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APPLICATION OF HIGH-RESOLUTION SEQUENCE STRATIGRAPHY TO TIDALLY INFLUENCED UPPER MISSISSIPPIAN CARBONATES, ILLINOIS BASIN

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ABSTRACT: The Upper Mississippian (Chesterian) Ste. Genevieve and Paoli Formations of the Illinois basin are composed of five carbonate and mixed carbonate-siliciclastic fourth-order (100–500 k.y.) sequences. 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 as many as nine regionally developed parasequences. Carbonate parasequences have patchy updip carbonate eolianites at the base, overlain by laterally discontinuous ooid tidal ridge and skeletal bank facies and laterally extensive muddy carbonate caps. Mixed carbonate-siliciclastic parasequences are composed of skeletal limestone capped by fossiliferous shale. Carbonate tidal-flat facies are rare in the succession.

The sequences were produced by moderate–amplitude, fourth-order (100–500 k.y.) glacio-eustatic sea-level changes, whereas the parasequences were likely produced by fifth-order (10–100 k.y.) glacio-eustatic sea-level changes during a time of transition from the greenhouse conditions of the Early Mississippian to the icehouse conditions of the late Paleozoic. This transition is reflected in upward change from carbonate-dominated, disconformity-bounded sequences with little lowstand erosion at the base to mixed carbonate-clastic sequences with significant lowstand erosion at the top of the study interval and in overlying sequences. These changes were likely caused by an increase in the magnitude of sea-level fluctuations coupled with increasingly humid wet-dry seasonal climate. The sequences may be stacked into one or two third-order (0.5–5 m.y.) composite sequences that are bounded by relatively well developed unconformities. However, the magnitude of the fourth-order signal makes third-order bundling less clear than in strata deposited during greenhouse times of lower-amplitude fourth- and fifth-order sea-level fluctuations. The dominance of fourth-order disconformity-bounded sequences and the lack of regional tidal-flat facies reflect moderate-amplitude eustasy associated with the transition from global greenhouse climates of the Early Mississippian to times of abundant global ice in the late Paleozoic.

INTRODUCTION

Previous work on the lower Chesterian Ste. Genevieve to Paoli Formations 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 Dodd et al. (1996), among others. The lateral heterogeneity of the ooid grainstone reservoirs has made the Ste. Genevieve a "correlation nightmare" in the subsurface (Zuppann, 1993). Most paleosols were considered to be local in extent and to have formed as a result of local island shoaling (Droste and Carpenter, 1990). Recent work has shown that some sequences and parasequences bounded by paleosols and eolianite horizons can be correlated both locally and regionally (Smith et al., 1995; Smith and Read, 1995; Smith, 1996; Dodd et al., 1996).

This paper demonstrates the value of combining detailed sedimentology with a basin-wide stratigraphic approach. Only through sedimentologic study of all rock types and exposure surfaces can the sequence stratigraphy be accurately interpreted, and only through basin-wide stratigraphic study can local features be distinguished from laterally extensive marker horizons and genetically significant surfaces. Detailed lithologic logs of 57 outcrops and cores were used to construct regional lithologic cross sections of the Ste. Genevieve Formation and the Paoli Formation (and its time-equivalent units) along more than 600 km 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 that 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-siliciclastic sequences deposited updip on a gently dipping ramp. This study also provides insight into the stratigraphic signature of the transition from global greenhouse conditions of the Early Mississippian to times of abundant global ice in the Pennsylvanian and Early Permian.

GEOLOGIC SETTING

Tectonic and Paleogeographic Setting

The Illinois basin was located between 5° and 15° 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 an extensive carbonate ramp that was continuous from Pennsylvania to western Canada (Craig and Connor, 1979) (Fig. 1A). The ramp was separated from the Antler foredeep of

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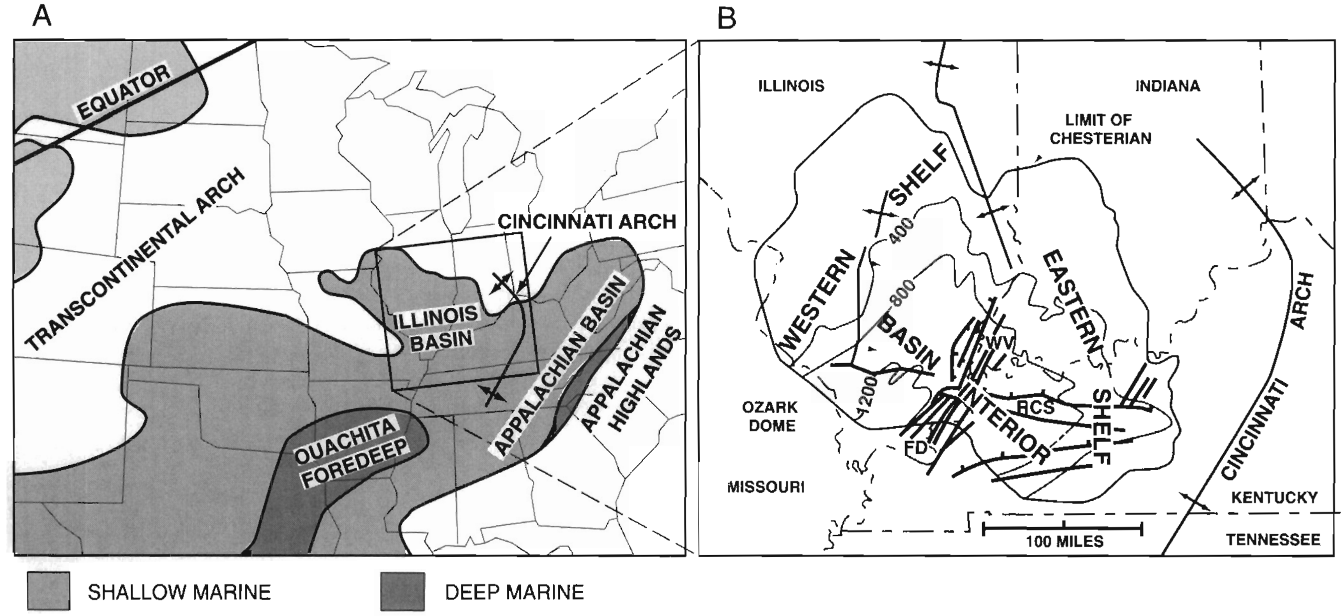


FIG. 1.—(A) Paleogeographic map showing regional setting of the Illinois basin (modified from Craig and Connor, 1979). (B) Tectonic map of Illinois basin. Isopachs in feet for the interval from the Bethel to the Mississippian/Pennsylvanian boundary (from Swann, 1963); 800-ft isopach is approximate boundary of basin interior. Faults: Wabash Valley fault system (WV); Rough Creek–Shawneetown fault system (RCS); Fluorspar District fault complex (FD).

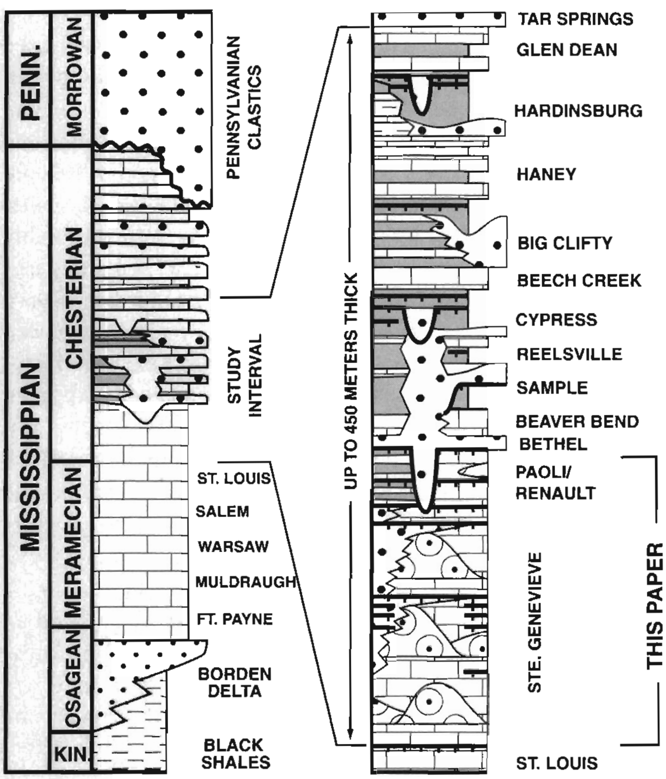


FIG. 2.—Left column is a generalized stratigraphic column for the Mississippian in the Illinois basin. Right column shows more detailed stratigraphy of lower to middle Chesterian units. This paper concentrates on the Ste. Genevieve to Paoli interval. For explanation of patterns of units, see Figure 4.

the western United States by the Transcontinental arch, which was emergent throughout the Chesterian (Craig and Connor, 1979). The Illinois basin lay at least 300 km updip from the ramp margin that may have started to develop along 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). The term shelf is used by local geologists to distinguish areas of lower subsidence rates from the more rapidly subsiding basin interior. The eastern shelf formed along the western flank of the Cincinnati arch, and the western shelf formed over the DuQuoin monocline. The basin interior appears to be centered on the Wabash Valley and Fluorspar district fault systems that trend northeast and may have been active during the Late Mississippian. Despite the varying subsidence rates, facies interpretations suggest that shallow-marine to nonmarine conditions prevailed throughout the Illinois basin during the early and middle Chesterian.

