

# The early Mesozoic Birdsboro central Atlantic margin basin in the Mid-Atlantic region, eastern United States: Discussion

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Fail (2003) provides an important contribution to our understanding of the geology and evolution of the Triassic-Jurassic basins of the central Atlantic margin that formed during the breakup of Pangea (e.g., Olsen, 1997). In particular, Fail draws attention to the importance of fluvial and alluvial inputs from the southeastern and northwestern margins of the Birdsboro basin (defined by Fail as the inferred continuous depositional basin consisting of the present-day Newark, Gettysburg, Culpeper, and Barbourville basins; Fig. 1) in influencing its stratigraphic architecture. He also highlights the hypothesis that substantial postrift erosion has modified the Birdsboro basin. Furthermore, Fail states that the subbasins of the Birdsboro basin are not rift basins because “the deformation (basin tilting, faulting, folding, and other structural changes) occurred after the filling of the basins. That is, the graben structure was superimposed on a preexisting, elongate, nonfaulted basin” (p. 406). In this discussion, we use seismic-reflection, outcrop, and core data to show that the Newark basin (Fig. 1) is a rift basin and that its border fault controlled sedimentation throughout most or all of its depositional history. Although the Newark basin was physically connected to other parts of the Birdsboro basin throughout much of its history, its depocenter was distinct from adjacent depocenters for most of this time. These depocenters and the intervening intrabasinal highs were structurally controlled and may have influenced the location of fluvial and alluvial systems on the northwestern basin margin.

## SEISMIC-REFLECTION DATA

Fail (1973, 1988, 2003) argues that the uniform dip of bedding ( $10^{\circ}$ – $15^{\circ}$ ) over large regions of the Newark basin indicates postdepositional faulting. Fail (2003) also claims that

seismic-reflection profiles “merely show the present cross-sectional geometry of the basin remnants [and] reveal nothing of the timing of faulting.” We agree that seismic-reflection profiles show the present cross-sectional geometry. We argue, however, that seismic-reflection profiles constrain the timing of faulting by revealing the presence or absence of synfaulting units (growth beds). Typically, growth beds exhibit thickness changes, fanning geometries (increasing dip with depth), and facies changes adjacent to the fault that controlled their deposition.

Seismic line NB-1 crosses the southwestern Newark basin (Figs. 1 and 2A; see Bally et al., 1990, for the complete uninterpreted seismic-reflection profile). Our interpretation of the seismic line (Fig. 2B) honors all available surface geology and drill-hole data (e.g., Ratcliffe et al., 1986; Olsen et al., 1996). The seismic data and interpretation show that the oldest synrift unit (not exposed at the surface) and the overlying Stockton Formation (exposed at the surface) thicken markedly toward the border fault. Restoration to the end of Stockton deposition (Fig. 2C) shows that bedding dips during the deposition of the oldest synrift units progressively increased with depth from  $0^{\circ}$  to  $\sim 5^{\circ}$ . The thickening toward the border fault and the subtle fanning with depth indicate that border faulting and deposition were coeval in the Newark rift basin. Only a few degrees of the present-day  $10^{\circ}$ – $15^{\circ}$  dip of the Stockton Formation occurred prior to or during its deposition. Thus, we infer that most, but not all, of the tilting occurred after the deposition of the Stockton Formation. The relatively shallow dips and subtle fanning of the oldest synrift strata in the Newark basin are similar to that observed in many other modern and ancient rift basins (e.g., Morley, 1999; Withjack et al., 2002).

Seismic line NB-1 does not image enough of the Lockatong and Passaic formations to determine whether or not they are growth deposits. However, seismic line 85SD10 (see Fig. 1 for location) does show that the Lockatong Forma-

tion thickens toward the border-fault system (Fig. 2D), indicating that it is a growth deposit (Reynolds, 1994). As discussed below, core and outcrop data also provide critical information about the tectonic environment during the deposition of the Lockatong and Passaic formations.

