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ANATOMY AND EVOLUTION OF THE TRIASSIC-JURASSIC CONTINENTAL RIFT SYSTEM, EASTERN NORTH AMERICA

Roy W. Schlische
Department of Geological Sciences, Rutgers University, Piscataway, New Jersey

Abstract. Mesozoic rift basins in eastern North America formed during continental extension associated with the separation of North America and Africa. These basins locally overprint the Appalachian orogen and involve the extensional reactivation of Paleozoic faults. Half graben are thought to have formed where Mesozoic extension was subperpendicular to orogenic strike. Transtensional basins formed where the extension was more oblique. Segmented border fault systems and predominantly synthetic intrabasinal faults characterize the half graben. These basins resemble elongate synclines in longitudinal section; this geometry resulted from border-fault displacement that was greatest near the center of the fault and decreased toward both ends. Large-scale segmentation of some border fault systems resulted in the formation of multiple synclinal subbasins separated by transverse anticlines at segment boundaries, where fault displacement was less. As displacement increased on individual fault segments, the faults and associated basins grew in length, perhaps linking originally isolated basins. Smaller-scale fault segmentation resulted in the formation of relay ramps, rider blocks, and transverse folds. Some transverse synclines are located near the centers of fault segments, and related anticlines are located at segment boundaries. Adjacent half-graben units within larger rift zones do not alternate polarity along strike and are generally not linked via accommodation zones, as in the East African rift system. Strike-slip-dominated basins are characterized by a network of strike-slip and normal faults, and are shallower and narrower than dip-slip-dominated basins.

INTRODUCTION

Fault-bounded sedimentary basins, typically half graben, are a fundamental manifestation of continental extension [e.g., Bally, 1982; Wernicke and Burchfiel, 1982; Anderson et al., 1983; Jackson and McKenzie, 1983; Gibbs, 1984; Rosendahl, 1987] and are also the prevalent architecture of transtensional regimes, at least within the Tanganyika-Rukwa-Malawi system in East Africa [Rosendahl et al., 1992; Scott et al., 1992]. Numerous basins formed along the margins of the incipient North Atlantic Ocean during the Mesozoic breakup of Pangea (Figure 1a) [Van Houten, 1977; Froelich and Olsen, 1984; Tankard and Balkwill, 1989]. In eastern North America, basins crop out over a distance of >1700 km; other basins are concealed beneath coastal plain deposits and the continental shelf (Figure 1b) (see recent summaries by Froelich and Robinson [1988], Manspeizer [1988], and Sheridan and Grow [1988]. The exposed basins are filled with thousands of meters of exclusively nonmarine strata [e.g., Smoot, 1985; Manspeizer et al., 1989; Olsen et al., 1989; Smoot , 1991] as well as tholeiitic lava flows and diabase intrusions [e.g., Froelich and Gottfried, 1988; Puffer and Philpotts, 1988; Manspeizer et al., 1989; Puffer and Ragland, 1992], most of which crystallized at ~200 Ma [Sutter, 1988; Dunning and Hodych, 1990]. Biostratigraphic dating indicates that preserved basin strata range in age from Middle Triassic to Early Jurassic (Table 1) [Cornet and Olsen, 1985; Olsen et al., 1989].

Rift basins in eastern North America locally overprint the Appalachian orogen and exhibit parallelism with Paleozoic contractional structures (Figure 1b) [e.g., Lindholm, 1978; Swanson, 1986; Ratcliffe et al., 1986]. The exposed rift basins are situated landward of the hinge zone of the continental margin; this region experienced considerably less crustal thinning than did the seaward region [Klitgord et al., 1988]. Consequently, many of these landward basins were not deeply buried by postrift strata and therefore remain accessible today.

This paper examines the similarities and differences in the architecture of exposed Mesozoic rift basins in eastern North America. Particular attention is given to the geometry of border fault systems and associated structures, especially along-strike variability. The relationship between structural geology and stratigraphy is used to infer the evolution of the rift basins. Finally, the nature of the linkage of basins within the rift complex is examined and compared to that of the well-studied East African rift system.

NOMENCLATURE

The border fault system (BFS) of a half graben refers to the network of normal faults bounding the asymmetric basin (Figure 2a); movement on these faults was largely responsible for the formation of the basin. Because the exact slip direction on most of the boundary faults is difficult to determine and some boundary faults likely experienced significant strike slip, BFS is used in the general sense for the primary basin-bounding fault system regardless of the nature of slip along it.

Longitudinal structures or profiles are oriented parallel to the BFS; transverse structures or profiles are oriented perpendicular to the BFS (Figure 2c).

Because the BFS commonly consists of multiple faults, the term segment refers to an individual fault within the BFS. As the geometry of the faults at depth is poorly constrained, segmentation is based exclusively on the map view trace of the BFS. Following the scheme used to define the segmentation of seismically active faults [e.g., Zhang et al., 1991], segment boundaries are marked by changes in strike as well as offsets or overlaps (Figures 2b and 2c). For overlapping faults, a common segment boundary may be placed in the center of the region of overlap. The relatively unfaulted blocks of rock located between overlapping fault segments are called ramps by Kelley [1979], fault bridges by Ramsay and Huber [1987] and relay ramps by Larsen [1988], Morley et al. [1990], and Peacock and Sanderson [1991]. Where such structures occur in extensional basins, synrift strata unconformably overlap prerift rocks on relay ramps (Figure 2c) [e.g., Larsen, 1988]. In regions of overlapping or subparallel faults, the bottom of the basin may step down along progressively more basinward-situated faults. The blocks between the faults are termed riders by Gibbs [1984] (Figure 2a).

