Structural sequences and styles of subsidence in the Michigan basin

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ABSTRACT

Subsidence in the Michigan basin produced ~5 km of sedimentation over a period of more than 200 m.y. during Paleozoic time. Utilizing well-log correlations and constrained by compaction corrections and estimates of paleobathymetry, we recognize four different styles of subsidence in the basin: trough-shaped, regional tilting, narrow basin-centered, and broad basin-centered. Subsidence began as a trough-shaped, northerly extension of the Illinois basin during Late Cambrian to Early Ordovician time. This was followed by narrow, basin-centered subsidence in Early to Middle Ordovician time. Basin-centered subsidence ceased for ~30 m.y. during long-wavelength (>1000 km) eastward tilting in Middle to Late Ordovician time, a pattern incompatible with thermal-contraction subsidence models. Basin-centered subsidence resumed in Silurian time, but with a broader distribution. A second episode of narrow, basin-centered subsidence occurred in latest Silurian through Middle Devonian time and was replaced by broad, basin-centered subsidence at the end of Middle Devonian time. The geometry of Upper Devonian and younger Paleozoic deposits suggests another eastward-tilting event, but results remain inconclusive due to erosion of strata and uncertainties in their paleobathymetry. In addition to these subsidence patterns, two distinct unconformity styles are present: basin-wide and marginal erosion. There is no evidence for significant basin-centered unconformities as predicted by purely thermal mechanisms. A history of episodic subsidence reactivations is interpreted as the result of a stress-induced, crustal-weakening mechanism for the narrow, basin-centered subsidence, whereas broad basin-centered subsidence is interpreted as thermal contraction related to lower crustal attenuation during the narrow-subsidence episodes. Recently proposed dynamic topography related to initiation of Ordovician subduction provides a driving mechanism for long-wavelength eastward tilting. Together with a temporal correlation to Appalachian tectonism, these mechanisms provide a plate tectonic framework for the history of the Michigan basin.

INTRODUCTION

The Michigan basin is a nearly circular, intracratonic basin 400 km in diameter and 5 km deep with only minor structural disruption (Fig. 1), yet it remains without a definitive origin (Leighton, 1996). The geometry of the Michigan basin has led to numerous proposals for basin subsidence mechanisms, including thermal contraction following development of an isolated "hot spot" (Sleep, 1971; Nunn and Sleep, 1984; Houseman and England, 1986), metamorphic phase changes in the crust (Haxby et al., 1976; Middleton, 1980; Ahern and Dikeou, 1989; Hamdani et al., 1991), lithospheric stretching (McKenzie, 1978; Klein and Hsui, 1987), free thermal convection (Nunn, 1994), and intraplate stress mechanisms (DeRito et al., 1983; Lambeck, 1983; Howell and van der Pluijm, 1990). Contention among these proposals centers mainly on irregular basin subsidence rates (Sleep and Sloss, 1978; Quinlan, 1987; Howell and van der Pluijm, 1990).

Decompacted basement subsidence curves for three representative wells from the western, central, and eastern portions of the basin document the temporal and areal distribution of subsidence variability (Fig. 2). Sequence A displays rapid subsidence in the western and central basin wells, but is barely present on the eastern margin. Sequence B has a pronounced subsidence reactivation in the basin center, but is completely absent along the eastern margin due to nondeposition and erosion. Subsidence rates increase eastward for sequence C, and all three wells show significant subsidence for sequence D.

This irregular basin subsidence history may hold the key to resolving the origin of the Michigan basin. Subsidence curves from other cratonic settings, such as the Illinois, Paris, and North Sea basins (Heidlauf et al., 1986; Brunet and LePichon, 1982; Sclater and Christie, 1980), show rapid subsidence in their early history followed by a long period of gradually declining subsidence, which is typical of thermal contraction following rapid lithospheric attenuation (e.g., McKenzie, 1978). In contrast, the Michigan basin reveals a number of subsidence reactivations and cessations (Sleep and Sloss, 1978).

Subsidence rates, however, provide only a part of the information. Additional evidence is derived from the style of subsidence in the basin. A cursory examination shows that the Michigan basin has a very simple geometry; it is nearly circular in shape, has relatively smooth structural contours on its sedimentary horizons, and has only minor syndepositional faulting that has an insignificant impact on the basin-scale geometry of stratigraphic units. In the most recent full compilation of stratigraphic data, Fisher et al. (1988) concluded that continuous, basin-centered subsidence appears to be compatible with thermal contraction subsidence mechanisms, although they added that some shifts in sediment depocenters require different subsidence mechanisms.

In contrast, Howell and van der Pluijm (1990) argued that the early history of the Michigan basin included significant changes in basin subsidence style (shape of the basin), which were accompanied by changes in subsidence rate (Fig. 2) and a response of the depositional systems to the pattern of subsidence. These changes in subsidence style provide constraints on the possible subsidence mechanisms for the basin (cf. Coakley et al., 1994; Coakley and Gurnis, 1995). In addition, recent advances in crustal rheology and mantle dynamics reveal that these two

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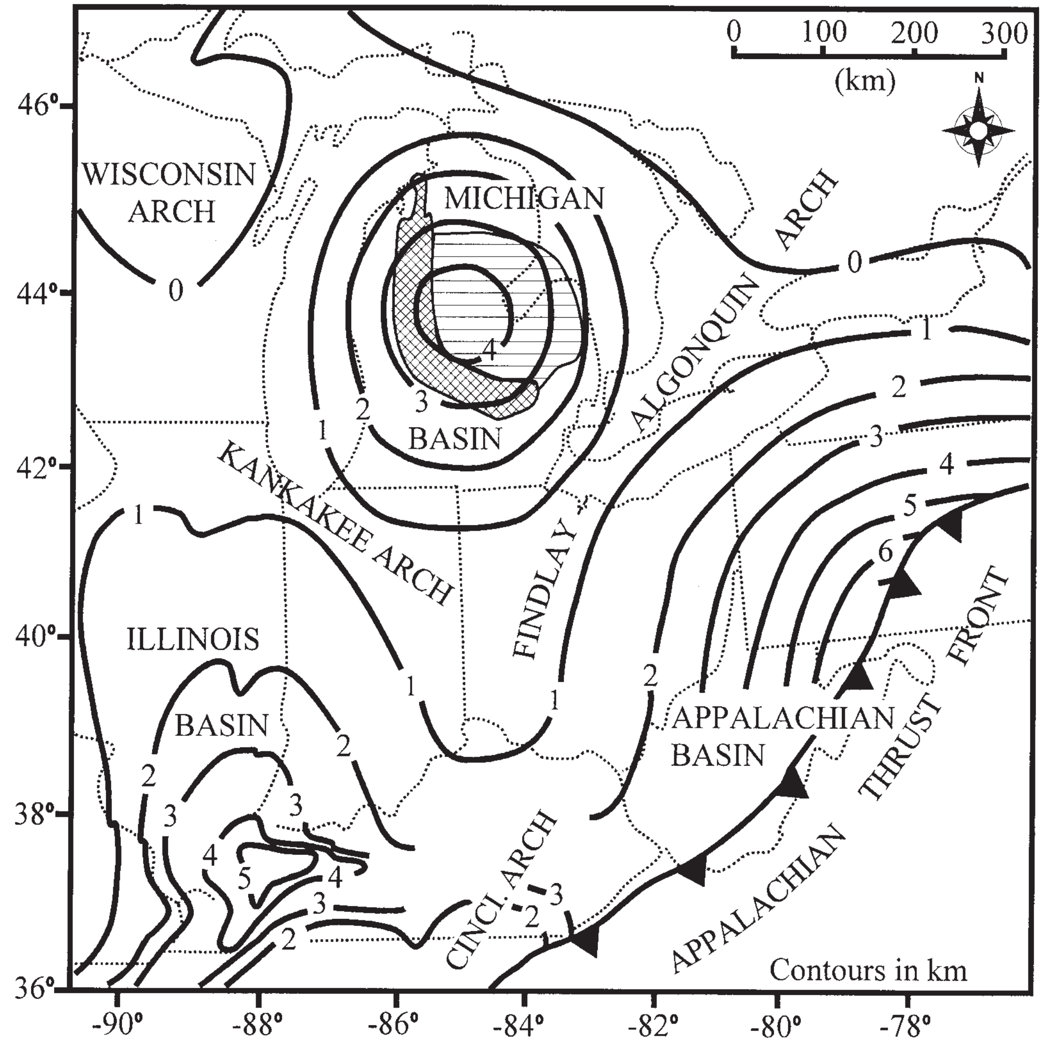

Figure 1. Total thickness of Phanerozoic sedimentary rocks in the Michigan basin region (contours in kilometers). Area with fine grid represents positive gravity anomaly in central Michigan corrected for basin fill (after Haxby et al., 1976). Striped area is location of Keweenawan rift based on drilling and Bouguer gravity (after Hinze et al., 1975).

factors exert significant influence on the stratigraphic signature of platform regions through their roles in continental deformation (Brace and Kohlstedt, 1980; Molnar, 1988), epeirogeny, and eustasy (Gurnis, 1992a, 1992b).

This paper presents a set of new correlations and interpretations for the stratal geometries of the Michigan basin, building on data presented by Howell and van der Pluijm (1990). These include new data made available since previous compilations (Lilienthal, 1978; Fisher et al., 1988), new correlations made possible by improved well-logging techniques, and new stratigraphic groupings that differ from those used in previous studies (Fisher et al., 1988). The correlations are presented in a framework of structural sequence stratigraphy, which offers an alternative framework with which to analyze basin evolution.

STRUCTURAL SEQUENCES

Howell and van der Pluijm (1990) grouped strata from the early history of the Michigan basin into stratal sequences that were defined by significant changes in basin-subsidence patterns. This represents a different type of sedimentary sequence than the unconformity-bounded depositional sequences of Mitchum (1977) and Van Wagoner et al. (1988) and the genetic sequences of Galloway (1989). A structural sequence is defined here as a succession of sedimentary strata in a basin that is bounded by distinctive changes in the pattern of basin)subsidence. These sequences may correspond to unconformity-bounded depositional sequences, or to groups of depositional sequences termed composite sequences (Mitchum and Van Wagoner, 1991), but they need not correspond directly to any of these other sequence types. This definition of a structural sequence corresponds more closely to the original working definition of a sequence: "a sequence comprises an assemblage of strata exhibiting similar responses to similar tectonic environments over wide areas, separated by objective horizons without specific time significance" (paraphrased by Wheeler [1958] from Sloss et al. [1949]).

To a first-order approximation, isopach maps can be used to define structural sequences (Kay, 1945), assuming that the changes in thickness of the strata under consideration are due to lateral subsidence rate variations during deposition. Calvert (1974), however, recognized that isopach maps record the structural deformation of a basin between the time of deposition of the lowermost and the uppermost strata of the interval examined only if certain reliability criteria are met. Compaction of the interval of interest, syndepositional compaction of the underlying units, and the paleobathymetries of both included and bounding strata must be considered.

METHODOLOGY

Petrophysical logs from more than 500 wells drilled in the Michigan basin for petroleum exploration and subsurface waste disposal were correlated on the basis of lithostratigraphy (Howell, 1993). Isopach maps presented in this paper include the locations of wells used to constrain the contours. New stratigraphic correlations have been obtained from this analysis and compared with those of Lilienthal (1978), Fisher et al. (1988), Catacosinos and Daniels (1991), Catacosinos et al. (1991), and Coakley et al. (1994).

