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Middle Devonian sedimentary cycles and sequences in the northern Appalachian Basin

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ABSTRACT

At least three time orders of sedimentary cyclicity are observed in Middle Devonian (Eifelian-Givetian) foreland basin deposits of New York State, Ontario, and Pennsylvania. These cycles are widespread, suggesting a eustatic origin. However, regional variations in thickness, facies content, and symmetry indicate that dynamics of sediment supply, subsidence rate, and resulting availability of accommodation space controlled the appearance of preserved eustatic cycles in a foreland basin.

Small-scale (fifth- and sixth-order) cycles, here termed parasequences, correspond to couplets of shale and concretionary limestone beds in western New York (offshore) facies and to asymmetric, upward-coarsening cycles in east-central New York and central Pennsylvania (nearshore) deposits. These cycles commence with shell pavements that record marine flooding surfaces with associated sediment starvation. The aggradational interval is capped by a shell bed that reflects shell and/or sand concentration due to winnowing conditions associated with sediment bypass.

Cyclothem-scale (fourth-order) cycles, parasequence sets, are cycles that include bundles of several parasequences. Parasequence sets contain a greater net range of facies and are characterized by multiple flooding surface shell beds typically capped by a widespread condensed interval representing a surface of maximum flooding. Similarly, where strongly asymmetrical cycles are developed in areas of high sediment supply, two or more component shell beds may become stacked or removed by erosion owing to lack of accommodation space. Third-order cycles, herein designated large-scale sequences, contain several fourth-order cycles. These correspond approximately (but not precisely) to several formations currently recognized in the Middle Devonian section. Widespread bounding unconformities mark the bases of several of these cycles, and several major bone- and pyrite-rich beds correspond to the maximum flooding surface.

Enigmatic beds recording abrupt shallowing, sedimentary condensation, and subsequent deepening are observed to punctuate the long, late-highstand shoaling phase of many cycles. These units, herein designated precursor beds, may be due to superposition of smaller cycles over larger, but are more likely due to complex downlap processes associated with lateral basin migration and depocenter shifts.

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INTRODUCTION

An important promise of the sequence stratigraphic paradigm is that sequence boundaries may serve to subdivide the stratigraphic record into process-related units which have chronostratigraphic significance and correlate with global sea-level events (see Vail et al., 1977, 1991; Haq et al., 1987; Van Wagoner et al., 1988, 1990). Detailed global correlations have been achieved for many passive margin sequences, from which the sequence approach was perfected. Eustasy has also been found to be important in explaining cyclic facies repetitions in epeiric seas, particularly in the North American midcontinent (Busch and Rollins, 1984; Heckel, 1986; Johnson et al., 1985; Ross and Ross, 1988). However, application of the sequence approach is more challenging in foreland basins where tectonics exert strong control on sediment supply and subsidence patterns (Jordan and Flemings, 1991). Processes of lithospheric flexure near craton margins may leave a major overprint on facies successions, modifying or possibly obscuring the cyclic motif (see Quinlan and Beaumont, 1984; Ettensohn et al., 1988; Klein, 1989; Jordan and Flemings, 1991).

Foreland basins yoked to collisional thrust loading events develop thick sedimentary successions and may record the effects of eustasy, isostasy, and basin dynamics. Herein we apply the principles of sequence stratigraphy to a cyclic succession of mixed marine siliciclastic and carbonate deposits of eastern North America comprising the late Eifelian to Givetian (Middle Devonian) Hamilton and Tully Groups. These strata accumulated in an elongate trough, the Appalachian foreland basin, that shifted cratonward during an episode of tectonic thrust loading (Fig. 1). Thus, they provide a detailed case study of sedimentary cyclicity in the context of tectonic basin modification and major orogenic sediment supply.

Within these strata erosional unconformities, corresponding to both relative sea-level lowstands and to sea-level rise events, are recognizable and correlatable. The distribution of such features is region specific within the basin and reflects local episodic changes in sediment supply and subsidence. Sequences and their component parasequences correlate circumbasinally, but vary in thickness, appearance, and component facies as a result of these local factors.

In previous papers (Brett and Baird, 1985, 1986), we described marine cyclothems within the Middle Devonian Hamilton Group. These cycles range from a few decimeters to about 15 m (0.5–50 ft) in thickness that are correlatable throughout much of the New York outcrop belt (Fig. 2) and adjoining Pennsylvania. These cycles display a consistent succession of lithofacies and biofacies in a given part of the basin. In offshore, relatively sediment starved areas of Ontario, Ohio, and western New York, the cycles are dominated by alternations of black to gray mudrock and thin calcareous beds. More proximal to siliciclastic sources, cycles in central New York and Pennsylvania are

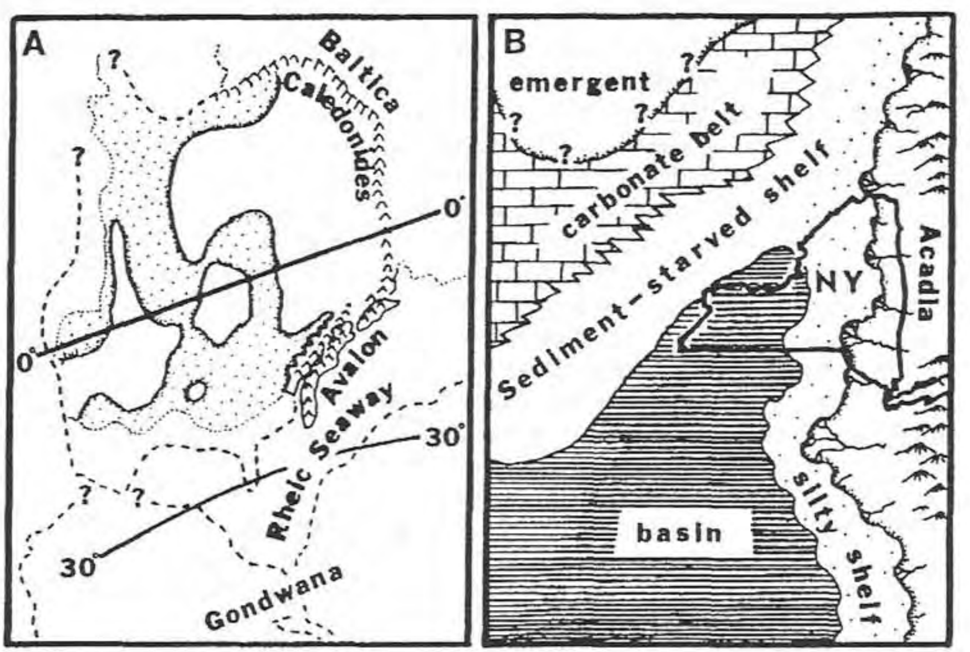


Figure 1. A: Inferred Middle Devonian paleogeography of Laurentia (ancestral North America) during deposition of the Givetian Hamilton Group. B: Paleogeography of New York State and adjacent regions. Note region of differential subsidence at the northern end of the Devonian foreland basin. This basin separates a western shelf region characterized by reduced terrigenous sediment input and minor carbonate production from a central New York shelf region characterized by higher turbidity and significant sediment bypass into the trough. Maximum water depth conditions and relative sediment aggradation rates occurred in western Pennsylvania and west-central New York.

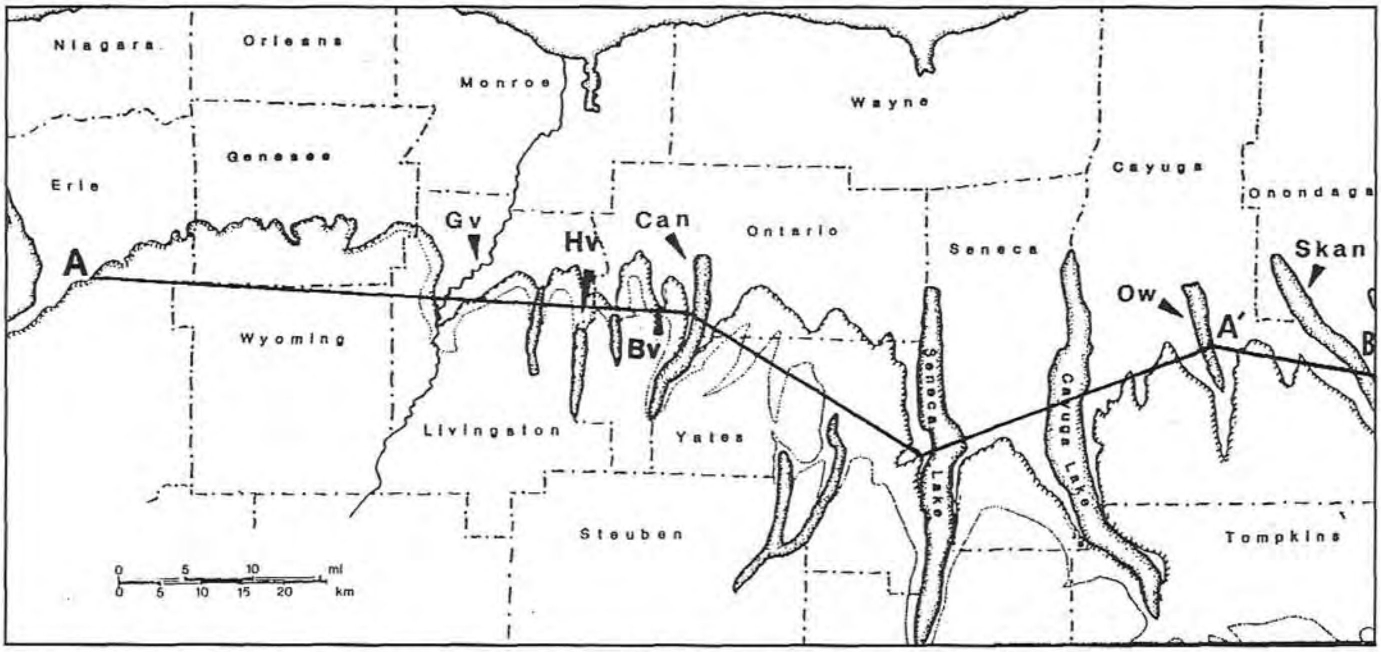


Figure 2. Study area in western and central New York State. Outcrop belt of the Middle Devonian (Givetian) Hamilton Group is marked. Abbreviations: Gv = Genesee Valley; Hv = Honeoye Valley; Bv = Bristol Valley; Can = Canandaigua Lake; Ow = Owasco Lake; Skan = Skaneateles Lake. Line A-A′ shows approximate cross sections of Figures 3 and 4; line A′-B shows cross section of Figure 8.

manifest as coarsening-upward shale to siltstone or sandstone successions. Upward-coarsening (shoaling) cycles occur in areas proximal to sediment sources where rapid sediment influx locally exceeded subsidence during times of relative sea-level drop. The effects of submarine erosion and sediment bypass probably explain the asymmetry of Middle Devonian upward-coarsening cycles throughout central New York and Pennsylvania.

Herein, we recognize a hierarchy of cycles comparable to that discussed by Busch and Rollins (1984) for the mid-continent Carboniferous, and Brett et al. (1990a) for the Silurian of the Appalachian basin. These cycles are differentiated in the context of sequence stratigraphy as sixth-order parasequences, fourth-order parasequence sets or subsequences, and third-order depositional sequences. We outline the generalized facies succession and motif of each scale of cycle, and describe the predictable facies successions in cycles in response to regionally variable patterns of sediment supply and turbidity within the foreland basin. In particular, we show that cycle thickness, cycle symmetry, and facies content are dynamically controlled by the interaction of water depth, sediment supply, and wave energy.

We discuss the formation of erosion surfaces and condensed sections (through either sediment bypass or through sediment starvation) in predictable parts of cycles. Diastems and associated lag deposits are discussed in a context of sea-level change and variation in sediment supply. A new concept, the *precursor bed*, is advanced and explained as a regionally extensive condensed sedimentary unit developed during late highstand sedimentation.

GEOLOGICAL SETTING

Deposits discussed include the Middle Devonian Hamilton Group (see Cooper, 1930, 1933, 1957; Rickard, 1975, 1981; Brett and Baird, 1985, 1986), composed predominantly of marine siliciclastic sediments with a minor carbonate component, and the overlying carbonate-dominated Tully Limestone (see Heckel, 1973). These late Middle Devonian (late Eifelian to Givetian) sediments include marine shelf to basin facies which compose the older portions of the Catskill siliciclastic wedge (Woodrow, 1985; Woodrow et al., 1988). This clastic wedge (Fig. 1) records the erosional product of the second tectophase of the Acadian orogeny (Ettensohn, 1985, 1987).

Tectonic loading during the Devonian Acadian orogeny generated a foreland basin and exported sediment from tectonic source areas to various depocenters within the basin (Kent, 1985; Ettensohn, 1985, 1987; Ettensohn et al., 1988).

During Hamilton Group deposition, the foreland basin was being modified by Acadian tectonics; we observe a progressive westward shift of the structural trough (Fig. 3). Initially, during Marcellus deposition, the depocenter lay in the Hudson Valley region with deepest water areas slightly to the west in east-central New York; during deposition of the Skaneateles and Ludlowville Formations, this trough was centered in the central Finger Lakes region (Figs. 2 and 3); by the time deposits of the medial part of the Moscow Formation (Kashong Shale) were accumulating, the axis and associated depocenter had shifted to the Genesee Valley region (Fig. 3); during late Moscow (Windom Shale) deposition,

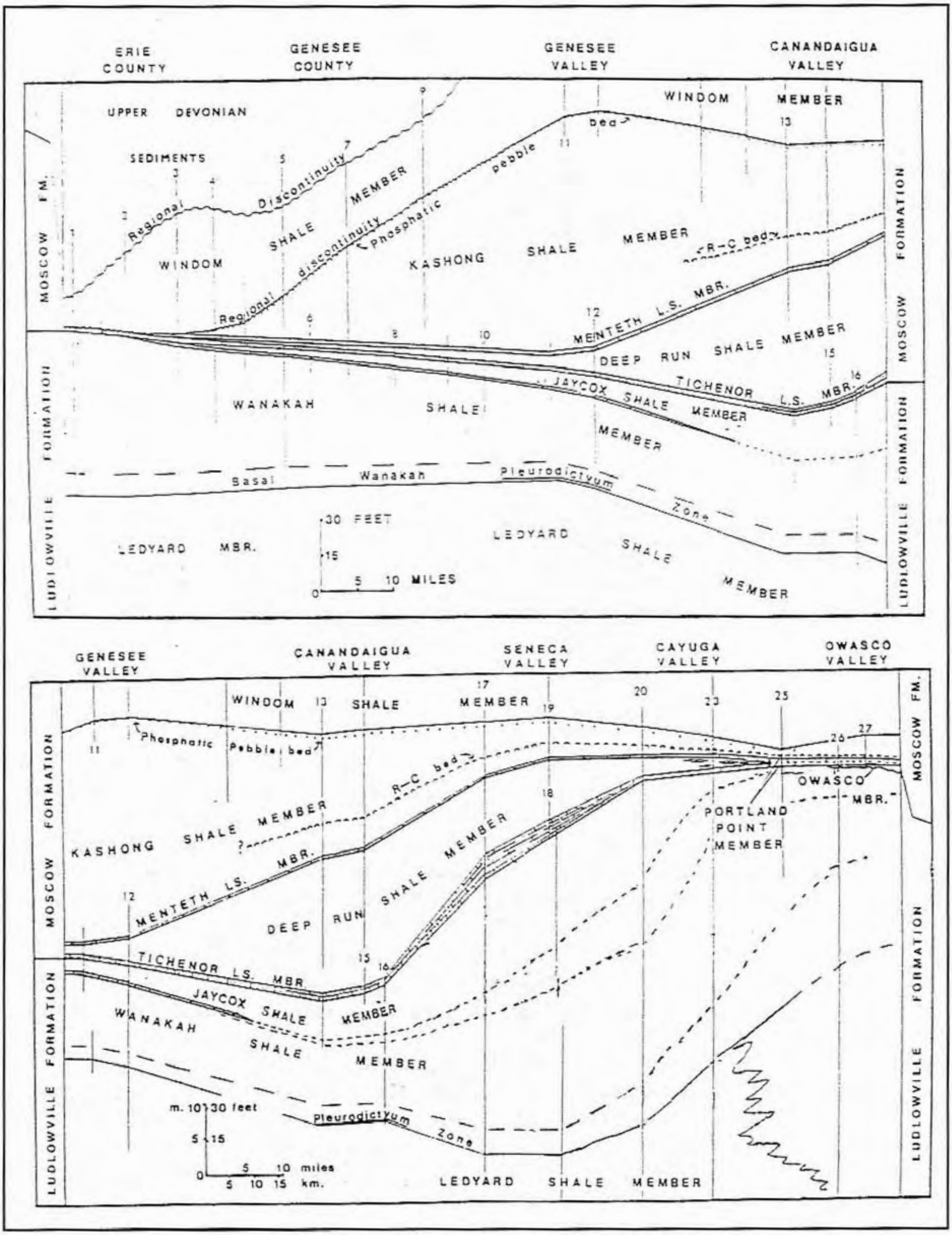


Figure 3. Westward migration of the depocenter during deposition of the middle to late Givetian Ludlowville and Moscow Formations of the Hamilton Group. Vertical lines represent measured sections; for descriptions of numbered localities see Baird (1979). Upper figure shows western sections from Erie County to Canandaigua Lake. Lower figure shows eastern sections from Genesee Valley to Owasco Lake; note slight overlap; see Figure 2 for cross section line. Note thin marker beds that serve as timelines bracketing terrigenous sediment wedges; these thin beds reveal the westward isostatic migration of the trough in a time-lapse fashion (modified from Baird, 1979).

a prominent trough was positioned in Erie County near Buffalo (Baird and Brett, 1981; Brett and Baird, 1982, 1986; Brett et al., 1990b). This pattern corresponds to the model of Ettensohn et al. (1988), which incorporates the Hamilton Group as the second of several collisional tectophases triggered by Acadian collision events. The Hamilton tectophase was associated with isostatic flexural deepening of the basin, probably due to thrust loading, as well as lateral shifts in basin axis. Isostatic shifting of basin axis and depocenter positions thus complicates interpretation of representative facies motifs in New York Devonian deposits.

