Structural and stratigraphic development of the Newark extensional basin, eastern North America: Evidence for the growth of the basin and its bounding structures

ROY W. SCHLISCHE *Department of Geological Sciences, Rutgers University, New Brunswick, New Jersey 08903*

ABSTRACT

The Newark basin is a half graben, bounded on its northwestern margin by a system of mostly normal-slip faults, many of which are reactivated Paleozoic thrust faults and were active at least sporadically during the deposition of the entire basin fill. As a result of along-strike variations in displacement on the border-fault system, subsidence increased from the lateral edges toward the center of the basin and from the hinged margin toward the border-fault system. Some intrabasinal faults also formed during the deposition of units currently preserved in their hanging walls. Along-strike variations in the density of intrabasinal faults probably reflect changes in the dip angle of the border-fault system; variations in the amount of extension accommodated on these intrabasinal faults resulted in the formation of a transfer fault. Along-strike variations in fault displacement probably produced the transverse folds in the hanging walls immediately adjacent to border faults and major intrabasinal normal faults; some folds formed syndepositionally. Progressively younger strata onlap "basement" rocks along the hinged margin of the basin, and younger splays of the border-fault system propagated into the footwall; the basin therefore grew deeper, wider, and longer through time.

INTRODUCTION

The Newark basin is perhaps the most extensively studied of the early Mesozoic basins of eastern North America that formed during the breakup of Pangea (Fig. 1). The basin is ~190 km long, excluding the narrow neck connecting it with the Gettysburg basin to the west, and a maximum of 50 km wide. In general terms, the basin is bounded on its northwestern and northern margins by a system of mostly normal faults (subsequently referred to as the border-fault system (BFS) (Figs. 1 and 2). Along the southeastern and southern margins, Mesozoic rocks rest unconformably on Precambrian and Paleozoic "basement" rocks; in places, Mesozoic rocks are buried by Cretaceous coastal-plain strata. Although disrupted by intrabasinal faults and warped by transverse folds developed in the hanging walls of major normal faults, strata generally dip toward the BFS (Figs. 3 and 4). The Newark basin is therefore a classic half graben, the fundamental expression of continental extension.

Erosional truncation has exposed virtually the entire basin fill as well as the bounding and internal structures (faults, dikes, joints, and folds). Key stratigraphic markers can be traced great distances laterally in the basin (Olsen, 1988); the same chronostratigraphic units are exposed in each of the major intrabasinal fault blocks. Consequently, one can examine how the same unit varies in thickness and facies with structural position in the basin and thus address the age of formation of the BFS, intrabasinal faults, and folds. Excellent chronostratigraphic control coupled with a 25-m.y.-long synrift record allows one to document the evolutionary changes in basin geometry, including changes in the depth, width, and length of the basin, and the critical effect this had on the large-scale stratigraphic development of the basin. This is difficult to obtain in most extensional basins; virtually the entire synrift basin fill is buried by the modern depositional surface and/or hundreds of meters of water in active or recently active extensional settings and by thousands of meters of post-rift strata on ancient passive margins.

GEOLOGIC BACKGROUND

Nine formations are recognized in the Newark basin and consist of six sedimentary units comprised of exclusively terrestrial deposits and three basalt flow formations (Table 1). Basin strata are also intruded by numerous quartz-normative tholeiitic diabase sills and dikes (Puffer and Philpotts, 1988). The sedimentary rocks are dated biostratigraphically, principally by pollen and spores (Cornet and Olsen, 1985), and range in age from Carnian (Late Triassic) to Sinemurian (Early Jurassic). The Triassic-Jurassic boundary is placed in the uppermost Passaic Formation, a few meters to tens of meters—depending on position in the basin—below the first basalt flow (Olsen and others, 1990).

As indicated in Table 1, most of the strata were deposited in lakes and exhibit a pervasive hierarchy of cyclicity in sediment fabrics, color, and total organic carbon content. As originally proposed by Van Houten (1962) and quantified by Olsen (1986) through Fourier analysis, the cyclicity resulted from basin-wide fluctuations in lake level driven by the Milankovitch orbital cycles of precession (~20 k.y.) and eccentricity (~100 and ~400 k.y.). All cyclical lacustrine sections in the Newark basin subjected to Fourier analysis yield the same cycle periods in time (Olsen and others, 1989), even though the thicknesses of the cycles vary with structural position in the basin.

The number and hierarchy of fixed-period cycles interbedded with the lava flows limit the duration of the extrusive interval to ~600 k.y. (Olsen and Fedosh, 1988; Olsen and others, 1989). This igneous activity occurred at 201 Ma, based on isotopic dates for the Palisades sill (Sutter, 1988; Dunning and Hodych, 1990), which is physically connected to lava

Geological Society of America Bulletin, v. 104, p. 1246–1263, 13 figs., 1 table, October 1992.

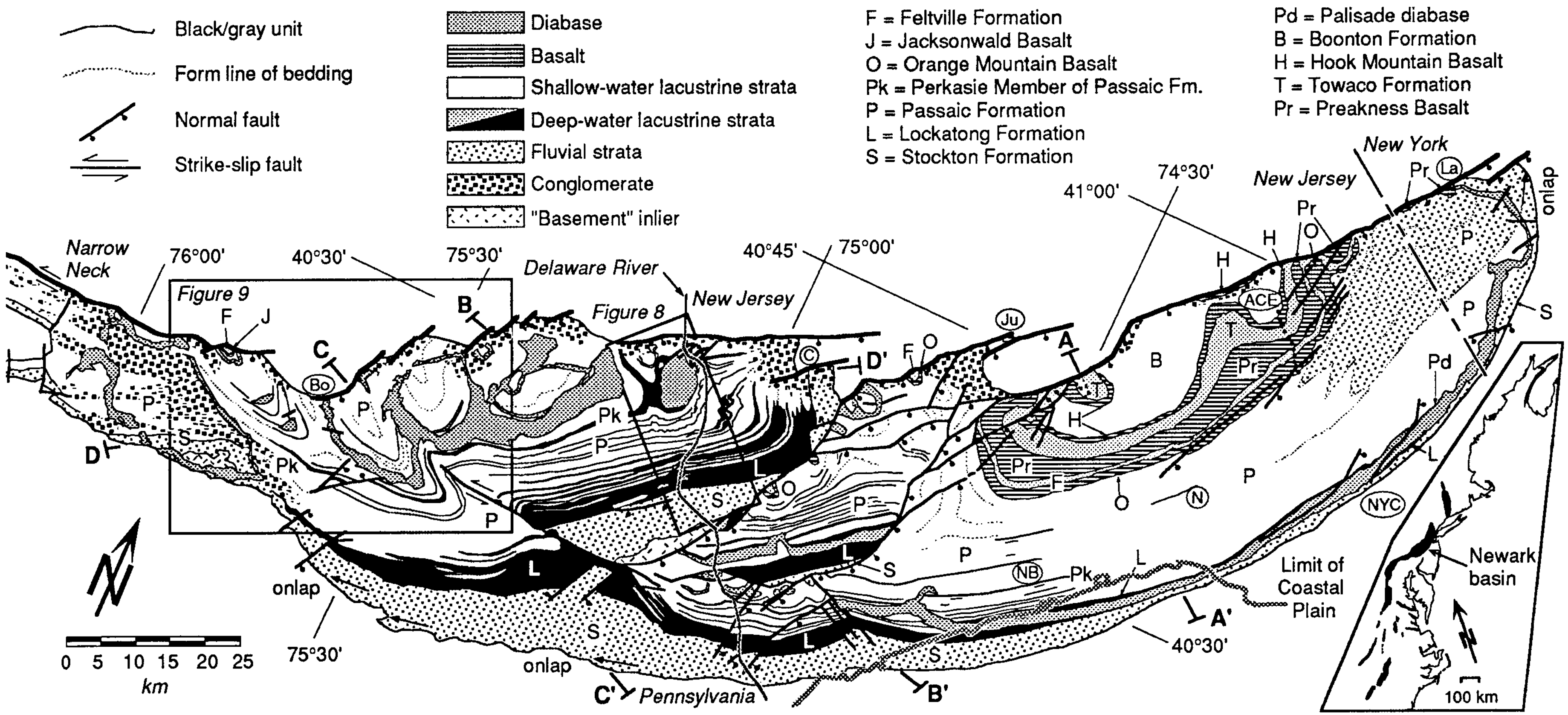


Figure 1. Geologic map of the Newark basin, illustrating locations of cross sections in Figure 3. Abbreviations for geographic locations are circled: ACE, Army Corps of Engineers core holes; Bo, Boyertown; C, Clinton; Ju, Jutland; La, Ladentown; N, Newark; NB, New Brunswick; and NYC, New York City. Map compiled from Berg (1980), Olsen (1980b), Ratcliffe and others (1986), Lyttle and Epstein (1987), Parker and others (1988), Ratcliffe and Burton (1988), Schlische and Olsen (1988), P. E. Olsen (1987–1990, personal commun.), H. Houghton (1990, personal commun.), and mapping by P. E. Olsen, M. Levy, and the author. Inset shows location of Newark basin with respect to other exposed early Mesozoic basins in eastern North America. Modified from Schlische (1990).

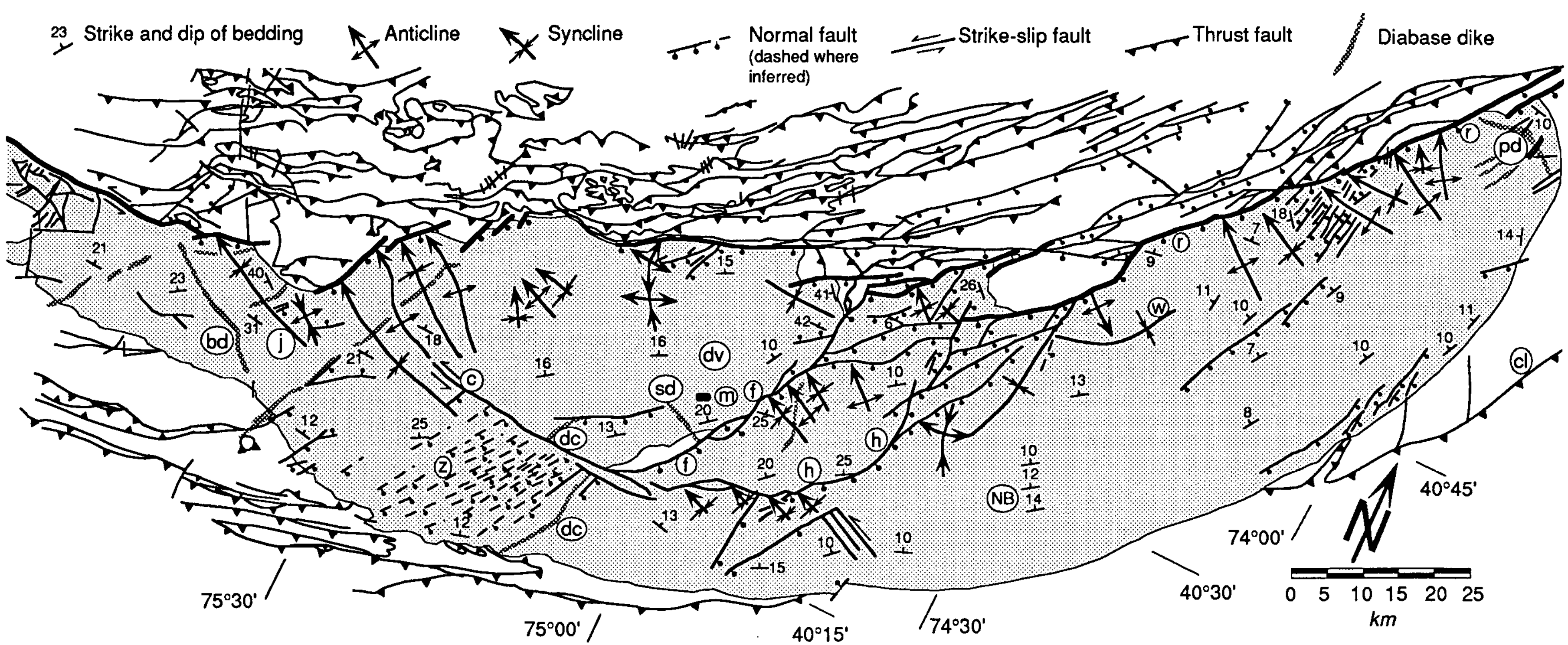


Figure 2. Structural map of the Newark basin. Abbreviations for specific structures are bd, Birdsboro dike; c, Chalfont fault; cl, Cameron's line; dc, dike apparently offset by Chalfont fault; f, Flemington-Furlong fault; h, Hopewell fault; j, Jacksonwald syncline; pd, dike-like extension of the Palisades sill; r, Ramapo fault; sd, Solebury dike; w, Watchung syncline; and z, zone of intense normal faulting south of the Chalfont fault (shown diagrammatically). Abbreviations for localities: dv, Delaware Valley region, and NB, New Brunswick region. The locality of the joint survey (Fig. 6C) is marked by the letter m. See Figure 1 for references.

