution of biogenic aragonite in shallow Ordovician calcite seas: Lethaia, v. 37, p. 417–427.

Wilson, M.A., Palmer, T.J., Guensburg, T.E., Finton, C.D. and Kaufman, L.E., 1992, The development of an Early Ordovician hardground community in response to rapid sea-floor calcite precipitation: Lethaia, v. 25, p. 19–34.

Wilson, M.A., Wolfe, K.R. and Avni, Y., 2005, Development of a Jurassic rocky shore complex (Zohar Formation, Makhtesh Qatan, southern Israel): Israel Journal of Earth Sciences, v. 54, p. 171–178.

Young, T.P., 1992, Ooidal ironstones from Ordovician Gondwana: a review: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 99, p. 321–347.