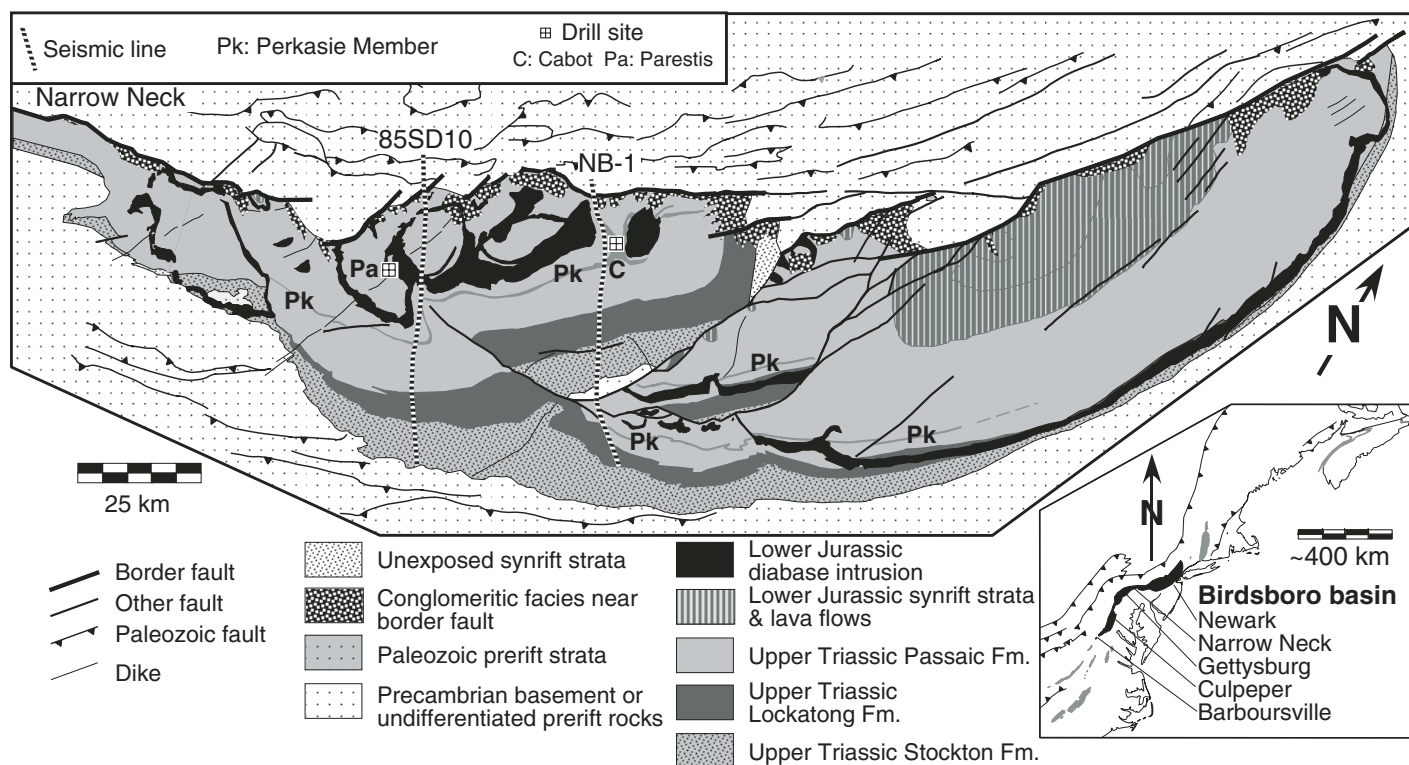
## CORE AND OUTCROP DATA

Core and outcrop data (Olsen et al., 1996) show that the Lockatong Formation thickens toward the border fault. The Skunk Hollow and Tohickon members are present in the Nursery core and outcrops at Byram, New Jersey (see Fig. 3A for location). The correlative units (constrained by cyclostratigraphy and magnetostratigraphy) are  $\sim 30\%$  thicker at Byram compared to the Nursery site (Fig. 3B) and exhibit a fanning geometry. The Lockatong Formation also changes thickness along strike: the lower Lockatong Formation is  $\sim 28\%$  thicker at the Nursery site than at the Princeton site (Olsen et al., 1996, their Fig. 14), located 36 km farther toward the northeastern lateral edge of the basin. The map width of the Lockatong Formation (Fig. 3A) varies sympathetically with the core thickness data. Although some of the changes in map width are due to variations in dip magnitude, most of the changes result from thickness changes. Thus, the Lockatong Formation thickens from the northeastern and southwestern edges of the basin toward the basin depocenter as well as from the southeastern margin toward the northwestern border-fault margin.

Core and outcrop data show similar thickness changes in the Passaic Formation (Schlische, 1992). More importantly, core and outcrop data show clear fanning geometries. For example, the overlap section of the Rutgers and Somerset cores (Fig. 3E) shows that equivalent units thicken by 14% between the Rutgers and Somerset sites, separated by 19 km (Fig. 3A). Each unit thickens toward the Somerset site, defining a fanning geometry. Furthermore, lacustrine facies accumulated in deeper water

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**Figure 1.** Geologic map of the Newark basin showing location of seismic-reflection profiles and drill sites discussed in text. Inset shows location of Newark basin (part of the larger Birdsboro basin of Faill, 2003) in eastern North America. Simplified from Schlische (1992) and Olsen et al. (1996).

in the Somerset core compared to the Rutgers core (i.e., correlative facies are finer grained and darker colored in the Somerset core compared to the Rutgers core). The Perkasie Member, which was recovered in the bottom of the Rutgers core and in the top of the Titusville core, correlates to outcrop sections traced for more than 125 km across the Newark basin (Fig. 3A; Olsen et al., 1996). Correlative units from the Rutgers site are 34% thicker at the Titusville site (Fig. 3D), located 33 km closer to the basin center. Correlative units at Tylersport are 32% thicker at Milford (Fig. 3C), located 25 km closer to the basin center. Correlative units thicken by 70% from the Rutgers site to Milford (Fig. 3E) and by 30% from the Titusville site to Milford, a distance of 25 km measured perpendicular to the border-fault system (Fig. 3A). The Milford section, which is closest to the border-fault system, contains the coarsest clastic rocks (gravel to cobble-sized grains).

The thickness changes and fanning geometries based on the seismic, core, and outcrop data indicate that the Stockton, Lockatong, and Passaic formations are growth deposits. Basin subsidence varied both perpendicular and parallel to the border-fault system, generally con-

sistent with the three-dimensional deformation produced by displacement on an isolated normal fault growing in length through time (Schlische, 1992, his Fig. 12). However, this simple fault-growth model does not account for some of the complexity present in the Newark basin. For example, the border fault is not a single fault but is composed of multiple segments, most of which are reactivated Paleozoic and older faults (e.g., Ratcliffe et al., 1986) (Figs. 1 and 2B). Nonetheless, the large-scale basin geometry and variations in thickness described above indicate that this segmented border-fault system behaved like a single fault (Schlische, 1992, his Fig. 12) for much of its history.

An additional departure from the simple fault-growth model is that the border fault of the Newark basin is not isolated but is connected to the border faults of other subbasins of the Birdsboro basin. The thickness and facies changes in the Lockatong and Passaic formations indicate that the depocenter of the Newark basin was distinct from the other depocenters of the Birdsboro basin for much of its history. Therefore, the Narrow Neck between the Newark and the Gettysburg basins was an intrabasin high during sedimentation (e.g., Anders and Schlische, 1994).

As discussed by Ratcliffe and Burton (1985), the Narrow Neck intrabasin high is most likely related to the attitude of the border fault with respect to the extension direction (WNW-ESE) during rifting. This extension direction yielded a smaller component of dip-slip displacement on the E-striking border fault of the Narrow Neck compared with the NE-striking border-fault system of the Newark basin. A reduced dip-slip displacement yielded reduced basin subsidence, produced the intrabasin high, and caused growth strata to thin toward this region. The situation is different at the northeastern lateral edge of the basin. Here, displacement on the border-fault system simply decreased to zero.

Although the majority of our discussion has focused on the evidence of syndepositional faulting, the Newark basin, like many other rift basins of eastern North America, has undergone significant postdepositional extensional deformation or postrift shortening deformation (Faill, 1973, 1988; Withjack et al., 1998; Schlische et al., 2003). Uplift associated with postrift shortening resulted in erosion (>5 km in places; e.g., Steckler et al., 1993; Malincono, 1999) of the Newark basin, significantly reducing the size of the basin.