To determine the preliminary breakdown of strata into structural sequences, isopach maps from 43 stratigraphic intervals were examined in conjunction with well-log cross sections, lithologies, and core data. Possible depositional environments were considered for each unit and lithologic facies changes examined for supporting evidence, leading to estimates of maximum and minimum probable paleobathymetries. In order to minimize possible effects of paleobathymetry on subsidence analysis, structural sequence boundaries were chosen only at intervals where the estimated paleobathymetry is small (i.e., relatively shallow water depths, <30 m). Thin stratigraphic intervals were not considered for structural sequence status when the uncertainty in paleobathymetry constituted a significant proportion of the thickness of the proposed sequence; uncertainty in paleobathymetry of strata constitutes one of the important limitations of fine-scale stratigraphic resolution of geodynamic behavior (Calvert, 1974; Célérier, 1988). Intervals were also examined for evidence of gradational facies contacts, synsedimentary faulting, basinal extent of erosional unconformities, and postdepositional dissolution of soluble lithologies, each of which could introduce uncertainty. Lithostratigraphic contacts defined by time-transgressive facies changes produce stratigraphic surfaces that are typically unacceptable as structural sequence boundaries; isopach maps derived from stratigraphic units with facies contacts for boundaries introduce multiple uncertainties into a structural analysis, including both paleobathymetric variability and diachroneity of the boundary. Isopach maps were compared for significant changes in relative shape, which were weighed against the possible uncertainties listed here, and stratigraphic

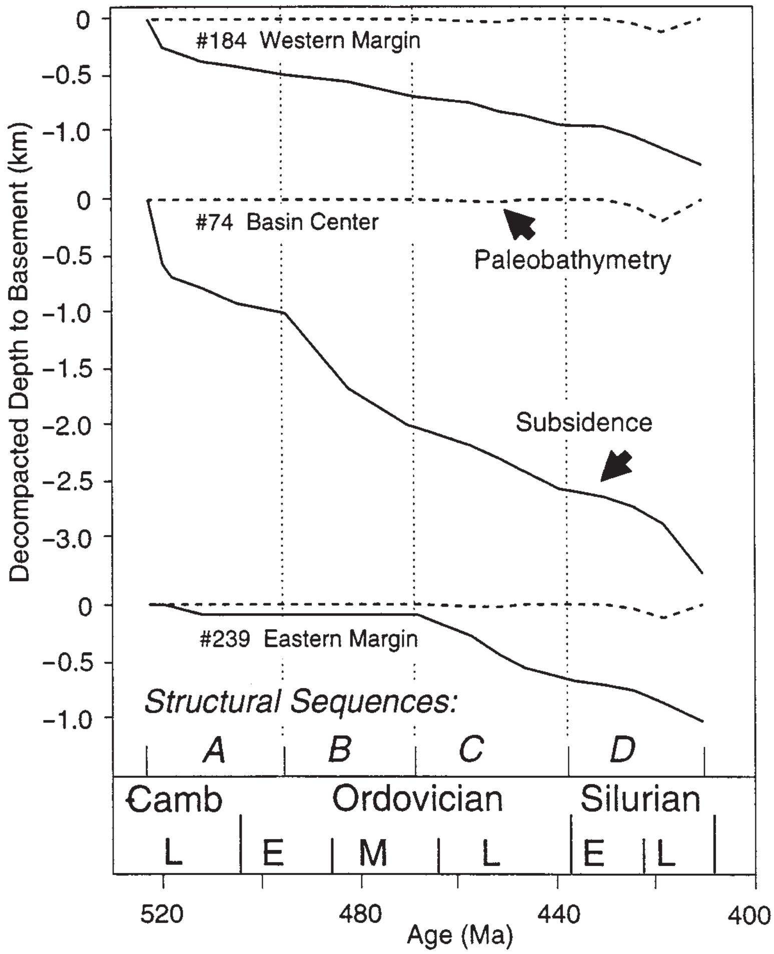

Figure 2. Decompacted basement subsidence curves for three Michigan basin wells for the first 120 m.y. of the basin's history. Note that subsidence patterns differ across the basin for each stratigraphic interval identified as a structural sequence. Numeric ages follow Fisher et al. (1988) and Smith et al. (1993). The bottom curves (solid lines) represent the depth to basement through time after correcting for paleobathymetry (uppermost, dashed curves) and compaction. Well numbers refer to listings in Howell (1993): #184—Amoco Schiller 1-10, Oceana County; #74—Hunt-Martin 1-15, Gladwin County; #239—MichCon Osterland 1-14, St. Clair County.

intervals were grouped informally into provisional structural sequences based on isopach geometries. Each structural sequence was examined internally for consistency of the interpreted subsidence geometry (e.g., basin-centered versus wedge-shaped basin subsidence patterns).

Basement subsidence was quantified by decompacting each structural sequence and the underlying sequences, and calculating the change in basement depth during deposition of each sequence. This change in depth to basement (hereafter referred to as ΔDB) is used to define the subsidence for each structural sequence, rather than a derivative of the ΔDB such as backstripped residual subsidence. Local backstripping (Steckler and Watts, 1978) essentially produces a similar shape as ΔDB, but a smaller amplitude due to the "stripping" of the sedimentary load in the basin. Airy assumptions (local isostatic balance) are probably unrealistic in a small, flexural basin such as the Michigan basin, where the basin margins are partially depressed by the sediment loading in the center of the basin (Haxby et al., 1976; Nunn and Sleep, 1984). Thus, local residual subsidence calculations will overestimate tectonic subsidence on the basin margins and underestimate it in the basin center. Lithology-dependent decompaction methods used in this analysis follow those of Angevine et al. (1990). We note that uncertainty in compaction parameters was found to have little effect on our analysis; sensitivity testing revealed small differences in overall shape of the basin despite a wide range of decompacted subsidence rates (Howell, 1993).

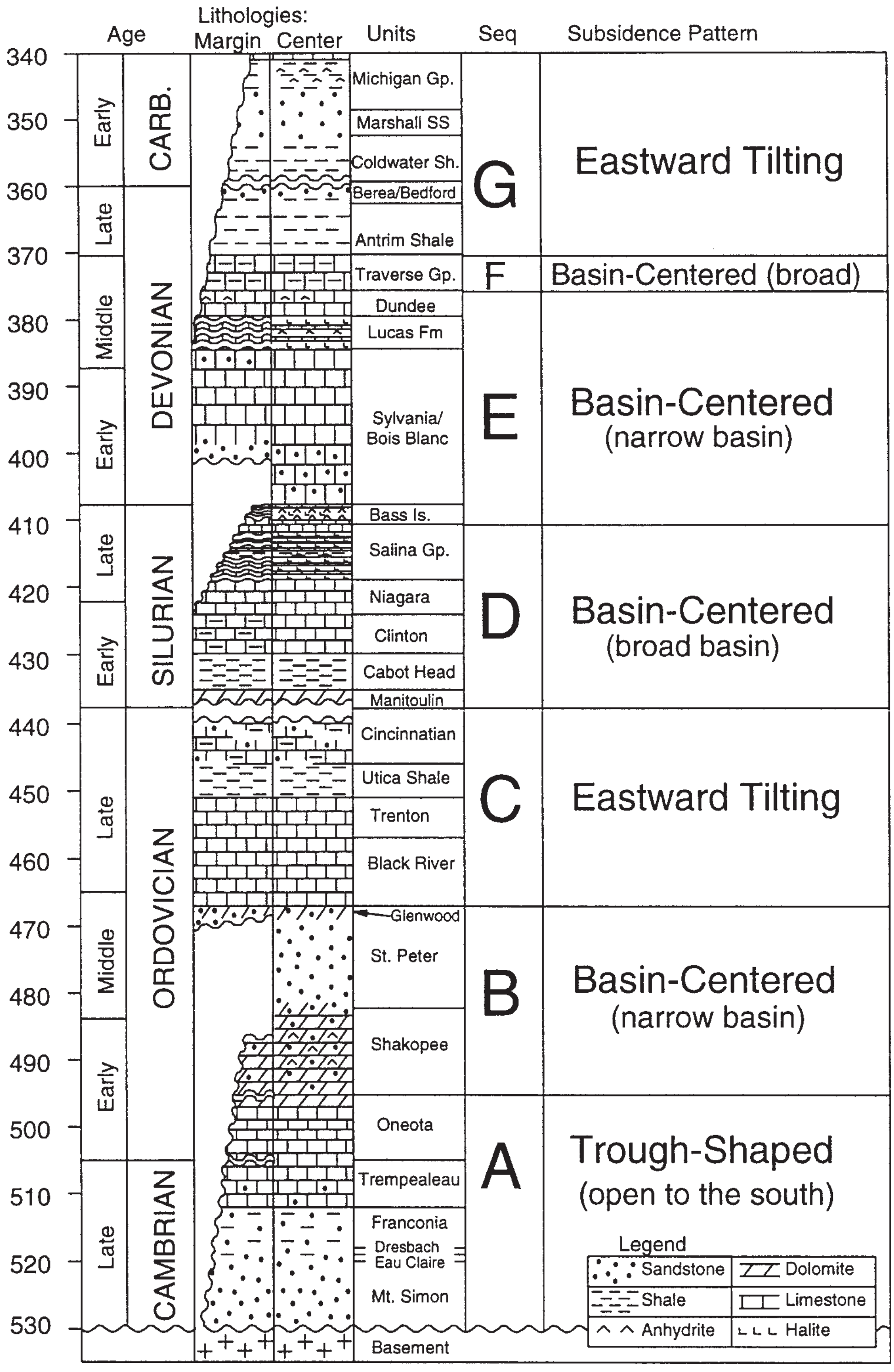

Figure 3. Stratigraphic column for the Michigan basin showing stratigraphic extent of structural sequences, unconformities, and inferred subsidence patterns. Note that Sloss sequence boundaries do not correspond directly with structural sequence boundaries.

STRUCTURAL SEQUENCES IN THE MICHIGAN BASIN

Strata of the Michigan basin are dominated by limestone and dolomite; there are significant siliciclastic and evaporite components. Figure 3 summarizes the stratigraphic succession in the central basin and on the basin margins, and includes the temporal ranges of the structural sequences described herein. The subsidence patterns labeled for each structural sequence are inferred from the shapes of the ΔDB maps and supporting evidence within the sequences.

For each of the six structural sequences we present an isopach map, a ΔDB map, and two cross sections, along with discussion concerning the internal consistency of these sequences. To facilitate comparison of changes in basin subsidence patterns, each ΔDB map is normalized to its maximum thickness (100 * thickness/maximum mapped thickness).