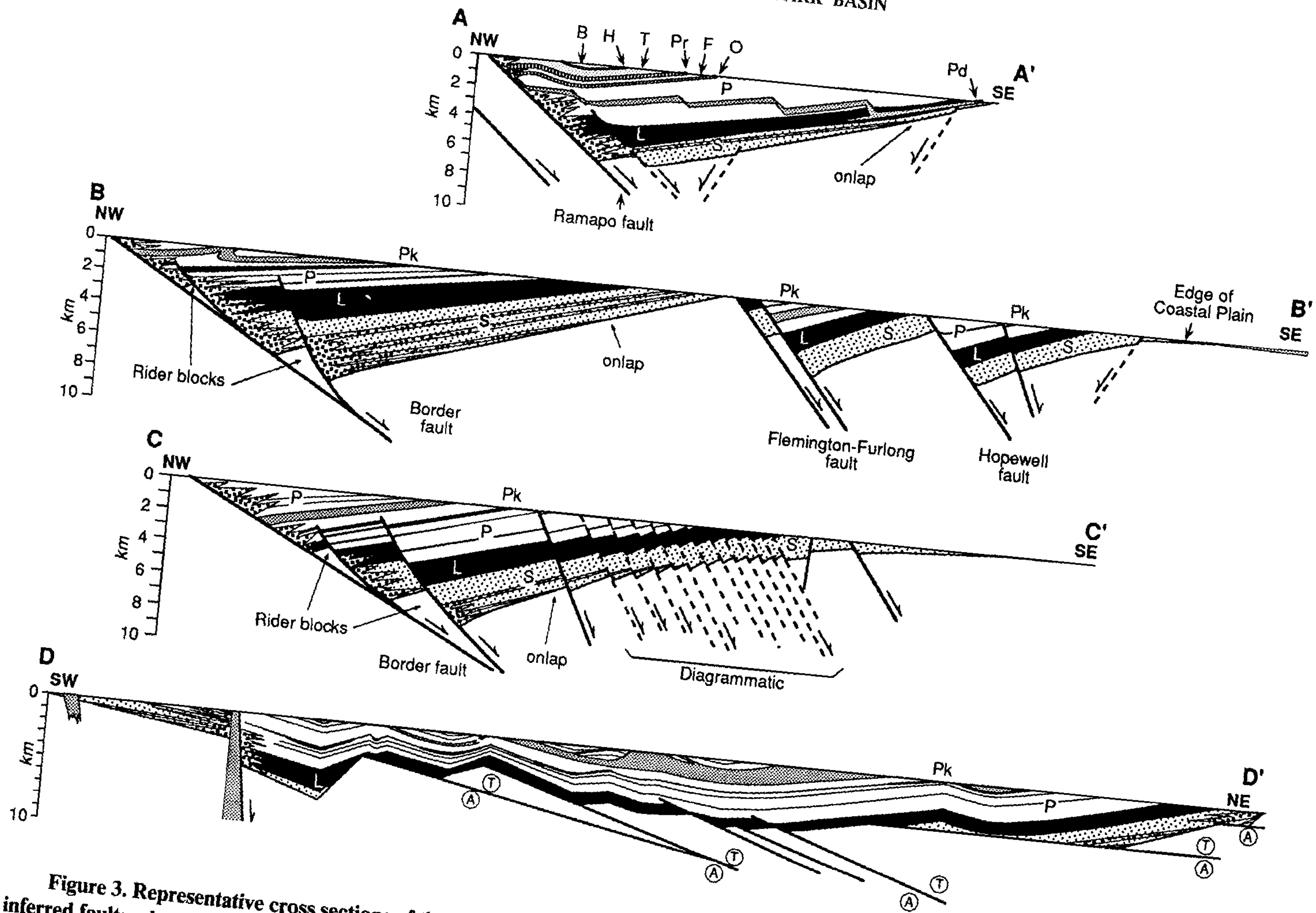


Figure 3. Representative cross sections of the Newark basin (see Fig. 1 for locations). Abbreviations as in Figure 1. Dashed lines represent inferred faults; closely spaced faults in section C–C′ are schematic. Transverse sections illustrate asymmetric nature of the basin. Rider block fault splays are based on along-strike projections of adjacent border-fault segments. The shallow dip of Triassic strata near the southeast ends of sections B–B′ and C–C′ is an apparent dip. Longitudinal section D–D′ shows the overall synclinal shape of the basin. The association of the BFS segments (projected down-dip from the surface) with particular folds appears fortuitous, given the obliqueness of the line of section with respect to the fold axes and the strike of the fault segments. T is toward the reader, A is away. Scale for D–D′ is different than in other sections. No vertical exaggeration in any section.

flows in the northern part of the basin (La in Fig. 1; Ratcliffe, 1988). For the purposes of this paper, all igneous rocks are considered to have been emplaced penecontemporaneously.

STRUCTURAL ARCHITECTURE

Basin-bounding and Intrabasinal Faults

For much of its length, the border-fault system consists of a number of fault segments, ranging in length from <1 km to 40 km, which generally display a right-stepping relay geometry (Fig. 2). At fault overlaps known as "relay ramps" (terminology of Larsen, 1988), Mesozoic strata rest unconformably on pre-rift rocks. Rider blocks (terminology of Gibbs, 1984) are also present along the BFS. Their three-dimensional geometry is illustrated in a block diagram (Fig. 5) of the area near Clinton, New Jersey (C in Fig. 1). In general, southeastward-situated fault segments become blind and lose displacement to the southwest; progressively older strata are absent in northwestward-situated rider blocks (for example, the Passaic Formation rests directly on "basement" at relay ramp in Fig. 5 and at Jutland, Ju in Fig. 1). Such relations also have been inferred from seismic reflection data (Reynolds and others, 1990) and indicate that younger faults propagated into the footwall, widening the basin through time. The presence of rider blocks causes the deepest part of the basin to be shifted basinward from the BFS (Figs. 3 and 5).

Individual fault segments of the BFS strike north-northeast, northeast, and east, respectively, in the northeastern, central and southwestern, and narrow neck parts of the basin (Fig. 2). The Mesozoic BFS closely parallels Paleozoic faults of Appalachian affinity, and the basin itself follows prominent oroclinal bends that formed in the Permo-Carboniferous (Miller and Kent, 1986). The near-surface dip of border faults, determined by drilling, generally decreases from the northeastern terminus of the basin (70°SE) to the Delaware River region (25°–30°SE), a trend mimicked by Paleozoic thrust faults in the footwall of the BFS (Ratcliffe and Burton, 1985; Figs. 3 and 4). These observations and cores of the fault-zone rocks

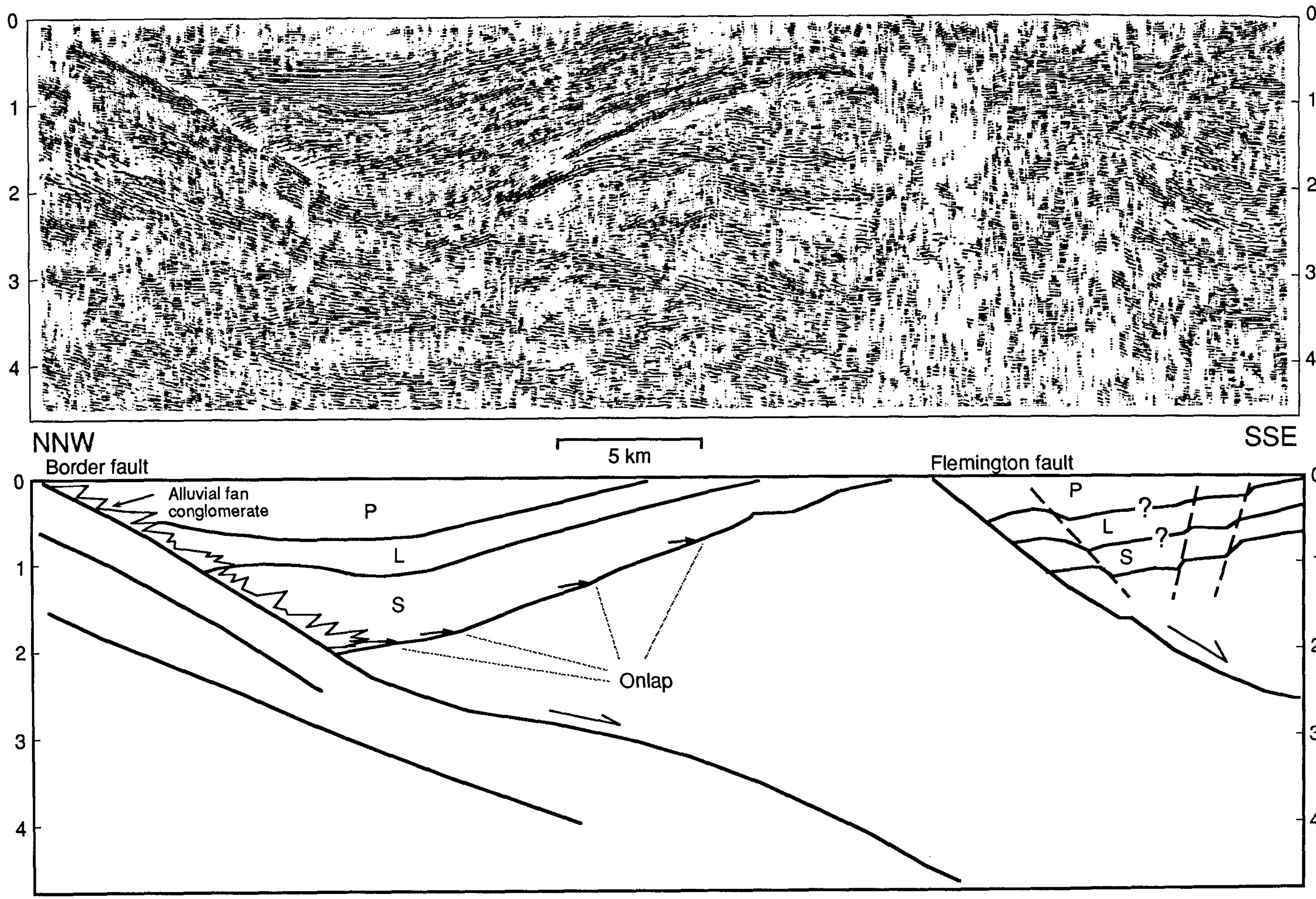


Figure 4. The NORPAC NB-1 seismic reflection profile across the central Newark basin (southwest of the Delaware River). Top: automatic line drawing of the time-migrated profile, modified from a figure presented by Costain and Çoruh (1989); actual data presented in Bally and others (1990). Two-way travel time is twtt, in seconds. Bottom: interpretation of the profile. Note the hanging-wall onlap within the Stockton Formation, the thickening of the Stockton Formation (S) toward the border fault, and the presence of alluvial-fan conglomerate in the Passaic (P), Lockatong (L), and Stockton Formations at depth.

TABLE 1. STRATIGRAPHY OF THE NEWARK BASIN

Unit	Thickness (m)*	Age	General description
Boonton Formation	500	Hettangian-Sinemurian	Mostly red, some gray and black (rarely deep-water), cyclical lacustrine clastic and carbonate rocks; minor red fluvial sandstone and alluvial fan conglomerate
Hook Mountain Basalt	110	Hettangian	Quartz-normative tholeiitic basalt flows
Towaco Formation	340	Hettangian	Mostly red, some gray and black (commonly deep-water), cyclical lacustrine clastic and carbonate rocks; minor red fluvial sandstone and alluvial fan conglomerate
Preakness Basalt	250	Hettangian	Quartz-normative tholeiitic basalt flows and minor interbedded strata
Feltville Formation	170	Hettangian	Mostly red, some gray and black (commonly deep-water), cyclical lacustrine clastic and carbonate rocks; minor red fluvial sandstone and alluvial fan conglomerate
Orange Mountain Basalt	150	Hettangian	Quartz-normative tholeiitic basalt flows and very minor interbedded strata
Passaic Formation	3300	L. Carnian-Hettangian	Mostly red, shallow-water, cyclical lacustrine clastic rocks; some gray and black cyclical lacustrine clastic rocks with minor limestone, particularly in lower part; minor red fluvial sandstone and alluvial fan conglomerate
Lockatong Formation	1100	L. Carnian	Mostly gray and black (commonly deep-water), fine-grained, cyclical lacustrine carbonate and clastic rocks; some red, fine-grained lacustrine and very minor fluvial clastic rocks
Stockton Formation	1800	M.-L. Carnian	Mostly coarse brown, fine red, and very minor gray fluvial clastic rocks

Note: modified from Olsen and others, 1989.
*Estimated maximum from outcrop width and dip magnitude.

indicate that many of the individual border faults, even those currently dipping at <30°, represent reactivated Paleozoic faults (Ratcliffe, 1971, 1980; Ratcliffe and Burton, 1985; Ratcliffe and others, 1986). Seismic profiles indicate that the shallow fault dip of the BFS persists southwest of the Delaware River region (Reynolds and others, 1990); however, west of Boyertown (Bo in Fig. 1), individual faults in outcrop steepen to 70°–80° (Lucas and others, 1988; Olsen and others, 1989). Dip slip and right-oblique slip occurred on the BFS in the central and northeastern basin, respectively (Ratcliffe and Burton, 1985); left-oblique slip occurred at the eastern end of the narrow neck (Lucas and others, 1988).

Border faults are commonly depicted as strongly listric, soling into shallow subhorizontal detachments (Hutchinson and Klitgord, 1988; Bell and others, 1988; Manspeizer, 1988). Border faults imaged on seismic reflection profiles in the Delaware River region (Ratcliffe and others, 1986; Costain and Çoruh, 1989; Bally and others, 1990) are planar to 2-sec two-way travel time (~6 km depth, assuming a seismic velocity of 6 km/sec). The border fault has a pronounced kink below the hanging-wall cutoff of the "basement"-sediment contact on the time-migrated NB-1 line (Fig. 4; Costain and Çoruh, 1989). Velocity pull-ups beneath Mesozoic basins, however, commonly give the impression of listric geometry (Unger, 1988). Until a depth-migrated version of the profile becomes available, the issue of the geometry of the BFS at depth will remain unresolved.

Given the variations in the surface dip of the BFS, cross sections through the northeastern, central, and southwestern parts of the Newark basin appear considerably different (Fig. 3). In the northeast, the BFS consists of a moderately dipping, continuous fault segment (the Ramapo fault; r in Fig. 2), throw is maximized, and few if any rider blocks are present. In the central basin, the dip of the BFS is 25°–30°, and the basin is cut by two major intrabasinal normal fault systems. In the southwest, the BFS has the shallowest dip, rider blocks are common, and the basin is dissected by numerous intrabasinal faults.

Nearly all of the intrabasinal faults are northeast-striking normal faults synthetic to the BFS; a few faults are east to east-southeast striking (Fig. 6A). The largest intrabasinal faults (in terms of stratigraphic separation) are the Flemington-Furlong, Hopewell, and Chalfont faults.

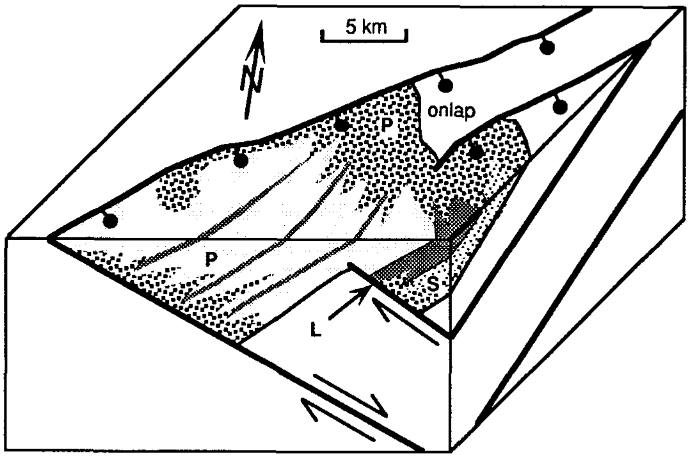


Figure 5. Block diagram illustrating rider block geometry near Clinton, New Jersey (C in Fig. 1). The southeast border-fault splay becomes blind and loses displacement to the southwest; displacement also decreases to the northwest. Older strata are absent from northwest rider block. Abbreviations as in Figure 1.