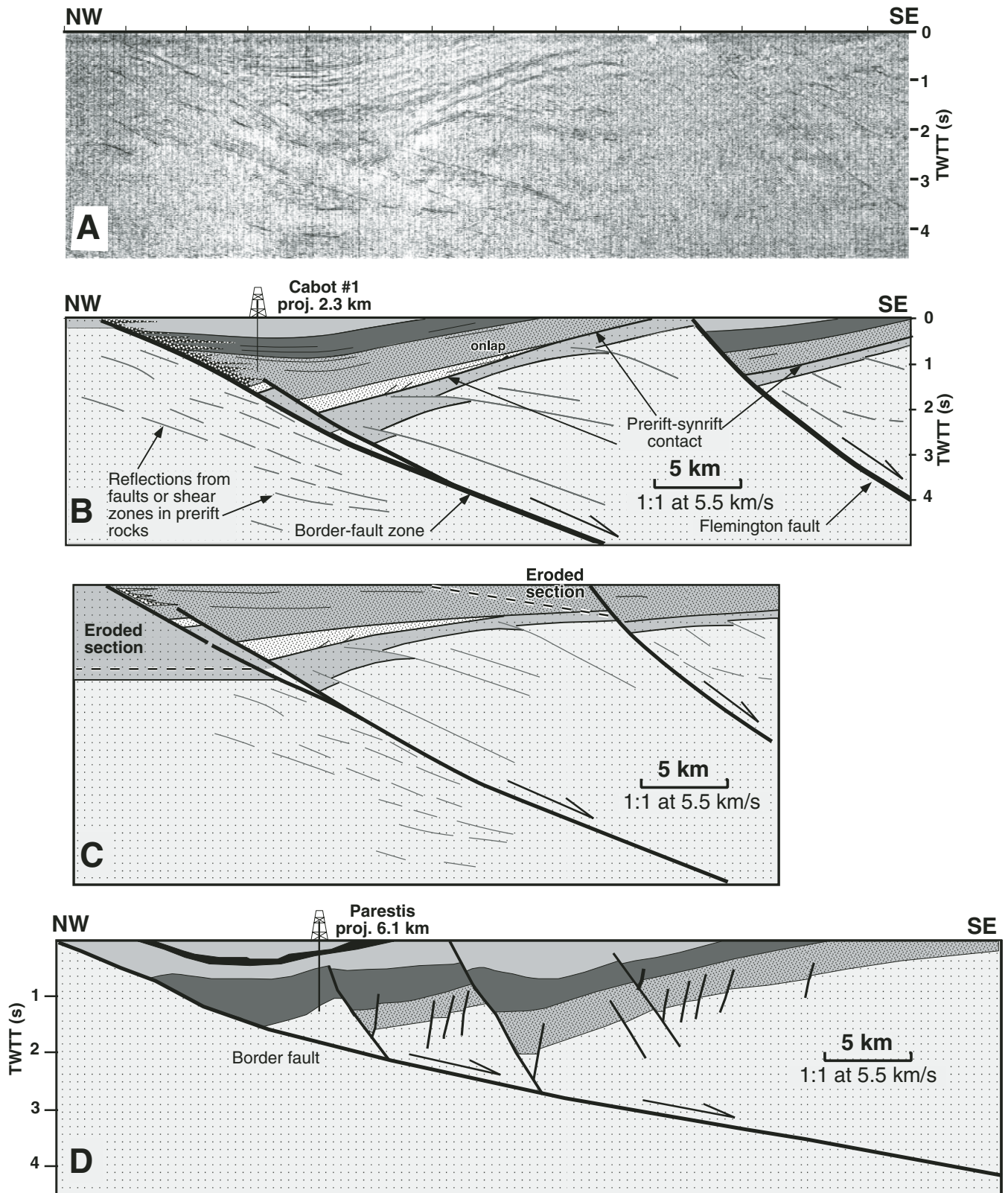


Figure 2. (A) Seismic-reflection profile NB-1; see Figure 1 for location. The seismic-reflection profile is displayed with no vertical exaggeration for a seismic velocity of 5.5 km/s. Modified from Bally et al. (1990). (B) Interpretation of line NB-1 constrained by surface geology and the nearby Cabot #1 well. See Figure 1 for legend. (C) Restoration of NB-1 to end-Stockton deposition. We decompacted the section and used vertical simple shear to make the top of the Stockton Formation horizontal. (D) Interpretation of seismic line 85SD10 (see Fig. 1 for location and legend) constrained by surface geology and the nearby Parestis well. Adapted from Reynolds (1994).