Ramping margin refers to the relatively unfaulted basin margin that generally dips toward the BFS (Figure 2a). The intersection of this margin with the Earth’s surface is the prerift-synrift contact. Prerift rocks form the “basement” to the basins; synrift rocks refer to the basin fill.

Several basins are composed of smaller or structurally distinct subbasins. The Fundy basin (Figure 3a) is subdivided into the northeast trending Fundy and Chignecto subbasins and the east trending Minas subbasin. The Deerfield and Hartford subbasins form the Connecticut Valley basin (Figure 4a). The Deep River basin (Figure 6c) consists of the Durham, Sanford, and Wadesboro subbasins.

RIFT BASIN ARCHITECTURE

The majority of the exposed basins in eastern North America are north-northeast to northeast trending half graben or contain half-graben type of subbasins (Table 1); exceptions include the east trending Minas subbasin and Narrow Neck. Geologic maps of nine basins are presented in Figures 3-6. Schematic transverse and longitudinal cross sections of these basins are depicted in Figures 7 and 8.

Half Graben

Basin Geometry. The Connecticut Valley, Pomperaug, Newark, Gettysburg, Culpeper, Richmond, Danville-Dan River, and Deep River basins as well as the combined Fundy-Chignecto subbasin vary in length from 11 to 356 km, in maximum width from 4 km to 77 km, and in estimated maximum depth from 1.5 to 7 km (Table 1). Shorter basins or subbasins tend to be narrower and shallower than longer basins or subbasins (Figures 7 and 8). In transverse section, these basins or subbasins display the classic asymmetric half-graben geometry (Figure 7). With the exception of the Connecticut Valley, Pomperaug, and Deep River basins in which the strata are east-southeast to southeast tilted, the majority of the basins have west-northwest to northwest tilted strata.

The trace of the prerift-synrift contact for the Pomperaug, Newark, Gettysburg, Culpeper, and Richmond basins is broadly concave toward the BFS. In addition, the dip direction of the basin fill is perpendicular to the BFS near the center of the basin but oblique near the ends. Both observations suggest that in longitudinal section these basins resemble elongate synclines, some of which are asymmetric (Figure 8). A large-scale synclinal geometry for the Newark basin may be inferred from Triassic units that thicken from the lateral edges of the basin toward its center [Olsen, 1980a; Olsen, 1988; Schlische, 1992]; for example, the outcrop width of the Lockatong Formation increases markedly toward the basin center (Figure 4b). The inferred simple synclinal geometry is, however, complicated by intrabasinal faults (Figure 8, section c-c’). The synclinal geometry of the Richmond basin is corroborated by gravity and magnetic modeling [Mickus et al., 1988]. On the basis of seismic reflection data, the Chignecto subbasin (Figure 3a) is interpreted to shoal toward the northeast [Schlische, 1990].

The Connecticut Valley, Danville-Dan River, and Deep River basins consist of multiple synclinal subbasins separated by transverse anticlines (Figure 8). The trace of the prerift-synrift contact is broadly concave toward the BFS for each subbasin; narrower transverse anticlines are associated with convex traces. Inliers of prerift rocks crop out within the transverse anticline between the Deerfield and Hartford subbasins (Figure 4a). Seismic refraction data indicate that the Hartford subbasin is deepest near the Massachusetts-Connecticut border (Figure 4a) and shoals northward toward the transverse anticline and southward toward the edge of the subbasin [Wenk, 1984b; Wise, 1992]. Paleoflow structures indicate that streams in the Hartford subbasin flowed southward and those in the Deerfield subbasin flowed northward away from the transverse anticline

[Hubert et al., 1992]. Lacustrine strata are apparently absent in “the narrows” between the northern and southern subbasins of the Danville-Dan River basin (Figure 6b) as well as in the Colon cross structure between the Durham and Sanford subbasins of the Deep River basin (Figure 6c), suggesting that these areas stood higher than the flanking subbasins. Synrift strata are thinnest within the Pekin cross structure of the Deep River basin (Figure 6c) and thicken toward the Sanford subbasin to the northeast and toward the Wadesboro subbasin to the southwest [Zablocki, 1959]; however, there is no appreciable change in dip direction across the structure [Randazzo and Copeland, 1976]. Differential subsidence is likely responsible for the geometry of the subbasins and intervening anticlines.

Border fault systems and associated structures. Mesozoic rift basins closely parallel and overprint the Appalachian orogen. Consequently, many of the border faults are inferred to have reactivated preexisting weaknesses, mainly Paleozoic thrust faults [Swanson, 1986] (see Table 1 for more recent citations). The shallow dip of some of the border faults (Table 1) is also consistent with reactivation of gently dipping Paleozoic structures.