Stratigraphic Setting

The Ste. Genevieve and Paoli Formations lie at the top of a thick Middle Mississippian carbonate succession that includes the Osagean to Meramecian Muldraugh, Warsaw, Salem, and St. Louis Formations (Cluff and Lineback, 1981) (Fig. 2). The St. Louis Formation directly underlies the study interval (Figs. 2, 3) and is primarily composed of shallow-water muddy carbonates and skeletal limestone (Cluff and Lineback, 1981). The base of the Ste. Genevieve is picked at a biostratigraphic zone boundary that coincides with a regional disconformity that commonly occurs within a muddy carbonate unit overlying the Lost River Chert of Elrod (1899). In outcrops and cores, the Lost River

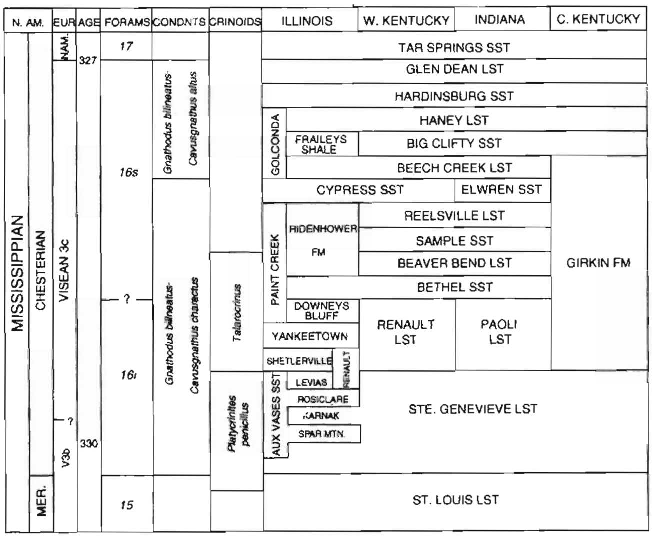


FIG. 3.—Biostratigraphy and correlation chart for Ste. Genevieve to Glen Dean interval (modified from Sable and Dever, 1990, and Maples and Waters, 1987). MER. = Meramecian, SST = sandstone, LST = limestone, Fm = formation.

Chert is a distinctive marker bed that occurs in the upper part of a laterally extensive grainstone unit in which the outlines of fenestrate bryozoans are commonly preserved within chert nodules. Conodont, foraminifer, and coral biostratigraphic zone boundaries occur at this unconformity (Rexroad and Fraunfelter, 1977). Maples and Waters (1987) moved the Meramecian/Chesterian provincial series boundary to this unconformity because the fauna above the boundary more closely resembles the fauna in the middle and upper Chesterian than the fauna in the underlying St. Louis Formation. Thus, age of the Ste. Genevieve to Paoli interval is early Chesterian or late Visean (middle V3b to early V3c).

The Ste. Genevieve Formation (Fig. 3) is primarily composed of ooid grainstone (a common reservoir facies), skeletal grainstone, commonly dolomitized muddy carbonates, and numerous disconformities that can be recognized by the presence of caliche, breccia, and eolianites. The boundary between the middle and upper Ste. Genevieve and the Aux Vases Sandstone in the western part of the basin was thought to be gradational (Cluff and Lineback, 1981) but has recently been interpreted to be primarily erosional (Leetaru, 1997). The source of the Aux Vases Sandstone lay to the northwest; the formation 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 Sandstone (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 interval 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, 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 is generally overlain 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. The Girkin Limestone includes the Paoli, Beaver Bend, Reelsville, and Beech Creek Limestones, each separated by well-developed carbonate paleosol horizons. The Paoli interval is overlain by middle and upper Chesterian mixed carbonate-clastic sequences (Swann, 1963, 1964; Smith, 1996).

CONSTRUCTION OF LITHOFACIES CROSS SECTIONS

Cross sections through the study interval have been constructed on the basis of 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, type and shape, 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 by 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 maps (Fig. 4).

LITHOFACIES AND DEPOSITIONAL ENVIRONMENTS

Characteristics of major rock types are summarized in Table 1. The relative abundance of facies at each location (Fig. 5), field observations, and subsurface data from the literature (Choquette

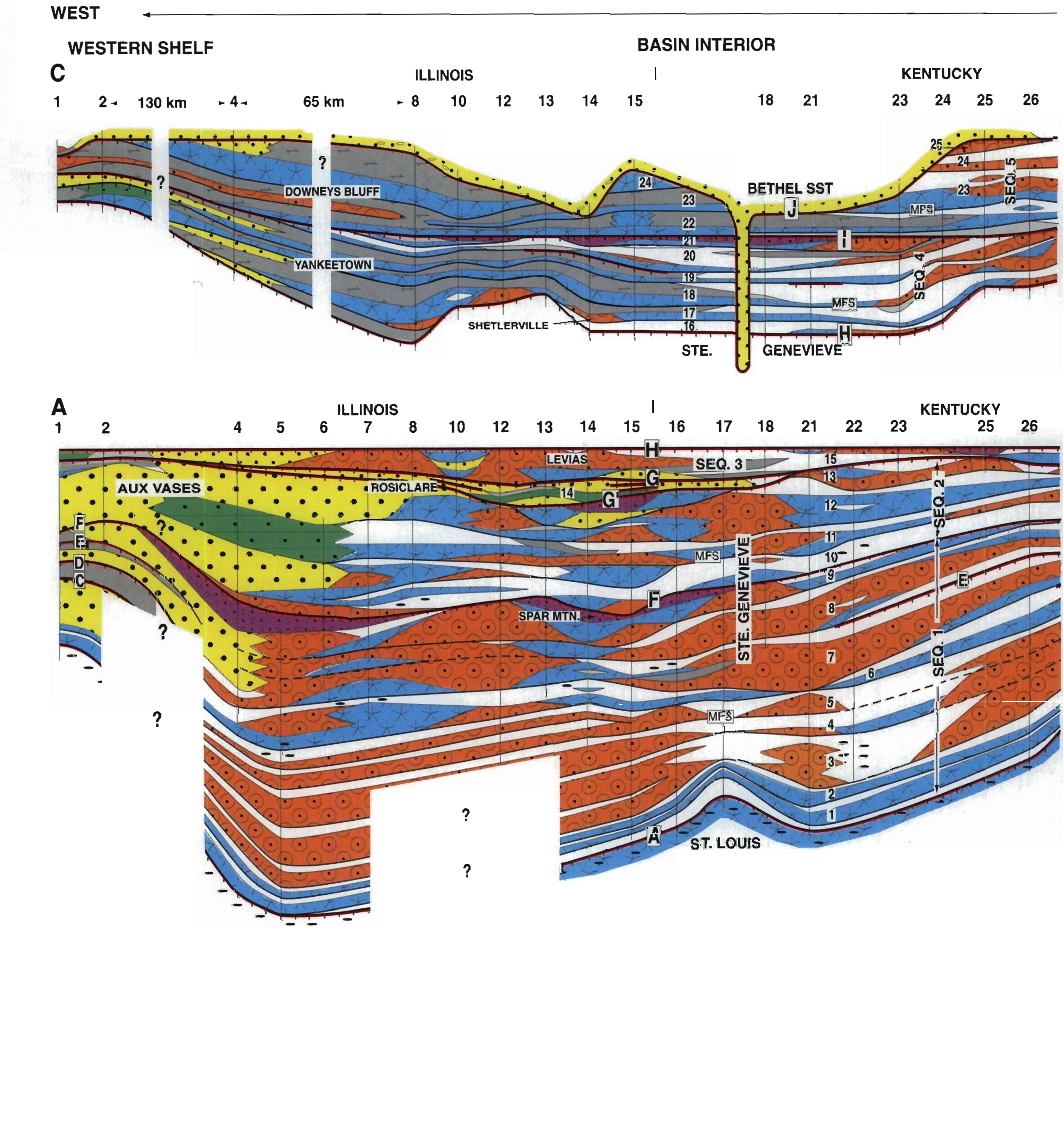
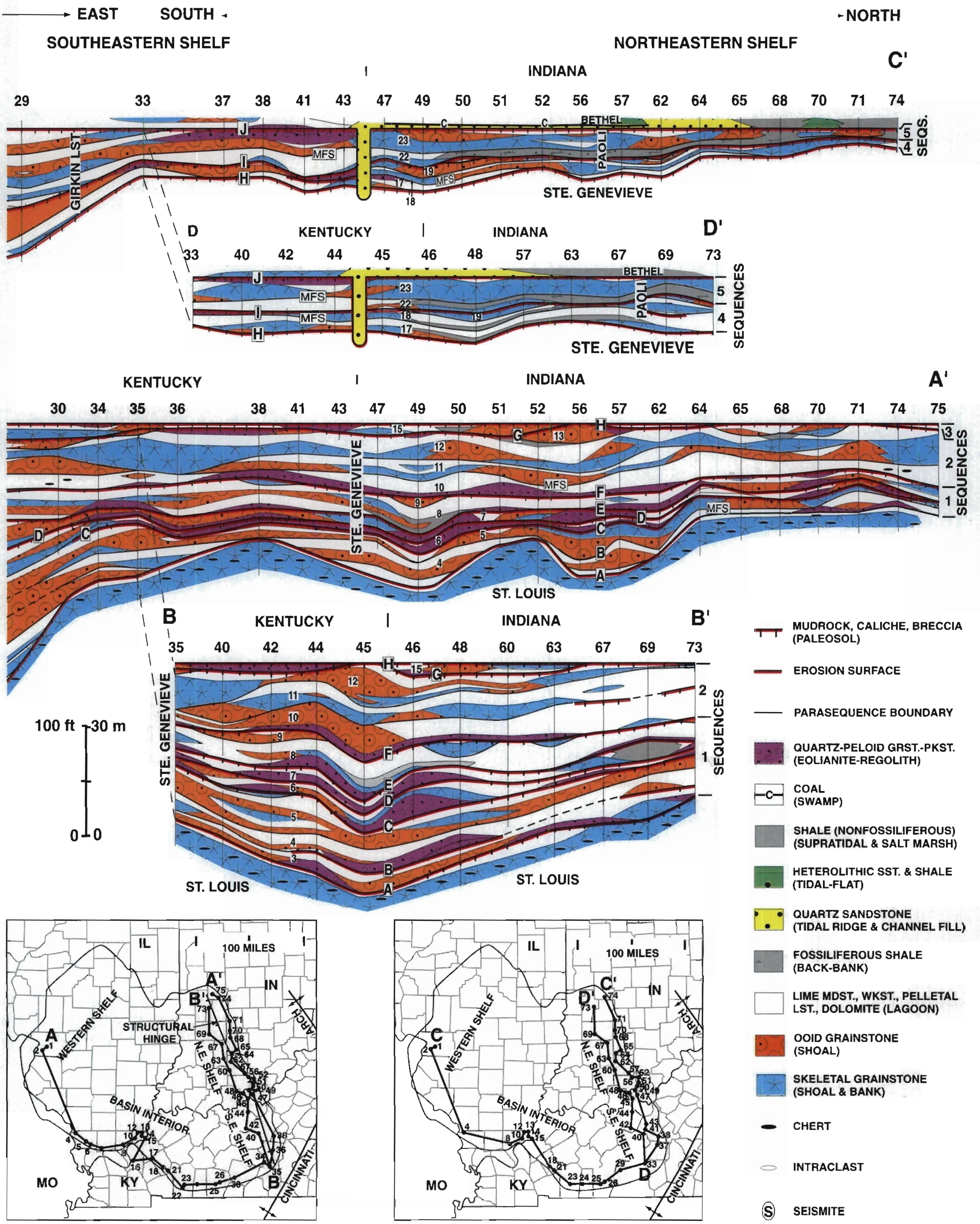


