High-Resolution Sequence Stratigraphy of Late Mississippian (Chesterian) Mixed Carbonates and Siliciclastics, Illinois Basin

by

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(ABSTRACT)

Eight 4th-order (~400 k.y.) disconformity-bounded mixed carbonate-siliciclastic sequences were deposited in the tectonically-active, tide-dominated Illinois basin during the Late Mississippian greenhouse to icehouse transition. Detailed, lithologic cross-sections were constructed through the Chesterian Ste. Genevieve through Glen Dean interval which show an upward change in character from carbonate-dominated sequences bounded by caliche and breccia paleosols to mixed-carbonate siliciclastic sequences bounded by red, slickensided mudrock paleosols and incised valleys. The 4th-order sequences are composed of 5th-order parasequences that can be correlated basin-wide. Parasequences in the basal, dominantly carbonate sequences are composed of patchy ooid grainstone tidal ridge reservoir facies which interfinger with skeletal limestone and are capped by laterally extensive muddy carbonate units. Parasequences in the overlying mixed carbonate siliciclastic interval commonly have basal quartz sandstone valley fill and tidal sand ridge reservoir facies overlain by skeletal limestone and shale-dominated siliciclastics.

The sequences can be bundled into sequence pairs and composite sequences. Composite sequences are composed of 4 sequences and are bounded by better developed disconformities that commonly coincide with biostratigraphic zone boundaries. High energy reservoir facies are widespread in transgressive sequence tracts and late highstand

sequence tract (where present) and confined to updip areas in the early highstand sequence tracts.

Increasing amplitude 4th-order glacio-eustasy produced the sequences and caused the upward increase in incised valleys and deeper water carbonate deposition. Parasequences were produced by 5th-order glacio-eustatic sea-level fluctuations (20-100 k.y.). Sequence pairs and composite sequences were produced by 3rd-order sea-level fluctuations possibly in combination with local tectonics. Spatial and temporal variations in differential subsidence between the eastern and western shelves and the more rapidly subsiding basin interior caused variations in onlap/offlap geometries of sequences and parasequences. Increasingly wetter wet-dry seasonality caused an upward increase in siliciclastic influx and concurrent decrease in ooid deposition. The increasing-amplitude eustasy and progressively more humid climate were caused by the onset of continental glaciation on Gondwana.

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Chapter 1: Introduction

This dissertation presents results of a study of Late Mississippian (Early to Middle Chesterian) carbonate and mixed carbonate-clastic sequences of the Illinois Basin. The study is based on regional lithologic cross-sections which incorporate detailed logs of 39 outcrops and 36 cores from the Mississippian outcrop belt in Indiana, Kentucky and Illinois.

Chapter 2 presents the high-resolution sequence stratigraphy of four disconformity-bounded 4th-order sequences (numbered 1 to 4) in the Early Chesterian Ste. Genevieve and Paoli Formations. Sequences 1 and 2 are composed of ooid grainstone reservoir-bearing carbonate parasequences and are bounded by basin-wide caliche and breccia paleosol horizons . Sequences 3 and 4 are composed of carbonate and skeletal limestone-shale parasequences, and are bounded by regional caliche, breccia and slickensided mudrock paleosol horizons. These sequences were likely formed by moderate amplitude 4th-order glacio-eustatic sea-level fluctuations and modified by temporal and spatial variations in differential subsidence in the basin.

Chapter 3 presents the high-resolution sequence stratigraphy of five disconformity-bounded mixed carbonate-clastic sequences (numbered 5 to 9) in the Bethel through Glen Dean Formations. Sequences 5 to 9 are bounded by paleosols that can commonly be traced into incised valleys up to 75 meters deep. Sequences consist of tide-dominated quartz sandstone reservoir facies in the early transgressive systems tract overlain by offshore carbonate facies deposited during maximum flooding, and late highstand siliciclastics.

This paper differs from previous work in that it combines detailed sedimentology with a basin-wide stratigraphic approach to define regional disconformity-bounded sequences and their internal parasequences. This approach enables better understanding of the distribution and compartmentalization of ooid grainstone and quartz sandstone reservoirs facies, and provides a framework for future studies at the field-scale.

Chapter 2: High-Resolution Sequence Stratigraphy of Early Chesterian Ste. Genevieve and Paoli Formations, Illinois Basin

ABSTRACT

Four carbonate and mixed carbonate-siliciclastic 4th-order sequences (~400 k.y.) were deposited in the tectonically-active, tide-dominated Illinois Basin and comprise the Late Mississippian (Chesterian) Ste. Genevieve through Paoli interval. Interpretive cross-sections based on detailed measured sections of 57 closely spaced outcrops and cores show that sequence-bounding carbonate and siliciclastic paleosol horizons and distinctive marker beds can be correlated basin-wide, within the available biostratigraphic framework. Each sequence is composed of 4 to 9 regionally developed parasequences, some of which also are disconformity-bounded. Carbonate parasequences have patchy updip carbonate eolianites at the base, overlain by laterally discontinuous ooid tidal ridge/skeletal bank facies and laterally extensive muddy carbonate caps. Mixed carbonate-clastic parasequences are composed of skeletal limestone capped by fossiliferous shale. Carbonate tidal flat facies are rare in the succession.

Spatial and temporal variations in differential subsidence between the Eastern Shelf and the more rapidly subsiding Basin Interior had a major impact on distribution of parasequences within the sequences. Sequences deposited during periods of high differential subsidence have multiple, onlapping parasequences in the transgressive systems tract (TST) and progradationally offlapping parasequences in the highstand systems tract (HST). Sequences deposited during periods of moderate differential subsidence show only minor onlap in the TST, whereas HST parasequences are developed basin-wide and disconformity-bounded updip. Sequences deposited during times of low differential subsidence, there is negligible onlap in the TST and late HST parasequences are bounded by basin-wide disconformities.

The sequences were produced by moderate amplitude 4th-order glacio-eustatic sea-level changes. Parasequences were produced by higher frequency 5th-order glacio-eustatic sea-level changes. The four sequences are stacked into two 3rd-order sequence pairs and a single 3rd-order composite sequence which are bounded by well-developed exposure horizons. High-energy reservoir facies are widespread in the transgressive sequence tract of the composite sequence, but are confined to the shelves in the highstand sequence tract. The amplitude of the 4th-order sea-level changes increased through time based on the deep incision associated with the top of the study interval as well as in the overlying sequences. The dominance of 4th-order disconformity-bounded sequences and the lack of regional tidal flat facies reflect moderate-amplitude eustasy associated with the transition from greenhouse times of the Early Mississippian to times of abundant global ice in the Late Paleozoic.

INTRODUCTION

Previous work on the Early Chesterian Ste. Genevieve to Paoli interval in the Illinois Basin includes regional surface and subsurface mapping of limestone, sandstone and shale formations by Butts (1917), Weller and Sutton (1940), Swann and Atherton (1948), McFarlan et al. (1955), Swann (1963, 1964), Potter (1962), Sable and Dever (1990) and Droste and Carpenter (1990) and detailed, local sedimentologic studies by Kissling (1967), Choquette and Steinen (1980), Cluff and Lineback (1981), Liebold (1982), Merkley (1991), and Hunter et al. (in review), among others. The lateral heterogeneity of the ooid grainstone reservoirs has made the Ste. Genevieve a "correlation nightmare" in the subsurface (Zuppann, 1993) and most paleosols have been considered to be local in extent (Droste and Carpenter, 1990). Recent unpublished work has shown that some sequences and parasequences bounded by paleosols and eolianite horizons can be correlated both locally and regionally (Smith and Read, 1994; Al-Tawil et al., 1994; Smith et al., 1995; Smith and Read, 1995; Hunter et al., in review).

This paper demonstrates the value of combining detailed sedimentology with a basin-wide stratigraphic approach. Detailed lithologic logs of 57 outcrops and cores were used to construct regional lithologic cross-sections of the Ste. Genevieve to Paoli interval along more than 600 kilometers of the eastern, southern and western outcrop belts of the Illinois Basin. The cross-sections show that, despite the lateral heterogeneity of much of the stratigraphy, several sequence-bounding paleosol horizons can be correlated over much of the basin within a framework of biostratigraphic and lithologic markers (Smith and Read, 1995; Smith et al., 1995). The paleosols provide additional lithologic markers which make it possible to correlate parasequences within the sequences. The cross-sections provide an opportunity to better understand the relative effects of eustasy, tectonics and tides and to test sequence stratigraphic concepts on carbonate and mixed-carbonate clastic sequences deposited updip on a gently dipping ramp.

GEOLOGIC SETTING

Tectonic and Paleogeographic setting

The Illinois Basin formed over a failed rift zone (Burke and Dewey, 1973) where incipient rifting began in the late Proterozoic (Bond et al., 1986) and normal faulting and extension occurred into the Middle and Late Cambrian (Kolata and Nelson, 1990). Three phases of subsidence occurred in the Illinois Basin during the Paleozoic: 1) exponentially decreasing thermal subsidence from the Late Cambrian to the Middle Ordovician; 2) slow linear subsidence from the Late Ordovician to the Early Mississippian; and 3) an enigmatic four-fold increase in the subsidence rate from the Middle Mississippian to the Late Pennsylvanian (Heidlauf et al., 1986). This Late Paleozoic subsidence event likely occurred as a response to continental collision and thrust-loading in the Appalachians to the east (DeRito et al., 1983). The increase in horizontal stress caused by continental

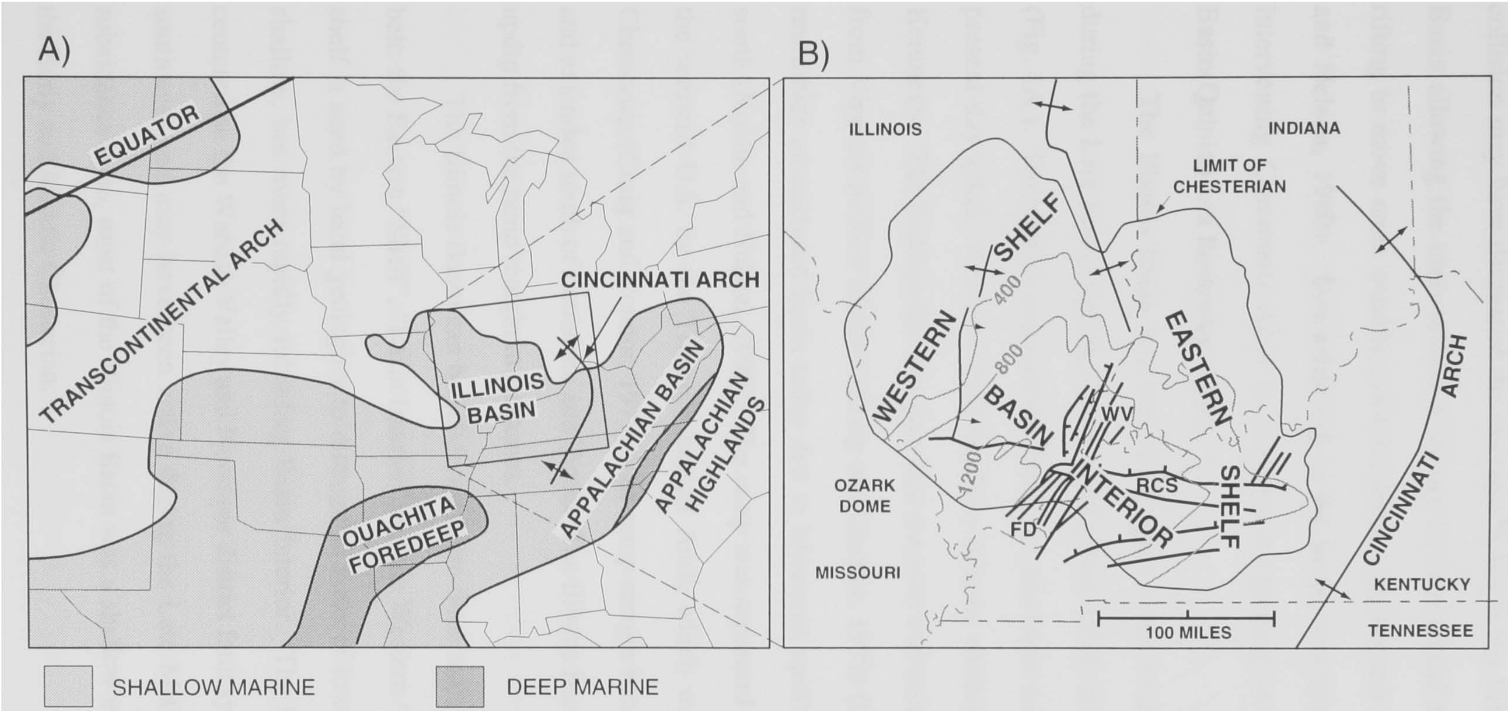


Fig. 1. A)Paleogeographic map showing regional setting of Illinois Basin (modified from Craig and Connor, 1979). B) Tectonic map of Illinois Basin.Isopachs in feet for Bethel to top of Mississippian (from Swann, 1963). 800 ft. isopach is approximate boundary of Basin Interior. WV = Wabash Valley Fault System; RCS = Rough Creek-Shawneetown Fault System; FD = Fluorspar District Fault Complex.

collision may have decreased the viscosity in the lower lithosphere beneath the Illinois Basin allowing the mass excess of denser mafic material emplaced in the crust during rifting to move more quickly toward isostatic equilibrium (DeRito et al., 1983; Kolata and Nelson, 1990). Downwarping in the basin was accompanied by uplift on the intervening Cincinnati Arch in response to loading in the Appalachian Foreland Basin(Quinlan and Beaumont, 1984).

The Illinois Basin was located between 5 and 15 degrees south of the equator during the Late Mississippian (Craig and Connor, 1979; Scotese and McKerrow, 1990) (Fig. 1A.). Throughout much of the Paleozoic, the basin was an embayment open to the present-day south, which covered southern Illinois, southwestern Indiana and western Kentucky. The embayment was part of a massive carbonate ramp that was continuous from Virginia to New Mexico (Craig and Connor, 1979) (Fig. 1A). The Illinois Basin resembles an enclosed basin today due to Mesozoic uplift of the Pascola Arch to the south (Kolata and Nelson, 1990). The ramp was separated from the Antler foredeep of the western U.S. by the Transcontinental Arch, which was emergent throughout the Chesterian (Craig and Connor, 1979). The ramp margin bordered the Ouachita Foredeep and extended south of Louisiana and Texas. The Illinois Basin lay at least 300 kilometers updip from the ramp margin in Arkansas.

The Illinois Basin can be subdivided into three main provinces along the outcrop belt: the Eastern "Shelf", "Basin Interior" and the Western "Shelf" (Fig. 1B). The term shelf is used by local geologists to distinguish areas of lower subsidence rates from the shallow, but more rapidly subsiding Basin Interior. The Basin Interior appears to be centered on the Wabash Valley and Fluorspar district fault systems which trend northeast-southwest and may have been active during the Late Mississippian. Despite varying subsidence rates, most of the Illinois Basin was a shallow epicontinental sea throughout the Early and Middle Chesterian.

Stratigraphic Setting

The carbonate ramp developed in the Middle Mississippian on and south of the Early Mississippian, siliciclastic Borden Delta which prograded from the north and west over the Illinois and Appalachian basins (Cluff and Lineback, 1981). During the Middle Mississippian, topography on and in front of the delta was filled by quartzose carbonates (Fort Payne Formation) and shallow-water ramp carbonates (Osagean to Meramecian Muldraugh, Warsaw, Salem and St. Louis Formations) (Cluff and Lineback, 1981) (Figs. 2 and 3). The St. Louis Formation directly underlies the study interval and is primarily composed of shallow-water muddy carbonates and skeletal limestone (Cluff and Lineback, 1981). The base of the Ste. Genevieve has been variously picked at different stratigraphic horizons (Sable and Dever, 1990; Droste and Carpenter, 1990; Rexroad and Fraunfelter, 1977). In this paper, the boundary is picked at a regional disconformity commonly within a muddy carbonate unit overlying the Lost River Chert of Elrod (1899) which appears to be coincident with conodont, foramanifera and coral biostratigraphic zone boundaries (Rexroad and Fraunfelter, 1977). The Lost River Chert occurs in the upper part of a laterally extensive skeletal grainstone unit and is composed of abundant chert nodules in which the outlines of fenestrate bryozoans are commonly preserved, and is a distinctive marker in outcrop and core. Maples and Waters (1987) moved the Meramecian-Chesterian Series boundary to this horizon because the fauna above the boundary more closely resemble fauna in the Middle and Upper Chesterian than the underlying St. Louis Formation. Thus, the Ste. Genevieve to Paoli interval is Lower Chesterian or Upper Visean (Middle V3b to Early V3c) in age.

The Ste. Genevieve Formation (Fig. 2) is primarily composed of ooid grainstone (a common reservoir facies), skeletal grainstone, muddy carbonates and numerous disconformities which can be recognized by the presence of caliche, breccia and eolianites. The middle and upper Ste. Genevieve Formation interfingers with the Aux

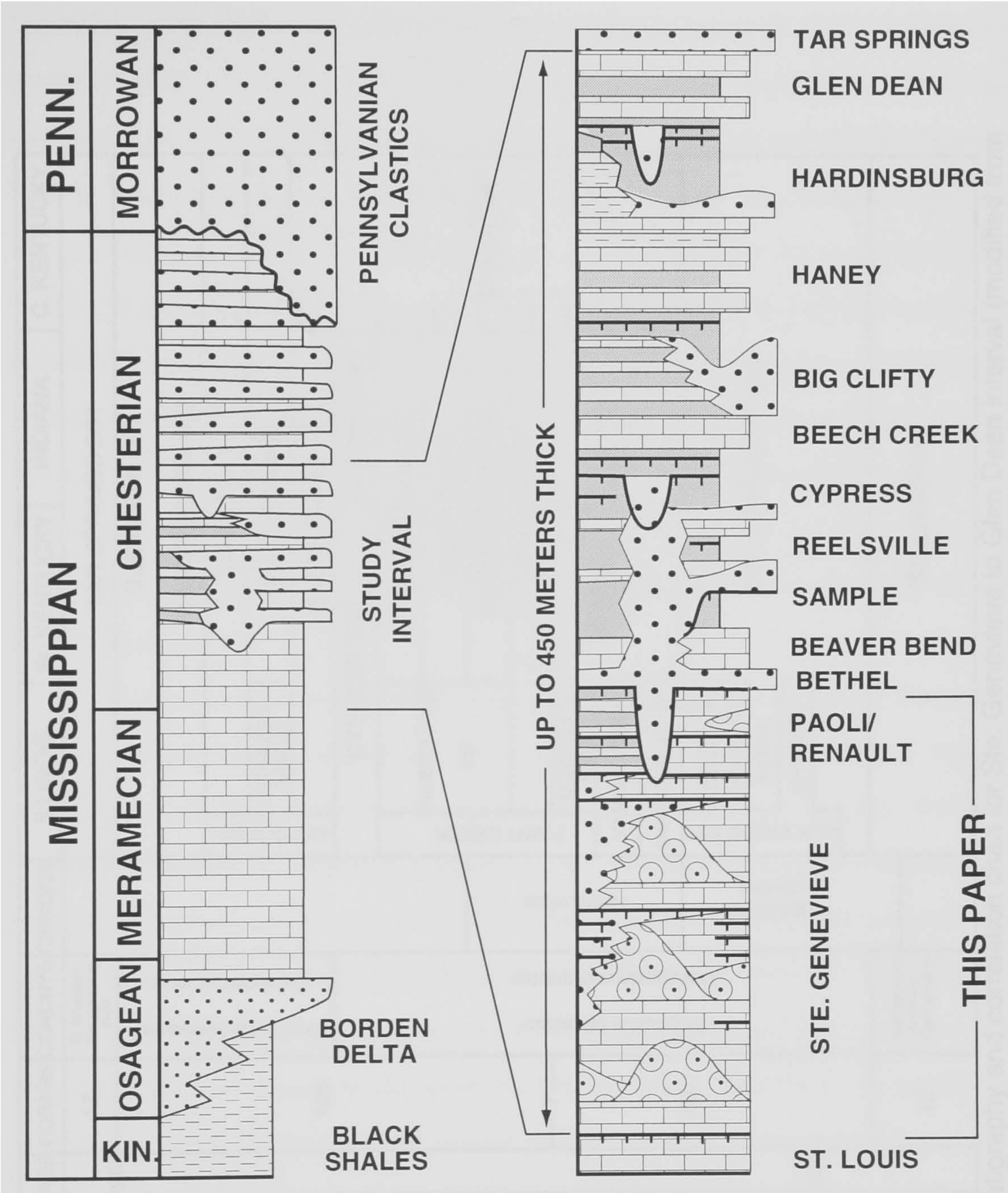


Fig. 2. Left-hand column is a generalized stratigraphic column for Mississippian of the Illinois Basin. Right-hand column shows more detailed stratigraphy of the Lower to MiddleChesterian units. See Fig. 4 for legend.

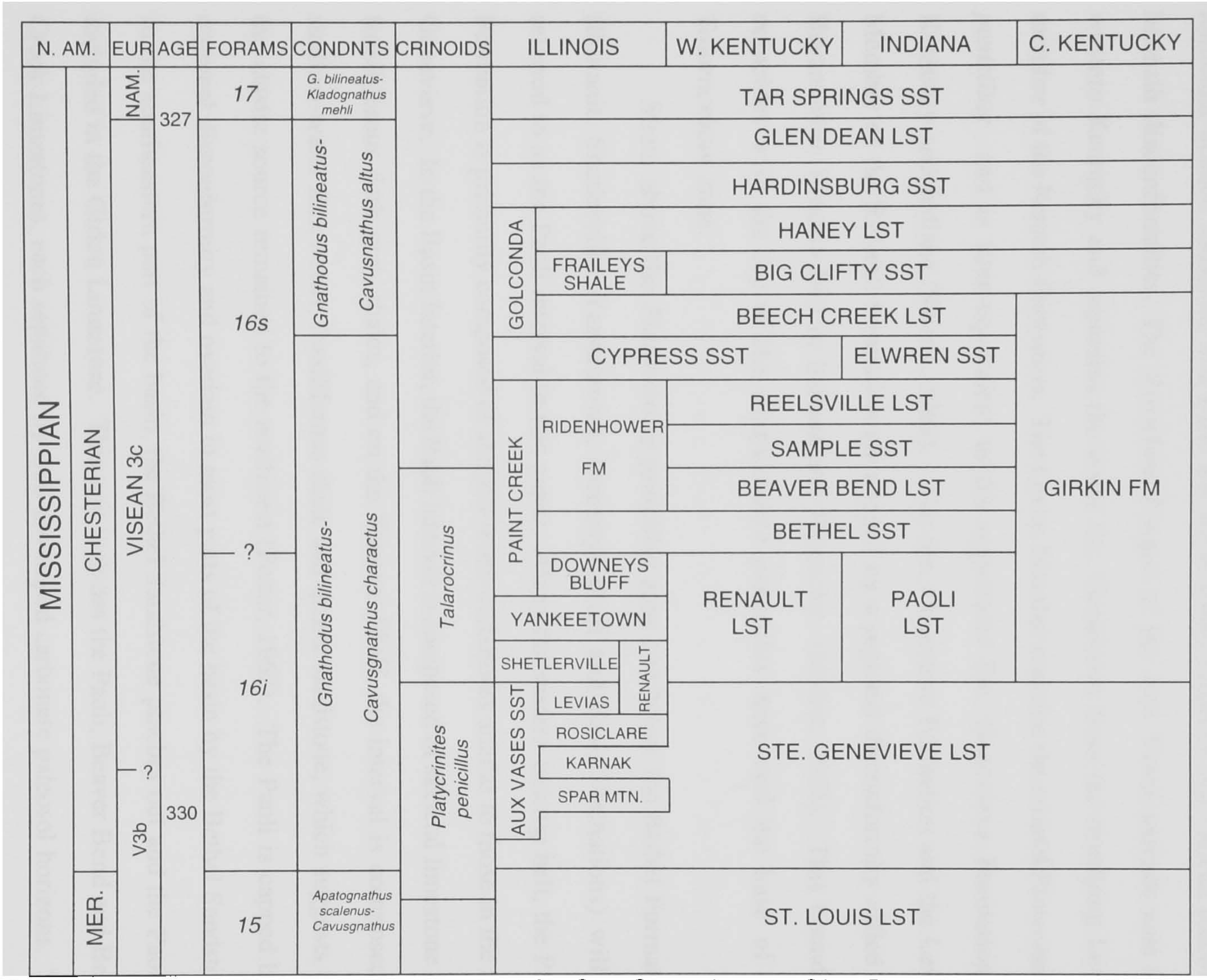


Fig. 3. Biostratigraphy and correlation chart for Ste. Genevieve to Glen Dean interval (modified from Sable and Dever, 1990 and Maples and Waters, 1987).

Vases Sandstone to the west which was derived from a source to the west/northwest (Potter, 1962). The Aux Vases is composed of quartz sandstone, flaser, wavy and lenticular bedded sandstone and shale and slickensided mudrock and caliche paleosols beneath disconformities. The Rosiclare Tongue of the Aux Vases extends east into western Kentucky and separates the lower Ste. Genevieve from the overlying Levias member of the Renault Formation. The Levias Member contains the crinoid *Platycrinites penicillus* and is time-equivalent to the uppermost Ste. Genevieve Formation in Kentucky and Indiana (Swann, 1963). The Ste. Genevieve Formation and the Levias Member of the Renault Formation are capped by a regional disconformity called the Bryantsville Breccia Bed in Indiana and Kentucky (Mallot, 1952). This boundary coincides with the top of the *Platycrinites penicillus* zone and the base of the *Talarocrinus* zone.

Strata above the *Platycrinites penicillus* zone and below the Bethel Formation (Renault, Shetlerville, Yankeetown, Downeys Bluff and Paoli Formations) will be referred to as the Paoli interval in this paper. Along the eastern outcrop belt, the Paoli Formation is primarily composed of shallow water carbonates similar to those in the Ste. Genevieve. In the Basin Interior, the Paoli interval is composed of skeletal limestone and fossiliferous dark gray shales, and on the Western Shelf, the interval is composed of shallow-water limestones, fossiliferous shale and quartz sandstone, which suggests that the clastic source remained to the northwest (Potter, 1962). The Paoli is capped by a regional disconformity and overlain in most parts of the basin by the Bethel Sandstone. In the southeastern part of the basin, the Bethel Sandstone pinches out and the Paoli is included in the Girkin Limestone. This unit includes the Paoli, Beaver Bend and Beech Creek Limestones, each separated by well-developed carbonate paleosol horizons. The Paoli Limestone is overlain by Middle and Upper Chesterian, mixed carbonate-clastic sequences (Swann, 1963, 1964; Chapter 3).

CONSTRUCTION OF LITHOFACIES CROSS-SECTIONS

Cross-sections through the study interval have been constructed based on closely spaced measured sections of core and outcrop (Fig. 4). Each core and outcrop was logged bed-by-bed and the thickness, color, grain size and shape, grain types, depositional texture, biota, sedimentary structures and evidence for subaerial exposure were noted for each unit. In outcrop, units were traced laterally to determine facies relationships where possible. The Ste. Genevieve cross-sections, A-A' and B-B', (Fig. 4) are hung on the disconformity that coincides with the top of the *Platycrinites penicillus* zone and the Bryantsville Breccia Bed in Indiana and western Kentucky. The overlying Paoli cross-sections, C-C' and D-D', (Fig. 4) have been hung on the disconformity at the top of the interval except where pre-Bethel erosion occurred. In these locations, the cross-section was hung on a regional disconformity (I) within the Paoli. Correlations were made using distinctive marker beds such as the Lost River Chert Bed, numerous paleosol horizons, and laterally extensive muddy carbonate and shale units.

Parasequences have been numbered to help describe the cross-sections. We are not suggesting that this numbering system be adopted by others because it is based on interpretive cross-sections and subject to change with the addition of new data. For clarity, measured sections have been equally spaced on the cross-sections, except in areas of major changes in thickness or larger than usual gaps between sections where they are spaced farther apart. Actual distances between sections can be determined from the location map (Fig. 4).

LITHOFACIES

Characteristics of major rock types are summarized in Tables 1 and 2 and a schematic representation of the depositional environments on the ramp is presented in Figure 5. Environmental interpretations were partly based on facies distribution on the ramp (Fig. 6). Facies will be discussed from offshore to onshore for both the carbonate-

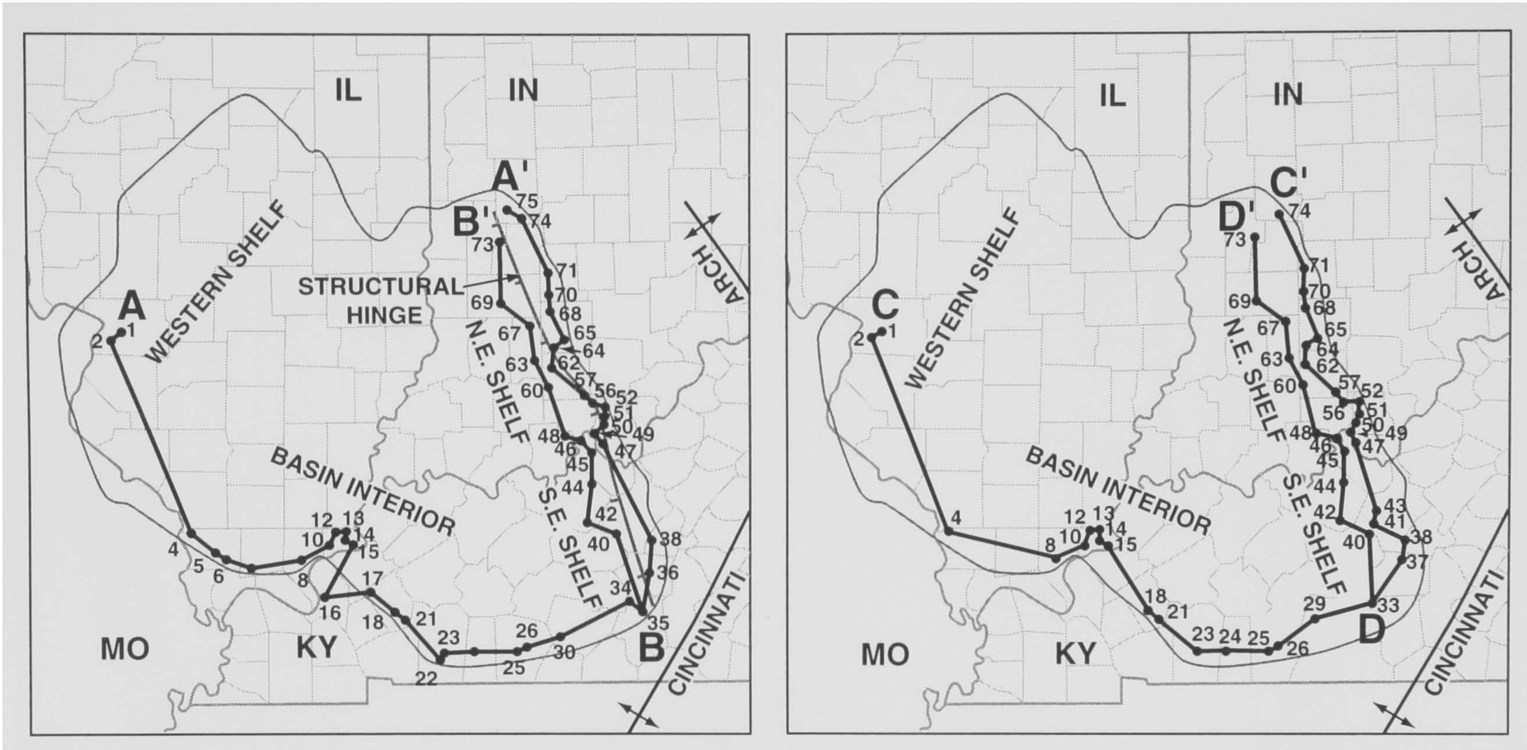


Fig. 4. Interpretive cross-sections of sequences 1 through 4 in the Ste. Genevieve through Paoli interval (on following page). Sections are equally spaced except where there are major gaps between sections. See location map for actual distance between sections. Disconformities are lettered from A to J, parasequences are numbered from 1 to 25. Bethel channel is not drawn to scale; its actual depth is 75 meters (Friberg et al., 1969). A-A' includes the Ste. Genevieve and trends west to east from location 1 to location 35, and south-north from location 35 to location 75. Cross-section B-B' shares location 35 and trends south to north parallel to and downdip of A-A'. Cross-section C-C' trends west to east from location 1 to location 33 and south to north from location 33 to location 74. Cross-section D-D' shares location 33 and trends south- north parallel and downdip of cross-section C-C'.

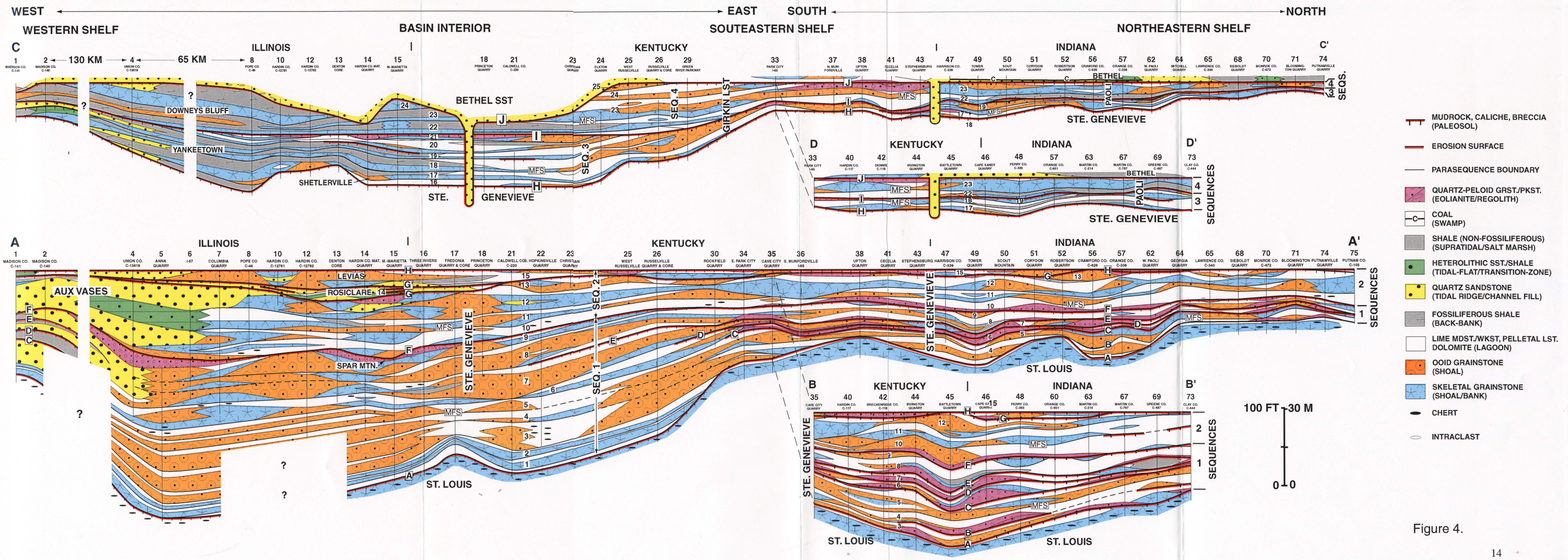

WEST
EAST
SOUTH
NORTH
WESTERN SHELF
BASIN INTERIOR
SOUTHEASTERN SHELF
NORTHEASTERN SHELF
C
C'
ILLINOIS
KENTUCKY
INDIANA
130 KM
65 KM
DOWNEYS BLUFF
YANKEETOWN
SHETLERVILLE
BETHEL SST
BETHEL
STE. GENEVIEVE
GIRKIN LST
PAOLI
SEQ. 4
SEQ. 3
SEQ. 2
SEQ. 1
SEQS.
MFS
D
D'
SEQUENCES
A
A'
AUX VASES
LEVIAS
ROSICLARE
SPAR MTN.
ST. LOUIS
B
B'
100 FT
30 M
0
MUDROCK, CALICHE, BRECCIA (PALEOSOL)
EROSION SURFACE
PARASEQUENCE BOUNDARY
QUARTZ-PELOID GRST./PKST. (EOLIANITE/REGOLITH)
COAL (SWAMP)
SHALE (NON-FOSSILIFEROUS) (SUPRATIDAL/SALT MARSH)
HETEROLITHIC SST./SHALE (TIDAL-FLAT/TRANSITION-ZONE)
QUARTZ SANDSTONE (TIDAL RIDGE/CHANNEL FILL)
FOSSILIFEROUS SHALE (BACK-BANK)
LIME MDST./WKST, PELLETAL LST. DOLOMITE (LAGOON)
OOID GRAINSTONE (SHOAL)
SKELETAL GRAINSTONE (SHOAL/BANK)
CHERT
INTRACLAST

Figure 4.

Table 1. Carbonate Ramp Facies

Facies	**Quartz-peloid grainstone (eolianite)**	**Muddy carbonates (lagoonal/ intershoal)**	**Ooid grainstone (tidal bars)**	**Skeletal grainstone/ packstone (bank to shallow shelf)**
Occurrence	Dune-forms; units commonly thicken and thin over short distances; underlie and overlie disconformities at tops and bases of sequences, respectively.	Lagoons and local intershoal swale-fills common regional traceable sheets capping parasequences.	Dip-parallel oolitic bar-forms (Zuppann, 1993); commonly occur at base of parasequences in Ste. Genevieve and Paoli of eastern shelf.	Interfingering with ooid grainstone at base of parasequences; Common throughout study interval.
Lithology	Quartz-peloid grainstone	Dolomitized and undolomitized pelletal limestone,skeletal wackestone, lime mudstone	Ooid and ooid-skeletal grainstone/ packstone	Skeletal and skeletal-ooid grainstone/ packstone
Color	Dark gray	Dolomitized units are tan, undolomitized units light gray brown to dark gray	White to light gray	Light to dark gray
Depositional texture and grain type	Grainstone; fine to medium, tightly packed and well sorted within laminae; dominated by peloids, rounded lithoclasts, sub angular quartz; lesser ooids (whole and broken), echinoderm grains.	Lime mudstone/wackestone; composed of mostly lime mud and 20-100 µm pellets; common whole thin-shelled fossils.	Grainstone, rare packstone; radial and concentric, fine to medium grained ooids with variable amounts of skeletal grains; ooid nuclei include skeletal grains, lithoclasts, quartz and pelloids;	Grainstone/packstone; fine to very coarse, subangular, skeletal fragments. Grades from grainstone to packstone with increasing water-depth. Micritization common.
Bedding and sedimentary structures	Mm-scale inverse-graded laminations form cross beds which dip <20°; cross-beds dip multi-directionally but dominantly to the north(Hunter, 1993). Rare preservation of dune crests.	Massive to burrowed, rare fenestrae, laminations and mudcracks.	Cross-bedded, less commonly thick bedded and massive; foresets dip up to 25°.	Thick bedded and massive, less commonly cross-bedded;
Terrigenous clastics admixed or interbedded	Average 15% very fine quartz sand on eastern shelf; up to 75 % quartz sand in basin interior; less quartz in units more distant from Aux Vases.	Green shale, quartz silt and quartz sand common.	Quartz grains form nuclei of ooids above some quartz sandstone units.	Quartz sand and shale very rare.
Biota	Abraded, rounded skeletal grains common.	Restricted lime mudstones: ostracodes and gastropods; In intershoal swales: bryozoans, brachiopods and echinoderms.	Echinoderm and brachiopod fragments common.	Abundant echinoderms; common brachiopods, bryozoans; rare mollusks, foramanifera, trilobites.
Diagenetic features	Fine-grained equant calcite cement lines small pores	Microcrystalline dolomite abundant; burial dolomite common; chert common in Ste. Genevieve.	fine to coarse equant calcite cement partially or completely fills voids; primary interparticle porosity.	Fine to coarse equant isopachous and syntaxial overgrowth calcite cements abundant.