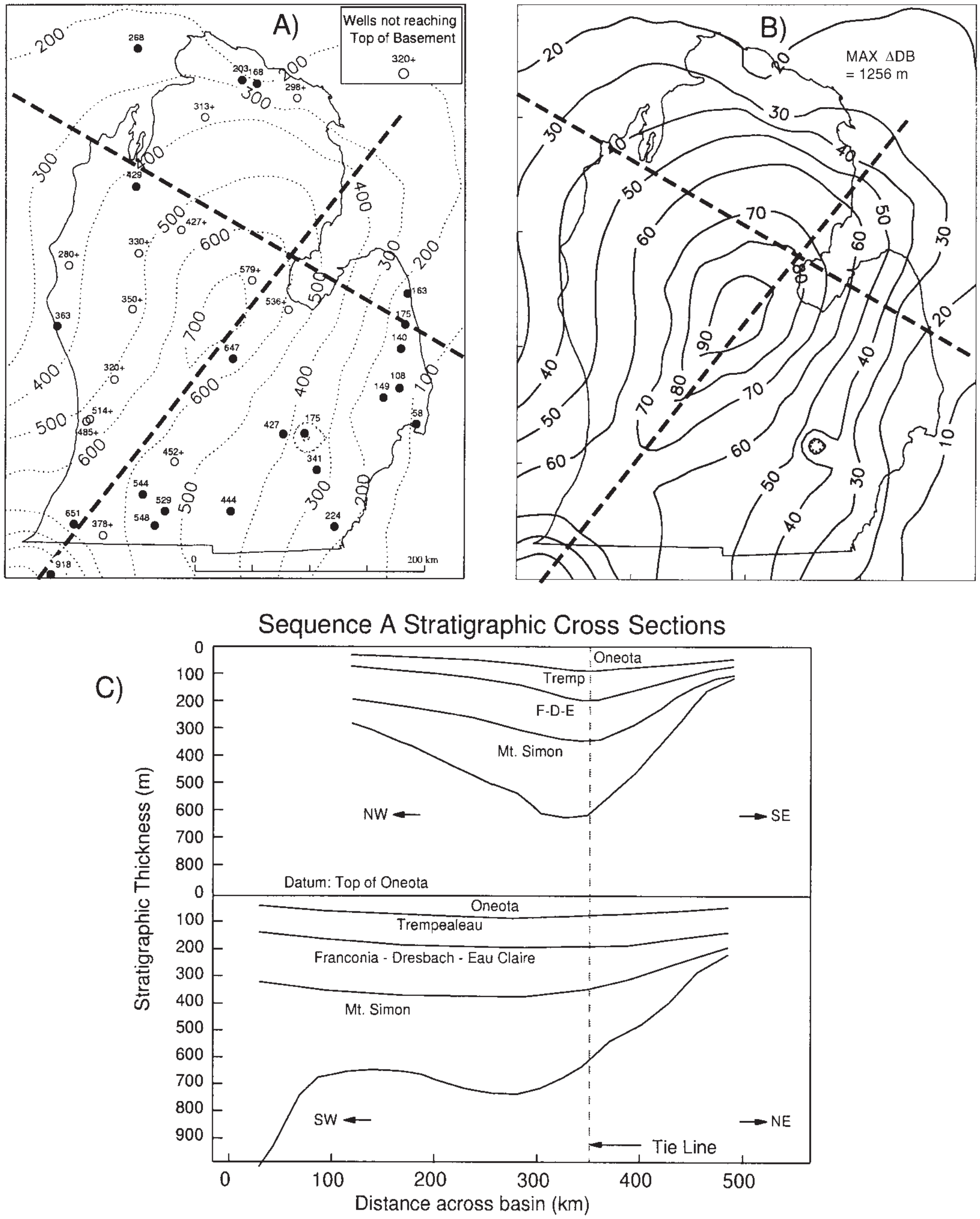

Figure 4. (A) Isopach map of sequence A, Upper Cambrian Mt. Simon Sandstone to Lower Ordovician Oneota Formation. Isopach values are in meters for Figures 4–10. (B) Map of basement subsidence (ΔDB) during deposition of Sequence A. Contours are normalized (100 basis) to facilitate comparison with other ΔDB maps for Figures 4–10. (C) Stratigraphic cross sections for the units composing sequence A, with section locations marked on the isopach and ΔDB maps. Note that closure within the central Michigan area is more pronounced on the ΔDB map than the isopach due to greater decompaction corrections in central Michigan (greater current burial depth).

Sequence A (Cambrian–Lower Ordovician)

The oldest Phanerozoic rocks in the basin were deposited in a trough-shaped depression that was open to the south (sequence A, Upper Cambrian–Lower Ordovician, Fig. 4). These rocks were deposited under relatively shallow water conditions (Catacosinos and Daniels, 1991); maximum water depths were probably tens of meters. The basal Mt. Simon Sandstone is a widespread Upper Cambrian unit that thickens southward into the northern Illinois basin, as does the shale of the overlying Eau Claire Formation. The boundary between the sandstone of the Mt. Simon and shales of the Eau Claire is a facies contact; Catacosinos and Daniels, 1991). This is overlain by thin Dresbach (Galesville) sandstones, which are also in apparent facies contact with both the Eau Claire and the overlying Franconia shales and carbonates. Log correlations suggest that the contact is gradational between Franconia carbonates and the Trempealeau Formation. These gradational contacts and facies changes prohibit the use of these individual formations as reliable indicators of basin subsidence change; boundaries for structural sequences should be as nearly synchronous as possible. The strong, persistent gamma-ray log marker that separates the Lower Ordovician Oneota Formation (Prairie du Chien Group) from the Trempealeau (Lilienthal, 1978) may be an ash bed; however, no core information is available to constrain its lithologic nature.

The southward-thickening, trough-shaped character of the ΔDB map (Fig. 4B) for this interval is strongly influenced by the thick Mt. Simon sandstones (Fig. 4C). The Franconia–Dresbach–Eau Claire interval also thickens to the south, but less dramatically. The Trempealeau and Oneota strata thicken very gradually toward central Michigan and become sandier northward. The younger units also over-

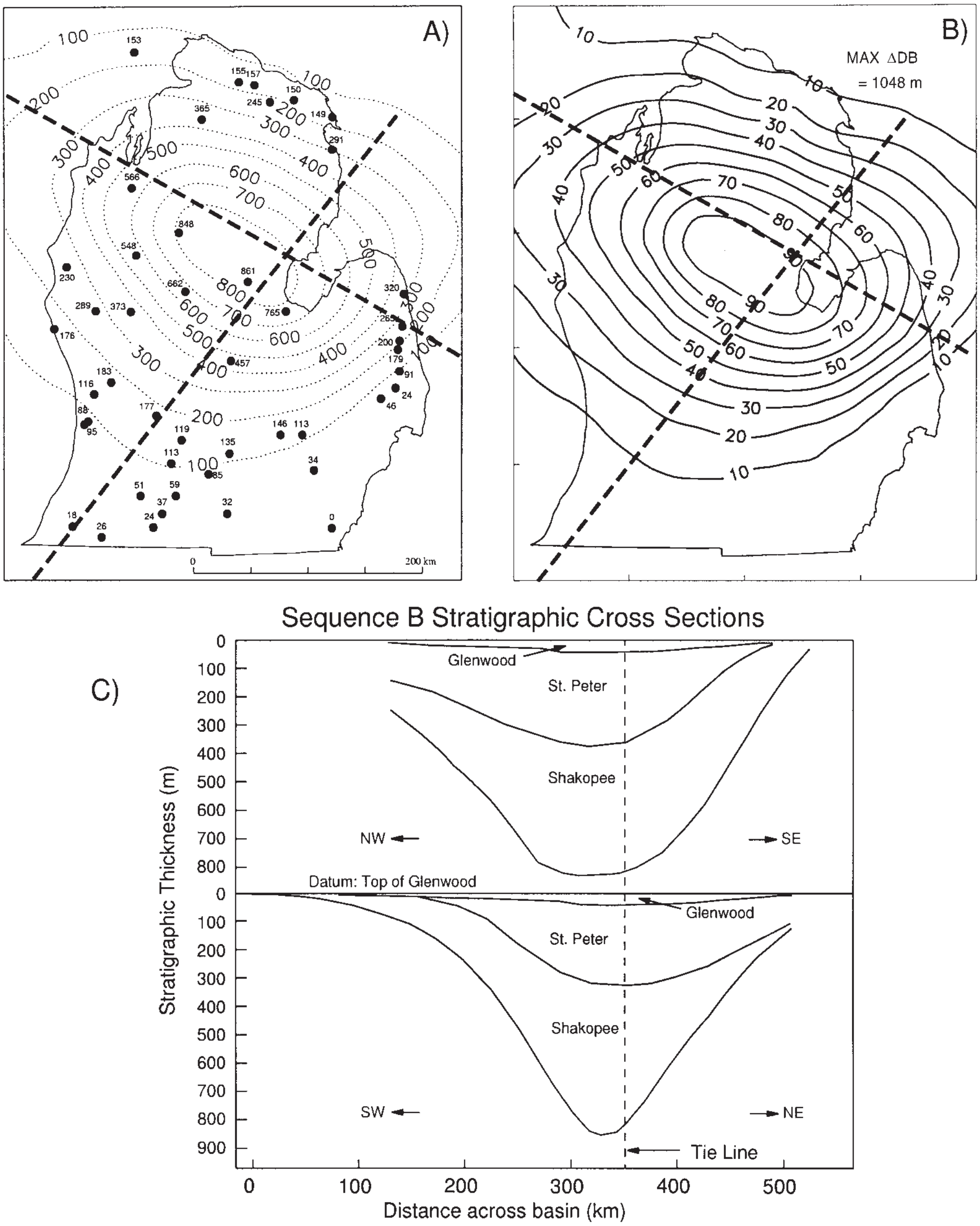

Figure 5. (A) Isopach map of sequence B, Lower Ordovician Shakopee to Middle Ordovician Glenwood. (B) Map of basement subsidence (ΔDB) during deposition of sequence B. (C) Stratigraphic cross sections for the units composing sequence B. Isopach and ΔDB maps both display broad basin subsidence.

step the Mt. Simon to the east on the Findlay arch and to the north, indicating a gradual broadening of the basin. Overall, the basin has the appearance of a trough-shaped depression that is a northward extension of the Illinois basin (Sleep et al., 1980). Although the stratigraphic units composing sequence A do not have a uniform style of trough-shaped thickening, they are considered as a single structural sequence because (1) they appear to show a gradual change in basin shape, (2) they cannot be distinguished as separate structural sequences due to their thinness (relative to uncertainties such as paleobathymetry), and (3) there is a distinctive change in basin shape between the Oneota and the overlying Shakopee Formations.

Sequence B (Lower Middle Ordovician)

Sequence B (Early–Middle Ordovician, Fig. 5) consists of three stratigraphic intervals, and represents the beginning of significant basin-centered subsidence and separation of the Michigan basin from the Illinois basin to the south. The lowest unit is the Shakopee Formation (upper portion of the Prairie du Chien Group), an ~500-m-thick succession of thin-bedded, shaly dolomites and sandstones containing as much as 30% anhydrite as thin beds in the central basin (Foster formation of Fisher and Barratt, 1985). This unit represents shallow-subtidal, intertidal-mudflat, and evaporitic environments (Fisher and Barratt, 1985). The Shakopee Formation is overlain by the St. Peter Sandstone, an ~300-m-thick, quartzose sandstone in Michigan containing thin interbeds of shaly dolomite. The upper portion of the St. Peter formation is highly bioturbated, indicating normal-marine, shallow-water conditions, whereas the lower half of this sandstone has very limited bioturbation, suggesting restricted (possibly hypersaline) environmental conditions similar to those of the underlying Shakopee Formation. Cored intervals from numerous wells in the central part of the basin document the shallow-water nature of this unit, which is a significant reservoir for natural gas (Fisher and Barratt, 1985; Harrison, 1987; Barnes et al., 1992). The St. Peter sandstones grade upward into shaly, burrowed dolomites and sandstones of the Glenwood Formation, a thin (<30 m) unit that represents continued shallow-water deposition.

This thick succession of shallow-water sandstones and carbonates accumulated in a narrow, rapidly subsiding depression (Fig. 5B), marking the onset of significant basin-centered subsidence in Michigan. On the basin margins, however, and throughout eastern North America, significant erosion and dissolution of carbonates occurred during this time, forming the post-Sauk unconformity of Sloss (1963). Examination of cores and well logs in this study and others revealed little physical evidence and no paleontological evidence of this unconformity within the deep central basin (Fisher and Barratt, 1985; Catacosinos and Daniels, 1991; Smith et al., 1993), although some have suggested that it may be present between the Shakopee Formation and the St. Peter Sandstone, particularly along the southeastern flank of the basin (Algonquin arch: Harrison, 1987; Barnes et al., 1992; Nadon and Smith, 1992). Because the surrounding region lacks correlative facies to the thin-bedded Shakopee Formation and the basal, nonburrowed sandstones of the St. Peter Sandstone, Fisher and Barratt (1985) suggested that the Michigan basin continued to subside during early Middle Ordovician time while uplift and erosion occurred on the basin margins. Uplift and erosion are well documented on the Kankakee arch (southern Michigan, northern Indiana, and Illinois; Fig. 1) where pre–St. Peter formation erosion locally eliminated as much as 200 m of Lower Ordovician and Upper Cambrian rocks, regions later filled in with St. Peter Sandstone (Buschbach, 1964; Willman et al., 1975), and to a lesser extent this pattern extends into Wisconsin (Smith et al., 1993). The thickness pattern of the St. Peter formation is well constrained across the northern basin by more than 100 wells, but the more limited data shown in Figure 5A represent the paucity of wells that completely penetrate the sequence B section.