The duration of the Hamilton strata is somewhat equivocal (Rickard, 1975, 1981). The lower part of the Marcellus Formation (Union Springs–Cherry Valley Members) and the underlying Seneca Member of the Onondaga Limestone represent the last three of five conodont zones in the Eifelian Stage (Fig. 4). In terms of thickness and number of major sedimentary cycles, the Seneca–lower Marcellus seems to constitute about a third of the Eifelian, or about 2 to 2.5 m.y. (using estimates of 6 to 8 m.y. for duration of the Eifelian; McKerrow et al., 1985; Gale, 1985; Harland et al., 1990). A precise date of 390 ± 0.5 Ma from a K-bentonite bed at the base of the Seneca Member (Roden et al., 1990) suggests that the Eifelian-Givetian boundary (at or slightly above the Cherry Valley Limestone in the lower part of the Marcellus Formation) should be about 387 to 388 Ma; this value is in reasonable accord with a U/Pb date of 384 ± 9 Ma obtained from middle Marcellus black shales in Pennsylvania (Bofinger and Compston, 1967). The upper part of the Hamilton Group (upper Marcellus, Skaneateles, Ludlowville, and Moscow Formations) represents the upper *ensensis* (or *hemiansatus* zone, as now defined; House and Walliser, 1993) and most of the *varcus* conodont zones (Klapper, 1981), and thus probably about two-thirds of the Givetian Stage. Nearly all sources list the age of the Givetian-Frasnian boundary as 374–377 Ma (McKerrow et al., 1985; Gale, 1985; Harland et al., 1990). Thus a duration of 9 to 10 m.y. for the late Eifelian to Givetian Hamilton Group and Tully Limestone seems reasonable. House (1992, 1995) estimated the duration of the Givetian Stage as 6.5 to 7.0 m.y.; using cyclostratigraphy and the inference of axial precession (~19 ka) for decimeter scale limestone-marl rhythms in pelagic carbonates in England, France, and Morocco. House (1995) estimated the duration of Givetian conodont zones. For the Hamilton-Tully portion of the Givetian he indicates the following durations: *hemiansatus* zone (upper Oatka Creek Shale): 0.43 m.y.; lower *varcus* subzone (= ?Skaneateles, Ludlowville, and lower Moscow formations): 4.05 m.y.; middle *varcus* subzone (= upper Moscow–lower Tully Limestone): 0.92 m.y.; upper *varcus* subzone (= uppermost Tully): 0.21 m.y.; these estimates suggest a total duration of about 5.6 m.y. for the upper Hamilton and Tully. Considering that the late Eifelian may span about a quarter of the Givetian, the interval considered in this chapter (Hamilton-Tully) is probably 7 to 8 m.y. in duration.

During the Middle Devonian, deposition in central and western New York (Fig. 2) was characterized by an eastward-thickening siliciclastic wedge, the stratigraphic units of which can be correlated cratonward across the foreland basin allowing for study of a spectrum of facies from platform carbonates to nearshore siltstones and sandstones (Rickard, 1975, 1981) (Figs. 4 and 5). Regional thin carbonate units, first successfully used to subdivide the Hamilton by Cooper (1930, 1957), form the shallowest-water facies of third-order regressive-transgressive cycles (Brett and Baird, 1985; Brett et al., 1990b); bioclastic limestone beds (e.g., Stafford and Centerfield Limestones; Mayer, 1985; Savarese et al., 1986; Fig. 2) in western New York are found to be stratigraphically equivalent to nearshore calcareous siltstone and sandstone units (Mottville, Chenango) in east-central New York (Gray, 1984; Grasso, 1986; Brett and Baird, 1985, 1986; Fig. 4A).

In western New York (Finger Lakes area–Lake Erie region; Fig. 4) the lower part of the Hamilton Group, comprising the Marcellus and Skaneateles Formations, is dominated by dark gray to black shales with thin intervals of gray mudstone and concretionary limestone (Fig. 4). In contrast, the upper Hamilton Group, represented by the Ludlowville Formation and overlying Moscow Formation, is characterized by variably fossiliferous, medium gray mudstone deposits with numerous thin limestone beds and concretion bands (Fig. 4). The regional stratigraphy of the Hamilton, as a whole, has a conspicuous "layer cake" aspect along the western outcrop belt, with numerous limestone beds, shell layers, discontinuities, and concretion bands being traceable across the entire width of the region (~350 km). This pattern probably reflects parallelism of the modern outcrop belt with depositional strike along the northwest shoulder of the basin (Fig. 5; Brett and Baird, 1982; Baird and Brett, 1983; Brett, 1986; Landing and Brett, 1991; and papers therein).

In the west-central Finger Lakes trough region (Seneca Lake–Owasco Lake region; Figs. 2 and 5) there are notable variations in the thickness of component divisions, and conspicuous lateral facies changes are observed at numerous levels (see Brett et al., 1986b; Mayer et al., 1994). Facies record generally deeper water conditions relative to east-central and western New York deposits and discrete carbonate beds are usually poorly developed. The lower Hamilton Group is dominated by black shale deposits. Drab gray mudstone and silty mudstone deposits with dispersed fossils are typical of upper Hamilton depocenter facies, particularly in the Ludlowville Formation. However, the mudstone intervals are bounded by thin, condensed shell-rich beds and discontinuities that allow for regional subdivision of the seemingly monotonous shale sequence into mappable packages (Baird and Brett, 1981, 1986; Baird, 1981; Brett et al., 1986a, 1986b; Figs. 3, 5, and 6).

Although facies transitions between the western New York shelf and central Finger Lakes tough area deposits are extended and gradual, a distinctly narrower facies transition into shallower central New York platform deposits marks the southeastward margin of the central Finger Lakes trough (Fig. 5; Baird and Brett, 1981; Brett et al., 1986a, 1990a). Basin-axis mudstones in the central Cayuga Valley and Owasco Valley area grade southeastward to siltstone and sandstone units in the southern Cayuga and Skaneateles valleys. This change occurs over several tens of kilometers distance, at numerous stratigraphic levels, and involves significant reduction of unit thickness as coarsening occurs in the upslope direction (Figs. 3 and 5). Facies strike is generally northeast-southwest with inferred downslope gradients toward the northwest (Baird, 1981; Baird and Brett, 1986; Brett et al., 1986b).

The central New York shelf (easternmost Finger Lakes region to Hamilton type area) appears to mark a transition east-

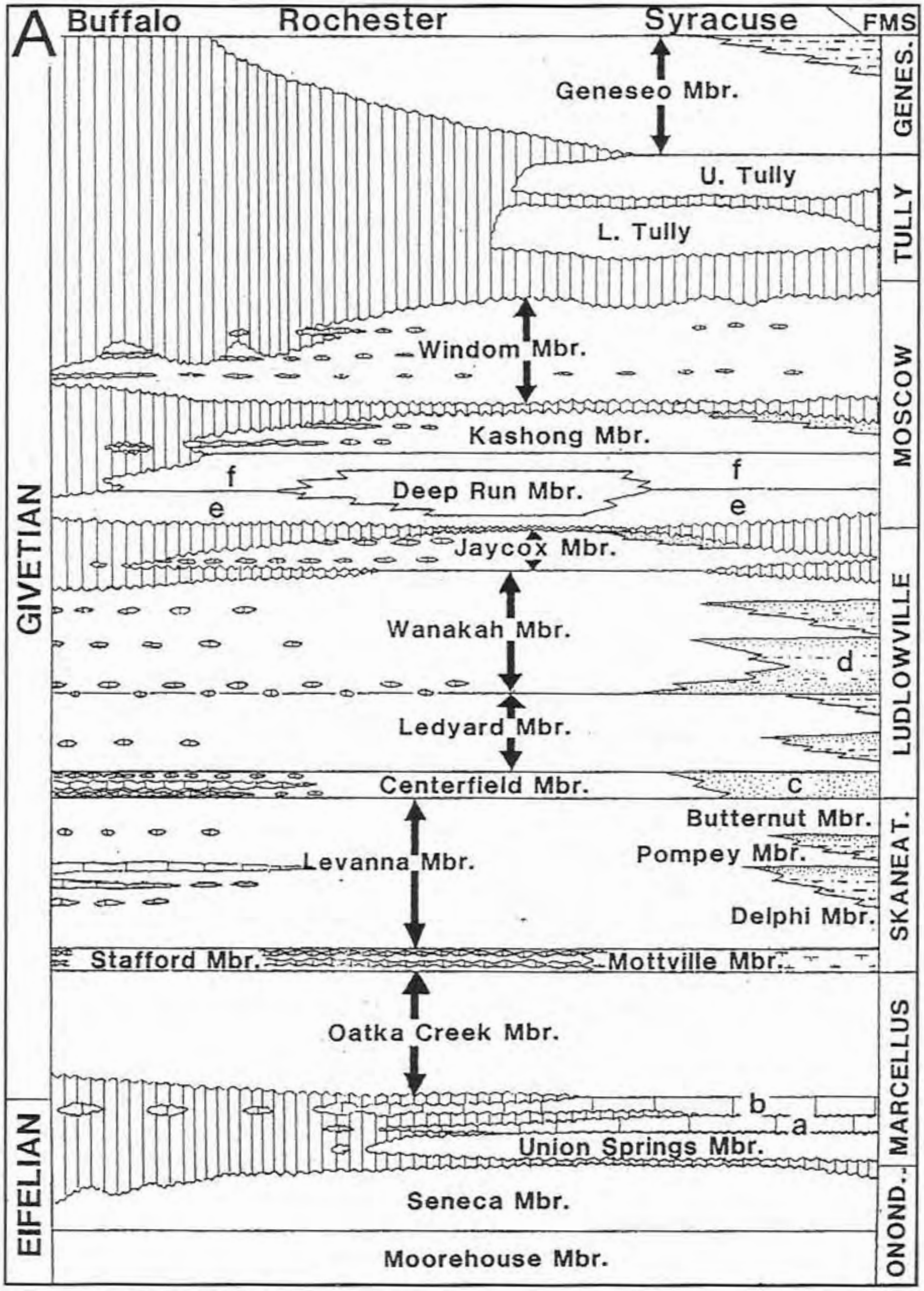
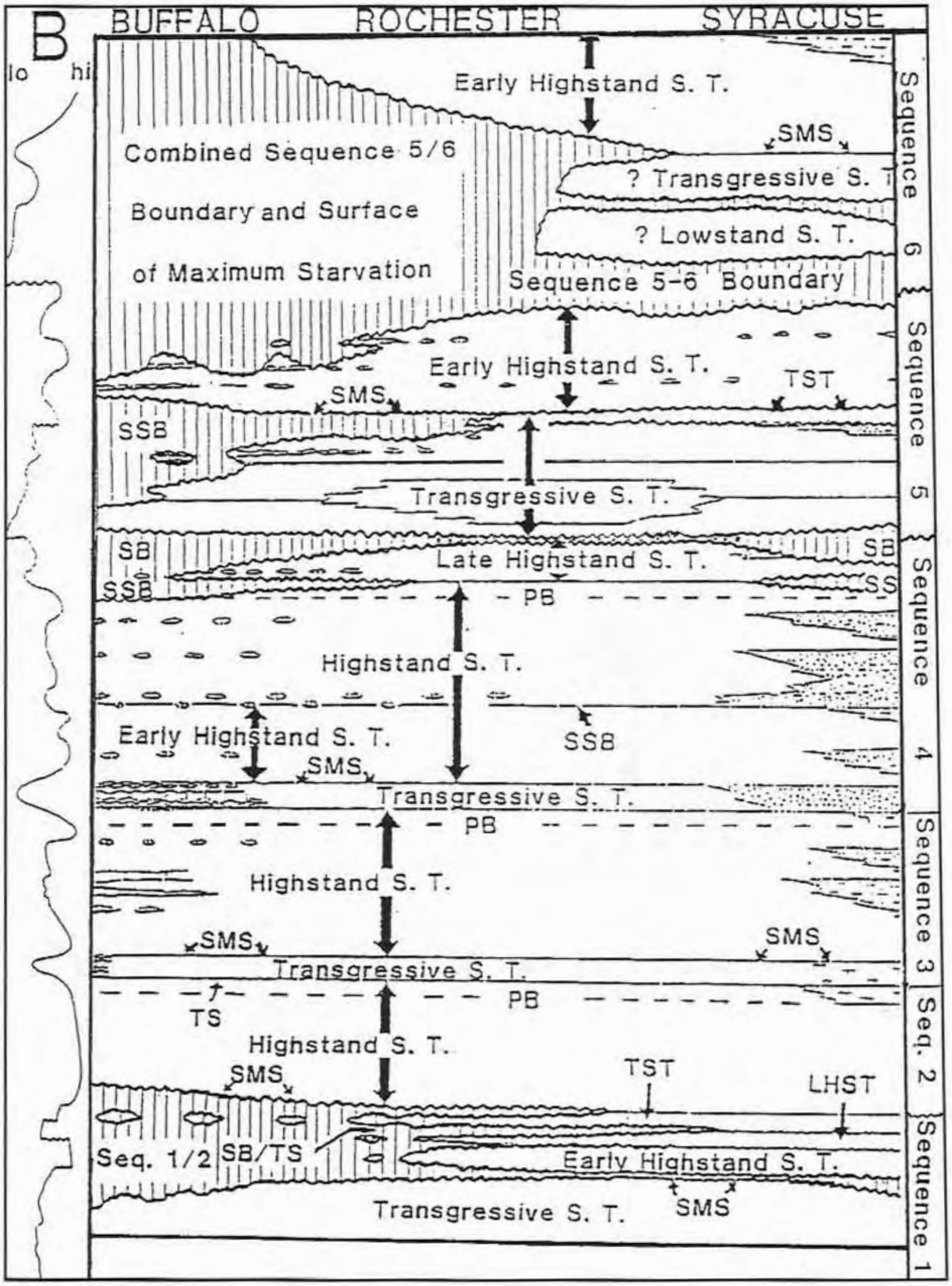


Figure 4 (on this and facing page). Time-rock diagram of upper Eifelian to upper Givetian deposits (upper Onondaga Limestone–Geneseo Shale interval) emphasizing sequence stratigraphic divisions across western New York. A: East-west time-rock diagram showing interval from upper Moorehouse Member in the Onondaga Limestone to the upper Geneseo Shale Member. Major lowstand unconformities (sub-Tichenor and sub-Tully disconformities) as well as major and minor unconformities ward to more platform-like character during Skaneateles, Ludlowville, and Moscow deposition. Although there is a steady southeastward thickening of the total section, several coarser intervals actually thin or maintain steady thicknesses toward the east. Terrigenous facies dominate with a preponderance of silty mudstone and muddy siltstone. Although bands of small calcareous concretions are observed in some shales, limestone beds are rare. Shell beds, however, are conspicuous and common and numerous condensed intervals and discontinuities appear to be present (Figs. 5 and 6).



associated with maximum flooding events (sub-Union Springs, sub-Oatka Creek, sub-Windom, sub-Geneseo) are shown. Lettered units include: a, Chestnut Street Limestone Bed; b, Cherry Valley Limestone Member; c, Chenango Sandstone Member; d, Ivy Point Siltstone Member; e, Tichenor Limestone Member; f, Menteth Limestone Member. B: Inferred sequence stratigraphy for deposits shown in A. Abbreviations: S.T. = systems tract; SMS = surface of maximum starvation; TST = transgressive systems tract; LHST = late highstand systems tract; TS = transgressive surface; PB = precursor bed; SB = sequence boundary; SSB = subsequence boundary.