Table 2. Mixed Carbonate-Siliciclastic Ramp Facies

Facies	**Shale (marginal marine/marsh)**	**Heterolithic (tidal flat and quartz sandstone-fossiliferous shale transition)**	**Quartz Sandstone (tidal bars, channel fills)**	**Fossiliferous Shale (carbonate-clastic transition)**	**Skeletal Limestone (bank, sheet)**
Occurrence	Underlies and overlies mudrock paleosols	Commonly overlies quartz sandstone and underlies salt-arsh and paleosols- more common in Bethel to Hardinsburg interval	Quartz sandstone is common throughout the Aux Vases in the western part of the basin, but rare elsewhere.	Green fossiliferous shale occurs in Ste. Genevieve of basin interior and Paoli of eastern shelf; dark gray shale occurs in basin interior Renault-Yankeetown Formations and also caps some parasequences.	Common in basin interior and western shelf in Paoli interval. Common at bases of parasequences with fossiliferous shale caps.
Lithology	Shale	Ranges from 10% to 85% quartz sand or silt interlaminated with shale	Quartz sandstone can have up to 15% shale	Fossiliferous shale and argillaceous skeletal wackestone	Skeletal grainstone/ packstone with up to 15 % shale
Color	Dark gray to olive green	Sand-white/green, shale- dark gray/green	White to light gray/green	Olive green and dark gray	Dark gray
Depositional texture and grain type	More micaceous to northwest	Very fine- to fine-grained quartz sand and quartz silt	Very-fine to medium-grained, subangular, well-sorted	Wackestone/fissile shale; whole brachiopods and bryozoans.	Grainstone/ packstone; fine to very coarse, subangular, skeletal fragments; micritization common.
Bedding and sedimentary structures	Poorly fissile	Flaser-, wavy- and lenticular-bedding and tidal rythmites (Huff, 1993).	Cross-bedded and massive bedded, flaser-bedding common	Fissile, flat-bedded	Thick bedded and massive
Biota	Plant fragments common in Bethel to Glen Dean interval, not in Aux Vases	Rare echinoderm fragments in transition facies.	Rare echinoderm fragments, could be transported or eroded from underlying limestone	Bryozoans, brachiopods and echinoderms	Abundant echinoderms; brachiopods, bryozoans; common mollusks, foramanifera; rare trilobites.

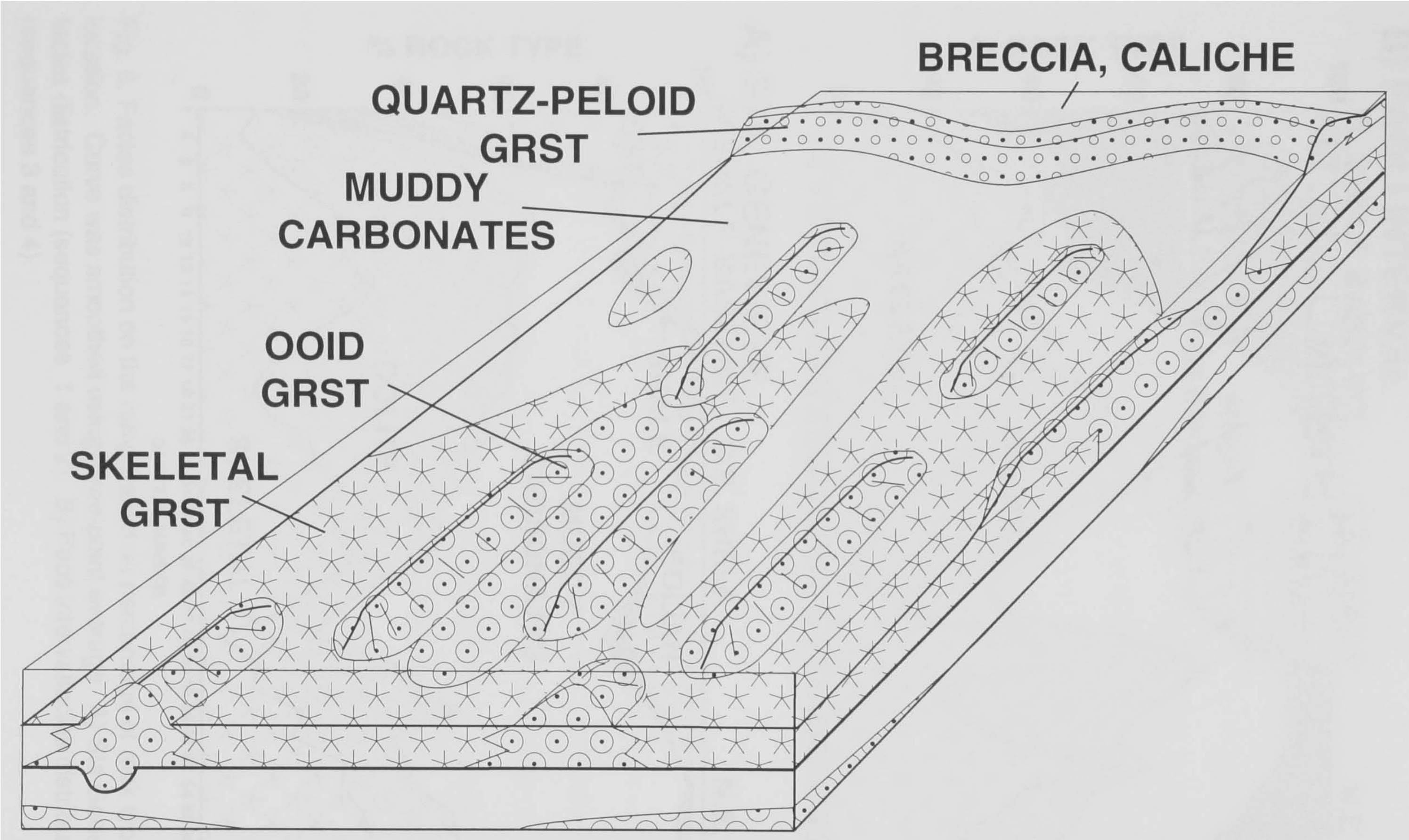


Fig. 5. Schematic depositional environment for carbonate-dominated ramp in Ste. Geneiveve Formation (not to scale).

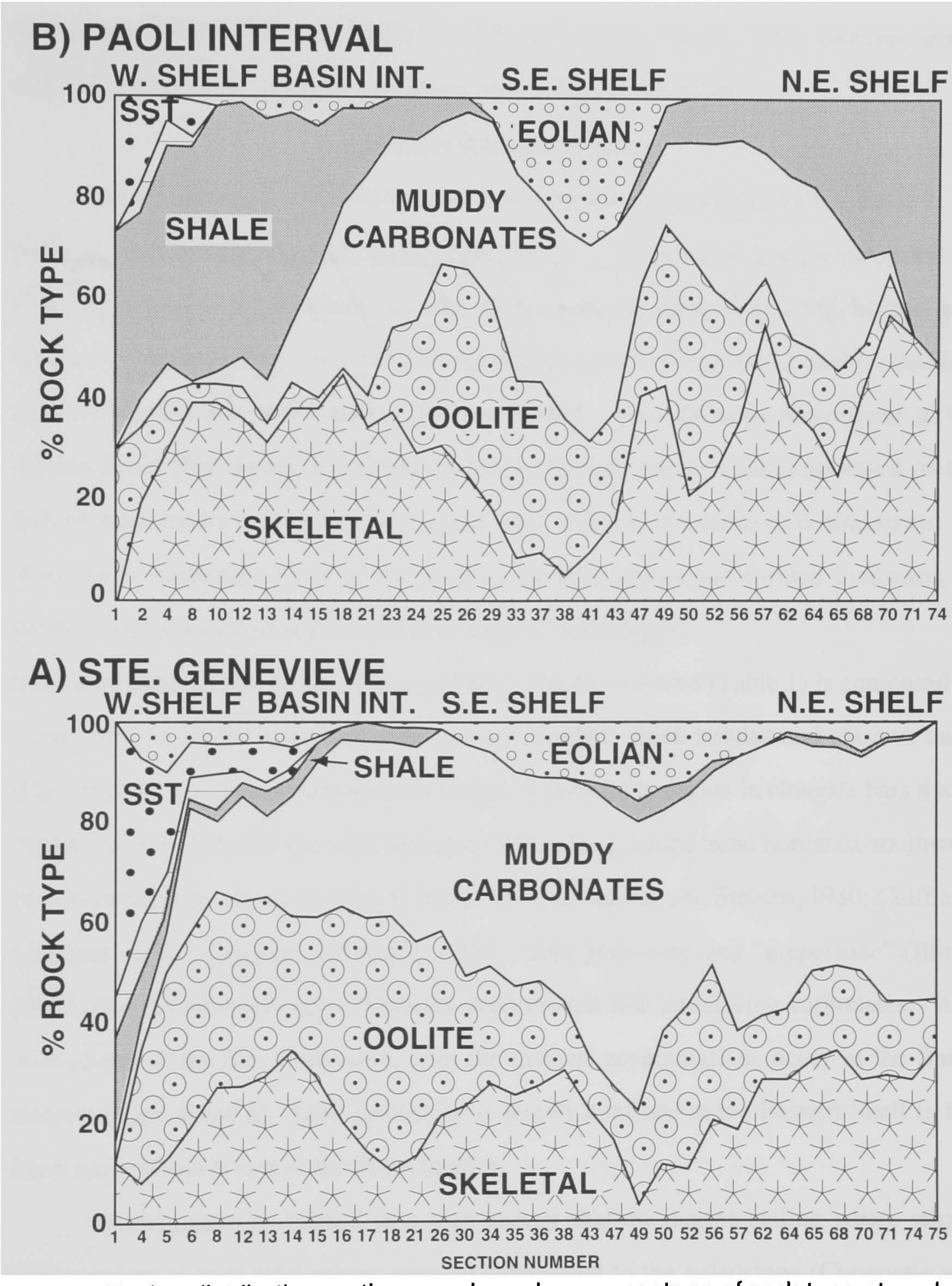


Fig. 6. Facies distribution on the ramp based on percentage of rock type at each location. Curve was smoothed using three-point average. A) Ste.Genevieve facies districution (sequences 1 and 2). B) Paoli intervalfacies distribution (sequences 3 and 4).

dominated ramp which was located in the eastern part of the basin and the mixed carbonate-siliciclastic ramp located in the western part of the basin.

Carbonate-Dominated Ramp

Facies deposited on the carbonate-dominated ramp are summarized in Table 1.

Skeletal Grainstone/Packstone (Bank/Open shelf).- The skeletal grainstone/packstone (Table 1) consists of medium to very coarse-grained echinoderm, brachiopod and bryozoan fragments (Fig. 7A). Most of the skeletal grainstone/packstone is thick-bedded, massive-bedded and less commonly cross-bedded. This facies is widespread in the Illinois Basin (Fig. 6) and formed in moderate- to high- energy conditions based on the lack of mud, medium- to very coarse-grain size, and the abundance of echinoderms. In dominantly carbonate parts of the section, skeletal grainstone formed landward, in between and seaward of ooid shoals in a range of water depths.

Ooid grainstone (Tidal Ridge, Channel Fill).- Ooid grainstone (Table 1) is composed of white, fine- to medium-grained ooids in cross-bedded, thick-bedded and massive units (Fig. 7B) and is an important reservoir facies. It commonly occurs in elongate bars 4 to 8 meters thick, 0.25 to 0.5 km wide and up to 10 km long, which trend northeast-southwest on subsurface porosity maps (Fig. 8) (Carr, 1973; Choquette and Steinen, 1980; Cluff and Lineback, 1981; Zuppann and Keith, 1988). Ooid grainstone and "grapestone" (Illing, 1954) also fill channels up to 4 meters deep eroded into underlying carbonates. It is widespread in the Ste. Genevieve, with the thickest accumulations occur in the Basin Interior (Figs. 4 and 6). Ooid grainstone is largely confined to the Eastern Shelf in the Paoli interval and is rare in the Basin Interior.

Ooid grainstone formed in a high-energy shallow-marine setting where strong tidal currents shaped tidal ridges oriented subparallel to the paleoslope (Choquette and Steinen, 1980) and cut tidal channels into underlying strata.

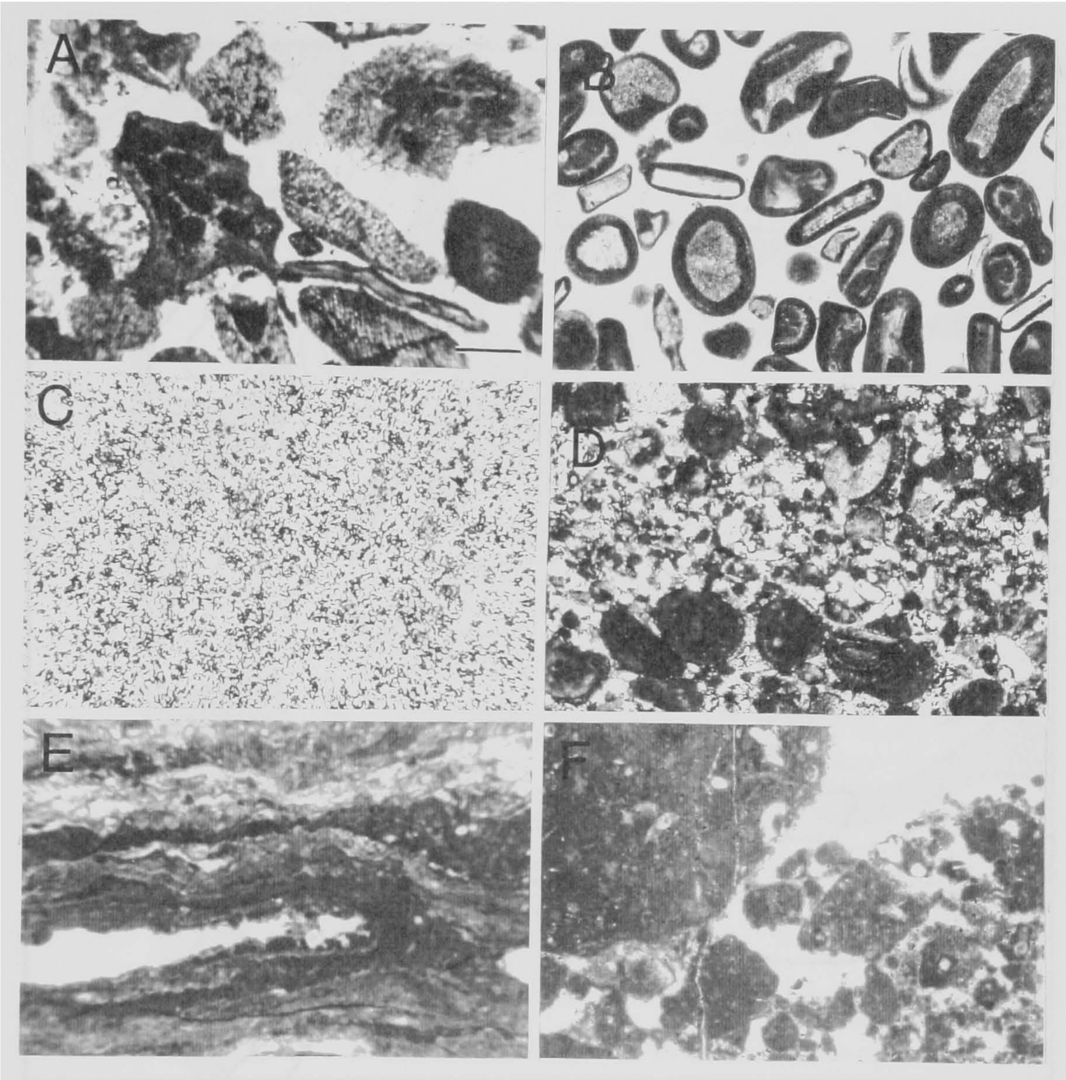


Fig. 7. Photomicrographs of important rock types. A) skeletal granstone; B) ooid grainstone; C) microcrystalline dolomite; D) quartz peloid grainstone (eolianite); E) caliche; F) breccia. Scale on A is 0.25 mm and same for all photographs.

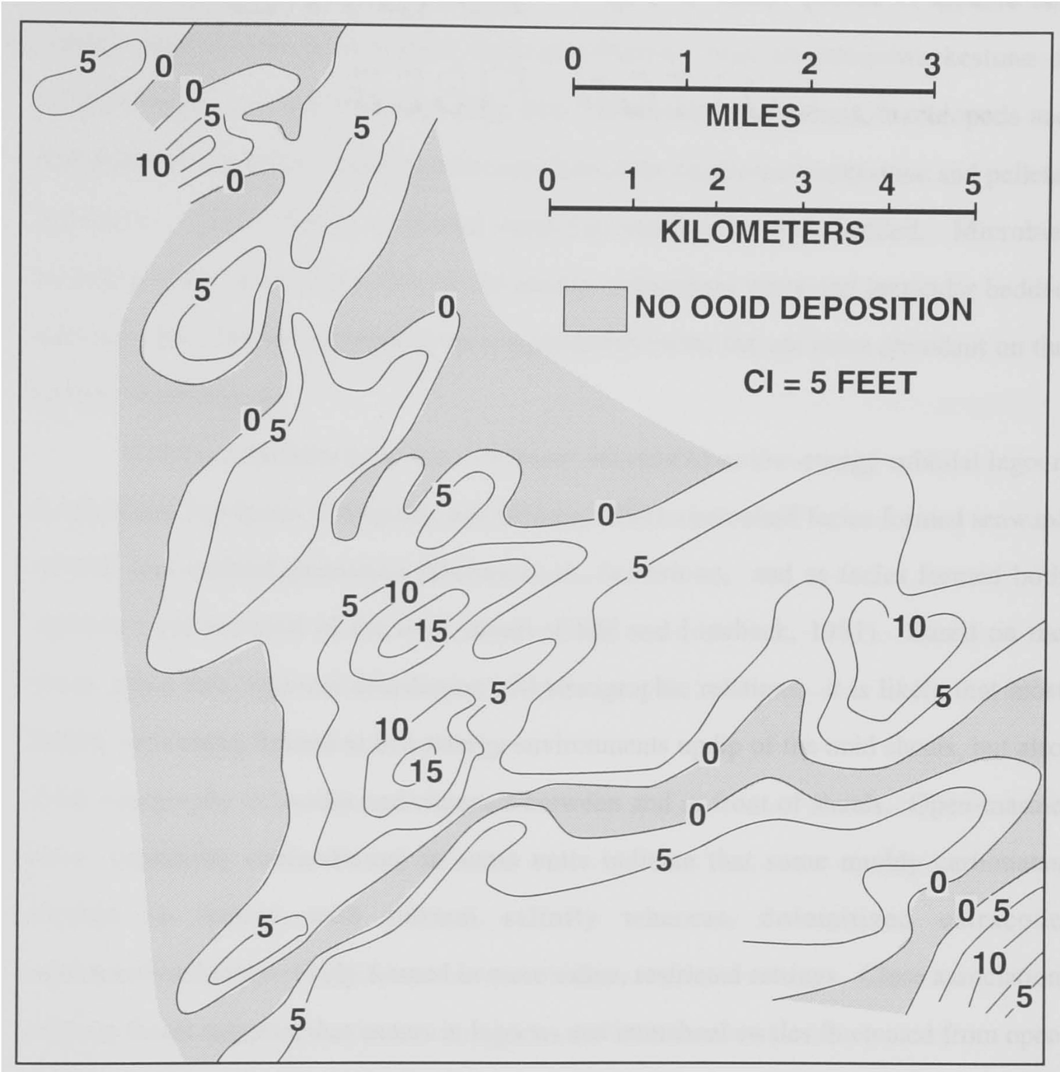


Fig. 8. Isopach map (thickness in feet) showing ridge-like form of ooid grainstone in the Ste. Genevieve Limestone if east central Illinois near the LaSalle Aniticline (modified from Choquette and Steinen, 1980 - isopach of "C-horizon" only). Northeast-southwest trending tidal ooid ridges formed parallel to paleolope and were shaped by strong tidal currents.

Muddy Carbonates (Lagoon/Intershoal).- Muddy carbonates (Table 1) include tan, earthy, microcrystalline dolomite, light gray-brown lime mudstone/wackestone or medium brown pelletal limestone (Fig. 7C). Ostracodes, bryozoans, brachiopods and fine-grained echinoderm fragments are common in the mudstone/wackestone and pelletal limestone. Most muddy carbonate units are massive or thick-bedded. Microbial laminites make up less than 10% of the muddy carbonates; wavy and lenticular bedded units are rare. Muddy carbonates are a widespread facies but are more abundant on the Eastern Shelf (Fig. 6).

Muddy carbonates have been variously interpreted as low-energy subtidal lagoon or nearshore bay facies (Choquette and Steinen, 1980), open-shelf facies formed seaward of ooid and skeletal grainstone (Hunter et al., in review), and as facies formed both landward and seaward of the ooid shoals (Cluff and Lineback, 1981). Based on the fauna, grain size, regional distribution and stratigraphic relations, it is likely that most muddy carbonates formed in low-energy environments updip of the ooid shoals, but also filled topography in swales and channels between and in front of shoals. Open-marine biota (especially echinoderms) in some units indicate that some muddy carbonates formed in waters with normal salinity whereas, dolomitized ostracode mudstone/wackestone likely formed in more saline, restricted settings. Close association of these facies suggests that waters in lagoons and intershoal swales fluctuated from open marine to restricted. The lack of laminites suggests that microbial mat-covered carbonate tidal flats and were rare despite the arid Late Mississippian climate.

Quartz-Peloid Grainstone (Eolianite/Beach).- Quartz-peloid grainstone (Table 1) is composed of well-sorted, non-fossiliferous, very-fine- to medium-grained rounded peloids, broken ooids and subangular quartz grains (Fig. 7D). Sedimentary structures include razor-sharp millimeter-scale laminae with inverse grading, preserved barchan dune crests, rhizoliths, low-angle cross-bedding (5 to 25° dips), and subhorizontal bedding (1 to 5° dips) (Hunter, 1993; Dodd et al., 1993). Quartz peloid-grainstone commonly overlies or is capped by caliche and breccia and occurs in beds that are up to 6 meters thick and that thicken and thin over short distances. Quartz-peloid grainstone is most common on the central Eastern Shelf in the Ste. Genevieve and on the Southeastern Shelf in the Paoli interval, but also occurs beneath disconformities in the Basin Interior.

Cross-bedded, quartz-peloid grainstone units are eolianites, based on the inverse grading, sorting, rhizoliths, rounded broken ooids, the lack of fossils or coarse grains, and the close association with carbonate paleosols (Merkely, 1991; Hunter, 1993; Dodd et al., 1993). Some sub-horizontally laminated quartz-peloid grainstone with coarse-grained fossiliferous laminae may be beach facies. Subhorizontal inverse graded units lacking coarse laminae characterize eolian translatent strata, formed by climbing wind ripples (Hunter, 1993). Quartz-peloid grainstone eolianites have also been reported in time-equivalent strata in Kansas (Abegg, 1994), Eastern Kentucky and West Virginia (Al-Tawil and Read, 1996).

Paleosols.- Paleosols on the carbonate units include quartz-peloid packstone, caliche and brecciated carbonate. Quartz-peloid packstone consists of dark gray weathering peloids, very fine- to fine-quartz sand and ooids in a clay matrix. Grains commonly are abraded and coated with caliche, and pebble-sized lithoclasts eroded from the subjacent bed are common. This facies occurs at the base or top of quartz-peloid grainstone beds, overlying breccia and caliche or as a veneer on marine carbonates.

Late Mississippian caliche (Fig. 7E) is gray, brown or tan, massive to laminated cryptocrystalline calcium carbonate that commonly has rootlet-casts filled with sparry calcite (Walls et al., 1975; Liebold, 1982; Ettensohn et al., 1988). Caliche coats grains, lines vertical fractures and bedding planes, or forms thick accumulations of laminar caliche beneath disconformities. Laminar caliche is up to 30 cm thick, and contains alternating hues of light and dark brown wavy-banded micrite in subhorizontal layers and stringers that extend up to 4 meters below the disconformity. Silcrete is present but is less common than caliche.

Brecciated carbonate/regolith forms argillaceous, brown-weathering recessive intervals characterized by angular to subrounded fragments of limestone forming fitted-fabrics in which fractures are filled with sparry calcite and internal sediment (Fig. 7F). It commonly is associated with thick laminar caliche and silicified limestone. Tepee-structures up to a meter wide and 30 cm high occur at the tops of some well-developed breccia/regolith profiles. Breccia is best developed in the Bryantsville Breccia Bed, but is common at each of the sequence boundaries.

Quartz-peloid packstone, caliche and breccia are paleosols that formed during subaerial exposure of carbonate rocks in a wet-dry climate. In the early stages of carbonate paleosol development, quartz-peloid packstone formed when eolian peloids, quartz sand and ooids accumulated along with clay carried by wind or sheet-wash. Some lithoclasts eroded from underlying carbonates were transported and deposited in swales between eolian dunes and on exposure surfaces (Hunter, 1993); other lithoclasts may be in-situ pedogenic concretions of caliche analogous to calcrete ooids or pisolites (Read, 1974). With continued exposure, caliche-coated grains and interstitial fine-grained carbonate was precipitated to form gray brown, quartz-peloid packstone paleosols. With time, the sediments were cemented, and laminar caliche lined vertical fractures, joints, and bedding planes. In the most mature profiles, exposed carbonates were brecciated, and laminar caliche crusts and subhorizontal layers of caliche formed up to 4 meters below the exposed surface along joints and bedding planes (Walls et al., 1975; Ettensohn, 1975; Harrison and Steinen, 1978; and Ettensohn et al., 1988).

Mixed Carbonate-Siliciclastic Ramp

Facies deposited on the mixed carbonate clastic ramp are summarized in Table 2. Interpretations of these facies are based in part on study of the mixed carbonate-siliciclastic facies in the overlying Bethel to Glen Dean interval (Chapter 3).

Skeletal Grainstone/Packstone (Open Shelf/Bank).- This facies is similar to the skeletal grainstone/packstone found in the carbonate dominated shelves but is commonly darker, has higher terrigenous clay content and is thin- to thick-bedded but rarely cross-bedded (Table 2). Where associated with siliciclastics, this facies likely formed in a storm-dominated environment offshore from the siliciclastic source and in a deeper, more distal environment than in the Ste. Genevieve. This is supported by the lack of cross-bedding and interfingering ooid grainstone.

(Fossiliferous) Shale (Low-Energy Shelf).- Dark gray to black fossiliferous shale with whole bryozoans and brachiopods (Table 2) occurs in the Yankeetown and Downeys Bluff Formations in the Basin Interior. Graded beds of skeletal limestone are common in most fossiliferous shale units. Dark gray fossiliferous shale grades laterally into muddy carbonates to the east, and quartz sandstone to the northwest. Sandy, olive green fossiliferous shale with bryozoans, brachiopods and echinoderms interbedded with limestone occurs in the Ste. Genevieve of the Basin Interior and the Paoli Formation of the Eastern Shelf. Fossiliferous shale commonly overlies, underlies and interfingers with muddy carbonates.

Fossiliferous shale is the most distal of the siliciclastic facies and represents a gradation between carbonate and siliciclastic environments. Dark gray to black units formed in quiet, poorly oxygenated water downdip of quartz sandstone. Common, thin, graded beds of skeletal debris were deposited during storms in the otherwise low-energy setting (Treworgy, 1985; Harris, 1992). It grades laterally into muddy carbonates, but may have formed in a slightly deeper water environment based on its storm-influenced rather than tidally-influenced sedimentary structures. Green fossiliferous shale interbedded with limestone formed in quiet, more oxygenated water and the interbedding could be due to periodic influx of skeletal debris carried in by storms or tides.

Quartz Sandstone and (Sand Ridges/Shallow marine).- Quartz sandstone is fine-grained, subangular and occurs in bi-directionally cross-bedded, thick-bedded and massive-bedded units (Table 2). Quartz sandstone commonly occurs in northeast-southwest trending convex up elongate bars up to 10 meters thick, based on isopach maps (Seyler, 1986; Huff, 1993; Leetaru, 1991; Udegbunam et al., 1993). It also fills lithoclastic conglomerate-floored channels up to 5 meters deep and less commonly occurs in sheets up to 10 meters thick (Rice et al., 1993). Quartz sandstone occurs in the Aux Vases Formation of the Western Shelf and in the Spar Mountain and Rosiclare Members of the Aux Vases Sandstone that extend into the Basin Interior. In the Yankeetown Formation, quartz sandstone is restricted to the northwestern part of the basin (Swann, 1963).

Quartz sandstone was deposited in a tide-dominated shallow marine environment based on abundant channel-form structures with tidal fills, bi-directional cross-bedding, and sand-ridges oriented parallel to the paleoslope (Seyler, 1986; Cole, 1990; Huff, 1993). Although they are somewhat smaller, the sand-ridges resemble those found in tide-dominated environments such as the North Sea (Off, 1963; Huff, 1993). Quartz sandstone and common sandy conglomerates fill channels scoured by strong tidal currents. These channels were not formed by fluvial processes during lowstand because they are neither filled with fluvial facies nor can they be traced laterally into exposure surfaces. Channels incised during sea-level lowstands in the overlying Bethel, Sample and Hardinsburg Formations can be traced laterally into paleosol horizons.

Heterolithic Facies (Tidal Flat/Transition Zone).- Flaser-, wavy- and lenticular-bedding and rhythmic interlamination are common in mixed sandstone and shale heterolithic units (Table 2). Some interlaminated sandstone and shale units are tidal rhythmites (Huff, 1993). Flaser-, wavy-, and lenticular-bedded sandstone and shale form on siliciclastic tidal flats and in the transition zone between sand-ridges or barrier islands and offshore muds (van Stratten, 1954; Reinick and Wunderlich, 1968). Based on its stratigraphic distribution, this facies is interpreted to have formed both on tidal flats and in the transition zone seaward of quartz sand-ridges.

Paleosols.- Paleosols on the mixed carbonate-siliciclastic ramp are most commonly blocky, slickensided mudrock clay paleosols which formed on exposed shale beds, but caliche and breccia are also common exposure features on quartz sandstone. The red, green, or dark gray mudrock paleosols are brecciated, root disrupted and commonly have brecciated carbonate nodules and tepee structures (Ambers and Petzold, 1992). These paleosols and associated disruption features formed on exposed siliciclastic facies in an arid climate (Ambers and Petzold, 1992). The brecciated carbonate nodules developed where thin dolomite beds, which had formed in water-filled swales on the exposed surface, were subjected to prolonged exposure in an arid to semi-arid climate (Ambers and Petzold, 1992).

Relative Ranking of Disconformities

Disconformities (lettered from A to J on cross-sections) can be crudely ranked in terms of the relative duration of exposure, estimated from the relative abundance of exposure features (Fig. 9). From immature to most mature, the exposure features are: 1) eolianite; 2) "dirty" packstone and minor caliche; 3) thick laminar caliche extending deep beneath the surface; and 4) brecciated or karsted limestone. Blocky, slickensided mudrock paleosols and incised valleys will be treated separately because they are difficult to rank relative to carbonate exposure features.

Disconformities H and J, on which the cross-sections are hung, are the best developed and likely represent the most widespread and longest period of emergence (major disconformities). Disconformity H has the highest occurrence of caliche and

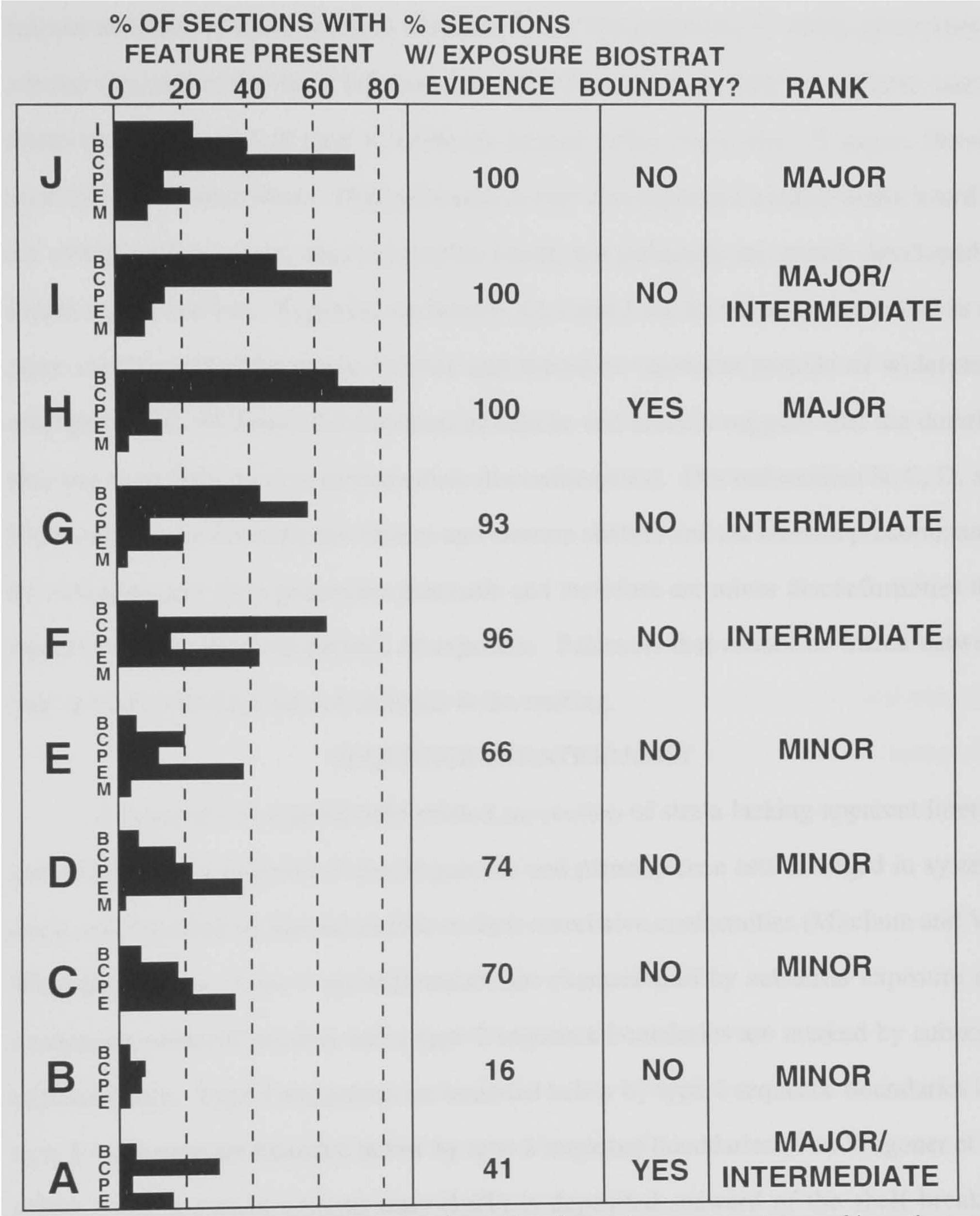


Fig. 9. Relative Ranking of disconformities based on percentage of locations measured with breccia (B), caliche (C), dirty packstone (P), eolianites (E) and slickensided mudrock (M). Incision (I) only occurs on the uppermost boundary and is a better measure of how far sea-level fell rather than the duration of exposure. Disconformities that coincide with biostratigraphic boundaries may represent relatively long-term exposure.

breccia and there is major incision associated with Disconformity J. Although incision is a better measure of sea-level fall than duration of exposure, it likely would have taken a relatively long period of time to erode an incised valley more than 75 meters through underlying carbonate strata. Disconformity A may also represent a major hiatus based on the evidence for a major biostratigraphic break, but paleosols are poorly developed or absent at this horizon. Exposure surfaces F, G, I and J can be traced into sections in the more rapidly subsiding Basin Interior and therefore represent periods of widespread emergence, but the lesser development of caliche and breccia suggests that the duration may not have been as long (intermediate disconformities). Disconformities B, C, D, and E can only be traced along the eastern and western shelves and are marked predominantly by eolianites and dirty packstone paleosols and therefore are minor disconformities that represent relatively short periods of exposure. Paleosols that cannot be traced between four or more locations are not included in the ranking.

SEQUENCE STRATIGRAPHY

A sequence is a genetically related succession of strata lacking apparent internal unconformities, composed of parasequences and parasequence sets arranged in systems tracts and bounded by unconformities or their correlative conformities (Mitchum and Van Wagoner, 1991). Type 1 unconformities are characterized by subaerial exposure and concurrent subaerial erosion and a type 2 sequence boundaries are marked by subaerial exposure only. Type 1 sequences are bounded below by type 1 sequence boundaries and type 2 sequences are bounded below by type 2 sequence boundaries (Van Wagoner et al., 1988). The lowstand systems tract (LST) is deposited seaward of the shelf break or depositional shoreline break and is composed of one or more progradational to aggradational parasequence sets. The transgressive systems tract (TST) is composed of one or more retrogradational parasequence sets and is capped by the maximum flooding surface (MFS). The overlying highstand systems tract is composed of one or more aggradational to progradational parasequence sets (Mitchum and Van Wagoner, 1991). There are four disconformity-bounded sequences in the Ste. Genevieve to Paoli interval which are each composed of 4 to 9 parasequences.