The Flemington-Furlong and Hopewell faults (f and h in Fig. 2) divide the central Newark basin into three fault blocks in which the Triassic stratigraphic section is repeated. The throw on the Flemington-Furlong fault is at least 4 km (see Fig. 3), and that on the Hopewell fault is estimated to be 2–3 km, based on the non-unique separation of stratigraphic units. Using the apparent offset of a fold that was assumed to have been once continuous across the Hopewell fault and the apparent right-lateral offset of the BFS along the postulated northeastward continuation of both the Hopewell and Flemington-Furlong faults, Sanders (1962) argued for as much as 19 km of strike slip. The offset in the BFS, however, is due to the presence of relay ramps; thus the intrabasinal faults could not have experienced significant strike slip. Coring indicates that the Flemington-Furlong fault dips 47°–50° southeast and experienced predominantly dip slip with a right-slip component; extension was southeast directed (Ratcliffe and Burton, 1988). Minor faults in the hanging wall of the Flemington-Furlong fault support a similar conclusion (Hozik, 1985).

The east-striking Chalfont fault (c in Fig. 2) is exposed from the Flemington-Furlong fault in the east to its termination in the west apparently in the hinge of an anticline. The fault has been proposed to be a north-dipping reverse fault that is part of a fault-propagation fold (Faill, 1988), a southeast-dipping normal fault (Willard and others, 1959), and a left-lateral fault on the basis of its alignment with the left-lateral faults of the narrow neck (Manspeizer, 1988). Sinistral slip is also supported by (1) apparent sinistral offset of stratigraphic contacts; (2) sinistral offset of a once continuous subvertical diabase dike (dc in Fig. 2); (3) drag folds consistent with left slip; and (4) minor, steeply dipping, left-slip fault zones subparallel to the Chalfont fault (Schlische and Olsen, 1988).

The fault block to the south of the Chalfont fault (z in Fig. 2) contains numerous northeast-striking, southeast-dipping, planar normal faults (Fig. 7; Watson, 1958). The density of intrabasinal faults and the resultant horizontal extension in the southern fault block decrease to the west, where both the displacement on the Chalfont fault and the fault itself die out. The fault block to the north is relatively unfaulted internally. The Chalfont fault, therefore, is a transfer fault (terminology of Gibbs, 1984) separating two regions that experienced different amounts of extension; the fault dies out where the difference in extension also diminishes. Because the southern block locally had been extended more than the northern block, the sense of slip was left lateral. As several gently northwest-

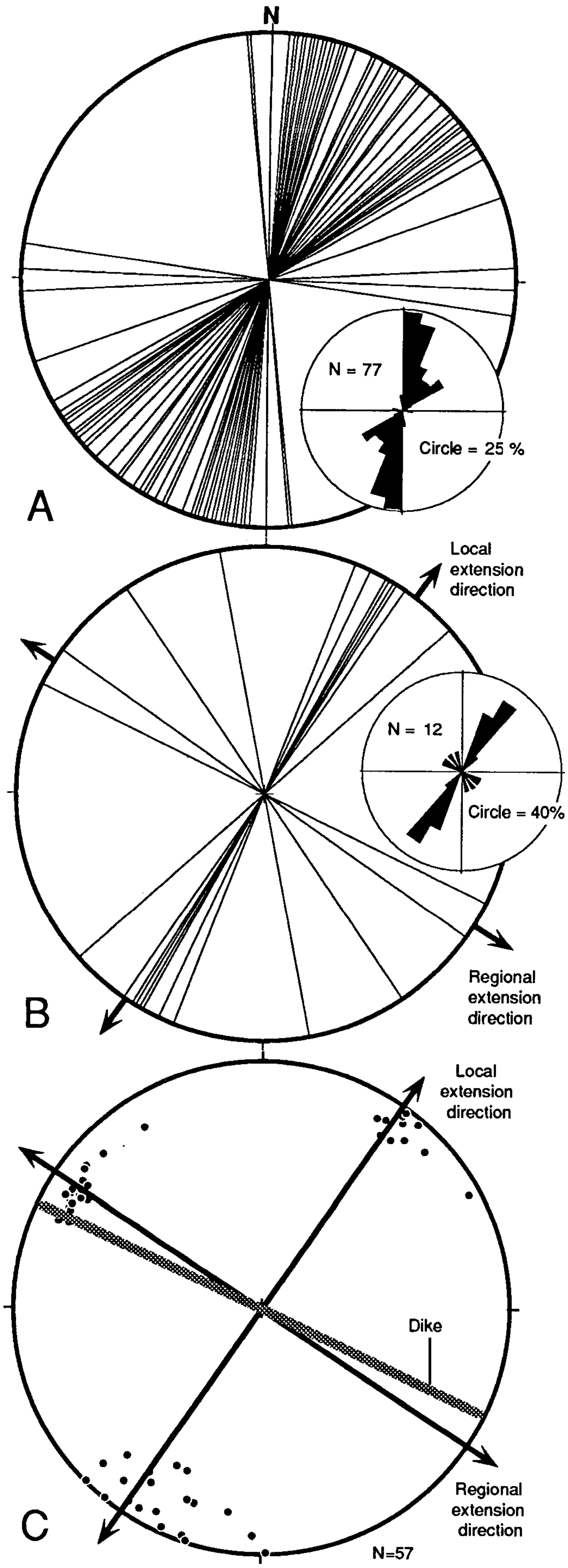


Figure 6. (A) Strikes of mapped faults of the Newark basin. Inset shows resulting rose diagram. (B) Strikes of dikes within the Newark basin and inferred directions of extension; inset shows resulting rose diagram. (C) Equal-area, lower-hemisphere stereographic projection of poles to orthogonal joint sets (with inferred extension directions) from Stockton Formation from quarry along NJ 29 north of Stockton (m in Fig. 2) with trend of nearby Solebury dike shown for reference.

plunging folds are present in this and other parts of the Newark basin (Figs. 1 and 2; see discussion below), it is likely that the association of the Chalfont fault with the anticline at its tip is coincidental, although one cannot dismiss the possibility that some degree of folding is related to movement on the fault.

Stratal Geometry

The average dip of bedding is 10°–15° northwest. Systematic variations in dip angle are difficult to observe because of the effects of intrabasinal faulting and folding. In relatively structurally uncomplicated fault blocks, such as the footwall of the Flemington-Furlong fault along the Delaware River (dv in Fig. 2), certain trends do emerge, however. This region was divided into five equal-sized areas, and dip values were numerically averaged (Fig. 8). With the exception of the area located adjacent to the BFS, where strata were affected by folding, faulting, and diabase intrusion, average dips generally decrease toward the northwest, the direction in which strata become younger, indicating that the BFS was active during deposition of the Stockton, Lockatong, and Passaic Formations. A similar decrease has been observed in the New Brunswick, New Jersey, area (NB in Fig. 2).

The attitude of bedding generally defines a monoclinal structure dipping toward the BFS, although folding complicates this pattern. In addition, the general northwesterly dip of strata gives way to a southwesterly dip near the relay ramps of the BFS and at the northeastern termination of the basin; a northeasterly dip occurs along the hinged margin at the southwest end of the basin. Overall basin geometry consists of a relay series of bowl-shaped structures plunging toward the BFS (see section D–D′, Fig. 3).

Folds and Associated Structures

Two types of folds are present adjacent to major intrabasinal and border faults of the Newark basin: (1) those with axes generally parallel to the associated faults (drag folds and forced folds) and (2) those with axes generally perpendicular to the associated faults (transverse folds). Northeast-trending drag folds are present adjacent to the Flemington-Furlong fault (Ratcliffe and Burton, 1988) and the BFS along the Delaware River near Milford, New Jersey (Olsen and others, 1989). The generally northeast-trending Watchung syncline (w in Fig. 2; see also section A–A′, Fig. 3) may be a large forced fold (terminology of Robson, 1971) formed above an upward-propagating fault.

Transverse folds are best developed in the hanging wall of the BFS, but they are also present in the hanging walls of the Flemington-Furlong and Hopewell faults (Fig. 2). The faults themselves do not appear to be folded, nor are folds present in the footwall. Furthermore, the folds decrease in amplitude and die out completely away from the faults. The folds thus appear to be related to the normal faults responsible for basin subsidence.

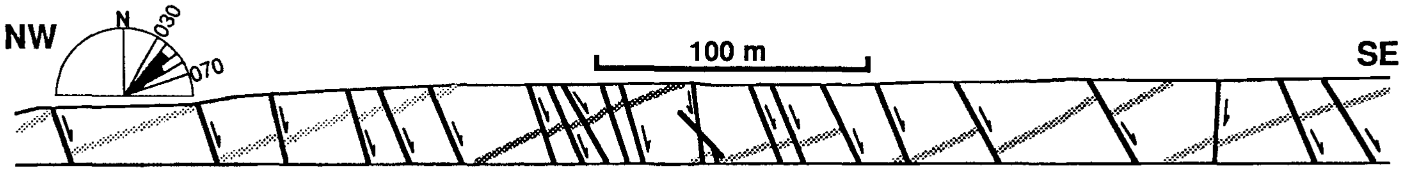


Figure 7. High-density normal faulting in Reading Railroad cut south of Gwynedd, Pennsylvania (z in Fig. 2). Stippled lines represent bedding. Rose diagram shows fault trends. Reverse fault near southeast end of section cannot be explained by rotation of an original normal fault and may indicate a small amount of post-rift shortening. Data and sketch modified from Watson (1958).

In southeastern Pennsylvania, transverse folds and diabase intrusions are interrelated. Quartz-normative diabase plutons (inferred to have been intruded at ~201 Ma) are generally concordant and sill-like in the synclines but commonly discordant to dike-like in the anticlines (Fig. 9). Two concordant diabase bodies restricted to the fold axis of the Jacksonwald syncline resemble phacoliths and indicate that they were intruded during or after some folding of the enclosing Passaic Formation (Manspeizer, 1988; Schlische and Olsen, 1988). The composition of the diabase also varies with respect to the folds (A. J. Froelich, 1989, personal commun.). Denser cumulates (for example, bronzite) are found within the structurally lower hinges of synclines (for example, the phacoliths of the Jacksonwald syncline), whereas less dense late-stage differentiates (for example, granophyre) are preferentially found on the limbs of synclines and the structurally higher crests of anticlines, suggesting folding during intrusion.

The Jacksonwald Basalt, which is equivalent to the Orange Mountain Basalt (Olsen and others, 1990) and is inferred to have been emplaced at approximately the same time as the intrusion of the phacoliths, accumulated on a relatively flat (unfolded) surface. In the northern Newark basin, however, the Ladentown lavas (La in Fig. 1) may have been "extruded across previously folded and dissected Triassic strata" (Ratcliffe, 1980, p. 293), but the bulk of the folding post-dated extrusion (Ratcliffe, 1988).

Several suites of minor structures are associated with the transverse folds in the southwestern Newark basin. Deformed mudcracks from the Jacksonwald syncline have been stretched parallel to the fold axis (Lucas and others, 1988) and/or shortened perpendicular to the fold axis. Tetrapod footprints from the syncline have been similarly deformed (Schlische and Olsen, 1988; Olsen and others, 1989; Silvestri, 1991). The overall lack of discrete surfaces of deformation (solution surfaces, cleavage planes) in the rocks containing the deformed footprints from outcrops in Douglasville, Pennsylvania (Fig. 9), even at the thin-section level of examination, suggest that deformation occurred shortly after sedimentation while the strata could still deform ductilely.

A northwest-striking, spaced pressure-solution cleavage interpreted to be axial planar is present in mudstone in some parts of the Jacksonwald syncline (Lucas and others, 1988) and locally present in other folds. Because the axial planar cleavage does not fan around the axis of the Jacksonwald syncline, it formed late in the history of folding (Lucas and others, 1988). Northeast-striking extension veins that cut the cleavage are the youngest structures to have formed in the syncline (Lucas and others, 1988).

Dikes and Joints

Several diabase dikes intrude strata of the Newark basin and adjacent "basement" rocks. Outside the basin, all of the dikes strike northeast. Most of the dikes within the basin are also northeast striking and are interpreted to reflect the regional northwest-southeast extension direction (Fig. 6B). In addition, west-northwest–striking dikes or dike-like features within the basin appear to define a northeast-southwest local extension direction. Two of the northwest-striking dikes are found at the lateral edges of the basin: the dike-like extension of the Palisades sill and the Birdsboro dike marking the boundary between the main Newark basin and the narrow neck (bd in Figs. 2 and 9). A third major northwest-striking dike (the Solebury dike, sd in Fig. 2) intruded Paleozoic carbonate rocks of the "basement" inlier as well as the Stockton and lower Lockatong Formations in the footwall of the Flemington-Furlong fault. Diabase intrusions are locally discordant, northwest-striking dike-like features within the transverse fold adjacent to the BFS in the Jacksonwald area of Pennsylvania (Fig. 9).