On the basis of seismic reflection profiles, the border faults of the Fundy subbasin [Schlische, 1990], Culpeper basin [Manspeizer et al., 1989], and Sanford subbasin [Olsen et al., 1990] are interpreted to be listric. However, velocity pull-ups may cause a planar normal fault to appear listric [Withjack and Drickman Pollock, 1984; Unger, 1988]. The border faults of other basins have been inferred to be listric (Table 1), based mainly on the presence of hanging-wall rollover folds. However, rollover folding may also be produced by nonuniform displacement in the volume surrounding a planar normal fault [Barnett et al., 1987; Gibson et al., 1989].

Syndepositional border faulting is indicated by a progressive decrease in dip in progressively younger synrift strata; this feature has been observed in the Connecticut Valley [Wise, 1992], Newark [Schlische, 1992], and Richmond [Venkatakrishnan and Lutz, 1988] basins. An increase in thickness of synrift strata toward the BFS indicates that border faulting produced an asymmetric basin during sedimentation. Figure 3c shows the pronounced thickening of all formations from the ramping margin toward the BFS of the Fundy subbasin. Similar thickening trends were observed in the Chignecto subbasin [Withjack et al., 1991], Hartford subbasin [Olsen, 1988; Olsen et al., 1989; McDonald and LeTourneau, 1990; Hubert et al., 1992], Newark basin [Olsen, 1980a; Olsen et al., 1989; Schlische, 1992], Richmond basin [Olsen et al., 1989], and Sanford subbasin [Olsen et al., 1990]. The presence of alluvial fan conglomerate adjacent to virtually all BFSs (Figures 3-6) is also consistent with syndepositional faulting.

The BFSs of the Mesozoic rift basins are commonly segmented. Large-scale segmentation of some BFSs appears to correlate with the geometry of subbasins and transverse anticlines in some basins. For example, the Durham subbasin of the Deep River basin is associated with a north-northeast striking BFS, whereas the Sanford subbasin is bounded by a northeast striking BFS; the change in strike occurs at the Colon cross structure (Figure 6c). A somewhat smaller-scale segmentation involving overlapping fault strands is associated with relay ramps in the Fundy, Newark, Culpeper, Richmond, and Deep River basins (Figures 3-6). Exposures of prerift rocks at relay ramps suggest minimal subsidence between the overlapping fault segments.

Seismic reflection and well data have identified rider blocks along the BFS margin of the southwest Newark basin (Figure 7, section G-G’) [Reynolds et al., 1990], yet some of the fault splays have no known surface expression. These splays are situated approximately along strike from fault segments associated with relay ramps, suggesting that some fault segments continue in the subsurface. Stratigraphic relations (Figure 7, section G-G’) indicate that younger faults progressively propagated into the footwall block, widening the basin in the process [Schlische, 1992]. Geophysical data from the Hartford subbasin [Wenk, 1984a], Gettysburg basin [Sumner, 1977], and Sanford subbasin [Lai et al., 1985] suggest that the deepest parts of these basins are located substantially basinward of the surface trace of the BFSs. Unless the BFSs have much shallower dips and much higher displacements than those indicated in Table 1 and shown in Figure 7, the geophysical data are consistent with step-faulted BFS margins (rider blocks).

In addition to the basin-scale folds described above, smaller-scale transverse folds are common in a number of basins or subbasins [e.g., Wheeler, 1939] (Table 1). In the southern part of the Hartford subbasin (Figure 4a), three large synclines (A, B, and C) are outlined by the map pattern of Jurassic basalts. The anticline between synclines C and B appears to correspond to a change in strike of the BFS, a possible fault segment boundary. Syncline B contains two smaller-wavelength synclines separated by an anticline, also corresponding with a change in strike of the BFS. The transverse folds in the northeast parts of the Newark basin (Figure 4b) clearly decrease in amplitude away from the BFS. In the southwest part of the same basin, synclines are associated with the centers of fault segments, whereas anticlines are associated with the overlap regions of segments, suggesting that the folds are fault-related. Stratigraphic units thicken in synclinal troughs and thin in anticlinal crests, indicating that the folds formed syndepositionally [Schlische, 1992]. In the Culpeper basin, two northwest plunging synclines (labeled A and B in Figure 5c), each of which contain several shorter-wavelength folds, are defined by the map pattern of the basalt flows. A tight anticline between synclines A and B coincides with a BFS segment boundary and small relay ramp. The main subbasins of the Danville-Dan River basin contain a series of northwest plunging transverse folds possibly related to differential downwarping along the BFS [Thayer, 1970].

Intrabasinal faults and dikes. The mean strikes of intrabasinal faults and dikes listed in Table 1 were derived from the map traces of the faults and dikes shown in Figures 3-6. A majority of the intrabasinal faults are northeast striking and synthetic to the BFSs (see Figure 7). In the Hartford subbasin (Figure 4a), northeast striking intrabasinal normal faults are oblique to the generally north-northeast striking BFS. The few northeast striking segments of the BFS are reported to have formed initially in the Mesozoic coeval with the intrabasinal faults [de Boer and Clifford, 1988] and thus better reflect Mesozoic extension than do the reactivated segments of the BFS. Intrabasinal faults are also oblique to the BFS in the Newark basin (Figure 4b). Antithetic faults generally have less throw than the synthetic faults (an exception occurs in the Richmond basin) and are most common along the hinged margins of the basins (Figure 7). Stratigraphic relations based on drill core data indicate that intrabasinal faults in the Newark [Schlische, 1992] and Richmond [Olsen et al., 1989] basins were syndepositionally active. Transverse folds are associated with some intrabasinal faults of the Connecticut Valley, Newark, and Deep River basins.