A parasequence is a relatively conformable succession of genetically related beds and bedsets bounded by marine flooding surfaces and their correlative surfaces (Mitchum and Van Wagoner, 1991). Facies within parasequences commonly exhibit a shallowing upward trend and their boundaries are marked by rapid deepening events (c.f. James, 1984; Goodwin and Anderson, 1985; Mitchum and Van Wagoner, 1991). We have only designated units that are regionally traceable as parasequences. Parasequences are 1 to 8 meters thick on the Eastern Shelf, where some parasequences are disconformity-bounded, and 4 to 12 meters thick in the Basin Interior where they are generally conformable except near some upper sequence boundaries. Most parasequences can be traced across the Eastern Shelf and into the Basin Interior using regional disconformities and extensive muddy carbonate marker beds. Some parasequences are confined to local topographic lows and form multilateral, discontinuous units, especially beneath sequence boundaries.

Typical parasequences consist of, from top to base:

6) disconformity or conformable boundary;

5) locally developed quartz-peloid grainstone (eolianite) units;

4) laterally extensive muddy carbonate or fossiliferous shale units;

3) laterally discontinuous ooid grainstone ridges and channel fills which overlie, underlie and pass laterally into skeletal grainstone sheets and banks;

2) locally developed quartz-peloid grainstone (eolianite);

1) disconformity or conformable boundary.

Most parasequences are asymmetric with high-energy grainstone bases capped by lower energy muddy carbonate/fossiliferous shale which suggests rapid flooding followed by upward shallowing. Rare symmetrical parasequences have thin basal muddy carbonate units. Some laterally extensive muddy units that appear to be at the base of parasequences are updip extensions of parasequences that have basal grainstone units downdip (see parasequence 10 - Fig. 4).

The facies within sequences 1 through 4 vary from the Western Shelf to the Basin Interior to the Eastern Shelf and will be discussed individually to illustrate their differences and similarities. Only brief descriptions of the Western Shelf strata will be given due to a lack of data.

Sequence 1

Sequence 1 is bounded at the base by disconformity A and at the top by disconformity F. It includes the Fredonia member and the lower McCloskey oolite reservoirs of the Ste. Genevieve, the lower half of the Aux Vases Formation on the Western Shelf and is capped by the Spar Mountain Member of the Aux Vases in the Basin Interior. Sequence 1 is 27 meters thick on the Western Shelf, thickens up to 63 meters thick in the Basin Interior, thins to 20 meters near the Cincinnati Arch and 9 meters on the Northeastern Shelf.

The lower boundary is a type 2 sequence boundary that coincides with a major biostratigraphic zone boundary at the St. Louis-Ste. Genevieve contact (Rexroad and Fraunfelter, 1977; Maples and Waters, 1987). Caliche and eolianites mark the boundary in the Basin Interior and the downdip sections of the Eastern Shelf (cross-section B-B'). Updip on the Eastern Shelf the boundary is picked at an undulose surface in a dolomite bed overlying the Lost River Chert Bed (Merkely, 1991; Hunter et al., in review; Rexroad, pers. comm., 1996). Well-developed breccia and conglomerate have been reported at the basal boundary in Missouri and in the Appalachian basin (Weller and Sutton, 1940; Al-Tawil, 1996). Although it appears that there was topography on the Lost River Chert on cross-section A-A' on the northern part of the Eastern Shelf, the cross-section is composed of locations that are on either side of the hinge of a northwest-southeast trending structural high that caused thinning of up to 16 meters over 35 km in the basal part of sequence 1 (see location map, Fig. 4). Locations 29, 35, 36, and 40-46 were on the structural high and sections 26, 32, 33, 37, 38 and probably 39 were on the more rapidly subsiding basin-ward side of the hinge. This structure parallels the northwest-southeast trending portion of the Cincinnati Arch and the Lasalle Anticlinal Belt (Fig. 1B).

The TST of sequence 1 is composed of four onlapping parasequences (1-4) (Fig. 4). Parasequences 1 and 2 are thin, regional skeletal grainstone-muddy carbonate parasequences that onlap the margin of the Eastern Shelf and are only present in the Basin Interior. Parasequences 3 and 4 are well defined ooid grainstone-muddy carbonate units in the downdip sections of the Eastern Shelf (cross-section B-B') that onlap the structural high in the northeastern part of the basin. In the Basin Interior, it is difficult to correlate these parasequences because they are commonly amalgamated in thick ooid grainstone and muddy carbonate deposits (locations 14, 17 and 22). The MFS is picked at the base of the grainstone unit in parasequence 5 because it is the first laterally extensive grainstone to be deposited on the structural high in the northeastern part of the basin.

The HST is composed of five through-going parasequences (5-9). Parasequence 5 can be correlated across the Eastern Shelf but is difficult to correlate in parts of the Basin Interior because it is amalgamated with parasequences 3 and 4 in sections 14, 19 and 22 and correlates laterally with two parasequences at locations 4, 9, 11 and 26. Parasequences 5, 6, 7 and 9 are capped by off-lapping disconformities on the Eastern Shelf and are primarily composed of muddy carbonates and eolianites. These HST parasequences are conformable in much of the Basin Interior and are locally amalgamated into thick ooid grainstone units (sections 4 and 9). Parasequence 9, which underlies the sequence boundary, consists of thin multilateral grainstone units confined to local lows on the Eastern Shelf , and is capped by an eolianite in the Basin Interior. The eolianite is the Spar Mountain Member of the Ste. Genevieve which passes westward into marine siliciclastics which prograded east at the top of the sequence (Swann, 1964). On the Western Shelf, the HST is composed of siliciclastic parasequences bounded by 4 paleosols which may correlate with the four regional disconformities on the Eastern Shelf (disconformities C to F). The sequence is capped by disconformity F.

Sequence 2

Sequence 2 is bounded at the base by disconformity F and at the top by disconformity H. It is composed of the Karnak, Joppa Members of the Ste. Genevieve, the Rosiclare Member of the Aux Vases and the Levias Member of the Renault Formation. Sequence 2 is 16 meters thick on the Western Shelf, thickens to 34 meters in the Basin Interior and thins to 13 to 15 meters on the Eastern Shelf. Sequence 2 is composed of tidally-influenced siliciclastics on the Western Shelf which cannot be subdivided without a more extensive data set.

Disconformity F is an intermediate-rank paleosol that commonly overlies eolian facies in the Basin Interior and underlies eolianites on the Eastern Shelf (Fig. 9). The TST of sequence 2 consists of parasequence 10 which is composed of patchy eolianites and grainstone capped by a 1 to 10 meter thick muddy carbonate unit that can be correlated over most of the Basin Interior and Eastern Shelf. The MFS is picked at the base of parasequence 11 because it is the first laterally extensive grainstone unit to be deposited on the Eastern Shelf. The HST is composed of 5 parasequences (11-15). Parasequences 11 and 12 are generally conformable grainstone-muddy carbonate units that can be correlated across most of the Basin Interior and Eastern Shelf. On the Eastern Shelf, the muddy cap to parasequence 11 is absent in many sections which suggests either non-deposition or erosion at the base of parasequence 12. Parasequence 13, which underlies disconformity G', consists of multi-lateral grainstone units and is laterally discontinuous across the basin. Parasequence 14 is a disconformity-bounded parasequence composed of tidally-influenced siliciclastics and multiple local disconformities that is only present in the Basin Interior and is capped by a well-developed unconformity (disconformity G'). The HST is capped by a basin-wide disconformity bounded parasequence (15) predominantly composed of grainstone and minor siliciclastics in the Basin Interior, eolianites on the Southeastern Shelf and muddy carbonates and grainstone on the Northeastern Shelf. Parasequence 15 may be an updip extension of a sequence that is better developed downdip because it has well-developed paleosols at the base and top. The lower boundary caps siliciclastics of the Rosiclare Tongue of the Aux Vases that prograded from the west up to the top of parasequence 14 and then backstepped with flooding of parasequence 15. However, because it is thin and composed of a single parasequence, it will be treated as a disconformity-bounded parasequence at the top of sequence 2. The sequence is capped by disconformity H.

Sequence 3

Sequence 3 is bounded at the base by disconformity H and at the top by disconformity I. It includes the lower part of the Paoli Formation, the basal disconformity-bounded unit in the Girkin Limestone on the Eastern Shelf and the Shetlerville member of the Renault Formation and the Yankeetown Formation in the central and western parts of the basin. The sequence is 6 meters thick on the Western Shelf, thickens to 20 meters in the Basin Interior and thins to 4 meters near the Cincinnati Arch and 2 meters in the northeastern sections.

Disconformity H is a major disconformity (type 2) characterized by accumulations of breccia and caliche up to 4 meters thick and called the Bryantsville Breccia Bed in Indiana and Kentucky (Mallot, 1952; Liebold, 1982). Disconformity H coincides with the top of the *Platycrinites penicillus* zone and the base of the *Talarocrinus* zone (Swann, 1963). The TST of sequence 3 is composed of parasequences 16 and 17. Parasequence 16 is a thin parasequence, confined to the most rapidly subsiding parts of the Basin Interior, which onlaps the shelves and a local tectonic high in the Basin Interior. Parasequence 17 has a patchy grainstone unit at the base which pinches out onto tectonic highs, and is capped a laterally extensive muddy carbonate/fossiliferous shale unit than can be correlated across most of the Basin Interior and Eastern Shelf. On the Western Shelf, the TST is absent except at section 3 where parasequence 16 is composed of quartz sandstone and parasequence 17 is composed of a fossiliferous shale capped by quartz sandstone. The Yankeetown Shale which occurs near the top of the sequence grades updip into quartz sandstone to the northwest (Swann, 1963).

The MFS is picked at the base of parasequence 18 because it is the first laterally extensive grainstone unit to be deposited on the eastern and western shelves. The HST consists of parasequences 18 to 21 in the Basin Interior but only parasequences 18 and 19 were deposited on the shelves. On the Eastern Shelf, parasequence 18 is a grainstone-muddy carbonate parasequence with the exception of a few downdip sections, where it is composed of muddy carbonates and green fossiliferous shale. Parasequence 19 is composed of multi-lateral grainstone deposits that underlie the sequence boundary on the Eastern Shelf. In the Basin Interior, parasequences 18 and 19 have skeletal grainstone/packstone bases with fossiliferous shale caps that grade laterally to muddy carbonates in the eastern Basin Interior. Ooid grainstone is rare in the Basin Interior, but does occur at the base of the HST near tectonic highs. Parasequence 19 is capped by a local disconformity at sections 10, 11 and 12. On the Western Shelf, parasequence 18 is composed of skeletal grainstone overlain by tidally-influenced siliciclastics and capped by a paleosol and parasequence 19 is composed of quartz siltstone. Late HST parasequences 20 and 21 progradationally offlap and are only present in the Basin Interior. Sequence 3 is capped by disconformity I.

Sequence 4

Sequence 4 is bounded at the base by disconformity I and at the top by disconformity J. It includes the upper part of the Paoli Formation, a second disconformity bounded unit in the Girkin Limestone and the Downeys Bluff Formation in the Basin Interior and Western Shelf. Sequence 4 is 6 meters thick on the Western Shelf, thickens to 20 meters in the eastern Basin Interior and thins to 7 meters near the Cincinnati Arch and to less than 3 meters on the Northeastern Shelf. The sequence may have been thicker in the Basin Interior but was eroded during pre-Bethel incision.

Disconformity I is an intermediate type 2 unconformity that overlies a thin eolianite in the Basin Interior (Fig. 9). The TST is composed of parasequence 22 which consists of a thin basal shale in the eastern Basin Interior, overlain by a grainstone unit that is very patchy on the Eastern Shelf and a laterally extensive fossiliferous shale or muddy carbonate cap. The maximum flooding surface is picked at the base of the grainstone unit in parasequence 23 because it is the first laterally extensive grainstone unit to flood the shelves. The HST is composed of parasequences 23 to 25 in the Basin Interior, but only parasequence 23 was deposited on the shelves. Parasequence 23 is composed of a thick skeletal/oolitic grainstone unit overlain by a patchy muddy carbonate unit on the Eastern Shelf, ooid/skeletal grainstone capped by fossiliferous shale and tidally influenced siliciclastics on the Western Shelf and skeletal limestone capped by fossiliferous shale in the Basin Interior. Parasequences 24 and 25 are only preserved basin-ward of the shelves suggesting that progradational offlap occurred at the top of sequence 4. These parasequences are composed of thick accumulations of ooid grainstone along the margin of the Eastern Shelf and skeletal limestone and shale in most of the Basin Interior. A widespread eolianite overlies the sequence on much of the Southeastern Shelf. The sequence is capped by a type 1 unconformity (disconformity J) marked by erosion of the underlying strata throughout much of the Basin Interior and an incised valley up to 75 meters deep that trends northeast-southwest from south-central Indiana to western Kentucky (Friberg et al., 1969). Where there is no incision on the eastern and western shelves, the sequence is bounded by a well-developed paleosol.

Composite-Sequences and Sequence-pairs

A composite-sequence is a succession of genetically related sequences in which the individual sequences stack into lowstand, transgressive and highstand sequence sets (Mitchum and Van Wagoner, 1991). Ross and Ross (1988) included all of the Ste. Genevieve, Paoli, and overlying Bethel to Ridenhower/Reelsville interval in one large scale sequence analogous to a 3rd-order composite-sequence (Fig. 13). Because the Illinois Basin was situated far updip of the shelf edge, it is not possible to trace sequence boundaries to determine which extends the furthest downdip. Thus, composite-sequence boundaries are distinguished from higher frequency sequence boundaries by the depth of incision and/or degree of paleosol development (Fig. 9). Using these criteria, two orders of lower frequency sequences occur in the Ste. Genevieve through Paoli interval which are also present in the overlying Bethel to Glen Dean interval (sequences 5 to 8 - Chapter 3). Sequences 1 through 8 are bundled into sequence-pairs and composite- sequences which are made up of 4 sequences each (Fig. 10). Sequences 1 through 4 are bundled into a composite-sequence (Fig. 11) that is composed of two sequence-pairs.

Composite-Sequences.- The basal unconformity for the Ste. Genevieve to Paoli composite-sequence occurs at the St. Louis-Ste. Genevieve boundary (disconformity A) and coincides with a boundary picked by Ross and Ross (1988). The transgressive sequence tract is composed of sequences 1, 2 and the TST of sequence 3 (Fig. 11). The transgressive sequence tract is dominated by high-energy ooid/skeletal grainstone and quartz sandstone. Major onlap occurs in the basal sequence which is composed of high-energy facies. The maximum flooding surface for the composite-sequence is picked to coincide with the MFS for sequence 3, because it marks the first time grainstone was

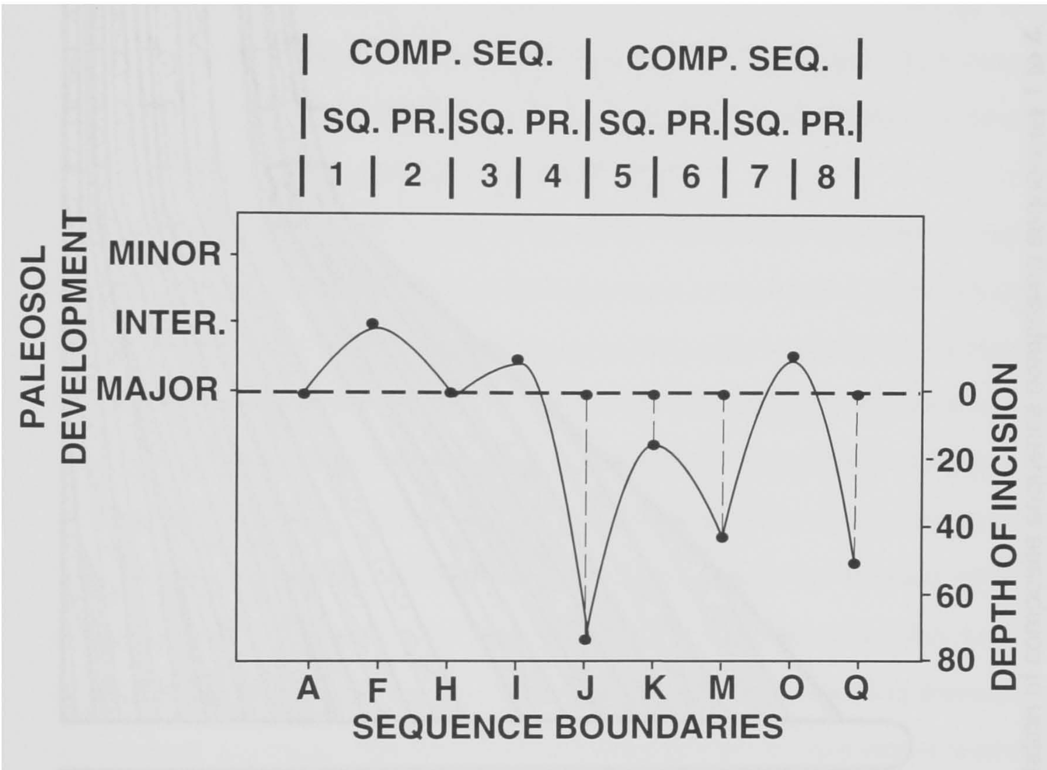


Fig. 10. Relationship between sequences, sequence pairs and composite sequences for Sequences 1 through 8. Sequences 1 through 4 are discussed in this paper, and sequences 5 through 8 are discussed in Chapter 3. Sequence pairs are bounded by major disconformities. Composite sequences are bounded by major disconformities that coincide with major biostratigraphic zone boundaries and show maximum incision.

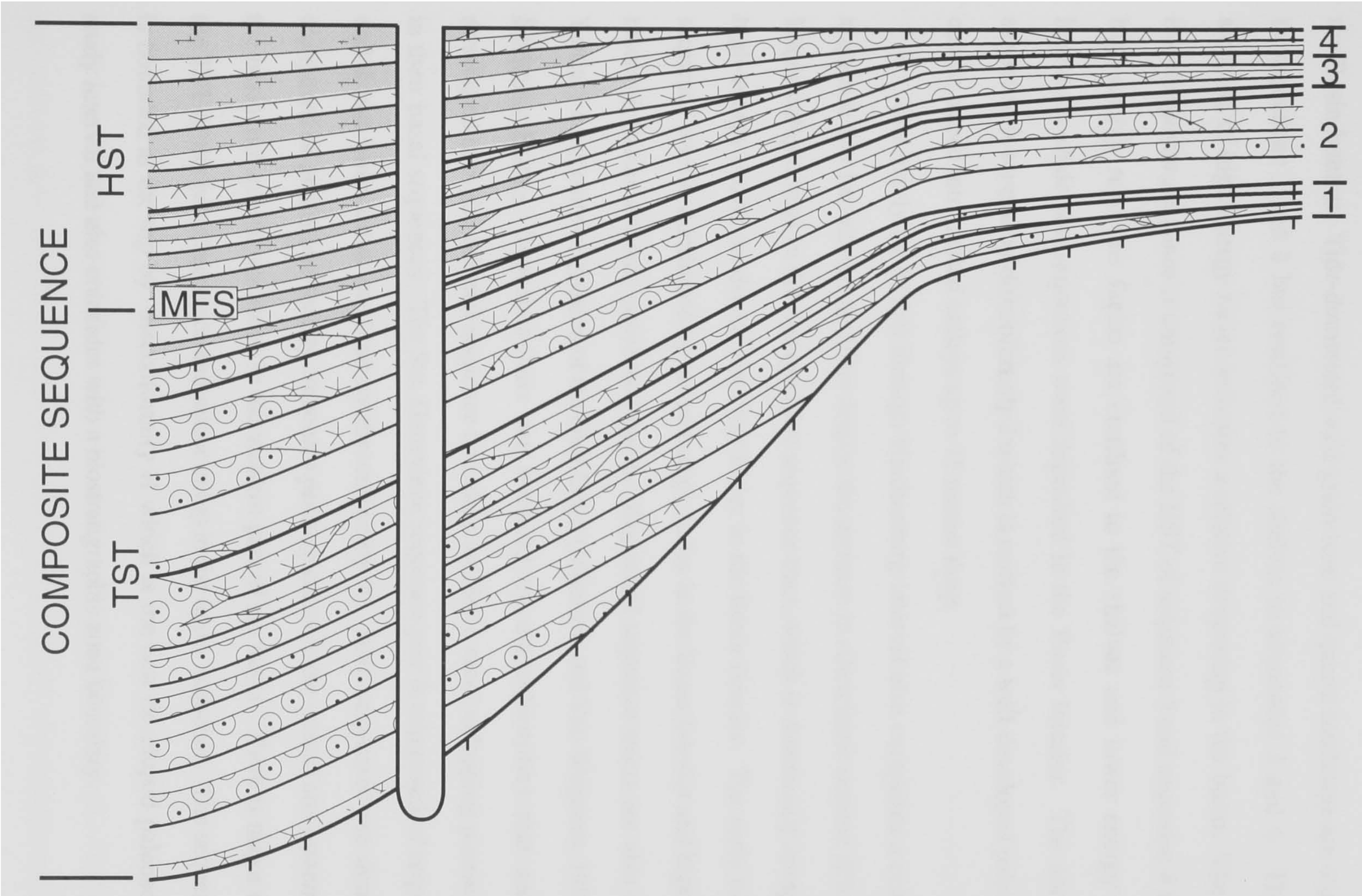


Fig. 11. Schematic representation of composite sequence comprising sequences 1 to 4. See Fig 4. for legend.

deposited on both shelves, despite high differential subsidence and an overall deepening the Basin Interior. Tide-dominated ooid grainstone and quartz sandstone are widespread in sequences 1 and 2 but confined to the shelves in sequences 3 and 4. This back-stepping of high-energy facies indicates a relative deepening in the basin. The HST of the composite-sequence is composed of the HST of sequence 3 and sequence 4 in which high-energy reservoir facies are confined to the shelves and lower energy skeletal limestone-shale parasequences were deposited in the Basin Interior. The composite-sequence is capped by disconformity J which is marked by a well-developed paleosol that can be traced into incised valleys up to 75 meters deep.

The overlying Bethel through Hardinsburg interval also comprises a composite-sequence that has similar qualities despite the increase in siliciclastic content (Chapter 3). Major onlap occurs in the transgressive sequence tract, which is dominantly composed of high-energy quartz sandstone reservoir facies in the Basin Interior. The early highstand sequence tract is dominated by lower energy facies in the Basin Interior and high-energy reservoir facies are only preserved updip. Highstand sequence tracts are also shaly in Tertiary composite-sequences of the Gulf Coast (Mitchum and Van Wagoner, 1991).

Sequence-Pairs.- Sequence-pairs are bounded by disconformities that are better developed than the internal sequence boundary and have more onlapping parasequences in their basal sequences. The Ste. Genevieve sequence-pair is composed of sequences 1 and 2 and bounded at the base by disconformity A. Both sequences are dominantly composed of grainstone-muddy carbonate parasequences in the central and eastern part of the basin and quartz sandstone in the western part of the basin. There is major onlap in the TST of the basal sequence and minor onlap in the upper sequence. The sequence-pair is bounded at the top by disconformity H which is the best developed paleosol in the study interval and also coincides with a biostratigraphic zone boundary.

The Paoli sequence-pair is bounded at the base by disconformity H and is composed of higher frequency sequences 3 and 4. These sequences are dominantly composed of limestone shale parasequences. There is more onlap in the TST of the basal sequence and both sequences have offlapping HST parasequences. The Paoli composite-sequence is bounded at the top by disconformity J which is a type 1 sequence boundary.

DISCUSSION

Duration of Sequences and Parasequences

The most recent ages for the duration of the Mississippian Period come from SHRIMP dating of volcanic rocks interbedded with marine sediments in Australia (Roberts et al., 1995). Brachiopod-biostratigraphy was used to correlate with European and North American conodont- and foramaniferal- zones which tied the SHRIMP ages to radiometric dates from Europe (Roberts et al., 1995). Based on foramaniferal- and conodont-zones, the Ste. Genevieve to Glen Dean interval is uppermost Visean (middle of V3b to V3c) (Baxter and Brenckle, 1982). This interval has a duration of roughly 3 m.y. (Roberts et al., 1995), but is poorly constrained due to the difficulties inherent in correlating between brachiopod-, conodont-, and foramaniferal-zones on different continents and the error bars on the radiometric dates.

There are four disconformity-bounded sequences in the Ste. Genevieve to Paoli interval and four in the overlying Bethel to Glen Dean interval (Chapter 3) which make eight sequences. Dividing eight sequences into 3 million years gives an average period of 375 k.y. per sequence which is suggestive of the long-term Milankovitch eccentricity signal (~414 k.y.) (Berger, 1988) and comparable to the calculated duration for some of the Pennsylvanian cyclothems (Heckel, 1986). Such 4th-order (0.1 to 0.5 m.y.) sequences are compatible with times of abundant global ice (Weber et al., 1995). If the approximate 400 k.y. duration for the sequences is correct, the sequence-pairs then formed in ~800 k.y. and the composite-sequences in ~1.6 m.y., both being in the range of 3rd-order sequences (0.5 to 5 m.y.) (Weber et al., 1995). Calculated average duration of the preserved parasequences would be about 40 k.y. in sequence 1, 70 k.y. in sequences 2 and 3 and 100 k.y. in sequence 4. This suggests that the preserved parasequences are dominantly 5th-order (10 to 100 k.y.). .

Climatic Influence

Increasingly humid wet-dry seasonality prevailed throughout the Ste. Genevieve to Paoli interval as indicated by the presence of caliche, breccia and red, slickensided mudrock paleosols. Evaporites, which are common in the underlying St. Louis Formation, are absent in the Ste. Genevieve to Paoli interval, which suggests that the climate became more humid in the Early Chesterian. The duration and amount of rainfall in the wet season must have been relatively small based on the abundance of ooids, dolomite and eolianites, all of which are most common in areas with arid and semi-arid climates (Witzke, 1990; Ettensohn et al., 1988; Ambers and Petzold, 1992). The overall upward increase in siliciclastic sedimentation up through the study interval is related to the source area becoming more humid with time. Increased influx of shale into the basin in sequences 3 and 4 coincides with a decrease in ooid production in the basin, possibly related to increased fresh-water influx from rivers to the northwest. This would have prevented marine waters from becoming super-saturated with respect to calcite/aragonite. However, the abundance of ooids, caliche and breccia on the Eastern Shelf in sequences 3 and 4 suggest that the Eastern Shelf remained a relatively dry climate. The trend of an upward, increasingly humid wet-dry seasonality continued into the overlying sequences 5 to 9 which have more siliciclastics as well as thin coal beds (Chapter 3).

Swann (1964) suggested that climate change in the source area was the primary control on carbonate-siliciclastic rhythmic sedimentation. He envisaged that arid climate led to increased siliciclastic influx and humid climate caused a decrease in siliciclastic influx as vegetation prevented erosion. However, disconformities that bound mixed carbonate-clastic sequences in the western part of the basin can be traced to the eastern part of the basin, where they bound sequences entirely composed of carbonates. This suggests that climate and variations in sediment supply did not control sequence development, because progradation of siliciclastics on the western side of the basin would not have caused exposure on the eastern side of the basin. Similarly, the limestone-shale parasequences in sequences 3 and 4 can also be traced laterally into carbonate parasequences suggesting that climate was not the primary control on parasequence development either.

Ramp-Slopes

Ramp-slopes may be estimated by dividing the shelf to basin thickness difference of an individual sequence by the number of parasequences in the sequence. This gives average slopes of 2.1 cm/km for sequence 1, 1.2 cm/km for sequence 2, 1 cm/km for sequence 3 and 1.5 cm/km for sequence 4. The maximum slope in the basin occurred along the margins of the shelves, where the gradient increased to as much as 5 cm/km based on thickness difference between sections on either side of the slope-break. Even at 5 cm/km, the slope was very gentle (0.003°), which would account for the similarity of facies deposited in the central and eastern parts of the basin in the Ste. Genevieve Formation. Although facies on the Eastern Shelf differ from the Basin Interior in sequences 3 and 4, this is primarily due to influx of siliciclastics from the west rather than an increase in the ramp-slope. The ramp-slope was sufficient at times, however, to allow onlap and progradational offlap of parasequences.

Tidal Energy

Strong tidal currents influenced sedimentation periodically in the Illinois Basin during the Chesterian, and formed: 1) bi-modal cross-bedding in ooid and skeletal grainstone (Carr, 1973); 2) elongate ooid shoals oriented parallel to inferred tidal currents

(Fig. 7) (Carr, 1973; Zuppann and Keith, 1988; Zuppann, 1993); 3) tidal channels filled with ooid grainstone and intraclasts; and 4) tidal sand-ridges, tidal channels, tidal rhythmites and flaser, wavy and lenticular bedding in the time-equivalent Aux Vases Formation in the western part of the basin (Cole, 1990; Huff, 1993).

Ooid- and quartz sand-ridge deposition occurred over a widespread area in the grainstone-dominated intervals of the parasequences in the Aux Vases and Ste. Genevieve. Similar Holocene ooid-ridges are found at Schooner Cays and the Tongue of the Ocean in the Bahamas, where the tidal range is 1 meter and ooid production and deposition occurs in a 10 to 20 km wide belt along the platform margin (Ball, 1967). If ooid deposition occurred in a similarly narrow belt in the Illinois Basin, deeper-water facies should have been deposited downdip while ooids were deposited updip; such deeper-water facies are not evident. Ooid grainstone distribution within parasequences indicates that Late Mississippian tidal ooid-ridges must have formed in a belt several times wider than those in the Bahamas. However, it is possible that ooid-ridges may be diachronous regressive sand bodies deposited in a narrower belt while strong tidal currents and wave sweeping caused non-deposition downdip until shoals prograded. In either case, very strong tidal currents would have been required to either make ooid-ridges over such an extensive area or to prevent open-marine deposition downdip of the ooid shoals.

Strong tidal influence on the gently sloping ramp caused much lateral heterogeneity of facies in which reservoir facies thicken, thin and pinch out into skeletal grainstone and muddy carbonates and reappear along the same horizon. Facies differentiation was due to local tidal energy rather than water depth or regional position on the ramp. If the ramp slopes had been steeper, or if the ramp were wave-dominated, strike-parallel facies belts would have formed. The dip-parallel orientation of the tidal sand-ridges focused and increased the velocity of the tidal currents which may have allowed synchronous ooid formation over a large area. The lack of a continuous barrier allowed open marine organisms such as echinoderms to live landward of, and between ooid shoals. During times of maximum tidal influence, the central and eastern parts of the basin were a mosaic of ooid-ridges and skeletal grainstone sheets which passed into quartz sandstone ridges and tidally influenced sandstone and shale on the western side of the basin.

The energy necessary to form ooid and quartz sand-ridges, tidal rhythmites and flaser-, wavy- and lenticular bedding across much of the Illinois Basin probably would probably have required at least meso-tidal conditions (2-4 meter range) (Kvale, 1996). Modern tidal rhythmites have only been found in meso- and macrotidal settings such as the Bay of Fundy (Kvale, 1996). However, the lack of updip carbonate tidal flat successions in the study interval, analogous to modern macrotidal siliciclastic settings argues against macrotidal conditions. The shallow water depth on the gentle ramp slope and the focusing of flow between ooid-ridges and sand-ridges increased the velocity of the tidal currents in the area of tidal sand-ridge deposition. The currents diminished onto the inner platform updip where there was less topography and facies showing little tidal-influence were deposited.

Tectonics

The Basin Interior generally subsided faster than the Eastern and Western Shelves, but relative subsidence rates between the shelves and the Basin Interior were not constant. This is indicated by dividing the average thickness of a given time slice in the Basin Interior by the average thickness on the Eastern Shelf near the shelf edge. Larger values indicate greater differential subsidence between the Basin Interior and the Eastern Shelf. The calculated basin/shelf ratios in sequence 1 are 27 in the TST and 3 in the HST; in sequence 2 they are 1.5 in the TST and 2.3 in the HST; in sequence 3 they are 9.5 in the TST and 7 in the HST; and in sequence 4 they are 3 in the TST and 4.5 in the HST

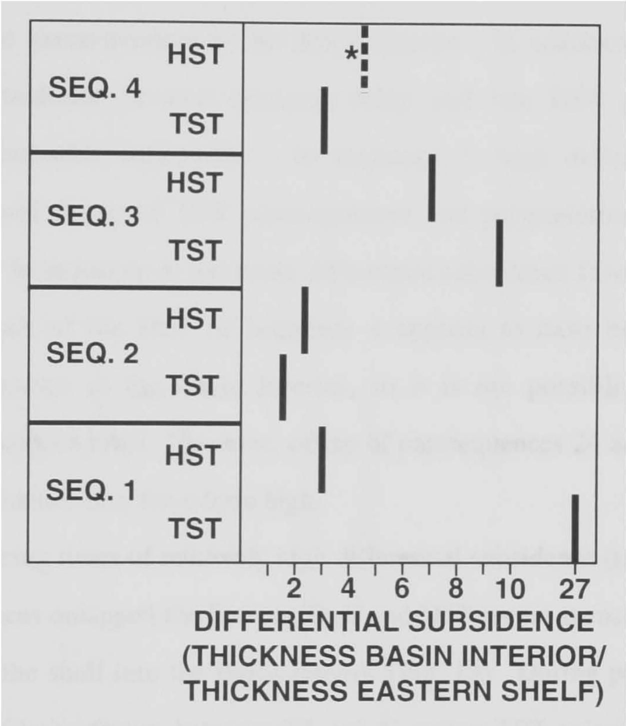


Fig.12. Plot of differential subsidence through time. Differential subsidence ratio is calcualted by dividing the average thickness ofeach timeslice in the basin interiorby the average thickness on the margin of the eastern shelf. *The HST of sequence 4 is substantially eroded in the basin interior, and it is therefore not possible to determine the thickness ratio, but the presence of progradationally offlapping parasequences suggests that the basin was undergoing high differential subsidence.

(Fig. 12). If compaction is ignored, which would make the disparity greater in sequences 3 and 4 due to the amount of shale in the Basin Interior, thickness ratios are proxies for differential subsidence ratios.

High differential subsidence during deposition of the TST of Sequence 1 caused four TST parasequences to onlap the Eastern Shelf. During deposition of HST parasequences, differential subsidence between the basin and shelf had decreased to 3, and, as a result, these parasequences are disconformity-bounded on the Eastern Shelf and are conformable parasequences in the Basin Interior. In sequence 2, relatively low differential subsidence favored minimal onlap and late HST parasequences are disconformity-bounded basin-wide. In sequence 3, high differential subsidence throughout caused onlap of TST parasequences and progradational offlap of HST parasequences. In sequence 4, moderate differential subsidence favored minor onlap in the TST. Much of the HST of sequence 4 appears to have been eroded during subsequent exposure in the Basin Interior, so it is not possible to determine the differential subsidence ratio. However, offlap of parasequences 24 and 25 suggests that differential subsidence may have been high.

Thus, during times of relatively high differential subsidence (ratios of 5 or more), TST parasequences onlapped the Eastern Shelf and HST parasequences progradationally offlapped from the shelf into the Basin Interior (Fig. 13). During periods of moderate differential subsidence (ratios between 2.5 and 5), minor TST onlap occurred and HST parasequences are disconformity-bounded on the shelf and passed into conformable parasequences downdip (Fig. 13). During times of relatively low differential subsidence (less than 2.5) there was negligible onlap of TST parasequences and late HST parasequences were disconformity-bounded basin-wide (Fig. 13).

Temporal variations in subsidence in the Illinois Basin may have been related to phases of thrust-loading in the Appalachian Basin during onset of the Late Paleozoic

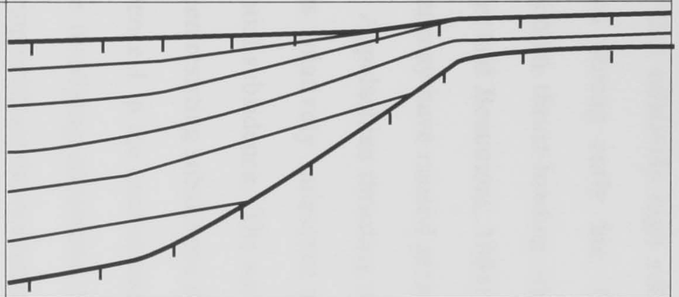
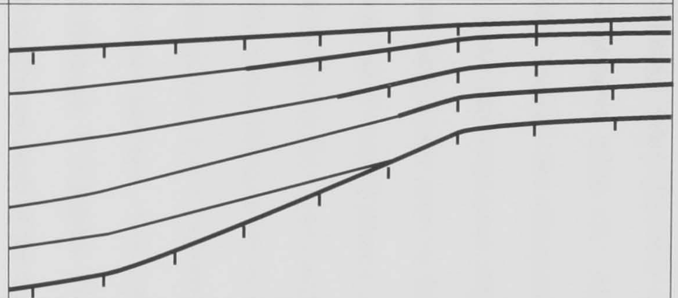
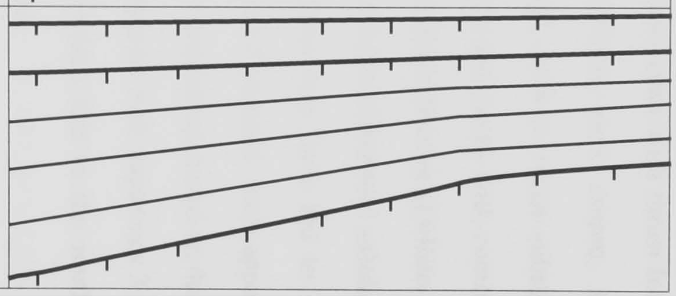
DIFFERENTIAL SUBSIDENCE	PARASEQUENCE DEVELOPMENT	OCCURRENCE	DSITRIBUTION OF PARASEQUENCES
HIGH	ONLAP OF TST PARASEQUENCES PROGRADATIONAL OFFLAP OF HST PARASEQUENCES	TST OF SEQUENCE1 ALL OF SEQUENCE 3 HST OF SEQUENCE 4	
MODERATE	MINOR ONLAP OF TST PARASEQUENCES HST PARASEQUENCES DISCONFORMITY-BOUNDED UPDIP ONLY	HST OF SEQUENCE 1 TST OF SEQUENCE 4	
LOW	NEGLIGIBLE TST ONLAP LATE HST PARASEQUENCES DISCONFORMITY-BOUNDED BASIN-WIDE	ALL OF SEQUENCE 2	

Fig. 13. Effects of variations in differential subsidence between the eastern shelf and the basin interior on parasequence distribution in the sequences.