Although it may appear that a simple drop in eustatic sea level, such as that postulated for the end of the Sauk sequence of Sloss (1963), could cause a basin-centered subsidence pattern, this is not the case for structural sequence B. There are two reasons for this: (1) the accumulation rates of the Shakopee and St. Peter intervals are far greater than that of the underlying Cambrian units (Fig. 2), indicating that there was a substantial change in basin-centered subsidence prior to the end of Sauk deposition, and (2) even a substantial (100 m) drop in sea level could, with total erosion of exposed strata, cause only about a 200 m differential in thickness from basin margin to center. On the basis of the evidence cited here, we consider it more plausible that uplift of the basin margin was coeval with increased subsidence and sedimentation in the basin center. The resulting isolation of the basin also has correspondence in the change from open-marine conditions to the more restricted facies of the Shakopee and lower St. Peter Sandstone in Michigan's central basin.

Sequence C (Middle Upper Ordovician)

Basin-centered subsidence ceased in Middle Ordovician time, replaced by eastward tilting toward the Taconic margin of North America (sequence C, Middle to Late Ordovician, Fig. 6), Regional eastward tilting dominated the entire eastern United States during this period (Howell and van der Pluijm, 1990; Coakley et al., 1994), and may have extended northward across much of eastern Canada and the Hudson Bay basin (Cook and Bally, 1975; Quinlan and Beaumont, 1984; Beaumont et al., 1988). In Michigan, the Black River and Trenton Group carbonates thin markedly to the north and west from their depocenters in the southeastern portion of the state. These thickness changes, however, are due in part to a facies change from shallow-water wackestones and packstones in the area of the depocenter to starved, deeper water mudstones and wackestones in the central basin (Fig. 6C; Wilson and Sengupta, 1985; Howell, 1993). Subsequent shales and shaly carbonates of the Utica Formation and Cincinnatian strata (undifferentiated) filled in the moderate paleobathymetry (<60 m water depths) that resulted from these facies changes; thus, the return to a basin-centered isopach pattern for the Utica and Cincinnatian units does not reflect a return to basin-centered subsidence, but rather the infilling of an existing bathymetric low (Howell, 1993).

This tilting episode was not evident in the previous collections of isopach maps (Fisher et al., 1988) primarily because the St. Peter Sandstone, with its decidedly basin-centered isopach pattern, was included in the interval with these later Ordovician carbonates (Tippecanoe I sequence of Fisher et al., 1988). Fisher et al. (1988) and Coakley et al. (1994) suggested that the succession of thickness variations in this sequence is suggestive of basinal subsidence migrating eastward through time (Black River and Trenton carbonates), followed by a return to basin-centered subsidence (Utica Shale). However, correction for compaction and paleobathymetry reveals that there is little or no basin-centered subsidence component during sequence C deposition (Fig. 6B), and the carbonate depocenters are primarily shallow-water platforms, not structural basin centers (Howell, 1993). The Cincinnatian units are capped by a regional erosional unconformity that is well documented across the basin and margins (Taconic disconformity of Wheeler, 1963; Lilienthal, 1978; Howell, 1993).

Sequence D (Lower Upper Silurian)

Silurian units of sequence D record a pattern of broad basin-centered subsidence (Fig. 7). The Manitoulin Dolomite is a thin (<40 m), basal Silurian, shallow-water carbonate unit that fills topographic lows on the Taconic disconformity, and contains local biohermal mounds in the southern part of the state. The Manitoulin Dolomite is overlain by the Cabot Head Shale, a thin (<50 m) marine shale that is found throughout the basin. In the southern part of the state, these units have a reciprocal thickness variation where thin Cabot Head shales overlie locally thickened Manitoulin bioherms. The overlying Clinton Group is represented by a very thin, shaly interval in the southern part of the state, which thickens into a shallow-water, limestone bank as much as 130 m thick on the northern side of the basin, and a thin, deeper water, mudstone and wackestone facies in the central basin (Harrison, 1985). This basin paleobathymetry was enhanced in Middle Silurian time with development of an extensive Niagara Group carbonate bank around the basin perimeter, and a deeper water area in the central basin. Pinnacle reefs were abundant on the slopes of the carbonate bank and help constrain basinal water depths to 200–300 m (Leibold, 1992).

The Niagaran paleobathymetric depression was filled by the end of A-2 carbonate (Salina Group) deposition, as evidenced by the presence of widespread sabkha deposits at the top of this unit (Leibold, 1992). Subsequently, the basin continued to subside and fill with evaporites and thin carbonates of the Salina Group during Late Silurian time. Postdepositional dissolution of evaporites on the southern basin flank produced an artificially thin isopach interval, which leads to a local underestimation of the original ΔDB (Fig. 7B) for this structural sequence. However, the overall pattern of basin-centered subsidence remains evident, with a much broader distribution of subsidence than is present in sequence B (saucer shaped for sequence D versus bowl shaped for sequence B). Thick units within this sequence that have shallow-water indicators near the top and bottom also display the same overall, saucer-shaped subsidence pattern as the entire sequence (e.g., Salina B unit, F-unit; Leibold, 1992).

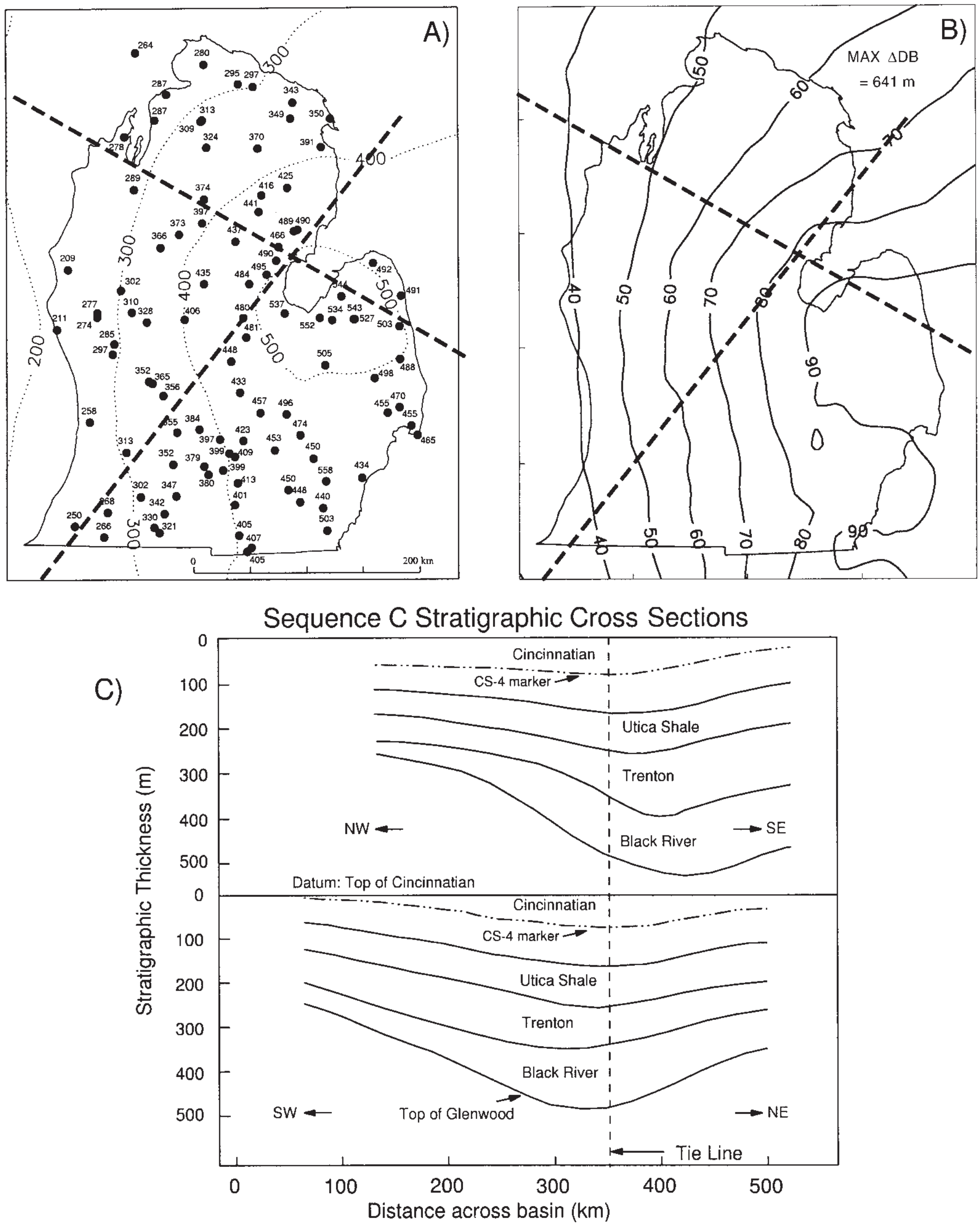

Figure 6. (A) Isopach map of sequence C, Middle Ordovician Black River to Upper Ordovician Cincinnatian. (B) Map of basement subsidence (ΔDB) during deposition of sequence C. (C) Stratigraphic cross sections for the units composing sequence C. Lack of closure in southeastern Michigan on ΔDB map suggests that subsidence due to compaction of underlying units in the central basin is the cause of isopach closure and eastward shift of the depocenter.

Sequence E (Uppermost Silurian–Middle Devonian)

The Bass Island carbonates and evaporites represent a return to narrow (bowl shaped), basin-centered subsidence in latest Silurian time, which continued into Middle Devonian time (sequence E, Fig. 8). In the central basin, these restricted-marine deposits (to 40% anhydrite by thickness with a few thin halite beds) grade upward into nonfossiliferous carbonates and quartzose sandstones of the Bois Blanc and Sylvania Sandstone formations. These two units are distinct on the basin margins, but are interbedded and stratigraphically inseparable in the deep central basin. An extensive unconformity present on the basin margins near the Silurian-Devonian boundary (Fig. 3, post-Tippecanoe unconformity of Sloss, 1963) cannot be recognized on the basis of well correlations in the central basin despite extensive coverage, although some evidence of disconformable lithologic relationships has been reported from cores (Gardner, 1974). This relationship of an extensive margin unconformity with no measurable erosion in the central basin is similar to that of the post-Sauk unconformity within sequence B (Fig. 3).