The strike of diabase dikes changes from northeasterly in the northern part of the rift system to northwesterly in the southern part (Table 1) [e.g., McHone, 1988]. In the Connecticut Valley and Newark basins, dikes and intrabasinal faults are subparallel (Figure 4). Northwest striking dikes of the Dan River basin are not offset along the BFS [Thayer, 1970] (Figure 6b). Vertical northwest striking dikes were intruded after rotation of the basin strata along the BFS of the Deep River basin (Figure 6c) [Randazzo and Copeland, 1976; Bain and Harvey, 1977]; these dikes therefore record northeast-southwest extension that postdated basin formation and sedimentation.

Strike-Slip Basins

The east trending Minas subbasin and Narrow Neck are highly oblique to the trend of half-graben basins, and their border faults likely experienced a large component of strike slip. The Minas subbasin is bordered to the north by the Minas fault zone (Mfz), which merges westward with the BFS of the Fundy subbasin (Figure 3a) [Olsen and Schlische, 1990; Schlische, 1990; Withjack et al., 1991]. The Mfz experienced right slip during the Carboniferous [Keppie, 1982] and was reactivated in left-oblique slip during the subsequent northwest-southeast extension that created the Fundy and Chignecto half graben [Keppie, 1982; Olsen and Schlische, 1990]. In the Five Islands region of Nova Scotia, the BFS is composed of east striking left-lateral faults and northeast striking normal faults (Figure 3b) [Olsen and Schlische, 1990] that resemble a left-slip extensional strike-slip duplex [Woodcock and Fischer, 1986]. Stratigraphic thickness varies considerably among the fault blocks, suggesting syndepositional faulting but is consistently smaller than in the dip-slip-dominated Fundy subbasin (Figure 3c). Since the same stratigraphic units are present in the Minas and Fundy subbasins, both the dip-slip and strike-slip margins of the two subbasins were active coevally [Olsen and Schlische, 1990].

The Narrow Neck is bounded by and contains east and northeast striking faults (Figure 5b). The east striking faults are probably sinistral strike-slip faults based on (1) west-northwest plunging folds that intersect east striking faults at angles of 10°-15° [McLaughlin, 1963], (2) the narrow width of the Narrow Neck (7 km), and (3) the thinner stratigraphic succession within the Narrow Neck compared with the Newark and Gettysburg basins. Sinistral faults have been mapped near the eastern end of the Narrow Neck [Lucas et al., 1988]. The northeast striking faults are thought to be normal faults.

Stratigraphy

The oldest strata in all basins are fluvial (Figures 3-7 and Table 1). Basin-wide fluvial sedimentation implies that the basins received an excess supply of sediment relative to basin capacity [Schlische and Olsen, 1990; Schlische, 1991]. In all basins the fluvial deposits are overlain by basin-wide lacustrine deposits or lava flows intercalated with lacustrine strata; these deposits signify a partially sediment-starved basin [Schlische and Olsen, 1990; Schlische, 1991]. Early Jurassic-age lava flows and lacustrine strata are found in the Fundy, Connecticut Valley, Newark, Gettysburg, and Culpeper basins but are absent in the southern basins. Jurassic deposits may have accumulated and been subsequently eroded; however, the absence of Norian (latest Triassic) deposits in the Richmond and Deep River basins would also require substantial erosion of Triassic deposits. Alternatively, Jurassic strata may never have accumulated in the southern basins if they stopped subsiding prior to Early Jurassic time. This notion is supported by the observation that Early Jurassic dikes cut across the grain of the southern basins and are thus postrift structures. A Jurassic basalt flow sequence of the South Georgia basin (Figure 1b) is also apparently a postrift unit [McBride et al., 1989].

Outcrop data from the Fundy subbasin [Hubert and Forlenza, 1988; Olsen et al., 1989], seismic reflection data from the Newark basin [Schlische, 1992], and drill hole data from the Richmond basin [Olsen et al., 1989] show that progressively younger synrift strata onlap prerift rocks along the ramping margin. Hanging wall onlap has also been observed on seismic reflection data from offshore basins and suggests that the basins progressively increased in width [Schlische and Olsen, 1990]. Longitudinal pinchout at the lateral ends of the northeast Newark and southern Dan River basins (Figures 4b and 6b) is a map view expression of longitudinal onlap, suggesting that the basins lengthened through time.

Synthesis

A generalized model of a Mesozoic extensional basin in eastern North America is shown in Figure 9a. A typical basin is a half graben with one margin bordered by a fault system that apparently shows predominantly normal dip slip and that was at least sporadically active during synrift sedimentation. The BFS generally parallels the grain of older Paleozoic structures; consequently, most of the border faults are known or inferred to be reactivated structures [e.g., Swanson, 1986]. Where border faults strike oblique to the Mesozoic extension direction, a strike-slip component of faulting is common [Ratcliffe and Burton, 1985].