Alleghenian Orogeny. Backstripping of time-equivalent units in southwest Virginia reveals that relatively high subsidence rates, possibly associated with thrust-loading, occurred during early Ste. Genevieve time (A. Al-Tawil, pers. comm, 1995). Appalachian thrust-loading likely caused the Cincinnati Arch to resist subsidence (Quinlan and Beaumont, 1984) while the horizontal stress associated with continental collision may have caused greater subsidence rates in the Basin Interior (DeRito et al., 1983). Appalachian thrusting may have caused phases of high differential subsidence, whereas relatively quiescent periods in the Appalachians may have led to lower differential subsidence. The northwest-southeast limb of the Cincinnati Arch appears to have been resisting subsidence during early Ste. Genevieve deposition, based on thinning of sequence 1 in the northeastern most sections. This contrasts with sequences 3 and 4 and the overlying sequences 5 through 7 (Chapter 3) which thin to the southeast, suggesting uplift on the northeast-southwest limb of the arch.

Phases of tectonic activity and quiescence appear to have occurred over periods that roughly coincide with the duration of the sequence-pairs (Fig. 12). Each sequence pair was deposited during a period of high differential subsidence at the base of which decreased to low or moderate differential subsidence at the top. High differential subsidence either during unconformity development or early in the deposition of the composite-sequences favored development of onlap relations across the sloping basin margin. Lower differential subsidence in the late highstand of sequence-pairs may have caused sequence boundaries to be better developed in the Basin Interior, but would not have caused better paleosol development updip. The only mechanism that could have caused long-term basin-wide exposure at the sequence boundaries is eustatic sea-level fall.

Eustasy and Sequence/Parasequence Development

Basin-wide paleosols and facies stacking patterns in the Mississippian succession can be used to construct a detailed relative sea-level curve (Fig. 14). Ross and Ross (1988) constructed an onlap curve for the Late Paleozoic but only correlated large-scale sequences (Fig. 14). They picked a larger-scale sequence boundary at the St. Louis-Ste. Genevieve contact and the Ste. Genevieve/Paoli interval only makes up part of one of their sequences. Swann (1963) constructed a transgression-regression, or relative sea-level, curve based on the seaward extent of siliciclastics within the Illinois Basin (Fig. 14). Swann's curve roughly parallels this paper's 4th-order sequence curve, because the siliciclastic units commonly prograded toward the tops of the sequences (Fig. 14). However, the Swann (1963) curve does not show a regression at the St. Louis-Ste. Genevieve boundary, because there was no siliciclastic influx associated with it.

Sequence Development.- Sequence-bounding disconformities formed during periods of high, moderate and low differential subsidence in the Illinois Basin. The disconformities can be physically traced using distinct marker beds and biostratigraphy from the Illinois Basin across the Cincinnati Arch to the Appalachian Basin (Smith et al., 1995). Disconformity H is coincident with a biostratigraphic zone boundary in both the Illinois and Appalachian Basins. Furthermore, the duration of the sequences roughly coincides with the Milankovitch long-term (~400 k.y.) eccentricity signal. All of this suggests that the sequences were produced by 4th-order (400 k.y.) glacio-eustatic sea-level changes rather than local tectonics. Disconformity-bounded sequences with similar characteristics occur in time-equivalent strata in Kansas (Abegg, 1994), Great Britain (Walkden, 1987), and Kazakstan (Lehmann et al., 1996). Comparison with sea-level curves for these intervals may confirm high-frequency global sea-level change during the Late Mississippian.

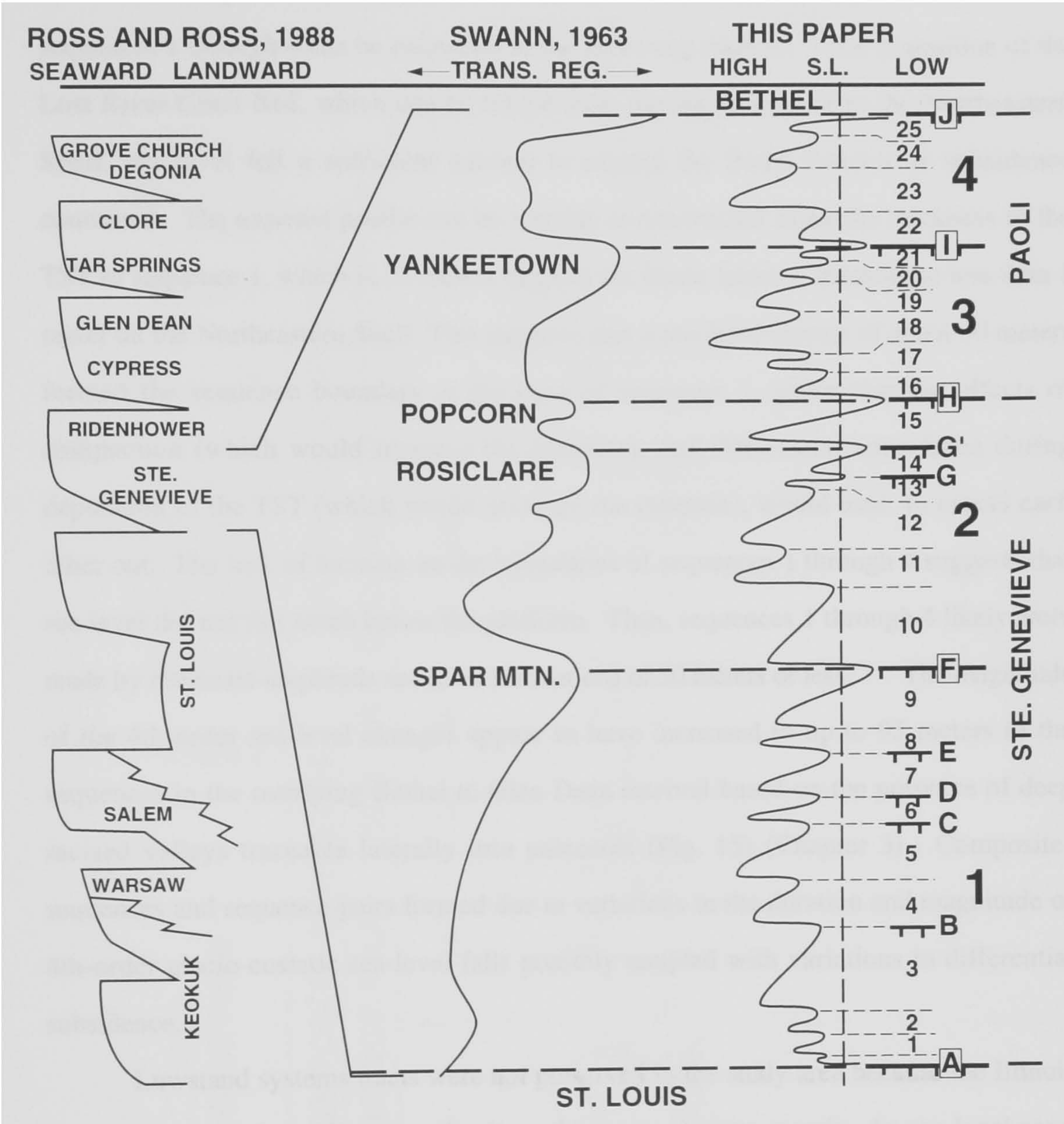


Fig. 14. Comparison of onlap curve from Ross and Ross (1988) (left), transgression-regression curve of Swann (1963 (center)), and relative sea-level curve prepared for this paper (right).

The magnitude of sea-level fluctuations that produced the type 2 boundaries of sequences 1 through 4 can be estimated in the following manner. After deposition of the Lost River Chert Bed, which can be traced from the Basin Interior to the Northeastern Shelf, sea level fell a sufficient amount to expose the Basin Interior as subsidence continued. The exposed profile can be roughly reconstructed using the thickness of the TST of sequence 1, which is 27 meters thick in the Basin Interior, thinning to less than 1 meter on the Northeastern Shelf. This suggests that a sea-level change of about 30 meters formed the sequence boundary at the base of sequence 1, given that the effects of compaction (which would increase the estimate), and differential subsidence during deposition of the TST (which would decrease the estimate), would tend to cancel each other out. The lack of incision on the boundaries of sequences 1 through 4 suggests that sea level did not fall much below the platform. Thus, sequences 1 through 4 likely were made by moderate-amplitude sea-level fluctuations of 30 meters or less. The magnitude of the 4th-order sea-level changes appear to have increased to up to 95 meters in the sequences in the overlying Bethel to Glen Dean interval based on the presence of deep incised valleys traceable laterally into paleosols (Fig. 15) (Chapter 3). Composite-sequences and sequence-pairs formed due to variations in the duration and magnitude of 4th-order glacio-eustatic sea-level falls possibly coupled with variations in differential subsidence.

Lowstand systems tracts were not preserved in the study area because the Illinois Basin was more than 300 km updip from the regional ramp-margin. As sea level rose, transgressive systems tracts (TST) composed of one to four parasequences onlapped the shelves. Initial widespread flooding caused the uppermost of these parasequences to extend onto the Eastern Shelf where they typically consist of laterally extensive muddy carbonate units. Maximum flooding generated extensive grainstone units over the Eastern Shelf, when water depths increased enough to accommodate the tidal wedge and

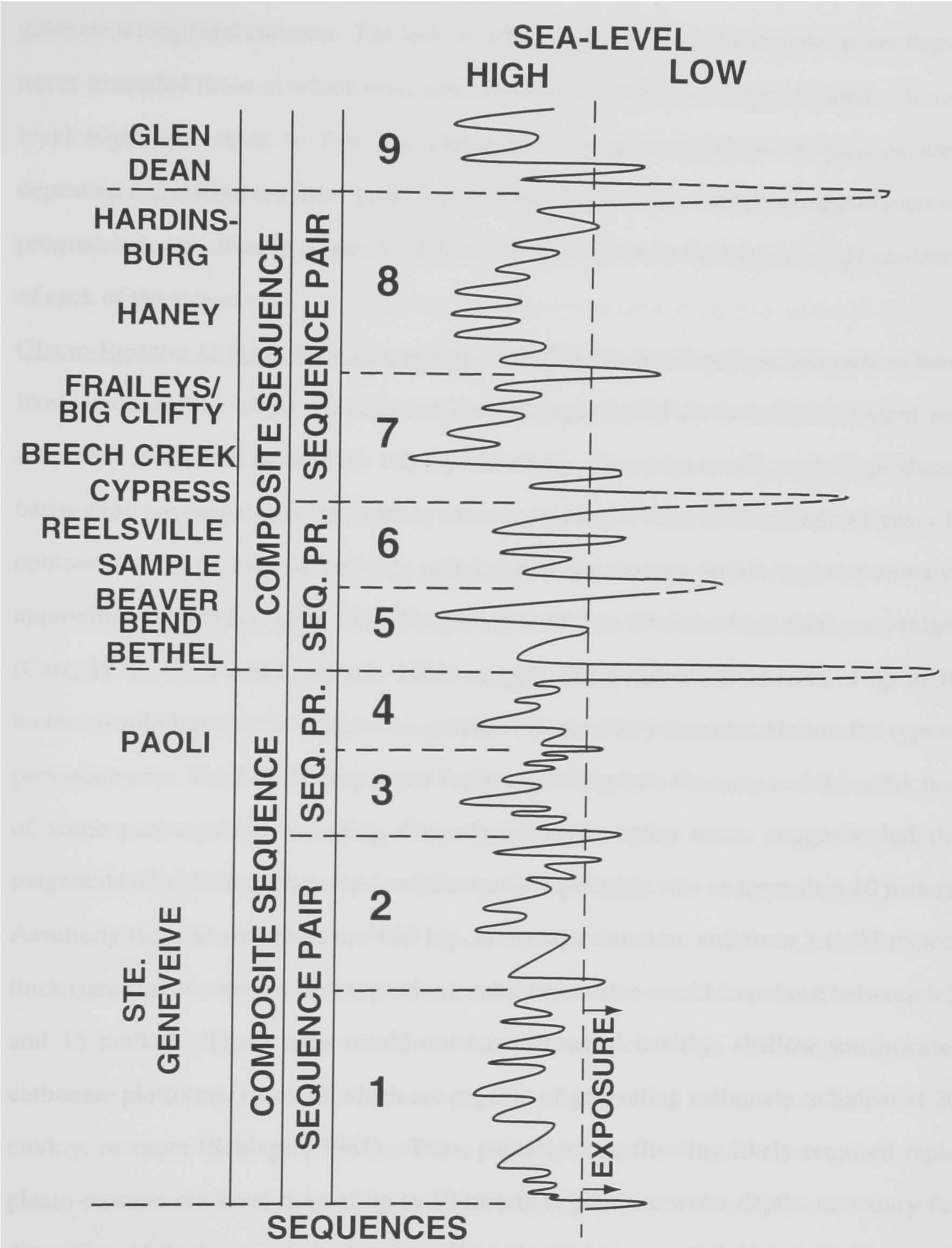


Fig. 15. Composite sea-level curve for Ste. Genevieveto Paoli (sequences 1 to 4 - this chapter) and Bethel to Glen Dean interval (sequences 5 to 9- chapter 2). Note suggestion of increasing-amplitude up section..

generate strong tidal currents. The lack of deeper water facies indicates that water depths never exceeded those at which ooid and skeletal grainstone lithotopes formed. At sea-level highstand, three to five aggradational to progradational parasequences were deposited as relative sea level peaked and began to fall. The degree of aggradation vs. progradation was directly related to differential subsidence in the basin during deposition of each of the sequences.

Glacio-Eustatic Origin of Parasequences.- Regionally developed parasequences were likely produced by glacio-eustatic sea-level changes based on their lateral extent and frequency (estimated to be 40 to 100 k.y. duration). Tectonics could not have produced basin-wide parasequences that appear to have formed in tens of thousands of years in contrast to the periods of tectonic activity and quiescence which had durations of approximately 800 k.y. (Fig. 12). The presence of 8 to 10 meter high tidal sand-ridges (Carr, 1973; Cluff and Lineback, 1981) suggest that rapid sea-level rises of up to 10 meters would have been required to provide the necessary accommodation for typical parasequences. The lack of deep-water facies extending onto the ramp and the restriction of some parasequence-bounding disconformities to updip areas, suggests that the magnitude of high-frequency sea-level fluctuations probably was no more than 10 meters. Assuming that the sequences are 400 k.y. in average duration, and from 2 to 63 meters thick (ignoring the effects of compaction), subsidence rates would have been between 0.5 and 15 cm/k.y. These rates would not have drowned healthy, shallow warm-water carbonate platforms, many of which are capable of generating carbonate sediment at 20 cm/k.y. or more (Schlager, 1981). Thus, parasequence flooding likely required rapid glacio-eustatic sea-level rises of up to 10 meters to generate water depths necessary for deposition of thick grainstone deposits. If sea-level rises occurred during the first 15 % of the parasequence duration (similar to Pleistocene sea-level rises), then rise rates would have been between 60 and 300 cm/k.y. for parasequences with durations of 20 to 100 k.y.

These rise rates would have been able to incipiently drown the ramp. Sea-level rises much less than 10 meters would have tended to generate meter-scale tidal flat cycles on the inner ramp similar to those described from greenhouse platforms (Goodwin and Anderson, 1985; Koerschner and Read, 1989; Goldhammer et al., 1990), which are notably absent on the Mississippian platform .

Parasequence Development.- Carbonate parasequence deposition was initiated by high-frequency sea-level rises that flooded disconformable surfaces and conformable surfaces formed on shallow-water mudstone that had shoaled to, or near to, sea level (Fig. 16 A). High-frequency (20 to 100 k.y.), rapid, eustatic sea-level rise coupled with background subsidence, generated water depths that were sufficient for the tidal wedge to develop and generate widespread high-energy tidal currents and wave-reworking. These high-energy conditions allowed deposition of skeletal- and ooid grainstone in discontinuous sheets and tidal ridges. Ooid and skeletal grainstone/packstone accumulated to form a broadly undulating submarine topography of ridges and passes. Muddy carbonates and shale were deposited on more shallowly flooded updip parts of the ramp, and eolianites were deposited at the shoreline. Tidal energy appears to have diminished into updip parts of the ramp where muddy carbonates were deposited in widespread lagoons. As ooid and skeletal grainstone filled part of the accommodation space on the ramp, tidal energy diminished and muddy lagoonal facies and fossiliferous shale prograded to form widespread sheets that capped and filled swales between carbonate sand-ridges. In areas with low accommodation rates, parasequences were capped by patchy eolianites and disconformities that pinched out into areas of higher accommodation where relative sea-level fall did not expose sediments for a period of time sufficient to develop disconformities. In these areas, renewed flooding generated conformable parasequences.

Shallowing-upward trends within the individual carbonate parasequences are complicated by the gentle dip of the ramp and the strong tidal currents, which generated

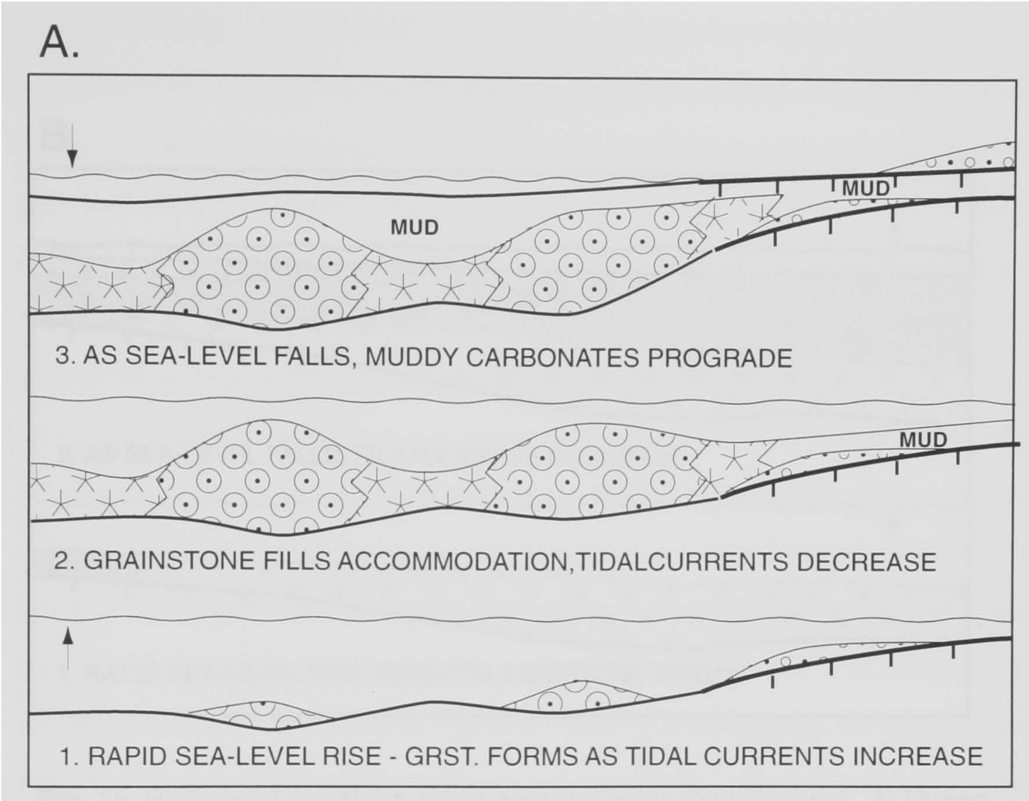


Fig. 16. A) Development of carbonate parasequences. 1) Rapid glacio-eustatic sea-level rise promotes tidal energy increase and deposition of ooid ridges and skeletal banks. 2) Ooid and skeletal grainstone fill much of the accommodation space, and tidal energy diminishes. 3) Muddy carbonates prograde, filling topography and forming laterally extensive sheets.

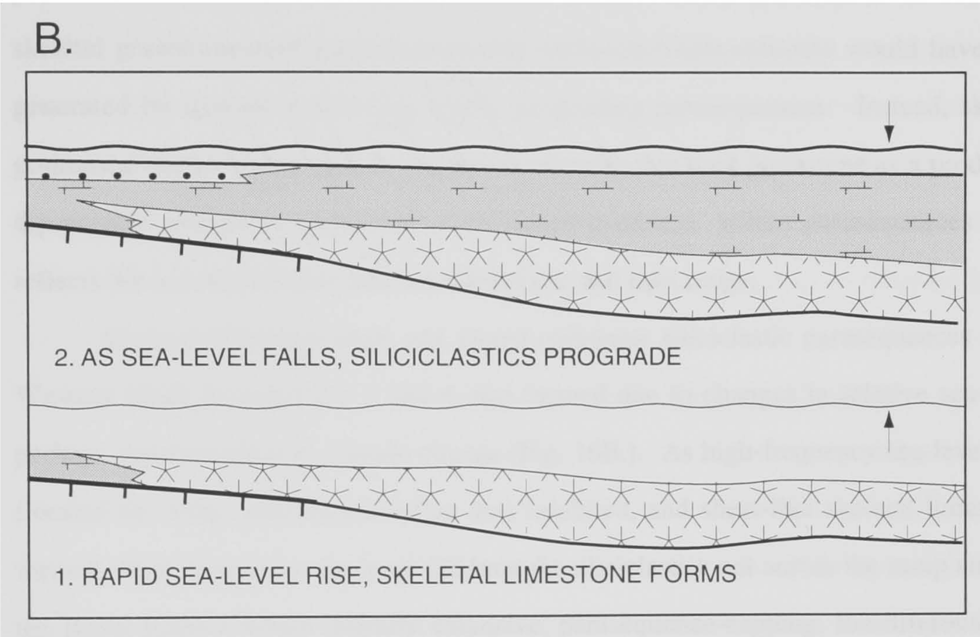


Fig. 16. B)Skeletal limestone-shale parasequence development. 1) During sea-level rise, siliciclastic source is shut down and carbonate-producing organisms thrive and make skeletal limestone. 2) As sea-level falls, shale progrades from west/northwest and caps parasequences. In the most updip sections, sandstone progrades over shale.

the dip-parallel orientation of the ooid shoals. Thus, facies were as much dependent on local tidal energy as water depth. If ooid shoals had formed wave-dominated strike-parallel barriers (rather than dip-parallel tidal ridges), a more predictable succession of skeletal grainstone-ooid grainstone-muddy carbonate/shale-eolianite would have been generated by upward shallowing within prograding parasequences. Instead, skeletal grainstone occurs above, below and passes laterally into ooid grainstone as a product of dip-parallel grainstone geometries. Grainstone thickness within parasequences likely reflects the accommodation due to sea-level rise and tidal range.

Skeletal limestone-shale and mixed carbonate siliciclastic parasequences of the Western Shelf in sequences 3 and 4 also formed due to changes in relative sea-level, perhaps supplemented by climate change (Fig. 16B.). As high-frequency sea-level rises flooded the ramp, siliciclastic influx was inhibited, and sheet-like skeletal limestones formed. High-frequency sea-level fall brought siliciclastics out across the ramp and into the Basin Interior where laterally extensive, parasequence-capping, fossiliferous shale was deposited in low-energy, shallow-water conditions. Fossiliferous shales pass laterally away from the siliciclastic source into muddy carbonate units which cap parasequences in the eastern part of the basin. Towards the siliciclastic source, sandstone and heterolithic facies capped parasequences on the western shelf.

Apparently symmetrical parasequences with muddy carbonates at the base and top (Pryor et al., 1990) can be traced downdip where the basal muddy units commonly form the tops of parasequences with basal, patchy, high-energy grainstone units. Thus the updip basal mudstones are actually the updip, low-energy tongues of downdip parasequences. At the other extreme, some well developed updip parasequences merge downdip into thick ooid grainstone or muddy units (e.g. parasequences 3, 4 and 5) which represent amalgamated parasequences. These developed where ooid grainstone and muddy carbonate deposition persisted through time, even though sea level was changing.

Other apparently amalgamated parasequences could be due to erosion of muddy carbonate caps during subsequent transgressions, leaving stacked grainstone units.

Autocyclic shallowing associated with local island development formed local parasequences within thicker, regionally traceable parasequences (Fig. 17). Such autocycles are clearly evident with the detailed, close-spaced sections but would not have been separable from regional parasequences on the basis of a single stratigraphic section. These autocycles appear to have shallowed-up into low energy facies, before the regional ramp shallowed up and underwent renewed local sea-level rise or an increase in tidal energy.

Stratigraphic Signature of the Greenhouse to Icehouse Transition

The Early Mississippian Period was dominated by precessionally-driven, low amplitude, high-frequency sea-level fluctuations, which suggests that it was relatively ice-free (Vaughn and Adams, 1984; Adams et al., 1990; Read and Horbury, 1993). The stratigraphy of the updip parts of large carbonate platforms that developed during greenhouse times of little global ice is dominated by meter-scale peritidal cycles capped by regional tidal flat facies, especially in highstand systems tracts of sequences (Goodwin and Anderson, 1985; Koerschner and Read, 1989; Goldhammer et al., 1990; Elrick and Read, 1991; Montanez and Read, 1993). These peritidal cycles may be bundled into obliquity or eccentricity cycle sets (Goldhammer et al., 1990). Disconformities are poorly developed on these greenhouse platforms except at 3rd-order sequence boundaries.

In contrast, the Ste. Genevieve to Paoli parasequences generally lack intertidal-supratidal caps, and disconformities are developed on subtidal lagoonal and shoal facies of 4th-order, and some updip 5th-order units. Similar disconformity-bounded sequences occur in time-equivalent strata in Great Britain (Walkden, 1987), which suggests that the

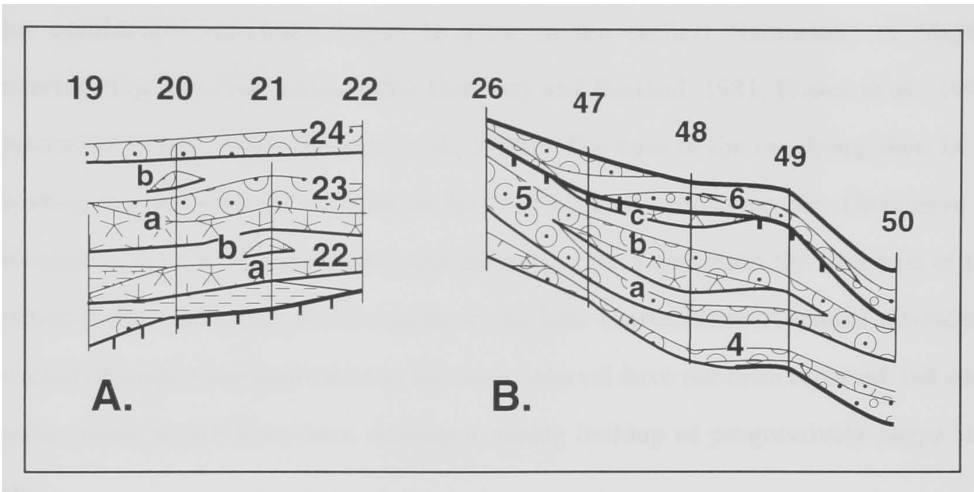


Fig. 17. Autocycles: location numbers are at the top, parasequences numbered as they are in the cross-sections. A) Autocycles 22b and 23b occur at single locations and likely formed due to local sediment buildup. B) Autocycles 5b and 5c developed over three different locations and likely represent local island shoaling.

lack of tidal flat facies and abundant disconformities are likely related to moderate amplitude 4th-order eustasy.

By the Pennsylvanian, the Gondwana glaciation was well advanced and high-amplitude sea-level fluctuations were likely driven by the 40 k.y. obliquity, and 100 and 400 k.y. eccentricity orbital variations (Heckel, 1986; Goldhammer et al., 1991). The first ice-rafted deposits of the Late Paleozoic glaciation of Gondwana suggest that the major continental ice-sheets began to grow in the earliest Namurian or Middle Chesterian (Fig. 18) (Garrasino, 1981; Hambrey and Harland, 1981; Frakes et al., 1992; Caputo and Crowell, 1985; Roberts et al., 1995). The base of the overlying Glen Dean Formation is equivalent to the base of the Namurian, therefore the Ste. Genevieve to Paoli interval is Late Visean (Baxter and Brenckle, 1982) and spans the later part of the transition from a relatively ice-free globe to the Late Carboniferous/Permian glaciation. Ice-rafted deposits time-equivalent to the study interval have not been reported, but such deposits likely would have been destroyed during buildup of progressively larger ice-sheets.

The waxing and waning of progressively larger ice-sheets may have caused the apparent increase in the amplitude of sea-level fluctuations forming the sequences (Fig. 13). Sequences 1 through 4 are bounded by type 2 boundaries with evidence for sea-level fluctuations of ~30 meters or less. In contrast, many of the younger sequences have type 1 boundaries and show evidence for up to 95 meter sea-level changes (Chapter 3, this volume). Time-equivalent strata in Great Britain show a similar upward change from type 2 to type 1 boundaries attributed to increasing amplitude of relative sea-level fluctuations coupled with tectonics (Walkden, 1987). This increasing amplitude is likely related to the onset of continental glaciation on Gondwana.

Global ice-house times such as the Pennsylvanian and Pleistocene are characterized by high-frequency 4th-order sequences (Heckel, 1986; Goldhammer et al.,

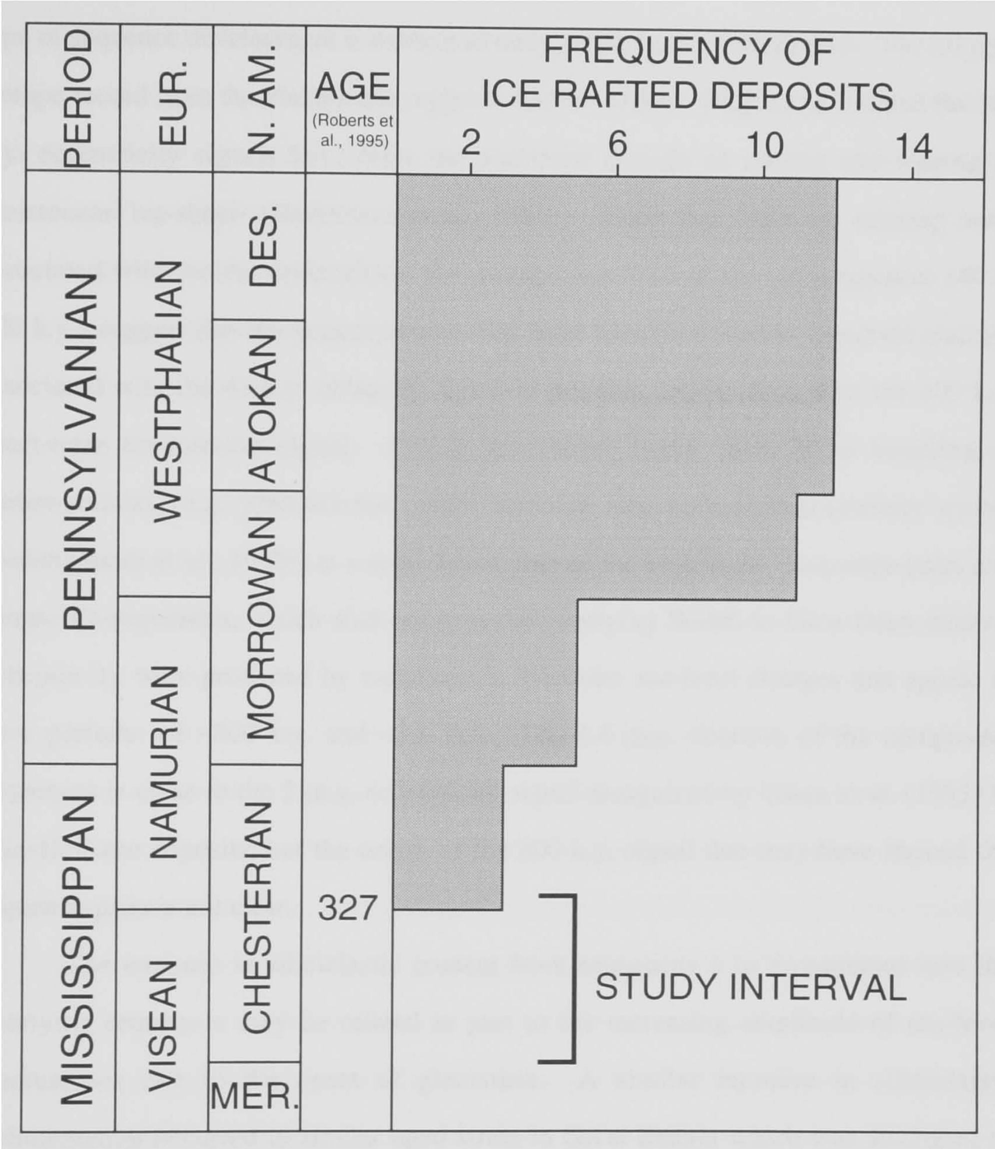


Fig. 18. Global climate setting during Late Mississippian and Pennsylvanian Periods. Graph shows the number of occurrences of ice-rafted deposits world-wide leading up to Late Paleozoic glaciation of Gondwana (modified from Frakes et al., 1992; Age from Roberts et al., 1995). The study interval occurs just below the first recorded glacial deposits.

1991; Weber et al., 1995; Mitchum and Van Wagoner, 1991) This paper shows that this type of sequence development is evident as early as the Late Mississippian. The Oxygen isotope record from the Pleistocene suggests that the 41 k.y. obliquity period and the 100 k.y. eccentricity signals have been the dominant periods of waxing and waning of Pleistocene ice-sheets (Ruddiman et al., 1986). Given that there are missing beats associated with the disconformities, the average durations of the parasequences (40 to 100 k.y.) suggest that the parasequences may have been produced by sea-level changes associated with the 41 k.y. obliquity signal in possible combination with the 100 k.y. short-term eccentricity signal. Unlike greenhouse times, there is no bundling of parasequences (e.g. precessional cycles bundled into obliquity/eccentricity cycles (Goldhammer et al., 1990)) at a scale below that of the sequences. Sequence-pairs and composite-sequences, which also occur in the overlying Bethel to Glen Dean interval (Chapter 3), were produced by variations in 4th-order sea-level changes that appear to have periods of ~800 k.y. and ~1.6 m.y. The 1.6 m.y. duration of the composite-sequences is close to the 2 m.y. eccentricity signal recognized by Olsen et al. (1993) in Triassic Lake deposits, but the origin of the 800 k.y. signal that may have formed the sequence-pairs is not clear.

The increase in siliciclastic content from sequences 1 to 4 continues into the overlying sequences may be related in part to the increasing amplitude of sea-level fluctuations due to the onset of glaciation. A similar increase in siliciclastic sedimentation occurred in similar aged strata in Great Britain which was attributed to increasing amplitude eustasy and local tectonics (Walkden, 1987). Increasing amplitude eustasy caused greater base-level falls that in turn caused greater erosion in the headlands and increased siliciclastic sediment supply. At the same time, synchronous climate change and increasingly humid wet-dry seasonality would have increased clastic influx during lowered sea level.

Controls on Reservoir Distribution

Reservoir facies in the Early Chesterian Ste. Genevieve/Aux Vases to Paoli interval are most commonly tidally-influenced ooid grainstone and quartz sandstone (Cluff and Lineback, 1981; Seyler and Cluff, 1990; Huff, 1993). These reservoir facies are widespread in the basin throughout the transgressive sequence tract in the composite-sequence and confined to the shelves in the highstand sequence tract (Fig. 11). Ooid grainstone reservoirs occur near the base of parasequences and were shaped by strong tidal currents during sea-level high stands. These tidal currents caused lateral compartmentalization of ooid grainstone and quartz sandstone reservoirs into tidal sand ridges and tidal channel-fills. Ooid grainstone reservoirs are laterally sealed by tight crinoidal skeletal grainstone and muddy carbonates and vertically sealed by muddy carbonate caps. Where present, caliche seals both ooid grainstone and quartz sandstone reservoirs.

Application of high-resolution sequence stratigraphy in individual fields will further improve understanding of reservoir distribution and compartmentalization. Paleosols and regional muddy carbonate units provide stratigraphic markers beyond the resolution of the formational stratigraphy which can be used to better correlate horizons of patchy reservoir facies. If a single core were drilled in each field to determine the elevation and signature of the paleosols on spectral-gamma ray logs, paleosols likely could be traced using spectral gamma-ray well-logs (Carr, 1995). These carbonate and mixed carbonate-siliciclastic sequences provide a well-exposed reservoir analog for other Late Mississippian strata, such as the giant Carboniferous oil-fields in Kazakstan.

CONCLUSIONS

1. Regional disconformities were previously thought to be of relatively local extent in the Ste. Genevieve to Paoli interval in the Late Mississippian (Chesterian) of the Illinois

Basin. However, these disconformities, which probably are the result of moderate-amplitude 4th-order eustasy, can be mapped regionally to define a high-resolution sequence stratigraphic framework of 4th-order sequences and component parasequences. The framework yields a much better understanding of the vertical and lateral distribution of reservoir facies and their seals.

2. Spatial and temporal variations in subsidence between the shelves and shallow "Basin Interior" controlled trends in thickness of facies and reservoir units. This differential subsidence had a strong influence on regional onlap/offlap patterns and parasequence stacking. However, glacio-eustasy was the dominant control on sequence and parasequence development.

3. Four disconformity-bounded 4th-order sequences (2 to 63 meters thick) can be traced across the Illinois Basin in the Ste. Genevieve to Paoli interval. The sequences generally are bounded by type 2 sequence boundaries marked by caliche, breccia, red mudrock paleosols and carbonate eolianites. However, a type 1 sequence boundary with up to 75 meters of incision caps sequence 4. Lowstand systems tracts are not preserved due to the updip position on the ramp. Transgressive systems tracts are relatively thin, and are composed of one to four onlapping parasequences. Maximum flooding surfaces are picked at the base regional grainstones which extend over the Eastern Shelf. Sequences are mainly composed of highstand systems tract, aggradational to progradational parasequences. The sequences correlate with those in the Appalachian Basin indicating a eustatic origin. The magnitude of sea-level changes was 30 meters or less for the type 2 sequence boundaries but increased to up to 95 meters for the type 1 boundary at the top of the interval. The increasing magnitude of sea-level fluctuations was probably driven by waxing and waning of progressively larger ice-sheets during the onset of Gondwana glaciation.

4. Parasequences contain the individual reservoirs, and consist of, from top to base: 4) a capping disconformity or conformable boundary; 3) muddy carbonate/fossiliferous shale caps; 2) ooid (reservoir) and skeletal grainstone; and 1) a basal disconformity or conformable boundary. These parasequences differ from typical carbonate cycles developed during global greenhouse conditions in their lack of tidal flat caps and common disconformities. Strong tidal currents formed laterally discontinuous dip-parallel ooid grainstone ridges and channel-fills (reservoirs) that pass laterally into skeletal grainstone sheets and banks. Grainstones were capped by muddy carbonates which were deposited in laterally extensive sheets and swale fills. This generated highly compartmentalized reservoirs. Without the sequence framework defined by regional tracing of disconformities, reservoirs appear to be randomly distributed. The sequence and parasequence framework more clearly show how individual reservoirs are related. The parasequences can be correlated between areas with vastly different subsidence rates, thus they are likely formed by 5th-order sea-level fluctuations, superimposed on the 4th-order eustatic signal.