Two orthogonal systems of joints that mutually cut one another

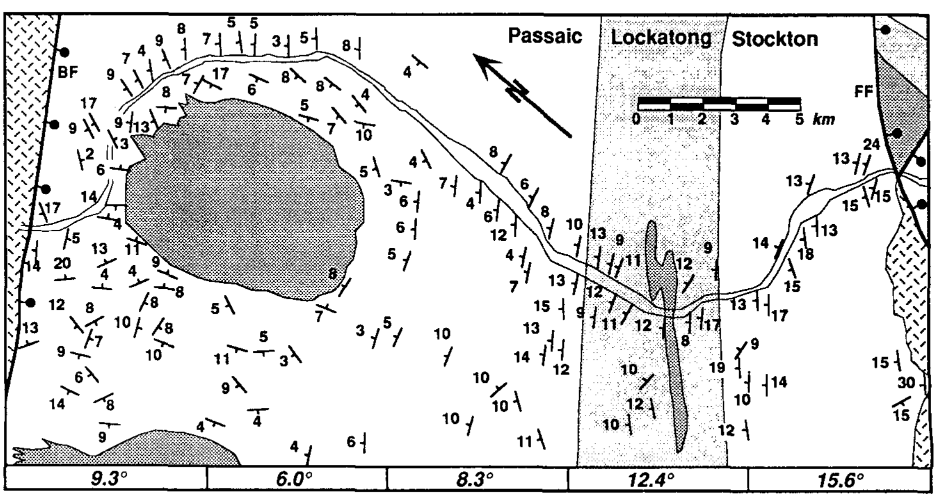


Figure 8. Geologic map of the Delaware River region (see Fig. 1 for location) showing attitude of bedding reported in Willard and others (1959). Magnitudes of dips were averaged for five areas of equal width. Note the general decrease in dip toward the northwest. Dark stipple represents diabase intrusions; light stipple represents the Lockatong Formation; cross hachures represent "basement"; BF is the border fault; FF is the Flemington-Furlong fault.

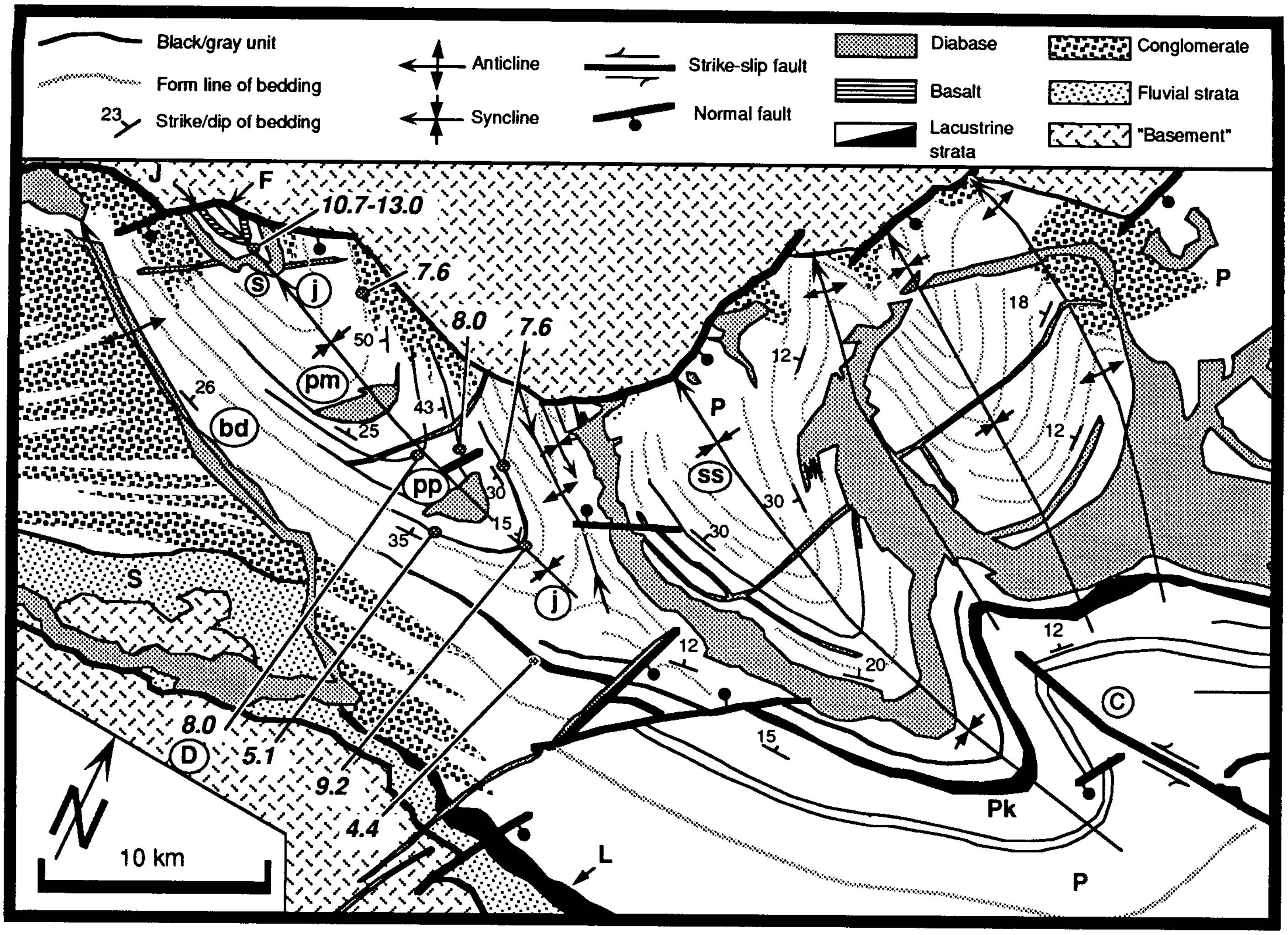


Figure 9. Geologic map of the southwest corner of the Newark basin (see Fig. 1 for location), illustrating the relations among folds, igneous rocks, and lacustrine-cycle thicknesses (numbers in italics). Abbreviations as in Figure 1 and as follows: bd, Birdsboro dike; c, Chalfont fault; D, Douglasville footprint locality; j, Jacksonwald syncline; pm, Monocacy Station phacolith; pp, Pottstown phacolith; s, Jacksonwald sill; and ss, Sassamansville syncline.

generally parallel the two main dike trends (Fig. 6C; Schlische and Olsen, 1988; Schlische, 1990) and may have formed coevally. Northeast-striking joints are commonly very well developed in hornfels of some diabase intrusions (Schlische and Olsen, 1988; Olsen and others, 1989). The density of the predominantly northeast-striking joints diminishes abruptly away from the diabase and within the diabase itself, suggesting that the joints may be related to hydrofracturing associated with elevated fluid pressures at the time of intrusion in a stress field in which the effective minimum principal stress was subhorizontal and southeast directed.

STRATIGRAPHIC ARCHITECTURE

In this section, elements of the stratigraphy of the Newark basin that shed important light on its tectonic development are examined. These include (1) conglomerate that accumulated on the downthrown sides of normal faults, (2) transverse and lateral changes in thickness and facies of time-equivalent units, and (3) onlap of strata onto "basement" rocks.

Conglomerate Adjacent to Normal Faults

Alluvial-fan conglomeratic strata typically accumulate on the downthrown sides of terrestrial normal faults that were active during sedimentation. Such coarse-grained deposits are present along much of the length of the border-fault system (Arguden and Rodolfo, 1986). Because only the youngest strata are usually exposed adjacent to the BFS of an asymmetric half graben, exposed conglomeratic strata are most prevalent in the Passaic, Feltville, Towaco, and Boonton Formations. It is not obvious whether the underlying older units (the downdip ends of the exposed Lockatong and Stockton Formations) in these areas also contain conglomerate. At a relay ramp where the Lockatong and Stockton Formations are in contact with the BFS (C in Fig. 1; Fig. 5), however, both formations are coarsely conglomeratic and indicate that a significant source of relief existed in the footwall during their deposition. These conglomeratic strata pass laterally into finer-grained facies. On the NB-1 seismic profile, reflectors lose their coherency adjacent to the BFS (Fig. 4), which may indicate

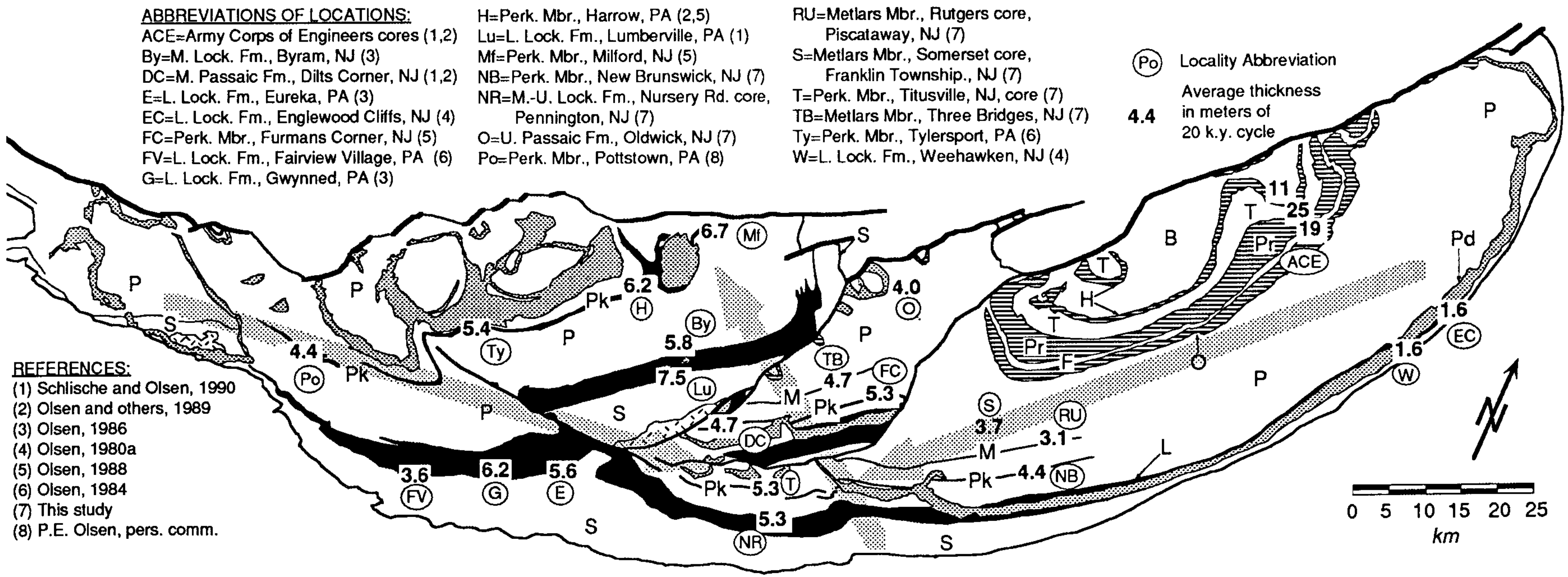
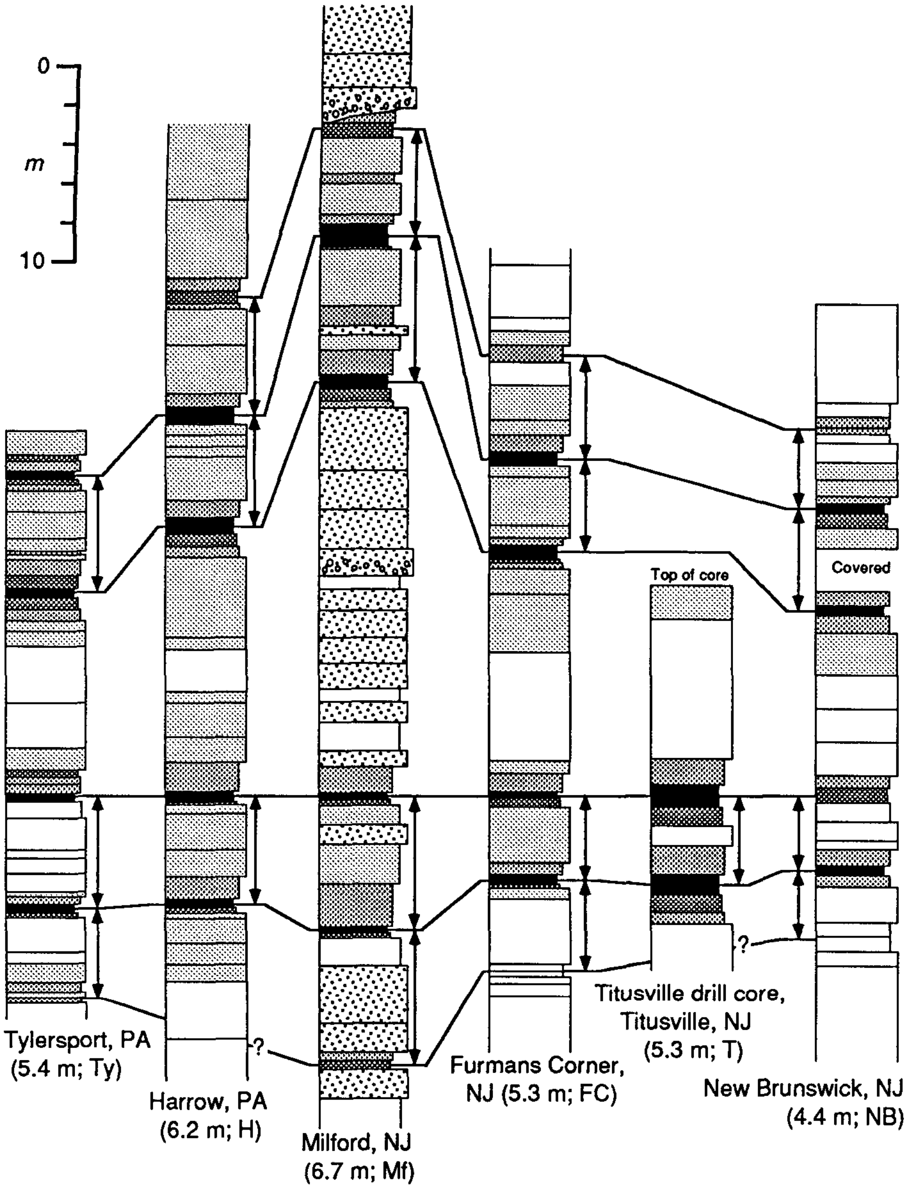


Figure 10. Map illustrating the variation in thickness of lacustrine cycles. Arrows indicate general directions of thickening within time-equivalent units. M is Metlars Member of Passaic Formation; other abbreviations for stratigraphic units as in Figure 1.



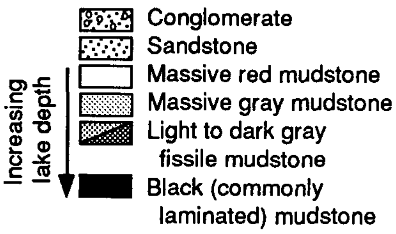


Figure 11. Generalized measured sections of the Perkasie Member of the Passaic Formation, illustrating variations in thickness and facies with position in the basin. Refer to Figure 10 for locations and references. Arrows delineate precession cycles; numbers in parentheses give average precession-cycle thickness.

the presence of alluvial-fan conglomerate. If correct, then alluvial-fan deposits would be present in the Passaic, Lockatong, and Stockton Formations at depth. Very coarse-grained facies have not been observed in association with the major intrabasinal faults.