The BFS is commonly segmented, with segment boundaries marked by changes in fault strike or fault overlap. Relay ramps and rider blocks occur between overlapping fault segments. Fault-normal anticlines tend to occur at or near segment boundaries, whereas synclines are found at or near the centers of fault segments. The ramping margin is generally unfaulted or contains only minor, mostly antithetic faults. Intrabasinal faults are typically synthetic to and may be slightly oblique to the BFS. Antithetic intrabasinal faults generally display less throw than the synthetic faults. Dikes are subparallel to intrabasinal normal faults and may be oblique to border faults.

In transverse profile, the idealized basin has a classic half-graben geometry. Total basin depth and the thickness of stratal units increase toward the BFS, reflecting fault-controlled basin asymmetry. In longitudinal profile, the basin is synclinal, with total basin depth and synrift stratal thickness increasing from the lateral edges to the center of the basin. The typical Triassic stratigraphy consists of a basal fluvial unit overlain by lacustrine strata, with the deepest lakes occurring near the base of the lacustrine succession; this is overlain by an Early Jurassic-age sequence of lava flows and intercalated lacustrine strata overlain by shallow lacustrine and fluvial strata.

Major exceptions to the general basin model outlined are as follows: (1) In the southern basins, diabase dikes are perpendicular to or highly oblique to the basin margins, and Jurassic strata are absent; (2) there are apparently no Triassic lacustrine strata in the Connecticut Valley basin; (3) some basins consist of multiple synclinal subbasins separated by transverse anticlines (Figure 9b); and (4) basins dominated by strike slip are shallower and have a thinner stratigraphic section than dip-slip basins (Figure 9c); if the Minas subbasin is representative of this structure, then such basins are characterized by mosaics of strike-slip faults that are oblique to the regional extension direction and normal faults that are perpendicular to the regional extension direction.

DISCUSSION

Growth of Normal Faults and Basin Evolution

Many of the rift basins or subbasins in eastern North America approximate elongate, in some cases asymmetric, synclines in longitudinal section (Figures 8 and 9a). These structures suggest differential subsidence, which was highest near the center of the basin or subbasin and decreased along strike. Because most of the subsidence in these basins was fault-controlled, the synclinal geometry likely resulted from along-strike variations in displacement along the BFSs: fault displacement was greatest near the center of the BFS and systematically decreased toward its ends. Similar displacement variations occur on faults ranging in length from a few centimeters to tens of kilometers, indicating that this is a scale-invariant feature of normal faults [Chapman et al., 1978; Muraoka and Kamata, 1983; Barnett et al., 1987; Walsh and Watterson, 1987; Peacock and Sanderson, 1991; Dawers and Anders, 1992].

A positive correlation between fault length and maximum displacement indicates that faults grow in length as cumulative displacement increases [e.g., Watterson, 1986; Walsh and Watterson, 1988; Marrett and Allmendinger, 1991; Cowie and Scholz, 1992a, b, c; Dawers and Anders, 1992]. A similar relationship among basin length, width, and depth for Mesozoic basins (Table 1) suggests basin growth. In a growing basin, younger synrift units should progressively onlap prerift rocks (Figure 10a). Longitudinal onlap is expressed in map view by longitudinal pinchout (Figure

10a, stage 4), which has been observed in the Newark and Dan River basins. If the rate at which basin capacity increases due to basin growth is faster than the rate of sediment infilling, there may be a transition from an oversupply of sediment (fluvial sedimentation) to partial sediment starvation (lacustrine sedimentation) (Figure 10a). Fluvial-lacustrine transitions occur in all Mesozoic basins (Table 1).

Normal fault systems and rift basins associated with large normal faults are commonly segmented [Schwartz and Coppersmith, 1984; Peacock and Sanderson, 1991; Nelson et al., 1992]. In eastern North America, the largest-scale segmentation is expressed by multiple subbasins within larger basin complexes. The evolution of such basins may involve linkage of originally isolated subbasins (Figure 10b). As fault slip increased, the faults lengthened and the isolated basins grew in size and eventually merged. Older synrift units are expected to form restricted sequences in each subbasin, whereas younger units may be deposited basin-wide.

Basin growth and subbasin linkage have been inferred for other continental rift systems. Lake Baikal occupies the oldest part of the Baikal rift system, and the age of normal faulting and associated sedimentation becomes younger to the northeast and southwest [Logatchev and Florensov, 1978]. Within Lake Baikal itself, the northern subbasin may be younger than, and have been originally isolated from, the central subbasin [Hutchinson et al., 1992]. Coalescence of subbasins has also been inferred from stratigraphic and structural relations in the East African rift system [Burgess et al., 1988; Ebinger, 1989b; Sander and Rosendahl, 1989; Rosendahl et al., 1992] and the Rio Grande rift [Chapin, 1979].