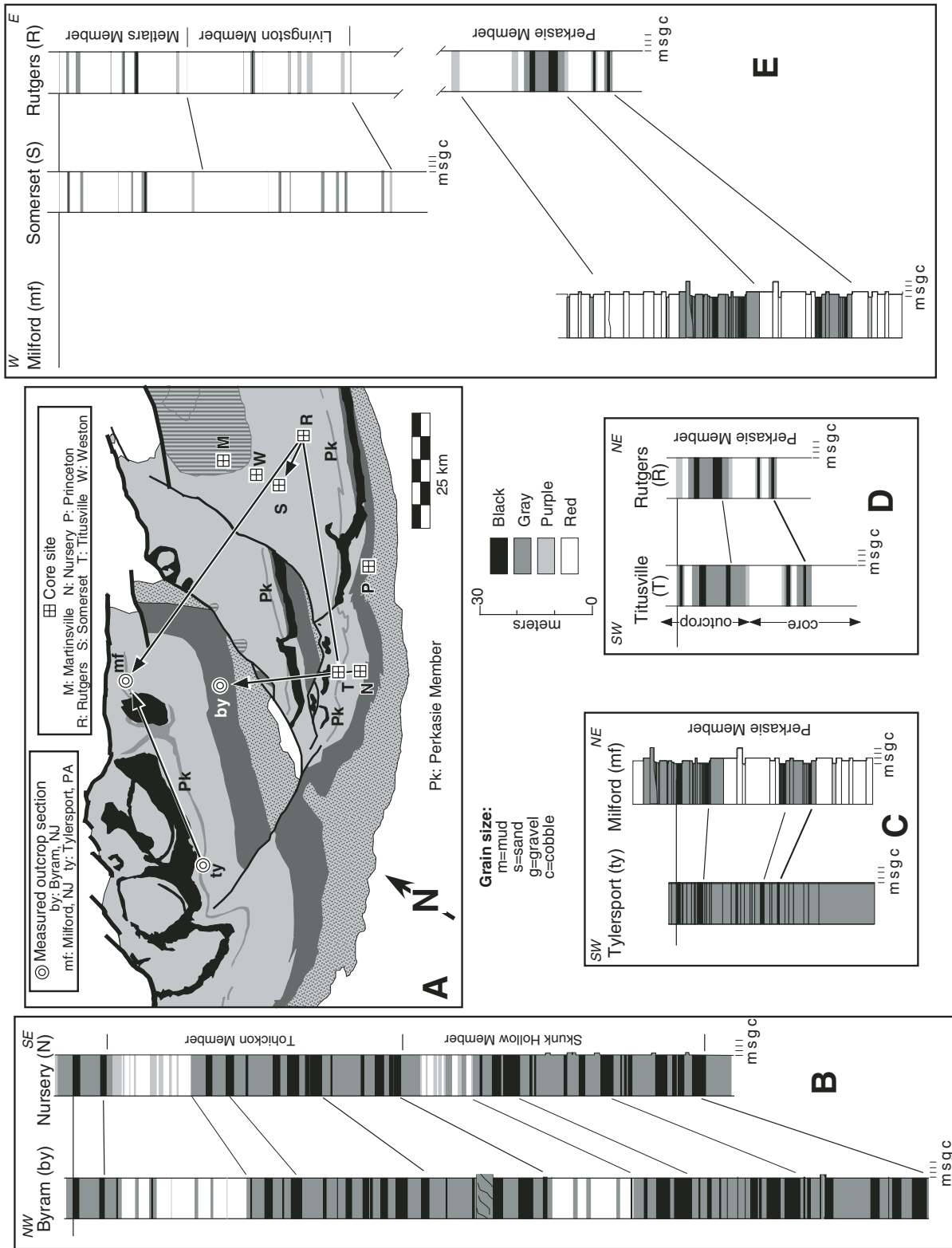
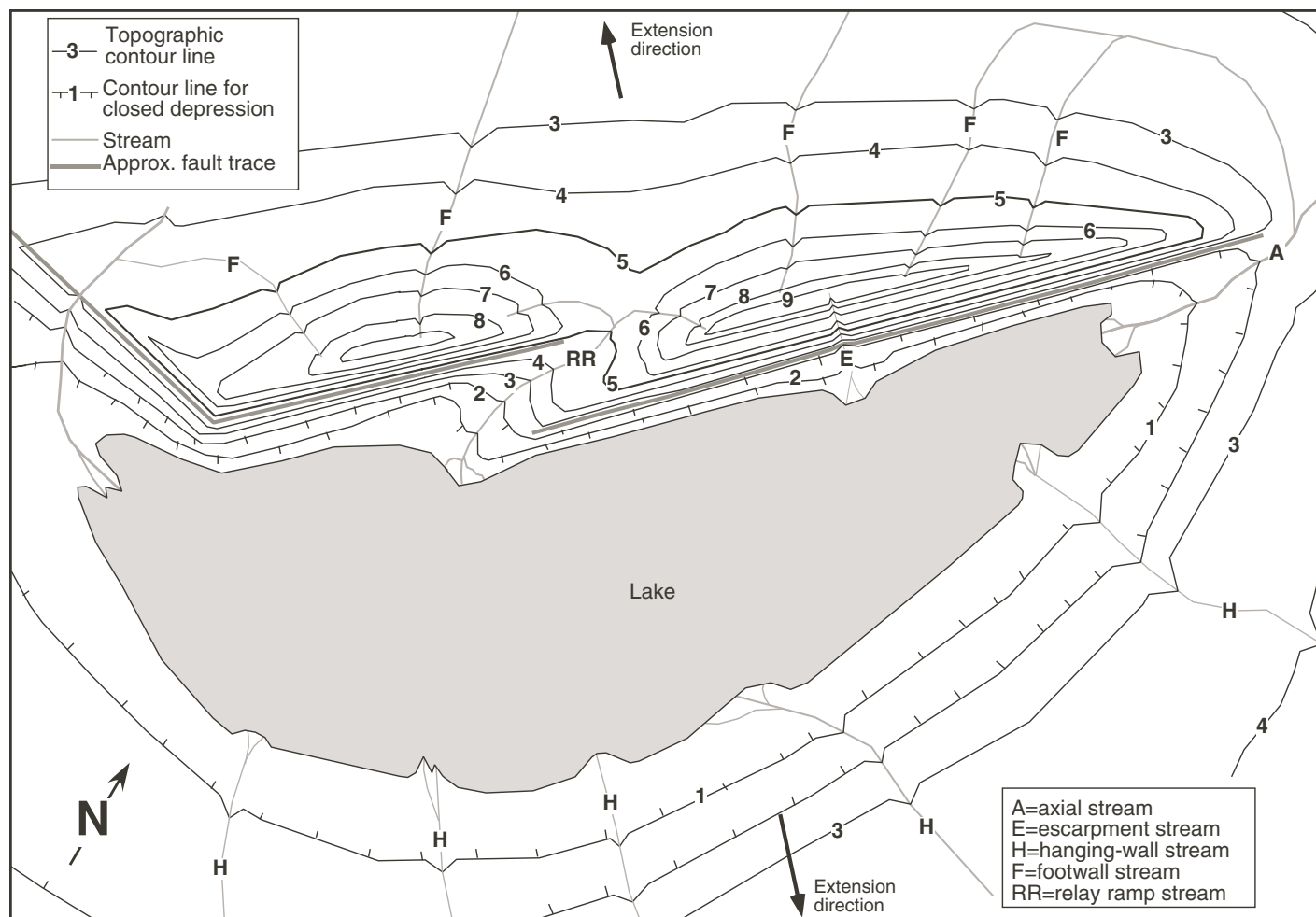


Figure 3. Thickness changes, facies changes, and fanning geometries in the Lockatong and Passaic formations. Adapted from Schlische (2003) based on data in Olsen et al. (1996). (A) Simplified geologic map of the central Newark basin (NJ—New Jersey; PA—Pennsylvania) (see Fig. 1 for legend), showing locations of core sites and outcrop sections as well as directions of thickening (arrows). (B) Part of the correlative sections of the middle Lockatong Formation in the Nursery core and the Byram outcrop section. Although the two sections are located in different fault blocks, the observed thickening trends are inconsistent with dip-slip movement on the intrabasinal faults during the deposition of this stratigraphic interval. (C) Correlative sections of the Perkasio Member of the lower Passaic Formation at Tylersport and Milford. (D) Correlative sections of the Perkasio Member in the Rutgers and Titusville cores. (E) Part of the correlative sections of the middle Passaic Formation in the Rutgers and Somerset sections and correlative sections of the Perkasio Member in the Rutgers core and Milford outcrop section. The upper and lower parts of the Rutgers core shown in this figure are separated by ~530 m of intervening section.



**Figure 4.** Idealized topography associated with footwall uplift and hanging-wall subsidence (partially filled with a lake) produced by a basin-bounding fault system. Total fault displacement decreases toward the northeastern end of the fault system; dip-slip displacement is lower on the E-striking fault segment that is oblique to the regional extension direction (WNW-ESE). Major streams flowing directly into the basin are hanging-wall, axial, and relay-ramp streams. Adapted from Schlische (2003).

#### IMPLICATIONS FOR SEDIMENT TRANSPORT

The topography produced by faulting and basin formation not only controls the thickness and facies of lacustrine deposits but also influences the patterns of fluvial and alluvial sedimentation (e.g., Cohen, 1990; Gawthorpe and Hurst, 1993; Gawthorpe and Leeder, 2000). In the Newark basin, conglomeratic deposits along the border-fault system are relatively small and discontinuous in map view (Fig. 1), consistent with footwall rivers that flowed away from the basin (Fig. 4). However, some of the largest conglomeratic deposits are associated with relay ramps along the border-fault system (Schlische, 1992), where fault segmentation produced footwall elevation lows that allowed streams to flow directly from the footwall block into the basin

(Fig. 4). Paleocurrent and provenance data indicate that the hanging-wall block to the southeast of the basin and the lateral ends of the basin provided most of the coarser sediment in the Newark basin (see references in Faill, 2003), consistent with faulting-induced topography (Fig. 4).