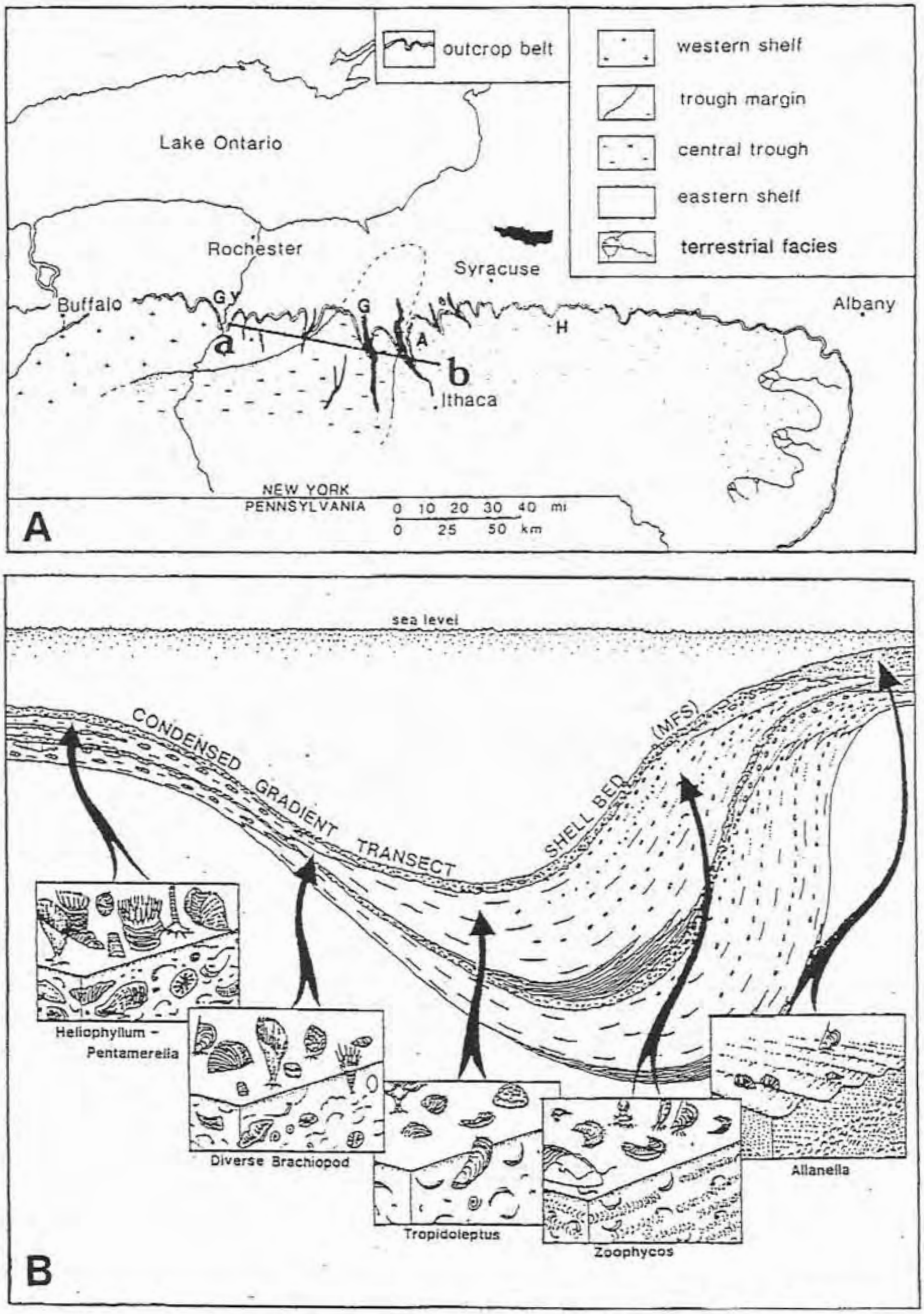


Figure 5. A: Geological setting of the northern Appalachian basin during the Givetian. Note position of Finger Lakes trough (subsidence) relative to present outcrop belt. Line a-b is cross section shown in B. Capital letters designate areas referred to in Figure 10 and in the text: Gv = Genesee Valley; G = Geneva; A = Aurora; H = Hamilton. B: Schematic cross-section through the Finger Lakes trough showing two idealized sediment wedges typical of parasequences in the Ludlowville Formation. Both cyclic sediment bundles are bounded by thin, condensed deposits that are regionally extensive. On the western shelf (to left) regressive, cycle-capping deposits are represented by limestone beds. On the central New York shelf (to right) the corresponding regressive facies is fossil-rich siltstone and sandstone. Note thin shell beds above marine flooding surfaces (MFS) at bases of parasequences. Deeper water and more rapid sediment aggradation are characteristic of the trough center. West to east paleoenvironmental and taphonomic conditions affecting bottom organisms are summarized in the series of insets; sessile filter-feeding benthos dominate in the west; mobile taxa and sparse filter feeders typify the trough, and a low-diversity mix of taxa characterize shifting sandy substrates in central New York. Communities named were discussed in more detail in Brett et al. (1990b).

In essence, the Finger Lakes trough separates two platform regions with greatly differing terrigenous sediment-supply budgets (Fig. 5). This results in a cratonic, sediment-starved suite of facies west of the trough, which is characterized by numerous shell beds and pervasive occurrence of sessile filter-feeding organisms. By contrast, on the sediment-supply side of the trough we observe a greater percentage of deposit-feeding organisms. Although both separated regions were platform-like in character, we show that combined differences in sediment supply and subsidence rate produced significantly different facies successions and cycle architecture at all scales.

HAMILTON SEDIMENTARY CYCLES

Herein we present a three-fold hierarchy of cycles in the Hamilton Group, Tully Group, and basal part of the Genesee Formation of western and central New York State (Figs. 5 and 6). The small-scale (fifth- and sixth-order) cycles correspond to parasequences, although they are not distinctly upward shallowing in all areas. At the next level, bundles of parasequences corresponding to fourth-order cycles compose parasequence sets, or subsequences. We recognize a still higher level of distinctly asymmetrical third-order cyclicity, which is represented by sequences (Vail et al., 1991; Brett et al., 1990a). We present examples of each of these cycles from Hamilton and Tully-basal Genesee deposits. The patterns described may have general application in recognition of multiscale eustatic sea-level changes within foreland basins.

Small-scale cycles: Parasequences

The smallest recognized cycles are typically less than 0.5 m (1.5 ft) thick in western New York and range up to a few meters thick in the central part of the state (Brett and Baird, 1986). In relatively shallow water environments represented by western facies they consist of subsymmetrical alternations of gray mudstone and thin, nodular to concretionary, bioturbated, calcareous mudstones or argillaceous limestones. The latter may contain very abundant skeletal debris, as well as thin obrution deposits ("smothered bottom assemblages"). Excellent examples include the well known "trilobite beds" of the Wanakah and Windom Shales in western New York (see Brett et al., 1986b; Miller, 1986, 1991, for details; Fig. 7). In deeper portions of the basin, equivalent beds may appear as thin styliolinid- and small mollusk-rich, calcareous, gray, and commonly concretionary beds that alternate with black shales.

Farther to the southeast, correlative cycles display distinctly asymmetrical, coarsening-upward patterns resembling typical

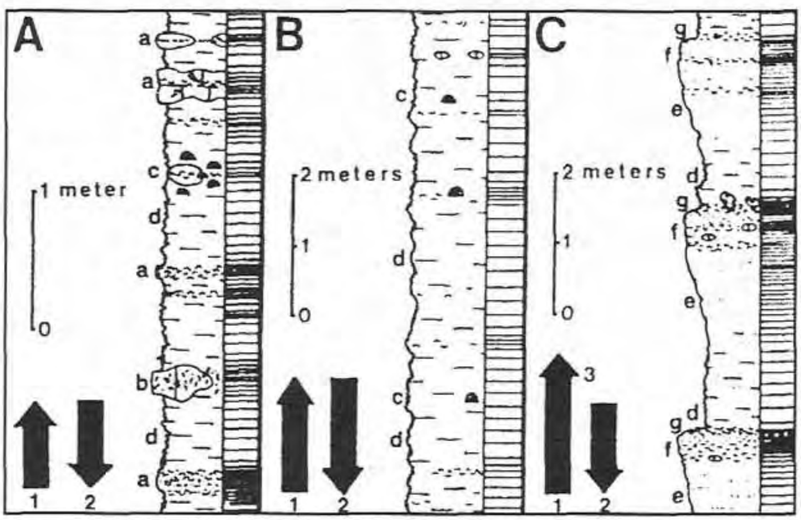


Figure 6. Schematic representation of small-scale (sixth-order) cycles in the Hamilton Group at three positions along a west-east transect. Large arrows indicate relative sediment aggradation (1) and subsidence rates (2). Spacing of horizontal bars in right columns indicates relative time richness of different intervals. Units include (a) shell beds; (b) burrowed concretionary beds; (c) layers rich in small corals (*Pleurodictyum*); (d) gray mudstone; (e) *Zoophycos*—bioturbated silty mudstone; (f) shell-rich siltstone and sandstone; (g) reworked skeletal debris beds correlative with a, at flooding surfaces of parasequences (base of parasequences [BOP] shell beds). A: Cycles in western New York calcareous facies; note that flooding surfaces are marked by shell rich beds (a); concretions (b, c) may occur midway through cycles. B: Very subtle cycles in mudstones of west-central New York; only expression of cycle tops are thin *Pleurodictyum* coral horizons; subsidence and sedimentation are both higher than in A and approximately in balance. C: Distinctly asymmetrical shallowing-upward cycles in central New York or Pennsylvania; note that sediment progradation exceeds subsidence, that such cycle caps represent winnowed silts or sands (f) deposited near wavebase; and that flooding surfaces are accentuated by shell lags (g). Modified from Brett and Baird (1986).

parasequences (Vail et al., 1991) or punctuated aggradational cycles (PACSs of Goodwin and Anderson, 1985; Fig. 6). Typical small-scale cycles in central New York display a sharp base marked by a skeletal concentration (coquina) up to a few centimeters thick (Figs. 7 and 8; see Linsley, 1991). These beds correspond to "base of parasequence shell beds" (BOP) discussed by Bannerjee and Kidwell (1991) and Kidwell (1991), and they correlate with the nodular to concretionary, argillaceous limestones of the western cycles. They carry a varied fauna of brachiopods, mollusks, pelmatozoans, trilobites, small corals, and

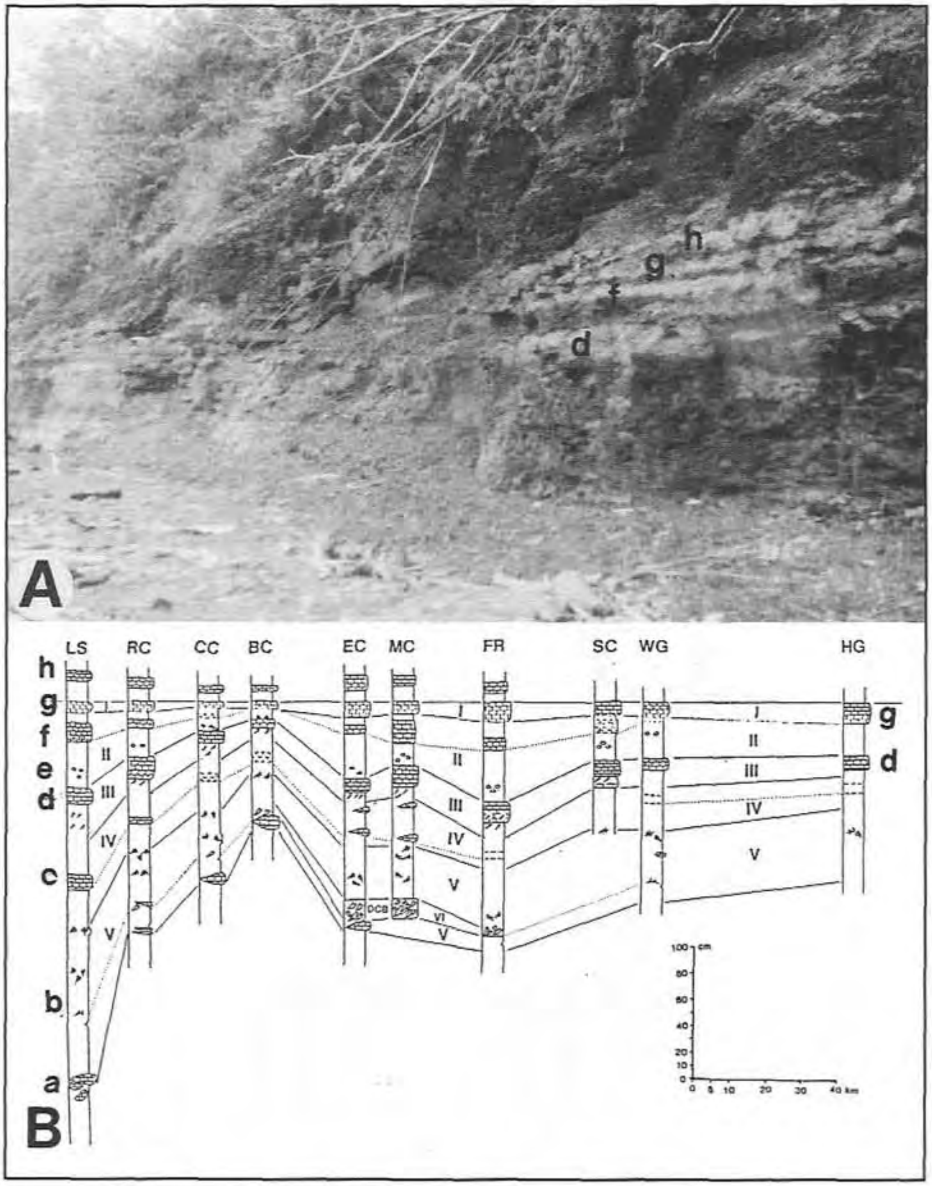


Figure 7. Small-scale cycles in western New York. A: Lower Wanakah Shale "trilobite beds"; Lake Erie shore north of Eighteenmile Creek; Wanakah, Erie County, New York. B: Correlated stratigraphic sections of upper transgressive interval of the lower cycle of the Wanakah Member (Ludlowville Formation) showing position of key beds that mark tops of parasequences (TOP), and drawn with "Bidwell Bed" (upper trilobite bed of Grabau, 1898) as datum. Correlated marker beds (solid lines) are interpreted as isochrons. Dotted lines mark correlated beds within taphofacies. Roman numerals refer to small-scale cycles. Marker beds, indicated by letters, include: a, Lakeview or *Nephriticeras* bed; b, Darien coral bed; c, Fargo bed; d, Murder Creek (trilobite) bed; e, *Athyris* bed; f, unnamed middle trilobite bed; g, Bidwell (trilobite) bed; h, unnamed upper trilobite bed. Localities are as follows: LS, Lake Erie shore; RC, Rush Creek; CC, Cazenovia Creek; BC, Buffalo Creek; EC, Elevenmile Creek; MC, Murder Creek; FR, Francis Road; SC, Salt (Bidwell) Creek; WG, Wheeler Gully; and HG, Hopewell Gully. Modified from Miller (1991).

bryozoans (Brett et al., 1986b). The varied modes of preservation range from articulated trilobites or crinoid calyces to fragmented, corroded shell debris; small phosphatic nodules may also occur. Following Kidwell (1991), we interpret these skeletal concentrations as accumulations during intervals of siliciclastic sediment-starvation associated with minor rises in relative sea level; hence they overlie marine flooding surfaces. In certain cycles (e.g., Staghorn Point cycle of Otisco Member, Ludlowville Formation) thickets of rugose corals up to several meters thick developed on the tops of siltstone platforms (Fig. 8). These regional coral biostromes record clear, relatively shallow water during initial sea-level rise. The dramatic increase of these stenotopic organisms probably resulted from siliciclastic sediment starvation during deepening.

The basal shell beds pass upward into moderately fossiliferous mudstones with abundant soft-substrate-adapted bivalves and brachiopods (*Mucrospirifer, Tropidoleptus*, chonetids; see Newman et al., 1992, for details). The upper parts of small-scale cycles are often expressed as bioturbated calcareous silty mudrock with a sparse but relatively diverse bivalve and brachiopod fauna; pipe-like concretions commonly enclose vertical pyritized burrows. Localized clusters of brachiopods, bryozoans, and pelmatozoans are observed in the upper parts of some cycles. In more basinal areas the correlative upper beds of cycles may be represented by gray bioturbated mudstone with small concretions.