5. Mississippian carbonate reservoir facies of this age world-wide are likely to show similar high-frequency sequence stratigraphy because of the global eustatic control during the time of transition into the Late Paleozoic glaciation.

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Chapter 3: High-Resolution Sequence Stratigraphy of Late Mississippian (Chesterian) Mixed Carbonate-Siliciclastic Bethel to Glen Dean Interval, Illinois Basin

ABSTRACT

Five mixed carbonate-siliciclastic 4th-order sequences (~400 k.y.) were deposited in the tectonically-active, tide-dominated Illinois Basin in the Late Mississippian (Chesterian) Bethel to Glen Dean interval. Sequence-bounding paleosol horizons and incised-valleys can be correlated basin-wide within a framework of biostratigraphy and distinctive marker beds on interpretive cross-sections constructed using detailed logs of 33 closely spaced outcrops and cores. Sequences have one or two transgressive-regressive parasequences in their transgressive systems tracts. Transgressive-regressive parasequences consist of basal incised-valley-filling quartz sandstone reservoir facies overlain by offshore skeletal limestone, which marks maximum flooding and shallows up into shale and heterolithic tidal flat facies. Highstand systems tracts have the upper half of transgressive-regressive parasequences at the base overlain by 1 to 6 regressive parasequences. Regressive parasequences consist of basal offshore skeletal limestone capped by laterally extensive shale and heterolithic facies.

The sequences are bundled into sequence-pairs and a composite-sequence. The transgressive sequence tract of the composite-sequence has sequences with abundant high-energy reservoir facies downdip bounded by type 1 unconformities. In the early highstand sequence tract, sequences are composed of high-energy reservoir on the shelves, low-energy shale and limestone dominate and are bounded by type 1 unconformities. The late highstand sequence tract consists of siliciclastic tidal flat facies that are capped by a type 1 unconformity that marks the top of the composite-sequence.

Moderate- to high-amplitude 4th-order glacio-eustatic sea-level changes produced the sequences, higher frequency 5th-order glacio-eustatic sea-level changes produced the parasequences and 3rd-order sea-level changes produced the composite sequences and sequence pairs. The onset of major continental glaciation on Gondwana caused the moderate to high-amplitude 4th-order eustasy and a shift toward more humid wet-dry seasonal climate in the Middle Chesterian. Spatial and temporal variations in differential subsidence between the Eastern Shelf and the more rapidly subsiding Basin Interior had a major impact on lithofacies distribution and onlap/offlap geometries in sequences and parasequences. Phases of tectonic activity and quiescence probably were caused by thrust-loading in the Appalachian Basin to the east.

INTRODUCTION

Late Mississippian (Chesterian) rhythmically interbedded carbonate and siliciclastic strata of the Illinois intracratonic basin provide an excellent opportunity to study the relative effects of eustasy, tectonics and sediment supply on vertical and lateral distribution and compartmentalization of reservoirs and the extent to which sequence stratigraphic concepts can be applied in such a setting. These units developed during Late Mississippian continental collision in the Appalachians to the east, which caused spatial and temporal variations in differential subsidence in the Illinois Basin. The Late Mississippian also was a time of transition between Early Mississippian greenhouse conditions and global glaciation in the Pennsylvanian/Permian when buildup of ice-sheets on Gondwana caused increasingly higher amplitude 4th-order eustatic sea-level and climate changes.

Previous work on the Middle Chesterian Bethel to Glen Dean interval in the Illinois Basin includes regional surface and subsurface mapping of limestone, sandstone and shale Formations by Butts (1917), Weller and Sutton (1940), Swann and Atherton

(1948), McFarlan et al. (1955), Swann (1963, 1964), Potter (1962), Kissling (1967), Specht (1985), Sable and Dever (1990) and Droste and Keller (1995?). Detailed sedimentologic studies have been done on parts of the study interval by Kissling (1967), Friberg et al. (1969), Visher (1980), Harris (1992), Ambers and Petzold (1992), Ambers and Robinson (1992), and Cole and Nelson (1995) among others. Swann (1963, 1964) suggested that interbedding of carbonate and siliciclastic formations was caused by periodic climate changes in the source area which influenced sediment supply under relatively static sea level. Paleosols have been considered to be local features formed due to island shoaling or autocyclic delta-switching (Droste and Carpenter, 1990; Ettensohn, 1975). Deep paleo-valleys filled with quartz sandstone cut through underlying strata were thought to represent submarine channels (Friberg et al., 1969), areas of continuous sandstone deposition that formed in facies with limestone (Laurin, 1988) or distributary channels of prograding river-dominated deltas (Swann, 1964). Treworgy (1985, 1988) traced a sequence bounding paleosol and invoked a eustatic model for stratigraphy the Golconda Formation. Ambers and Robinson (1992) documented channel-fills in the Sample Formation traceable laterally to paleosol horizons and interpreted them to be incised-valleys cut during sea-level lowstand.

In this paper, detailed sedimentologic studies and regional mapping of disconformities are used to define the basin-wide high-resolution sequence stratigraphy of the Bethel to Glen Dean interval. Regional cross-sections along more than 600 kilometers of the eastern, southern and western outcrop belts of the Illinois Basin were constructed using detailed logs of 33 outcrops and cores. Despite the lateral heterogeneity of individual units, several sequence-bounding paleosols, which commonly pass laterally into incised-valleys, can be correlated over much of the basin within a framework of biostratigraphic and lithologic markers (Smith and Read, 1995; Smith et al., 1995). Recognition of the through-going disconformities allows correlation at the parasequence scale. The cross-sections provide a data set with which the relative effects of eustasy, tectonics, sediment supply and tidal energy can be assessed. The data set also allows refinement of sequence stratigraphic concepts for mixed-carbonate clastic sequences deposited updip on a gently dipping ramp.

GEOLOGIC SETTING

Tectonic and Paleogeographic Setting

The Illinois Basin formed over a failed rift zone (Burke and Dewey, 1973) where incipient rifting began in the late Proterozoic (Bond et al., 1986) and normal faulting and extension occurred into the Middle and Late Cambrian (Kolata and Nelson, 1990). Three phases of subsidence occurred in the Illinois Basin during the Paleozoic: 1) exponentially decreasing thermal subsidence from the Late Cambrian to the Middle Ordovician; 2) slow linear subsidence from the Late Ordovician to the Early Mississippian; and 3) an enigmatic four-fold increase in the subsidence rate from the Middle Mississippian to the Late Pennsylvanian (Heidlauf et al., 1986). This Late Paleozoic subsidence event likely occurred as a response to continental collision and thrust-loading in the Appalachians to the east. The increase in horizontal stress caused by continental collision may have decreased the viscosity in the lower lithosphere beneath the Illinois Basin allowing the mass excess of mafic material emplaced in the crust during rifting to move more quickly toward isostatic equilibrium (DeRito et al., 1983; Kolata and Nelson, 1990). Downwarping in the basin was accompanied by uplift on the intervening Cincinnati Arch in response to loading in the Appalachian Foreland Basin (Quinlan and Beaumont, 1984).

The Illinois Basin was located between 5 and 15 degrees south of the equator during the Late Mississippian (Craig and Connor, 1979; Scotese and McKerrow, 1990) (Fig. 1A.). Throughout the Paleozoic, the basin was an embayment open to the present-day south, which covered present-day southern Illinois, southwestern Indiana and western

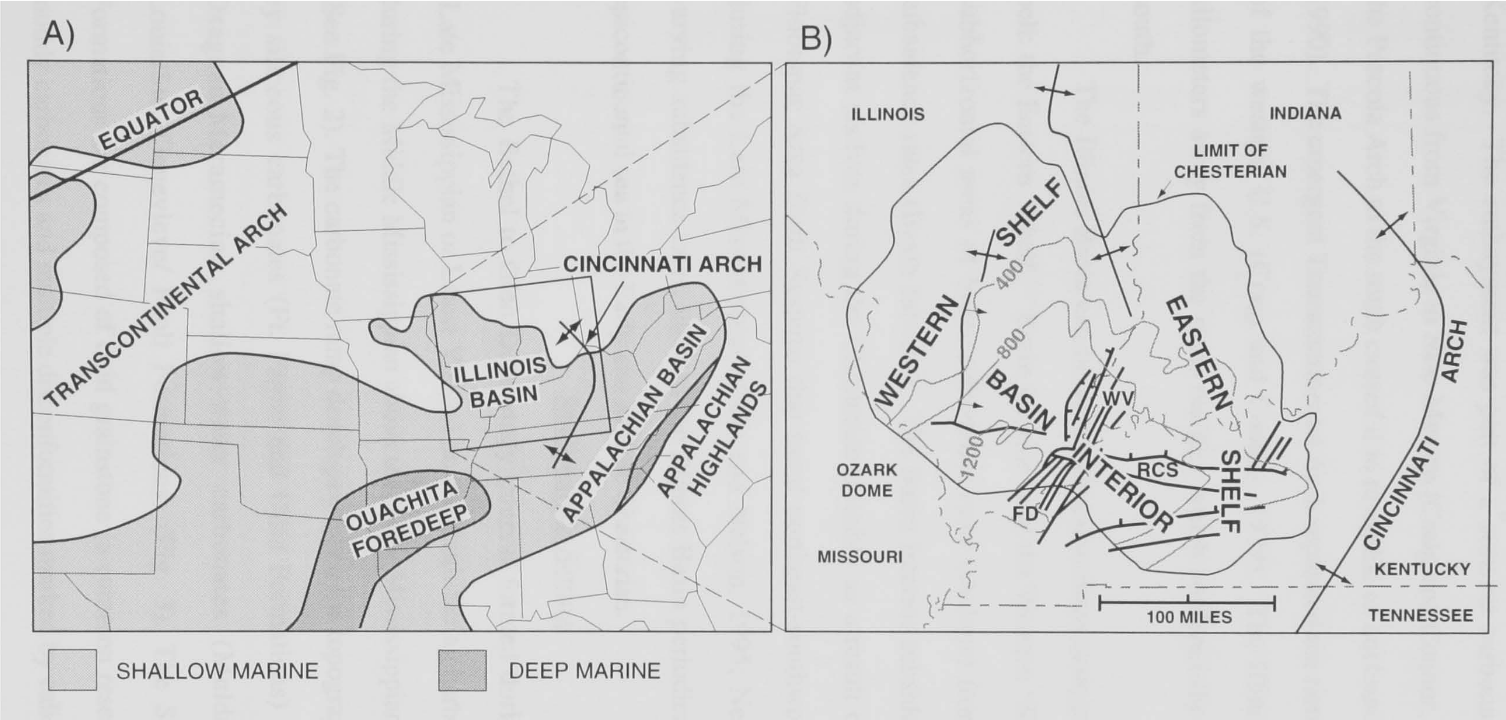


Fig. 1. A)Paleogeographic map showing regional setting of Illinois Basin (modified from Craig and Connor, 1979). B) Tectonic map of Illinois Basin.Isopachs in feet for Bethel to top of Mississippian (from Swann, 1963). 800 ft. isopach is approximate boundary of Basin Interior. WV = Wabash Valley Fault System; RCS = Rough Creek-Shawneetown Fault System; FD = Fluorspar District Fault Complex.

Kentucky. The embayment was part of a massive carbonate ramp that at times was continuous from Virginia to New Mexico (Craig and Connor, 1979). Mesozoic uplift of the Pascola Arch to the south caused it to resemble an enclosed basin (Kolata and Nelson, 1990). The emergent Transcontinental Arch separated the ramp from the Antler foredeep of the western U.S. (Craig and Connor, 1979). The Illinois Basin lay at least 300 kilometers away from the ramp margin, which bordered the Ouachita Foredeep to the south.

The Illinois Basin can be subdivided into three main provinces along the outcrop belt: the Eastern "Shelf," "Basin Interior" and the Western "Shelf" (Fig. 1B) delineating subhorizontal areas of lower subsidence rates (shelves) from internal areas of higher subsidence rates (Basin Interior). The Basin Interior subsided at a faster rate than the adjacent shelves during the Chesterian possibly as a result of the Wabash Valley and Fluorspar Area Fault Systems that trend northeast-southwest and likely were active during the Late Mississippian (Cole and Nelson, 1995; Nelson, 1996). Despite the varying subsidence rates, most of the Illinois Basin periodically lay beneath a shallow epicontinental sea in the Early and Middle Chesterian.

Stratigraphic Setting

The Bethel to Glen Dean study interval formed during the Middle Chesterian (Late Mississippian or Latest Visean) on the south-facing carbonate ramp that developed during the Middle Mississippian above the Early Mississippian siliciclastic Borden Delta (See Fig. 2). The carbonate ramp developed when the topography on the delta was filled by siliceous carbonates (Ft. Payne and Ullin Formations) which shallowed up into Osagean-Meramecian shallow-water carbonates (Muldraugh/Warsaw/Salem/St. Louis/Ste. Genevieve/ Paoli Formations) (Fig. 3). The Ste. Genevieve and Paoli Formations are composed of ooid grainstone (a common reservoir), skeletal grainstone, muddy carbonates and multiple disconformities marked by caliche, breccia and eolianites.

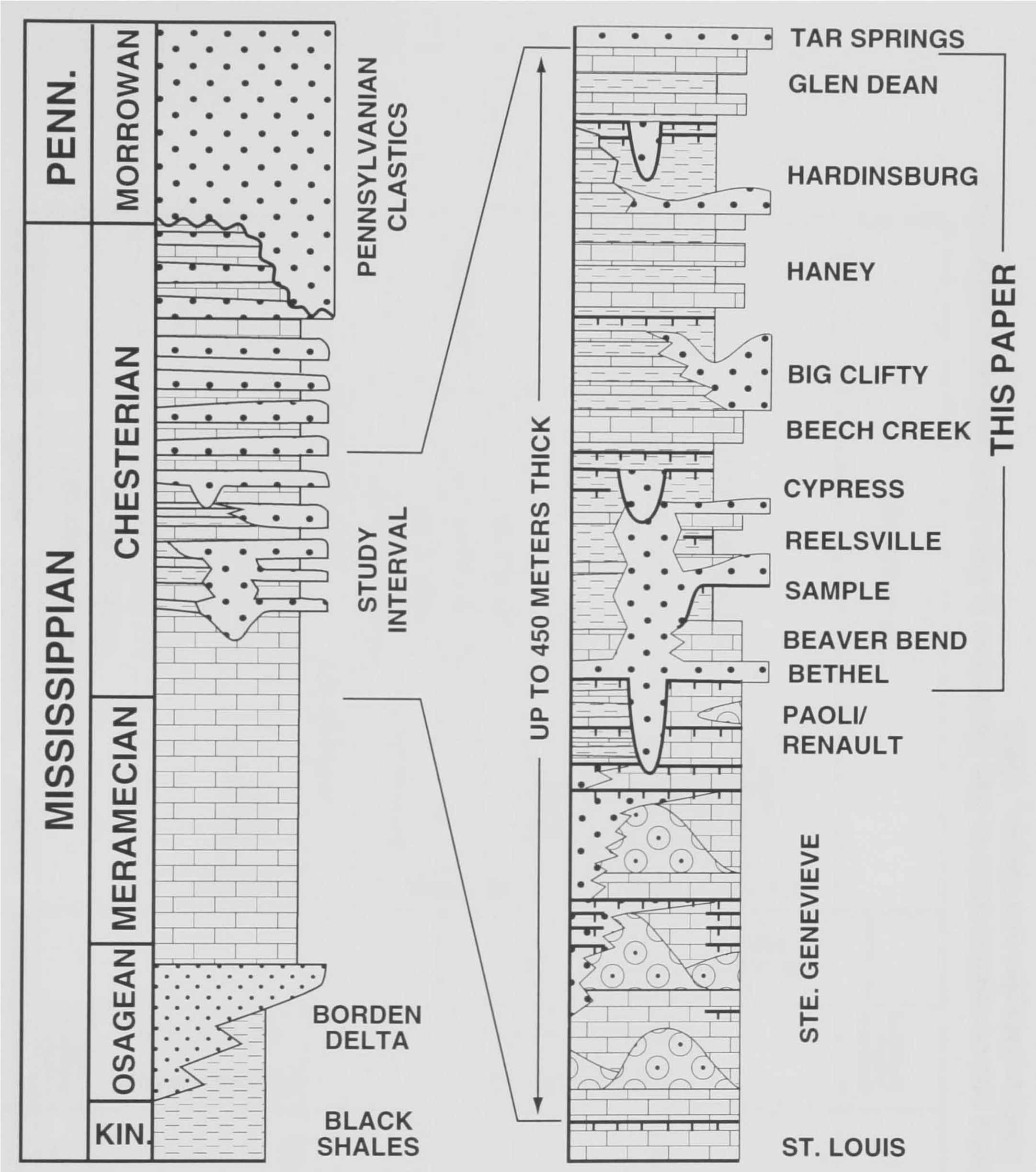


Fig. 2. Left-hand column is a generalized stratigraphic column for Mississippian of the Illinois Basin. Right-hand column shows more detailed stratigraphy of the Lower to MiddleChesterian units.

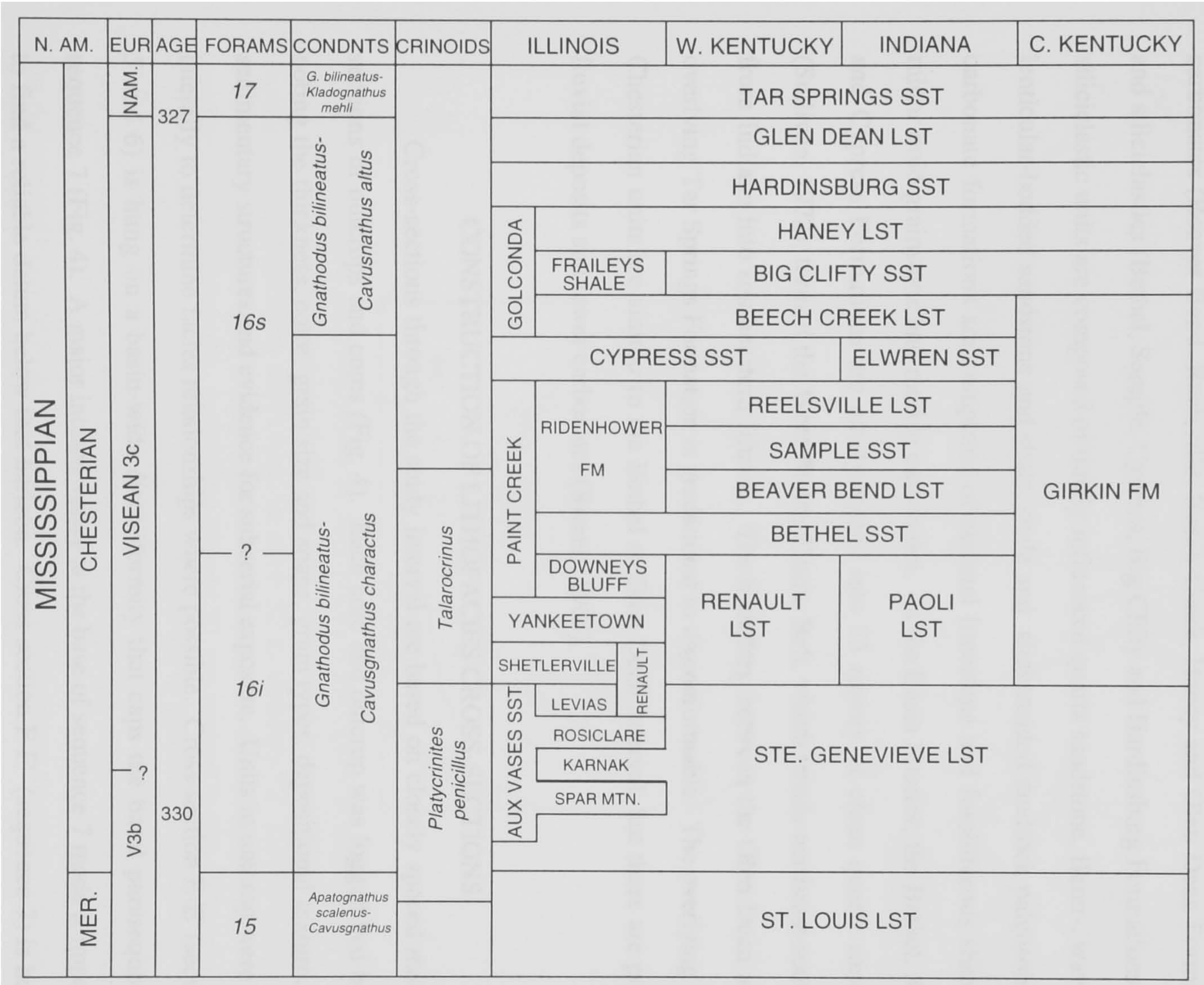


Fig. 3. Biostratigraphy and correlation chart for Ste. Genevieve to Glen Dean interval (modified from Sable and Dever, 1990 and Maples and Waters, 1987).

There is a major unconformity between the Bethel Formation and the underlying Paoli Formation marked by well-developed paleosols and incised-valleys up to 75 meters deep (Friberg et al., 1969). The Bethel to Glen Dean interval is composed of interbedded carbonates (Beaver Bend, Reelsville, Beech Creek, Haney and Glen Dean Formations) and siliciclastics (Bethel, Sample, Cypress, Big Clifty and Hardinsburg Formations). The siliciclastic units are composed of tidally influenced quartz sandstone, flaser-, wavy- and lenticular-bedded sandstone and shale, shale and slickensided mudrock paleosols. The carbonate formations are composed of skeletal limestone and fossiliferous shale with minor ooid grainstone and muddy carbonates. In the Basin Interior, the Bethel, Sample and Cypress Formations are amalgamated into 75 meters of clean quartz sandstone (Sullivan, 1972), termed the West Baden Clastic Belt, which trends northeast-southwest from Indiana into southeastern Illinois. The boundary between the Glen Dean and the overlying Tar Springs Formation is gradational to disconformable. The overlying Upper Chesterian units are similar to the Bethel to Glen Dean interval, but there are possible fluvial deposits and fewer carbonates (Swann, 1963).

CONSTRUCTION OF LITHOFACIES CROSS-SECTIONS

Cross-sections through the study interval are based on closely spaced measured sections of outcrops and cores (Fig. 4). Each core and outcrop was logged bed-by-bed, noting the thickness, color, grain size and shape, grain types, depositional texture, biota, sedimentary structures and evidence for subaerial exposure. Units in outcrop were traced laterally to determine facies relationships where possible. Cross-section E-E' (sequences 5 and 6) is hung on a basin-wide disconformity that caps the basal parasequence in sequence 7 (Fig. 4). A major incised-valley at the base of sequence 7 made it impossible to find a reliable datum below that horizon. Cross-section F-F' (sequence 7) is hung on

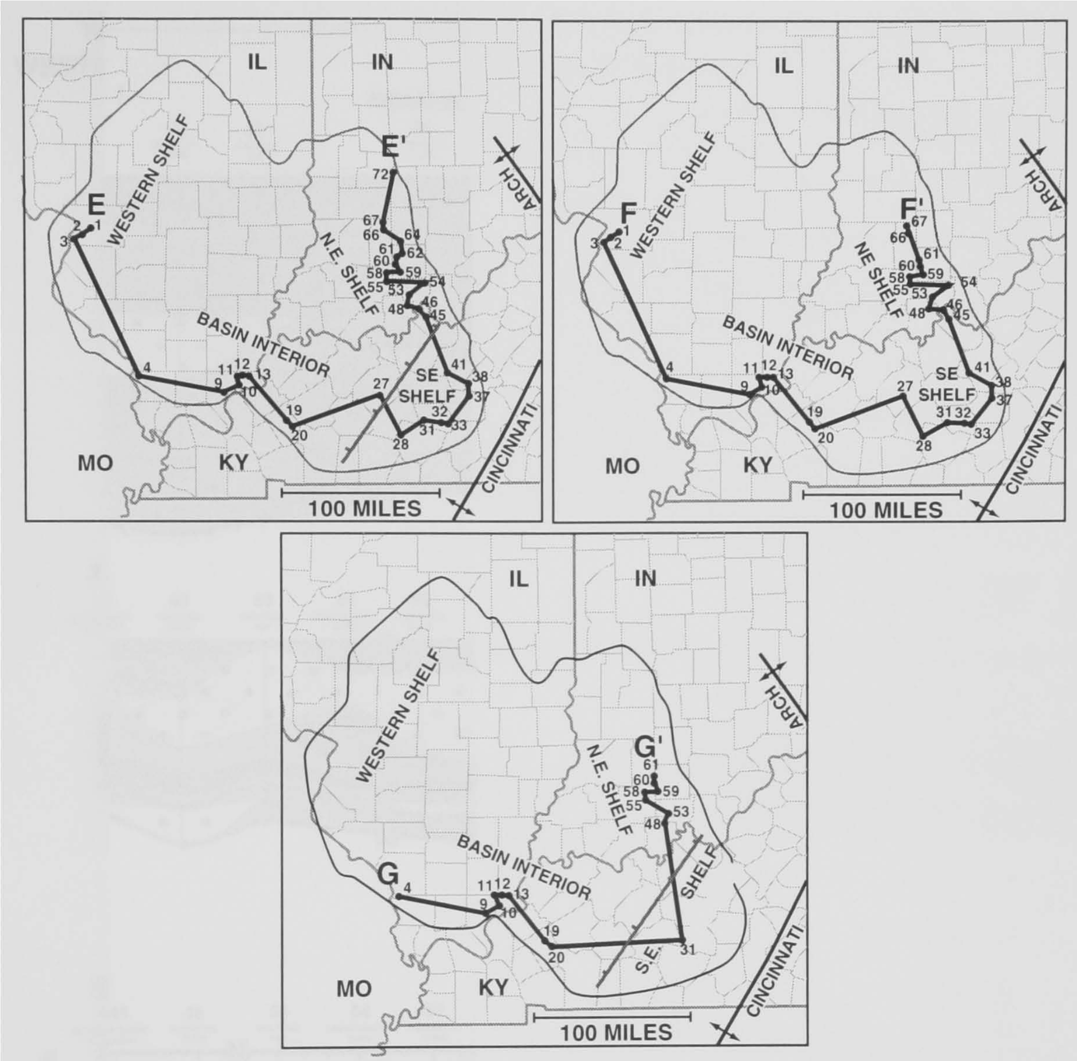
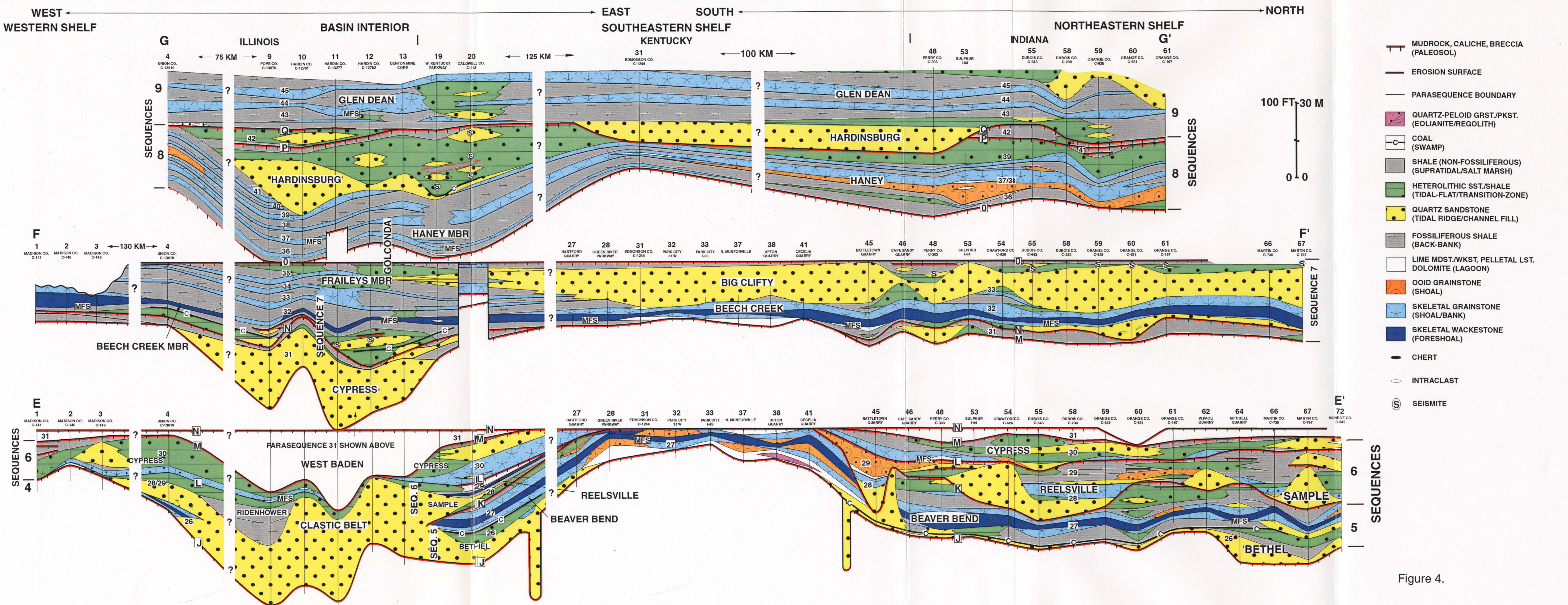


Fig. 4. Interpretive cross-sections of sequences 5 through 9 in the Bethel to Glen Dean interval. Cross sections E-E' and F-F' trend west to east from location 1 to location 33 and south to north from location 33 to location 72. Cross-section G-G' trends west to east from location 4 to location 31 and south to north from location 31 to location 61. Locations 33 and 31 are closest to the Cincinnati Arch and thinning at these locations is due to proximity to the Arch. The cross-sections do not go over the arch. Locations are equally spaced except where there are major gaps between sections. See location maps for actual distance between sections. Disconformities are lettered from J to Q and parasequences are numbered from 26 to 45. Bethel channel is not drawn to scale; its true depth is 75 meters. Cross-sections on following page.



WEST
EAST
WESTERN SHELF
BASIN INTERIOR
SOUTHEASTERN SHELF
SOUTH
NORTH
NORTHEASTERN SHELF
ILLINOIS
KENTUCKY
INDIANA
G
G'
F
F'
E
E'
75 KM
125 KM
100 KM
130 KM
SEQUENCES
GLEN DEAN
HARDINSBURG
HANEY
HANEY MBR
FRAILEYS MBR
GOLCONDA
BIG CLIFTY
BEECH CREEK
BEECH CREEK MBR
CYPRESS
SEQUENCE 7
PARASEQUENCE 31 SHOWN ABOVE
WEST BADEN
RIDENHOWER
CLASTIC BELT
SAMPLE
BETHEL
REELSVILLE
BEAVER BEND
SEQ. 6
SEQ. 5
MFS
100 FT 30 M
MUDROCK, CALICHE, BRECCIA (PALEOSOL)
EROSION SURFACE
PARASEQUENCE BOUNDARY
QUARTZ-PELOID GRST./PKST. (EOLIANITE/REGOLITH)
COAL (SWAMP)
SHALE (NON-FOSSILIFEROUS) (SUPRATIDAL/SALT MARSH)
HETEROLITHIC SST./SHALE (TIDAL-FLAT/TRANSITION-ZONE)
QUARTZ SANDSTONE (TIDAL RIDGE/CHANNEL FILL)
FOSSILIFEROUS SHALE (BACK-BANK)
LIME MDST./WKST, PELLETAL LST. DOLOMITE (LAGOON)
OOID GRAINSTONE (SHOAL)
SKELETAL GRAINSTONE (SHOAL/BANK)
SKELETAL WACKESTONE (FORESHOAL)
CHERT
INTRACLAST
SEISMITE

Figure 4.

the basin-wide paleosol at the top of the sequence (Fig. 4). Cross-section G-G' (sequence 8 and the base of sequence 9) is hung on the base of the Glen Dean Limestone.

Initial correlations were made using distinctive marker beds such as paleosols and laterally extensive limestone and fossiliferous shale units. In outcrop, limestone units commonly overlie paleosol horizons except in cases where incision occurred. In these cases, marine, quartz sandstone, incised-valley-fills commonly grade upward into the overlying limestone unit. Similar transitions observed in cores are interpreted to be marine incised-valley-fills. Sequences are numbered 5 to 9, parasequences have been numbered 26 to 45 and disconformities have been labeled J to Q (sequences 1 to 4, parasequences 1 to 25 and disconformities A to Q occur in the underlying Ste. Genevieve to Paoli interval (Chapter 2)).

LITHOFACIES

Lithofacies are summarized in detail in Table 1. Most siliciclastics were deposited in nearshore tide-dominated settings and most carbonates were deposited in offshore storm-dominated settings (Fig. 5). From offshore to onshore the lithofacies in the Bethel to Glen Dean interval include:

Skeletal Wackestone (Foreshoal).- Open-marine skeletal wackestone (Table 1) commonly consists of whole and fragmented skeletons of echinoderms, brachiopods, bryozoans with less common gastropods, mollusks, corals and foramanifera in a lime mud matrix that commonly is pelletal or slightly argillaceous (Fig. 6A) (Harris, 1992). This facies forms laterally extensive sheet-like units in the Beaver Bend and Beech Creek Formations that can be correlated across the Eastern Shelf and into the Basin Interior. In the Fraileys, Haney and Glen Dean Formations, skeletal wackestone is patchy and only present downdip. The predominance of this facies in the Basin Interior and its relatively high fossil diversity suggest that it formed offshore in a sub-wave base, low-energy

Table 1. Mixed Carbonate-Siliciclastic Ramp Facies

Facies	**Shale (marginal marine/marsh)**	**Heterolithic (tidal flat and quartz sandstone-fossiliferous shale transition)**	**Quartz Sandstone (tidal bars, channel fills)**	**Fossiliferous Shale (carbonate-clastic transition)**	**Skeletal Limestone (bank, sheet)**
Occurrence	Underlies and overlies mudrock paleosols	Commonly overlies quartz sandstone and underlies salt-arsh and paleosols- more common in Bethel to Hardinsburg interval	Quartz sandstone is common throughout the Aux Vases in the western part of the basin, but rare elsewhere.	Green fossiliferous shale occurs in Ste. Genevieve of basin interior and Paoli of eastern shelf; dark gray shale occurs in basin interior Renault-Yankeetown Formations and also caps some parasequences.	Common in basin interior and western shelf in Paoli interval. Common at bases of parasequences with fossiliferous shale caps.
Lithology	Shale	Ranges from 10% to 85% quartz sand or silt interlaminated with shale	Quartz sandstone can have up to 15% shale	Fossiliferous shale and argillaceous skeletal wackestone	Skeletal grainstone/ packstone with up to 15 % shale
Color	Dark gray to olive green	Sand-white/green, shale- dark gray/green	White to light gray/green	Olive green and dark gray	Dark gray
Depositional texture and grain type	More micaceous to northwest	Very fine- to fine-grained quartz sand and quartz silt	Very-fine to medium-grained, subangular, well-sorted	Wackestone/fissile shale; whole brachiopods and bryozoans.	Grainstone/ packstone; fine to very coarse, subangular, skeletal fragments; micritization common.
Bedding and sedimentary structures	Poorly fissile	Flaser-, wavy- and lenticular-bedding and tidal rythmites (Huff, 1993).	Cross-bedded and massive bedded, flaser-bedding common	Fissile, flat-bedded	Thick bedded and massive
Biota	Plant fragments common in Bethel to Glen Dean interval, not in Aux Vases	Rare echinoderm fragments in transition facies.	Rare echinoderm fragments, could be transported or eroded from underlying limestone	Bryozoans, brachiopods and echinoderms	Abundant echinoderms; brachiopods, bryozoans; common mollusks, foramanifera; rare trilobites.

Table 1. Mixed Carbonate-Siliciclastic Ramp Facies

Facies	**Shale (marginal marine/marsh)**	**Heterolithic (tidal flat and quartz sandstone-fossiliferous shale transition)**	**Quartz Sandstone (tidal bars, channel fills)**	**Fossiliferous Shale (carbonate-clastic transition)**	**Skeletal Limestone (bank, sheet)**
Occurrence	Underlies and overlies mudrock paleosols	Commonly overlies quartz sandstone and underlies salt-arsh and paleosols- more common in Bethel to Hardinsburg interval	Quartz sandstone is common throughout the Aux Vases in the western part of the basin, but rare elsewhere.	Green fossiliferous shale occurs in Ste. Genevieve of basin interior and Paoli of eastern shelf; dark gray shale occurs in basin interior Renault-Yankeetown Formations and also caps some parasequences.	Common in basin interior and western shelf in Paoli interval. Common at bases of parasequences with fossiliferous shale caps.
Lithology	Shale	Ranges from 10% to 85% quartz sand or silt interlaminated with shale	Quartz sandstone can have up to 15% shale	Fossiliferous shale and argillaceous skeletal wackestone	Skeletal grainstone/ packstone with up to 15 % shale
Color	Dark gray to olive green	Sand-white/green, shale- dark gray/green	White to light gray/green	Olive green and dark gray	Dark gray
Depositional texture and grain type	More micaceous to northwest	Very fine- to fine-grained quartz sand and quartz silt	Very-fine to medium-grained, subangular, well-sorted	Wackestone/fissile shale; whole brachiopods and bryozoans.	Grainstone/ packstone; fine to very coarse, subangular, skeletal fragments; micritization common.
Bedding and sedimentary structures	Poorly fissile	Flaser-, wavy- and lenticular-bedding and tidal rythmites (Huff, 1993).	Cross-bedded and massive bedded, flaser-bedding common	Fissile, flat-bedded	Thick bedded and massive
Biota	Plant fragments common in Bethel to Glen Dean interval, not in Aux Vases	Rare echinoderm fragments in transition facies.	Rare echinoderm fragments, could be transported or eroded from underlying limestone	Bryozoans, brachiopods and echinoderms	Abundant echinoderms; brachiopods, bryozoans; common mollusks, foramanifera; rare trilobites.