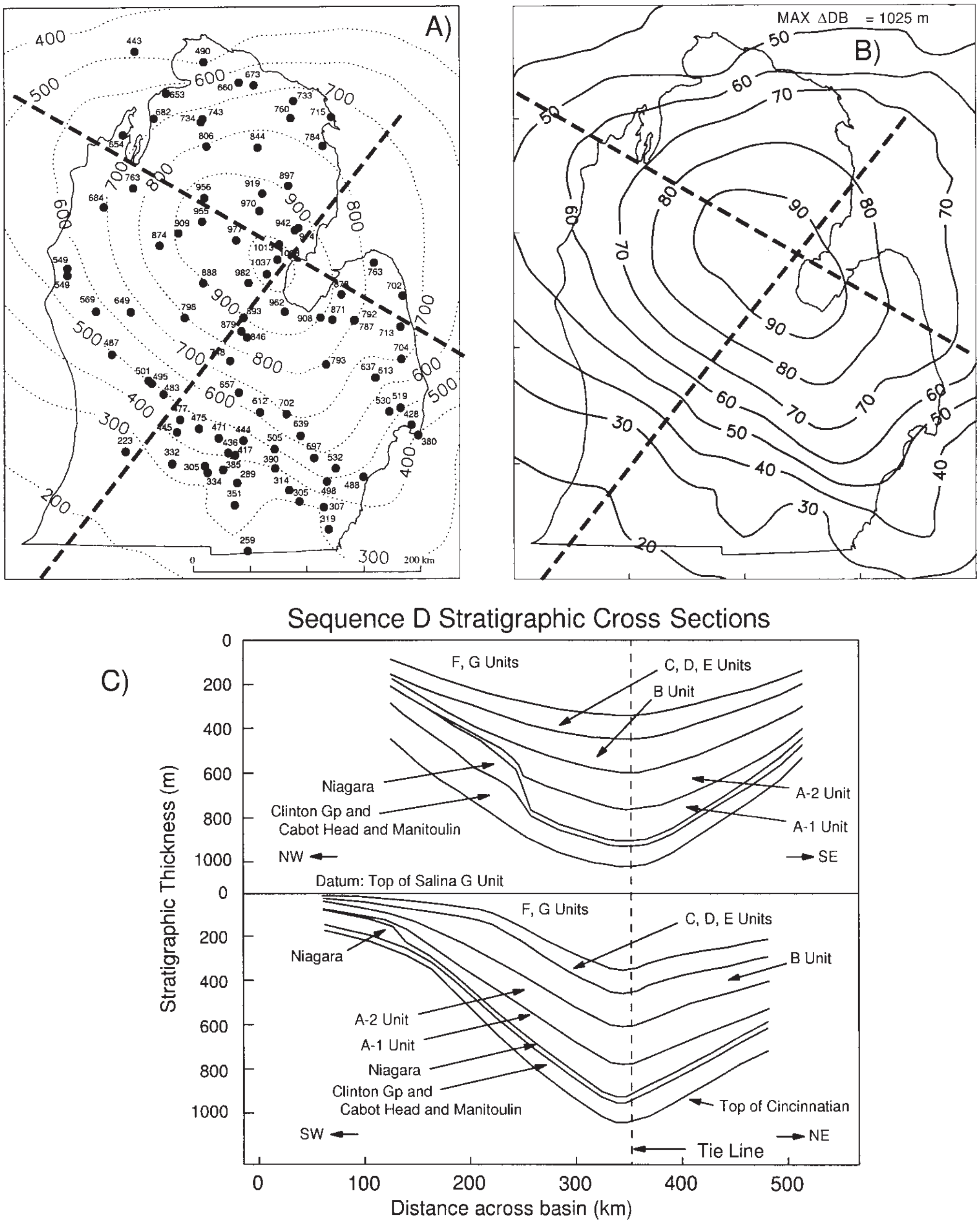

Figure 7. (A) Isopach map of sequence D, Lower Silurian Manitoulin to Upper Silurian Salina G unit. (B) Map of basement subsidence (ΔDB) during deposition of sequence D. (C) Stratigraphic cross sections for the units composing sequence D. Isopach and ΔDB maps both display broad basin subsidence. The tightening of the contours on the southwestern flank of the basin reflects extensive halite dissolution in the Salina A-2, B, and F units. Location of Niagara reef growth is apparent on the cross sections.

The Sylvania–Bois Blanc deposits grade upward into fossiliferous limestones of the Amherstberg Formation (lower Detroit River Group), which represents open-marine conditions with a rich coral and brachiopod assemblage, but no clear paleobathymetric gradients (Gardner, 1974). These carbonates are overlain by the Lucas Formation, a succession of shallow, restricted-marine and intertidal carbonates and evaporites. The sequence is capped by the Dundee Formation, a marine carbonate that has been extensively studied for its role as a major oil reservoir in the basin. The Dundee consists of lagoonal and sabkha carbonates and anhydrite in the western half of the basin, and a facies change to open-marine wackestones and mudstones in the eastern portion of the central basin (Gardner, 1974). Although the paleobathymetry at the top of the Dundee Formation is uncertain, the paleorelief is small relative to the thickness of this structural sequence (~50 m maximum water depth versus 962 m maximum measured thickness), and therefore introduces no significant uncertainty in the ΔDB calculations for this interval. The individual thickness patterns of the Sylvania–Bois Blanc, Lucas, and Dundee intervals (Fig. 8C) are all compatible with the overall narrow basin pattern displayed in Figure 8B.

Sequence F (Uppermost Middle Devonian)

The Middle Devonian Traverse Group carbonates and shales record another episode of basin-centered subsidence, but with a broad distribution (sequence F, Fig. 9). The paleobathymetry of the Bell Shale at the base of this sequence is uncertain, and this unit may belong to the underlying sequence E (narrow basin-centered subsidence pattern); however, because it is a thin unit and is considered part of the Traverse Group, it is treated within sequence F. The Traverse Group carbonates are considered moderately shallow water, marine deposits throughout

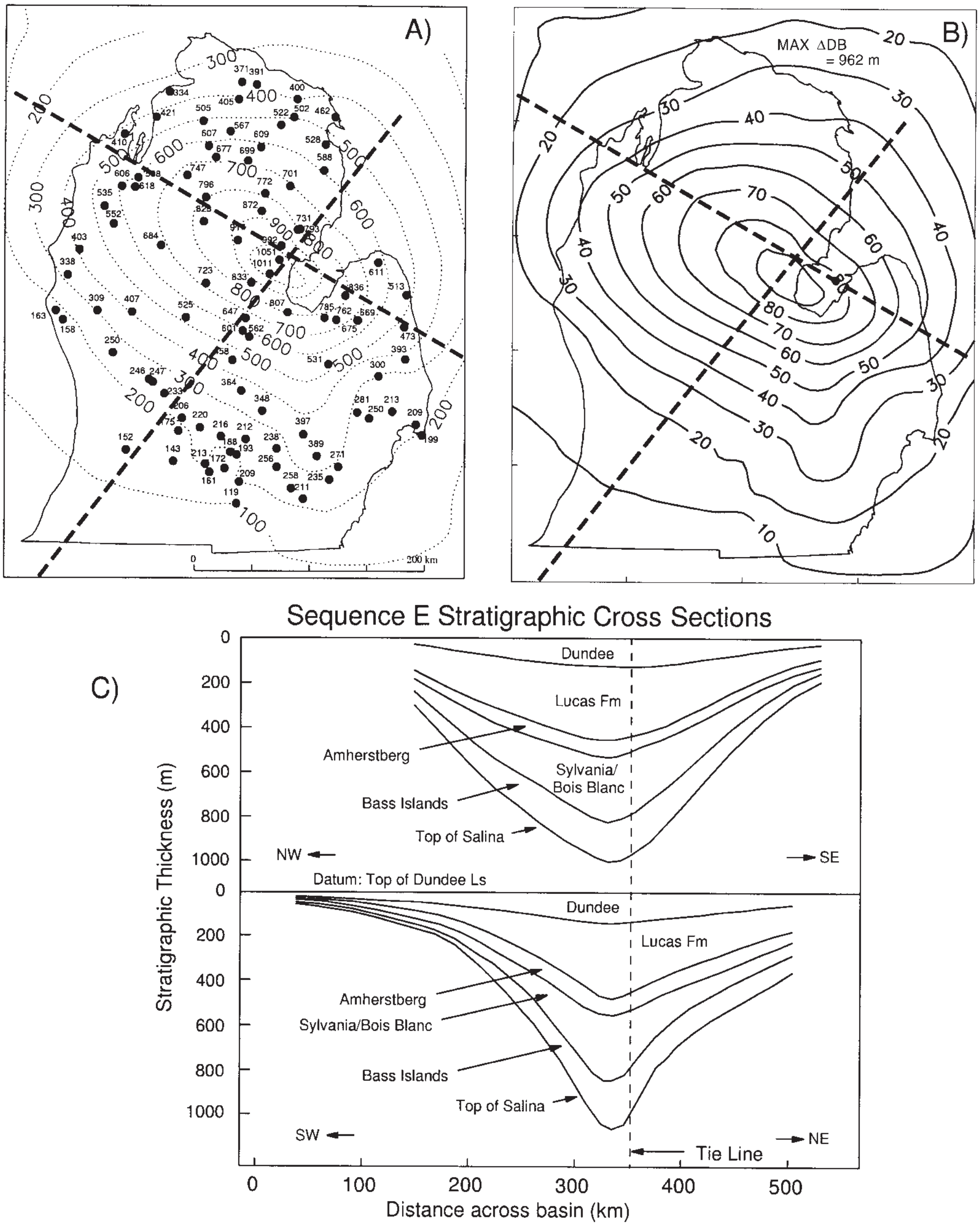

Figure 8. (A) Isopach map of sequence E, Upper Silurian Bass Island to Middle Devonian Dundee. (B) Map of basement subsidence (ΔDB) during deposition of sequence E. (C) Stratigraphic cross sections for the units composing sequence E. Isopach and ΔDB maps both display narrow basin subsidence. Although halite dissolution may be significant in the Lucas evaporites, all units display similarly narrow, basin-centered isopach patterns.

the basin on the basis of their faunal content, the presence of local bioherms, and the very gradual nature of facies changes across the basin (Gardner, 1974; Lilienthal, 1978; Fisher et al., 1988). The shale content of the Traverse Group increases eastward across the basin from ~20% shale in western Michigan to as much as 80% shale in the eastern region (Gardner, 1974). This structural sequence has the broadest (most saucer shaped) distribution of the sequences described here. Although there are few data available to constrain the paleobathymetry of the uppermost Traverse carbonates, even a modest paleobathymetry (e.g., ~50 m deeper in the basin center or eastern portion) would cause little change in the overall basin subsidence pattern. Much deeper basinal paleobathymetry (~150 m) would have been required for the broad pattern described here to result from a narrow subsidence pattern similar to that of the underlying sequence E, but such substantial paleobathymetries are commonly recognized by the presence of basin margin clinoforms (Sarg, 1988), and are not present here.

Upper Devonian–Mississippian Strata (Provisional Sequence G)

Upper Devonian and Mississippian strata are extensively eroded around the basin margin, but the uneroded remnant in central Michigan suggests a final eastward-tilting event (provisional sequence G, Fig. 10). These units do not completely qualify for structural sequence status because the shape of the ΔDB across the entire basin is unclear due to the limited lateral extent of the strata and uncer-