Variations in Thickness and Facies

The thickness of fixed-period lacustrine cycles varies considerably throughout the Newark basin—both along strike and in a direction perpendicular to the BFS. The Perkasie Member of the Passaic Formation has been traced throughout the basin (Olsen, 1988) and crops out in every major fault block (Figs. 10 and 11). The average thicknesses of the 20,000-yr precession cycles are 4.4 m at its northeasternmost outcrop (NB in Figs. 10 and 11), 5.3 m in the hanging wall of the Hopewell fault in the Titusville drill core (T), 5.3 m in the central fault block (FC), 6.2 m and 6.7 m adjacent to the BFS near the Delaware River (H and Mf), 5.4 m at Tylersport (Ty), and 4.4 m near the southwest edge of the basin (Po). Cycles within the Metlars Member of the Passaic Formation thicken from 3.1 m in the Rutgers core (RU in Fig. 10) to 3.7 m in the Somerset core (S) to 4.7 m at Three Bridges (TB).

The Passaic Formation displays significant lateral changes in facies. In the central basin, the unit consists of mostly red cyclical lacustrine strata with some gray and black cyclical lacustrine strata (deposited under the deepest water). The frequency of gray and black units decreases to the northeast (Fig. 1). Fine-grained lacustrine rocks give way to coarser fluvial facies toward the northeast end of the basin and toward the narrow neck, where conglomerate is prevalent (Figs. 1 and 9).

Average cycle thicknesses within the lower Lockatong Formation exhibit considerable along-strike variability: 1.6 m along the hinged margin near the northeast end of the basin (W and EC in Fig. 10), 5.6 and 6.2 m at Eureka and Gwynned, Pennsylvania (E and G), and 3.6 m near the southwest edge of the basin (FV). Cycles within the middle Lockatong Formation thicken in the dip direction: 5.3 m in the Nursery Road drill core (NR in Fig. 10) and 5.8 m along the classic Delaware River exposures (By). The trends in cycle thickness are mimicked by the outcrop width of the Lockatong Formation (Fig. 10), which is greatest in the central basin and decreases to zero laterally. These changes cannot be explained by changes in dip, but the large map width south of the Chalfont fault is related to normal faulting (Fig. 7).

In the northeast part of the basin, lacustrine cycles in Englewood Cliffs (EC in Fig. 10) contain a lower percentage of strata inferred to have been deposited under deep water than at the more southern exposures in Weehawken (W), reflecting a shoaling of the basin toward its northern terminus (Olsen, 1980a). In fact, farther to the north, those attributes that make the Lockatong a unique formation—gray to black, often laminated to microlaminated mudstone—disappear completely, and time-equivalents of the Lockatong Formation are mapped as Stockton and Passaic Formations (Parker and others, 1988).

The basin-wide changes in thickness and facies recorded in lacustrine strata of the Lockatong and Passaic Formations clearly indicate that thickness and paleo-lake depth increase from the lateral edges toward the center of the basin and from the hinged margin toward the BFS and/or intrabasinal fault systems (Fig. 10). Similar variations in thickness and facies have been documented in the overlap sections of six cores recently obtained from the northeastern fault block of the Newark basin in New Jersey (Schlische and others, 1991). Details of the cored sections will be presented elsewhere.

Early Jurassic formations, cored by the Army Corps of Engineers (ACE in Fig. 10; Fedosh and Smoot, 1988), record markedly higher cycle thicknesses than any of the Triassic strata: 19 m for the Feltville Formation, 25 m for the Towaco Formation, and 11 m for the Boonton Formation (Olsen and others, 1989).

Distinctive lacustrine units have been extensively mapped and cycle thicknesses determined within the Jacksonwald syncline (Fig. 9). Precession cycles of the Metlars Member average 9.2 m in thickness in the hinge of the syncline and 7.6 m and 5.1 m on the northern and southern limbs, respectively. The deepest-water part of the cycles contains finely laminated claystone in the hinge; laminated claystone is absent, and sandstone is much more abundant on the southern limb, which is situated closer to the edge of the basin. The uppermost unit of the Passaic Formation, located only a few meters below the Jacksonwald Basalt and very close to the BFS as well as the hinge of the syncline, has an average cycle thickness of 10.7–13 m (Olsen and others, 1990). Furthermore, seismic reflection profiles from the southwestern Newark basin are interpreted to indicate that the Lockatong Formation is thicker in the hinges of synclines and thinner in the hinges of anticlines (Reynolds and others, 1990; Schlische and Reynolds, 1992). All evidence strongly suggests structural control of sedimentation.

Cycle thicknesses and facies vary considerably throughout the Newark basin, but care must be taken in comparing cycle thicknesses in different age units. Although one might expect cycle thicknesses to increase toward the BFS, that prediction is not always borne out. In the Delaware River region, the cycle thickness is 7.5 m in the lower Lockatong Formation (Lu in Fig. 10) and 5.8 m in the middle Lockatong Formation (By). In the Rutgers #1 core (RU), precession cycles are 4.4 m at the bottom (Perkasie Member) but only 3.1 m at the top (Metlars Member). Within the central fault block, cycles average 5.3 m in the Perkasie Member (FC in Fig. 10), 4.7 m in the Metlars Member (TB), and 4.0 m in the upper Passaic Formation (O). These changes reflect the evolutionary development of the basin—the interplay among tectonically controlled basin capacity, sediment supply rates, and climate (Schlische and Olsen, 1990)—and not the along-strike and transverse variations existing at any particular time.

The fluvial Stockton Formation lacks Milankovitch-period cycles. Nonetheless, the outcrop-based thickness of the entire formation varies from 1,800 m in the footwall of the Flemington-Furlong fault (Olsen, 1980b) to 900 m in the hanging wall of the Hopewell fault (Willard and others, 1959). The mapped width also decreases from the center of the basin to the northeast and southwest. On the NB-1 line (Fig. 3), the interpreted Stockton Formation thins away from the BFS.

Onlap Geometry

All along the unfaulted margins of the Newark basin, Mesozoic strata rest unconformably on Paleozoic and Precambrian "basement" rocks. The Stockton Formation onlaps "basement" rocks along northeastern and southern margins of the basin (McLaughlin, 1945; McLaughlin and Willard, 1949; Willard and others, 1959), with younger Stockton strata progressively onlapping structurally higher "basement" rocks laterally. At the northeastern termination of the basin, strata of the Passaic Formation rest directly on "basement" (Fig. 1). The Stockton Formation is interpreted to onlap the "basement" rocks of the ramped hanging-wall margin on the NB-1 line (Fig. 3). As the Stockton Formation was deposited by through-flowing streams, implying that the basin was filled to its lowest outlet (Schlische and Olsen, 1990), the onlap does not represent the progressive infilling of a pre-existing depression. Rather, the basin itself was growing both wider and longer (Fig. 12A).

TECTONOSTRATIGRAPHIC EVOLUTION

In this section, structural and stratigraphic observations are combined to examine the following: (1) timing of border and intrabasinal faulting relative to sedimentation, (2) timing of folding relative to sedimentation, (3) growth of the basin through time and its effect on the large-scale stratigraphic development, and (4) the overall history of the Newark basin.

Border Fault System

Faill (1973, 1988) argued that if border faulting were syndepositional, then one would expect to find a progressive decrease in dip in younger strata. Faill saw no such systematic variation in dip; in fact, he observed that the youngest strata commonly had the highest dips. He also found little correlation between conglomerate and the BFS, and he concluded that most border faults within the Newark basin postdated sedimentation. The observations presented in this paper suggest otherwise. In structurally uncomplicated areas, younger (unfolded) strata do dip less than older strata (Fig. 8). Progressive changes in dip in other areas are probably complicated by transverse and drag folding adjacent to the BFS. Conglomeratic strata are almost always present along the BFS and in all sedimentary formations. Perhaps the best evidence of syndepositional faulting is the increase in thickness of many units toward the BFS. Hence, beginning with Stockton deposition, strata accumulated in an asymmetric basin controlled by at least sporadic syndepositional border faulting.

In addition to thinning away from the BFS in a transverse direction, time-correlative units thin appreciably toward the lateral edges of the basin; taken to the extreme, the basin itself terminates along strike to the northeast. If these variations in thickness of units are a proxy for variations in basin subsidence, then subsidence was greatest near the center of the basin adjacent to the BFS and decreased away from this region in all directions.

The aforementioned variations in subsidence are similar to variations in displacement that occur in the volume of rock surrounding a blind normal fault (Barnett and others, 1987; Walsh and Watterson, 1987, 1988). A normal fault may be modeled as an approximately elliptical surface in which maximum displacement is located at the center of the fault and decreases to zero at the tip-line loop. Displacement also decreases

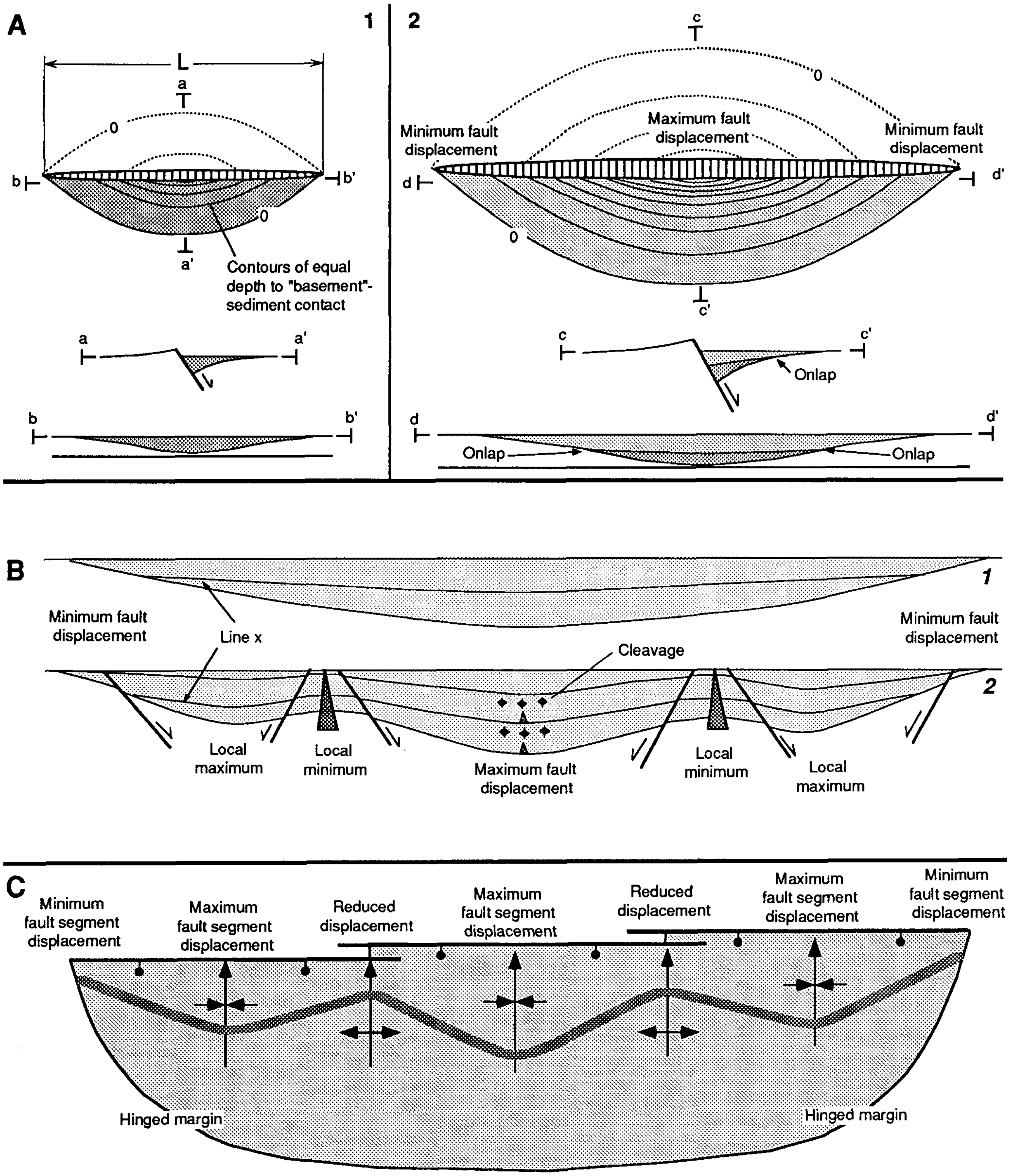


Figure 12. (A) Variations in subsidence in a half-graben basin bounded by a normal fault on which displacement is greatest at the center of the fault and decreases laterally, shown at two different stages. Amount of hanging-wall subsidence (solid) and footwall uplift (dotted) shown by contour lines. Basin grows in depth, width, and length through time as maximum displacement on the fault accumulates. Synrift stratal units (stippled) onlap "basement" rocks transversely and laterally. Modified from Schlische (1991). (B) Formation of transverse folds by along-strike variations in fault displacement. Both cross sections are drawn parallel to BFS. Length of line x must increase as a result of a variable displacement on the border fault. Also shown is the possible distribution of structures associated with fault-displacement folds; dark stipple represents diabase intrusions. (C) Map-view relationship between border-fault segmentation and transverse folds. The position of the hinged margin is unaffected by the segmentation and seems to reflect the variations in displacement on the entire BFS.