FIG. 4.—Interpretive cross sections of sequences 1 through 5 in the Ste. Genevieve to Paoli interval. Measured sections are equally spaced except where there are major gaps between sections (see location maps for actual horizontal spacing). Disconformities are lettered from A to J, and parasequences are numbered from 1 to 25. The Bethel channel is not drawn to scale; its actual depth is 75 m (Friberg et al., 1969). Cross section A-A' includes the Ste. Genevieve and trends west to east from location 1 to location 35 and south to north from location 35 to location 75. Cross section B-B' shares location 35 and trends south to north parallel and downdip of cross section 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 to north parallel and downdip of cross section C-C'. MFS = maximum-flooding surface, GRST. = grainstone, PKST. = packstone, MDST. = mudstone, WKST. = wackestone, LST. = limestone.


EAST
SOUTH
NORTH
SOUTHEASTERN SHELF
NORTHEASTERN SHELF
C'
INDIANA
29 33 37 38 41 43 47 49 50 51 52 56 57 62 64 65 68 70 71 74
BETHEL
PAOLI
GIRKIN LST.
MFS
STE. GENEVIEVE
SEQS.
D
KENTUCKY
INDIANA
D'
33 40 42 44 45 46 48 57 63 67 69 73
SEQUENCES
KENTUCKY
INDIANA
A'
30 34 35 36 38 41 43 47 49 50 51 52 56 57 62 64 65 68 70 71 74 75
ST. LOUIS
B
KENTUCKY
INDIANA
B'
35 40 42 44 45 46 48 60 63 67 69 73
100 ft 30 m
0 0
MUDROCK, CALICHE, BRECCIA (PALEOSOL)
EROSION SURFACE
PARASEQUENCE BOUNDARY
QUARTZ-PELOID GRST.-PKST. (EOLIANITE-REGOLITH)
COAL (SWAMP)
SHALE (NONFOSSILIFEROUS) (SUPRATIDAL & SALT MARSH)
HETEROLITHIC SST. & SHALE (TIDAL-FLAT)
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
SEISMITE
IL
IN
MO
KY
100 MILES
WESTERN SHELF
STRUCTURAL HINGE
N.E. SHELF
S.E. SHELF
BASIN INTERIOR
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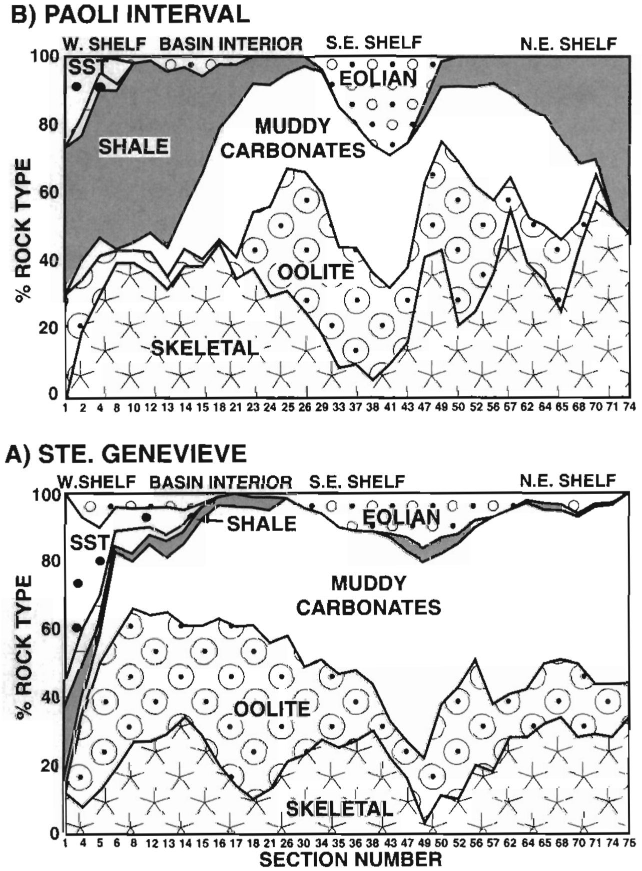

FIG. 5.—Facies distribution on the ramp based on percentage of rock type at each location. Curves smoothed by using three-point average. (A) Ste. Genevieve facies distribution (sequences 1 and 2); (B) Paoli interval facies distribution (sequences 3 and 4).

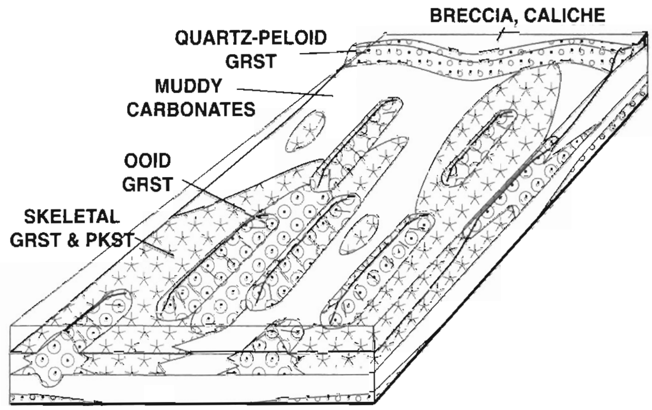

FIG. 6.—Schematic depositional environment for carbonate-dominated ramp in the Ste. Genevieve Formation (not to scale). See Figure 4 for legend and rock types.

and Steinen, 1980; Cluff and Lineback, 1981; Zuppann and Keith, 1988) were used to determine relative facies distribution and depositional environments for carbonate-dominated and mixed carbonate-siliciclastic parts of the ramp (Fig. 6). The ramp slope was very gentle (average 1.5 cm/km [0.001°] for the basin and no more than 5 cm/km [0.003°] along the margins of the shelves [Smith, 1996]), which accounts for the similarity of facies deposited in the central and eastern parts of the basin in the Ste. Genevieve Formation (Smith, 1996). Although the facies on the eastern shelf and the basin interior are different in sequences 4 and 5, this variation is primarily due to the northwestern location of the clastic source rather than a change in ramp slope.

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 of the wet season must have been relatively short during Ste. Genevieve time, on the basis of the abundance of ooids, dolomite, and eolianites, all of which are most common in areas with arid and semiarid 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's 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. The trend of increasingly humid wet-dry seasonality continued into the overlying sequences, which have more siliciclastics as well as thin coal beds (Smith, 1996).

Carbonate-Dominated Ramp

The Ste. Genevieve Formation and most of the Paoli interval of central Kentucky and southeastern Indiana were deposited on a carbonate-dominated part of the ramp (Fig. 6) (facies summarized in Table 1).

Skeletal grainstone and packstone.—
Skeletal grainstone and packstone (Fig. 7A) formed in moderate- to high-energy conditions on the basis of the lack of mud, the medium to very coarse grain size, and the abundance of echinoderms. In this setting, skeletal grainstone formed landward, in between and seaward of ooid shoals in water depths that rarely exceeded 10 m (Smith, 1996).

Ooid grainstone.—
This reservoir facies (Fig. 7B) 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 locally cut tidal channels into underlying strata. Ooid grainstone is widespread in the Ste. Genevieve, with the thickest accumulations in the basin interior, but largely confined to the eastern shelf in the Paoli interval.

Tidal influence on grainstone units.—
Strong tidal influence on the gently sloping ramp caused much lateral heterogeneity of facies. Ooid grainstone reservoir