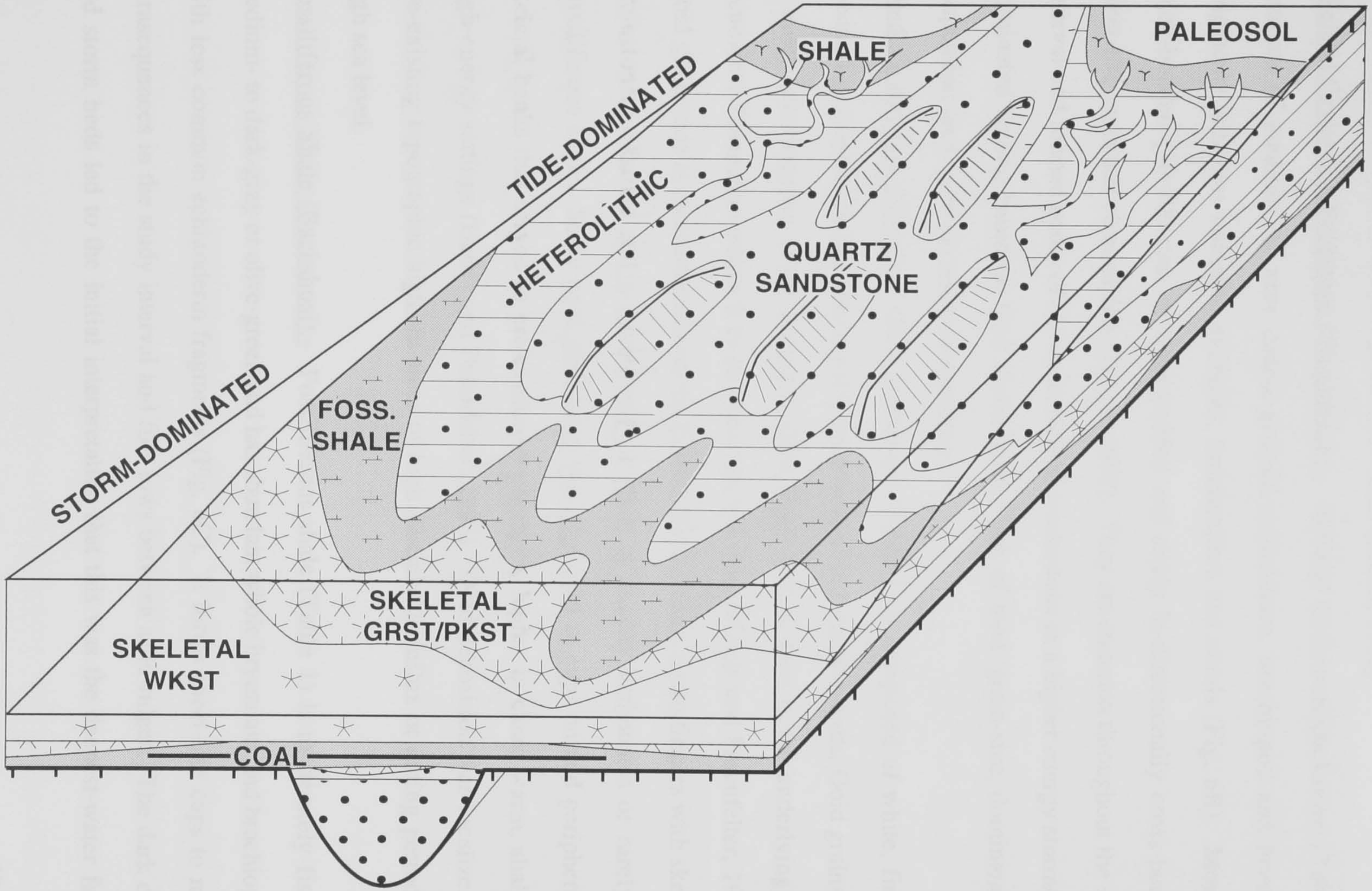


Fig. 5. Schematic representation of depositional environments for Bethel to Glen Dean Interval (not to scale).

environment seaward of skeletal grainstone banks (Harris, 1992).

Skeletal Grainstone/Packstone.(Shoal/Bank).- Skeletal grainstone/packstone (Table 1) consists of medium- to very coarse-grained echinoderm, brachiopod and bryozoan fragments with less common mollusks, foramanifera and corals (Fig. 6B). Most are thick-bedded (10-100 cm) or cross-bedded and rarely bi-directionally cross-bedded, sheet-like units and channel-fills (Harris, 1992). They are common throughout the study interval and formed updip of the open skeletal wackestone in a higher energy storm/tide-dominated setting based on their distribution, lack of mud, grain-size, dominance of suspension feeding fauna and cross-bedding.

Ooid Grainstone (Shoal).- Ooid grainstone (Table 1) is composed of white, fine to medium-grained ooids in cross-bedded, thick bedded and massive units. Ooid grainstone is much less common in the Bethel to Glen Dean interval than in the underlying Ste. Genevieve Formation where it is an important reservoir (Harris and Fraunfelter, 1993). Ooid grainstone is most common on the Eastern Shelf, where it interfingers with skeletal grainstone/packstone and is commonly overlain by muddy carbonates or rarely by fossiliferous shale. Some ooid grainstone was deposited as caps on and peripheral to skeletal banks that served as pre-existing topographic highs in clear, warm, shallow, high-energy settings (Harris and Fraunfelter, 1993). Ooid deposition was localized on pre-existing topographic highs in areas of high siliciclastic influx or during periods of high sea level.

Fossiliferous Shale (Backshoal).- Fossiliferous shale (Table 1) is moderately fissile, medium- to dark-gray or olive-green and has abundant whole bryozoans and brachiopods with less common echinoderm fragments (Fig. 6C). It forms sheet-like caps to many parasequences in the study interval and fills lows between sand-ridges. The dark color and storm beds led to the initial interpretation that this was the deepest-water facies

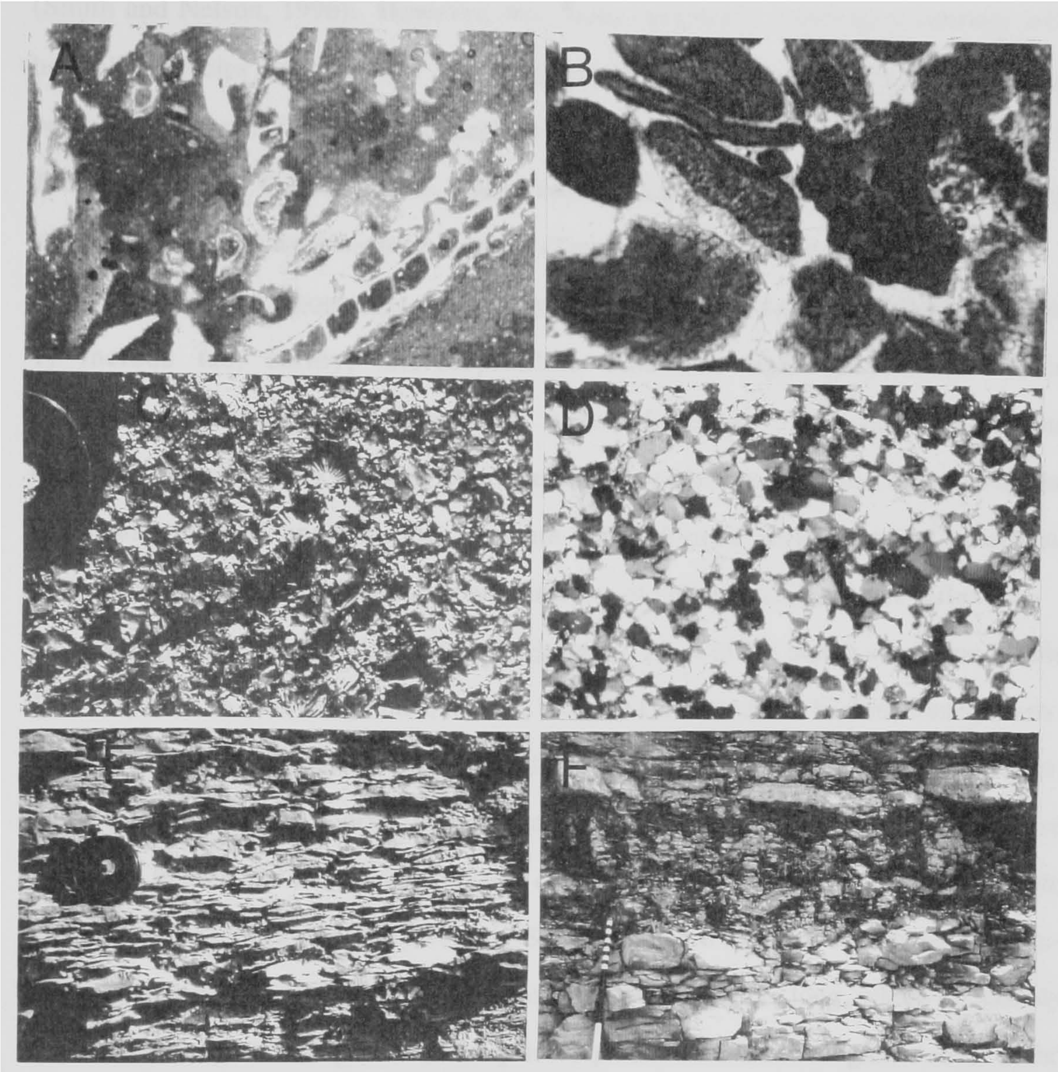


Fig. 6. Photographs of important rock types. Scale in A, B and C is 0.25 mm. A) Open skeletal wackestone; B) skeletal grainstone; C) fossiliferous shale (lens cap for scale); D) very fine-grained quartz sandstone; E) heterolithic facies (lens cap for scale); F) paleosol with stacked dolomitic nodules (staff marked in feet).

(Smith and Nelson, 1996). However, this facies commonly interfingers laterally with heterolithic facies and underlies paleosol horizons. This suggests a backshoal setting between offshore carbonates and nearshore siliciclastics (Harris and Fraunfelter, 1993; Treworgy, 1985). Common, thin, graded beds of skeletal debris were deposited during storms in the otherwise low-energy setting (Treworgy, 1985; Harris, 1992).

Restricted Muddy Carbonates (Lagoon/Intershoal).- Restricted muddy carbonates (Table 1) include tan, earthy, microcrystalline dolomite, light gray-brown lime mudstone/wackestone or medium-brown pelletal limestone. Most muddy carbonate units are massive or thick-bedded. Restricted muddy carbonates are most common on the Southeastern Shelf and laminites are rare. They formed in low-energy, restricted environments behind the shoals and banks in areas of little to no siliciclastic influx.

Quartz Sandstone (Sand Ridges/Shallow Marine).- Most of the quartz sandstone (Table 1) is very fine- to medium-grained, mature, burrowed and occurs in bi-directionally cross-bedded, trough cross-bedded, thick-bedded and massive-bedded units (Fig. 6D) (Potter, 1963; Cole and Nelson, 1995). It occurs in ridge-like units and channel-fills, which are important hydrocarbon reservoirs and less commonly in sheets (Seyler and Cluff, 1990; Zuppann and Keith, 1988; Grube, 1992). Sand- ridges are elongate convex-up barforms up to 18 meters thick, 2 kilometers wide and several kilometers long, which commonly trend northeast-southwest parallel to the paleoslope and have been recognized in all of the siliciclastic formations (Fig. 7) (Off, 1963; Specht, 1985; Zuppann and Keith, 1988; Grube, 1992; Cole and Nelson, 1995). Channel-fill sandstones are tidally-influenced and commonly contain marine skeletal debris and trace fossils (Friberg et al., 1969; Visher, 1980; Ambers and Robinson, 1992).

Quartz sandstone was deposited in a tide-dominated shallow-marine environment, indicated by bi-directional cross-bedding, dip-parallel orientation of sand-ridges, marine trace fossils, and abundant tidal sedimentary structures (Specht, 1985; Grube, 1992). The

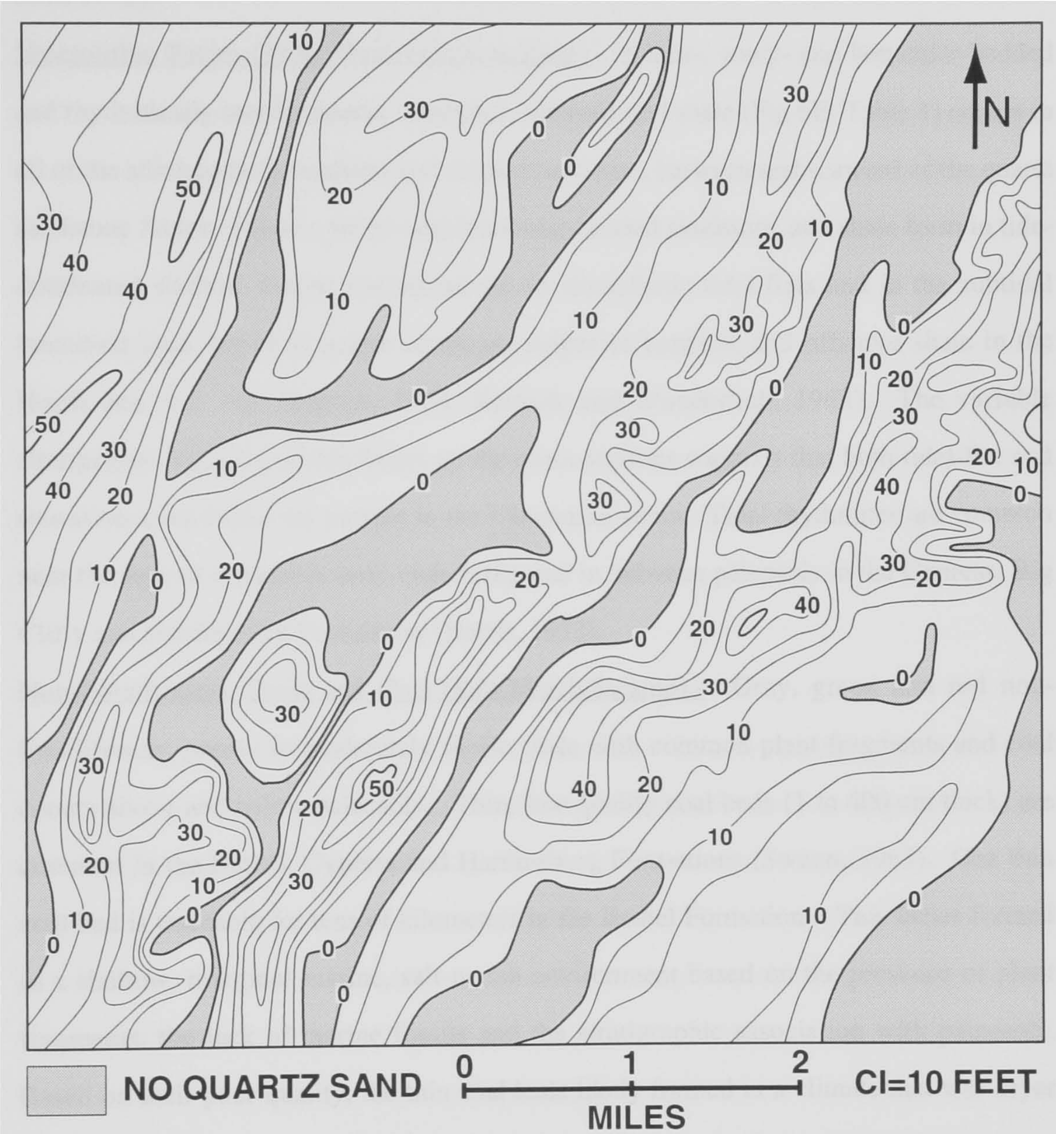


Fig. 7. Isopach map (thickness in feet) of total sand thickness in Big Clifty Formation of southwestern Indiana (modified from Specht, 1988). Northeast-southwest trending sand-ridges are more than 50 feet thick.

convex up barforms resemble tidal sand-ridges in macrotidal settings in the North Sea (Off, 1963).

Heterolithic Facies (Tidal-Flat/Transition-Zone).- Flaser-, wavy- and lenticular-bedded and rhythmically interlaminated sandstone, siltstone and shale (Fig 6E: Table 1) occurs in all of the siliciclastic formations and formed landward, between and seaward of the quartz sandstone facies. Flaser-, wavy- and lenticular-bedded sandstone and shale form in tide-dominated shallow marine environments on siliciclastic tidal-flats and in the subtidal transition-zone between quartz sandstone ridges or barriers, and offshore shale in the North Sea (c.f. van Stratten, 1954; Reinick and Wunderlich, 1968). The variable stratigraphic position of this facies on the cross-sections suggests that both tidal-flat and transition-zone facies are present in the Chesterian strata. Tidal rhythmites are common near the tops of sequences both underlying and in between paleosols in the Cypress, Big Clifty and Hardinsburg Formations (Harris, 1992).

Non-Fossiliferous Shale and Coal (Salt-Marsh/Swamp).- Gray, green and red non-fossiliferous, poorly to moderately fissile shale with common plant fragments and coal occurs above and below paleosols. Thin, low-quality coal beds (1 to 400 cm thick) are common in the Bethel, Cypress and Hardinsburg Formations (Swann, 1963). One thin coal bed is traceable for tens of kilometers in the Bethel Formation. This facies formed in a shallow, marginal marine, salt-marsh environment based on the presence of plant fragments, the lack of marine fossils and the stratigraphic association with paleosols. Based on their poor quality, the thin coal beds likely formed in a climate that was dryer than later in Pennsylvanian when much thicker coals formed (Rowley et al., 1985).

Paleosols.- During exposure, blocky, slickensided mudrock formed on shale and caliche and breccia developed on carbonates and less commonly on quartz sandstone. Mudrock paleosols are red, green, or dark gray and commonly brecciated, root-disrupted and may contain brecciated carbonate nodules (Fig. 6F). Dolomite formed in swales on the exposed surface and was brecciated during subsequent exposure (Ambers and Petzold, 1992). These blocky, slickensided mudrock paleosols likely formed during periods of prolonged exposure in an arid or seasonal semi-arid climate (Ambers and Petzold, 1992).

Caliche paleosols are gray, brown and tan cryptocrystalline calcium carbonate and commonly have rootlet casts filled with sparry calcite (Walls et al., 1975; Ettensohn et al., 1988). The caliche coats grains, lines vertical fractures and bedding planes or forms thick accumulations of laminar caliche beneath disconformities. Brecciated carbonate/regolith forms argillaceous, brown- weathering recessive intervals, with angular to subrounded fragments of limestone in fitted-fabrics cemented by sparry calcite, commonly associated with thick laminar caliche and silicified limestone. Tepee structures up to a meter wide and 30 cm high are common near the tops of well-developed breccia/regolith profiles.

SEQUENCE STRATIGRAPHY

A sequence is a genetically related succession of strata lacking apparent internal unconformities, composed of parasequences and parasequence sets arranged in systems tracts and bounded by unconformities or their correlative conformities (Mitchum and Van Wagoner, 1991). A parasequence is a relatively conformable succession of genetically related beds and bedsets bounded by marine flooding surfaces and their correlative surfaces (Mitchum and Van Wagoner, 1991). Facies within parasequences commonly show a shallowing-upward trend and parasequence boundaries are marked by rapid deepening events (Goodwin and Anderson, 1985; Mitchum and Van Wagoner, 1991). A depositional sequence consists of a <u>lowstand systems tract</u> (LST), which is deposited seaward of the shelf break or depositional shoreline break and made up of one or more progradational parasequence sets; a <u>transgressive systems tract</u> (TST), which consists of one or more retrogradational parasequence sets and capped by the maximum flooding surface (MFS) at the change from retrogradational to aggradational or progradational parasequence sets; and a highstand systems tract (HST), which is composed of one or more aggradational to progradational parasequence sets (Mitchum and Van Wagoner, 1991).

The Bethel to Glen Dean interval contains five sequences numbered 5 to 9; sequences 1 to 4 are in the underlying Ste. Genevieve to Paoli interval (Chapter 2). The sequences generally consist of: 1) a sequence boundary with common incised-valleys; 2) early TST parasequences filling incised-valleys and consisting of fining-upward from quartz sandstone to heterolithic facies and coal; 3) late TST and early HST limestone-shale parasequences, with the MFS commonly occurring at or near the base of carbonate units; and 4) late HST heterolithic siliciclastics capped by a sequence boundary.

Parasequences in the study interval include: 1) transgressive-regressive parasequences that have sandy siliciclastic bases overlain by carbonate units and shale-dominated siliciclastics and occur in transgressive systems tracts; and 2) regressive parasequences with skeletal limestone bases and shale-dominated siliciclastic caps that are confined to highstand systems tracts. Most parasequences are regionally traceable across the Eastern Shelf and into the Basin Interior.

Sequence 5

Sequence 5 is bounded at the base by disconformity J, which developed on top of the Ste. Genevieve to Paoli interval containing sequences 1 to 4 (Chapter 2). It is bounded at the top by disconformity K and includes the Bethel, Beaver Bend Formation and basal Sample Formations (Fig. 4). The sequence is up to 25 meters thick on Northeastern Shelf, thins to 7 meters on the Southeastern Shelf near the Cincinnati Arch (location 33) and thins to zero on the Western Shelf. The top of the sequence cannot be recognized in the central Basin Interior where the Bethel, Sample and Cypress Formations merge to form the West Baden Clastic Belt. On the Southeastern Shelf, the uppermost disconformity-bounded carbonate unit with the crinoid *Talarocrinus* within the Girkin Formation (Fig. 3) is equivalent to the Beaver Bend (G. Dever, K.G.S., pers. comm., 1994). On the Western Shelf, the Ridenhower Formation commonly has at least two limestone units, the lower may be traceable into the Beaver Bend Limestone (Swann, 1963). Because there is no biostratigraphic control on the Ridenhower limestones, correlations within sequences 5 and 6 between the Eastern and Western Shelves are speculative.

The basal unconformity (J) is a type 1 sequence boundary marked by paleosols on most of the Eastern and Western Shelves and up to 75 meters of incision into underlying carbonates in a channel extending from south-central Indiana to western Kentucky (Fig. 8A) (Reynolds and Vincent, 1967; Friberg et al., 1969). The transgressive systems tract (TST) is composed of transgressive-regressive parasequence 26 and the lower half of transgressive-regressive parasequence 27. The valley-fill of parasequence 26 is composed of quartz sandstone with unidirectional planar and trough cross-bedding that dips in the downstream direction to the southwest in much of the valley-fill (Friberg et al., 1969). The unidirectional dips of the cross-bedding suggest that the basal sandstone may be fluvial. Quartzose-skeletal limestone and heterolithic facies occur near the top of the channel-fill indicating a marine origin (Friberg et al., 1969). Friberg et al. (1969) interpreted this as a submarine channel-fill based on incision of only 10 to 15 meters elsewhere in the basin, the lack of any other evidence of exposure and the presence of limestone within the channel-fill. However, the fact that the incised-valley can be traced laterally into a regional paleosol horizon and the presence of possible fluvial facies at the base of the channel-fill suggests that the valley was eroded in a subaerial environment. A thin, laterally extensive coal bed caps parasequence 26 in the eastern side of the basin. On the Northeastern Shelf and in the eastern Basin Interior, parasequence 27 is a transgressive-regressive parasequence that consists of marine siliciclastics overlain by

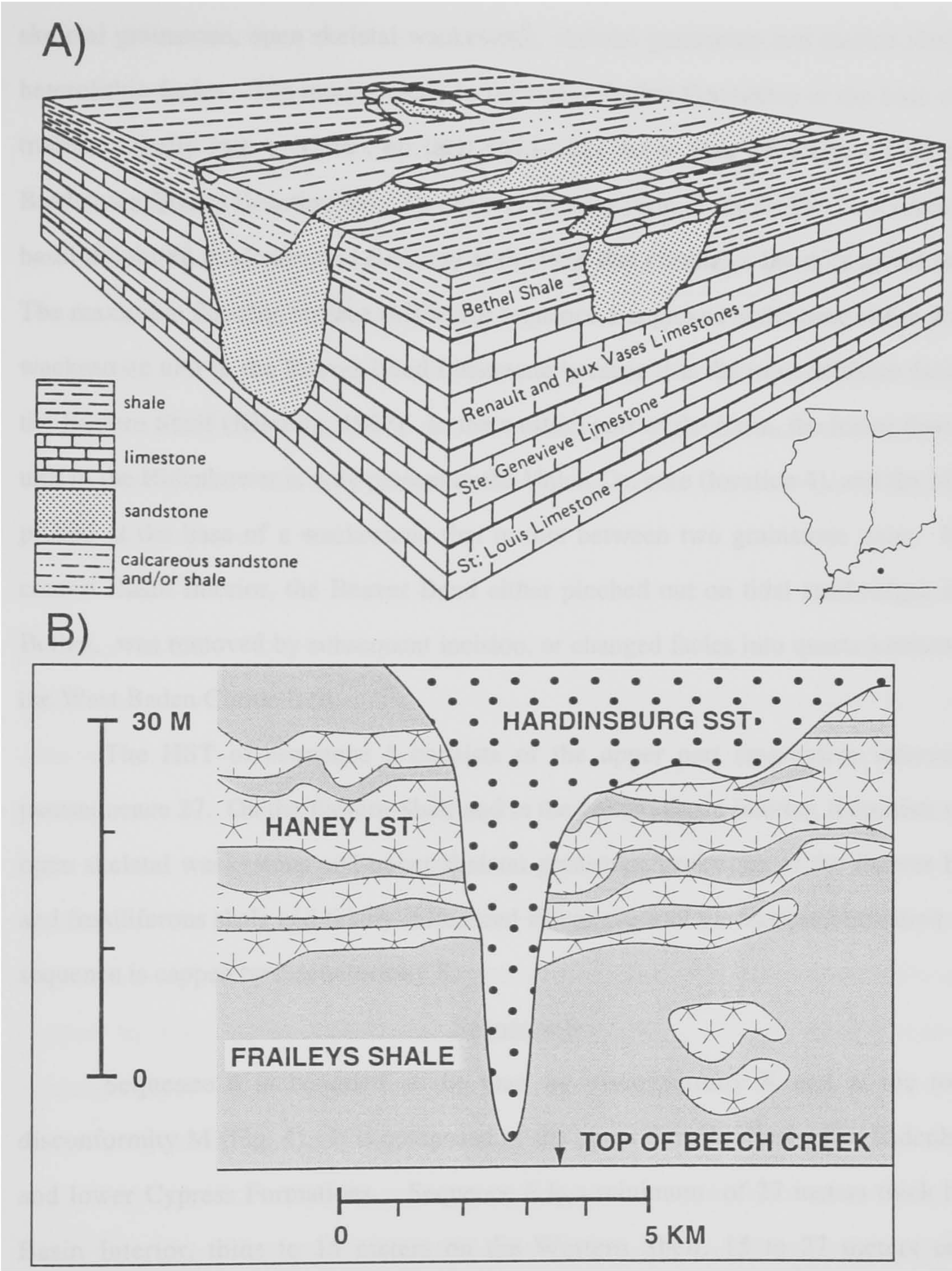


Fig. 8. A) Incised valley underlying Sequence 5 that cuts through more than 75 meters of the underlying Paoli and Ste. Genevieve Formations (from Seyler and Cluff, 1990 after Reynolds and Vincent, 1967). B) Incised Valley at the base of Sequence 9 cuts down more than 50 meters into the lower Hardinsburg, Haney and Fraileys Members of the Golconda Formation (modified from Treworgy, 1988).

skeletal grainstone, open skeletal wackestone, skeletal grainstone and marine shale and heterolithic facies. The marine shale in the upper Bethel Formation at the base can be traced laterally into a convex-up tidal sand-ridge more than 18 meters thick at the Battletown Quarry (location 45). The Beaver Bend portion of the Girkin Limestone has a basal dolomitic mudstone and eolianite at the base that correlates laterally to the Bethel. The maximum flooding surface (MFS) for sequence 5 is picked at the base of the skeletal wackestone unit in the Beaver Bend Limestone because it is the most offshore facies on the Eastern Shelf (Kissling, 1967). In the western part of the basin, the lower limestone unit in the Ridenhower is only present in the Union Co. core (location 4), and the MFS is picked at the base of a wackestone that occurs between two grainstone units. In the central Basin Interior, the Beaver Bend either pinched out on tidal sand-ridges of the Bethel, was removed by subsequent incision, or changed facies into quartz sandstone of the West Baden Clastic Belt.

The HST of sequence 5 consists of the upper part (regressive interval) of parasequence 27. On the Eastern Shelf and in the eastern Basin Interior, it consists of the open skeletal wackestone and upper skeletal grainstone/packstone of the Beaver Bend, and fossiliferous shale and tidally-influenced siliciclastics of the Sample Formation. The sequence is capped by disconformity K.

Sequence 6

Sequence 6 is bounded at the base by disconformity K and at the top by disconformity M (Fig. 4). It is composed of the upper Sample, Reelsville, Ridenhower and lower Cypress Formations. Sequence 6 is a minimum of 27 meters thick in the Basin Interior, thins to 15 meters on the Western Shelf, 15 to 27 meters on the Northeastern Shelf and pinches out completely on the Southeastern Shelf.

The basal sequence boundary (disconformity K) is a slickensided mudrock paleosol that can be traced laterally into incised-valleys where at least 15 meters of incision occurred on the Northeastern Shelf (Kissling, 1967; Ambers and Robinson, 1992). Paleosols occur at the base of incised-valleys at locations 45 and 67 on the Northeastern Shelf. A well-developed caliche and breccia with tepee structures developed on the Southeastern Shelf that remained emergent throughout sequence 6 and in the early part of sequence 7. In the Basin Interior, multiple incised-valleys occur within the West Baden Clastic Belt and one of these likely correlates with the sequence boundary (J. Nelson, I.S.G.S., pers. comm., 1996). On the Western Shelf, disconformity K may occur at the base of the fossiliferous sandstone in the Union Co. core (location 4) and is coincident with disconformity J farther to the northwest, where sequence 5 pinches out (locations 1 to 3).

The TST includes transgressive-regressive parasequence 28 and regressive parasequence 29. On the Northeastern Shelf, the base of parasequence 28 consists of tidally-influenced quartz sandstone units up to 12 meters thick that partially fill incised-valleys. In outcrop, these quartz sandstone channel-fills are commonly flaser-bedded and contain abundant marine trace fossils (Ambers and Robinson, 1992). Some channels were not completely filled with quartz sandstone, and skeletal grainstone/packstone of the lower Reelsville Formation was deposited in remaining topographic lows. In the eastern Basin Interior, parasequence 28 is composed of a basal fossiliferous heterolithic bed capped by a quartz sandstone and a fossiliferous shale. Transgressive-regressive parasequence 29 has a thin quartz sandstone at the base capped by an argillaceous skeletal limestone/fossiliferous shale and non-fossiliferous shale with patchy paleosols near the top. The paleosols cannot be correlated regionally and are therefore treated as caps of local autocycles within parasequence 29. Parasequences 28 and 29 are unrecognizable in the West Baden Clastic Belt and cannot be distinguished from one another in the western Basin Interior due to lack of data. Parasequence 29 is capped by a well-developed paleosol (disconformity L) across the eastern side of the basin. Disconformity L is overlain by a carbonate unit, composed of skeletal grainstone/packstone on the Eastern Shelf, skeletal wackestone in the central Basin Interior and quartzose intraclastic skeletal packstone on the Western Shelf, that may be equivalent to the Turkey Fork Limestone Member of the Cypress Formation of Kissling (1967).

The MFS for sequence 6 is picked at the base of the limestone in parasequence 30 because it is the first carbonate unit in sequence 6 to be deposited on the Western Shelf and in most of the West Baden Clastic Belt. The HST of sequence 6 consists of regressive parasequence 30 that is composed of the Turkey Fork Limestone at the base overlain by heterolithic tidal flat facies with patchy quartz sandstone channel-fills. The sequence is capped by disconformity M.

Sequence 7

Sequence 7 is bounded at the base by disconformity M and at the top by disconformity O (Fig. 4). It includes the upper Cypress Formation, the Beech Creek Limestone and most of the Big Clifty Sandstone in Indiana and the Beech Creek and Fraileys Members of the Golconda Formation in Illinois and western Kentucky. The sequence is up to 68 meters thick in the Basin Interior, thins to 18 meters on the Western Shelf, 35 meters on the Northeastern Shelf and 21 meters on the Southeastern Shelf. However, most of the difference in thickness occurs in the upper Cypress; the interval between the Beech Creek and the top of the sequence is of approximately equal thickness across the basin.

Sequence 7 is bounded at the base by a type 1 unconformity (disconformity M) marked by a regional paleosol on the Eastern and Western Shelves and up to 45 meters of incision in the Basin Interior. This boundary may coincide with a conodont biostratigraphic zone boundary between the Reelsville and the Beech Creek (Collinson et al., 1962). The TST includes a disconformity bounded transgressive-regressive parasequence (31) and the transgressive interval of transgressive-regressive parasequence

32. Parasequence 31 is up to 35 meters thick in the Basin Interior but thins to less than five meters on the western and northeastern shelves and pinches out completely on the Southeastern Shelf. In the Basin Interior, it is composed of a thick quartz sandstone marine channel-fill that passes laterally into storm-bedded fossiliferous shale (locations 1, 2, 19 and 20) and is overlain by heterolithic facies. On the Northeastern Shelf, the parasequence consists of shale and tidal rhythmites capped by a paleosol that can be correlated basin-wide and traced into a 6 meter incised-valley on the Northeastern Shelf (disconformity N). In the Basin Interior, the base of the overlying parasequence 32 consists of basal siliciclastic tidal flat facies and shale with poorly developed coal and paleosol horizons and NE-SW trending lenticular quartz sandstone bodies in the Basin Interior that are important hydrocarbon reservoirs (Grube, 1992). These strata are overlain by the Beech Creek Limestone that is composed of a patchy basal skeletal grainstone unit overlain by a laterally extensive open skeletal wackestone and a second skeletal grainstone/packstone unit on the Eastern and Western Shelves (Harris, 1992). The grainstone units pinch out in the Basin Interior where the Beech Creek thins and is composed of the open skeletal wackestone (Harris, 1992). The MFS of sequence 8 is picked at the base of the open skeletal wackestone, which is the most offshore facies deposited on the Eastern Shelf (Harris, 1992). The Beech Creek is the first unit to be deposited on the Southeastern Shelf since sequence 5.

The HST is composed of the regressive interval of transgressive-regressive parasequence 32 and regressive parasequences 33 to 35 except on much of the Eastern Shelf where it consists of the Big Clifty Sandstone. In the Basin Interior, the HST consists of limestone-shale parasequences 32 to 35 which can be correlated regionally in the Beech Creek and Fraileys Shale Members of the Golconda Formation. The fossiliferous shale parasequence caps are progressively more quartzose toward the top of the Fraileys. On the Eastern Shelf, the regressive part of parasequence 32 consists of

Beech Creek open skeletal wackestone and grainstone overlain in places by dark gray shale. In most locations on the Eastern Shelf, the Big Clifty Sandstone lies directly on top of the Beech Creek Limestone. Only in a few locations is the base of parasequence 33 is preserved (locations 54 and 55). Tidal channel-fills occur at the base of the Big Clifty Sandstone (Visher, 1980). The Big Clifty Sandstone may not have started prograding from the east until after deposition of parasequence 33, and may have eroded underlying shaly units. The Big Clifty Sandstone is a single regressive unit consisting of tidal channels-fills and tidal sand-ridges up to 16 meters thick (Visher, 1980; Specht, 1985; Treworgy, 1985; Harris, 1992). The sequence is capped by disconformity O.

Sequence 8

Sequence 8 is bounded at the base by disconformity O and at the top by disconformity Q (Fig. 4). It is composed of the uppermost Fraileys/Big Clifty, Haney and lower Hardinsburg Formations. The sequence is thickest in the study interval and is up to 50 meters thick in the Basin Interior, 30 meters thick on the Northeastern Shelf and 18 meters thick near the Cincinnati Arch.

The basal unconformity is a type 2 unconformity (disconformity O) marked by a basin-wide paleosol horizon (Treworgy, 1985). The TST of sequence 8 is composed of transgressive-regressive parasequence 36 that consists of a thin shale at the top of the Fraileys/Big Clifty, overlain by a skeletal packstone that is confined to the Basin Interior and a basin wide fossiliferous shale unit of the Haney Limestone. The MFS is picked at the base of parasequence 37, the first laterally extensive limestone unit to be deposited across the Eastern Shelf

The HST of sequence 8 is composed of up to 6 regressive parasequences (37 to 42). Although these units have not been previously recognized as parasequences, subsurface mapping has illustrated that limestone-shale parasequences in the Haney Limestone can be correlated for tens of kilometers (Treworgy, 1985, 1988). The skeletal limestone units are dominantly grainstone/packstone on the Eastern Shelf, packstone/wackestone in the Basin Interior and mixed on the Western Shelf (Treworgy, 1985). In the western part of the basin, parasequences 37 to 41 are present in Union and Pope Co., Illinois (locations 4 and 9) where the Hardinsburg thins and pinches out. Where the Hardinsburg is thicker in the central and eastern parts of the basin, only parasequences 37 to 39 are present. In the Basin Interior, HST parasequences 37 and 38 are discrete parasequences but on the Eastern Shelf, they are amalgamated into an ooid grainstone at the base overlain by a skeletal grainstone/packstone unit. Locally, ooid grainstone units occur in bars with muddy carbonates filling swales (location 53) (Fig. 4). In the central and eastern parts of the basin, parasequence 39 consists of skeletal limestone overlain by siliciclastics of the lower Hardinsburg Formation. The overlying Hardinsburg Formation is composed of heterolithic sandstone/shale tidal flat facies and quartz sandstone lenses on the Northeastern Shelf, and a basal quartz sandstone unit overlain by tidal flat facies in the Basin Interior. One plant fossil/coal horizon near the base of the Hardinsburg can be traced in western Kentucky (locations 19 and 20) and several other coal and plant fossil horizons have been reported from the Hardinsburg (Swann, 1963). Parasequences 40 and 41 cannot be traced within the Hardinsburg Formation in the Basin Interior or on the Eastern Shelf, but may be related to horizons where channels and/or sand-ridges or coals occur. Carbonate parasequences 40 and 41 formed in the western part of the basin while siliciclastics were deposited in the rest of the basin. There is a disconformity-bounded parasequence (42) at the top of the sequence that occurs between disconformities Q and R and consists of siliciclastic tidal flat facies. The sequence is capped by disconformity R.

Sequence 9

Sequence 9 was only partially studied, but is bounded at the base by disconformity Q and consists of the upper Hardinsburg, Glen Dean Limestone and the

Tar Springs Formation (Fig. 4). The Glen Dean is roughly equal thickness across the basin except in places on the Northeastern Shelf where erosion occurred at the base of the Tar Springs.