Finally, cycles in central New York may be capped by physically laminated siltstone and fine sandstone (Fig. 9). These beds are generally thin (<30 cm [1 ft]), and may display coquinites of robust brachiopods (e.g., *Allanella, Spinocyrtia*),

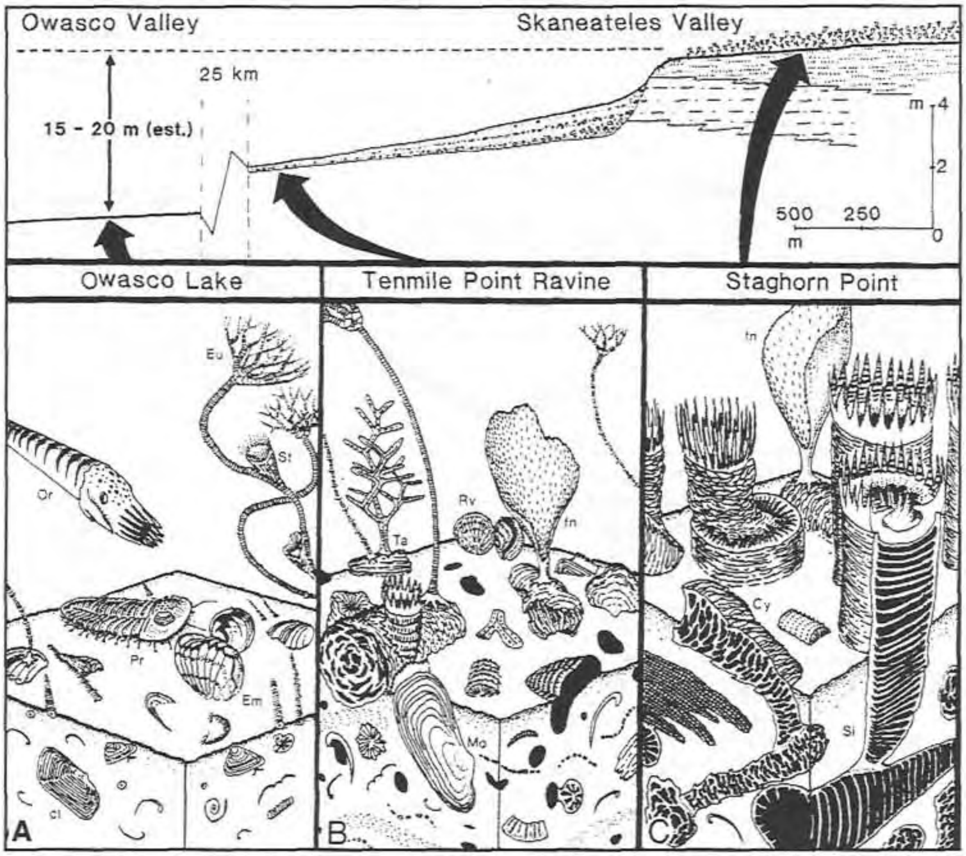


Figure 8. Northwest to southeast (basin to shelf margin) paleoenvironmental setting of the Staghorn Point coral bed (Otisco Member; Ludlowville Formation; sequence 4). Prolific growth of the stenotopic corals was a result of sediment starvation (low turbidity) during a transgression event. Spectacular differential erosion at the Staghorn shelf margin, timed with the sediment cutoff, led to formation of a submarine escarpment up to several meters in height with continued erosion, which ate into the coral-capped platform bordering the scarp. C shows in situ coral thicket. B is storm-washed debris apron of broken corals at base of escarpment. Degraded large corals and phosphatic debris were encrusted by deeper water organisms. A, lower left, shows equivalent downslope biota in minimally oxic water near the center of the Finger Lakes trough. Abbreviations: Corals: Cy = *Cystiphylloides* sp.; Si = *Siphonophrentis gigantea*; st = *Sterolasma rectum*; bryozoans: fn = fenestellids; Ta = *Taeniopora exigua*; brachiopods: Em = *Eumetabolotoechia multicosta*; Rv = *Rhipidomella vanuxemi*; trilobite: Pr = *Phacops rana*; bivalves: Cl = *Cypricardella bellistriata*; Mo = *Modiomorpha* sp.; nautiloid: Or = orthocerid; crinoid: Eu = *Eutaxocrinus*. From Brett et al. (1990b).

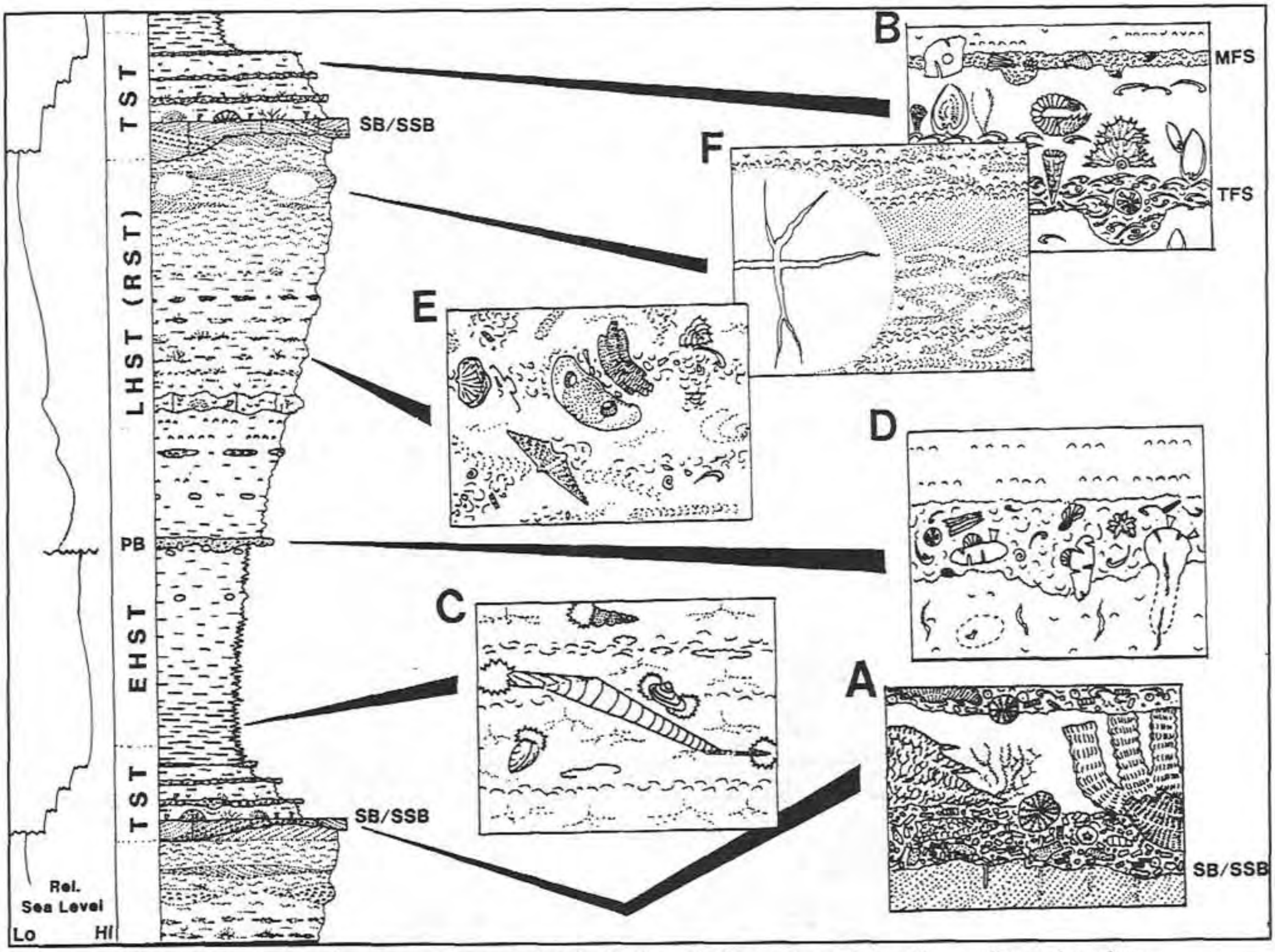


Figure 9. Idealized cycle subsequence typical of the Ludlowville Formation or lower Moscow Formation of central New York showing variation in sedimentological and taphonomic conditions. Cycles of this type (fourth-order magnitude) are dominated by the lower (regressive) part of the cycle owing to the effect of high terrigenous input (rapid aggradation to storm-wave base followed by erosion and/or sediment bypass by winnowing) followed by sediment starvation with minimal aggradation during deposition of the transgressive systems tract (TST). Key phases of cyclic sediment deposition (see insets A to F), commencing with initial transgression include: (A) beginning of transgressive systems tract (TST) above sequence or subsequence boundary unconformity (SB/SSB), a regionally widespread grainstone blanket followed by an alternation of condensed bioclastic debris beds. Robust shallow-water communities of rugose corals, brachiopods, bryozoans, and crinoids are often well preserved in rapidly deposited mud burial layers; TFS = transgressive flooding surface; MFS = maximum flooding surface. (B) Condensed maximum flooding interval is characterized by progressively deeper water taxa and series of marine flooding surfaces (MFS) which often culminate in a diastemic surface of maximum starvation (SMS), marked by phosphatic or pyritic lag debris. Between the shelly debris layers are intervals of aggradational and often obrutionary mudstone. (C) Early highstand (EHST) mudstone deposits yielding offshore dysoxic, muddy substrate association of mobile deposit feeders and scavengers (nuculids, gastropods, trilobites) and diminutive opportunistic brachiopods. Strong diffusion gradients around localized centers of sulphate reduction led to formation of numerous early diagenetic (uncompressed) pyritic nodules and steinkerns. Sediment regime was low energy and aggradational. (D) Problematic, highly condensed and diastemic precursor bed (PB) marks beginning of late highstand (regressive) systems tract. Erosion and winnowing are indicated by bored and bioencrusted, reworked concretions. This type of bed may be due to superposition of a lower-order regression event on the larger cycle or it may be due to an intrinsic, dynamic process associated with the main cycle (see text). The precursor bed is followed by a minor transgressive pulse. (E) Fossiliferous silty mudstone and siltstone facies deposited during sea-level fall; note numerous storm-scour events and pervasive bioturbation. Fossils are commonly degraded and disarticulated and obrution layers are very rare. Unstable substrate and high turbidity account for low faunal diversity. (F) Late highstand (LHST) or regressive (RST) systems tract. Note sequence or subsequence boundary (SB/SSB) at top. Interval comprises shallowing-upward siltstone to fine-grained sandstone; note winnowed storm shell lags and large late diagenetic septarial concretions; regressive maximum records highest energy conditions (approximate lower limit of fair weather wavebase).

bivalves, and (some) scattered corals. These are the equivalents of top of parasequence (TOP) shell beds (Banerjee and Kidwell, 1991). They were produced by episodes of winnowing associated with sediment bypass.

Still more proximal representatives of the same small-scale cycles from the central Pennsylvania fold belt were discussed by Duke and Prave (1991). These display very thin, basal shell or hematitic ooid beds and pass up into bioturbated siltstones, hummocky cross-stratified fine sandstones, and, in some cases, trough cross-stratified, medium to coarse sandstone. Coarse-cycle capping beds may again display sharp, crosscutting surfaces that reflect intervals of winnowing and bypass; sandstone beds in this position may exhibit *Skolithos* burrows.

Detailed correlations of decimeter- to meter-scale parasequence in the Appalachian Hamilton Group cycles (Miller, 1991) suggest an allocyclic forcing function. Flooding (deepening) events may record either minor widespread pulses of basin subsidence or minor eustatic sea-level rises, or both. As predicted from the sequence stratigraphy paradigm, the most widespread and easily correlated parts of the cycles are the thin basal shell beds. These base of parasequence beds certainly appear to reflect marine flooding events associated with widespread sedimentary condensation.

The middle and upper portions of the cycles in central New York and Pennsylvania display an upward-coarsening motif and, by inference, record shallowing-upward successions (Fig. 8). Sedimentation rates apparently increased upward to a point at which winnowing and bypass became dominant processes. This threshold may reflect the position of average storm wave base. The shallowing was apparently due to a combination of relative sea-level drop and sediment progradation. Input of sediment greatly enhanced the shallowing process, particularly in eastern (shoreward) sections. However, correlative very thin cycles (to decimeter scale) in sediment-starved western sections also show minor biofacies shifts that may be attributed to slight fluctuations in actual water depth. These small-scale cycles have been previously interpreted as sixth- to fifth-order cycles with durations of ~20 to 100 ka (Brett and Baird, 1986).

Fourth-order cycles (Parasequence sets or subsequences)

The focus of much previous study was on somewhat thicker and more complex cyclic intervals of strata, ranging from about 3 to more than 15 m (10–50 ft) thick (Brett and Baird, 1985, 1986; Savarese et al., 1986; Miller, 1986, 1991). Several of these cycles include previously named thin calcareous members of the Hamilton Group (e.g., Stafford-Mottville Members, Grasso, 1986; Centerfield-Chenango Members, Gray, 1984; Savarese et al., 1986) or informal submembers (e.g., lower Wanakah Shale, Miller, 1986, 1991; Fig. 4). About 17 such cycles have been identified in the Hamilton Group.

Each of these cyclic intervals typically contains several of the decimeter- to meter-scale parasequences described in the previous section; hence these larger intervals correspond to parasequence sets of sequence stratigraphers. We previously emphasized the similarities of these cycles to midcontinent Pennsylvanian cyclothems (Brett and Baird, 1985, 1986). Estimated durations of these cyclothems are on the order of 400,000 yr (Heckel, 1986; Klein, 1989). From the number of fourth-order cycles within the Hamilton Group and Tully Limestone (about 19–20 cycles) deposited in the 7–8 m.y. duration (late Eifelian to late Givetian Stage), we estimate similar durations. Previously, cyclothems had been termed fifth-order cycles (Busch and Rollins, 1984); however, the general consensus now appears to favor grouping cycles of 0.1 to 1.0 m.y. as fourth-order cycles (Vail et al., 1991; Goldhammer et al., 1994) and that convention is followed herein.

Where most completely developed, fourth-order cycles display a spectrum of faunal change from relatively deep water (~50–100 m [160–320 ft]), dysoxic, black shale facies to shallow (10–15 m [35–50 ft]), near wave base carbonates or nearshore sandstones that carry *Skolithos* ichnofacies (see Vogel et al., 1986, for discussion of water depths). A biofacies spectrum ranges from very low diversity leiorhynchid brachiopod faunas, through small ambocoeliid-chonetid assemblages, to more diverse brachiopod- or coral-rich associations (Brett et al., 1990b). On basin margins they may display distinctly sequence-like stacking patterns, with sharp erosion surfaces truncating shallowing-upward facies and overlain by transgressive lag deposits. Many fourth-order sequences (subsequences) are thinner and display conformable bases over a broader area of the basin than do larger (third-order) sequences (see below).

These cycles, like the parasequences, display consistent lateral changes. In western New York, where cycles are thin, the vertical facies succession reflects an upward-shallowing succession in the lower half of the cycle followed by a symmetrical upward-deepening facies progression in the upper half. The cycles, as exemplified by the Stafford and Centerfield Members (Figs. 9 and 10), commence with black or dark gray, platy shales with leiorhynchid brachiopods and pass upward into gray shaley mudstones, typically with more diverse assemblages of ambocoeliid and chonetid brachiopods, small bivalves, gastropods, and phacopid trilobites. These give way, in turn, to more calcareous mudstone with thin, shell-rich beds carrying moderate- to high-diversity brachiopod associations (e.g., *Athyris-Mediospirifer* and *Pseudoatrypa* associations of Brett et al., 1990b) with the small rugose corals (*Stereolasma* and *Amplexiphyllum*), bryozoans, pelmatozoans, and trilobites (see discussion and illustration of associations in Brett et al., 1990b). Cycles may shallow further into coral-rich mudstone or crinoidal packstone. The transgressive or deepening half-cycles may be similar in thickness and display a mirror-image biofacies progression (e.g., Centerfield Limestone; Savarese et al., 1986; Fig. 10). Transgressive facies may also contain two or more minor cycles (parasequences) expressed as multiple carbonate beds, as exemplified by the succession of concretionary trilobite-rich limestone bands within the transgressive