Border Fault Segmentation, Relay Ramps, Transverse Folds, and Rider Blocks

Large-scale border fault segmentation is thought to be responsible for the formation of multiple subbasins. As shown in the idealized model in Figure 11, additional segmentation of the BFS may be responsible for the presence of relay ramps, rider blocks, and transverse folds. Consider a number of fault segments arranged in a relay series in which the ends of the fault segments overlap. If each fault segment displays the characteristic along-strike variations in displacement, then displacement maxima occur near the centers of fault segments, and displacement minima occur at the segment boundaries (the zones of overlap). This results in a set of fault-displacement folds, with synclinal fold traces located at displacement maxima and anticlinal traces at displacement minima. Relay ramps form in the regions between the overlapping segments (Figure 11). Many of the transverse folds in the southwest Newark basin (Figure 4b) are related to fault segments arranged in relay.

BFS segmentation responsible for the transverse folds usually has no basin-wide expression. In the southwest Newark basin (Figure 4b), many folds and BFS segments are present, yet the trace of the prerift-synrift contact is unaffected by this segmentation. Instead, the trace of this contact suggests that the BFS of the Newark basin can be treated as a single fault with maximum displacement near its center. This also appears to be the case in the southern Connecticut Valley and central Culpeper basins and is depicted in the idealized model in Figure 9a. There are three possible explanations for these observations:

1. The current segmented BFS developed relatively late in the history of the basin, modifying an earlier BFS that was largely responsible for the geometry of the prerift-synrift contact. In the southwest Newark basin, stratigraphic data indicate that parts of the BFS were once situated more basinward [Schlische, 1992].

2. The transverse folds are controlled by displacement variations on the nearest BFS segment, whereas the trace of prerift-synrift contact reflects the sum of displacements on all faults, large and small, within and bounding the basin. In the North Sea, a large percentage of regional extension is accommodated on small faults [e.g., Marrett and Allmendinger, 1992].

3. The BFS segmentation responsible for the transverse folds is a relatively near-surface phenomenon. At depth, individual segments may merge into a master fault (Figure 11); the overall geometry of the basin thus reflects displacement variations on the composite fault system. More data on the subsurface geometry of BFSs are needed to test this hypothesis, but kinematic linkage among normal fault splays has been demonstrated in the North Sea [Walsh and Watterson, 1991].

Differential displacement along complex normal faults has been proposed to account for the transverse folds in the Connecticut Valley [Wheeler, 1939; Withjack and Drickman Pollock, 1984], Newark [Wheeler, 1939], and Dan River basins [Thayer, 1970]. Wheeler [1939] emphasized the role of fault-surface irregularities in generating the folds; he argued that synclines and anticlines formed at recesses and salients, respectively, in the fault surface. Some of the smaller-wavelength folds from the southern Hartford subbasin (Figure 4a) appear to fit this model rather well. Wheeler's model is not incompatible with the segmentation model for producing transverse folds discussed above because salients delimit fault segment boundaries along active normal faults in the Basin and Range [e.g., Machette et al., 1991].

Previously published folding mechanisms unrelated to differential displacement along normal faults include strike-slip faulting [Manspeizer, 1981], synrift shortening subparallel to the BFS [Ratcliffe and Burton, 1985], and postrift regional shortening [Sanders, 1963]. Schlische [1992] rejected these models for the Newark basin because (1) en echelon folds intersect strike-slip faults at angles of <45° [Christie-Blick and Biddle, 1985], not the 70°-90° typically observed; (2) diabase dikes apparently synchronous with folding indicate extension parallel to the BFS; and (3) some folds are clearly synrift structures.

The formation of some rider blocks can be attributed to the relay geometry of border fault segments as shown in Figure 11. This geometry implies that segment B continues southwestward in the subsurface. Segments B and C may also appear to extend northeastward (as shown by the dashed lines in Figure 11) if these segments reactivated Paleozoic faults. However, the extent of Mesozoic reactivation is limited to the areas around the relay ramps because the presence of prerift rocks in the relay ramps implies minimal subsidence. Rider blocks also formed as a result of progressive incisement of the footwall block, as observed in the Newark basin. This type of rider block may have developed in response to gravitational collapse of the uplifted footwall block [Gibbs, 1984], perhaps aided by preexisting structural weaknesses within the footwall block [Schlische, 1992]. This mechanism contrasts with the rolling-hinge model of Buck [1988], which predicts progressive hanging wall incisement of fault splays.

Intrabasinal Faults

Intrabasinal faults synthetic to the BFS are considerably more common than antithetic faults. There are two possible explanations.

1. Normal faults that dip in the same general direction are less likely to crosscut at depth. Consequently, a larger number of faults can be active at the same time in order to accommodate extension [e.g., Jackson and McKenzie, 1983].

2. As with border faults, intrabasinal faults may have originated as preexisting structures in prerift rocks. As intrabasinal faults propagated upward through the basin fill, the faults may have changed orientation in order to become normal to the extension direction. Thus, the strikes of intrabasinal and border faults may differ, and this is commonly observed (Table 1).