The sequence is bounded at the base by a type 1 unconformity marked by a well-developed paleosol interpreted to be traceable laterally into incised-valleys cutting down into the Haney (Fig. 8B). Incised valleys occupied by thick quartz sandstones in the Hardinsburg Formation cut through more than 50 meters of the lower Hardinsburg, and the Haney and Fraileys Members of the Golconda Formation (Potter, 1963; Treworgy, 1988; Droste and Horowitz, 1988). Conodont, coral and foramaniferal biostratigraphic zone boundaries between the Haney and Glen Dean Limestones may coincide with this disconformity (Maples and Waters, 1987; Collinson et al., 1962).

The TST is composed of a transgressive-regressive parasequence (43) which includes siliciclastics filling incised-valleys of the Hardinsburg, a patchy skeletal grainstone unit confined to the Basin Interior, and a fossiliferous shale that can be correlated basin-wide. The MFS for sequence 9 occurs at the base of parasequence 44, which contains the first grainstone/packstone unit to be deposited basin wide in the sequence. The HST consists of parasequences 44 and 45 and siliciclastics of the overlying Tar Springs Formation. Channels filled with heterolithic facies cut through the parasequences and fossiliferous shale units within the Glen Dean Limestone. The Tar Springs Formation (not shown on cross-sections) commonly fills valleys up to 12 meters deep incised into the underlying Glen Dean, and has patchy coals and paleosols near its top (Swann, 1963).

Composite-Sequences and Sequence-Pairs

A composite-sequence is a succession of genetically related sequences in which the individual sequences stack into lowstand, transgressive and highstand sequence sets (Mitchum and Van Wagoner, 1991). Ross and Ross (1988) picked two large-scale sequences in the study interval (Fig. 13). The first is bounded at the base by the Ste. Genevieve-St. Louis boundary and at the top by the unconformity within the Cypress (base of sequence 7). Their second large-scale sequence includes the Cypress through Glen Dean and is capped by the disconformity at the base of the Tar Springs Sandstone.

Because the Illinois Basin was situated far updip from the regional ramp margin, it is not possible to trace the downdip extent of sequence boundaries, so composite-sequence boundaries are distinguished from higher frequency sequence boundaries by depth of incision and/or degree of paleosol development (Fig. 9). Two orders of lower frequency sequences are apparent in the Ste. Genevieve through Glen Dean interval. Sequences 1 through 8 are bundled into both sequence-pairs that are themselves bundled into composite-sequences which consist of groups of four sequences (Fig. 9).

Composite-Sequence. The composite-sequence in the Bethel to Glen Dean interval (Fig. 10) is bounded at the base by disconformity J. It may be bounded at the top by disconformity Q which occurs within the Hardinsburg, or at the base of incised-valleys in the overlying Tar Springs as suggested by Ross and Ross (1988). The Hardinsburg boundary has up to 50 meters of incision and likely coincides with a major biostratigraphic zone boundary between the Haney and Glen Dean (Maples and Waters, 1987). Incision at the base of the Tar Springs is not as deep, but it is more widespread and is overlain by a greater volume of quartz sandstone. Because the Hardinsburg boundary has greater depth of incision and likely coincides with a major biostratigraphic zone boundary, it is picked as the composite sequence boundary and the composite-sequence thus includes sequences 5 through 8 (Fig. 9). The lowstand/transgressive sequence sets cannot be differentiated and are composed of sequences 5, 6 and the TST of sequence 7. Sequences within the lowstand/transgressive sequence set have incised-valleys at the base and onlap the Southeastern and Western Shelves. They are primarily composed of transgressive-regressive parasequences that are dominated by high-energy

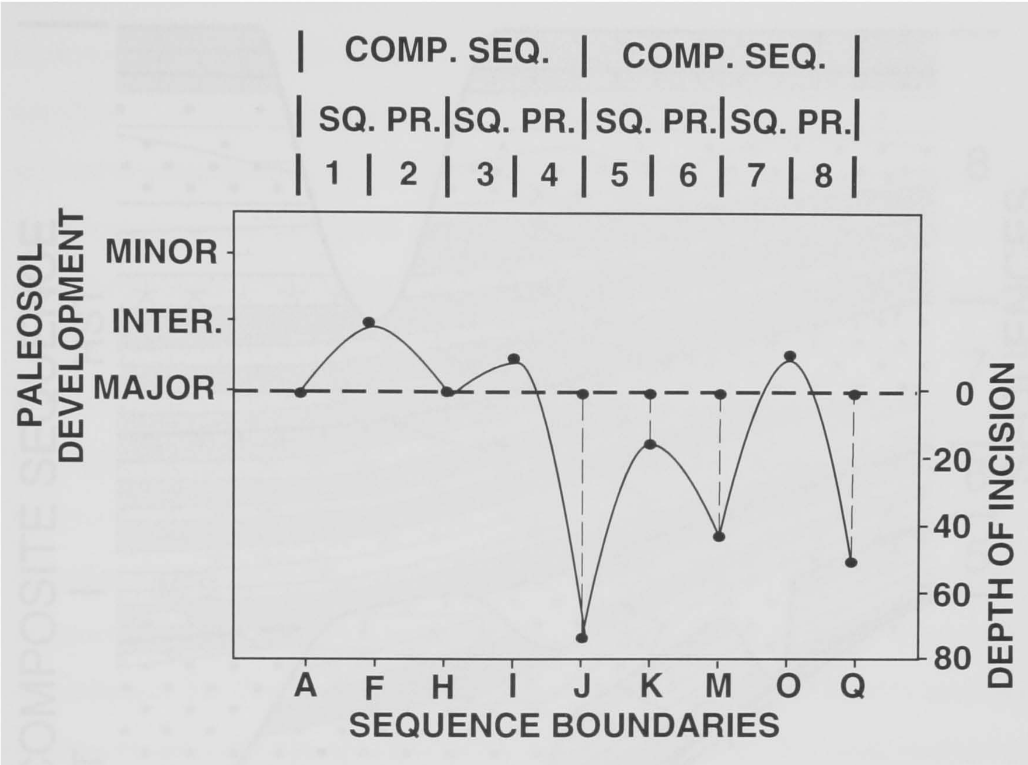


Fig. 9. Relationship between sequences, sequence pairs and composite sequences for Sequences 1 through 8. Sequences 1 through 4 are discussed in Chapter 2, and sequences 5 through 8 are discussed in this paper. Sequence pairs are bounded by major disconformities. Composite sequences are bounded by major disconformities that coincide with major biostratigraphic zone boundaries and show maximum incision.

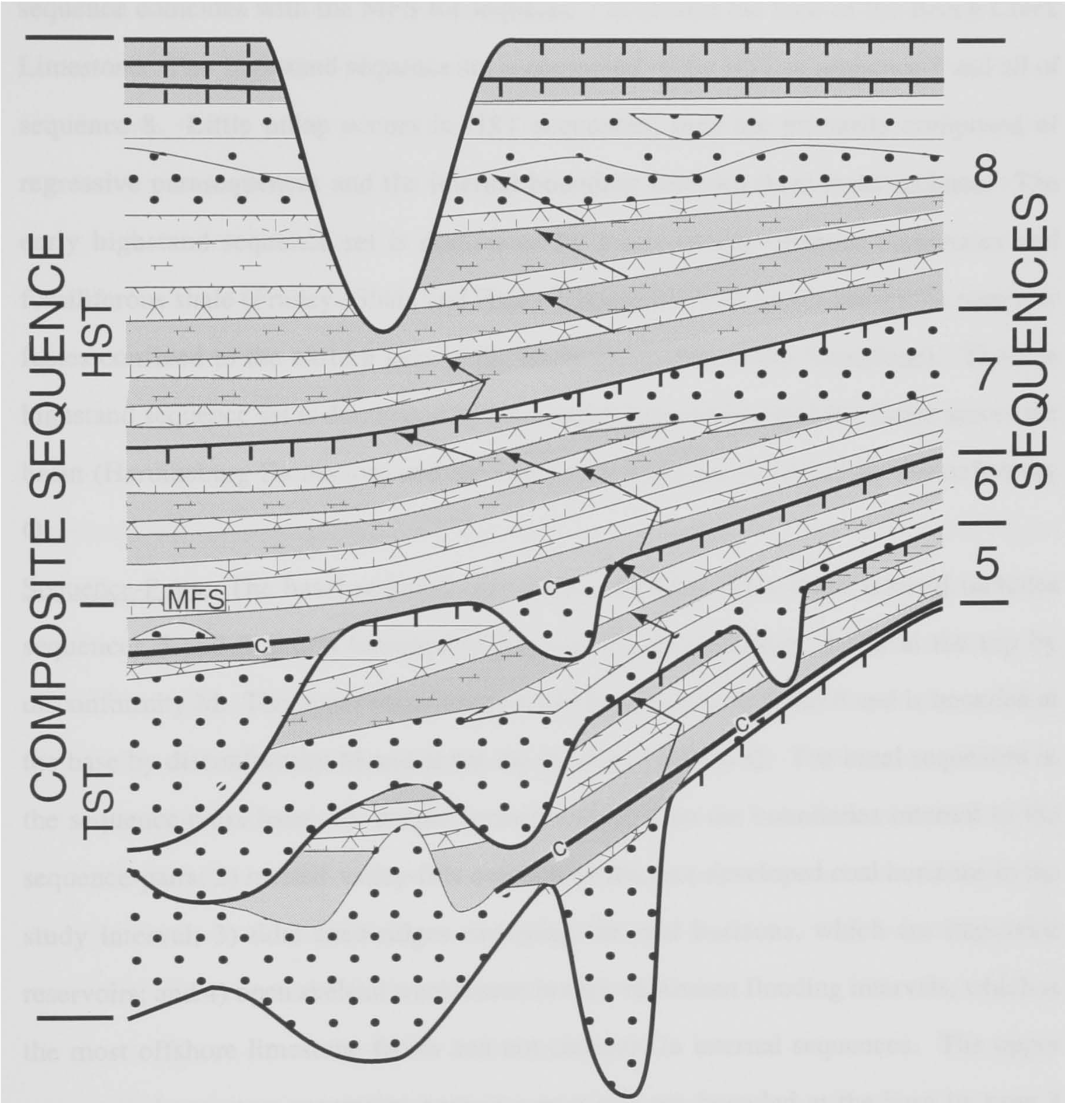


FIG. 10. Schematic cross-section of composite sequence composed of sequences 5 through 8. Arrows denote transgressive-regressive and regressive parasequences. See Fig. 4 for legend.

quartz sandstone reservoir facies. The maximum flooding surface for the composite-sequence coincides with the MFS for sequence 7 at or near the base of the Beech Creek Limestone. The highstand sequence set is composed of the HST of sequence 7 and all of sequence 8. Little onlap occurs in HST sequences, they are primarily composed of regressive parasequences and the internal boundary between them lacks incision. The early highstand sequence set is dominated by lower-energy offshore carbonates and fossiliferous shale (Fraileys Shale and Haney Limestone) with quartz sandstone reservoir facies confined to the eastern side of the basin updip (Big Clifty Sandstone). The late highstand sequence set is dominated by low-energy heterolithic tidal flat facies across the basin (Hardinsburg Formation) and the composite sequence is capped by disconformity Q.

Sequence-Pairs. The basal sequence-pair in the Bethel to Glen Dean interval includes sequences 5 and 6 and is bounded at the base by disconformity J and at the top by disconformity M. The upper sequence-pair includes sequences 7 and 8 and is bounded at the base by disconformity M and at the top by disconformity Q. The basal sequences in the sequence-pairs have: 1) deeper incised-valleys than the boundaries internal to the sequence-pairs; 2) incised-valley-fills overlain by the best-developed coal horizons in the study interval; 3) tidal sand-ridges overlying the coal horizons, which are important reservoirs; and 4) open skeletal wackestone in their maximum flooding intervals, which is the most offshore limestone facies and not common in internal sequences. The upper sequences have more regressive parasequences and are bounded at the base by type 2 sequence boundaries or shallower incised valleys.

DISCUSSION

Duration of Composite-Sequences, Sequences and Parasequences

The best ages for the Mississippian come from SHRIMP dating of volcanics interbedded with marine sediments in Australia (Roberts et al., 1995). Roberts et al.

(1995) used brachiopod-biostratigraphy to correlate with European and North American conodont- and foramaniferal- zones, and tied the SHRIMP ages to radiometric dates from Europe. Based on foramaniferal- and conodont-zones, the Ste. Genevieve to Glen Dean interval is uppermost Visean (middle of V3b to V3c) (Baxter and Brenckle, 1982). This interval has a duration of roughly 3 m.y. (Roberts et al., 1995), but is poorly constrained due to the difficulties inherent in correlating between brachiopod-, conodont-, and foramaniferal-zones on different continents and the error bars on the radiometric dates.

There are four disconformity-bounded sequences in the Ste. Genevieve to Paoli interval (Chapter 2) and four in the overlying Bethel to Glen Dean interval. Eight sequences divided into 3 million years gives an average period of 375 k.y. for the sequences which roughly coincides with the long-term Milankovitch eccentricity signal (~414 k.y.) (Berger, 1988). Dominance of 4th-order sequences is common in times of abundant global ice such as the Pennsylvanian and Pleistocene (Heckel, 1986; Weber et al., 1995). If the ~400 k.y. duration for the sequences is correct, the sequence-pairs formed in ~800 k.y. and the composite-sequences formed in ~1.6 m.y. Both the composite sequence and sequence pairs would be in the range of 3rd-order units (0.5-5 m.y.) (Weber et al., 1995). There are 2 to 7 parasequences in each sequence with average durations between 57 and 200 k.y. which makes them 4th-order (100-500 k.y.) and 5th-order (10-100 k.y.) units (c.f. Weber et al., 1995).

Ramp Slope

The cross-sections and other published work suggest that the Illinois Basin was a gently sloping ramp throughout the Bethel to Glen Dean interval (Treworgy, 1985, 1988). The steepest slope on the ramp was between the Southeastern Shelf and the Basin Interior during deposition of sequence 6 and the TST of sequence 7, which are absent on the Southeastern Shelf but up to 32 meters thick in the Basin Interior (location 20). If this total thickness difference is divided by the 4 parasequences in this interval, the thickness difference per parasequence is 8 meters. Division of the average difference in thickness per parasequence (8 meters) by the distance between the shelf and basin margin (110 kilometers) gives an average slope of 7 cm/km (0.004°). The ramp had its lowest slope during deposition of the HST of sequence 7 when it was less than 1 cm/km. The gentle ramp slope caused major shifts in facies with minor relative sea-level change.

Tidal Energy

Chesterian siliciclastics were initially interpreted to be fluvial, deltaic and shallow marine facies that were deposited in a fluvially-dominated bird-foot delta setting analogous to the modern Mississippi Delta (Swann, 1964; Potter, 1963; Seyler, 1982). However, recent studies have re-interpreted most of the siliciclastics as tide-dominated deltaic and shallow marine facies (Visher, 1980; Treworgy, 1985; Specht, 1985; Cole and Nelson, 1995). Tidal sand-ridges occur in macrotidal estuaries and shallow shelf settings such as the North Sea and the Bay of Fundy (Off, 1963; Hayes, 1975; Dalrymple et al., 1990). Marine sand-filled channels are common in tide-dominated deltas, but rare in river-dominated deltas (Coleman, 1976). Lenticular sand-bodies within heterolithic facies near the top of several siliciclastic formations are interpreted to be tidal channel deposits; analogous facies are common in tide-dominated delta successions (Coleman, 1976). Modern tidal rhythmites are only found in meso- to macrotidal intertidal and subtidal settings (Kvale, 1996). Modern siliciclastic tidal flats commonly occur in tide-dominated deltas and estuaries where there is mixing of sandy offshore facies and muddy salt-marsh or delta-plain facies. The presence of these tidally-influenced features suggests that the Illinois basin was at least meso-tidal for much of the Middle Chesterian. Tidal currents probably were strongest in incised-valleys and in areas of sand-ridge and swale topography. The strong tidal currents formed, and were intensified by, the sand-ridges which focused tidal flow into intervening swales and channels. The great width of the ramp (>400 km) and the funnel-like bathymetry of the Illinois Basin likely served to increase the tidal range and bottom current velocity in the Late Mississippian (c.f. Klein, 1977).

Minor Middle Chesterian ooid grainstone and bi-directional cross-bedding in some skeletal grainstone units updip on the ramp suggest that some carbonates were also tidally-influenced. However, fossiliferous shale, skeletal wackestone and some of the skeletal packstone/grainstone units in the Basin Interior have graded beds caused by storms rather than tides (Treworgy, 1985; Harris, 1992). This suggests that most of the carbonate units were deposited in deeper-water, more offshore settings where tidal currents were diminished (Fig. 5).

Climatic Influence

Chesterian climate was dominated by increasingly humid, wet-dry seasonality. The Late Middle Mississippian St. Louis Formation has widespread evaporites which suggest an arid climate (Shaver et al., 1986). The Early Chesterian Ste. Genevieve Formation has widespread caliche and breccia, which are indicative of wet-dry seasonal climate and abundant eolianites, ooid grainstone and dolomite which suggest a dominantly dry climate. However, there are no evaporites in the Ste. Genevieve which suggests that the climate was wetter than it was during deposition of the underlying St Louis Formation. In the Bethel to Glen Dean study interval, slickensided mudrock paleosols, caliche and breccia also suggest wet-dry seasonality, but increased siliciclastic influx, decreased ooid production and the presence of thin coals and minor karst suggest that it was wetter than the underlying Ste. Genevieve (Cecil, 1990). However, the lack of thick coals or plant fossils indicates that the climate was not as humid as later in the Pennsylvanian.

Swann (1964) suggested that the alternation of carbonate and siliciclastic units was due to cyclic climate change under relatively static sea level. In his model, siliciclastics were carried form the source area during arid periods and carbonates were deposited during humid periods when dense vegetation prevented erosion upstream. In contrast, Droste and Horowitz (1990) suggested that siliciclastic influx occurred during humid phases and carbonate deposition occurred during more arid periods. Cecil (1990) suggested that a combination of climate and relative sea-level changes produced the Pennsylvanian cyclothems which overlie the studied interval. In the Pennsylvanian,. carbonate deposition occurred during arid and semi-arid periods, maximum siliciclastic influx occurred during periods of wet-dry seasonality, and peat deposition occurred during the most humid phases (Cecil, 1990). Based on the sedimentological paleoclimatic indicators in the study interval, Late Mississippian climate in the Illinois Basin did not undergo such radical shifts and remained seasonally wet-dry throughout. Subtle variations in the duration of the wet season may have caused greater siliciclastic influx in some of the siliciclastic units, but a model of cyclic climate change alone cannot explain the basin-wide extent of the paleosols, the depth of incision in many of the incised-valleys or the extent of marine inundation required to deposit deeper water carbonates on the shelves.

Wet-dry seasonality was likely caused by monsoonal circulation which arose as Pangea was assembled on the equator (Rowley et al., 1985; Parrish et al., 1986; Witzke, 1990). The intra-tropical convergence zone of low pressure and high rainfall shifted far to the north during northern hemisphere summer and far to the south during southern hemisphere summer, leaving the equatorial region dry most of the year (Rowley et al., 1985; Parrish et al., 1986). This circulation pattern likely was best developed during pre-Chesterian deposition of the St. Louis Formations and progressively decreased through the Late Mississippian as global cooling started. Onset of glaciation in the southern hemisphere likely contributed to the progressively more humid climate in the Late Mississippian as buildup of large ice-sheets on Gondwana prevented the intra-tropical convergence zone from shifting as far to the south as it did in the Middle Mississippian.

Although continental collision to the east was in its early stages, uplift of the equatorial Appalachian Plateau may have created a zone of low pressure which locked the intra-tropical convergence zone into the low latitudes for progressively longer periods (Rowley et al., 1985).

Tectonics

Evidence for Tectonic Activity.- Subtle variations in differential subsidence between the Basin Interior and the shelves, apparent syn-sedimentary faulting in some units, and the presence of seismically disturbed bedding, or "seismites," suggest that the Illinois Basin was tectonically active during the Late Mississippian. Thickness ratios from the Western Shelf: Basin Interior: Southeastern Shelf are 1.5:10:1 for the interval spanning sequences 5, 6 and the TST of 7; 1:1:1 for the HST of sequence 7; and 1.4:3:1 for sequence 8. Neglecting compaction effects, thickness variations are proxies for variations in differential subsidence between the Basin Interior and the shelves. Uplift on the shelves and/or downwarping in the basin caused relatively high differential subsidence (>5:1) during deposition of sequences 5, 6 and the TST of 7. This was followed by a period of low differential subsidence in the HST of sequence 7, and a subsequent increase to moderate differential subsidence in sequence 8. The Eastern and Western Shelves subsided differently relative to each other; sequence 5 is present on the Southeastern Shelf but absent on the Western Shelf and sequences 6 and the TST of 7 are present on the Western Shelf but absent on the Southeastern Shelf.

Rapid subsidence probably was accompanied by normal faulting in the Basin Interior during deposition of sequences 5, 6 and the TST of 7, which are amalgamated in the West Baden Clastic Belt. This belt directly overlies the northeast-southwest trending Dixon Springs Graben, which is part of the Fluorspar Area Fault Complex and extends over part of the Wabash Valley Fault Zone (Cole and Nelson, 1995). Low differential subsidence appears to have been accompanied by block-faulting in the Fluorspar Area

Fault Complex in the HST of sequence 7 and at the base of sequence 8 (Fig. 4). Uplift of a horst in the middle of the HST of sequence 7 is indicated by the absence of parasequences 34 and 35 at location 20. A second up-thrown fault block occurs at the base of sequence 9 at location 11 where only two of the four parasequences in the Haney Limestone are preserved (Fig. 4). Renewed subsidence in the Basin Interior was accompanied by normal faulting in the late HST of sequence 8. The Hardinsburg Sandstone thickens by more than 25 meters over less than 2 km in the area of the Dixon Springs Graben without complementary thinning of the underlying Haney Limestone, which suggests syn-sedimentary normal faulting during Hardinsburg time (Nelson, 1996). It is likely that the Late Mississippian faulting was along older, reactivated northeast-southwest trending basement faults (Nelson, 1995).

This syn-sedimentary faulting generated seismites when earthquakes caused sediment liquefaction or dewatering, which distorted original bedding and lamination in strata most likely within 15 meters of the sediment-water interface (Kuenen, 1958; Seilacher, 1984; Scott and Price, 1988). Seismites in the study interval consist of ball and pillow structures and contorted bedding in the heterolithic facies. A seismite horizon in the uppermost Cypress Formation can be traced between locations 11 and 12 in the Fluorspar District and could be related to faulting and major thickening of the section at location 12 (Fig. 4). A second seismite horizon at the top of the Big Clifty Sandstone can be traced in southern Indiana between six locations over a distance of 75 kilometers (location 48 to location 67, Fig. 4). There is also a seismite horizon in the Hardinsburg Formation that can be traced between locations 19 and 20. Earthquakes with magnitudes between 6 and 8 probably would have been required to produce these liquefaction features (Kuribayashi and Tatsuoka, 1975)

Tectonic Control on Clastic Input.- During periods of high and moderate differential subsidence, the thickest accumulations of sandstone appear to occur in northeast-southwest trending belts, which parallel and overlie the Wabash Valley Fault System and the Fluorspar Area Fault Complex (Cole and Nelson, 1995; Droste and Horowitz, 1990). During West Baden Clastic Belt deposition (sequences 5, 6 and TST of 7), high differential subsidence may have locked the tidal delta into position in the northeastern part of the basin and funneled quartz sandstone into tectonic lows in the area of the fault zones (Droste and Horowitz, 1990). Low differential subsidence occurred during deposition of the HST of sequence 7 and possibly the HST of sequence 8 as the Southeastern Shelf, which was emergent throughout sequences 6 and the early TST of 7, began to subside at a rate similar to that of the Basin Interior. This renewed subsidence on the Eastern Shelf, along with possible elevation of fault-bounded blocks in the Basin Interior, caused the tide-dominated delta to switch to the east and quartz sandstone deposition was concentrated on the eastern side of the basin (Big Clifty Sandstone). Differential subsidence increased in the late highstand of sequence 8, when normal faulting resumed in the Basin Interior and the tidal delta was again locked into the northeastern part of the basin during Hardinsburg Sandstone deposition (Droste and Horowitz, 1990).

<u>Tectonics and Composite-Sequence Development.-</u> Phases of tectonic activity and quiescence occurred at the same frequency as the composite sequence (~1.6 m.y.), but were out of phase with composite sequence development. Periods of high differential subsidence coincided with the deepest incision events at the composite-sequence and sequence-pair boundaries, and the single type 2 sequence boundary formed during a period of low differential subsidence. If tectonics was driving composite sequence development, low differential subsidence should have been associated with incised sequence boundaries. Tectonic activity apparently controlled the location of the incised-valleys, and may have accentuated local incision on basin-wide disconformities, but could not have formed the disconformities. Incised-valley formation in the Basin Interior

during periods of high-differential subsidence would require relative sea level to fall farther than during periods of lower differential subsidence.

Tectonics and Sequence and Parasequence Development.- During periods of high differential subsidence, sequences and parasequences onlapped and progradationally offlapped the shelves. For example, sequence 5 onlaps the Western Shelf because it was uplifted or resisting subsidence relative to the rest of the basin. Sequence 6 and parasequence 31 at the base of sequence 7 onlap the Southeastern Shelf because it was uplifted or resisting subsidence at that time. The basal siliciclastic unit of sequence 5 onlaps the Southeastern Shelf and the siliciclastic unit at the top of sequence 5 progradationally offlaps the Southeastern Shelf during moderate differential subsidence between the Southeastern and Northeastern shelves. Minimal onlap occurs at the bases of sequences 8 and 9, which were deposited during periods of low differential subsidence.

Tectonic Linkage Between Illinois and Appalachian Basins.- Tectonic movement on the Southeastern Shelf was likely related to activity on the northeast-southwest trending part of the Cincinnati Arch (Fig. 1 and location map, Fig. 4). During deposition of the Bethel to Glen Dean interval, the Southeastern Shelf subsided at a slower rate than the Northeastern Shelf which is opposite to the subsidence pattern in the underlying Ste. Genevieve to Paoli interval (Chapter 2). Chesterian sections on the Southeastern Shelf and western Appalachian Basin are similar in thickness and sequence development from sequence 3 to the base of the Big Clifty Sandstone in sequence 7 (Smith et al., 1995). In contrast, the overlying Big Clifty Sandstone is up to 20 meters thick on the Southeastern Shelf (Fig. 4) and absent in the Appalachian Basin. Thus, the Southeastern Shelf of the Illinois Basin and the western Appalachian Basin likely subsided synchronously until the deposition of the Big Clifty Sandstone, when the Southeastern Shelf started to subside more rapidly. There must have been at least 20 meters of relief on the Cincinnati Arch at this time, which prevented siliciclastic influx into the Appalachian Basin. The overlying

Haney and Hardinsburg Formations can be correlated across the Cincinnati Arch, suggesting re-establishment of the link between the Southeastern Shelf and the western Appalachian Basin over the Cincinnati Arch. The laterally extensive seismite at the top of the Big Clifty Sandstone may be related to this tectonic adjustment.

Variations in the rates of subsidence in the Illinois Basin probably were related to phases of thrust-loading in the Appalachian foreland basin. Backstripping of time-equivalent units in southwest Virginia reveals that relatively high subsidence rates associated with thrust-loading commenced during early Ste. Genevieve time and continued through the study interval (Al-Tawil. pers. comm., 1995). Appalachian thrust-loading likely caused the Cincinnati Arch to resist subsidence (Quinlan and Beaumont, 1984) while the horizontal stress associated with continental collision may have caused greater subsidence rates in the Illinois Basin interior (DeRito et al., 1983). Uplift on the northeast-southwest trending Cincinnati Arch and greater subsidence in the Basin Interior likely initiated reactivation of normal faults in the northeast-southwest trending Wabash Valley Fault System and Fluorspar Area Fault Complex (Fig. 1B). Phases of high differential subsidence would have occurred as a response to Appalachian thrusting and lower differential subsidence would have occurred during periods of relative quiescence. Variations in tectonic subsidence changed the ramp slope by a few cm/km at most, but it was sufficient to focus deposition of siliciclastics whose loading effects accentuated this differential subsidence.

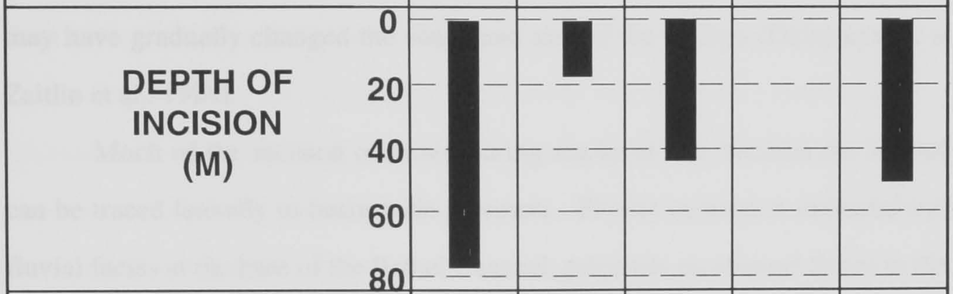
<u>Eustasy and Sequence/Parasequence Development</u>

The regional and interbasinal extent of the disconformities and their development during periods of high, moderate and low differential subsidence indicate that glacio-eustatic sea-level changes produced the sequences. These sea-level changes would have been relatively large to cause incision of up to 75 meters followed by deposition of deeper water, offshore wackestone on the previously emergent shelves. A glacio-eustatic origin is supported by evidence of glacial deposits on several Gondwana continents that have been dated as earliest Namurian which occurs immediately above the study interval (Roberts et al., 1995; Frakes et al., 1992). Similar mixed carbonate-clastic, disconformity-bounded sequences with incised-valleys occur in time-equivalent strata in Great Britain (Walkden, 1987), but these have not been correlated with their North American counterparts at the precision necessary to prove global eustatic control.

Magnitude of 4th-Order Sea-Level Changes.- The depth of incision associated with each sequence boundary is the only evidence available to determine the magnitude of the sea-level falls (Fig. 11). To do this, it is first necessary to distinguish lowstand incised-valley systems from mid-highstand erosion by prograding tide-dominated deltas. Highstand submarine erosion occurred at the base of the Big Clifty Sandstone in sequence 7, when strong tidal currents caused incision into underlying shale and limestone during tide-dominated delta progradation over the Eastern Shelf (Visher, 1980). These tidal channels cannot be traced laterally into paleosols and therefore developed in a subaqueous environment.

Incised valleys cut during sea-level lowstand are defined as fluvially-eroded, elongate topographic lows typically larger than a single channel form and characterized by an abrupt seaward shift in facies across a regionally mappable sequence boundary at its base (Zaitlin et al., 1994). Their fill typically begins to accumulate during the subsequent base-level rise (Zaitlin et al., 1994). The Late Mississippian lowstand channels and incised-valleys can be traced laterally to paleosols that formed on highs where no erosion occurred. As sea-level fell, fluvial channels likely occupied earlier shallow tidal channels on the prograded delta plain. These fluvial channels amalgamated and migrated laterally to form incised-valleys as they continued to cut farther into underlying strata. Paleosols at the bases of channels in the Sample Formation probably formed as these channels were abandoned during sea-level fall. Channel fills commonly

DISCONFORMITY	J	K	M	O	Q
SEQUENCE NO.	5	6	7	8	9



ESTIMATED SEA-LEVEL FALL (M)	75	15	45	10	50
ESTIMATED SEA-LEVEL RISE (M)	20	15	20	15	20
ESTIMATED SEA-LEVEL CHANGE (M)	95	30	65	25	70
BIOSTRATIGRAPHIC BOUNDARY?	NO	NO	YES	NO	YES
COMPOSITE SEQUENCE (CS) OR SEQUENCE PAIR (SP) BOUNDARY	CS		SP		CS

Fig. 11. Estimation of the magnitude of sea-level changes which produced 4th-order sequences 5 through 9, Bethel to Glen Dean interval.

show a deepening-upward facies trend as dominantly tidally-influenced quartz sandstone grades up into offshore carbonates and fossiliferous shale. The lack of fluvial facies in most valley-fills suggests that any channel-floor fluvial facies were reworked by strong tidal currents during transgression in tide-dominated estuaries, where strong tidal currents may have gradually changed the shape and size of the valleys (Dalrymple et al., 1990; Zaitlin et al., 1994).

Much of the incision occurred during sea-level fall, because the incised-valleys can be traced laterally to basin-wide paleosols. Fluvial incision is indicated by possible fluvial facies at the base of the Bethel Channel, paleosols on channel floors in the Sample Formation, and the narrow width of deep channels at the bases of sequences 5 and 9. Such narrow and deep incised-valleys are typical of updip areas, in contrast to downdip areas near base-level where valleys become wider (Schumm and Etheridge, 1994). The deepest incised-valleys in the Bethel and Hardinsburg Formations are less than one or two kilometers wide which suggests that they developed far updip under subaerial rather than submarine conditions (Fig. 8). However, linkage of the depth of the incised-valleys to the magnitude of sea-level fall is complicated by the possibility that some erosion may have occurred by fluvial or tidal downcutting below sea level. It is possible that strong tidal currents could have further eroded the channels during lowstand or early transgression. If sub-sea erosion was a major factor in some of the channel formation, then the estimates for sea-level fall may be too high. However, any component of sub-sea-level erosion may be balanced by the subsequent effect of compaction which reduced the thickness of the valley fills and adjacent sediments and caused underestimation of incision depth.

Subsequent sea-level rises can be estimated by the overlying limestone facies deposited on the shelves. A tidal sand-ridge up to 16 meters thick was deposited in the early TST of sequence 5 prior to carbonate deposition in the maximum flooding interval.

This suggests water depths of 15 meters for some skeletal grainstone/packstone and up to 20 meters for storm-influenced skeletal wackestone. Confirmation of these water depth estimates comes from sequence 7, where skeletal wackestone and grainstone/packstone are overlain by 16 meter thick tidal sand-ridges of the Big Clifty Sandstone which formed during progradation and sea-level fall (Specht, 1985).

Estimates for the magnitude of 4th-order sea-level fluctuations for sequences 5 through 9 are calculated in Figure 11. If the incised-valleys can be used as a measure of sea-level fall, and the limestones can be used to estimate water depth during maximum flooding, then 4th-order sea-level changes ranged from 95 meters in Sequence 5 to 30 meters in sequence 6, to 65 meters in sequence 7, to 25 meters for sequence 8 and 70 meters for sequence 9.

Comparison of Sequences with Pennsylvanian Cyclothems.- The HST of sequence 7 and the TST of sequence 8 resemble sequences 3 and 4 which likely were formed by low to moderate amplitude sea-level fluctuations (Chapter 2). These sequences are characterized by disconformities with little incision and they contain multiple regressive limestone-shale parasequences. Whereas minor transgressive deposits are preserved, they are not as thick as transgressive units in the sequences formed by higher-amplitude sea-level fluctuations.

In contrast, sequences 5 and 7 were formed by moderate to high amplitude sea-level fluctuations and resemble typical Pennsylvanian cyclothems (Fig. 12). Thick transgressive deposits are preserved at the bases of both the high-amplitude sequences and the Pennsylvanian cyclothems. Both these sequences and typical cyclothems have deep incised-valleys at the base filled with quartz sandstone. Both have coal beds overlying incised-valley-fills, which are overlain by marine siliciclastics and symmetrical carbonate units. Pennsylvanian cyclothems commonly have deeper-water black shale bounded by two carbonate units, whereas the Chesterian examples have

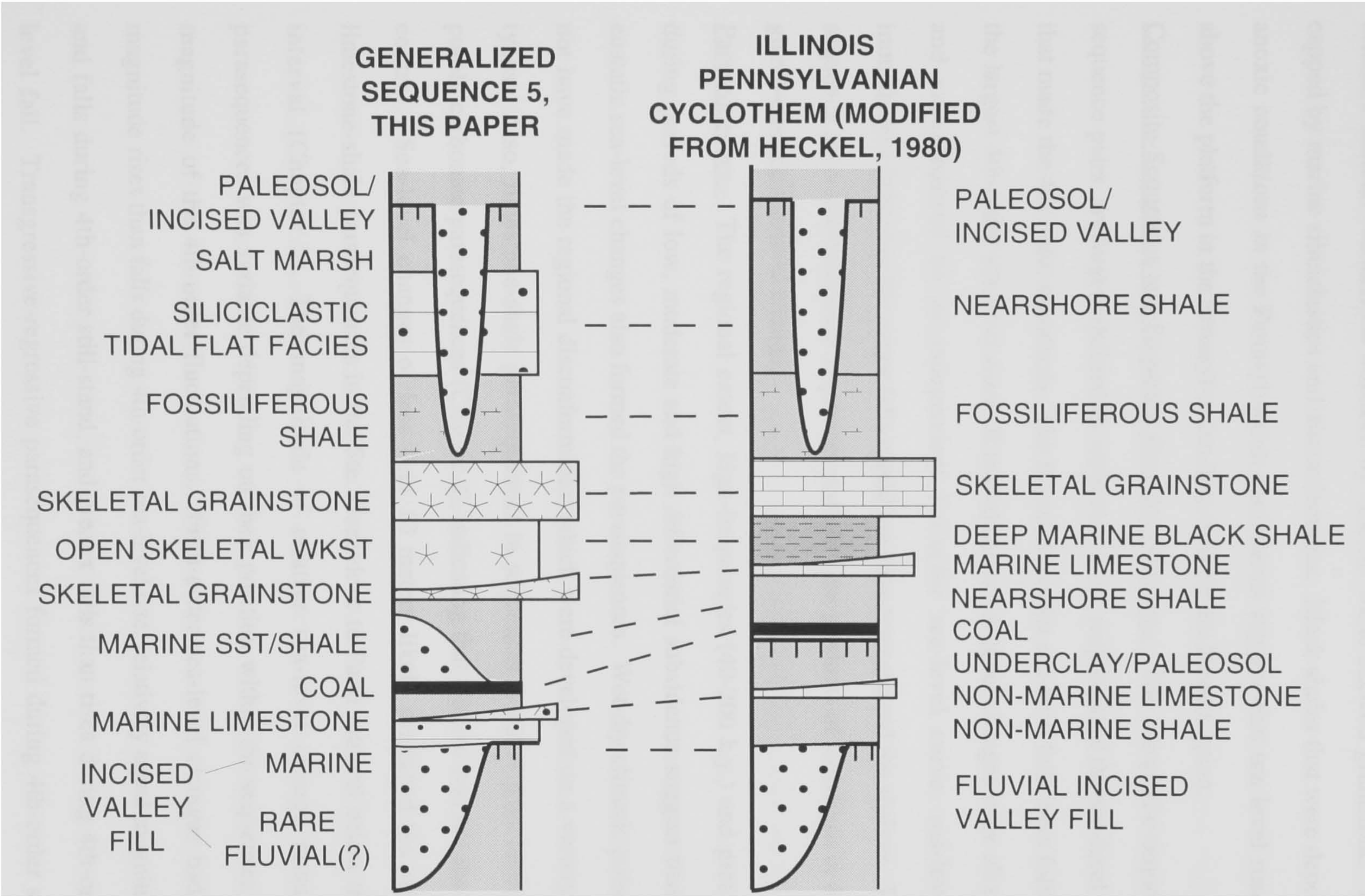


Fig. 12. Comparison of sequence 5 from this study with Pennsylvanian "complete Illinois cyclothem" (modified from Heckel, 1980).

offshore skeletal wackestone bounded by shallow water skeletal grainstone. Both are capped by marine siliciclastics and disconformities. Black shales that were deposited in anoxic conditions in the Pennsylvanian cyclothems suggest that sea level rose higher above the platform in the Pennsylvanian than it did in the Mississippian.