TABLE 1.—CARBONATE AND MIXED CARBONATE–CLASTIC RAMP FACIES

Lithology (Depositional Environment)	Blocky mudrock and caliche/breccia (paleosol)	Quartz-peloid grainstone (eolianite)	Muddy carbonates (lagoon and intershoal)	Ooid grainstone (tidal bar and shoal)	Skeletal grainstone and packstone (bank, shoal)	Quartz sandstone (tidal sand ridge, channel fill)	Fossiliferous Shale (carbonate-clastic transition)
Occurrence	Common throughout basin, better developed on shelves. Commonly underlies and overlies marginal marine shale with plant fragments.	Dune-forms; units commonly thicken and thin over short distances; underlie and overlie disconformities at tops and bases of sequences, respectively.	Commonly occur as regional, traceable sheets capping parsequences. Also lagoons and local intershoal swale-fills.	Dip-parallel oolitic bar-forms (Zuppann, 1993); commonly occur at base of parasequences in Ste. Genevieve and Paoli of eastern shelf.	Common throughout basin in all limestone units. Forms sheet-like units at base of most regressive parasequences.	Commonly fills incised valleys and forms in linear sand ridges.	Common throughout basin. Transition between siliciclastic and carbonate rock types.
Color	Caliche is light brown; Breccia zones are dark gray; Mudrock is red and maroon.	Dark gray.	Dolomitized units are tan; limestone units light gray brown to dark gray.	White to light gray.	Light to medium to dark gray.	White to light gray/green.	Olive green and dark gray
Depositional texture and grain type	Caliche is stringy to laminated micrite. Breccia has fitted fabric. Mudrock is blocky, brecciated, slickensided clay w/dolomitic nodules.	Grainstone; fine to medium, tightly packed and well sorted within laminae; dominated by peloids, rounded lithoclasts, subangular quartz; lesser ooids (whole and broken), echinoderm grains.	Mudstone and wackestone; predominantly composed dolomitic or 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 peloids.	Grainstone and packstone with up to 15% shale. Medium to very coarse grained.	Very-fine to medium-grained, subangular, well-sorted; may have up to 15% shale.	Wackestone/fissile shale; whole brachiopods and bryozoans.
Bedding and sedimentary structures	Mudrock occurs in breccia and calichified zones up to three meters thick. Massive beds up to one meter thick; rooting and reduction halos common.	Millimeter-scale inverse-graded laminations form cross-beds that dip <20%; cross-beds dip multidirectionally but dominantly to the north (Hunter, 1993).	Massive to burrowed, rare fenestrae, laminations, and mudcracks.	Cross-bedded, less commonly thick bedded and massive; foresets dip as much as 25%.	Thick-bedded, massive and cross-bedded.	Cross-bedded, flat-bedded and massive bedded, flaser-bedding common.	Fissile, flat-bedded.
Terrigenous clastics admixed or interbedded		Average 15% very fine quartz sand on eastern shelf; up to 75% quartz sand in basin interior and western shelf.	Green shale, quartz silt and quartz sand common.	Quartz grains form nuclei of ooids above some quartz sandstone units.	Quartz sand and silt are rare; thin green and gray shale beds and stringers are common.		
Biota		Abraded, rounded skeletal grains common.	Lagoonal facies: ostracodes and gastropods; Intershoal swales: bryozoans, brachiopods and echinoderms.	Echinoderm and brachiopod fragments common.	Abundant echinoderms, brachiopods, bryozoans, common mollusks and foraminifera.	Rare echinoderm fragments, could be transported or eroded from underlying limestone.	Bryozoans, brachiopods and echinoderms.

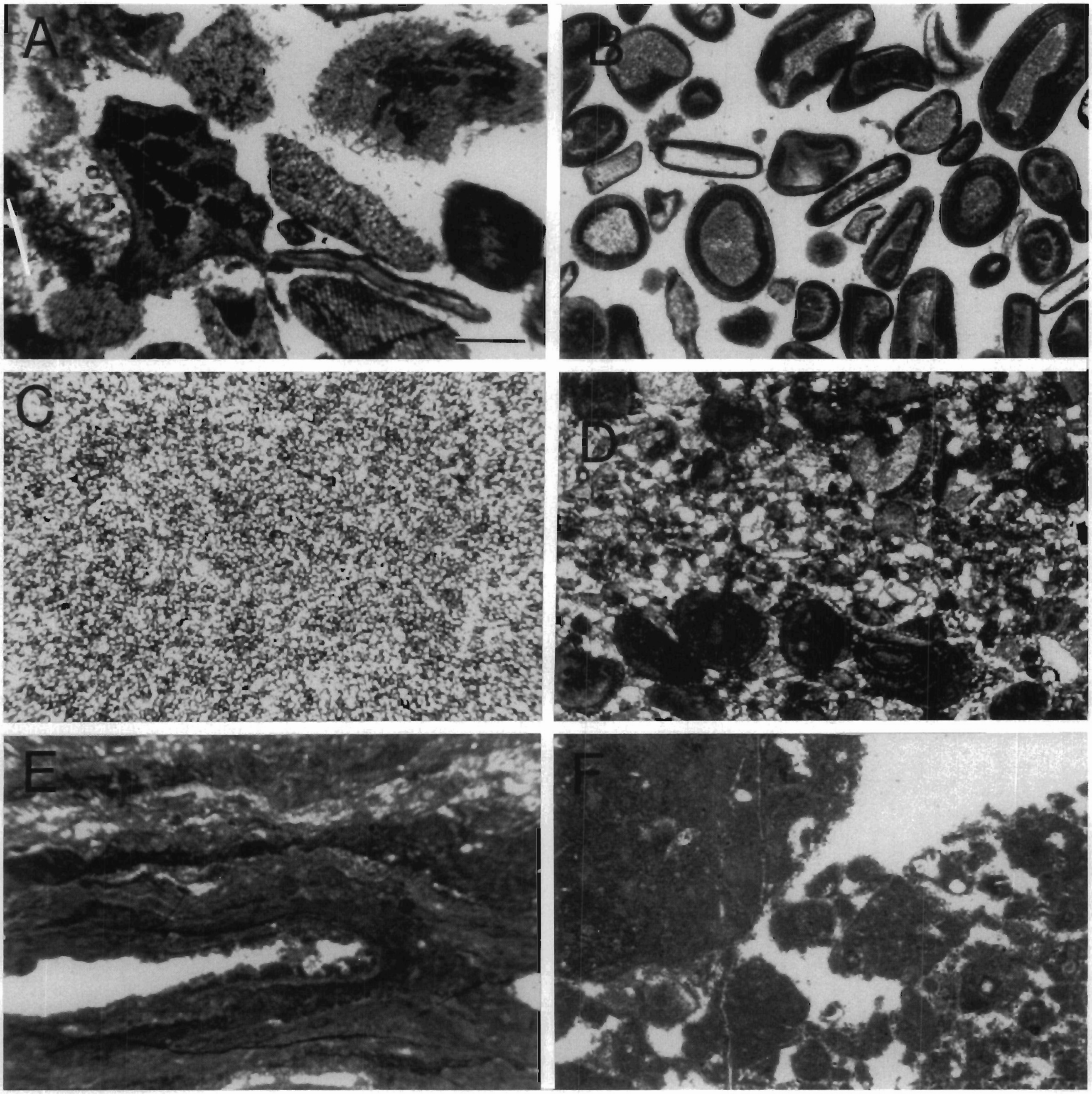

FIG. 7.—Photomicrographs of important rock types: (A) skeletal grainstone, (B) ooid grainstone, (C) microcrystalline dolomite, (D) quartz-peloid grainstone (eolianite), (E) caliche, (F) breccia. Scale on A represents 0.25 mm and is same for all photographs.

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 had been wave-dominated, then 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.

Muddy carbonates.—

These have been interpreted as low-energy subtidal lagoon facies (Choquette and Steinen, 1980), as open-shelf facies formed seaward of ooid and skeletal grainstone (Dodd et al., 1996), and as facies formed both landward and seaward of the ooid shoals (Cluff and Lineback, 1981). On the basis of the fauna, grain size, regional distribution, and stratigraphic relationships, 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-bearing mudstone and wackestone (Fig. 7C) 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 were rare despite the arid Late Mississippian climate.

Quartz-peloid grainstone.—

This rock type (Fig. 7D) is most common on the central eastern shelf in the Ste. Genevieve and on the south part of the eastern shelf in the Paoli interval, but also occurs beneath disconformities in the basin interior. The cross-bedded, quartz-peloid grainstone units are eolianites, on the basis of their inverse grading, sorting, rhizoliths, rounded broken ooids, lack of fossils or coarse grains, and close association with carbonate paleosols (Merkley, 1991; Hunter, 1993; Dodd et al., 1993). Some subhorizontally laminated quartz-peloid grainstone with coarse-grained fossiliferous laminae may be beach facies. 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). Quartz-peloid grainstone commonly overlies or is capped by caliche and breccia and occurs in beds that are as much as 6 m thick and that thicken and thin over short distances.

Carbonate paleosols.—

Quartz-peloid packstone, caliche, and breccia are paleosols that formed during subaerial exposure of carbonate rocks in a seasonally wet-dry climate. In the early stages of Late Mississippian carbonate paleosol development, quartz-peloid packstone formed when eolian peloids, quartz sand, and ooids accumulated along with clay carried by wind or sheetwash. 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 were precipitated to form gray-brown, quartz-peloid packstone paleosols. With time, the sediments were cemented, and laminar caliche (Fig. 7E) lined vertical fractures, joints, and bedding planes. In the most mature profiles, exposed carbonates were brecciated (Fig. 7F), and laminar caliche crusts and subhorizontal layers of caliche formed as much as 4 m below the exposed surface along joints and bedding planes (Walls et al., 1975; Ettensohn, 1975; Harrison and Steinen, 1978; Ettensohn et al., 1988).

Mixed Carbonate-Siliciclastic Ramp

The Aux Vases Formation and the Paoli interval of Illinois, western Kentucky, and western and central Indiana were deposited on a mixed carbonate-siliciclastic part of the ramp (Smith, 1996).

Skeletal grainstone and packstone.—

Where associated with siliciclastics, skeletal grainstone and packstone likely formed in a deeper, more distal environment than in the Ste. Genevieve on the basis of the lack of cross-bedding, lack of ooids, and common presence of graded beds that suggest storm influence. Skeletal grainstone and packstone grade laterally into fossiliferous shale that represents a gradation between carbonate and siliciclastic environments.

Fossiliferous shale.—

This rock type commonly interfingers with muddy carbonates to the east and grades laterally into tidally influenced heterolithic interlaminated sandstone and shale toward the northwestern siliciclastic source. The heterolithic facies formed both on tidal flats and in the transition zone seaward of quartz sandstone.

Quartz sandstone.—

Quartz sandstone was deposited in a tide-dominated shallow-marine environment on the basis of abundant channel-form structures with tidal fills, bidirectional 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 with conglomerate beds also commonly fills channels scoured by strong tidal currents.

Paleosols.—

On the mixed carbonate-siliciclastic ramp, paleosols are most commonly blocky, slickensided mudrock that formed from clay 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 (Smith, 1996).

Relative Ranking of Disconformities

Disconformities (lettered from A to J, oldest to youngest, on cross sections) can be crudely ranked in terms of the relative duration of exposure, estimated from the relative abundance of exposure features (Fig. 8). 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 karst-weathered limestone. Blocky, slickensided mudrock paleosols and incised valleys are treated separately because they are difficult to rank relative to other features that indicate carbonate exposure.