The density of intrabasinal faults apparent at the surface varies from basin to basin and even within a basin. This may in part represent mapping details; for example, faults that offset sedimentary rocks are more difficult to detect than those cutting basalt and diabase. In the Newark basin, the apparent frequency of intrabasinal faults increases in those parts of the basin where the BFS has a shallower dip [Schlische, 1992]. The orientation of the reactivated BFS with respect to the extension direction also may be a factor. Where extension is oblique, new faults oriented normal to the extension direction may have formed to more easily accommodate the extension.

Rift Basin Linkage

The Connecticut Valley basin, Newark-Gettysburg-Culpeper basins, and the Deep River basin may be considered to be rift zones (nomenclature of Rosendahl [1987]) in that they consist of multiple half graben. Although the polarity (sense of asymmetry) of the rift zones varies along the eastern North American rift system, individual half graben within a given rift zone exhibit no reversals in asymmetry. In contrast, the Tanganyika and Malawi rift zones each exhibit six to eight reversals in half-graben asymmetry along strike [Rosendahl, 1987; Ebinger, 1989b; Rosendahl et al., 1992]. The lack of polarity reversals within rift zones in eastern North America may be due to the existence of a precursor tectonic fabric of the Appalachians in eastern North America, which probably controlled the localization of border faults [Burgess et al., 1988; Reynolds and Schlische, 1989; Rosendahl, 1990]. No doubt, the explanation for this significant difference is more complicated: (1) the BFSs of some basins in the East African rift system may be reoccuppied thrust faults [Wheeler, 1989; Kilembe and Rosendahl, 1991]; (2) large shear zones may influence the locations of polarity reversals in East Africa [Versfelt and Rosendahl, 1989]; and (3) there is no known preexisting structural control for the border faults of the Deep River basin.

There appears to be little physical (“hard”) linkage among border faults bounding basins and subbasins in eastern North America. No known structures directly connect the southeast dipping border fault of the Newark basin and the west dipping border fault of the Hartford subbasin. Rather, displacement on the Newark basin border fault apparently dies out to the northeast, and that on the Hartford subbasin border fault apparently dies out to the south. Thus, no accommodation structure is required. Although the Hartford and Deerfield subbasins are connected, they are essentially overlapping zones of subsidence associated with two distinct BFS segments (similar to the model in Figure 10b) that may have only completely linked together in Early Jurassic time [Wise, 1992]. The subbasins of the Deep River basin are similarly “soft”-linked. This type of linkage may also apply to the Gettysburg and Culpeper basins, but erosion has removed the shallow basin in the zone of overlapping subsidence.

Physical linkage among the major faults of the Fundy basin is plausible. As the width of the Fundy subbasin is much greater than that of the Chignecto subbasin, it is likely that some dip-slip displacement of the Fundy subbasin BFS was transferred to the left-oblique slip Minas fault zone (Figure 3a). The Minas fault zone also appears to have linked the Fundy basin with the Chedabucto (Orpheus) basin and may once have extended into the Gibraltar region (Figure 1a) [Olsen and Schlische, 1990]. Hard linkage is also possible between the Newark and Gettysburg basins via the Narrow Neck.

In the East African rift system, adjacent half graben are commonly linked by accommodation zones, which are faults or fault zones that trend oblique to the main border faults and underwent a considerable component of strike slip [Rosendahl, 1987; Ebinger, 1989a, b; Morley et al., 1990]. These accommodation zones appear to be particularly well developed along adjacent half graben of opposite polarity (Figure 10c). The larger number of accommodation zones in the East African rift system compared to eastern North America may be related to the closer spacing of and more frequent polarity reversals between adjacent half graben in East Africa. Clearly, an oblique or strike-slip accommodation structure is required between two oppositely dipping, overlapping BFS segments on which there is significant displacement (Figure 10c). No large-scale “hard” linkage structures are required where half graben are sufficiently widely spaced, displacement goes to zero at the fault tips, and adjacent faults dip in the same direction (Figure 10b).

CONCLUSIONS

1. Most exposed rift basins in eastern North America are half graben with border faults that likely originated as Paleozoic structures. These structures formed where the Mesozoic extension direction was oriented at a high angle to preexisting structures. Preexisting faults oriented obliquely to the extension direction experienced significant components of strike slip.

2. Predominantly dip-slip border fault margins display considerable variability along strike and are characterized by segmented fault systems (sometimes arranged in a relay pattern), relay ramps, rider blocks, and transverse folds.

3. In addition to the classic half-graben geometry in transverse section, the basins are synclinal in longitudinal section. This geometry relates to variations in border fault displacement, with maximum slip near the center of the border fault system and decreasing toward its ends.

4. Some rift basins consist of multiple synclinal subbasins separated by transverse anticlines related to displacement variations on multiple fault segments. Originally isolated subbasins may have coalesced when their border fault tips propagated laterally as displacement increased.

5. Nearly all exposed basins contain a basal fluvial unit overlain by lacustrine strata. Progressively younger strata onlap prerift rocks on the hanging wall block. Both stratigraphic relations are consistent with fault and basin growth, which is responsible for the positive correlation among basin or subbasin length, width, and depth.

6. In many basins, border fault systems are segmented at a scale smaller than the smallest subbasin. Some border fault segments are arranged in a relay pattern, with relay ramps situated at the overlap sections. Local fault displacement maximum (transverse synclines) are found near the centers of fault segments, whereas local displacement minima (transverse anticlines) are found at segment boundaries. Rider blocks formed both at relay ramps and as a result of progressive footwall incisement, perhaps driven by gravitational collapse of the uplifted footwall block.