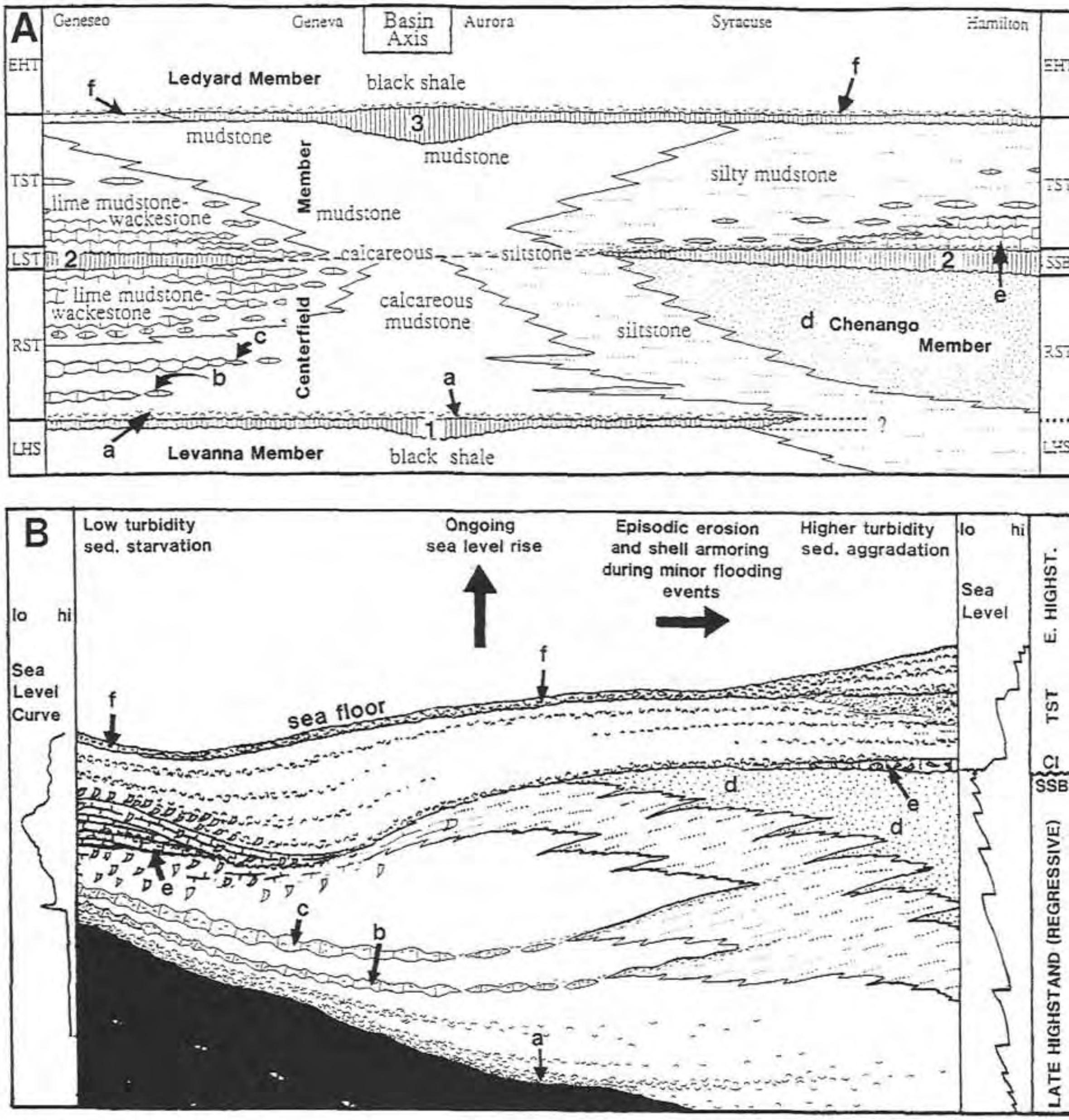


Figure 10. Fourth-order cycle (subsequence) in western and central New York region. A: East-west time-rock transect (approximately along section line a-b in Fig. 5) through middle Givetian Centerfield Member (base of Ludlowville Formation; see Fig. 4) showing position of associated unconformities (vertical rule) and condensed intervals. A first unconformity (1), marking the base of the Centerfield Member, corresponds to the basal contact of a precursor bed (a, Peppermill Gulf Bed). This bed is abruptly juxtaposed onto late highstand dark gray to black shales of the underlying Levanna Member. Maximum erosive downcutting is centered near the eastern margin of the basin. The lowstand maximum (LST) corresponds to a sequence bounding unconformity (2) which closes to conformity only near the basin axis. A third key unconformity (3) corresponds to a surface of maximum flooding at the base of a condensed shell bed (f, Moonshine Falls Bed) which floors early highstand dark Ledyard Shale above the Centerfield. Again, note that the maximum downcutting occurs just east of the basin axis. B: Schematic cross section of a fourth-order sedimentary cycle associated with deposition of the Centerfield Limestone. Note eastward thickening of the Centerfield and lithologic transition from carbonate facies (in west) to siltstones and sandstones in central New York. Note also a change from a symmetrical limestone-cored cycle in the west to an upward-regressive half cycle in the east owing to higher terrigenous supply in the latter region (see discussion in text). Also shown is an abrupt lower boundary for Centerfield regressive deposits that is marked by a complex and problematic "precursor bed." This bed records sedimentary condensation and local erosive scour timed with the beginning of the late highstand (regressive) phase of deposition (see discussion of "precursor beds"). Lettered horizons for both diagrams include (a) Peppermill Gulf bed; (b) Salt Creek bed; (c) Browns Creek (trilobite) bed; (d) Chenango Sandstone; (e) Stone Mill Limestone and laterally equivalent Centerfield main limestone (Schaeffer Creek bed); (f) Moonshine Falls phosphatic bed (see Figs. 4 and 9 for explanation of symbols for sequence stratigraphy; RST = regressive systems tract or late highstand [LHS]).

half cycle in the lower Wanakah Shale (Miller, 1986, 1991; Fig. 7).

In central basinal locations (i.e., in the Finger Lakes trough of Brett et al., 1986b; Mayer et al., 1994) the cycles are thicker, more diffuse, and also rather symmetrical. Dark gray to black shales pass upward into calcareous silty mudstones (Figs. 9 and 10). These latter may be more sparsely fossiliferous than counterparts to the west and carry a higher proportion of "mud-tolerant" bivalves or semiinfaunal brachiopods (e.g., *Tropidoleptus* biofacies of Brett et al., 1990b; Fig. 5). The shallowest part of the cycle is significantly expanded and is expressed as a more compact silty-calcareous bed or concretion horizon with a few corals. The transgressive half of the cycle is similar to the lower, but thinner and with more closely spaced shell-rich beds (often with *Tropidoleptus* or diverse brachiopod biofacies). In still more basinal positions cycles consist predominantly of dark dray to black shale with a leiorhynchid fauna that grades into axial gray ambocoeliid- or chonetid-bearing calcareous mudstones at the regressive "core" of the cycle.

In central New York and west-central Pennsylvania the equivalent cycles are distinctly asymmetrical and are typically capped by coarse siltstone and sandstone beds (Faill et al., 1978; Duke and Prave, 1991; Linsley, 1991; Figs. 5, 8, and 9). They display a hierarchical stacking of minor cycles in which successive parasequences display more prominent and coarser caps; the highest two or three parasequences may be amalgamated to form a single coarsening-upward cycle which usually weathers to form a prominent bench (Fig. 8). The top of this parasequence set is a fine- to medium-grained hummocky cross-stratified sandstone which commonly displays large calcareous concretions. Typically the sandstone is capped by a distinctive (BOP) shell bed that may carry a diverse brachiopod- or even a coral-rich fauna. This shell bed is followed by a relatively thin, complex, shelly interval bearing numerous thin, high-diversity shell beds (see below). Faunal composition may fluctuate significantly between adjacent beds owing to stacking of component smaller cycles. But, overall, there is an upward trend from moderate-diversity assemblages (chonetid, *Tropidoleptus* and *Mucrospirifer* biofacies of Brett et al., 1990b; see Fig. 5) to sparse assemblages of scattered large bivalves (*Ptychopteria*) and robust brachiopods (*Spinocyrtia, Allanella*). A few, isolated large corals, such as *Favosites hamiltoniae*, up to 50 cm (1.5 ft) across, may occur within cycle-capping calcareous siltstones or sandstones. These occurrences demonstrate that the sea floor was not only shallow, but also that water was clear enough to permit coral growth. Perhaps sufficient mud-silt bypass occurred to permit corals to colonize the sandy substrates.

The first transgressive beds above the sandstone caps of the coarsening-upward cycles (major BOP shell beds) may include crinoid- or coral-rich, arenaceous limestones (e.g., Stone Mill Limestone bed of the Chenango Member; Kramers, 1970; Gray, 1984; Fig. 10). A thin transgressive interval, less than 1 m to about 5 m thick (~3–16 ft), overlies the basal transgressive shell- or coral-rich bed. This interval is distinctly muddy and highly fossiliferous as opposed to the silty or sandy, and sparsely fossiliferous facies that typify the regressive deposits. Lithologically and faunally this interval displays lateral uniformity, resembling the coeval transgressive beds farther west (Fig. 10). These beds were deposited in an interval of siliciclastic sediment starvation during continued sea-level rise. Evidently, most of the coarse siliciclastic fraction was deposited in nearshore areas, probably estuaries and bays created by rising sea level (see McCave, 1969, 1973). However, finer-grained muds were distributed much more widely and uniformly, because the transgressive shales with their component beds display relatively little variation in thickness. Bands of calcareous nodules or argillaceous limestone may occur within this interval. These calcareous beds seem to represent enhanced sediment starvation due to minor fifth- and sixth-order flooding events (bases of parasequences) superimposed on the general transgression (see Figs. 9 and 10). These beds generally display a retrogradational facies succession; i.e., each successive parasequence-capping shell-rich carbonate bed appears to record deeper water conditions than the lower one (Miller, 1986, 1991; Figs. 7 and 9).

The top of the transgressive interval may be recognized as an abrupt shift from fossiliferous medium gray, calcareous mudstone to darker gray or even black shales (Figs. 9 and 10). This sharp contact, apparently recording a surface of maximum flooding, is overlain in many instances by a thin but distinctive condensed bed. Such beds, marking condensed intervals, are more complex than typical BOP shell layers of parasequences. They display evidence for prolonged time averaging of skeletal debris, such as mixed biofacies fossils or strongly corroded and abraded fossils (e.g., the RC shell bed of the Kashong Member; see Brett and Bordeaux, 1990, for discussion). The appearance of diaclasts, such as phosphatic prefossilized steinkerns and nodules or reworked carbonate concretions and pyrite, also indicate submarine erosion of older sediment under minimally oxygenated water (see Baird and Brett, 1991, for further discussion).

Highstand deposits above the maximum flooding surface are represented by relatively thick aggradational successions of dark gray mudrocks. Several subtle, mudstone-dominated parasequences may be present within these intervals; however, in general, these early highstand portions of the cycle are fairly monotonous and display little upward change (Figs. 9 and 11).

Only the late highstand or regressive facies display a distinct upward-shallowing pattern as discussed above. In western sections this shallowing is merely recorded by progressive change in biofacies (e.g., from leiorhynchid to ambocoeliid, diverse brachiopod, and finally coral-rich biofacies). In eastern sections this change is represented by coarsening-upward cycles (Figs. 5, 9, and 10). However, in both areas, the base of the regressive facies is commonly sharply set off from underlying dark shales by a condensed shell-rich bed. This bed is comparable to the precursor beds described below for larger-scale sequences.

Third-order cycles: Sequences

The Eifelian to late Givetian (Middle Devonian) interval (8–10 m.y.) constitutes the early phase or holostrome of Sloss's (1963) Kaskaskia supersequence (see Dennison and Head, 1975). It is bounded by major erosional unconformities, the Wallbridge unconformity below and the Taghanic unconformity (Johnson, 1970) above. The largest-scale stratal cycles within the Middle Devonian holostrome of the north-central Appalachian basin consist of sharply-bounded intervals ranging from 10 m to more than 500 m (35 ft to 1,600 ft) in thickness.

Each of these packages contains two to five component (fourth-order) cyclothems, and represents about 0.8 to 2.0 m.y. These values fall within the range of temporal magnitudes of many third-order sequences, as now recognized by seismic stratigraphers (Vail et al., 1991). In the Hamilton Group, these sequences closely, but not precisely, parallel formations previously defined in New York State, because Cooper (1930, 1933) used thin, widespread limestones to define the bases of these formations. These carbonates are interpreted herein as transgressive systems tracts of the third-order sequences.

In ascending order, the sequences discussed herein for the Middle Devonian Hamilton Group (Tully Limestone and overlying units using traditional lithostratigraphic terminology) are (a) upper Seneca member (Onondaga Formation), and lower Marcellus Formation (Union Springs and Stoney Hollow members); (b) Cherry Valley Limestone and upper members of Marcellus Formation; (c) Skaneateles Formation; (d) Ludlowville Formation; (e) Moscow Formation; and (f) Tully Formation–lower Genesee Formation (Fig. 4).

As thus defined, each sequence in western New York commences with a sharply-based, thin, relatively shallow water limestone or calcareous siltstone-sandstone corresponding to a transgressive systems tract. These are, in ascending order, (a) upper Onondaga Limestone; (b) Cherry Valley Limestone (Marcellus Formation); (c) Stafford Limestone–Mottville Siltstone (Skaneateles Formation); (d) Centerfield Limestone–Chenango Siltstone (Ludlowville Formation); (e) Tichenor–Portland Point Limestone (Moscow Formation); and (f) Lower Tully Limestone and eastern clastic equivalents (Fig. 4). In each case, the transgressive systems tract comprises the calcareous to condensed sandy facies of the lowest fourth-order cycle within the sequence. In some sequences two or more subtle cycles (subsequences) can be discriminated within the transgressive systems tract; e.g., three "beds" or subdivisions of the Cherry Valley Member (Griffing and Ver Straeten, 1991; Tichenor, Menteth, and R.C. (*Rhipidomella-Centronella* bed cycles in basal Moscow, Fig. 4).

Each of the Middle Devonian sequences is bounded by a sharp diastemic to erosional sequence bounding surface; however, lowstand deposits are absent, so these surfaces also coincide with transgressive surfaces (cf. Jordan and Flemings, 1991). Each sequence also contains a reasonably distinct surface of maximum starvation (SMS), typically marked by a thin phosphatic pebble bed or bone bed. For the Hamilton Group sequences these include (a) top of Onondaga bone bed; (b) top of Cherry Valley bone bed (Marcellus Formation); (c) unnamed top of Mottville surface (Skaneateles Formation); (d) Moonshine Falls phosphate bed (Ludlowville Formation); and (e) Little Beards phosphate bed (Moscow Formation; Fig. 4).

Highstand deposits include thin intervals of argillaceous limestone or very calcareous shale in the predominantly carbonate Onondaga and Tully Formations, and monotonous, sparsely to moderately fossiliferous, medium to dark gray mudstone in the Hamilton Group. In the siliciclastic-dominated Hamilton Group the highstand-regressive deposits are much thicker than the transgressive systems tracts and include two to four fourth-order cycles, corresponding to members or submembers. Thus, for example, the upper Marcellus sequence is subdivisible into

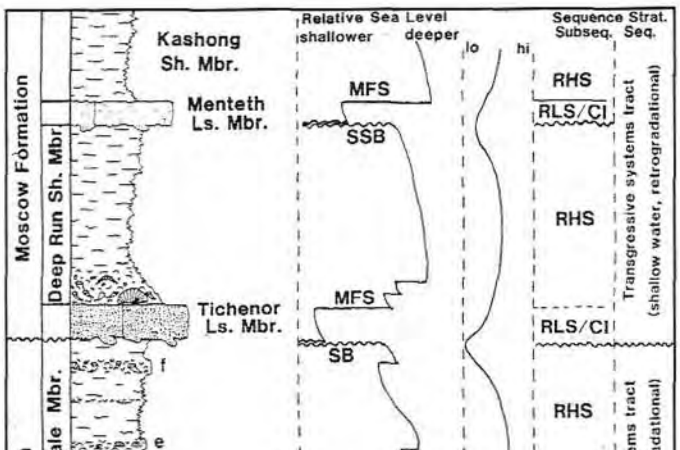


Figure 11. Multiscale (hierarchical) sedimentary cycles in the upper part of the Ludlowville Formation and the lower part of the Moscow Formation. Cycles are of varying temporal and process magnitude and are typically bounded by diastems and condensed intervals. The third-order sequence boundary (SB; base of Tichenor) is shown, as are subsequence (fourth-order) contacts (SSB; base Hills Gulch Bed, base Menteth Limestone). Maximum flooding surfaces (MFS) are represented by lag or condensed beds beneath shale units. At least one definite precursor bed (PB; Limerick Road Bed) is shown (see text). Abbreviations: RLS = relative lowstand; CI = condensed interval; RHS = relative highstand. Letters: a, Bloomer Creek Bed; b, dark shale yielding diminutive brachiopod fauna (uppermost Wanakah Member); c, Spafford Shale Member; d, Greens Landing Coral Bed; e, *Megastrophia* sponge-rich bed; f, Cottage City coral bed.

the Berne and Otsego members in eastern New York (Ver Straeten, 1994), with the former represented by a very thin (0.5 m [1–1.5 ft]) interval in western New York. The Skaneateles highstand is subdivided into Delphi Station, Pompey, and Butternut members; the Ludlowville highstand is subdivided into Otisco (Ledyard), lower Wanakah, upper Wanakah– Spafford, and Jaycox cycles; and the Moscow highstand comprises lower, middle, and upper cycles of the Windom Member (Figs. 4 and 11).

Although each of these larger-scale cycles displays at least a subtle deepening-shallowing pattern, as described above, the highest fourth-order cycle in each sequence is sharply set off from the others, such that the highstand can be subdivided into early, largely aggradational, and late progradational phases. As is the case within the intermediate-scale cyclothems, the late, or progradational, phase of each large-scale sequence is sharply demarcated at its base by a condensed "precursor" bed (see Figs. 9, 10, and 11). This is a thin but widespread fossil-rich bed, commonly with reworked concretions, that occurs abruptly above more sparsely fossiliferous, and probably deeper-water mudstone or shale facies. Hamilton sequences possess "precursor beds" (PB) in the upper portions of the highstand at the base of the final fourth-order cycle that culminated in the next sequence boundary (Fig. 4B). These are as follows: (a) lower Marcellus sequence: unnamed fossiliferous bed in the lower Stoney Hollow Member (Griffing and Ver Straeten, 1991); (b) upper Marcellus sequence: "Unit A" coral bed of the Mottville Member (Grasso, 1986); (c) Skaneateles sequence: Peppermill Gulf bed of the Chenango Sandstone and equivalent Centerfield Member (Gray, 1991; Fig. 10); and (d) Ludlowville sequence: Limerick Road bed of the Spafford Member (Mayer et al., 1994; Fig. 11). No precursor bed is recognized in the upper Moscow Formation, but this probably is due to strong erosional truncation of beds along the sub-Tully unconformity.

GENESIS OF KEY EROSION SURFACES AND CONDENSED BEDS IN MIDDLE DEVONIAN SEQUENCES

Condensed sections are relatively thin, time-rich intervals that are widespread and may correlate laterally with much thicker sedimentary successions (cf. Tucker, 1973; Baum and Vail, 1988). These intervals, which range from millimeters to a few meters in thickness, are characterized by multiply reworked fossils and clasts, commonly showing a mixture of biofacies; some fossils are heavily corroded and may be fragmented. They may show evidence of prefossilization, i.e., a phase of early diagenesis within sediment followed by exhumation. Diaclasts, such as hiatus concretions, tubular pyrite, and reworked fossils, may be common within condensed beds (Figs. 12, 13, and 14). Biostratigraphically condensed intervals may show a mixing of zonal fossils; they are enriched with zonally important index fossils, such as conodonts. Chemically resistant skeletal elements, such as phosphatized fish bones and teeth, are common in Middle Devonian condensed beds (Hussakoff and Bryant, 1918; Baird and Brett, 1986, 1991). The presence of certain early diagenetic minerals, such as phosphate and glauconite, is also indicative of sedimentary condensation.