Disconformities H and J, on which the cross sections are hung, are the best developed and likely represent the longest period of emergence (major disconformities). Disconformity H has the highest occurrence of caliche and breccia, and there is major incision associated with Disconformity J. Although depth of incision is a better measure of sea-level fall than of duration of exposure, it would have taken a relatively long period of time

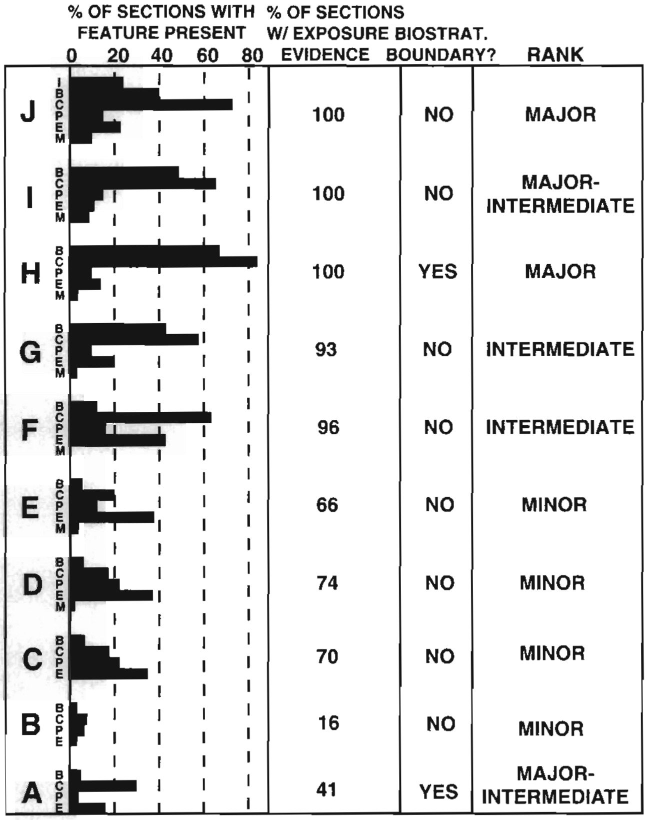

FIG. 8.—Relative ranking of disconformities based on percentage of locations measured with breccia (B), caliche (C), dirty packstone (P), eolianites (E), and slickensided mudrock (M). Incision (section I) only occurs on the uppermost boundary and is a better measure of how far sea level fell rather than the duration of exposure. Disconformities that coincide with biostratigraphic boundaries may represent relatively long-term exposure.

to incise a valley more than 75 m through underlying carbonate strata. Disconformity A may also represent a major hiatus because it shows 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.

SEQUENCE STRATIGRAPHY

A *sequence* is defined as a genetically related succession of strata lacking apparent internal unconformities and composed of parasequences and parasequence sets arranged in systems tracts and bounded by unconformities or their correlative conformities (Mitchum and Van Wagoner, 1991). It is not possible to meet all of these criteria with the disconformity-bounded units preserved in the Upper Mississippian strata of the Illinois basin because they were deposited hundreds of kilometers updip from the shelf edge. In this updip setting, in which lowstand systems tracts are absent and are represented only by unconformities, it is difficult to recognize parasequence sets, and transgressive and highstand systems tracts or entire sequences may be composed of as few as one parasequence. Stratigraphic units are called sequences in this study if they have maximum-flooding surfaces and are bounded by regionally extensive disconformities. On the basis of this definition, there are five disconformity-bounded sequences in the Ste. Genevieve to Paoli interval that are each composed of as many as nine parasequences.

A *parasequence* is defined as a relatively conformable succession of genetically related beds and bedsets bounded by marine flooding surfaces and their correlative surfaces (Mitchum and Van Wagoner, 1991). Most parasequences in the study interval can be traced across the eastern shelf and into the basin interior within a framework of disconformities and laterally extensive marker beds. Some parasequences are confined to local topographic lows and form multilateral, discontinuous units, especially beneath sequence boundaries. Parasequences are 1 to 8 m thick on the eastern shelf, where some parasequences are capped by disconformities, and 4 to 12 m thick in the basin interior where they are generally conformable.

Typical parasequences consist of, from base to top, (1) a disconformity and/or marine flooding surface; (2) locally developed muddy carbonates and/or quartz-peloid grainstone (eolianite); (3) laterally discontinuous ooid grainstone ridges and channel fills that overlie, underlie, and pass laterally into skeletal grainstone sheets and banks; (4) laterally extensive muddy carbonate or fossiliferous shale units; (5) locally developed quartz-peloid grainstone (eolianite) units; and (6) a disconformity and/or marine flooding surface. The facies within sequences 1 through 5 vary from the western shelf to the basin interior to the eastern shelf and are discussed individually to illustrate their differences and similarities.

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 and the lower half of the Aux Vases Formation on the western shelf, and it is capped by the Spar Mountain Member of the Aux Vases in the basin interior. Sequence 1 is 27 m thick on the western shelf, thickens to 63 m in the basin interior, thins to 20 m near the Cincinnati arch and 9 m on the northern part of eastern shelf.

On the updip parts of the eastern shelf (A–A′), the lower boundary is picked at an undulose surface in a dolomite unit overlying the Lost River Chert Bed that coincides with a major biostratigraphic zone boundary at the St. Louis/Ste. Genevieve contact (Rexroad and Fraunfelter, 1977; Maples and Waters, 1987; Merkley, 1991; Dodd et al., 1996). Caliche and eolianites mark the boundary in the basin interior and the downdip sections of the eastern shelf (cross section B–B′). 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 and Read, 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-trending structural high that caused thinning of as much as 16 m 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 basinward side of the hinge. This structure parallels the northwest-trending part of the Cincinnati arch and the Lasalle anticlinal belt (Fig. 1B).

The transgressive systems tract (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 present only 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 17, 22, and 26). The maximum-flooding surface (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 highstand systems tract (HST) is composed of five throughgoing 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 17, 22, and 26 and correlates laterally with two parasequences at locations 5, 12, and 14. Parasequences 5, 6, 7, and 9 are capped by offlapping 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 (locations 5 and 12). 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—which is the Spar Mountain Member of the Ste. Genevieve—passes westward into marine siliciclastics that prograded east at the top of the sequence (Swann, 1964). On the western shelf, the HST is composed of siliciclastic parasequences bounded by four paleosols that may correlate with the four regional disconformities on the eastern shelf (disconformities C to F). The sequence is bounded at the top by disconformity F. Leetaru (1997) mapped an erosional contact with as much as 24 m of relief at the contact between the siliciclastics of the Aux Vases and the carbonates of the Ste. Genevieve. However, this erosion is not related to the sequence-bounding disconformity F that can be traced through the middle of the siliciclastic unit. The erosion may be the result of scour caused by a prograding tidal delta in the late HST of the sequence. If this erosion were related to a sea-level lowstand, the sequence-bounding paleosol would be absent in locations where erosion occurred.

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 and Joppa Members of the Ste. Genevieve and the Rosiclare Member of the Aux Vases. Sequence 2 is 16 m thick on the western shelf, thickens to 30 m in the basin interior, and thins to 13 to 15 m on the eastern shelf. Sequence 2 is composed of tidally influenced siliciclastics on the western shelf that 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. 8). The TST of sequence 2 consists of parasequence 10 that is composed of patchy eolianites and grainstone capped by a 1–10-m-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 was the first laterally extensive grainstone unit to be deposited on the eastern shelf. The HST is composed of five 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. In many sections on the eastern shelf, the absence of the muddy cap to parasequence 11 suggests either nondeposition or erosion at the base of parasequence 12. Parasequence 13, which underlies disconformity G/G′, consists of multilateral grainstone units and is laterally discontinuous across the basin. The HST is capped by parasequence 14, a disconformity-bounded parasequence composed of tidally influenced siliciclastics and multiple local disconformities; the HST is only present in the basin interior and is capped by a well-developed unconformity (disconformity G). The contact between the Aux Vases and Ste. Genevieve is commonly erosional near the top of this sequence and has also been picked as a sequence boundary (Leetaru, 1997). However, the fact that the regional disconformity can again be traced through the siliciclastic unit suggests that any erosion occurred by tidal scour in the late part of the highstand rather than during sea-level lowstand.

Sequence 3

Sequence 3 is bounded at the base by disconformity G and at the top by disconformity H. This sequence is composed of a single disconformity-bounded parasequence at the top of the Ste. Genevieve Formation in Kentucky and Indiana and the Levias Member of the Renault Formation in Illinois. The lower boundary (disconformity G) is a well-developed, basin-wide paleosol that caps siliciclastics of the Rosiclare Tongue of the Aux Vases. Sequence 3 is predominantly composed of grainstone and minor siliciclastics in the basin interior, eolianites on the southern part of the eastern shelf, and muddy carbonates and grainstone on the northern part of the eastern shelf. The MFS is interpreted to be coincident with the lower sequence boundary. The upper sequence boundary (disconformity H) is the best-developed paleosol in the study interval, which is called the Bryantsville Breccia Bed in Indiana and central Kentucky. In earlier work (Smith, 1996), this unit was called a disconformity-bounded parasequence at the top of sequence 2, but is here reinterpreted as a separate sequence. The progradation of siliciclastics at the top of the underlying unit and the intensity of the brecciation at the top of the sequence suggest that it is a thin updip extension of a sequence that is better developed downdip.

Sequence 4

Sequence 4 is bounded at the base by disconformity H and at the top by disconformity I. Sequence 4 includes the lower part of the Paoli interval, 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 m thick on the western shelf, thickens to 20 m in the basin interior, and thins to 4 m near the Cincinnati Arch and 2 m in the northeastern sections.

Disconformity H is a major disconformity characterized by accumulations of breccia and caliche as much as 4 m 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 4 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 that pinches out onto tectonic highs and is capped by a laterally extensive muddy carbonate–fossiliferous shale unit that 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 that 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 multilateral grainstone deposits that underlie the sequence boundary on the eastern shelf. In the basin interior, parasequences 18 and 19 have skeletal grainstone or 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-highstand parasequences 20 and 21 progradationally offlap and are only present in the basin interior. Sequence 4 is capped by disconformity I.

Sequence 5

Sequence 5 is bounded at the base by disconformity I and at the top by disconformity J. In the basin interior and western shelf, sequence 5 includes the upper part of the Paoli interval, a second disconformity-bounded unit in the Girkin Limestone, and the Downeys Bluff Formation. Sequence 5 is 6 m thick on the western shelf, thickens to 20 m in the eastern part of the basin interior, and thins to 7 m near the Cincinnati arch and to less than 3 m on the northern part of the eastern shelf. The sequence may have been thicker in the basin interior but was eroded during pre-Bethel incision.