Composite-Sequences and Sequence Pair Development.- The composite-sequence and sequence-pairs are largely defined by variations in the amplitude of the sea-level changes that made the 4th-order sequences. The largest 4th-order sea-level falls were followed by the largest 4th-order sea-level rises. If the 4th-order sea-level changes were all the same and superimposed on an independent 3rd-order sea-level curve, sea-level rises immediately following the major falls would not have transgressed the shelves. Thus, the sequence pairs and composite sequences probably are products of variations in 4th-order glacio-eustatic sea-level change.

Parasequences.- The regional extent, high-frequencies (40-200 k.y.) and preservation during periods of low, moderate and high differential subsidence suggest that glacio-eustatic sea-level changes also formed the parasequences. Wet-dry climatic pulses could not have made the regional disconformities, which were developed on a variety of rock types. Also, limestone-shale parasequences in sequences 3 and 4 pass laterally into purely carbonate parasequences (Chapter 2), indicating that climate was not the primary control. Sea-level changes of less than 10 meters likely produced the regressive limestone-shale parasequences in the Ste. Genevieve to Paoli interval below the study interval (Chapter 2). The magnitude of relative sea-level change forming the parasequences was variable depending on their position within the sequences and the magnitude of the 4th-order fluctuations. Fifth-order sea-level changes had greater magnitude rises than falls during 4th-order sea-level rise, relatively equal magnitude rises and falls during 4th-order still-stand, and greater falls than rises during 4th-order sea-level fall. Transgressive-regressive parasequences formed during 4th-order sea-level rises and regressive limestone-shale parasequences formed in the early HST of the sequences during 4th-order still-stand.

Transgressive-regressive parasequences occur at the base of each of the sequences, but are best developed within the transgressive sequence tract of the composite-sequence. Preservation of transgressive units is most common in the TSTs of 4th-order sequences because incised-valleys provided accommodation space, and quartz sand was available for reworking, and the duration and magnitude of the 5th-order sea-level rises was amplified by superimposed 4th-order sea-level rises. Transgressive-regressive parasequences that formed in the early TST at the bases of sequences 5 and 7 were initiated as sea-level rose, and quartz sand that had been bypassing the shelf during lowstand was reworked and deposited in tidally-influenced estuarine/marine settings in incised-valleys. As valleys filled, any remaining quartz sand was shaped into tidal sand-ridges by strong tidal currents (c.f. Nio and Yang, 1991); as sea level reached its highstand, siliciclastic sediment supply decreased and fossiliferous shale, minor limestones and heterolithic facies were deposited followed by swamp deposition and disconformity development as sea level fell. In contrast, transgressive-regressive parasequences formed during the late TST/early HST do not fill deeply incised-valleys. These parasequences overlie early TST parasequences at the base of sequences 5 and 7 and occur at the bases of sequences 6 and 8. They were initiated as sea level started to rise, and heterolithic facies, shale and tidal sand-ridges were deposited, especially in tectonic lows. With continued transgression siliciclastics were ponded updip and limestones were deposited. With the largest sea-level rises, symmetrical limestones (basal grainstone, deeper water open skeletal wackestone, upper grainstone) were deposited. Maximum flooding surfaces for each of the sequences and parasequences occur at the base of the deepest water facies in these carbonate units. As sea level started to fall, carbonate units were capped by fossiliferous shale or heterolithic facies and overlain by one or more regressive limestone-shale parasequences.

Asymmetric regressive skeletal limestone-shale parasequences are best developed in the HSTs of sequences in the highstand sequence tract of the composite sequence. They formed when rapid, high-frequency sea-level rises halted siliciclastic influx, allowing carbonate producing organisms to flourish and skeletal limestone to be deposited (Chapter 2). As sea level slowly fell, shale prograded over the basin and capped the parasequences.

Sea-Level Curve for Bethel to Glen Dean Interval.- Incised valleys, paleosols and parasequences in the Bethel to Glen Dean interval can be used to construct a detailed relative sea-level curve (Fig. 13). Ross and Ross (1988) constructed an onlap curve for the Late Paleozoic but only correlated large-scale sequences. Their sequence boundaries occur at the base of sequence 7 and at the base of the Tar Springs which overlies the Glen Dean Formation. The base of the Tar Springs is unconformable, but deeper incised-valleys occur within the Hardinsburg suggesting that it may be a bigger fall. Swann (1963) constructed a transgression-regression, or relative sea-level, curve based on the seaward extent of siliciclastics within the Illinois Basin. Swann's curve roughly parallels the 4th-order sequence curve presented here because siliciclastic units commonly prograded toward the tops of the sequences and were deposited in the early transgressive systems tract at the base of sequences (Fig. 13). Swann also recognized minor transgressions and regressions within the Fraileys, Haney and Glen Dean which correlate to the limestone-shale parasequences. However, Swann (1963) did not pick a regression at the top of the Fraileys/Big Clifty (disconformity O) because the Big Clifty Sandstone did not prograde across the basin and basin-wide paleosols were not used in construction of the curve. The refined sea-level curve shown here could only be developed by recognition of parasequences, paleosols and incised-valleys.

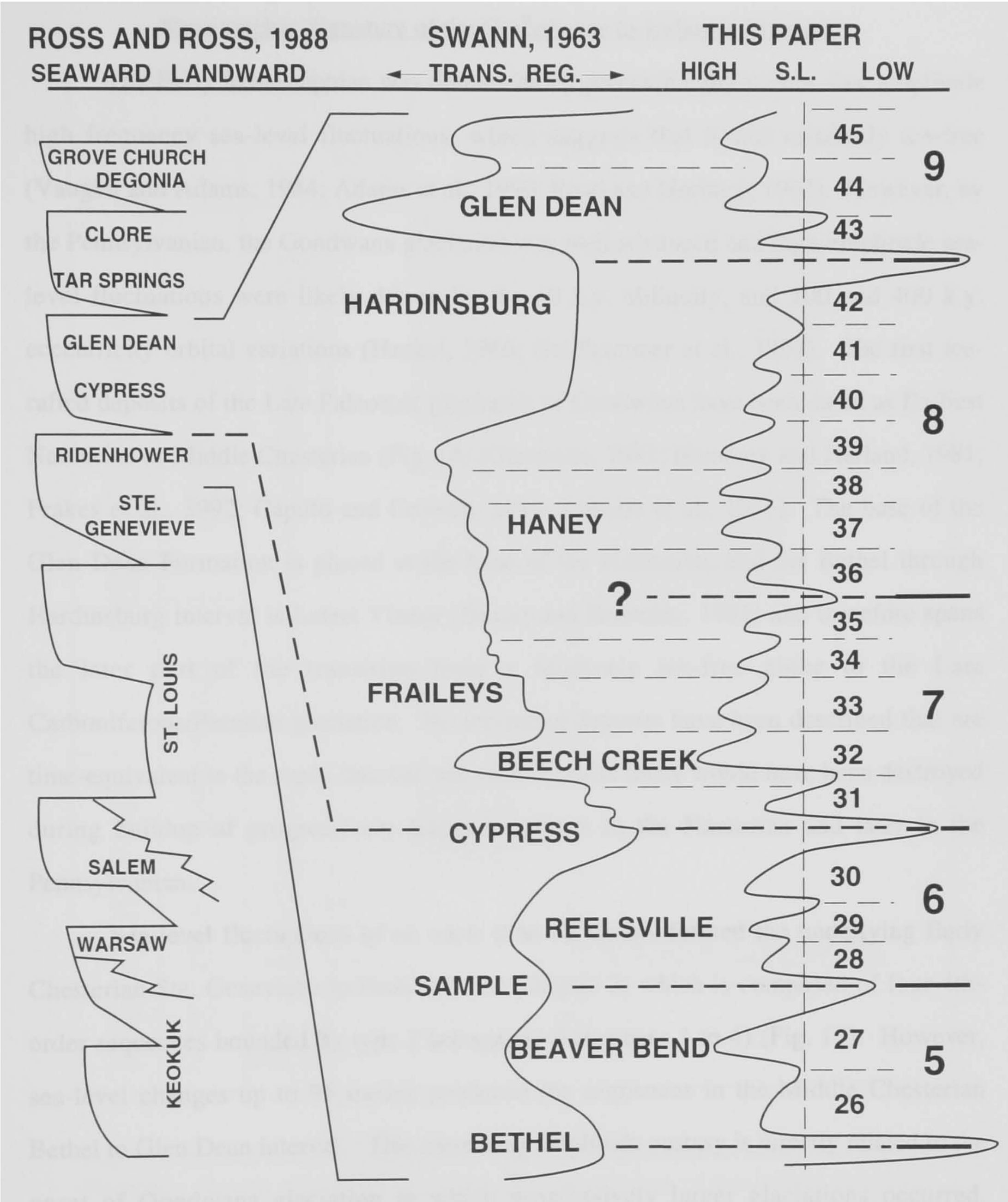


Fig. 13. Comparison of Mississippian onlap curve of Ross and Ross (1988)(left), transgression-regression curve of Swann (1963) (center), and relative sea-level curve prepared for this paper (right) .

Stratigraphic Signature of the Greenhouse to Icehouse Transition

The Early Mississippian was dominated by precessionally-driven, low amplitude high frequency sea-level fluctuations, which suggests that it was relatively ice-free (Vaughn and Adams, 1984; Adams et al., 1990; Read and Horbury, 1993). However, by the Pennsylvanian, the Gondwana glaciation was well advanced and high-amplitude sea-level fluctuations were likely driven by the 40 k.y. obliquity, and 100 and 400 k.y. eccentricity orbital variations (Heckel, 1986; Goldhammer et al., 1991). The first ice-rafted deposits of the Late Paleozoic glaciation of Gondwana have been dated as Earliest Namurian or Middle Chesterian (Fig. 14) (Garrasino, 1981; Hambrey and Harland, 1981; Frakes et al., 1992; Caputo and Crowell, 1985; Roberts et al., 1995). The base of the Glen Dean Formation is placed at the base of the Namurian, and the Bethel through Hardinsburg interval is Latest Visean (Baxter and Brenckle, 1982) and therefore spans the later part of the transition from a relatively ice-free globe to the Late Carboniferous/Permian glaciation. No ice-rafted deposits have been described that are time-equivalent to the study interval, but such deposits likely would have been destroyed during buildup of progressively larger ice-sheets in the Namurian and later in the Pennsylvanian.

Sea-level fluctuations of no more than 30 meters formed the underlying Early Chesterian Ste. Genevieve to Paoli interval (chapter 2) which is composed of four 4th-order sequences bounded by type 2 boundaries (sequences 1 to 4) (Fig. 15). However, sea-level changes up to 95 meters produced the sequences in the Middle Chesterian Bethel to Glen Dean interval. This increasing amplitude eustasy is directly related to the onset of Gondwana glaciation in which progressively larger glaciations occurred. Increasing amplitude 4th-order eustasy may have also affected 5th-order parasequence development, as transgressive-regressive parasequences are best developed at the bases of the sequences that were produced by higher amplitude sea-level changes.

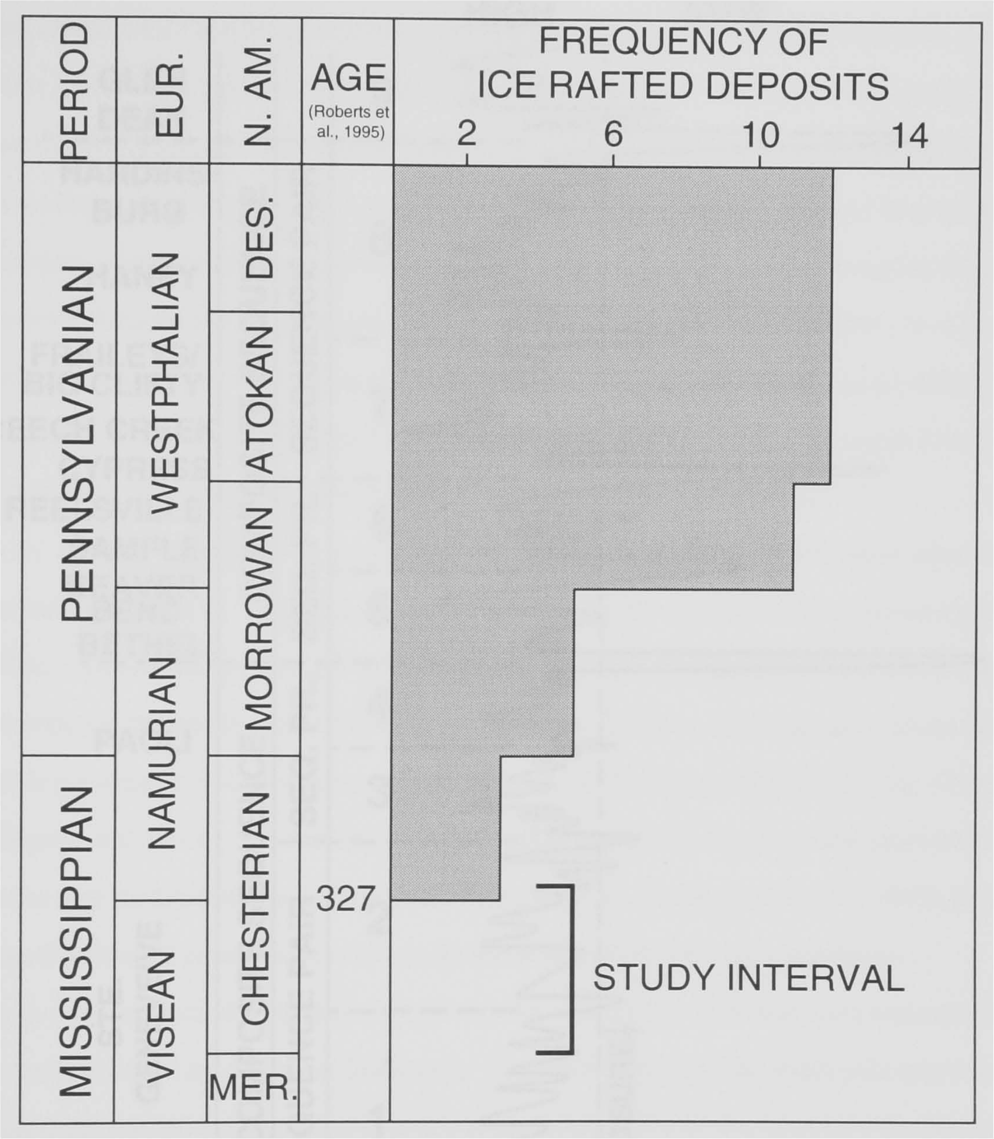


Fig. 14. Global climate setting during Late Mississippian and Pennsylvanian Periods. Graph shows the number of occurrences of ice-rafted deposits world-wide leading up to Late Paleozoic glaciation of Gondwana (modified from Frakes et al., 1992; Age from Roberts et al., 1995). The study interval occurs just below the first recorded glacial deposits.

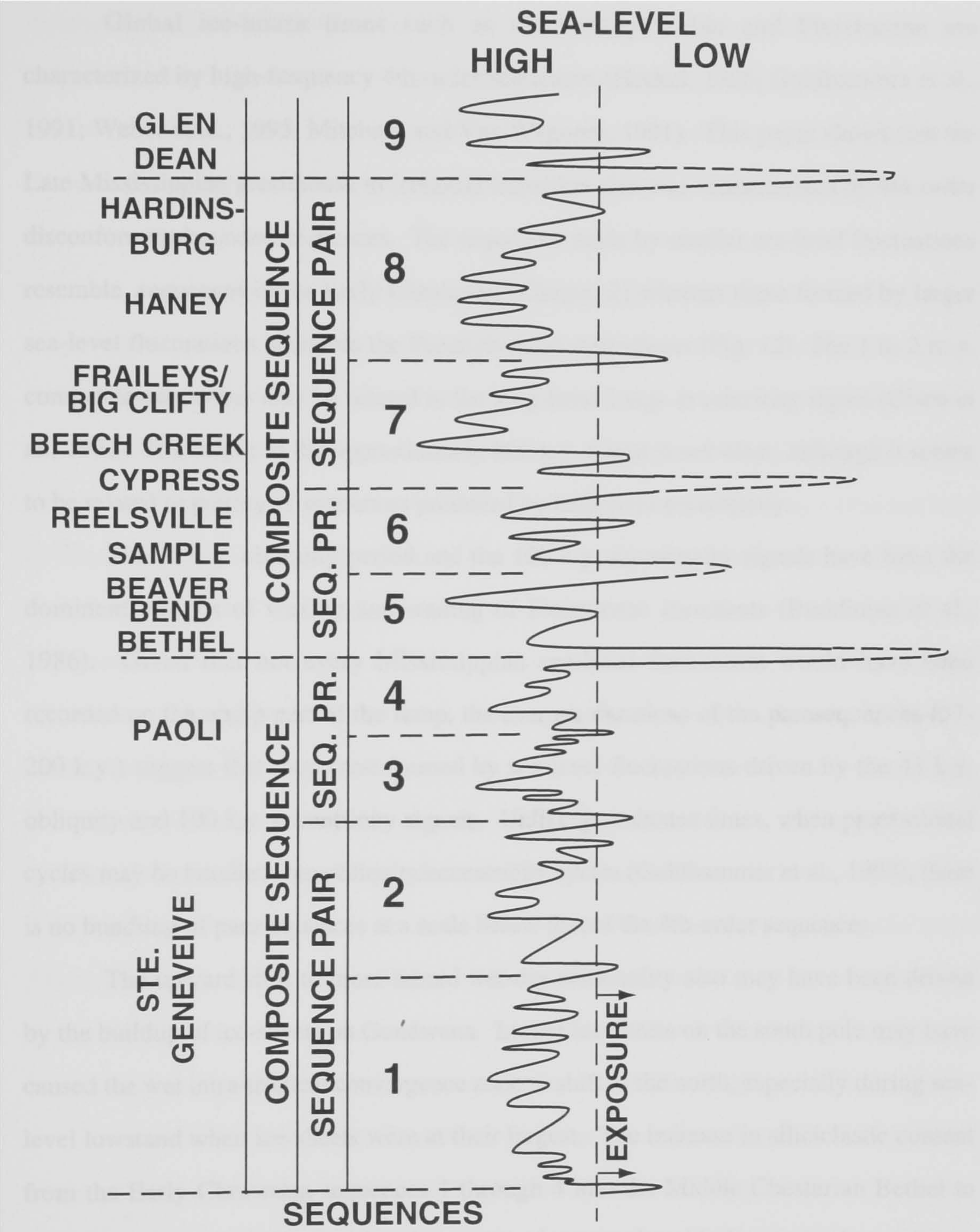


Fig. 15. Composite sea-level curve for Ste. Genevieveto Paoli (sequences 1 to 4 -Chapter 2) and Bethel to Glen Dean interval (sequences 5 to 9 - this chapter). Note suggestion of increasing-amplitude up section..

Global ice-house times such as the Pennsylvanian and Pleistocene are characterized by high-frequency 4th-order sequences (Heckel, 1986; Goldhammer et al., 1991; Weber et al., 1995; Mitchum and Van Wagoner, 1991). This paper shows that the Late Mississippian greenhouse to icehouse transition also was characterized by 4th-order disconformity-bounded sequences. The sequences made by smaller sea-level fluctuations resemble sequences in the Early Chesterian (Chapter 2) whereas those formed by larger sea-level fluctuations resemble the Pennsylvanian cyclothems (Fig. 12). The 1 to 2 m.y. composite-sequences may be related to the long-term 2 m.y. eccentricity signal (Olsen et al., 1993). The origin of the approximately 800 k.y. signal is not clear, although it seems to be related to pairing of sequences produced by long-term eccentricity.

The 41 k.y. obliquity period and the 100 k.y. eccentricity signals have been the dominant periods of waxing and waning of Pleistocene ice-sheets (Ruddiman et al., 1986). Given that not every Mississippian sea-level fluctuation would have been recorded on the updip part of the ramp, the average durations of the parasequences (57-200 k.y.) suggest that they were formed by sea-level fluctuations driven by the 41 k.y. obliquity and 100 k.y. eccentricity signals. Unlike greenhouse times, when precessional cycles may be bundled into obliquity/eccentricity cycles (Goldhammer et al., 1990), there is no bundling of parasequences at a scale below that of the 4th-order sequences.

The upward shift to more humid wet-dry seasonality also may have been driven by the buildup of ice-sheets on Gondwana. Larger ice-sheets on the south pole may have caused the wet intra-tropical convergence zone to shift to the north, especially during sea-level lowstand when ice-sheets were at their largest. The increase in siliciclastic content from the Early Chesterian sequences 1 through 4 into the Middle Chesterian Bethel to Glen Dean interval is likely related to onset of a more humid climate in the region in combination with increasing amplitude eustasy. A similar increase in siliciclastic sedimentation occurred in time-equivalent strata in Great Britain which was attributed to increasing-amplitude eustasy coupled with tectonics (Walkden, 1987). Increasing-magnitude eustasy caused greater base-level falls that in turn caused greater erosion in the headlands and increased siliciclastic sediment supply. At the same time, synchronous climate change and increasing humidity would have increased influx of clastics during lowered sea level.

Reservoir Distribution and Compartmentalization

Quartz sandstone tidal sand-ridges and incised-valley-fills are common reservoir facies in the Middle Chesterian Bethel to Glen Dean interval (Seyler and Cluff, 1990; Grube, 1992). Tidal sand-ridges were deposited in the late TST of sequences in the transgressive sequence tract and in the HSTs of sequences in the highstand sequence tract of the composite-sequence. Transgressive tidal sand-ridge reservoirs, which occur in the Bethel Sample and Cypress Formations, were deposited during periods of increasing accommodation and are sealed by backstepping, offshore fossiliferous shale and skeletal limestone. Highstand tidal sand-ridge reservoirs, which occur in the Big Clifty and lower Hardinsburg Formation, were deposited during a period of decreasing accommodation and were sealed by prograding shaly tidal flat and salt marsh facies. Narrow, elongate laterally discontinuous sand-ridges that are both vertically and laterally sealed by shale, such as those in the upper Cypress Formation, have been the most productive reservoirs (Grube, 1992). Thick sandstones, which may be composed of amalgamated parasequences such as the West Baden Clastic Belt, may be prone to leakage and have not been as productive as thinner sand bodies (Sullivan, 1972).

Reservoirs in incised-valley-fills at the base of sequences in the transgressive sequence tract of the composite-sequence, were deposited during periods of increasing accommodation, and are commonly sealed by backstepping offshore fossiliferous shale and skeletal limestone. The most prolific incised-valley-fill reservoir is the Bethel channel (Reynolds and Vincent, 1967) which is laterally sealed by limestone of the Ste.

Genevieve and Paoli Formations and capped by marine shale. Tidal channels cut into shaly tidal flat facies and filled with quartz sandstone during the late highstand of the sequences also may be important reservoir facies.

Recognition of paleosols and incised-valleys has provided a framework in which reservoir distribution and compartmentalization may be better understood in specific fields within the Illinois Basin. Paleosols serve as additional stratigraphic markers and may help in mapping reservoir horizons. These mixed carbonate-siliciclastic sequences provide a well-exposed analog for other Late Mississippian strata such as the giant Carboniferous oil-fields in Kazakstan.

CONCLUSIONS

1. Disconformities and incised-valleys, previously thought to be of relatively local significance, can be mapped regionally in the Late Mississippian Bethel to Glen Dean interval of the Illinois Basin. They help define a high-resolution sequence stratigraphic framework of 4th-order sequences and component parasequences, which yields a better understanding of the vertical and lateral distribution of reservoir facies and their seals.

2. Four disconformity-bounded 4th-order (~400 k.y.) sequences in the Bethel to Hardinsburg interval (up to 68 meters thick) can be traced across the Illinois Basin. Three of the sequences are bounded by type 1 sequence boundaries marked by red mudrock paleosols and incision of up to 75 meters. The other is bounded by a type 2 boundary marked by a basin-wide paleosol. Lowstand systems tracts are not preserved in the Illinois Basin due to the updip position on the ramp. Transgressive systems tracts consist of one or more transgressive-regressive parasequences. The maximum flooding surfaces for the sequences are picked at the base of the deepest water carbonate units to be deposited on the shelves. Early highstand systems tracts are composed of regressive limestone-shale parasequences, whereas late HST parasequences consist of prograding

siliciclastic tide-dominated deltaic facies. Correlation of the sequences across the Illinois Basin and into the Appalachian Basin indicates a eustatic origin. The amplitude of sea-level changes that produced the 4th-order sequences varied between 25 and 95 meters.

3. The sequences bundle into two sequence-pairs and a composite-sequence. Sequence-pairs have deeper incised-valleys and have more quartz sandstone in their basal sequences than the upper sequences which are shalier and dominated by regressive parasequences. The two sequence pairs are arranged into a composite sequence bounded by disconformities that can be traced into the deepest incised-valleys in the study interval. The composite sequence is characterized by incision at the bases of the lower 3 sequences, which are dominantly composed of transgressive-regressive parasequences, and an upper sequence bounded by a type 2 sequence boundary dominantly composed of regressive parasequences.

4. Individual reservoirs occur in both transgressive-regressive and regressive parasequences. Transgressive-regressive parasequences in the TST and early HST of sequences commonly consist of a basal quartz sandstone channel-fills and/or tidal sand-ridges (both important reservoirs) which deepen upward into skeletal limestone and are capped by heterolithic facies or shale. Sequences dominated by transgressive-regressive parasequences resemble Pennsylvanian cyclothems and likely were formed by relatively large sea-level changes. Asymmetric regressive parasequences in the HSTs of sequences commonly consist of laterally extensive skeletal limestone capped by fossiliferous shale. The parasequences can be correlated between areas with vastly different subsidence rates, and were likely formed by <10 meter 5th-order sea-level fluctuations, superimposed on the 4th-order eustatic signal.

5. Other Late Mississippian reservoir facies are likely to be composed of similar high-frequency sequences because of global glacio-eustasy operating during the transition into the Late Paleozoic glaciation.

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APPENDIX A. LOCATIONS OF OUTCROPS USED IN THIS STUDY

NAME, OPERATOR (if applicable)	LOCATION	UNITS DESCRIBED
ILLINOIS		
5. Anna Quarry	East side of Anna, IL, NW of junction of S.R. 51 and S.R. 146	St. Louis to Ste. Genevieve
6. I-57 Roadcut	milepost 27.4 on I-57 N, south of Anna exit	Ste. Genevieve to Aux Vases
7. Cypress Quarry, Columbia Quarries Inc.	S.R. 37, 3 miles south of Cypress	Ste. Genevieve to Aux Vases
14. Hardin Co. Materials, Pit 2	at intersection of Rts. 1 and 146, north of Cave-in-Rock, IL	Ste. Genevieve to Bethel
15. Martin-Marietta, Cave in Rock Quarry	on county roads ENE of Cave-in-Rock, IL	Ste. Genevieve to Bethel
KENTUCKY		
16. Three Rivers Quarry, Martin Marietta	U.S. 60, 6.25 mi. NE of Smithland, KY	Uppermost St. Louis to Aux Vases
17. Fredonia Quarry, Denny and Simpson	U.S. 641, 2.5 mi. south of Fredonia, KY	Ste. Genevieve
18. Princeton Quarry, Kentucky Stone Co.	S.R. 91, 2.75 mi. south of Princeton, KY	Ste. Genevieve to base of Bethel
19. Western Kentucky Parkway	WKP, 3.5 miles east of Princeton, KY	Cypress to Tar Springs
22. S. Hopkinsville Quarry, Rogers Group	U.S. Alt. 41, 4.5 mi. south of Hopkinsville, KY	Ste. Genevieve
23. Christian Quarries Quarry (abandoned)	U.S. 68, 0.5 miles east of Alt. 41, Hopkinsville, KY	upper Ste. Genevieve to Bethel
24. Todd Quarry, Kentucky Stone Co.	U.S. 68, 7 miles west of Elkton, KY	Renault
25. W. Russelville Roadcut	New roadcut for U.S. 68 bypass in west Russelville, KY	upper Ste. Genevieve to Bethel

NAME, OPERATOR (if applicable)	LOCATION	UNITS DESCRIBED
26. Russelville quarry and core, Kentucky Stone Co.	S.R. 79, 1 mi. NE of Russelville, KY	St. Louis through Renault
28. Green River Parkway (now Natcher Parkway)	milepost 9.1, north of Bowling Green, KY	Ste. Genevieve through Renault
29. Green River Parkway	milepost 9.5, north of Bowling Green, KY	Beaver Bend through Big Clifty
30. Rockfield Quarry, Kentucky Stone Co.	U.S. 68, Rockfield, KY	Ste. Genevieve
32. Park City 31W Quarry, Medusa Crushed Stone Co.	U.S. 31W, ~3 miles west of Park City, KY	Ste. Genevieve to Big Clifty
33. Park City, I-65 (partially	I-65 N, at milepost 48, north of Park City	Ste. Genevieve to Big Clifty exit covered)
34. E. Park City Quarry (abandoned)	Old Bardstown Road, 2 mi. east of Park City, KY.	Ste. Genevieve
35. Cave City Quarry, Scotty's Paving	Rt. 90, ~5 miles SE of Cave City, KY	St. Louis to Ste. Genevieve
36. S. Munfordville, I-65 roadcut	I-65 S, 1 mile south of Munfordville exit	Ste. Genevieve
37. N. Munfordville, I-65 roadcut	I-65 N,).5 mi. north of Munfordville exit	Ste. Genevieve to Big Clifty
38. Upton Quarry, Kentucky Stone Co.	2 miles east of Upton exit of I-65 on Quarry Road, Upton, KY	Ste. Genevieve to Big Clifty
41. Cecelia Quarry, Larry Glass Paving	S.R. 62, 3 miles west of Cecelia, KY	Ste. Genevieve to Beech Creek
43. Stephensburg Quarry (abandoned)	on dirt road 2 mi. N of Stephensburg, KY	Ste. Genevieve to Paoli
44. Irvington Quarry, Kentucky Stone Co.	S.R. 477, 2 mi. NW of Irvington, KY	Ste. Genevieve to Paoli

NAME, OPERATOR (if applicable)	LOCATION	UNITS DESCRIBED
45. Battletown Quarry, Kosmos Cement Co.	S.R. 228, 0.5 mi. NW of Battletown, KY	Ste. Genevieve to Big Clifty
INDIANA		
46. Cape Sandy Quarry, Pit 1, Mulzer Crushed Stone, Inc.	4 miles E. of Alton, IN	Ste. Genevieve to Big Clifty
49. Tower Quarry, Mulzer Crushed Stone, Inc.	2 mi. north of Leavenworth, IN	uppermost St. Louis to Sample
50. Scout Mountain Roadcut, I-64	milepost 99, 7 miles west of Corydon exit	Ste. Genevieve to Reelsville
51. Corydon, Corydon Crushed Stone Co.	2.5 mi. NW of Corydon, IN, off S.R. 135	St. Louis to Paoli
52. Robertson Crushed Stone, Inc.	S.R. 64, 1 mi. W of Depauw	St. Louis to Beaver Bend
53. Sulphur Interchange, I-64	Exit 86 off of I-64, SE and NW roadcuts	Reelsville to Hardinsburg
62. W. Paoli Quarry, Cave Quarries, Inc.	U.S. 150, 3.5 mi. NW of Paoli, IN	Ste. Genevieve to Reelsville
64. Mitchell Quarry, Rogers Group	S.R. 60, 5 mi. west of Mitchell, IN	Ste. Genevieve to Cypress
68. Sieboldt Quarry, Rogers Group	county roads, 5 mi. north of Oolitic, IN 2 mi. west of S.R. 37	Ste. Genevieve to Bethel
71. Bloomington Quarry, Rogers Group	Oard Rd, 4 mi. west of Bloomington, IN	Ste. Genevieve to Paoli
74. Putnamville Quarry, Kentucky Stone Co.	S.R. 343, 1 mi. west of Cloverdale, IN	Ste. Genevieve to Beaver Bend

APPENDIX B. CORES STORED AT GEOLOGICAL SURVEYS USED IN THIS STUDY

COUNTY	CALL N0.	LOCATION	OPERATOR	FARM	WELL NO
ILLINOIS STATE GEOLOGICAL SURVEY					
1. Madison	C-141	26-4N-8W	Madison Coal Corp.		No. 6
2. Madison	C-140	3-3N-8W	Madison Coal Corp.		No. 5
3. Madison	C-146	9-3N-8W	Madison Coal Corp.		No. 4
4. Union	C-13619	6-11S-2W	ISGS-COGEOMAP		CB-2
8. Pope	C-46	19-13S-6E	Abner Field	Abner Field	No. 1
9. Pope	C-13076	33-12S-7E	Ozark Mahoning		BM-225
10. Hardin	C-12781	21-12S-8E	Ozark-Mahoning		KT-19S
11. Hardin	C-13277	11-11S-7E	U.S. Bureau ofMines	Knox & Yingling Proj.	K-4
12. Hardin	C-12782	17-11S-9E	Ozark-Mahoning	U.S. Harris Creek Tract	USHC-55S
13. Hardin		9-11S-9E	Ozark-Mahoning	Denton Mine	DT-134
KENTUCKY GEOLOGICAL SURVEY					
17. Caldwell	C-218	21-G-21	KGS-USGS	Fredonia Valley Quarry	No. 1
20. Caldwell	C-216	2-I-20	KGS-USGS	Shrewsbury Collins	No. 1
21. Caldwell	C-220	2-I-20	KGS-USGS	Cobb	No. 1
31. Edmonson	C-1244		Marathon Oil	Lindsey	No. 22
40. Hardin	C-117	6-M-42	KGS-USGS	Summit Area	No. 1
42. Breckenridge	C-118	10-N-38	KGS-USGS	Dennis	No. 1
INDIANA GEOLOGICAL SURVEY					
47. Harrison	C-339	14-4S-2E	IGS	Harrison State Forest	SDH-152
48. Perry	C-303	10-4S-1W	IGS	Knieriem, Lafe	SDH-132
54. Crawford	C-626	7-2S-3W	IGS	Hy-Rock Products Co.	SDH-313
55. Dubois	C-643	14-2S-3W	IGS	Schroeder, Walter	SDH-320
57. Orange	C-338	15-1S-1E	IGS	Free, Robert	SDH-151
58. Dubois	C-230	11-1S-3W	IGS	Kalb, Ralph	SDH-83
59. Orange	C-625	30-1S-2W	IGS	US Army Corps of Engnrs.	SDH-311
60. Orange	C-651	34-1N-2W	IGS	Belcher, James	SDH-326
61. Orange	C-167	32-2N-2W	IGS	Hendricks, Harry	SDH-48
63. Martin	C-214	23-3N-3W	U.S. Gypsum	U. S. Gypsum	No. 10-60

65. Lawrence	C-340	27-4N-1W	IGS	Tow, Walter	SDH-153
66. Martin	C-735	28-5N-3W	U.S. Army Corps of Eng.	Naval Weapon Support Ctr.	No. C-3
67. Martin	C-797	15-5N-3W	USAE Waterway Exp. Sta.	NWSC Crane DBI	DB-1
69. Greene	C-487	29-7N-5W	IGS	Roudebush, Renos	SDH-208
70. Monroe	C-672	22-7N-2W	IGS	Munson, P.J.	SDH-330
72. Monroe	C-342	7-8N-2W	IGS	Leininger, Richard	SDH-155
73. Clay	C-444	23-11N-6W	IGS	Fisher, John	SDH-186
75. Putnam	C-105	16-13N-5W	IGS	Pennsylvania Railroad	SDH-8-A

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PERSONAL INFORMATION:

Born March 22, 1961, Norfolk, Virginia — Married, two children

EDUCATION:

Ph.D., Geology, in progress at Virginia Tech, August 1992-present

B.S., Geology, 1992, Temple University, May 1992

POSITIONS HELD:

Graduate Teaching Assistant — August 1992-present
Virginia Tech, Blacksburg, Va.

Undergraduate Teaching Assistant — May 1992-August 1992
Temple University, Philadelphia, Pa.

Ombudsperson/Student Advisor — August 1991-May 1992
Temple University Geology Department, Phila., Pa.

Partner — January 1986-January 1990
Widmer-Smith Construction Company, Phila., Pa.,
Duties performed included management of up to thirteen employees, contract negotiations, building design and construction management

TEACHING EXPERIENCE:

Sedimentology/Stratigraphy Laboratory
organized and taught laboratory/field course for senior level class that covered sedimentology, stratigraphy and petrology of clastics and carbonates.

Historical Geology Laboratory

Physical Geology Laboratory

Catastrophic Geology Laboratory

Geology for Engineers Laboratory

AWARDS:

Tillman Teaching Award, Virginia Tech, 1994 (for excellence in teaching)

Byron Cooper Award, Virginia Tech, 1994, 1995 (for research in Appalachians)

Otto Kurschner Award for Excellence in Science, Temple University, 1992
(Awarded to the student with the highest G.P.A. in all sciences at Temple University).

Temple Geology Department Award, Temple University, 1992.

RESEARCH GRANTS:

Petroleum Research Fund, 1995

Sigma Xi, 1993, 1995.

Geological Society of America, 1993, 1994.

Appalachian Basin Industrial Associates, 1993, 1994, 1995.

PUBLICATIONS:

Smith, L.B., and W.J. Nelson, 1996, High-resolution sequence stratigraphy of the hydrocarbon-bearing Ste. Genevieve-Downeys Bluff Formations, Illinois Geological Society Field Trip Guide, 19p.

Smith, L.B. and J.F. Read, 1995, Stratigraphic evidence for increasing amplitude 4th-order eustasy during Late Mississippian greenhouse to icehouse transition, Illinois basin (abstract): Geological Society of America, 1995 Annual Meeting, Abstracts with Programs, p.A332.

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