Sedimentary condensation in siliciclastic-dominated environments results from two major processes: (1) winnowing and bypass of fine-grained sediments due to persistent or episodic current action; and (2) sediment starvation in marine basins as a result of nearshore sediment entrapment or absence of sediment input (Kidwell, 1991). In carbonate depositional systems, condensation may result from a restriction or near elimination of carbonate production in the absence of input of extrabasinal sediment.

The sequence stratigraphic model predicts that rapid relative rises in sea level will be associated with sedimentary condensation in offshore areas. Siliciclastic sediment starvation will commonly result from rapid transgression, which drowns rivers, hence trapping siliciclastic sediment nearshore (Van Wagoner et al., 1988). In areas of shallow water where carbonate production may be high, siliciclastic sediment starvation could be associated with the development of limestone beds. Indeed, this model of nearshore siliciclastic alluviation has been proposed to explain thin continuous carbonate units in the Devonian rocks of New York State and elsewhere (see Johnson and Friedman, 1969; McCave, 1973). We have argued (Brett and Baird, 1985, 1986) that most thin carbonates in the Middle Devonian of New York State cannot represent maximum marine flooding, but rather appear to have been deposited during or immediately after major drops of sea level. Nonetheless, the very widespread Middle Devonian limestones used to define formation boundaries were probably deposited during initial parts of transgressive systems tracts (see Figs. 4 and 11).

Certain phosphatic, bone-rich, or conodont-rich carbonate beds also represent sediment-starved condensed intervals associated with maximum marine flooding as predicted in all sequence models. Condensed sections would not be predicted to occur near the tops of sedimentary sequences associated with sea-level fall. Nonetheless, as we illustrate in this chapter, a second category of condensed beds, closely resembling those associated with sediment starvation during sea-level rise, were formed during early phases of sea-level fall. Such condensed beds at the bases of coarsening-up cycles are the aforementioned "precursor beds," which mark the bases of late highstand (regressive) phases of sedimentary cycles (Figs. 9, 10, and 11). These beds are difficult to explain with existing models.

Surfaces of erosional truncation and condensed beds occur predictably within sedimentary cycles (Figs. 4, 10, and 11). Whereas condensed horizons are associated with rapid relative sea-level rise, erosion surfaces should be associated with rapid relative sea-level drop. The rate of fall may exceed local subsidence leading to exposure of formerly inundated areas and subaerial erosion. Furthermore, the lowering of relative sea level brings formerly deep basinal muds first into the range of shallow storm wave base, and eventually normal wave base. This shallowing may result in submarine erosional truncation of sediments, particularly muds, which accumulated initially below the reach of wave erosion. This may result in erosion and redeposition of muds and silts into deeper portions of the basin. The storm-generated currents themselves may produce erosive effects in some areas, particularly in regions of slopes, where minor bypass channels may develop that permit more rapid basinward sediment transfer of fine-grained sediments across or down gently sloping ramps and into basinal regions. Consequently, relatively thick and "time-poor" successions in depocenters should be expected to coincide with erosion surfaces on the basin margin.

The sequence model does not predict erosion associated with rapid transgression, except in shoreface areas, where ravinement may take place. In offshore areas, the raising of wave base and storm wave base should have the opposite effect of that seen during sea-level lowering (i.e., areas formerly affected by winnowing should be deepened to the point that they are no longer eroded and offshore areas should undergo little or no wave erosion). Again, however, we find that there are

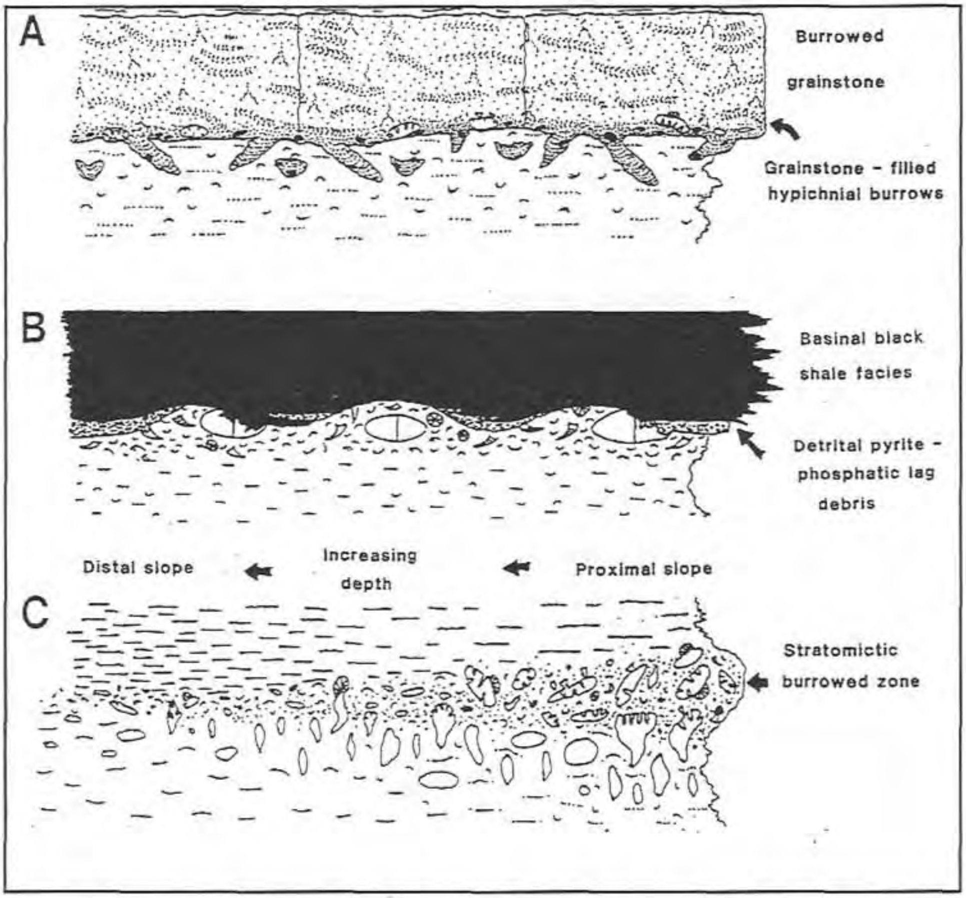


Figure 12. Discontinuities produced under differing conditions of submarine erosion. A: Contact beneath high-energy grainstone bed showing enlargement of domichnial burrows at limestone base by abrasion and infilling of the enlarged burrows by skeletal carbonate. Burrowing occurred during time of mudstone erosion and also during period of episodic accumulation of limesand. B: Black (organic-rich) shale-roofed unconformity produced by combined process of episodic abrasion and carbonate dissolution (corrasion) under conditions of sediment starvation on a near-anoxic submarine slope. Lag deposits of insoluble detrital pyrite, bone-conodont debris, phosphatic nodules, as well as quartz sand and gravel are typically concentrated in erosional runnels on such discontinuity surfaces (see Baird and Brett, 1986, 1991). This is an extreme expression of a surface of maximum flooding seen in some higher-order transgressive-regressive cycles (see text). C: Cryptic "shale on shale" discontinuity obscured by soft-sediment burrowing that was synchronous both with episodic scour events and with final discontinuity burial. *Stratomictic* contacts such as these are revealed by occurrences of reworked concretions and prefossilized fossil steinkerns which were loosened and exhumed by borrowing action and episodic scour events (see Baird, 1978, 1981; Baird and Brett, 1981); such discontinuities are typical of precursor beds.

many exceptions to this generalization, and indeed, major erosional truncation surfaces commonly appear to be associated with rapid relative sea-level rise and with condensed sections. In many instances, condensed beds associated with rapid sea-level rise and relative sediment starvation may be truncated, in turn, by erosion surfaces that underlie still deeper water facies, particularly black shale. Such erosion surfaces may be marked by unusual lag deposits, such as lenticular bodies of reworked phosphate nodules sedimentary pyrite, fish bones, and conodonts (Baird and Brett, 1986; Baird et al., 1988).

In the following sections we discuss patterns and processes involved in the genesis of three distinctive types of condensed beds and associated erosion surfaces in the Middle Devonian. From a sequence perspective these are (a) lowstand erosion surfaces and overlying transgressive lags; (b) surfaces of maximum sediment-starvation and associated lag beds formed during maximum marine flooding (bases of highstand deposits); and (c) mid-highstand condensed beds or "precursor beds" (Fig. 12).

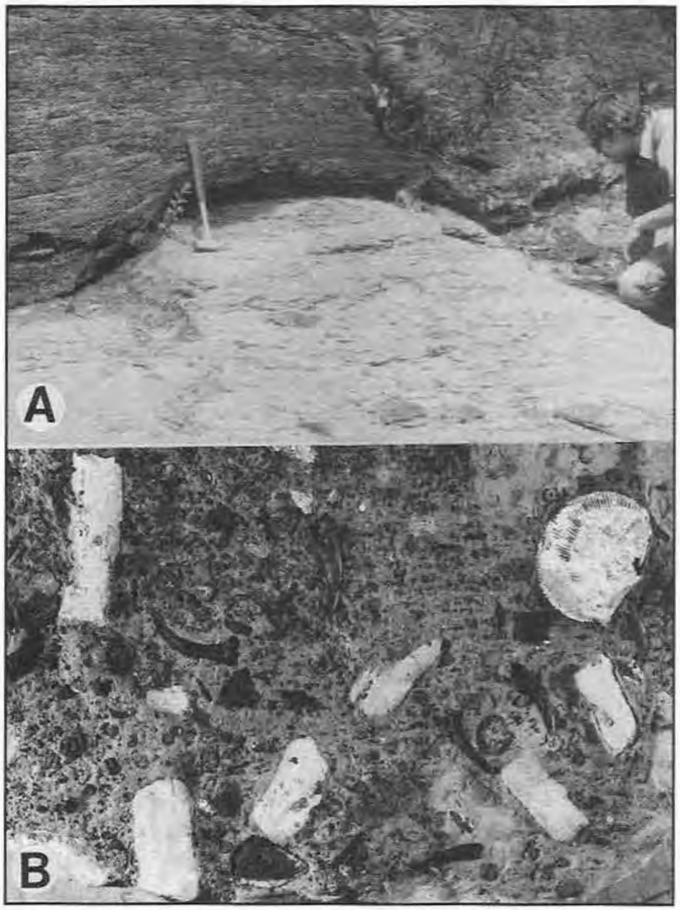


Figure 13. A: Conspicuously channeled discontinuity below black shale deposits of the medial Levanna Shale Member at Buffalo Creek near Buffalo, New York. These north-south–trending channels are probably the result of bottom currents; hammer gives scale. B: Pavement of teeth belonging to the crossopteryian fish genus *Onychodus* which occurs on a corrosional unconformity between the Onondaga Limestone (Seneca Member) and overlying black shale deposits of the Union Springs Member at the Jamesville Quarry at Jamesville, New York, near Syracuse. Episodic bottom scour and corrosion of exposed carbonate in a sediment-starved transgressive regime led to spectacular concentration of teeth, bones, and reworked burrow fills (white tubes). On this contact, reworked pyrite is rare, but the lag fraction is dominated by fish teeth and bones.

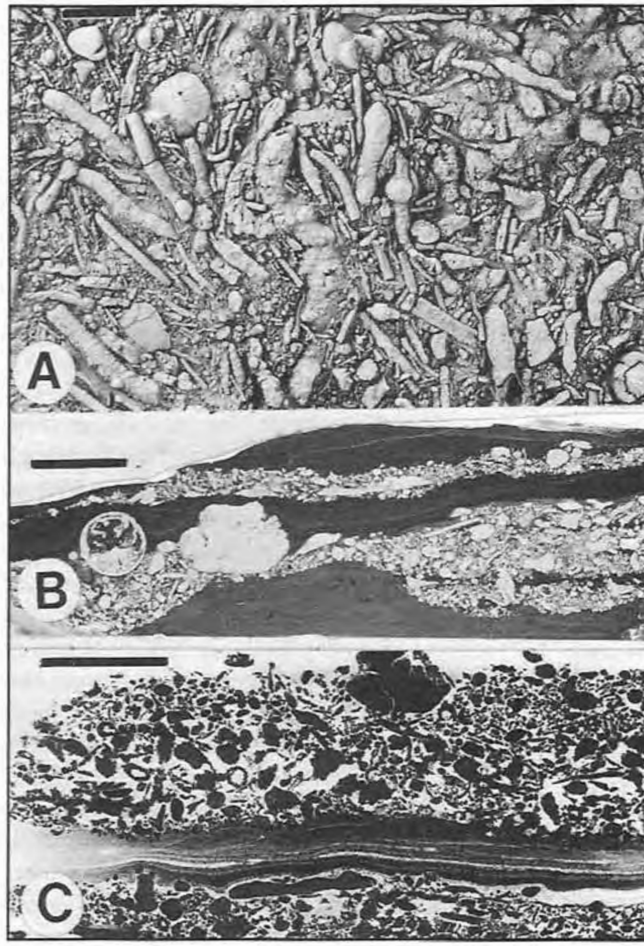


Figure 14. Detrital pyrite lag concentration in the Leicester Pyrite Member at the base of the Geneseo Black Shale, along the regional Taghanic unconformity in western New York. A: Top view of Leicester pyrite concentration showing abundance of broken, prefossilized, pyritic burrow tubes exhumed from the Windom Shale, which underlies the disconformity. Pyritic steinkern of distinctive upper Windom brachiopod *Allanella tullia* is visible in extreme lower right. Creek is near Richmond Mills, New York. Bar = 1 cm (2.5 in). B: Vertical cut and polished section showing Leicester debris both on the regional disconformity (lower debris band) and above it within the basal part of the Geneseo Shale (upper debris band); this shows that Leicester deposition occurred within the near-anoxic basin setting. Note the nautiloid steinkern in transverse view with geopetal fill. Cazenovia Creek at Spring Brook, Erie County, New York. Bar = 1 cm. C: Thin section of Leicester from same locality as B showing two Leicester debris layers separated by black, laminated, organic-rich Geneseo Shale. Note absence of mud between pyrite clasts. Bar = 0.5 cm (1.3 in). From Baird and Brett (1986).

Erosion and condensation associated with regression maxima (sequence boundaries)

The most important type of sharp erosional surface occurs at caps of shallowing-upward cycles. Such surfaces form at the lower contacts of thin, skeletal limestone beds that comprise basal units of Hamilton Group Formations (sequences). As noted, these limestones record the initial transgressive deposits of fourth- and third-order cycles (Figs. 4, 10, and 11). The erosion surfaces beneath such beds bevel underlying strata along basin margins. The discontinuity-floored, coarse crinoidal and shell-rich limestones appear to represent some of the shallowest (highest energy) deposits in the entire Devonian succession (see Fig. 4).

The most prominent erosion surface in the Hamilton Group is the lower contact of the Tichenor Limestone, which forms the base of the Moscow Formation (third-order sequence) by definition of Baird (1979). The Tichenor is a prominent and widespread, though very thin (10–40 cm [~4–16 in]) crinoidal packstone-grainstone unit (Figs. 4, 9, and 11). The Tichenor occurs immediately above shallow-water deposits (regressive or late highstand deposits) of the largest-magnitude shallowing event during deposition of the entire Hamilton Group (Figs. 4, 11). In the Seneca Lake region (basin center) the Tichenor is nearly conformable with underlying calcareous, silty mudstones of the upper Jaycox Member and with very similar facies of the overlying Deep Run Member (Baird, 1979). These facies appear to represent slightly lower energy mud deposits that accumulated below normal wave base, whereas the Tichenor packstones and grainstones reflect persistent physical reworking of carbonate gravels and sands in relatively high energy, near wave base environments. Tichenor crinoidal debris accumulated immediately after a period of maximum shallowing. The maximally regressive deposits may have been coarse siltstones or sandstones. Remnants of a sub-Tichenor siltstone have been discovered in one locality (Big Hollow Creek) near the upper Ludlowville depocenter between Seneca and Cayuga lakes. However, elsewhere the highest portions of the sub-Tichenor coarsening-upward cycle have been removed by an interval of submarine or possibly subaerial erosion. Hence, the Tichenor overlies an erosion surface that bevels the underlying upper Ludlowville siltstones and mudstones on either side of the basin axis. The amount of section removed beneath this surface increases both east and west of the Seneca Lake region; in such areas, the upper portions of the Jaycox Member (3 to 10 m [10 ft to 35 ft] of section) have been removed (Mayer et al., 1994, Fig. 11). Progressive westward overstep of distinctive marker beds in the underlying Jaycox Shale Member by the Tichenor has been demonstrated (Mayer, 1989; Mayer et al., 1994). In extreme western sections along Lake Erie, the Tichenor rests on Wanakah Shale with the Jaycox and uppermost Wanakah beds removed. The pattern of erosion reveals truncation of older sediments along the flanks (east- and west-dipping ramps) of the central Finger Lakes trough (Baird and Brett, 1981) during a period of relative sea-level lowstand. The fact that the base of the Tichenor is unconformable, even near the basin center in New York, reveals the severity of the regression. The eroded sediments have apparently been removed by southward current transport along the axis of the foreland basin.

The Tichenor is a condensed bed consisting of multiple reworked, broken, and abraded pelmatozoan ossicles, corals, and other skeletal debris (Figs. 9 and 12). In places, two or more graded beds can be observed within the Tichenor, suggesting amalgamation of sediment by severe storms during which much or all of the relict sediment blanket was reworked. The Tichenor may be slightly diachronous, becoming younger both eastward and westward from the axis of the Finger Lakes trough (Fig. 4, A and B).