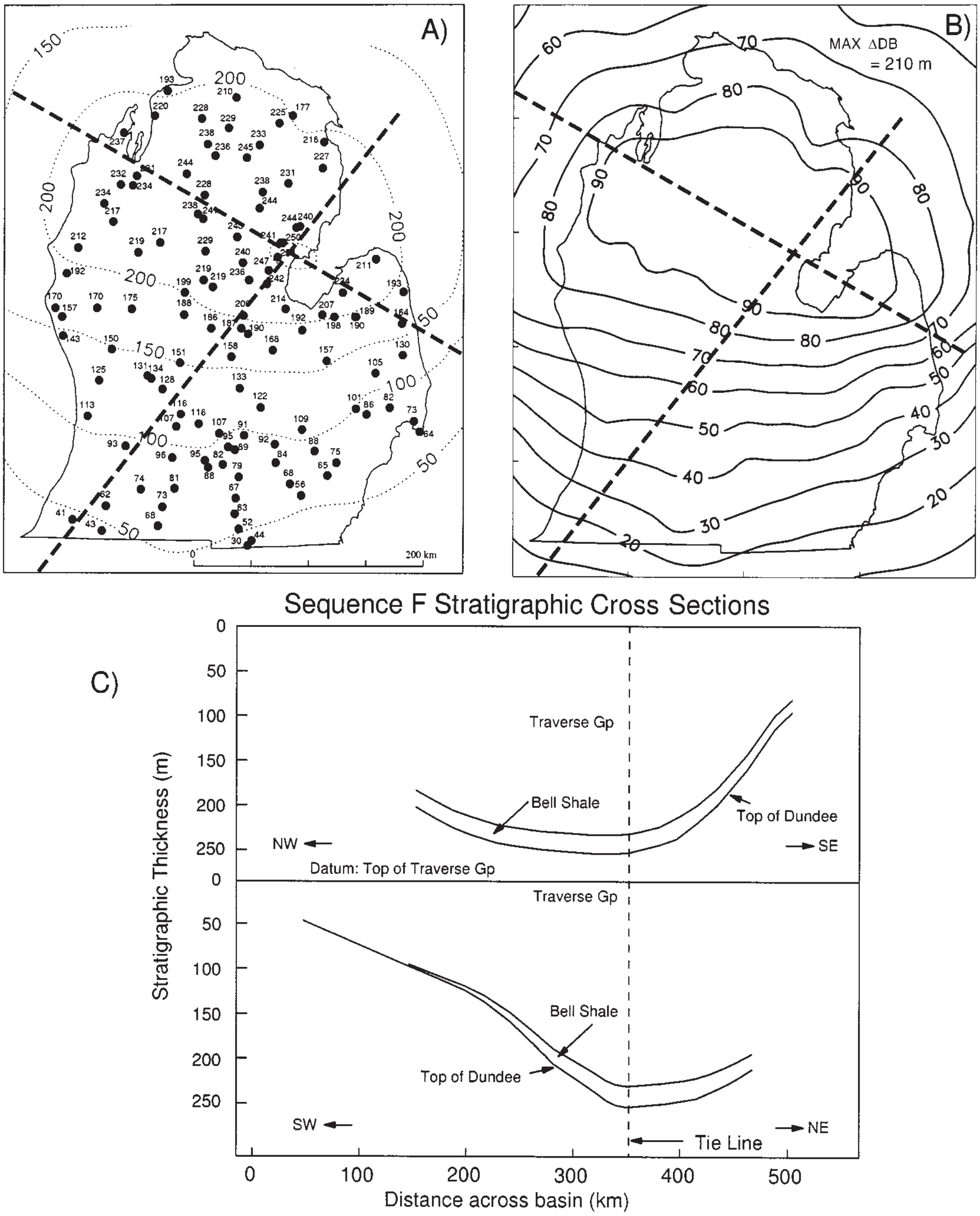

Figure 9. (A) Isopach map of sequence F, Middle Devonian Traverse Group, including the Bell Shale. (B) Map of basement subsidence (ΔDB) during deposition of sequence F. (C) Stratigraphic cross sections for the units composing sequence F. Isopach and ΔDB maps both display broad basin subsidence.

tainty associated with the paleobathymetry of some units.

The basal units of this interval consist of the Antrim and Ellsworth Shales. The Antrim Shale is very organic rich, with a highly radioactive gamma-ray log character, and is restricted to the eastern and central part of the basin. To the west, this unit grades into the Ellsworth Shale, which is less radioactive and thickens westward (Fig. 10C). Both units grade upward into interbedded shales and fine sandstones of the Bedford Shale and Berea Sandstone formations. These coarser detrital sediments represent the distal limit of a westward-prograding delta complex derived from the Alleghanian orogenic welt (Gutschick and Sandberg, 1991). An isopach map of the combined Antrim, Ellsworth, and Berea-Bedford interval displays a distribution that is thickest to the west and thinnest in the south-central portion of Michigan (Fig. 10A). This distribution is interpreted in terms of moderate paleobathymetry at the end of Berea-Bedford deposition (perhaps ~50–100 m deeper in the south-central area) superposed on a pattern of regional subsidence (Gutschick and Sandberg, 1991). The overlying Coldwater Shale and Marshall Sandstone interval thickens eastward (Fig. 10, B and C), suggesting another eastward regional tilting event toward the Alleghanian foreland. Basinwide distribution of the overlying Michigan Group and Pennsylvanian Saginaw Formation are too limited areally to delineate regional subsidence patterns and are not included in the analysis.

MODELS FOR BASIN SUBSIDENCE

There is no single mechanism that fully explains the origin and subsequent evolution of the

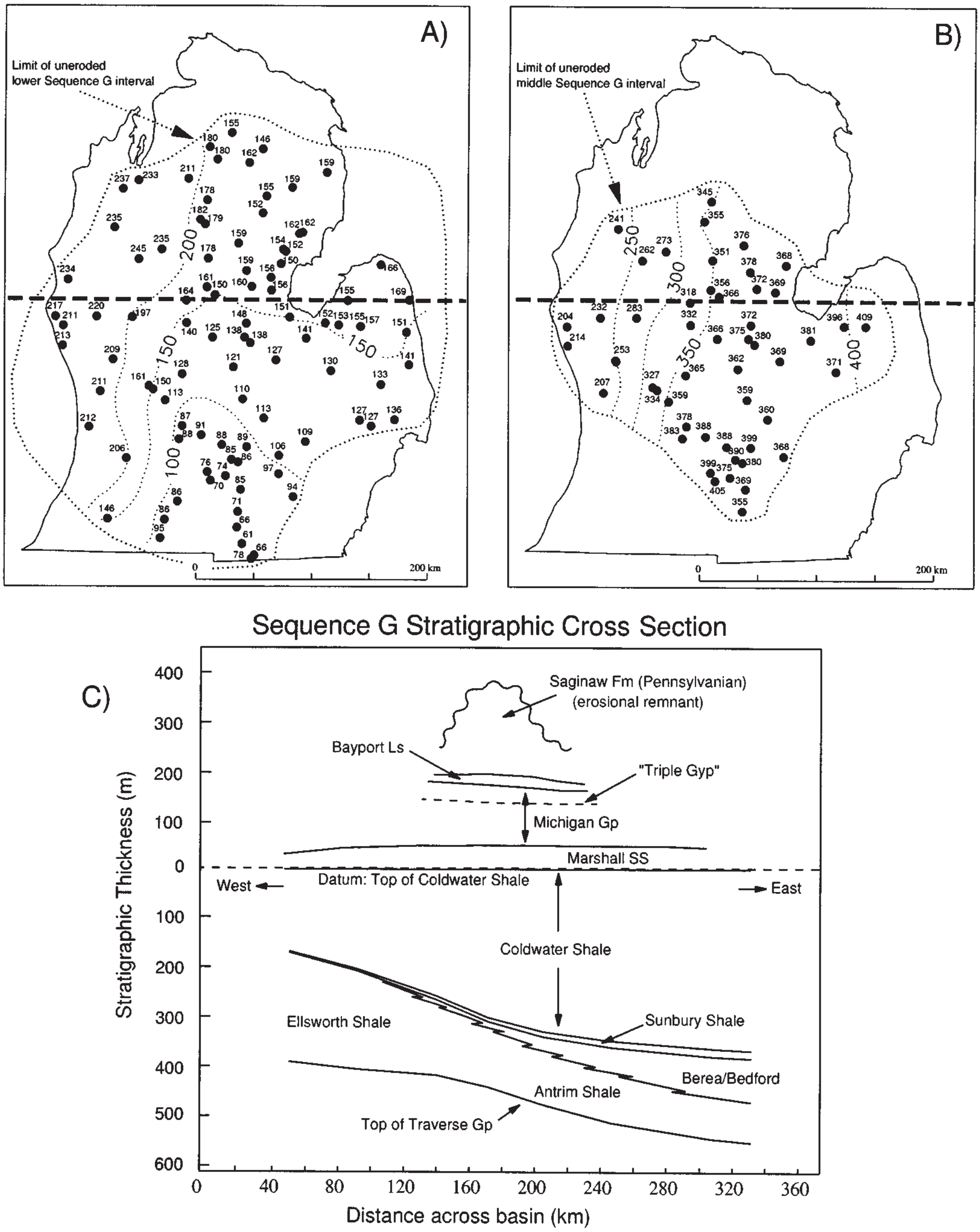

Figure 10. (A) Isopach map of lower portion of sequence G, Upper Devonian Antrim Shale, Ellsworth Shale, Bedford Shale, and Berea Sandstone. Thinning in basin center reflects deeper water conditions. (B) Isopach map of middle portion of sequence G, Marshall Sandstone and Coldwater and Sunbury Shales. Marshall is a coarser facies equivalent to a portion of the Coldwater, and does not include sandstones of the "Michigan stray." Overall eastward tilting pattern is evidence. (C) East-west stratigraphic cross section for the units composing sequence G, with section locations marked on the isopach maps (dashed lines).

Michigan basin. Rather, changing subsidence styles of the basin require several mechanisms to be active through the course of Paleozoic time, as documented by profiles of normalized ΔDB for each sequence (Fig. 11). The profiles illustrate distinct changes in subsidence style between sequences and highlight the significant difference between narrow and broad basin-centered patterns. Normalization to the maximum thickness along each profile facilitates comparison of profile shape and subsidence style.

The following is a summary of the subsidence patterns of the Michigan basin. Sequence G: Upper Devonian to Mississippian, eastward tilting(?); sequence F: Middle Devonian, basin centered, broad; sequence E: Lower Devonian, basin centered, narrow; sequence D: Silurian, basin centered, broad; sequence C: Upper Ordovician, eastward tilting; sequence B: Lower Ordovician, basin centered, narrow; sequence A: Cambrian, trough shaped.

The following sequence of mechanisms is suggested as a scenario that explains the major features of the basin.

A. Rift Development

Subsidence of the Michigan basin began in Late Cambrian time with deposition of the Mt. Simon Sandstone in a trough-shaped pattern, with a deeper depocenter to the southwest in northern Illinois (Buschbach, 1964). This pattern suggests that the eastern midcontinent region underwent a

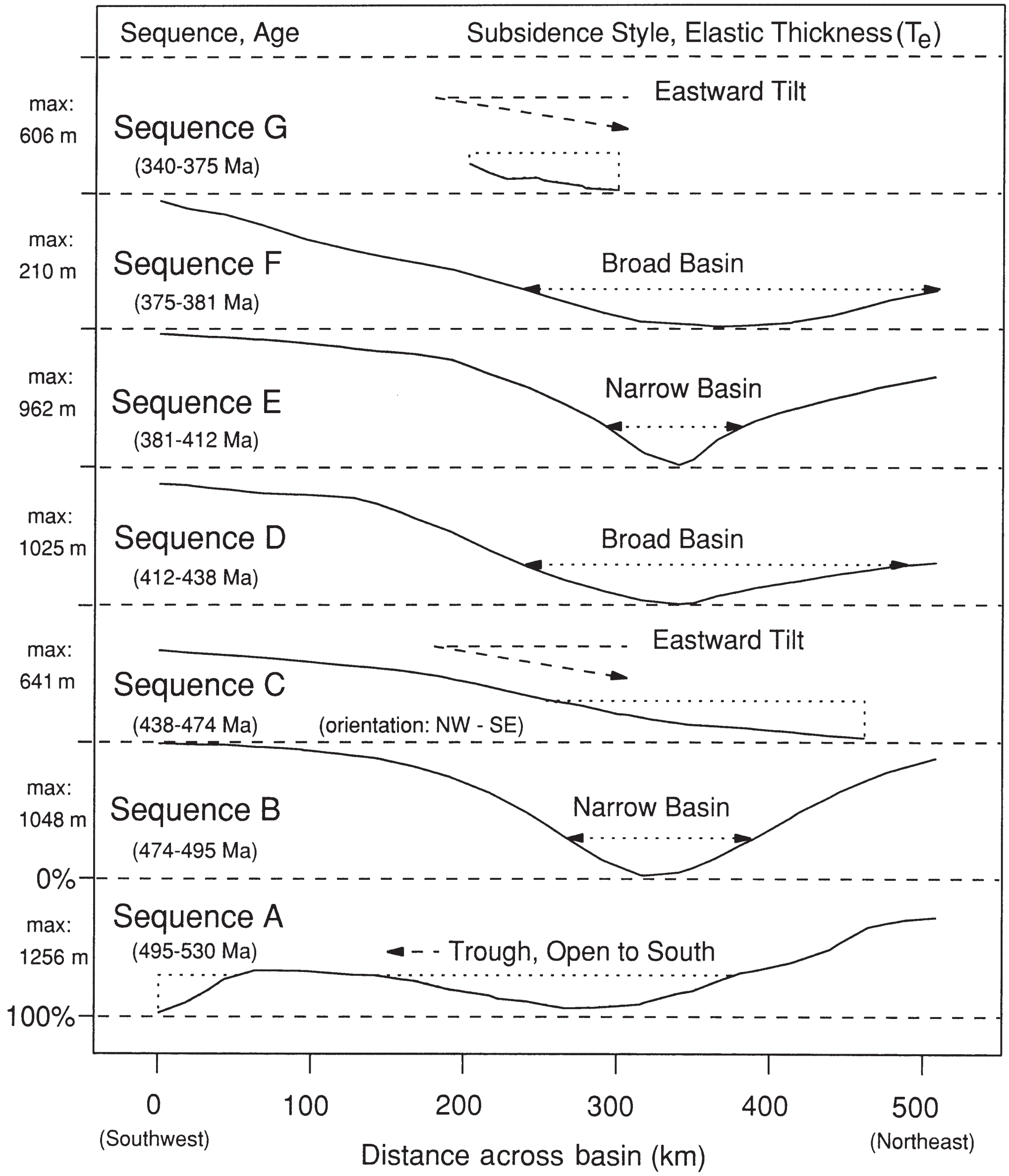

Figure 11. Summary profiles of ΔDB for sequences A–G (orientations northeast-southwest except as indicated), with style of subsidence labeled. Portions of profiles within dotted lines emphasize areas of basin with normalized ΔDB > 70%.