Faill (2003) proposed that the regional alluvial fans of the Mahwah and Hammer Creek fluvial systems controlled the large-scale stratigraphic architecture of the Newark basin. Although Faill (2003) argues that these fluvial systems were inherited from regional drainages established after the Alleghanian orogeny, we propose that structural controls on these fluvial systems during rifting were at least equally important. The Mahwah fluvial system was located near the northeastern axial end of the Newark basin, where footwall uplift was presumably minimal, allowing fluvial systems derived from the

uplifted footwall block to enter the basin from the northeast (see Fig. 4). The Hammer Creek fluvial system was located along the Narrow Neck between the Newark and Gettysburg basin. As discussed above, dip-slip displacement along the border fault was lower here during rifting. Therefore, footwall uplift was also lower, allowing major fluvial systems to enter the basin from the northwest (see Fig. 4).

In summary, we agree with Faill (2003) that postrift deformation and erosion have reduced the original maximum width of the Newark basin. However, synrift deformation was also important: seismic, core, and outcrop data demonstrate that border faulting and sedimentation occurred during the deposition of the Passaic, Lockatong, and Stockton formations as well as an unexposed unit below the Stockton Formation. The continuity of the Stockton Formation

and its equivalents throughout the Newark basin and into the Narrow Neck and Gettysburg basin indicate a continuous basin by the end of Stockton deposition (see Fig. 1 of Faill, 2003). Faill (2003) argues that the present-day distinctiveness of the subbasins of the Birdsboro basin (including the Newark and Narrow Neck basins) is entirely a result of postrift deformation. However, stratigraphic data (thinning of synrift strata toward the Narrow Neck, absence of deep-water lacustrine deposits in the Narrow Neck) indicate that this distinctiveness existed throughout deposition of the Locketong and Passaic formations and was structurally controlled. Furthermore, the major fluvial inputs to the Newark basin likely were influenced by structures active during rifting and sedimentation.

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# Reply

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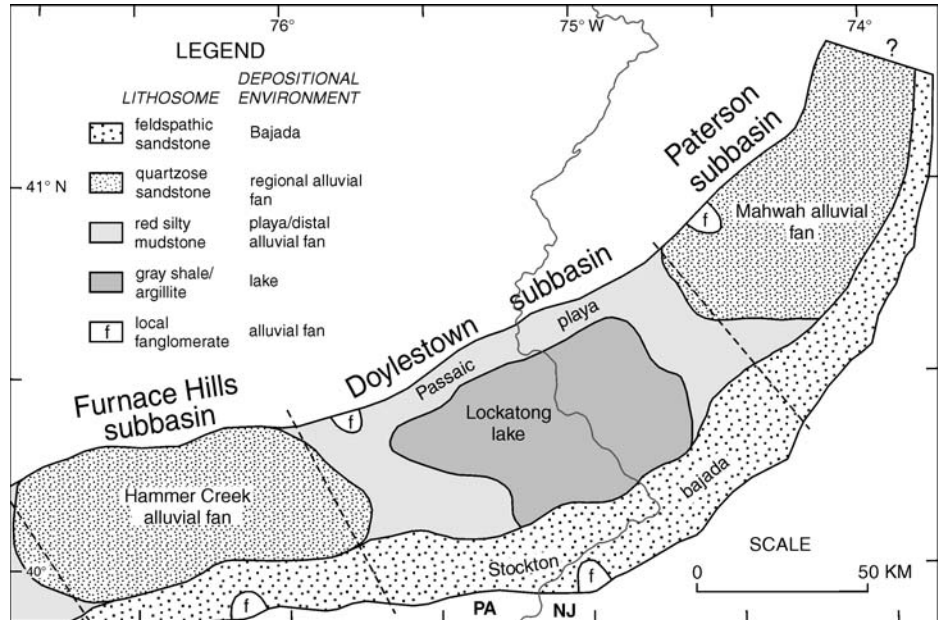
Pennsylvania Geological Survey, 3240 Schoolhouse Road, Middletown, Pennsylvania 17057-3534, USA

The discussion of the Birdsboro paper (Faill, 2003) offered by Schlische and Withjack (2005) presents their view that the Newark basin developed as a rift structure with syndepositional faulting, a hypothesis they support with outcrop, drilling, and seismic data. Unfortunately, much of their discussion is not relevant to the original paper because the origin and tectonics of the basin was specifically excluded from the Birdsboro paper.

The Birdsboro paper (Faill, 2003) examined the Late Triassic–Early Jurassic sediments contained in the four westernmost basin remnants (Barboursville, Culpeper, Gettysburg, and Newark) that lie in the mid-Atlantic region of eastern United States. This paper analyzed the lateral and vertical variations in the sediment content and makes inferences of their depositional environments and provenances. This analysis led to the suggestion that the sediments were deposited in a single, long, sinuous, northeast-trending trough that only approximately parallels the underlying Paleozoic structures of the Appalachian orogen.

The origin and tectonics of the Birdsboro basin were deliberately excluded in order to focus attention solely on the sediments. The post-depositional deformation of the Birdsboro basin (monoclinical tilt of bedding, folds, faults, etc.) was noted only to explain the contrast between the original single depositional basin and the four basin remnants (following erosion) we see today. An analogous event sequence is the postdepositional (Permian) transformation by the Alleghany orogeny of the Paleozoic central Appalachian basin into the Alleghany foreland (Appalachian Valley and Ridge and Allegheny Plateaus provinces).

Schlische and Withjack (2005) do not address the main points of the Birdsboro paper. Although they agree that postdepositional tectonism altered the Birdsboro basin, they argue that some of the deformation was syndepositional. Specifically, they posit that the basin developed as a rift during the late Triassic, with specific consequences in the depositional patterns within the Doylestown subbasin (Fig. 1, the southwestern part of the Newark remnant). For them, the northwest margin of the basin was



**Figure 1. Hypothetical map of the subbasins in the central and northern part of the Birdsboro basin, schematically illustrating the contrast in depositional environments and resultant lithosomes among the three subbasins. The rift model discussed by Schlische and Withjack (2005) applies primarily to the cyclic lacustrine and playa sediments in the Doylestown subbasin.**

an active, syndepositional normal fault and the basin formed on the subsiding hanging wall. This view is shared, or at least accepted in some form or another, by perhaps more than 99% of the geological community. I disagree with this view. Although basin tectonism was beyond the scope of the original paper, their arguments bear somewhat on sediment distributions and geometries, and therefore warrant responses. My objections to their view rest in three principal areas: (1) the presence of a syndepositional fault along the northwest margin; (2) the causes of the apparent thickening of strata toward a “depocenter”; and (3) the inapplicability of their model to other parts of the Birdsboro basin.