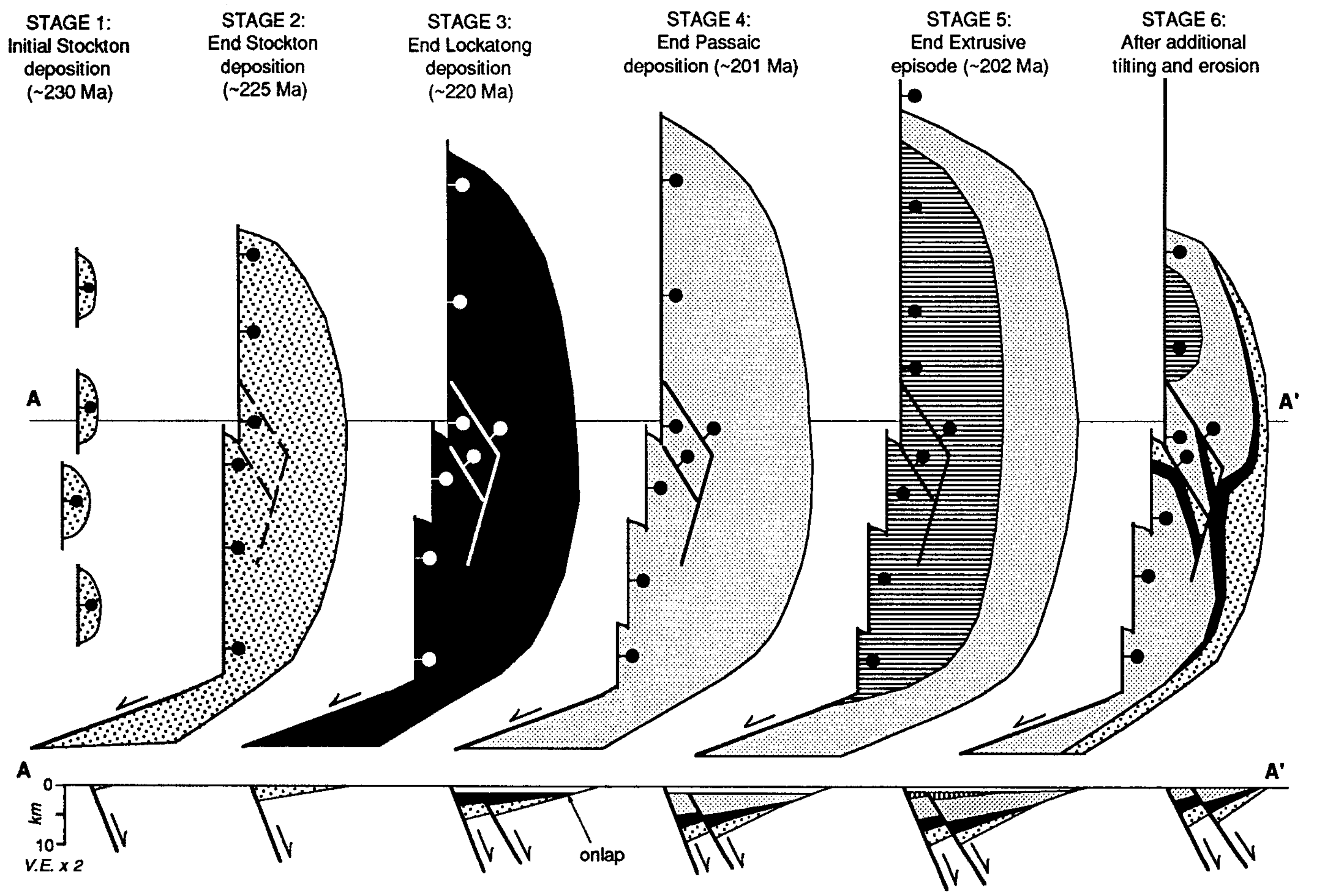


Figure 13. Tectonostratigraphic evolution of the Newark basin. Note the early coalescence of a number of isolated sub-basins, the increasing length and width of the depositional basin until the Early Jurassic extrusive interval, and the progressive footwall incisement of the border fault in the southeastern half of the basin. The units depicted (for example, the Lockatong Formation) are chronostratigraphic units and have no specific lithostratigraphic or facies significance. Ages derived from Schlische and Olsen (1990).

in a direction normal to the fault surface, producing a typical "reverse drag" profile, even on planar faults. Slight modification of this model is required for synsedimentary faults (Gibson and others, 1989), in which maximum displacement occurs where the center of the fault intersects the Earth's free surface. Based on the observed variations in stratal thicknesses and by analogy with the normal fault model, the BFS of the Newark basin experienced greatest fault displacement and resulting basin subsidence near the center of the BFS; both displacement and subsidence decreased toward the lateral ends and hinged margins of the basin (Fig. 12A).

The scenario presented above is complicated by two factors. (1) The BFS does not consist of one continuous fault but a number of fault segments. Nonetheless, the hinged margin of the basin (defined by the Stockton-"basement" contact) defines a broad syncline plunging toward the BFS; its geometry does not appear to have been influenced by any segmentation of the BFS but rather the large-scale variations in displacement on the entire BFS (Fig. 12C). (2) Basin subsidence was also influenced by the dip angle of the fault. Thus, the northeast part of the basin, although situated at the along-strike end of the basin, experienced appreciable subsidence because of the steeper dip and hence greater throw of the BFS.

The synclinal downwarping as a result of a variable displacement field on the BFS may have created bending stresses within the basin, particularly in its lower, convex-downward reaches (Fig. 12A). This may have established localized northeast-southwest extension, leading to the formation of the northwest-trending dikes and extensional joints within the basin. Alternatively, there may have been two mutually perpendicular directions of extension operating simultaneously.

According to the fault reactivation model of Ratcliffe and Burton (1985), a regional east-southeast extension direction, perpendicular to the average strike of dikes, accounts for the dip-slip motion observed on most northeast-striking border faults and the left-lateral slip on east-striking faults. The resulting variations in throw produced variations in subsidence: based on outcrop width (7 km) and an average dip of 20°, <3 km of strata is present in the narrowest part of the narrow neck, where strike slip predominated, in contrast to the >6 km present in the central Newark basin (Fig. 3).

Intrabasinal Faults

On the basis of outcrop studies alone, the intrabasinal faults appear to postdate the deposition of strata preserved in their hanging walls because none of the preserved strata (Stockton, Lockatong, Passaic, and Feltville

Formations) is known to contain any very coarse facies, and any thickening toward the intrabasinal faults may also be explained by the known thickening toward the BFS (Faill, 1988; Schlische and Olsen, 1988; Olsen and others, 1989; Schlische, 1990). Recent subsurface data, however, suggest otherwise. The percentage of thickening between overlap sections of the upper Lockatong Formation from the top of the Nursery Road core (NR in Fig. 10) to the bottom of the Titusville core (T) is 4.8% over a distance of 3.4 km in the dip direction. When this rate of thickening is linearly extrapolated from the Nursery Road site to the outcrops of the middle Lockatong Formation along the Delaware River (By), it predicts an average cycle thickness of 7.1 m, not the observed 5.8 m (By in Fig. 10). Either the rate of thickening decreased toward the BFS, which is unlikely in an asymmetric half graben, or rates of thickening were independent in each fault block, requiring that at least one of the intrabasinal normal faults was active during the deposition of the Lockatong Formation (Schlische and others, 1991). On the basis of seismic data, the Chalfont fault is interpreted to have been syndepositionally active (Reynolds and others, 1990). Although different fault blocks were subsiding independently and may have contained separate lakes, strata can be correlated across faults because the lacustrine cycles responded to the same climatic forcing.

The increasing density of intrabasinal faulting from northeast to southwest is apparently related to the decreasing dip of the BFS in the same direction (see Fig. 3). A larger number of intrabasinal faults may have accommodated the extension in areas where the original dip of the BFS was shallower. The BFS probably has also undergone some flexural rotation to shallower dip angles as a result of isostatic unloading of the footwall block (*sensu* Buck, 1988; Wernicke and Axen, 1988), which should have been greater in areas of higher fault displacement (for example, near the center of the basin).

Timing and Origin of Folding

The present body of evidence strongly suggests that at least some of the transverse folding in the Newark basin was synsedimentary. Lacustrine-cycle thicknesses are greater in the hinge of the Jacksonwald syncline than on the limbs (Fig. 9), and seismic lines are interpreted to reveal greater thicknesses of units in the hinges of synclines than on the fold limbs and in the hinges of anticlines; this is to be expected if folding and sedimentation were coeval. Structural thickening in fold hinges, when it occurs, should be observed in both synclines and anticlines. Some minor thrust faults are present within the hinge of the Jacksonwald syncline, but they account for a thickening of only a few millimeters.

The strained tetrapod footprints and desiccation cracks, at least in some areas, may have been ductilely deformed, implying that fold-related deformation possibly occurred shortly after sedimentation but prior to significant dewatering and/or lithification. The phacoliths within the hinge of the Jacksonwald syncline (Fig. 9) imply that some folding occurred during or before intrusion (Manspeizer, 1988; Schlische and Olsen, 1988). The relative conformity of diabase plutons in synclines and disconformity in the hinges of anticlines and the distribution of diabase types at various structural levels within the folds imply the presence of folds during intrusion.

Paleomagnetic work further defines the timing of the folding in the Jacksonwald syncline. Red beds contain two components of magnetization (Witte and others, 1991; Witte and Kent, 1991). The first component was acquired shortly after deposition and before significant folding, whereas a secondary overprint, dated at ca. 175 Ma, was completely post-folding and requires that deformation ceased prior to the onset of sea-floor spreading at 175 Ma. Paleomagnetic data from the Jacksonwald sill also support some folding prior to intrusion (Stuck and others, 1988).

The time of formation of transverse folds adjacent to dip-slip intrabasinal faults is unknown. Because it is likely, however, that at least one of these faults formed syndepositionally, then the folds may have formed at the same time.

The formation of the transverse folds in the Newark basin has been attributed to post-depositional regional shortening (Sanders, 1963), strike slip along the BFS (Manspeizer, 1980), the operation of a regional sinistral shear zone along the future Kelvin-Cornall fracture zone (Manspeizer, 1981), and a northeast-southwest–oriented compressive σ_2 during basin formation (Ratcliffe and Burton, 1985). These four models can be discounted because (1) some folds formed syndepositionally; (2) most folds are oriented nearly perpendicular to the associated faults, not the 45° or less that is required if the folds were related to strike slip on the faults (Christie-Blick and Biddle, 1985); (3) the associated faults experienced predominantly dip slip; and (4) northwest-striking, dike-like intrusions (for example, bd in Fig. 9) indicate northeast-southwest extension rather than shortening during "folding." Schlische and Olsen (1988) postulated that the folds formed in the concave-upward portion of a synformally downwarped basin, but this model requires external, and unaccounted for, lateral shortening.

Wheeler (1939) proposed that the transverse folds resulted from differential displacement along normal faults as a result of irregularities in the fault surface, which he called "fault-line deflections": synclines formed where the fault was convex toward the footwall, and anticlines formed where the fault was convex toward the hanging wall; the decrease in amplitude of the folding away from the fault would be a consequence of an "evening out" effect with distance from the fault-line deflections. Seismic reflection profiles, however, indicate that folds in the subsurface are not associated with any undulations along the fault surface (D. J. Reynolds, 1991, personal commun.).

As discussed above, normal faults display along-strike variations in displacement. The folds, therefore, may have formed as a result of variable amounts of slip along the associated faults: synclines formed in areas of locally higher fault slip, and anticlines formed in areas of locally lower fault slip (Fig. 12B). The segmented nature of the BFS in the Newark basin, and of normal faults in general (Schwartz and Coppersmith, 1984; Jackson and White, 1989; Peacock and Sanderson, 1991), makes this mechanism even more plausible (Fig. 12C). In the southwest part of the basin (Fig. 9), synclinal axes are located at the approximate centers of fault segments, where displacements are expected to be highest; anticlinal axes are situated where fault segments overlap or terminate, where displacements are expected to be lowest. Not all folds, however, appear to be related to fault segmentation. For example, the transverse folds along the Ramapo fault are not associated with any obvious segmentation, but the fault may be imprecisely mapped. Alternatively, second-order variations in fault displacement on longer fault segments may have produced the folds (Fig. 12B).

If the folds did result from along-strike variations in fault displacement, then the folds are only apparent shortening structures, and bed lengths must have increased over their pre-warping dimensions (Fig. 12B). Thus, some fold-axis parallel extensional structures (generally oriented northwest-southeast) are required. The dike-like diabase intrusions that are present in the hinges of the anticlines (Fig. 9) may have accommodated some of this northwest-southeast extension, but this model does not explain the formation of an axial planar cleavage (a result of true shortening) late in the history of folding. One possibility is that the cleavage is restricted to the inner arcs of clusters of beds in the synclines that buckled independently during warping and experienced localized shortening, with requisite extensional structures on the outer arcs (Fig. 12B). More data on the distribution of the cleavage and extensional structures are required to test this hypothesis.

Transverse folds are not unique to the Newark basin and are present within the early Mesozoic Fundy, Deerfield, Hartford, Gettysburg, Culpeper, Taylorsville, and Deep River basins of eastern North America (Schlische, 1990); the Mesozoic rift basins of eastern India (Withjack and Gallagher, 1983); the Railroad Valley basin of the Basin and Range of Nevada (Vreeland and Berrong, 1979); and the East African rift system (D. J. Reynolds, 1990, personal commun.). Transverse bedrock ridges or "basement" highs adjacent to fault-segment boundaries and valley lows associated with segment centers in the Basin and Range (Crone and Haller, 1991; Zhang and others, 1991) appear to be additional analogues.

Basin Growth and Effect on Stratigraphy

Onlap geometries indicate that the basin was growing wider and longer through time. This is a natural consequence of the accumulation of displacement on normal faults. In general, shorter length (the along-strike dimension) faults exhibit less maximum displacement than larger length faults; this is interpreted to reflect a growth sequence, with fault length increasing as displacement accumulates (Watterson, 1986; Walsh and Watterson, 1987, 1988; Cowie and Scholz, 1990). Gibson and others (1989) and Schlische (1991) extended these models to fault-bounded basins and concluded that basins ought to grow deeper, wider, and longer through time (Fig. 12A). When these model sedimentary basins are infilled, strata progressively onlap "basement" rocks of the hanging-wall block (Fig. 12A).

Given the segmented nature of the BFS and fault growth, it is possible that during the earliest stages of basin evolution, what was to become the Newark basin consisted of a number of border-fault segments with isolated depositional basins (Fig. 13). These basins must have quickly coalesced, because the Stockton Formation now exposed at the surface forms a continuous outcrop belt along the hinged margin of the basin. Bear in mind that onlap relationships require that the part of the Stockton Formation in contact with "basement" rocks along the hinged margin is younger than the oldest (buried) deposits.

Footwall-incised rider blocks (Figs. 3 and 5) also indicate that the basin widened through time. Footwall-propagating faults may have formed by gravitational collapse (Gibbs, 1984) to relieve excessive isostatically driven footwall topography (Schlische, 1990); this process may have been aided by the pervasive southeast-dipping "grain" of the footwall rocks inherited from the Appalachian orogenies.