The basal surface of the Tichenor Limestone is everywhere sharp and it locally displays large burrows up to 10 cm (4 in) across, and up to 1 m (3 ft) in length, protruding from the sole surface (Fig. 12). In places these burrows may extend, as discrete exichnial "tubes," downward into underlying shales. These "megaburrows" are apparently devoid of scratch marks, but otherwise they resemble the large burrows on the analogous basal erosion surface of the Middle Devonian Hungry Hollow Limestone in Ontario; these have been ascribed to *Cruziana* (probably burrows of the large trilobite *Dipleura*) by Landing and Brett (1987). Such burrows prove that the underlying Jaycox-Wanakah muds were overcompacted and firm; presumably, up to several meters of mudstone had been removed by pre-Tichenor erosion. The details of the basal surface are commonly obscured by crusts of late diagenetic pyrite that seem to have formed preferentially along the contact between the Tichenor and the underlying mudstone.

The erosion surface below the Tichenor is a third-order sequence bounding unconformity combined with a transgressive surface (Figs. 4 and 11). The Tichenor Limestone thus represents an initial transgressive lag deposit. The upper contact of the Tichenor is sharp, but conformable, with the overlying Deep Run Shale, perhaps as far west as Buffalo Creek in Erie County (Figs. 4 and 11). West of there, the overlying lower Moscow units, Deep Run Shale, Menteth Limestone, and Kashong Shale, grade into even more condensed carbonate layers until they are completely overstepped by a discontinuity below the Windom Shale Member. In western Erie County the top of the Tichenor is marked by an eroded hardground, particularly at localities near and along the Lake Erie shore. Hence the upper surface of the Tichenor is a disconformity representing depositional pinchout of muds onto the shallow western platform. Only the relatively major deepening event during deposition of the highest Hamilton Windom Shale allowed this disconformity to be overlapped and preserved (Fig. 4).

In addition to the sub-Tichenor unconformity, lowstand regressive erosion surfaces accompanied by transgressive limestone beds are seen at various other levels in the New York Devonian. For example, the base of the Jaycox Member fourth-order cycle, which immediately (and unconformably) underlies

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In addition to the sub-Tichenor unconformity, lowstand regressive erosion surfaces accompanied by transgressive limestone beds are seen at various other levels in the New York Devonian. For example, the base of the Jaycox Member fourth-order cycle, which immediately (and unconformably) underlies the Tichenor in western New York is represented by a thin (15–20 cm [~6–8 in]) crinoidal and coral-rich limestone unit (Hills Gulch bed) that resembles the Tichenor (Fig. 11). In the Genesee Valley area (Mayer et al., 1994), the Hills Gulch bed appears as a silty calcareous, moderately fossiliferous mudstone that conformably overlies soft gray shale and appears to cap a small cycle (Figs. 4 and 11). However, as the bed is traced westward from Livingston County (Genesee Valley area), it becomes a compact, silty, coral- and crinoid-rich packstone and, near Bethany, it begins to display a sharp, diastemic lower surface. Coarse crinoid and coral debris occurs in vaguely graded layers within the Hills Gulch bed (Mayer, 1989). The Hills Gulch horizon oversteps the Spafford Member near the Erie-Genesee county border, and the erosion surface below the Hills Gulch bed cuts into the underlying Wanakah Shale. The base of the Hills Gulch bed resembles that of the Tichenor, with large burrows extending down into the underlying shale. The Hills Gulch bed is instructive in that it demonstrates that the pattern of erosion evident in major sequences may also develop during lowstands of lesser fourth-order magnitude. This same pattern of erosional overstep near the basin margin is evident at many stratigraphic levels.

A final example of the regressive-type (lowstand) unconformity is the upper contact of the Moscow Formation at the top of the Hamilton Group (Fig. 4). In western New York this is a complex multiple event disconformity (the Taghanic unconformity), in part representing lowstand erosion (Johnson, 1970). Previous work has shown that the upper contact of the Windom Shale is a regionally angular unconformity which increases in magnitude to the northwest (Brett and Baird, 1982; Baird and Brett, 1986). This truncation surface can be traced into the Finger Lakes region, where it occurs beneath the 0–10-m-thick (0–35 ft), middle to late Givetian Tully Limestone (see Heckel, 1973), which intervenes between the Moscow and Genesee Formations (Fig. 4). From Canandaigua Lake westward to Erie County, this truncation surface at the top of the Windom Shale floors latest Givetian black shales of the Genesee Formation.

It is evident, therefore, that some or much of the Windom truncation surface was developed prior to deposition of the Tully. Basal Tully carbonates are massive micrites representing shallow subtidal (and possibly lagoonal) shelf conditions, as suggested by the occurrence of rare stromatolites (Heckel, 1973). Where the Windom truncation surface is overlain by the Tully Limestone, the contact, like that of the Tichenor, is marked by large tubular burrows, some of which extend downward 10–20 cm (4–8 in) into the Windom Shale (Fig. 12A).

In western New York, the entire Tully deposit has been removed by a still younger erosion surface; this latter discontinuity is of a very different nature and is associated with a period of sediment starvation during maximum highstand (Fig. 14). Nonetheless, this latest unconformity is combined with the pre-Tully sequence boundary unconformity in western New York (Figs. 4 and 14).

Sub-Tichenor, sub-Hills Gulch, and sub-Tully unconformities involved the removal of stiff compact mud. We cannot rule out that some of this erosion, at least in basin margin areas, may have been subaerial. However, at least later stages of sediment removal were surely submarine. Dislodgment and resuspension of the firm muds may have been aided by active burrowing by organisms adapted to excavation of compact mud. Vestiges of this fauna are seen in the megaburrows at the bases of limestone beds (Fig. 12). Abrasive scour of the sea floor may have been aided by tractional and saltational transport of skeletal sands and gravels during storms. In some cases, it is evident the pre-existing burrows or furrows were present and widespread prior to deposition of the lowest calcarenite sediments. In fact, basal sediments of the limestone may have accumulated as tubular tempestites, i.e., infillings of deep burrows (see Wanless, 1986).

Discontinuity surfaces associated with rapid sea-level rise and sediment-starvation

A second type of erosion surface is associated with phases of rapid deepening and the basal contacts of black shales (Fig. 13). The lower contacts of many Devonian black shales or lag beds are sharp, and regional truncation of the underlying strata is typical (Baird and Brett, 1991). Such erosion surfaces are nearly planar to gently undulatory and razor sharp. No burrowing of the underlying sediments is evident, indicating that submarine erosion took place under anoxic or at least dysoxic conditions (see Figs. 12–15). There is evidence for dissolution of carbonates at the surface, and the overlying thin lag beds commonly contain chemically resistant particles such as phosphate nodules, quartz grains, conodonts, fish bones and teeth, and, in several instances, diaclasts of reworked early diagenetic pyrite (Baird and Brett, 1986; 1991). Pelmatozoan ossicles may be present, but other carbonate detritus is rare or absent in the lag beds. Lag beds typically range from thin sheets only millimeters thick to pods or lenses up to 20 cm (8 in) thick and several meters across (Figs. 4 and 14). Internally, the lag beds may be laminated or vaguely cross-laminated and may display thin partings of black shale. Small lenses of debris may occur, up to several centimeters above the erosional base, interbedded with black shales. Pyritic lag beds, typified by the Leicester Pyrite (Fig. 14), have been described in detail elsewhere (Baird and Brett, 1986, 1991; Baird et al., 1988). Sharp discontinuities are characteristic of most Devonian black shale–limestone contacts. There is evidence for erosion and dissolution of underlying sediments, and thin lag beds are present above the erosion surfaces. Typically, the underlying beds are condensed outer shelf limestone facies.

One such contact of black shale on carbonate is the contact of the Union Springs (or Bakoven) black shale on the upper Onondaga Limestone (Seneca or Moorehouse members) (Fig. 4A). This contact is rarely exposed, but it is sharp and marked by concentrations of onychodid fish teeth, placoderm armor, and phosphate nodules on the upper surface of the

Onondaga (Fig. 13B). Where the succession is more nearly continuous, as in the Cayuga–Seneca Lake area, alternating 10–25-cm thick (4–10 in) micritic, styliolinid ("ribbon") limestones and dark gray shale beds (i.e., Seneca Member) underlie the black shale. Although these beds appear to be transitional between massive Onondaga carbonate and, overlying black, organic-rich shale, there is still a scattering of fish bones, quartz grains, and rare phosphatic granules above the uppermost ribbon limestone of the Seneca Member. East and west of this area, the basal contact of the Marcellus black shales is increasingly erosional and marked by a distinct bone bed. Both east and west of the central Finger Lakes trough, the transitional black micritic limestone-shale beds of the Seneca Member appear to be somewhat beveled by the erosion surface, such that the bone bed rests on the beveled upper surface of the underlying Moorehouse at some locations in the Hudson Valley.

A similar but more cryptic contact occurs at the upper surface of the Tully Limestone in the southern Cayuga Lake region. Here, again, the upper beds of the Tully (Fillmore Glen beds of Heckel, 1973) appear to be transitional, with decimeter-scale, dark, styliolinid limestones alternating with dark gray to black shales through about 1 m (3 ft) of section. The upper surface of the highest limestone bed displays a thin lag of styliolinids and silty, pyritic debris. As with the upper contact of the Onondaga Limestone, the transitional Fillmore Glen beds and successively lower units of the Tully Limestone are truncated by the erosion surface to the northwest and perhaps also to northeast of Cayuga Lake. Between Seneca and Canandaigua lakes, the lowest bed of the upper Tully member (Bellona Coral Bed) is beveled, and a thin lag of bone, crinoidal grains, and minor reworked pyritic burrow tubes (Leicester Pyrite Member) rests sharply on the planar upper surface (Figs. 14 and 15).

At Canandaigua Lake, the entire Tully is beveled and the overlying black Geneseo Shale rests sharply on the upper beds of the Windom Shale (Fig. 11). As noted above, some upper Windom beds had already been truncated beneath the sub-Tully sequence boundary, a lowstand erosion surface. Hence, the unconformity between the Windom (upper Hamilton Group) and Geneseo is actually a compound erosion surface marking the juxtaposition of the basal Tully, middle Tully, and upper Tully unconformities. How much additional erosion of Windom beds can be ascribed to the last erosion event is not clear. However, some additional scour of the bottom during latest Givetian (post-Tully) time must have taken place. Where the erosion surface underlies the Tully, it is irregular and evidently heavily burrowed, whereas the combined unconformity separating

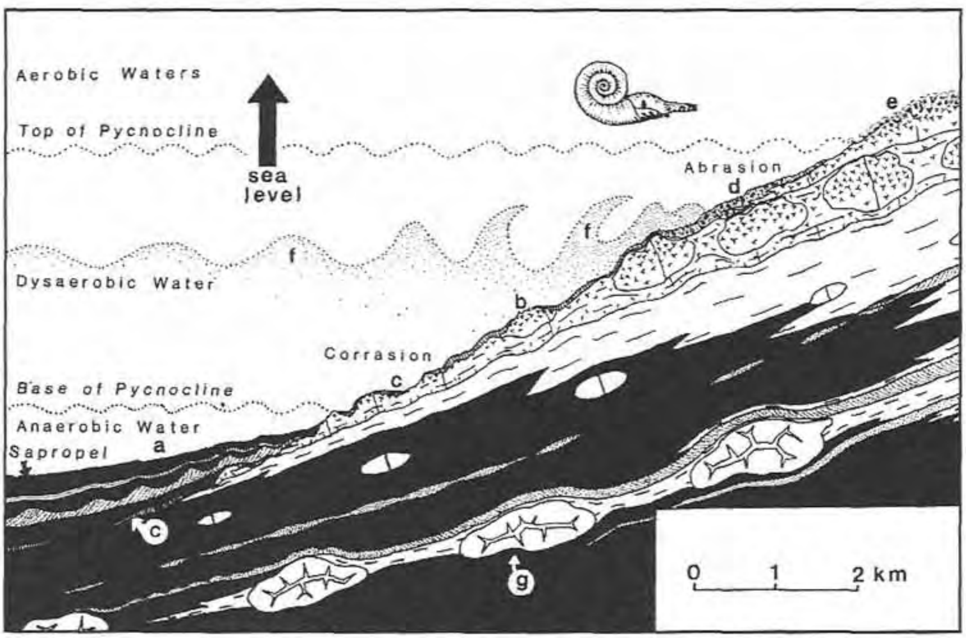

Figure 15. Hypothetical model for the genesis of corrasional discontinuities through impingement of shoaling internal waves against a sloped basin margin. Internal waves, generated within the pycnocline of a stratified basin, propagate along the temperature-density boundary and shoal where the boundary intercepts the sediment-starved basin margin slope. This causes minor erosion and serves to clear the substrate of loose sediment. Upslope migration of the zone of shoaling during transgression extends the discontinuity surface upslope and marginward while the downslope parts of that surface are buried by overlapping black mud deposits in the quiet dysoxic to anoxic zone below the pycnocline. This model may explain the frequent occurrence of discontinuities, often very widespread, beneath black shale deposits, and it explains the absence of transitional passage facies between shallow-water deposits below many of these contacts and deep-water anoxic facies above them (from Baird et al., 1988).

Geneseo from Windom Shale is nearly planar and knife sharp. This indicates a modification of the old sub-Tully erosion surface by later processes. The Tully carbonate was not merely removed by dissolution; physical abrasion was also involved. In addition, the contact between the Windom and Genesee, like that between Tully and Genesee, is marked by lenses of fish bones, conodonts, quartz pebbles, and, above all, reworked pyrite debris (Fig. 14). The pyrite lenses, termed Leicester Pyrite Member of the Genesee Shale, are clearly related to deposition of the black Geneseo muds, because they are interbedded with the black shale (Fig. 14). Conodonts indicate also that the lenses are slightly younger than the Tully Limestone (Huddle, 1981). The occurrence of reworked pyrite in the Leicester indicates that erosion and concentration of pyrite diaclasts, derived from the Windom mudstone, took place under low-oxygen to anoxic bottom conditions (see Baird and Brett, 1986; Baird et al., 1988).

We have identified several similar erosion surfaces at the junctions between black shales and underlying calcareous mudstones or concretionary carbonate beds. For example, we documented two thin limestone intervals (Fir Tree and Lodi limestones) within the Genesee Formation (latest Givetian to early Frasnian) that are beveled beneath black shales (Baird et al., 1988; Kirchgasser et al., 1988).

Any model for the genesis of black shale-roofed disconformities must explain the following features: (1) The erosion occurs during apparent intervals of sea-level rise rather than fall. (2) They are nearly planar or runnelled surfaces overlain by dark gray or black laminated shales. (3) The surfaces are frequently underlain by concretionary argillaceous micritic carbonate or calcareous mudstones with distinctive deeper water biofacies (auloporid corals, small brachiopods, plus pelagic forms such as styliolinids and cephalopods). (4) Surfaces are sharply overlain by lag deposits of geochemically resistant allochems such as phosphate, quartz, bone, and conodonts, but lacking most carbonates. (5) Reworked pyrite is commonly present, at least where pyritic mudstones, serving as source beds, underlie the unconformities; pyrite should be stable only under anoxic conditions. (6) Surfaces are sharp where overlain by black shales but seem to fade into conformity where overlain by gray dysaerobic deposits in the presumed upslope direction (Fig. 15).

From the standpoint of sequence stratigraphy, the major black shale–roofed disconformities are not sequence boundaries, but correspond to surfaces of maximum starvation or downlap surfaces of seismic stratigraphers. As noted above, both the upper Onondaga and Tully limestones appear to compose the bases of large-scale third-order sequences (Fig. 4). The upper nodular to tabular, argillaceous carbonates and interbedded shales appear to record a condensed section terminated by the abrupt change to deep-water black shales. Hence, the carbonates themselves compose a transgressive systems tract; this sharp upper contact is a sediment-starved surface recording early highstand (deepest-water) conditions. During the rapid transgressive phase, the carbonate factory was abruptly terminated and the offshore areas were also starved of siliciclastic sediment due to coastal entrapment. In the following highstand interval siliciclastics began to prograde seaward, both filling the basin and providing a sedimentary wedge that downlaps onto the surface of starvation (Fig. 15).

Relative sea-level rise also led to an elevation of the pycnocline (zone of density stratification), which caused flooding of formerly shallow water areas with deeper and slightly denser bottom water (Fig. 15). Relative sediment starvation and high productivity in surface waters enabled abundant organic matter to accumulate below the pycnocline, producing widespread oxygen deficiency on the sea floor. During shallow-water intervals, anoxic water was confined to basin centers, but in major transgressions it expanded onto the basin margin ramps and shelves.

This pattern fits with the typical sequence model for development of surfaces of maximum starvation. A critical adjunct to the model, however, is that substantial submarine erosion may occur during times of sea-level rise in addition to sedimentary condensation. The thin concretionary beds that underlie the black shales reflect the onset of sedimentary condensation. Once sediment starvation reaches a critical point, however, nondeposition may give way to erosion. One way of looking at this problem is that minor erosion, due to storm-generated currents, or deep-flowing density currents, plus chemical corrosion is always taking place. However, during times of sediment input this erosion is more than balanced by sedimentation. At times of total sediment starvation, ambient erosion processes may begin to play a more dominant role and the balance shifts from one of slow accumulation through nonsedimentation to erosion without any change in energy. Hence, the key to this type of erosion is lack of new sediment input.