Disconformity I is an intermediate unconformity that overlies a thin eolianite in the basin interior (Fig. 8). The TST is composed of parasequence 22 that consists of a thin basal shale in the eastern part of the 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 MFS is picked at the base of the grainstone unit in parasequence 23 because it is the first laterally extensive grainstone unit to cover 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 basinward of the shelves, suggesting that progradational offlap occurred at the top of sequence 5. 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 southern part of the eastern shelf. The sequence is capped by an unconformity (disconformity J) that consists of a well-developed paleosol on the eastern and western shelves that can be traced into an incised valley that trends southwest from south-central Indiana to western Kentucky and is as deep as 75 m (Friberg et al., 1969).

Composite Sequences

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). The updip location of the Illinois basin on the ramp makes it difficult to trace sequence boundaries to determine which extends the farthest downdip. In this setting, composite sequence boundaries can only be distinguished from higher-frequency sequence boundaries by the degree of paleosol development, depth of incision, and/or coincidence with biostratigraphic zone boundaries (Smith, 1996). On the basis of these criteria, the major disconformities at the base of sequence

1 (disconformity A), at the top of sequence 3 (disconformity H), and at the top of sequence 5 (disconformity J) may be composite sequence boundaries (Smith, 1996). In Smith (1996), sequences 1 through 5 were picked as a composite sequence on the basis of the development of the bounding unconformities (A and J) and apparent backstepping of ooid grainstone upward through the interval. Disconformity H was picked as a higher-frequency composite sequence boundary. Disconformities A and H have been interpreted to be composite sequence boundaries in the Appalachian basin (Al-Tawil, 1998), but J is not any better developed than the other sequence boundaries. Thus, the sequences are likely bundled into composites sequences, but the updip location of the Illinois basin and the intensity of the sequence development make it difficult to confidently pick composite sequence boundaries.

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 sedimentary rocks in Australia (Roberts et al., 1995). Brachiopod biostratigraphy was used to correlate with European and North American conodont- and foraminiferal-zones that tied the SHRIMP ages to radiometric dates from Europe (Roberts et al., 1995). On the basis of foraminiferal 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 approximately 3 m.y. (Roberts et al., 1995), but is poorly constrained because of the difficulties inherent in correlating between brachiopod, conodont, and foraminiferal zones on different continents and the error bars on the radiometric dates.

There are five disconformity-bounded sequences in the Ste. Genevieve to Paoli interval and four in the overlying Bethel to Glen Dean interval (Smith, 1996), for a total of nine sequences. Dividing ten sequences into 3 m.y. gives an average period of 333 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 fourth-order (0.1 to 0.5 m.y.) sequences are compatible with times of abundant global ice (Weber et al., 1995). Calculated average duration of the preserved parasequences would be about 37 k.y. in sequence 1, 55 k.y. in sequences 2 and 4, and 83 k.y. in sequence 5. These results suggest that the parasequences are dominantly fifth-order (10 to 100 k.y.) (cf. Weber et al., 1995). Any composite sequences would be third-order (0.5–5 m.y.) (cf. Weber et al., 1995).

Origin and Development of Sequences and Parasequences

Origin of sequences.—

The sequences and sequence boundaries in the Chesterian interval have been interpreted to have formed by local tectonics (Leetaru, 1997), cyclic climate change (Swann, 1963), and eustatic sea-level change (Smith et al., 1995; Smith, 1996). The conclusion that the sequence boundaries likely were caused by fourth-order (0.1–0.5 m.y.) eustatic sea-level changes is based on four facts: (1) by using distinct marker beds and biostratigraphy, the sequence-bounding unconformities in both siliciclastics and carbonates can be correlated across the Illinois basin and in some cases into the Appalachian basin (Smith et al., 1995); (2) the unconformities coincide with biostratigraphic zone boundaries; (3) disconformity-bounded sequences with similar characteristics occur in time-equivalent strata in Kansas (Abegg, 1994), Great Britain (Walkden, 1987), and Kazakhstan (Lehmann et al., 1996); and (4) the duration of the sequences roughly coincides with the Milankovitch long-term (~400 k.y.) eccentricity signal and the calculated period for the overlying Pennsylvanian cyclothems (Heckel, 1986). If the sequences were made by local tectonic uplift, the sequence boundaries would be confined to the area of uplift, but instead they are basin-wide. If the unconformities were caused by a migrating peripheral bulge, they would be diachronous, but the fact that the unconformities coincide with biostratigraphic zone boundaries suggests that they are relatively synchronous across the basin. Sequence-bounding disconformities formed during periods of high, moderate, and low differential subsidence in the Illinois basin (Smith, 1996). Regional uplift of the Illinois and Appalachian basins could have caused the unconformities, but there is no known mechanism that would cause rhythmic uplift and subsidence of the craton with a period on the order of 400 k.y. If the sequences were made by variations in siliciclastic sediment supply caused by climate change (Swann, 1963), the unconformities would be present only in areas of mixed carbonates and clastics, but they can be traced from mixed carbonate-clastic intervals into carbonate-dominated strata.

Sequence development.—

The magnitude of sea-level fluctuations that produced the 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 northern part of the eastern shelf, the sea-level fall was sufficient to expose the basin interior as subsidence continued. The exposed profile can be roughly reconstructed by using the thickness of the TST of sequence 1, which is 27 m thick in the basin interior, thinning to less than 1 m on the northern part of the eastern shelf. These data suggest that a sea-level change of about 30 m 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 may not have fallen far below the platform. The incision within the Aux Vases Formation may have been caused by prograding tidal deltas or local tectonics (Leetaru, 1997), but these erosion surfaces are not sequence boundaries because they do not correlate with paleosols in areas of no incision. Thus, sequences 1 through 4 likely were made by moderate-amplitude sea-level fluctuations of 30 m or less. In

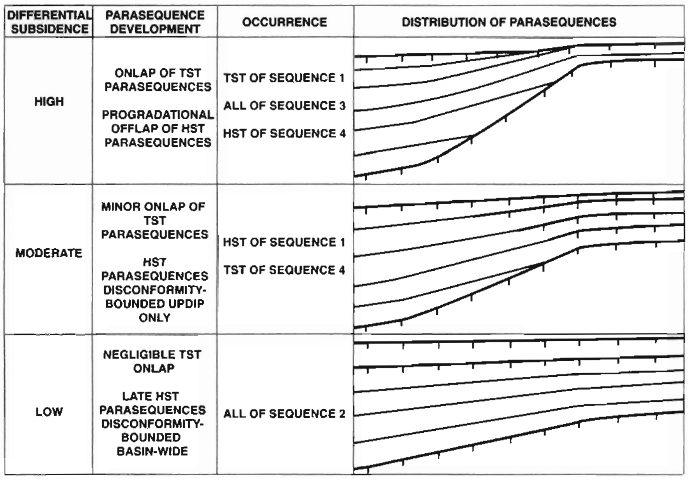

FIG. 9.—Effects of variations in differential subsidence of the eastern shelf and the basin interior on parasequence distribution within the sequences. TST = transgressive systems tract, HST = highstand system tract.

contrast, sequences in the overlying middle Chesterian mixed carbonate-clastic sequences also have incised valleys (indicating as much as 75 m of erosion) as well as deeper-water carbonate facies that together suggest that the magnitude of sea-level changes increased to as much as 30–95 m (Smith, 1996).

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 composed of one to four parasequences onlapped the shelves. Onlap was greatest in the transgressive systems tracts during periods of high differential subsidence between the shelves and the basin interior (Fig. 9) (Smith, 1996). 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 generate strong tidal currents. The lack of deeper-water facies indicates that water depths never exceeded those of ooid and skeletal grainstone lithotopes. In the HST, three to five aggradational to progradational parasequences were deposited as relative sea level peaked and began to fall. Progradational offlap of parasequences occurred in the HST during periods of high differential subsidence in the basin, and aggradation occurred during periods of relatively low differential subsidence (Fig. 9) (Smith, 1996).

Glacio-eustatic origin of parasequences.—
The lateral extent and frequency of the regionally developed parasequences indicate that these strata were produced by fifth-order (10–100 k.y.) glacio-eustatic sea-level changes. Neither local tectonics nor autocyclic processes could have produced basin-wide parasequences that appear to have formed in tens of thousands of years. The presence of tidal sand ridges 8–10 m high (Carr, 1973; Cluff and Lineback, 1981) suggest that rapid sea-level rises of as much as 10 m 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 not much more than 10 m.