Because the basin was deepening, widening, and lengthening through time, the change in the rate of increase in the volume of capacity of the basin was positive; that is, the incremental capacity of the basin increased through time (Schlische, 1991). This had a profound effect on the large-scale stratigraphy of the Newark basin, as originally discussed by Schlische and Olsen (1990). The through-flowing streams of the Stockton Formation were deposited in a basin in which incremental capacity was overwhelmed by the available volume of sediment such that excess sediment and water flowed out of the basin. The fluvial-lacustrine transition (Stockton-Lockatong) occurred when the incremental capacity of the basin exceeded the available volume of sediment; a lake then occupied the excess capacity of the basin. The decreasing lacustrine cycle thicknesses throughout the Triassic may represent a basin in which incremental capacity increased faster than the rate at which sediment was supplied to the basin such that sediment was distributed over a progressively larger depositional surface area. Alternatively, with incremental capacity constant or decreasing, the volume of sediment supplied to the basin may have decreased, perhaps as a result of increasing aridity through the close of the Triassic. Paleo-lakes during even the wettest of the Milankovitch cycles within the Passaic Formation appear to have been shoaling during this time, but this may also be a consequence of water distributed over a progressively larger volume due to basin growth.

The trend toward thinner and "drier" lacustrine cycles in younger strata is abruptly terminated in the earliest Jurassic strata underlying and interbedded with the lava flows. An increase in the extension rate may have asymmetrically deepened the basin, causing sediment and water to preferentially accumulate adjacent to the BFS, thereby increasing both cycle thickness and paleo-lake depth. The minimum heave (extension) on the BFS for the Newark basin (see section A–A′, Fig. 3) was ~2.8 km during the ~20 m.y. of Passaic deposition (based on cyclostratigraphy; Schlische and Olsen, 1990) but ~0.8 km during the ~0.6 m.y. of accumulation of the Early Jurassic lava flows and interbedded strata. Alternatively, a wetter Jurassic climate may have brought in more water and sediments to the basin, increasing both cycle thickness and paleo-lake depth. For this hypothesis to work, the actual extension rate during deposition of the Passaic Formation needs to have been greater than the minimum value reported above, and the basin needs to have become progressively more sediment starved through the Triassic in order to accommodate the deeper Jurassic lakes. Cycles within the post-extrusive Boonton Formation are thinner and accumulated in shallower water; the minimum extension rate was ~0.9 km/2 m.y.

History of the Basin

On the basis of the age of the oldest biostratigraphically dated strata currently exposed (Cornet and Olsen, 1985), extension began at the latest in late Carnian time and resulted in the brittle reactivation of several Paleozoic thrust faults as normal faults, creating several, possibly sedimentologically isolated, sub-basins in which fluvial deposits accumulated. The slipped areas of the individual faults quickly increased, and the sub-basins merged to form the Newark basin. As a result of footwall collapse indicated by rider blocks and as displacement on the border fault system increased, the basin grew in width, length, and depth through time (Fig. 13), resulting in the transition from fluvial deposition (Stockton Formation) to lacustrine deposition (Lockatong Formation). Hanging-wall onlap also indicates that the basin widened and lengthened through time. Variations in thickness of a given unit reflect the syndepositional activity and along-strike variations in displacement of the intrabasinal and BFS faults. The Lockatong and Passaic Formations in southeastern Pennsylvania record evidence of syndepositional folding, also in response to along-strike variations in fault displacement.

In latest Triassic time, the Newark basin may have experienced a vastly higher extension rate. This may have resulted in sufficiently rapid thinning of the crust to generate adiabatic melting in the mantle and result in extrusive and intrusive igneous activity in the basin 40,000 yr after the start of Jurassic time (Olsen and others, 1990). Magma intruded the hinge of already folded strata of the Jacksonwald syncline, forming phacoliths; at the same time, the Jacksonwald sill intruded partially folded strata, and the Jacksonwald Basalt flowed over a nearly flat surface. Igneous activity lasted only 600,000 yr (Olsen and Fedosh, 1988; Olsen and others, 1989), and the extension rate again decreased during post-extrusive Boonton Formation deposition.

Faulting, folding, and sedimentation probably continued until ~175

Ma, when continental rifting gave way to tectonically quiescent drifting (Klitgord and Schouten, 1986). Also at this time, a major hydrothermal event reset isotopic clocks (Sutter, 1988) and remagnetized red beds (Witte and others, 1991; Witte and Kent, 1991). This hydrothermal event may have been associated with the magmatism responsible for the emplacement of seaward-dipping extrusions at the newly formed continent-ocean boundary (Sheridan and others, 1991). No additional tilting or folding occurred after ~175 Ma (Witte and Kent, 1991). The Newark basin was then eroded from Middle Jurassic to Early Cretaceous time. Thermal maturity studies (Pratt and Burruss, 1988) and fission track analysis (Kohn and others, 1988) revealed that ~2 km of post–Early Jurassic strata were removed by erosion. During the Early Cretaceous period, high sea levels and/or flexural loading of the passive margin caused coastal-plain sediments to overlap the Newark basin, effectively blanketing the basin and preventing further erosion. Much of this coastal-plain cover was then stripped during the Pleistocene epoch.

DISCUSSION

Along-strike variations in fault displacement have been invoked in this paper to account for both the transverse folds and the overall plunging synclinal geometry of the basin. Enigmatically, the transverse folds appear to have been influenced by segmentation of the border fault system, whereas the larger scale basin geometry—represented by the position of the Stockton-"basement" contact—was not (Fig. 12C). One possibility is that some older BFS was mainly responsible for the larger basin geometry and that a younger BFS, which progressively evolved as a result of footwall incisement, most influenced the transverse folds. Alternatively, the transverse folds may reflect displacement variations on the nearest down-plunge fault segment, whereas the position of the hinged margin may reflect the sum of all fault displacements, including buried and intrabasinal faults. Anders (1991) noted that although many "basement" highs in the Basin and Range are associated with segment boundaries, the amount of footwall uplift is commonly unaffected across these boundaries. He proposed that "basement" highs form in areas of fault splaying so that displacement on any one fault and total basin depth are reduced; in contrast, footwall uplift reflects the sum of displacements on all faults, which presumably merge at depth.

Longitudinal hanging-wall onlap relations in the Newark basin and published fault-length–displacement relationships for dip-slip faults indicate that the basin and its BFS increased somewhat in length through time. If individual fault segments also increase in length through time, then the ends of faults (segment boundaries) may migrate. In the East African rift system, early isolated half graben within what is now Lake Malawi eventually merged as the tips of the boundary faults propagated laterally (Ebinger, 1989). Accommodation zones developed in areas where the propagating fault tips overlapped, and sediment accumulations are highest at the centers of border-fault segments (Ebinger, 1989).

Hanging-wall onlap and footwall-incised rider blocks indicate that the Newark basin lengthened and widened through time. Transverse onlap geometries have also been observed in the following half graben: Fundy, Richmond, Atlantis, Long Island, and Nantucket basins of eastern North America; Hopedale and Saglek of the Labrador margin; North Viking of the North Sea; Tanganyika of East Africa; Dixie Valley, Northern Fallon, Diamond Valley, Railroad Valley, and the Great Salt Lake basin of the Basin and Range (see Schlische and Olsen, 1990, for references). Basin lengthening, widening, and deepening imply that the incremental basin capacity should increase through time unless fault-displacement rates decrease (Schlische, 1991). This increase in incremental capacity predicts that any initial fluvial deposits (sediment supply exceeds incremental capacity) should be overlain by lacustrine deposits (incremental capacity exceeds sediment supply). Initial fluvial deposits succeeded by younger lacustrine strata are present in the following terrestrial extensional basins: Keweenawan of the North American midcontinent; Morondava of Madagascar; Mombasa of Kenya; Deep River, Dan River, Richmond, Culpeper, Gettysburg, Hartford/Deerfield, and Fundy of eastern North America; Sudan; and West African of Gabon and Angola (see Schlische and Olsen, 1990, and Lambiase, 1991, for references).

SUMMARY AND CONCLUSIONS

1. The Newark basin is a half graben, bordered on its northwestern margin by a series of normal faults, many of which have reactivated Paleozoic thrust faults. The border-fault system has a relay geometry and contains a number of rider blocks in which younger faults propagated into the footwall, possibly due to gravitational collapse of the isostatically uplifted footwall.
2. Strata of all ages thicken toward the BFS, where they contain conglomeratic facies, indicating that the border faults were active during sedimentation. In some structurally uncomplicated regions, younger strata dip less steeply than older strata, indicating syndepositional rotation of the hanging wall. Lacustrine strata thicken and were deposited under deeper water conditions from the lateral edges toward the center of the basin, indicating that the basin generally subsided more at its center than at its ends, reflecting along-strike variations in displacement on the entire BFS.
3. Some intrabasinal faults were at least sporadically active during the deposition of the units currently preserved in their hanging walls. Most intrabasinal normal faults are synthetic to the BFS.
4. The Chalfont fault is a left-lateral, intrabasinal transfer fault that separates regions of differential extension and dies out where the differential extension is zero. Variations in the density of intrabasinal faults are attributed to variations in the dip angle of the BFS, which generally decreases to the southwest.
5. Transverse folds, some of which formed syndepositionally, characterize the hanging walls of the border and intrabasinal normal faults and likely formed as a result of along-strike variations in fault displacement; border-fault segmentation apparently controlled the location of some of the folds.
6. Younger strata progressively onlap structurally higher "basement" rocks on the hanging wall and indicate that the basin grew in width and length through time; the BFS similarly increased in length.
7. As the basin deepened, widened, and lengthened, initial fluvial deposition gave way to lacustrine sedimentation in a sediment-starved basin.

ACKNOWLEDGMENTS

I thank P. Olsen, M. Anders, D. Reynolds, W. Burton, and M. Withjack for valuable discussions; P. Olsen and D. Reynolds for sharing unpublished data with me; M. Anders, N. Christie-Blick, J. Husch, M. Levy, P. Olsen, R. Shagam, A. Sylvester, D. Wise, and M. Withjack for critical reviews; and M. Levy, M. Angell, R. Hayden, S. Fowell, and L. Schlische for assistance in the field. Research was supported by the Donors of the American Chemical Society, administered by the Petroleum Research Fund (ACS 19878G2 to P. Olsen); the Nuclear Regulatory Commission (NRC-04-85-111-02 to L. Seeber and P. Olsen); the National Science Foundation (EAR-9017785); a Henry Rutgers Research Fellowship; and Grants-in-Aid-of-Research from Sigma Xi and the Geological Society of America.

REFERENCES CITED

Anders, M. H., 1991, Are normal fault rupture segments persistent?: Geological Society of America Abstracts with Programs, v. 23, p. A432.

Arguden, A. T., and Rodolfo, K. S., 1986, Sedimentary facies and tectonic implications of lower Mesozoic alluvial-fan conglomerates of the Newark basin, northeastern United States: Sedimentary Geology, v. 51, p. 97–118.

Bally, A. W., Withjack, M. O., Meisling, K. E., and Fisher, D. A., 1990, Seismic expression of structural styles: Geological Society of America Short Course Manual.

Barnett, J.A.M., Mortimer, J., Rippon, J. H., Walsh, J. J., and Watterson, J., 1987, Displacement geometry in the volume containing a single normal fault: American Association of Petroleum Geologists Bulletin, v. 71, p. 925–937.

Bell, R. E., Karner, G. D., and Steckler, M. S., 1988, Detachments during extension: Application to the Newark series basins: Tectonics, v. 7, p. 447–462.

Berg, T. M., compiler, 1980, Geologic map of Pennsylvania: Pennsylvania Topographic and Geologic Survey, scale 1:250,000.

Buck, W. R., 1988, Flexural rotation of normal faults: Tectonics, v. 7, p. 959–973.

Christie-Blick, N., and Biddle, K. T., 1985, Deformation and basin formation along strike-slip faults, *in* Biddle, K. T., and Christie-Blick, N., eds., Strike-slip deformation, basin formation, and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 37, p. 1–34.

Cornet, B., and Olsen, P. E., 1985, A summary of the biostratigraphy of the Newark Supergroup of eastern North America, with comments on early Mesozoic provinciality, *in* Weber, R., ed., Congresso Latinoamericano de Paleontologia México, III, Simposio Sobre Floras del Triasico Tardio, su Fitografia y Paleoecologia, Memoria, p. 67–81.

Costain, J. K., and Çoruh, C., 1989, Tectonic setting of Triassic half grabens in the Appalachians: Seismic data, acquisition, processing and results, *in* Tankard, A. J., and Balkwill, H. R., eds., Extensional tectonics and stratigraphy of the North Atlantic margins: American Association of Petroleum Geologists Memoir 46, p. 155–174.

Cowie, P. A., and Scholz, C. H., 1990, Fault growth and fault termination [abs.]: Eos (American Geophysical Union Transactions), v. 71, p. 631.

Crone, A. J., and Haller, K. M., 1991, Segmentation and coseismic behavior of Basin and Range normal faults: Examples from east-central Idaho and southwestern Montana, U.S.A.: Journal of Structural Geology, v. 13, p. 151–164.

Dunning, G. R., and Hodych, J. D., 1990, U-Pb zircon and baddeleyite age for the Palisade and Gettysburg sills of northeast United States: Implications for the age of the Triassic-Jurassic boundary: Geology, v. 18, p. 795–798.

Ebinger, C. J., 1989, Tectonic development of the western branch of the East African rift system: Geological Society of America Bulletin, v. 101, p. 885–903.

Faill, R. T., 1973, Tectonic development of the Triassic Newark-Gettysburg basin in Pennsylvania: Geological Society of America Bulletin, v. 84, p. 725–740.

——— 1988, Mesozoic tectonics of the Newark basin, as viewed from the Delaware River, *in* Husch, J. M., and Hozik, M. J., eds., Geology of the central Newark basin, field guide and proceedings: Meeting of the Geological Association of New Jersey, 5th, Rider College, Lawrenceville, New Jersey, p. 19–41.

Fedosh, M. S., and Smoot, J. P., 1988, A cored stratigraphic section through the northern Newark basin, New Jersey, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins of the eastern United States: U.S. Geological Survey Bulletin 1776, p. 19–24.

Gibbs, A. D., 1984, Structural evolution of extensional basin margins: Geological Society of London Journal, v. 141, p. 609–620.