change in locus of extension from the Middle Cambrian, east-west–oriented Rome trough arm of the Reelfoot rift in Kentucky (McGuire and Howell, 1963; Cable and Beardsley, 1984) to the more north-south orientation of the Illinois-Michigan extension (Sleep et al., 1980). This Late Cambrian event perhaps represents a final Cambrian extensional phase of the Reelfoot aulacogen, an interior counterpart to the divergent margin that developed on the eastern and southern margins (present coordinates) of the Laurentian continent (Rankin, 1976). Later units within sequence A, particularly the Trempealeau and Oneota Formations, display slight and gradual thickening toward the basin center and overstep the underlying units toward the basin margins. This pattern is compatible with thermal contraction following the moderate lithospheric extension suggested for the area at that time. This structural sequence thus contains a gradual change in the basin subsidence pattern, one that is consistent with an origin by lithospheric stretching.

The rifting discussed here is distinct in time and spatial pattern from the much earlier Keweenawan rift. The Keweenawan rift (ca. 1100 Ma, Fig. 1) is well defined by gravity and seismic data and extends northwest-southeast across the central basin (Catacosinos and Daniels, 1991), effectively orthogonal to the northeast-southwest–trending trough-shaped pattern of subsidence established for sequence A. The long wavelength and amplitude of the large, regional gravity anomaly shown in Figure 1 suggest that it is located deeper in the crust than the narrow, high-amplitude gravity anomaly of the Keweenawan rift (Haxby et al., 1976). Sleep et al. (1980) suggested a depth of origin in the middle crust for this anomaly, and we join these authors as we fail to conjecture as to the origin of this feature.

B. Initiation of Narrow Basin-Centered Subsidence

Basin-centered subsidence began with deposition of the Shakopee Formation (Lower to Middle Ordovician). Although the Shakopee-Oneota contact is unconformable in Wisconsin (Smith et al., 1993), there is no evidence for uplift and erosion of the underlying Oneota Formation within the central Michigan basin, as would be expected during lithospheric heating for thermal contraction and metamorphic phase change mechanisms proposed for the basin.

Lithospheric stretching (McKenzie, 1978;

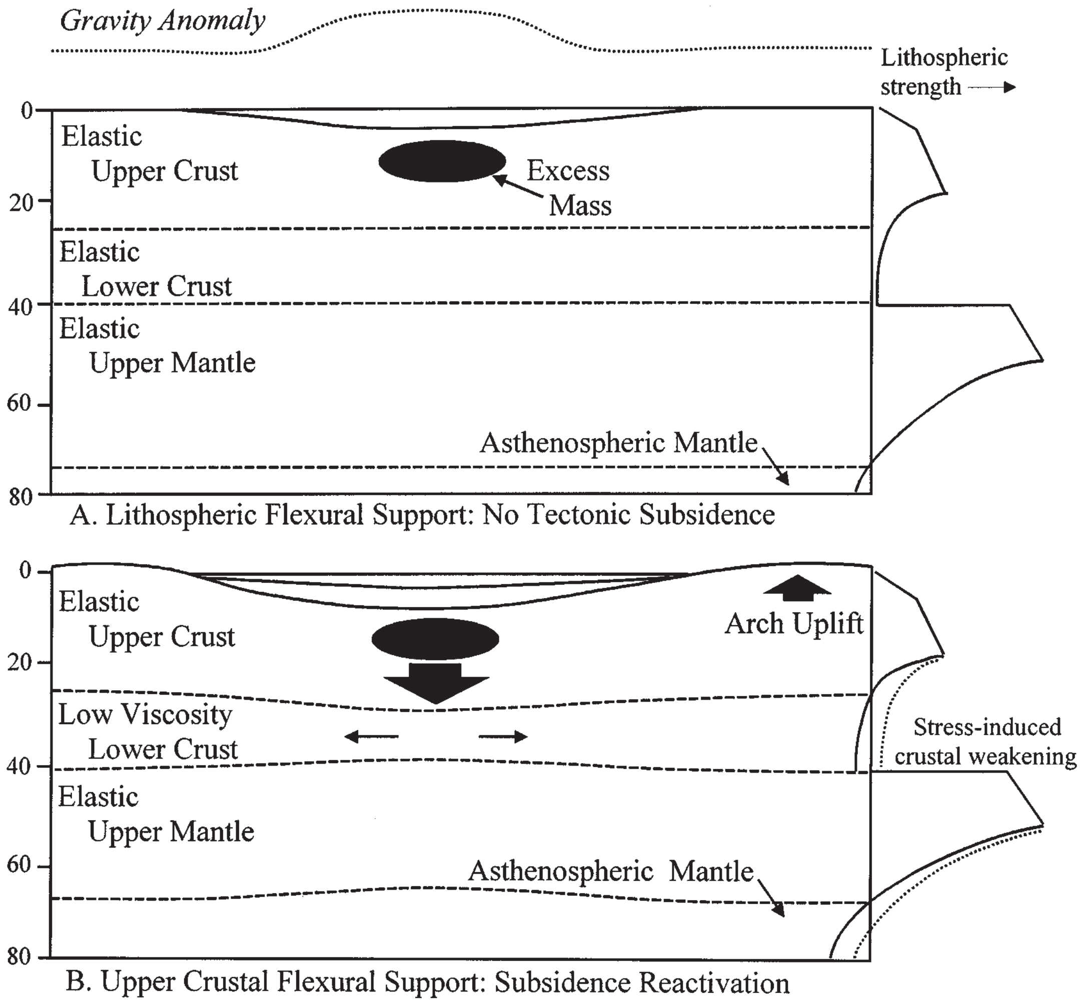

Figure 12. Model for subsidence caused by stress-induced crustal weakening. Each panel shows the location of an excess mass in the upper crust and behavior of different lithospheric layers; on the right is a stylized lithospheric strength profile (after Ellis, 1988). (A) Under low stress conditions, the excess mass is flexurally supported by the entire lithospheric thickness. (B) High stress levels cause a weakening of the lithospheric portions where strength is controlled by crystal plasticity. Here the lower crust becomes sufficiently viscous to flow and the lithosphere-asthenosphere boundary is elevated. This allows the excess mass to be supported by the upper crust alone, resulting in narrow (low rigidity), basin-centered subsidence accompanied by lower crustal flow toward the basin margins, causing uplift of the surrounding arches. The total amount of flexure shown in this figure is exaggerated for clarity; total sediment accumulation during the two episodes of narrow, basin-centered subsidence in the Michigan basin is <2 km.

Klein and Hsui, 1987) is ruled out by the absence of syndepositional faulting, whereas lithospheric flexure amplification due to in-plane stress (Cloetingh et al., 1985; Karner, 1986) and models of stress-induced basin reactivation (DeRito et al., 1983; Lambeck, 1983) also predict reactivation of the Illinois basin (there was no such event at this time; Heidlauf et al., 1986). A stress-induced crustal weakening mechanism introduced by Howell and van der Pluijm (1990) is compatible with the geologic evidence. In this model, a load in the middle to upper crust is supported flexurally by a thick, lithospheric plate (Fig. 12A). Basin-centered subsidence is initiated when high intraplate stresses weaken the crust enough to allow viscous flow in the lower crust (Fig. 12B).

Numerous studies of crustal rheology support this model of lower crustal weakness (Molnar, 1988) and quantitative models have shown the impact of intraplate compressive stresses on viscoelastic stress relaxation in the lower crust (Kusznir, 1982; Kusznir and Park, 1984). When the lower crust is thus weakened, the upper crust effectively becomes a thin plate, leading to flexural subsidence of the upper crust beneath the anomalous load. Lower crustal material flows away from the central basin subsidence toward the basin margins, resulting in uplift of the arches. The net result is that the crust subsides due to an effective attenuation of the lower crust. Isostatic equilibrium is maintained throughout this process, so there is no rebound effect when the stress is removed. Tectonic events capable of providing the horizontal stresses to drive subsidence may be found in a temporal correlation between early tectonic events in the Taconic orogeny and initiation of sequence B subsidence (Howell and van der Pluijm, 1990).

C. Dynamic Tilting of Eastern North America

Basin-centered subsidence abruptly ceased in the latter part of Middle Ordovician time. There is no clear evidence for a basin-centered subsidence component in any of the units composing sequence C. Rather, the Michigan basin along with most of eastern North America tilted eastward toward the developing Taconic orogenic welt. Because the wavelength of this tilting is so long (~1000 km) and includes preexisting basins and arches inland as far as Wisconsin and Iowa (Cook and Bally, 1975), flexural downwarping from marginal, supracrustal loading (Quinlan and Beaumont, 1984; Beaumont et al., 1988) is rejected as a potential mechanism, although it certainly played a role in subsidence of the more proximal portions of the foreland (Diecchio, 1993).

Gurnis (1992a) proposed that dynamic topography associated with subduction beneath Laurentia was responsible for uplift and subsequent tilting of the Appalachian region during Late Ordovician time. Howell (1993) and Coakley and Gurnis (1995) further explored this interpretation for the Ordovician history of the Michigan basin. In the Gurnis model, uplift of a viscous (nonflexural) forebulge associated with subduction accounts for the erosional post-Sauk unconformity in the eastern Appalachian region. As dip of the westward-subducting slab dynamically shallowed, tilting extended farther into the midcontinent. This tilting lasted until Late Ordovician collision of the Taconic arc with the Appalachian margin (Stanley and Ratcliffe, 1985), culminating in uplift and erosion across the foreland region. Polarity of the subduction under Laurentia as required in the Gurnis (1992a) model is opposite to that proposed by some workers (e.g., Jacobi, 1981; Hatcher, 1989; Bradley, 1989), but paleomagnetic and paleogeographic arguments support subduction under the Laurentian margin (van der Pluijm et al., 1990, 1995). Gurnis (1992a) initially proposed a 200–500 km lateral extent for this tilting event based on regional isopach maps of Cook and Bally (1975). Documentation in this study and Coakley and Gurnis (1995) indicates that tilting reached westward ~1000 km from the Ordovician margin of Laurentia. The stratigraphy within sequence C may provide strong constraints for future modeling efforts.