First, I am unaware of any incontrovertible evidence of syndepositional faulting along the northwest margin. Faults are present there, but the movement on them was Jurassic (or later) in age. They offset Jurassic sediments (Drake et al., 1996), they contain Jurassic basalt and diabase fragments (Ratcliffe and Burton, 1985;

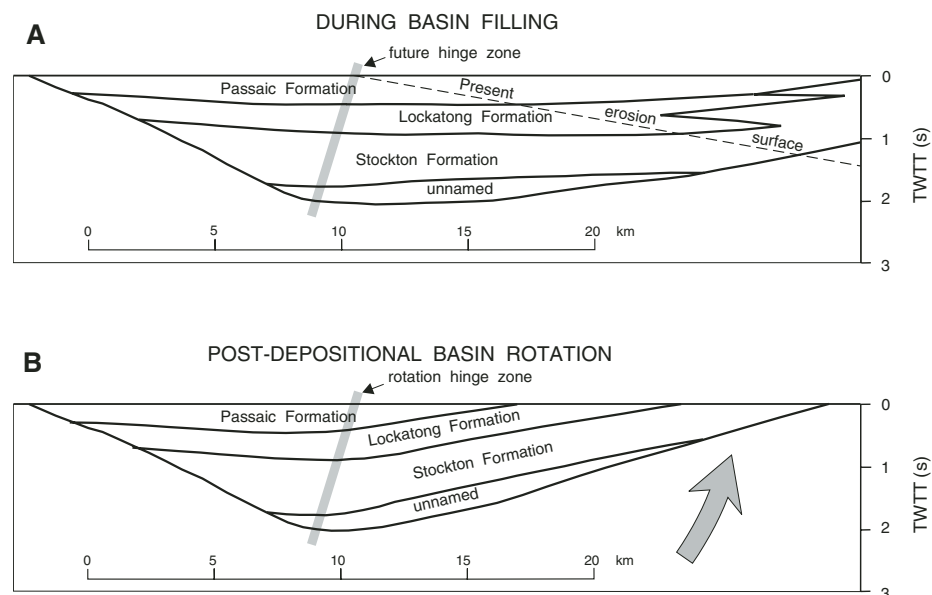
Ratcliffe et al., 1986), and they truncate or offset postdepositional (Jurassic) structures (e.g., the Jacksonwald syncline; Schlische, 1992, notwithstanding). Nowhere has definitive Triassic movement been demonstrated. The presence of coarse-grained fanglomerate deposits along the northwest margin have long, and often, been cited as evidence for syndepositional faulting there, in the Birdsboro basin and in basins elsewhere in the world. However, syndepositional faulting is not the only mechanism for producing margin fanglomerates (e.g., Faill, 1973, 1988). Texturally identical fanglomerates occur along the southeastern overlap margin of the Birdsboro basin. There, mechanical weathering of carbonate hills in an arid climate, colluviation, and occasional storms produced the debris flows that deposited the fanglomerates along the basin margin. The same process probably dominated the northwest margin as well. In other words, margin fanglomerates by themselves are *not sufficient* evidence for syndepositional faulting.

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The NB-1 seismic line (Costain and Çoruh, 1989; Schlische and Withjack, 2005; their Fig. 2A) provides a remarkable cross-sectional image of the Doylestown subbasin along the Delaware River. Superficially, it appears to show a half-graben, albeit one with an unusual geometry. For example, the basin's deepest point lies under the center of the basin, at the intersection of two opposing moderately dipping reflectors, presumed to be the basin floor (Costain and Çoruh, 1989). In the southeastern half of the basin, several higher reflectors that parallel the basin floor reflector are interpreted to represent sedimentary bedding. The persistent  $13^\circ$  dip through the Stockton, Lockatong, and lowest Passaic Formations (as identified in Schlische and Withjack, 2005, their Fig. 2B) is nearly identical to surface measurements. In the northwestern part of the basin, these same reflectors are subhorizontal to southeast dipping (up to  $6^\circ$ ) and terminate against the  $24^\circ$  southeast-dipping basin floor reflector. Shallow drilling (Ratcliffe et al., 1986) shows that, close to the northwest margin, the basin floor dips  $27^\circ$  to the southeast along a Jurassic fault (basalt clasts are in the fault gouge). A deeper drill hole, the #1 Parestis, encountered the basin floor at 2048 m  $\sim$ 7 km from the northwest margin (Reynolds, 1994), yielding a mean slope of  $16^\circ$ . This correspondence between the apparent dip ( $24^\circ$ ) in the seismic line and surface/drilling measurements indicates that the NB-1 seismic line (as presented) is a good approximation of the actual Doylestown subbasin cross section; however, this is not a typical cross section geometry for a half graben.

Indeed, there is no need for a syndepositional fault to explain this seismic line—other interpretations are equally valid. For example, the northwest margin, the southeast-dipping basin floor reflector, could be an overlap, as is the southeast margin (Fig. 2A). The parallel reflectors truncated against the basin floor reflectors represent the sediments that progressively overlapped both basin margins. The present subhorizontal attitude of these reflectors in the northwest part (Schlische and Withjack, 2005, their Fig. 2A) indicates that these beds were not monoclinaly rotated as were those in the rest of the basin. Only the southeastern part in their Figure 2A was rotated post-depositionally (post-Early Jurassic) (their Fig. 2B). In short, the seismic line requires no syndepositional fault along the northwest margin.

Second, Schlische and Withjack (2005) argue that growth beds, with a subtle fanning geometry in the oldest units, indicate syndepositional faulting. The seismic reflectors in this deepest part of the basin (below the bend in the “bedding” reflectors) are poor—even the basin