Assuming that the sequences are 400 k.y. in average duration, and from 2 to 63 m 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 as much as 10 m 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 m 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.—
Parasequence deposition was initiated by high-frequency sea-level rises that flooded both disconformable and conformable surfaces on shallow-water mudstone that had shoaled to, or near to, sea level (Fig. 10). 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 and packstone accumulated to form a broadly undulating submarine topography of ridges and passes. Tidal energy appears to have diminished in updip parts of the ramp where muddy carbonates were deposited in widespread lagoons. As ooid and skeletal grainstone filled much 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. Tidal-flat laminites were not deposited at the tops of most parasequences because sea-level fall caused rapid movement of the shoreline seaward across the shelf, stranding any tidal flats updip. In areas with low accommodation rates, parasequences were capped by patchy eolianites and disconformities. These disconformities 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 the dip-parallel orientation of the ooid shoals. Thus, facies were as much dependent

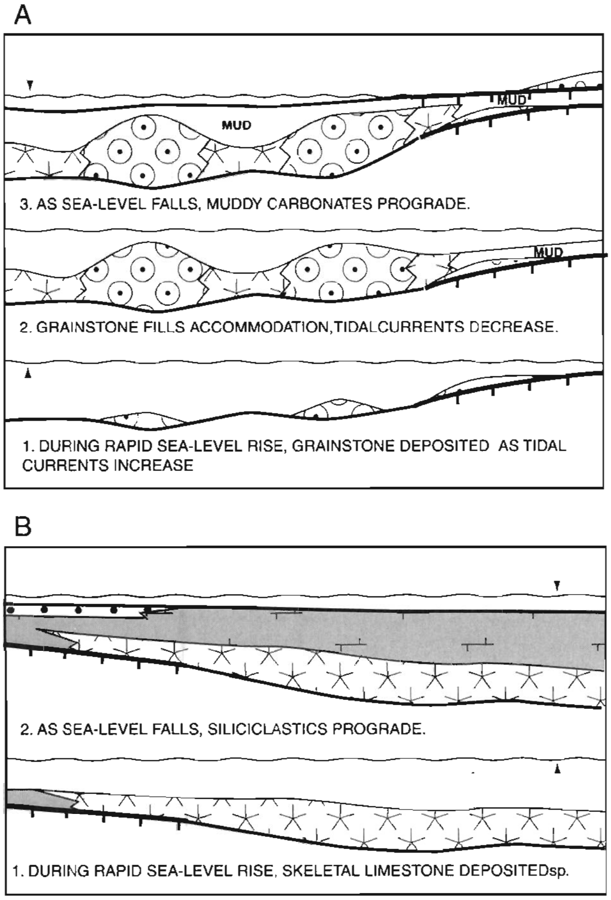

FIG. 10.—(A) Development of carbonate parasequences: (1) Rapid glacio-eustatic sea-level rise promotes increase in tidal energy and deposition of ooid ridges and skeletal banks. (2) Ooid and skeletal grainstone fills much of the accommodation space, and tidal energy diminishes. (3) Muddy carbonates prograde, filling topography and forming laterally extensive sheets. (B) Skeletal limestone-shale parasequence development: (1) During sea-level rise, successive siliciclastic layers are deposited farther updip toward the source, and carbonate-producing organisms thrive and make skeletal limestone. (2) As sea level falls, shale progrades from the west-northwest and caps parasequences. In areas close to the siliciclastic source, sandstone progrades over shale.

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 to ooid grainstone to muddy carbonate and shale to eolianite would have been generated by upward shallowing within prograding parasequences. Instead, skeletal grainstone occurs above and below and passes laterally into ooid grainstone as a product of dip-parallel grainstone geometries.

Most 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 many 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 mudstone units (e.g., parasequences 3, 4 and 5) that 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. Such autocycles are clearly evident with the detailed, closely spaced cross sections but could not have been distinguished from regional parasequences on the basis of a single stratigraphic section.

Late Mississippian Sea-Level Curve

Basin-wide paleosols and facies-stacking patterns in the Mississippian succession can be used to construct a detailed relative-sea-level curve (Fig. 11). Ross and Ross (1988) constructed an onlap curve for the late Paleozoic but only correlated large-scale sequences (Fig. 11). 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 on the basis of the seaward extent of siliciclastics within the Illinois basin (Fig. 11). Swann's curve roughly parallels this paper's fourth-order sequence curve, because the siliciclastic units commonly prograded toward the tops of the sequences. 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.

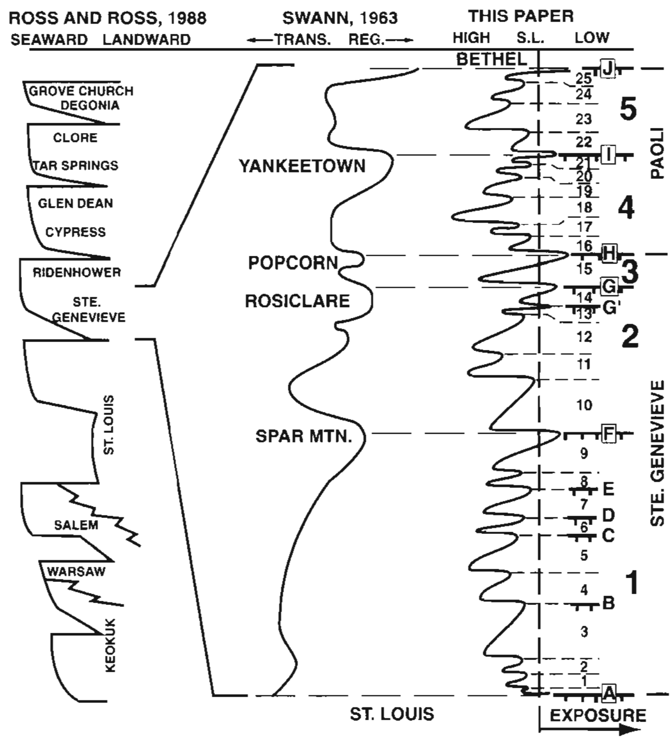

FIG. 11.—Comparison of onlap curve from Ross and Ross (1988), transgression-regression curve of Swann (1963), and relative sea-level curve prepared for this paper. Disconformities are lettered from A to J, sequences are numbered from 1 to 5, and parasequences are numbered from 1 to 25.

Stratigraphic Signature of the Greenhouse to Icehouse Transition

The first ice-rafted deposits of the late Paleozoic glaciation of Gondwana suggest that major continental ice sheets began to grow during the earliest Namurian or middle Chesterian (Fig. 12) (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 age of the Ste. Genevieve to Paoli interval is late Visean (Baxter and Brenckle, 1982). Thus the study interval spans the later part of the transition from a relatively ice-free globe in the Early to middle Mississippian to the Pennsylvanian–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 an increase in the amplitude of sea-level fluctuations forming these sequences and those in the overlying middle Chesterian interval (Fig. 13) (Smith, 1996). Sequences 1 through 4 are bounded by surfaces with minimal erosion and were likely formed by sea-level fluctuations of ~30 m or less. In contrast, sequences 5 through 10 have bounding surfaces with as much as 75 m of incision, which may have been produced by sea-level changes of as much as 95 m (Smith, 1996). Time-equivalent strata in Great Britain show a similar upward change from nonerosional to erosional boundaries that has been 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.

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). 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 third-order sequences (Goodwin and Anderson, 1985; Koerschner and Read, 1989; Goldhammer et al., 1990; Elrick, 1991; Read, 1995). 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 third-order sequence boundaries.

The parasequences formed during the transition period of the Late Mississippian covered by this study generally lack

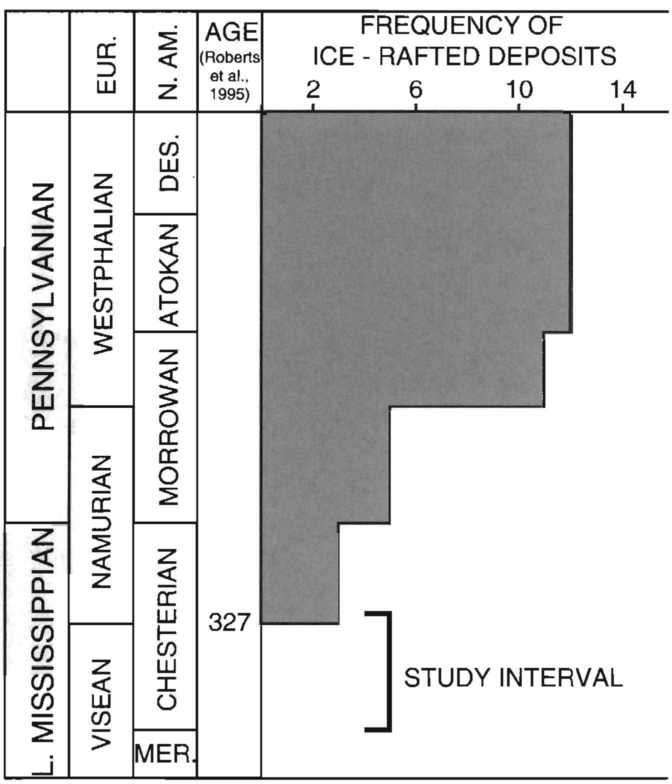

FIG. 12.—Global climate setting during Late Mississippian and Pennsylvanian. Graph shows the number of occurrences of ice-rafted deposits worldwide leading up to late Paleozoic glaciation of Gondwana (modified from Frakes et al., 1992; age from Roberts et al., 1995). The study interval occurs just below the first glacial deposits and during the transition from little global ice to major continental glaciation. MER. = Meramecian, DES. = Desmoinesian.

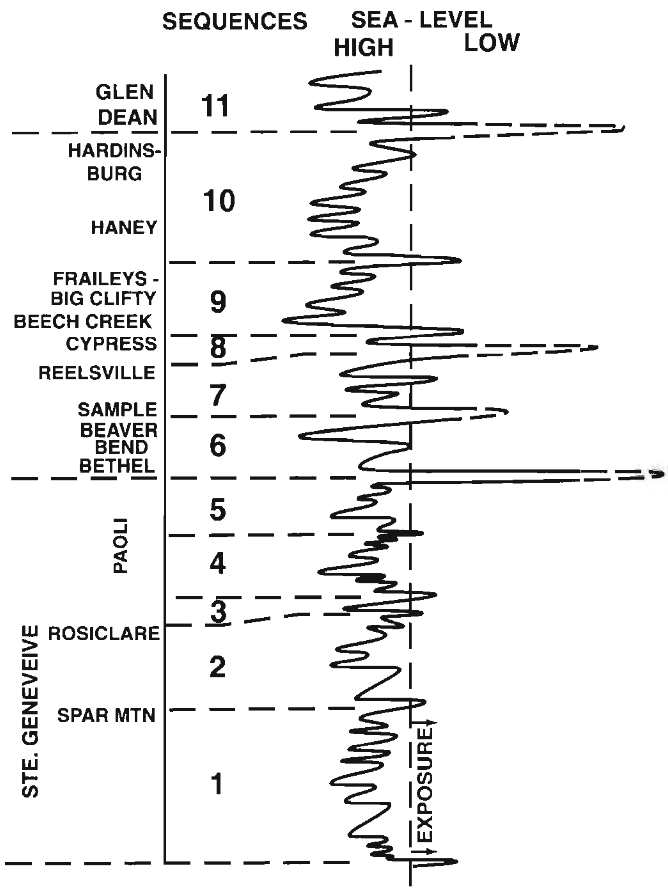

FIG. 13.—Composite sea-level curve for the Ste. Genevieve to Paoli interval and the Bethel through Glen Dean interval (sequences 6 to 10) (Smith, 1996). The amplitude of sea-level changes increased through the Chesterian in response to onset of major continental glaciation on Gondwana.

intertidal-supratidal caps, and disconformities are developed on subtidal lagoonal and shoal facies of fourth-order and some updip fifth-order units. Third-order composite sequences are difficult to recognize because of the frequency of the disconformities and the magnitude of the moderate-amplitude fourth-order sea-level changes.