Sequence C also demonstrates that the basin-centered subsidence in Early to Middle Ordovician time (sequence B) effectively ceased for a period of ~30 m.y., a result incompatible with thermal contraction as a mechanism for sequence B. Thermal contraction of continental lithosphere may last from 100 to 300 m.y., depending on lithospheric thickness (Ahern and Ditmars, 1985). Cessation of thermal contraction can only arise from a second heating event, timed to add just enough heat to the lithosphere to replace that being lost by conduction at the surface. This situation is unlikely by the lack of geologic evidence for either a first or a second heating event in the basin (Sleep and Sloss, 1978).

D. Broad Basin-Centered Subsidence

Basin-centered subsidence resumed in the Silurian (sequence D), but with a broad (saucer shaped) distribution. In Howell and van der Pluijm (1990), these styles of basin-centered subsidence were not distinguished and both sequences B and D were considered to have the same probable origin. However, quantification of the flexural behavior of these sequences (Howell, 1993), combined with recognition of two subsequent structural sequences of basin-centered subsidence, indicates that these represent different styles of subsidence with different origins.

Thermal contraction associated with the lower crustal thinning suggested for narrow basin-centered subsidence is a possible mechanism to explain this pattern of broad basin-centered subsidence. Thermal contraction subsidence due to uniform lithospheric attenuation directly follows an instantaneous stretching event (McKenzie, 1978). In the model presented here, however, attenuation is limited to the lower crustal region; for instantaneous attenuation in the lower crust only, mantle geotherms would be elevated an amount approximately equal to the crustal thinning minus the associated sediment loading, and the geothermal gradient across the thinned lower crust would be significantly increased. Simple calculations suggest that a thermal anomaly at depth causes a delay of tens of millions of years before peak heat flow (and thermal contraction subsidence) occurs: $t = l^2/\kappa = (3 \cdot 10^4)^2/(1 \cdot 10^{-6}) = 30$ m.y.; where t is the time delay, l is the length scale of interest (depth to the zone of lower crustal attenuation, in meters), and κ is the thermal diffusivity of crustal rocks (M^2s^{-1}; Turcotte and Schubert, 1982). This delay in thermal contraction approximates the duration of sequence C, where no significant component of basin-centered subsidence is recorded.

E. Renewed Narrow Basin-Centered Subsidence

A resurgence of narrow basin-centered subsidence from latest Silurian to Middle Devonian time (sequence E) suggests a renewal of elevated intraplate stress and subsidence due to lower crustal weakening. A possible tectonic event triggering increased lithospheric stress is the final accretion of the Avalonian microcontinent to Laurentia, which correlates with a change in the direction of Laurentian plate motion from southward to northward (van der Pluijm et al., 1990). Sequence E subsidence was accompanied by regional uplift and erosion on the arches flanking the basin in a fashion analogous to the unconformity architecture of sequence B, and there is no corresponding subsidence event in the Illinois basin (Heidlauf et al., 1986).

F. Renewed Broad Basin-Centered Subsidence

The style of subsidence changed in later Middle Devonian time (Traverse Group, sequence F) to broad basin-centered subsidence for ~4 m.y. Analogous to sequence D, narrow subsidence in sequence E should generate a similar component of broad subsidence in sequence F. The ~200 m thickness of sequence F, however, contrasts with the ~900 m of sedimentation due to thermal contraction that is suggested for this model. There is also a problem accounting for this subsidence pattern with such a model—why is there no lag time in the onset of thermal contraction, such as that proposed for sequence C? At this time we simply recognize the problem, but not its resolution.

G. Final Laurentia-Gondwana Collision

Late Paleozoic sedimentation in the Michigan basin and surrounding regions was strongly affected by the orogenic activity associated with final suturing of Laurentia and Gondwana. Clastic sediments dispersed cratonward across the foreland, far from their erosional origins. In Michigan, we find an unusual, irregular isopach pattern for the Late Devonian units (Fig. 10A), suggestive of an anomalous tilting to the west. However, this pattern of sediment accumulation may not truly reflect the shape of the basin due to the significant uncertainty associated with the paleobathymetry of the Devonian black shale units.

Bond and Kominz (1991) surveyed the residual subsidence of this interval for basins in western, southern and eastern North America and found a pattern of excess subsidence throughout the region, as compared to a benchmark section in Iowa that was assumed to be tectonically stable. They suggested that this unusual pattern of regional subsidence developed as a result of North America moving into a zone of dynamic downwelling and compression as the continents coalesced into a supercontinent configuration. The significant westward thickening shown here for these units strongly suggests that the subsidence at the time was neither a simple basin-centered nor an eastward-tilting model.

Devonian subsidence was succeeded by a final episode of eastward, continental tilting in the Mississippian, reflecting final collision between Laurentia and Gondwana and the formation of Pangea (upper sequence G; Fig. 10, B and C). Although Alleghanian supracrustal loading extended significantly farther westward than earlier Taconian thrusting, it is not clear that Mississippian tilting in Michigan is a flexural response to this load. Erosion of the basin margins, however, prohibits full examination of this problem because of the limited stratigraphic data preserved.

DISCUSSION

This study of Michigan basin stratigraphy presents further support for the geodynamic model presented by Howell and van der Pluijm (1990) for basin-centered subsidence. A significant question concerning this model is how to test it further. We suspect that seismic methods in search of a thinned and deformed lower crust would be inconclusive; the total amount of deformation is small relative to the thickness of the crust (<1 km vs. ~40 km; Zhu and Brown, 1986) and imaging the lower crust in any detail with reflection seismic tools is still a difficult task.

More productive at this time will be efforts to examine conceptual predictions about surface deformation made by this subsidence model, and to compare them with the stratigraphic record. The first prediction is that arches surrounding the basin should be uplifted during narrow, basin-centered subsidence due to outward flow of lower crustal material. This effect is seen along the Kankakee and Algonquin arches during Ordovician time (sequence B subsidence) and again in this area during Late Silurian to Early Devonian time (sequence E subsidence). Second, the shape of the basin during narrow, basin-centered subsidence should reflect flexure of a thin, crustal plate instead of a thick, lithospheric plate. Preliminary results of flexural modeling suggest effective plate thicknesses for sequences B and E of ~30 km, compared with ~70 km for sequence D and ~80 km for sequence F episodes of broad, basin-centered subsidence (Howell, 1993). These values suggest flexure of the entire lithosphere during broad, basin-centered subsidence, and flexural decoupling with only the upper crust (~30 km) supporting the surficial loading during narrow, basin-centered subsidence.

A third prediction of this model concerns the nature of surficial deformation that might occur across eastern Laurentia during an episode of stress-induced crustal weakening. If the lower crust becomes weakened over this broad region, then other areas with significant upper crustal loads may subside, and areas with extra light loads in the upper crust may undergo uplift and erosion. For example, eastern Pennsylvania has a significantly thickened lower Middle Ordovician carbonate succession and is apparently lacking a post-Sauk unconformity (Mussman and Read, 1986), and there is a significant Bouguer gravity anomaly in that area that does not match the typical foreland basin paired anomaly pattern (Karner and Watts, 1983). Thus eastern Pennsylvania, like the Michigan basin, may have undergone subsidence rather than exposure at the end of the Sauk sequence due to stress-induced crustal weakening. In the broader context of the post-Sauk unconformity, most of eastern Laurentia had accumulated thick deposits of Cambrian and Lower Ordovician sediments at the time of this event. We suggest that these thick sediments would produce a negative load (net buoyancy) in the upper crust over a broad region when the lower crust became weakened, and this would lead to modest uplift and erosion in synchrony with the subsidence of sequence B in the Michigan basin. Thus, erosion of the post-Sauk unconformity over the broad interior of eastern Laurentia might be related more to in-plane stresses associated with the initiation of the Taconic orogeny than to a migrating peripheral bulge (Jacobi, 1981; Mussman and Read, 1986). The extent of the post-Sauk unconformity goes well inboard on the craton, beyond the reach of a migrating peripheral bulge with any realistic lithospheric rigidities, but in-plane stresses may transfer ~2000 km into a plate interior (Craddock et al., 1993; van der Pluijm et al., 1997). Similarly, the narrow, basin-centered subsidence of sequence E corresponds in time to erosion of the widespread post-Tippecanoe sequence boundary of Sloss (1963). In-plane stress models may provide a new framework with which to examine the origin of the widespread Sloss sequences and bounding unconformities (cf. dynamic topography model of Burgess et al, 1997).

We reiterate here and extend our previous suggestion (Howell and van der Pluijm, 1990) that the initiation of Taconic orogenesis correlates temporally with initiation of sequence B. High levels of intraplate stress seem to be required to produce effects such as we describe here, levels that must be only rarely achieved by tectonic processes. Stresses associated with simple subduction of oceanic lithosphere beneath a continent seem insufficient to generate the widespread effects we document here (cf. Cloetingh and Wortel, 1986, Coblentz et al., 1995). During the main phases of Middle to Late Ordovician foreland basin formation and Taconic deformation in the northeastern United States, Laurentia is tilting eastward with no basin-centered subsidence in Michigan. Rather, we suggest that episodes of significant collision or plate reorganization (such as would likely occur at the very beginnings of significant orogenies) lead to tectonic conditions that impart high levels of in-plane stress to cratonic lithosphere.

We have demonstrated the utility of a structural sequence approach to examining basin subsidence history. By utilizing structural sequence criteria that carefully define basin deformation patterns and further demand consideration of the effects of paleobathymetry, compaction, and dissolution, the use of structural sequences provides a quality check on basin subsidence data sets that will aid the basin modeling process. Interpretation problems of several previous studies of the Michigan basin are largely resolved when the stratigraphy is examined in light of structural sequences, rather than using a predetermined stratigraphic grouping (Fisher et al., 1988) or if paleobathymetry had been explicitly considered (Coakley et al., 1994). Quinlan and Beaumont (1984) and Beaumont et al. (1988) produced a synthetic model of supracrustal loading on a viscoelastic plate that successfully reproduced the first-order stratigraphy of eastern North America. However, the modeling of Beaumont et al. (1988) achieved significant regional subsidence as far inland as eastern Wisconsin during Middle Ordovician time only through specifying an additional component of basin-centered tectonic subsidence in Michigan (the subsidence of the Michigan and Illinois basins was not explained geodynamically, but rather was simply added to the model based on empirical sediment thickness data). In large part, the authors specified this additional subsidence in their model in order to fit the isopach pattern of the entire lower Tippecanoe sequence of Sloss (1963), which includes the St. Peter Sandstone (sequence B, basin centered) with the Black River and Trenton Formations (sequence C, tilting). The subsidence due to supracrustal loading alone (without the Illinois or Michigan basins added) would have been insufficient to explain the observed stratigraphic thickness of the Black River–Trenton succession in the cratonward portion of their study area. High-resolution modeling points out this geodynamic dilemma, which has subsequently been used as evidence for subduction-related dynamic topography (Gurnis, 1992a; Howell, 1993; Coakley and Gurnis, 1995).

CONCLUSIONS

Analysis of structural sequences significantly aids interpretation of basin geometry through time, and has applicability in other cratonic areas. The Michigan basin underwent several distinctive changes in basin subsidence patterns, reflecting changes in the controlling subsidence mechanisms. These changes in basin subsidence, combined with the architecture of unconformities associated with the structural sequences, together provide important constraints on viable subsidence mechanisms for the Michigan basin. Although not a quantitative geodynamic model, the stress-induced subsidence mechanism of Howell and van der Pluijm (1990) is further supported by this expanded database for two episodes of narrow, basin-centered subsidence. A stress-induced crustal weakening mechanism for subsidence offers several opportunities for further testing and evaluation based on stratigraphic data. These findings, in combination with other models linking plate dynamics with stratigraphy of plate interiors, suggest that we can look at cratonic regions for clues to understanding the complex tectonic history of ancient plate interactions.

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