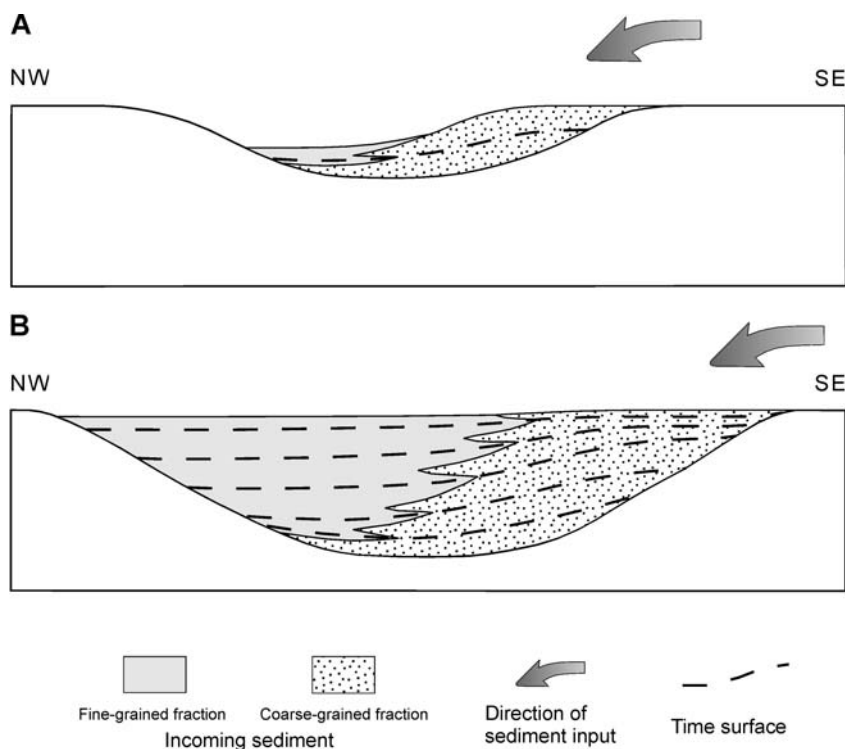
**Figure 2.** A non-rift representation of the Birdsboro basin, Doylestown subbasin, with an overlap on the northwest margin. Horizontal kilometer scale is approximately equal to vertical two-way travel time scale (TWTT). (A) Cross-sectional geometry of the basin at mid-Passaic time (age of youngest beds in the NB-1 seismic line). (B) Cross-sectional geometry of eroded basin following a post-depositional  $10^\circ$  rotation of the southern part of the basin. Note the similarity to the geometry of the NB-1 seismic line (Schlische and Withjack, 2005, their Fig. 2A).

floor loses definition (their Fig. 2A). Schlische and Withjack (2005) have drawn a fault splay (rider block) and a thickening of the oldest beds (including Stockton) in this area (their Fig. 2B). These features are not clearly evident in this “fuzzy” part of the seismic line. In addition, their drawn contacts dip toward the faults despite individual seismic reflectors being subhorizontal. I submit that this indistinct imagery is a very weak basis for hypothesizing syndepositional faulting, especially in the absence of any other definitive, confirming evidence.

Stratal thickening is a key argument for their rift hypothesis. Schlische and Withjack (2005) cite drill core and outcrop evidence of stratal thickening toward a “depocenter” near the northwest margin (their “border fault”). This evidence is based on the excellent lithic, stratigraphic, and magnetostratigraphic data accumulated and generated by Paul Olsen and his coworkers over the past two decades. Olsen et al. (1996) have demonstrated that not only can members of the Lockatong and Passaic formations be traced over large distances but also individual black shale units within the members can be identified and correlated. These unit boundaries are time surfaces, validated by magnetostratigraphy, which can be traced for more than 100 km across the Doylestown subbasin and even into adjacent subbasins. Schlische (1992)

and Schlische and Withjack (2005) cite selected thickness data from this work to illustrate thickening of units (by up to 34%) toward their “depocenter.” In effect, their argument is that the northwestward increase in spacing between time surfaces implies a differential subsidence that can best be explained by a rifting process, with basin floor rotation during syndepositional faulting along the basin margin.

However, the NB-1 seismic line does not bear this out. Despite the disclaimer by Schlische and Withjack (2005) that the line does not image enough of the Lockatong Formation to show its growth character, they draw the entire Lockatong as present (see their Fig. 2 especially 2B), which is imaged much better than the underlying Stockton Formation. Their cross section (their Fig. 2B) shows constant thickness of the Lockatong Formation from south of the Flemington fault northwestward to where the reflectors become horizontal. Northwestward from there, their Lockatong becomes slightly thinner, just where, according to their model, it should be thickening. The reflectors in the overlying lowest Passaic Formation also show no northwestward thickening. Now this is odd, for if individual cycles thicken to the northwest and the yet the formation doesn't, then either the seismic line presents an inaccurate image of the basin or other cycles within the formation must thin in that direction.



**Figure 3. A non-rift model for producing northwestward-thickening units (represented by time surfaces) in a non-faulted, bowl-shaped basin, with a sediment source to the southeast. (A) In the early stages of basin development, the sediments are predominantly coarse-grained and accumulate in a bajada on the southeast side of the basin. The paucity of fine-grained sediment results in a northwest-sloping sediment surface within the basin. (B) As the basin enlarges and the incoming sediment becomes muddier, the northwestern lacustrine/playa side of the basin fills more rapidly than on the bajada. As a result, the spacing of time surfaces (represented by traceable cyclic units) increases northwestward.**

An alternative explanation exists for this conundrum, one that does not require a rift interpretation for the basin. Suppose, for simplicity, a two-dimensional basin (oriented northwest-southeast) with a slowly subsiding bowl-shaped floor (Fig. 3 herein). When fluvial sediment is introduced from the southeast, the coarser fraction accumulates near the entry point and the winnowed finer-grained fraction spreads across the remainder of the basin. If the fluvial sediment were 75% coarse-grained, the proximal accumulation would be thicker, thereby producing a northwestward-sloping sediment surface (a time surface) within the basin (Fig. 3A herein). If, at some later time, the fluvial sediment mix changes to 75% fine-grained, then proximal accumulation would lessen and the distal thickness increase, with the finest-grained fraction settling in the deepest water as the darkest mud (Fig. 3B herein). Thus, with finer-grained mixes, the cycle time surfaces would diverge to the northwest, whereas with coarser-grained mixes, the time surfaces would converge. In addition, varia-

tions in the incoming mix would produce lateral interfingering of lithologies as well as thickness changes. Furthermore, a package of thickening and thinning cycles (making a formation) could end up with a constant overall thickness. And none of this requires any syndepositional faulting or any differential subsidence. Or, is the seismic line wrong—does the constant formation thickness mirepresent reality?

Expanding this model to three dimensions, the axial movement of fine-grained sediment from the Hammer Creek regional alluvial fan in the Furnace Hills subbasin and the Mahwah fan in the Paterson subbasin to the northeast (Fig. 1) is analogous. Schlische and Withjack (2005) cite such examples (their Figs. 3C and 3E). Indeed, the Lockatong Formation is respectively absent or has minimal thickness in the adjacent subbasins. But the total thickness along the Birdsboro basin is relatively constant, southwestward to the Culpeper remnant (Faill, 2003, Table 1). Within this total thickness, the spacing between individual time surfaces (unit thickness) can

vary from place to place and from time to time. As implied above, the fluctuation of time surface spacing (unit thickness) may be less a function of differential subsidence than of location in the basin, grain-size mix and composition of incoming sediment, and climate.