Global icehouse times such as the Pennsylvanian and Pleistocene were characterized by high-frequency fourth-order sequences (Heckel, 1986; Goldhammer et al., 1991; Weber et al., 1995; Mitchum and Van Wagoner, 1991; Read, 1995) This paper shows that this type of sequence development is evident as early as the Late Mississippian. The oxygen isotope record suggests that the dominant episodes of waxing and waning of Pleistocene ice sheets correlate with the 41-k.y. obliquity and the 100-k.y. eccentricity periods (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 during 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.

The increase in siliciclastic content from sequences 1 to 4 continues into the overlying sequences and 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 time-equivalent strata in Great Britain that 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.

CONCLUSIONS

Regional disconformities were previously thought to be of relatively local extent in the Ste. Genevieve to Paoli interval in the Upper Mississippian (Chesterian) of the Illinois basin. However, these disconformities, which probably are the result of moderate-amplitude fourth-order eustasy, can be mapped regionally to define a high-resolution sequence stratigraphic framework of fourth-order sequences and component parasequences.

Five disconformity-bounded fourth-order sequences (2 to 63 m thick) can be traced across the Illinois basin in the Ste. Genevieve to Paoli interval. The boundaries of the lower four sequences are marked by caliche, breccia, red mudrock paleosols, and carbonate eolianites. The boundary at the top of sequence 5 is marked by a well-developed paleosol that can be traced into an erosional unconformity with as much as 75 m of incision. Lowstand systems tracts are not preserved because of the updip position of the study area 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 of regional grainstones that 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 probably 30 m or less for the nonerosional sequence boundaries but increased to as much as 95 m to create the erosional unconformity 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.

Parasequences contain the individual reservoirs and consist of, from base to top, (1) a basal disconformity or conformable boundary, (2) ooid (reservoir) and skeletal grainstone, (3) muddy carbonate and/or fossiliferous shale caps; and (4) a capping 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 that were deposited in laterally extensive sheets and swale fills. This process 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 shows how individual reservoirs are related. The parasequences can be correlated between areas with vastly different subsidence rates; thus they were likely formed by fifth-order sea-level fluctuations, superimposed on the fourth-order eustatic signal.

Carbonate reservoir facies of Mississippian age worldwide 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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APPENDIX 1. OUTCROP AND CORE LOCATIONS

The following are the outcrop and core locations for the data in the cross sections in Figure 4. Abbreviations: Alt.—alternate route, Co.—county, Ill.—Illinois, Ind.—Indiana, Ky.—Kentucky, S.R.—state road.

1 Core, Madison Co., C-141, Illinois State Geological Survey Core Facility, Champaign, Ill.

2 Core, Madison Co., C-140, Illinois State Geological Survey Core Facility, Champaign, Ill.

4 Core, Union Co., C-13619, Illinois State Geological Survey Core Facility, Champaign, Ill.

5 Outcrop, Anna Quarry, East Side of Anna, Ill., northwest of junction of S.R. 51 and S.R. 146

6 Outcrop, I-57, milepost 27.4, south of Anna, Ill., exit

7 Outcrop, Cypress Quarry, Columbia Quarries Inc., S.R. 37, 3 mi (about 5 km) south of Cypress, Ill.

8 Core, Pope County, C-46, Illinois State Geological Survey Core Facility, Champaign, Ill.

10 Core, Hardin Co., C-12781, Illinois State Geological Survey Core Facility, Champaign, Ill.

12 Core, Hardin Co., C-12782, Illinois State Geological Survey Core Facility, Champaign, Ill.
13 Core, Hardin Co., from Ozark-Mahoning Fluorspar mine near Denton, Ill.
14 Outcrop, Hardin County Materials Company, Pit 2, junction of S.R. 1 and S.R.146, Cave-in-Rock, Ill.
15 Outcrop, Cave-in-Rock Quarry, Martin-Marietta, on county roads east-northeast of Cave-in-Rock, Ill.
16 Outcrop, Three Rivers Quarry, Martin-Marietta, U.S. 60, 6.25 mi (10 km) northeast of Smithland, Ky.
17 Outcrop, Fredonia Quarry, Denny and Simpson, U.S. 641, 2.5 mi (4 km) south of Fredonia, Ky.
18 Outcrop, Princeton Quarry, Kentucky Stone Company, S.R. 91, 2.75 mi. south of Princeton, Ky.
21 Core, Caldwell Co., C-220, Kentucky Geological Survey Core Facility, Lexington, Ky.
22 Outcrop, S. Hopkinsville Quarry, Rogers Group, U.S. Alt. 41, 4.5 mi (about 7 km) south of Hopkinsville, Ky.
23 Outcrop, Christian Quarries Quarry (abandoned), U.S. 68, 0.5 mi (0.8 km) east of Hopkinsville, Ky.
24 Outcrop, Todd Quarry, Kentucky Stone Company Quarry, U.S. 68, 7 mi (11 km) west of Elkton, Ky.
25 Outcrop, new road cut for U.S. 68 bypass, west of Russelville, Ky.
26 Outcrop, Russelville Quarry and Core, Kentucky Stone Company, S.R. 79, 1 mi (1.6 km) east of Russelville, Ky.
29 Outcrop, road cut at milepost 9.1 on Green River Parkway north of Bowling Green, Ky.
30 Outcrop, Rockfield Quarry, Kentucky Stone Company, U.S. 68, Rockfield, Kentucky
33 Outcrop, road cut on I-65 N at milepost 48 north of Park City, Ky., exit
34 Outcrop, E. Park City Quarry (abandoned), Old Bardstown Road, 2 mi (about 3 km) east of Park City, Ky.
35 Outcrop, Cave City Quarry, Scotty's Paving Company, Route 90, about 5 mi (8 km) east of Cave City, Ky.
36 Outcrop, road cut on I-65 south, 1 mi (1.6 km) south of Munfordville, Ky., exit
37 Outcrop, road cut on I-65 North, 0.5 mi (0.8 km) north of Munfordville, Ky., exit
38 Outcrop, Upton Quarry, Kentucky Stone Company, Quarry Road, Upton, Ky.
40 Core, Hardin Co., C-117, Kentucky Geological Survey Core Facility, Lexington, Ky.
41 Outcrop, Cecelia Quarry, Larry Glass Paving, S.R. 62, 3 mi (about 5 km) west of Cecelia, Ky.
42 Core, Breckenridge Co., C-118, Kentucky Geological Survey Core Facility, Lexington, Ky.
43 Outcrop, Stephensburg Quarry (abandoned), on dirt road 2 mi (about 3 km) north of Stephensburg, Ky.
44 Outcrop, Irvington Quarry, Kentucky Stone Company, S.R. 477, 2 mi (about 3 km) northwest of Irvington, Ky.
45 Outcrop, Battletown Quarry, Kosmos Cement Company, S.R. 228, 0.5 mi (0.8 km) northwest of Battletown, Ky.
46 Outcrop, Cape Sandy Quarry, Pit 1, Mulzer Crushed Stone Inc., 4 mi (about 6.5 km) east of Alton, Ind.
47 Core, Harrison Company, C-339, Indiana Geological Survey Core Facility, Bloomington, Ind.
48 Core, Perry Company, C-303, Indiana Geological Survey Core Facility, Bloomington, Ind.
49 Outcrop, Tower Quarry, Mulzer Crushed Stone Company, 2 mi (about 3 km) north of Leavenworth, Ind.
50 Outcrop, road cut at milepost 99 on I-64, 7 mi (about 11 km) west of Corydon, Ind., exit
51 Outcrop, Corydon Quarry, Corydon Crushed Stone Inc., 2.5 mi (4 km) northwest of Corydon, Ind.
52 Outcrop, Robertson Quarry, Robertson Crushed Stone Inc., S.R. 64, 1 mi (1.6 km) west of Depauw, Ind.
56 Core, Crawford Co., C-626, Indiana Geological Survey Core Facility, Bloomington, Ind.
57 Core, Orange Co., C-338, Indiana Geological Survey Core Facility, Bloomington, Ind.
60 Core, Orange Co., C-651, Indiana Geological Survey Core Facility, Bloomington, Ind.
62 Outcrop, W. Paoli Quarry, Cave Quarries Inc., U.S. 150, 3.5 mi (about 5.5 km) northwest of Paoli, Ind.
63 Core, Martin Co., C-214, Indiana Geological Survey Core Facility, Bloomington, Ind.
64 Outcrop, Mitchell Quarry, Rogers Group, S.R. 60, 5 mi (8 km) west of Mitchell, Ind.
65 Core, Lawrence Co., C-340, Indiana Geological Survey Core Facility, Bloomington, Ind.
67 Core, Martin Co., C-797, Indiana Geological Survey Core Facility, Bloomington, Ind.
68 Outcrop, Sieboldt Quarry, Rogers Group, county roads, 5 mi (8 km) north of Oolitic, Ind.
69 Core, Greene Co., C-487, Indiana Geological Survey Core Facility, Bloomington, Ind.
70 Core, Monroe Co., C-672, Indiana Geological Survey Core Facility, Bloomington, Ind.
71 Outcrop, Bloomington Quarry, Rogers Group, Oard Road, 4 mi (about 6.5 km) west of Bloomington, Ind.
73 Core, Clay Co., C-444, Indiana Geological Survey Core Facility, Bloomington, Ind.
74 Outcrop, Putnamville Quarry, Kentucky Stone Company, S.R. 343, 1 mi (1.6 km) west of Cloverdale, Ind.
75 Core, Putnam Co., C-105, Indiana Geological Survey Core Facility, Bloomington, Ind.