Part of the explanation for erosional loss of section may well involve dissolution of the older carbonates by corrosive (low pH, or carbonate undersaturated) bottom water. We have noted that carbonate grains are often conspicuously absent in the lag deposits that overlie disconformities (see Figs. 14 and 15). Nonetheless, there is also evidence for increased erosional energy in a distinctive pattern along basin margins because relatively heavy clasts, such as pyrite, were apparently dislodged from firm siliciclastic clays and concentrated on the sea bed. We argue that the erosion process is also concentrated along particular levels of the slope to basin profile and that it dies out upslope. These positions may be marked by thin pyrite lenses; in the latest Givetian Geneseo Member, we observed pyrite lag beds on either side of the basin center (Baird et al., 1988).

Furthermore, this erosive process occurs in settings that were primarily anoxic. Wave and/or current energy may be focused along internal water-mass boundaries (Fig. 15). If internal waves propagated along the pycnoclines, then these waves would impinge along the sea floor at the point where the pycnoclines intersected sloping basin-margin ramps. In such a case a topographically narrow belt of erosional energy would migrate upslope during periods of rapid sea-level rise, when the pycnocline also expanded upward, producing a transgressive unconformity overlain by diachronous, anoxic, laminated deposits (Fig. 15). This process would produce a pattern of erosion that would be most intense along lower to midslope regions of the basin margins and would die out upslope.

Hence, major erosion surfaces appear to have developed in deep water during times of relative sea-level rise. The truncation probably involved (1) sediment starvation in distal regions; (2) development of corrosive anoxic bottom waters that dissolved carbonates; and, (3) internal waves along the rising pycnocline that eroded the sea floor and concentrated pyritic lag deposits.

Discontinuity surfaces and condensed deposits associated with initial sea-level drop: "Precursor beds"

As noted, the Middle Devonian Hamilton Group has been subdivided into a series of shallowing-deepening cycles (see Brett and Baird, 1985, 1986). Many of the shallowing-upward half-cycles display abrupt bases along which beds of relatively condensed, shell-rich, and oxic facies sharply overlie dark gray to black offshore, dysoxic shale deposits (Fig. 12C). These sharply-based "precursor" beds are traceable for up to 200 km (125 mi) between the Erie County region and the central Finger Lakes area. These beds are normally a few centimeters to about 0.5 m (1.5 ft) thick and are characterized by shell-rich sediments; in some cases, their bases display evidence for submarine erosion (Fig. 16). Thus, for example, at the base of the well-known Centerfield Member a sharp surface separates a gray shell-rich bed (Peppermill Gulf Bed) from the underlying black *Leiorhynchus*-bearing Levanna Shale, and provides a reference horizon at which of the Centerfield lower boundary may be drawn (Gray, 1984, 1991; see Figs. 10 and 11). A comparable bed (Stafford-Mottville "A") occurs at the base of the widespread Stafford-Mottville limestone interval (Grasso, 1986). Minor cycles such as the lower Wanakah *Pleurodictyum* beds (see Miller, 1986, 1991) also display sharp condensed beds which are typically condensed shell-rich layers (Fig. 16).

The term "precursor" is used for those mid-highstand condensed beds because they appear to mark the beginnings of major shallowing cycles and because their faunas resemble those that occur near the cap of the shallowing cycle. In these senses the beds are the "harbingers" of major shallowing intervals.

The succession above a precursor bed consistently displays (a) a minor interval of less-fossiliferous shales somewhat resembling those below the beds, followed by (b) a distinct shallowing-upward trend in the highest fourth-order cycle; this trend culminates in relatively coarse regressive (lowstand?) deposits near the basin center but in a sequence-bounding erosion surface around the flanks of the basin (Fig. 4B).

Hence the apparent pattern of the late highstand phases is (a) an abrupt shallowing recorded in the thin condensed "precursor" bed; (b) a minor deepening interval immediately above this unit; and (c) a shallowing interval for the remainder of the cycle. The "precursor bed" is sharply based, highly fossiliferous, and is followed by an interval recording minor deepening and then shallowing. In these ways it resembles the major transgressive limestone of a sequence as a whole. In places, there is even evidence for erosion below the precursor beds in the form of reworked concretions, and there is little or no preserved record of a shallowing-upward cycle beneath the bed; hence it does not resemble a top of parasequence shell bed, because there is no parasequence below it. It may, in fact, represent the capping shell beds of two or more extremely abbreviated (erosionally "telescoped") parasequences (Fig. 10). However, the latter are not distinct except perhaps in more eastern outcrops, proximal to the sediment source area.

Precursor beds may contain faunal elements otherwise restricted to the caps of the shallowing cycles. These common elements may simply record the existence of a fauna that is relatively intolerant of turbidity and therefore developed primarily during times of siliciclastic sediment shut-off. However, such faunas are usually better developed and more diverse in the cap beds of shallowing upward cycles or bases of transgressions.

In western New York, precursor beds are recorded as thin, calcareous, shell-rich markers within shales; they are not particularly prominent, but they represent important condensed intervals in which reworked concretions may be common (Fig. 16). Precursor beds are particularly well developed in central and west-central New York, but tend to die out toward the east (see Pepper Mill Gulf bed; Figs. 10 and 11). A common feature of many precursor beds is that they display distinct lateral gradients in biofacies (Fig. 16). This suggests that they record isochronous intervals of sediment starvation.

The precursor beds appear to splay eastward into a series of less-distinct shell-rich beds capping minor decimeter-scale mudstone intervals. These latter shell beds exhibit an overall shallowing trend in that each successive bed appears to carry a shallower water biofacies than the preceding one. In sections farther offshore (westward), the individual shell beds may merge to form the precursor bed. The precursor bed thus appears to record an interval of sediment starvation and sediment downlap associated with one episode of abrupt relative shallowing (see Figs. 10B and 11). These condensed beds are widespread but less laterally persistent than those associated with the transgressive systems tracts (i.e., basal transgressive limestone and maximum flooding surface beds). Precursor beds splay out and become unrecognizable east of the west-central New York region, whereas transgressive condensed beds can be traced considerably farther east. In addition, whereas minor parasequences become increasingly condensed upward in the transgressive systems tract, those of the late highstand interval become less condensed upward; the highest one is a major thick interval that commences with the precursor bed itself. Finally, the overall pattern is reversed from that in a transgressive systems tract; i.e., it is progradational and upward shallowing rather than retrogradational and deepening.

The association of condensed shell-rich beds with sea-level drop is counterintuitive and demands an explanation. In the sequence stratigraphy model the precursor beds lie between early and late highstand, i.e., at the change from aggradational deep-water facies to progradational, upward-shallowing successions. These precursor beds are typically followed by thick sections of silty, sparsely fossiliferous mudstone that appear to record major influxes of siliciclastics associated with the continuing sea-level drop. The initial period of offshore sediment starvation (which is not as extensive as that for typical transgressive sediment-starved beds; see below) may record disequilibrium conditions associated with the rapid change in sea level. For example, Posamentier et al. (1988) argued that rapid drops in base level may cause rivers to regrade to an equilibrium profile. During this period of readjustment of stream beds, a considerable amount of sediment may actually be deposited in subaerial environments, a phenomenon referred to as "subaerial accommodation"; sediment is being trapped near the lower reaches of streams, resulting in a brief period of sediment starvation in offshore areas. Erosion associated with the bases of precursor beds may, in part, record lowering of storm wave base or increased scouring by gradient currents. Erosive surfaces of the bottoms of these beds are most accentuated in areas that appear to have had steep northwest-dipping slopes, as in the case of shell-nodule–rich diastemic beds in the Ludlowville Formation of the Cayuga Lake region (see Fig. 16; Baird, 1981; Baird and Brett, 1981).

Any model for the genesis of precursor beds must also explain the facts that these horizons display (a) sharp to disconformable bases and relative condensation; and (b) and abrupt

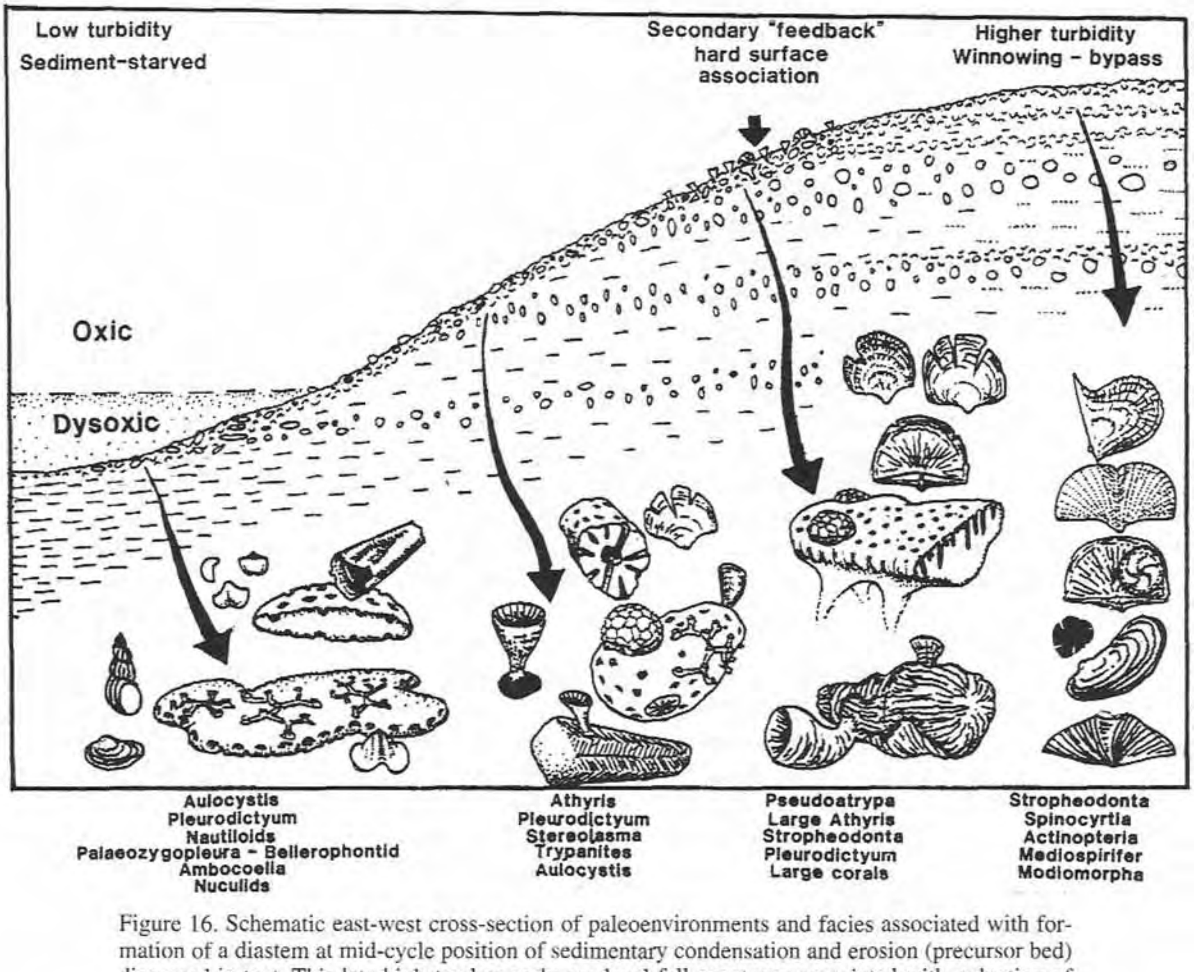

Figure 16. Schematic east-west cross-section of paleoenvironments and facies associated with formation of a diastem at mid-cycle position of sedimentary condensation and erosion (precursor bed) discussed in text. This late highstand or early sea-level-fall event was associated with reduction of sediment supply to a gently sloped sea floor. In the region of slope, intense bioturbation of surface mud plus episodic scour of this water-rich churned sediment resulted in erosion of the sea floor and sediment export into deeper basin areas. The record of this erosion is revealed by westward (basinward) overstep of concretion-rich intervals by the precursor bed and by the presence of abundant bioencrusted reworked concretions, prefossilized steinkerns, and degraded shell material within the precursor bed. However, no discrete erosion surface is preserved at the diastem position owing to the burrowing process. The presence of a northwest-sloping paleoslope is revealed by downslope faunal and lithologic gradient both within the precursor bed and in synjacent deposits.

change from dysoxic to oxic, presumably deeper to shallower water. Finally, we must account for the fact that these beds appear to be connected closely in time with the development of more major shallowing cycles. In short, these are condensed and sometimes erosionally based beds associated with initial sea-level drop or the beginnings of a shallowing cycle. However, it is also probable that such beds reflect a smaller-scale, shallowing-deepening cycle superimposed on the larger one. If so, some precursor beds may be thought of as accentuated caps of heretofore unrecognized small-scale cycles. Keith Miller (1988) suggested that the abruptness of the bases of these horizons may result from the constructive interference of two different sea-level-fall processes. For example, a small-scale Milankovich band–induced eustatic sea-level fall might be superimposed on a larger-scale cycle.

SUMMARY

Three scales of sedimentary cyclicity are recognized at present within the Middle Devonian strata (late Eifelian–Givetian) of central and western New York. These include small-scale (fifth- and sixth-order or PAC-like) cycles equivalent to parasequences of sequence stratigraphic terminology; cyclothem-scale (fourth-order) cycles, which we term parasequence sets or subsequences; and large-scale (third-order) cycles that correspond to sequences. In offshore, sediment-starved western sections, parasequences are expressed as symmetrical vertical alternations of concretionary mudstone and nonconcretionary mudstone or shale. However, each parasequence thickens eastward into trough or basin depocenters before thinning again into shallower, current-winnowed, turbid shelf settings in central New York. Here, parasequences transform into upward-coarsening (upward-shallowing) asymmetrical cycles, bounded by base of parasequence (BOP) transgressive shell pavements at their base and top of parasequence (TOP) shell beds associated with winnowed silts and sands near cycle tops.

Fourth-order subsequences contain bundles of fifth- and sixth-order parasequences. Because these units may contain numerous small-scale cycles, the variety of facies within them is greater, sometimes ranging from black shale at early highstand to cross-bedded sandstone at maximum shallowing. As with parasequences, there is an eastward change in cyclic motif from asymmetrical sequence-like motifs on the western ramp in western New York to Ontario to symmetrical to subsymmetrical thin cycles in offshore (sediment starved) western New York settings, through a depocenter interval where the cycle is symmetrical and thick, to a nearshore strongly asymmetrical, upward-coarsening motif that reflects sediment-bypass conditions on a shallow shelf. Unlike parasequences, fourth-order subsequences display complex (multiple) transgressive shell pavements climaxed by highly condensed phosphatic shell beds that record maximum flooding of the sea floor.

Five large-scale (third-order) sequences are recognized in the Hamilton Group of the study area. These are asymmetrical, thick complexes and are marked by a discontinuity and transgressive systems tract at their bases. Maximum flooding is marked by conspicuous, regionally extensive phosphate or bone beds. Aggradational highstand facies associated with this scale of cyclicity account for most of the thick offshore mudstone deposits that are observed in the Middle Devonian section. We predict that, in eastern New York, well-developed sequence bounding unconformities over nearshore marine or nonmarine deposits should be present, and that these surfaces should correlate to those below limestone beds that floor third- and fourth-order cycles in western New York.

Significant discontinuities occur, predictably, at three positions within Middle Devonian marine shale and carbonate sequences of western New York. Pronounced erosional unconformities, commonly with irregular, burrowed contacts, occur on the bases of shallow-water carbonate beds corresponding to rapid lowering of relative sea level. These contacts (e.g., basal Tichenor Limestone, basal Tully Limestone) correspond to sequence or subsequence boundaries. In some cases these disconformities are composites of lowstand submarine (and possibly subaerial) erosion surfaces and transgressive ravinements.

A second type of important discontinuity occurs at the bases of major highstand deposits, roofed by black shales. These are typically planar or gently furrowed disconformities that lack burrows. Such contacts are marked by thin concentrations of phosphate, pebbles, glauconite, conodonts, bones, and/or reworked pyrite. Such surfaces are associated with condensed sections formed during times of rapid sea-level rise; as such, they correspond to accentuated surfaces of maximum sediment starvation or downlap surfaces of sequence models. Erosional processes may include deep storm waves, density currents, or internal waves propagated along pycnoclinal water-mass boundaries.

The third category of discontinuity and condensed interval occurs at the boundary between early highstand (deepest water) deposits and progradational late highstand deposits. As such, they appear to be associated with the inflection point between latest phases of sea-level rise and early stages of regression. They are overlain by shallowing-upward intervals of gray mudstone to silty calcareous mudstone that may culminate in a sequence or subsequence bounding unconformity below shallow-water transgressive deposits. These mid-highstand beds commonly display reworked concretions and relatively shallow water biofacies. They appear to initiate cycles of rapid shallowing and mimic some characteristics of cycle-capping carbonates (limestones above sequence or subsequence boundaries); hence, we term such units "precursor" beds. Such beds are not predicted by existing sedimentary models. We infer that they may represent brief sediment-starved intervals during periods of sediment disequilibrium resulting from the lowering of sea level. Submarine erosion probably involves a combination of bioturbation and storm-generated gradient currents.

All three types of discontinuities and associated condensed beds provide outstanding regional and roughly isochronous markers that enable detailed physical correlations of strata. They also subdivide the stratigraphic column into bounded, genetically related bundles. Mapping of these discontinuities facilitates interpretation of sea-level variations and basin dynamics.

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