Schlische and Withjack (2005) argue that not only was the northwest margin a syndepositional fault, but the intrabasinal Flemington fault (and presumably the Chalfont, Furlong, and Hopewell intrabasinal faults) was also active during Stockton sedimentation (their Fig. 2C) and, presumably, thereafter. If the Birdsboro basin consisted of independently moving structural blocks (Schlische, 1992) then one would expect some effect of these syndepositional movements on the distribution of the sediments. Yet, the remarkable lateral continuity of the lacustrine units within the Lockatong and Passaic Formations (Olsen, 1988; Olsen et al., 1996) suggests otherwise. My question is: if a lacustrine environment is quiet and stable enough to produce a depositional cyclicity in response to periodic, astronomically induced climate changes, how can major, intrabasinal faults leave no sedimentary evidence of their syndepositional activity? To put it differently, if the basin consisted of several independently moving structural blocks, then the active faulting would have interrupted the lateral continuity of the cycles, thereby rendering the cycle thickening argument untenable. Given the stratal continuity, it would seem that the intrabasinal fault movements must have been postdepositional. In short, a long-term tectonic quiescence persisted throughout the basin, with gradual basinwide subsidence lasting for more than 20 m.y.

Third, the Schlische and Withjack (2005) comments neglect most of the Birdsboro basin—their rift model discussion applies primarily to the Doylestown subbasin (southwestern part of the Newark remnant; Fig. 1 herein). They do not address the sedimentology in the other parts of the basin. These other subbasins contain different lithosome sequences, which grade laterally into adjacent lithosome sequences with no abrupt breaks. One of the main points of the Birdsboro paper is that the differences in sequences reflect changes of sediment input and distribution along the length of a continuous basin. One most important component is the regional alluvial fan (the Cedar Mountain, Goose Creek, Hammer Creek, and Mahwah fans).

Schlische and Withjack (2005) do address the Hammer Creek and Mahwah fans, but they attribute a structural origin to them, following Anders and Schlische (1994). They argue that the two fans lie over structural highs, caused by the overlapping of the border faults to produce relay ramps. To create the structural highs, they infer a

decrease in fault displacement on each fault into the relay ramps that results in lesser subsidence, hence a thinning of stratigraphic units over the ramps. However, two aspects of the basin contradict this view—the location of the relay ramps, and the stratigraphic thicknesses.

Their relay ramps in the Doylestown subbasin lie a short distance northeast and southwest of Milford, New Jersey, close to their depocenter. Yet, their depocenter is just where their maximum subsidence occurred, with maximum, not minimal, accumulation of sediment (fanning growth beds). In addition, they argue that less sediment accumulated away from their depocenter, in the Hammer Creek alluvial fan in the Furnace Hills subbasin (the “narrow neck”) and the Mahwah fan in the Paterson subbasin (Fig. 1 herein), but no relay ramps are present adjacent to those fans.

Schlische and Withjack (2005) do not mention Table 1 in the original Birdsboro basin paper (Faill, 2003). That table summarizes the lithic character (lithosomes) and stratigraphic thicknesses along the basin (from Virginia to New York), with references to the data sources. It demonstrates that the total Late Triassic–Early Jurassic sediment accumulation was relatively constant at  $7 \pm 1$  km along the entire length of the basin. In short, there is no decreased subsidence in the alluvial fan areas. In that Schlische and Withjack (2005) did not take issue with that data nor even mention that table, I conclude that either they agree with it or they simply chose to ignore it.

This is not to say that units or formations do not thin toward the fans—some do, especially the Lockatong Formation. However, do all the cycles in that formation thin to zero, or do the number of gray, lacustrine cycles decrease toward the fans because of lateral gradation into a red, coarser-grained lithology? Schlische and Withjack (2005) do not address this question, yet it is an important one. Although the Lockatong lake environment persisted for several million years, the lake was eventually filled in and replaced by a playa environment (Passaic Formation). This change may simply reflect an increase of sediment into the basin, or increased aridity, or both. Whatever the reason, the gradational, inter-layered contact between the Lockatong and the Passaic Formations suggests the change was progressive, with the playa environment encroaching on the lake environment (with occasional temporary reversals). Within this depositional pattern of progressive shrinkage of the lake’s areal extent, it follows that the package of gray lacustrine cycles (the Lockatong Formation) would contain fewer cycles in its distal parts as compared with the number near the depocenter. In other words, the change in lithosomes and lithosome sequences

(formations) from one part of the basin to another was caused for reasons other than simple thinning of cycles away from a depocenter by differential subsidence.

Sediment provenance was discussed in the original Birdsboro basin paper (Faill, 2003) and has significant relevance to the rift model of Schlische and Withjack (2005). Perhaps 50% or more of the total sediment entered the Birdsboro basin from the northwest through the four regional alluvial fans (Faill, 2003). By inference, this volume of sediment requires an extensive drainage system developed over a large area to the northwest that slopes toward the basin. Yet, their rift model contradicts this inference by showing a footwall topography and drainage sloping away from the basin (their Fig. 4). Their minor footwall streams drain areas far too small to generate the sediment volume necessary to supply the regional alluvial fans. They do not suggest any other sources for the alluvial fan sediment.

Schlische and Withjack (2005) refer to the numerous small fanglomerate deposits along the northwest margin of the basin, in particular to those adjacent to a relay ramp (their Fig. 1). I agree that these debris-flow deposits entered through small, local drainages, but I consider their adjacency to a relay ramp coincidental—most such deposits occur far from any relay ramp. Indeed, the largest (by an order of magnitude) fanglomerate deposit in the basin, the Leesburg member of the Bull Run Formation in the northern part of the Culpeper remnant, is not associated with any relay ramp. As discussed above, fanglomerates are probably unrelated to any (hypothetical) syndepositional faulting. And, they are completely different from, and unrelated to, the regional alluvial fans.

In summary, the rift model as presented by Schlische and Withjack (2005) is related only to the Doylestown subbasin—they do not discuss the lithosome relationships throughout the remainder of the basin. But, even in the Doylestown subbasin, their model raises more questions than it answers. They have not demonstrated that syndepositional faulting actually occurred—instead, they only present evidence to support a preconceived fault model, and even some of the evidence they cite is inconsistent with their rift hypothesis.

Rift tectonism is not a good model for the Birdsboro basin, as seductively simple as such a model may seem. Many aspects of the basin are not explained by rifting. That is why the focus in the original Birdsboro basin paper (Faill, 2003) was only on the sediments, their distributions, and the inferences that could be made therefrom. Discussion of how the basin formed and the overall causative tectonics was eschewed in

order to avoid the introduction of preconceived notions (half-graben, rift tectonism). It is hoped that a better understanding of the sediment content and its distribution will help in developing a tectonic model that fully explains the origin, development, and subsequent history of the entire Birdsboro basin.

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