Gibson, J. R., Walsh, J. J., and Watterson, J., 1989, Modelling of bed contours and cross-sections adjacent to planar normal faults: Journal of Structural Geology, v. 11, p. 317–328.

Hozik, M. J., 1985, Evidence for dominance of normal dip-slip motion on segment of Flemington fault in Newark basin of New Jersey [abs.]: American Association of Petroleum Geologists Bulletin, v. 69, p. 1438.

Hutchinson, D. R., and Klitgord, K. D., 1988, Evolution of rift basins on the continental margin off southern New England, *in* Manspeizer, W., ed., Triassic-Jurassic rifting, continental breakup, and the origin of the Atlantic Ocean and passive margins: Amsterdam, The Netherlands, Elsevier, p. 81–98.

Jackson, J. A., and White, N. J., 1989, Normal faulting in the upper continental crust: Observations from regions of active extension: Journal of Structural Geology, v. 11, p. 15–36.

Klitgord, K. D., and Schouten, H., 1986, Plate kinematics of the central Atlantic, *in* Vogt, P. R., and Tucholke, B. E., eds., The western North Atlantic region: Boulder, Colorado, Geological Society of America, The Geology of North America, Volume M, p. 351–378.

Kohn, B. P., Lutz, T. M., Wagner, M. E., and Organist, G., 1988, Anomalous thermal regime during Mesozoic rifting, central Appalachian Piedmont [abs.]: Eos (American Geophysical Union Transactions), v. 69, p. 487.

Lambiase, J. J., 1991, A model for tectonic control of lacustrine stratigraphic sequences in continental rift basins, *in* Katz, B., ed., Lacustrine exploration: Case studies and modern analogues: American Association of Petroleum Geologists Memoir 50, p. 265–276.

Larsen, P.-H., 1988, Relay structures in a Lower Permian basement-involved extension system, East Greenland: Journal of Structural Geology, v. 10, p. 3–8.

Lucas, M., Hull, J., and Manspeizer, W., 1988, A foreland-type fold and related structures of the Newark rift basin, *in* Manspeizer, W., ed., Triassic-Jurassic rifting, continental breakup, and the origin of the Atlantic Ocean and passive margins: Amsterdam, The Netherlands, Elsevier, p. 307–332.

Lyttle, P. T., and Epstein, J. B., 1987, Geologic map of the Newark 1° × 2° quadrangle, New Jersey, Pennsylvania and New York: U.S. Geological Survey Miscellaneous Investigations Series Map I-1715, scale 1:250,000.

Manspeizer, W., 1980, Rift tectonics inferred from volcanic and clastic structures, *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: Annual Meeting of the New York State Geological Association, 52nd, Rutgers University, Newark, New Jersey, p. 314–350.

Manspeizer, W., 1981, Early Mesozoic basins of the central Atlantic passive margin, *in* Bally, A. W., ed., Geology of passive continental margins: History, structure and sedimentologic record (with special emphasis on the Atlantic margin): American Association of Petroleum Geologists Education Course Note Series 19, p. 4-1 to 4-60.

Manspeizer, W., 1988, Triassic-Jurassic rifting and opening of the Atlantic: An overview, *in* Manspeizer, W., ed., Triassic-Jurassic rifting, continental breakup, and the origin of the Atlantic Ocean and passive margins: Amsterdam, The Netherlands, Elsevier, p. 41–79.

McLaughlin, D. B., 1945, Type sections of the Stockton and Lockatong Formations, *in* Pennsylvania Academy of Sciences, Proceedings, v. 19, p. 102–113.

McLaughlin, D. B., and Willard, B., 1949, Triassic facies in the Delaware Valley: Pennsylvania Academy of Sciences, Proceedings, v. 23, p. 34–44.

Miller, J. D., and Kent, D. V., 1986, Paleomagnetism of the Upper Devonian Catskill Formation from the southern limb of the Pennsylvania salient: Possible evidence of oroclinal rotation: Geophysical Research Letters, v. 13, p. 1173–1176.

Olsen, P. E., 1980a, Fossil great lakes of the Newark Supergroup in New Jersey, *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: Annual Meeting of the New York State Geological Association, 52nd, Rutgers University, Newark, New Jersey, p. 352–398.

Olsen, P. E., 1980b, The latest Triassic and Early Jurassic formations of the Newark basin (eastern North America, Newark Supergroup): Stratigraphy, structure, and correlation: New Jersey Academy of Science Bulletin, v. 25, p. 25–51.

Olsen, P. E., 1984, Comparative paleolimnology of the Newark Supergroup: A study in ecosystem evolution [Ph.D. thesis]: New Haven, Connecticut, Yale University, 726 p.

Olsen, P. E., 1986, A 40-million-year lake record of early Mesozoic climatic forcing: Science, v. 234, p. 842–848.

Olsen, P. E., 1988, Continuity of strata in the Newark and Hartford basins of the Newark Supergroup, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins in the eastern United States: U.S. Geological Survey Bulletin 1776, p. 6–18.

Olsen, P. E., and Fedosh, M. S., 1988, Duration of the early Mesozoic extrusive igneous episode in eastern North America determined by use of Milankovitch-type lake cycles: Geological Society of America Abstracts with Programs, v. 20, p. 59.

Olsen, P. E., Schlische, R. W., and Gore, P.J.W., eds., 1989, Tectonic, depositional, and paleoecological history of early Mesozoic rift basins, eastern North America: International Geological Congress Field Trip T-351 Guidebook, Washington, D.C., American Geophysical Union, 174 p.

Olsen, P. E., Fowell, S. J., and Cornet, B., 1990, The Triassic-Jurassic boundary in continental rocks of eastern North America: A progress report, *in* Sharpton, V. L., and Ward, P. D., eds., Global catastrophes in Earth history: An interdisciplinary conference on impacts, volcanism, and mass mortality: Geological Society of America Special Paper 247, p. 585–593.

Parker, R. A., Houghton, H. F., and McDowell, R. C., 1988, Stratigraphic framework and distribution of early Mesozoic rocks of the northern Newark basin, New Jersey and New York, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins of the eastern United States: U.S. Geological Survey Bulletin 1776, p. 31–39.

Peacock, D.C.P., and Sanderson, D. J., 1991, Displacements, segment linkage and relay ramps in normal fault zones: Journal of Structural Geology, v. 13, p. 721–733.

Pratt, L. M., and Burruss, R. C., 1988, Evidence for petroleum generation and migration in the Hartford and Newark basins, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins of the eastern United States: U.S. Geological Survey Bulletin 1776, p. 74–79.

Puffer, J. H., and Philpotts, A. R., 1988, Eastern North American quartz tholeiites: Geochemistry and petrology, *in* Manspeizer, W., ed., Triassic-Jurassic rifting, continental breakup, and the origin of the Atlantic Ocean and passive margins: Amsterdam, The Netherlands, Elsevier, p. 579–605.

Ratcliffe, N. M., 1971, The Ramapo fault system in New York and adjacent New Jersey: A case of tectonic heredity: Geological Society of America Bulletin, v. 82, p. 125–141.

Ratcliffe, N. M., 1980, Brittle faults (Ramapo fault) and phyllonitic ductile shear zones in the basement rocks of the Ramapo seismic zones, New York and New Jersey, and their relationship to current seismicity, *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: Annual Meeting of the New York State Geological Association, 52nd, Rutgers University, Newark, New Jersey, p. 278–311.

Ratcliffe, N. M., 1988, Reinterpretation of the relationships of the western extension of the Palisades sill to the lava flows at Ladentown, New York, based on new core data, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins in the eastern United States: U.S. Geological Survey Bulletin 1776, p. 113–135.

Ratcliffe, N. M., and Burton, W. C., 1985, Fault reactivation models for the origin of the Newark basin and studies related to U.S. eastern seismicity: U.S. Geological Survey Circular 946, p. 36–45.

Ratcliffe, N. M., and Burton, W. C., 1988, Structural analysis of the Furlong fault and the relationship of mineralization to faulting and diabase intrusion, Newark basin, Pennsylvania, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins in the eastern United States: U.S. Geological Survey Bulletin 1776, p. 176–193.

Ratcliffe, N. M., Burton, W. C., D'Angelo, R. M., and Costain, J. K., 1986, Low-angle extensional faulting, reactivated mylonites, and seismic reflection geometry of the Newark basin margin in eastern Pennsylvania: Geology, v. 14, p. 766–770.

Reynolds, D. J., Olsen, P. E., Steckler, M. S., and Burgess, C. F., 1990, Structural framework of the Newark basin [abs.]: Eos (American Geophysical Union Transactions), v. 71, p. 1605–1606.

Robson, D. A., 1971, The structure of the Gulf of Suez (Clysmic) rift, with special reference to the eastern side: Geological Society of London Journal, v. 127, p. 247–276.

Sanders, J. E., 1962, Strike-slip displacement on faults in Triassic rocks in New Jersey: Science, v. 136, p. 40–42.

Sanders, J. E., 1963, Late Triassic tectonic history of northeastern United States: American Journal of Science, v. 261, p. 501–524.

Schlische, R. W., 1990, Aspects of the structural and stratigraphic development of early Mesozoic rift basins of eastern North America [Ph.D. thesis]: New York, Columbia University, 479 p.

Schlische, R. W., 1991, Half-graben basin filling models: New constraints on continental extensional basin development: Basin Research, v. 3, p. 123–141.

Schlische, R. W., and Olsen, P. E., 1988, Structural evolution of the Newark basin, *in* Husch, J. M., and Hozik, M. J., eds., Geology of the central Newark basin, field guide and proceedings: Annual Meeting of the Geological Association of New Jersey, 5th, Rider College, Lawrenceville, New Jersey, p. 43–65.

Schlische, R. W., and Olsen, P. E., 1990, Quantitative filling model for continental extensional basins with applications to the early Mesozoic rifts of eastern North America: Journal of Geology, v. 98, p. 135–155.

Schlische, R. W., and Reynolds, D. J., 1992, A new view of folding in early Mesozoic rift basins: Geological Society of America Abstracts with Programs, v. 24, p. 73.

Schlische, R. W., Olsen, P. E., Cornet, B., and Silvestri, S. M., 1991, Preliminary analysis of Newark basin drilling project cores: Implications for basin tectonics: Geological Society of America Abstracts with Programs, v. 23, p. A251.

Schwartz, D. P., and Coppersmith, K. J., 1984, Fault behavior and characteristic earthquakes: Examples from the Wasatch and San Andreas fault zones: Journal of Geophysical Research, v. 89, p. 5681–5698.

Sheridan, R. E., and seven others, 1991, EDGE deep seismic reflection study of the U.S. mid-Atlantic continental margin reveals relationships of Appalachian sutures to Mesozoic magmatic underplating: Geological Society of America Abstracts with Programs, v. 23, p. A309.

Silvestri, S. M., 1991, Ichnofauna of the last seven million years of the Triassic from the Jacksonwald syncline, Newark basin, Pennsylvania [M.S. thesis]: Newark, New Jersey, Rutgers University, 180 p.

Stuck, R. J., Vanderslice, J. E., and Hozik, M. J., 1988, Paleomagnetism of igneous rocks in the Jacksonwald syncline, Pennsylvania: Geological Society of America Abstracts with Programs, v. 20, p. 74.

Sutter, J. F., 1988, Innovative approaches to the dating of igneous events in the early Mesozoic basins, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins in the eastern United States: U.S. Geological Survey Bulletin 1776, p. 194–200.

Unger, J. D., 1988, A simple technique for analysis and migration of seismic reflection profiles from the Mesozoic basins of eastern North America, *in* Froelich, A. J., and Robinson, G. R., Jr., eds., Studies of the early Mesozoic basins in the eastern United States: U.S. Geological Survey Bulletin 1776, p. 229–235.

Van Houten, F. B., 1962, Cyclic sedimentation and the origin of analcime-rich Upper Triassic Lockatong Formation, west-central New Jersey and adjacent Pennsylvania: American Journal of Science, v. 260, p. 561–576.

Vreeland, J. H., and Berrong, B. H., 1979, Seismic exploration in Railroad Valley, Nevada, *in* Newman, G. W., and Goode, H. D., eds., Basin and Range symposium and Great Basin field conference: Denver, Colorado, Rocky Mountain Association of Geologists, p. 557–569.

Walsh, J. J., and Watterson, J., 1987, Distribution of cumulative displacement and of seismic slip on a single normal fault surface: Journal of Structural Geology, v. 9, p. 1039–1046.

Walsh, J. J., and Watterson, J., 1988, Analysis of the relationship between displacements and dimensions of faults: Journal of Structural Geology, v. 10, p. 239–247.

Watson, E. H., 1958, Triassic faulting near Gwynedd, Pennsylvania, *in* Pennsylvania Academy of Sciences, Proceedings, v. 32, p. 122–127.

Watterson, J., 1986, Fault dimensions, displacements, and growth: Pure and Applied Geophysics, v. 124, p. 365–373.

Wernicke, B., and Axen, G. J., 1988, On the role of isostasy in the evolution of normal fault systems: Geology, v. 16, p. 848–851.

Wheeler, G., 1939, Triassic fault-line deflections and associated warping: Journal of Geology, v. 47, p. 337–370.

Willard, B., Freedman, J., McLaughlin, D. B., Ryan, J. D., Wherry, E. T., Peltier, L. C., and Gault, H. R., 1959, Geology and mineral resources of Bucks County, Pennsylvania: Pennsylvania Geological Survey Bulletin C9.

Withjack, M. O., and Gallagher, J. J., 1983, The rifted basins of eastern India, *in* SEAPEX Proceedings, v. 6, p. 41–57.

Witte, W. K., and Kent, D. V., 1991, Tectonic implications of a remagnetization event in the Newark basin: Journal of Geophysical Research, v. 96, p. 19569–19582.

Witte, W. K., Kent, D. V., and Olsen, P. E., 1991, Magnetostratigraphy and paleomagnetic poles from Late Triassic–earliest Jurassic strata of the Newark basin: Geological Society of America Bulletin, v. 103, p. 1648–1662.

Zhang, P., Slemmons, D. B., and Mao, F., 1991, Geometric pattern, rupture termination and fault segmentation of the Dixie Valley–Pleasant Valley active normal fault system, Nevada, U.S.A.: Journal of Structural Geology, v. 13, p. 165–176.

Manuscript Received by the Society December 16, 1991
Revised Manuscript Received February 10, 1992
Manuscript Accepted February 14, 1992

Printed in U.S.A.