7. Intrabasinal faults are predominantly synthetic to border fault systems to better allow coeval activity on a number of faults and due to reactivation of preexisting structures in prerift rocks.

8. Half-graben units within a larger rift zone do not alternate asymmetry along strike. Adjacent half graben are generally not linked by accommodation zones. These notable differences with the East African rift system stem from the localization of North American basins along preexisting structures that generally dip in the same direction over large areas and the wider spacing of half-graben units.

9. Basin margins dominated by strike slip are characterized by mosaics of strike-slip and normal faults and less subsidence than dip-slip margins, although both formed coevally.

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REFERENCES

Anderson, R. E., M. L. Zoback, and G. A. Thompson, Implications of selected subsurface data on the structural form and evolution of some basins in the northern Basin and Range province, Nevada and Utah, *Geol. Soc. Am. Bull.*, *94*, 1055-1072, 1983.

Bain, G. L., and B. W. Harvey, Field guide to the geology of the Durham Triassic basin, in *Carolina Geological Society, 40th Annual Meeting Guidebook*, 83 pp., 1977.

Bally, A. W., Musings over sedimentary basin evolution, *Philos. Trans. R. Soc. London*, *A305*, 325-338, 1982.

Barnett, J. A. M., J. Mortimer, J. H. Rippon, J. J. Walsh, and J. Watterson, Displacement geometry in the volume containing a single normal fault, *AAPG Bull.*, *71*, 925-937, 1987.

Berg, T. M. (Compiler), Geologic Map of Pennsylvania, scale 1: 250,000, Pennsylvania Topogr. and Geol. Surv., Harrisburg, 1980.

Brown, P. M., et al., Geologic Map of North Carolina, scale 1:500,000, North Carolina Dep. Nat. Res. and Commun. Develop., Geol. Surv. Section, Raleigh, 1985.

Buck, W. R., Flexural rotation of normal faults, *Tectonics*, *7*, 959-973, 1988.

Burgess, C. F., B. R. Rosendahl, S. Sander, S., C. A. Burgess, J. Lambiase, S. Derksen, and N. Meader, The structural and stratigraphic evolution of Lake Tanganyika: A case study of continental rifting, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 859-881, Elsevier, New York, 1988.

Chandler, W. E., Jr., Graben mechanics at the junction of the Hartford and Deerfield basins of the Connecticut Valley, Massachusetts, 151 pp., Univ. of Mass., Dept. of Geology and Geogr., (Contrib. No. 33), Amherst, 1978.

Chapin, C. E., Evolution of the Rio Grande rift—A summary, in *Rio Grande Rift—Tectonics and Magmatism*, edited by R. E. Riecker, pp. 1-5, AGU, Washington, D. C., 1979.

Chapman, G. R., S. J. Lippard, and J. E. Martyn, The stratigraphy and structure of the Kamasia Range, Kenya Rift Valley, *J. Geol. Soc. London*, *135*, 265-281, 1978.

Christie-Blick, N., and K. T. Biddle, Deformation and basin formation along strike-slip faults, in *Strike-slip Deformation, Basin Formation, and Sedimentation*, edited by K. T. Biddle and N. Christie-Blick, *SEPM Spec. Publ. 37*, pp. 1-34, 1985.

Cornet, B., and P. E. Olsen, A summary of the biostratigraphy of the Newark Supergroup of eastern North America, with comments on early Mesozoic provinciality, in *Congresso Latinoamericano de Paleontologia Mexico, vol. 3, Simposio Sobre Floras del Triasico Tardio, su Fitografia y Paleoecologia, Memoria*, edited by R. Weber, pp. 67-81, 1985.

Cowie,. P. A., and C. H. Scholz, Displacement-length scaling relationship for faults: Data synthesis and discussion, *J. Struct. Geol.*, *14*, 1149-1156, 1992a.

Cowie, P. A., and C. H. Scholz, Growth of faults by accumulation of seismic slip, *J. Geophys. Res.*, *97*, 11,085-11,095, 1992b.

Cowie, P.A., and Scholz, C.H., Physical explanation for displacement-length relationship of faults using a post-yield fracture mechanics model, *J. Struct. Geol.*, *14*, 1133-1148, 1992c.

Dawers, N. H., and M. H. Anders, Displacement-length scaling for normal faults of the "Volcanic Tableland," eastern California, *Geol. Soc. Am. Abstr. Programs*, *24*, A156, 1992.

de Boer, J. Z., and A. E. Clifford, Mesozoic tectogenesis: Development and deformation of 'Newark' rift zones in the Appalachians (with special emphasis on the Hartford basin, Connecticut), in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 275-306, Elsevier, New York, 1988.

Donohoe, H. V., Jr., and P. I. Wallace, Geological Map of the Cobequid Highlands, Colchester, Cumberland, and Pictou Counties, scale 1:50,000, Nova Scotia Dep. of Mines and Energy, Halifax, 1982.

Dunning, G. R., and J. D. Hodych, 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*, *18*, 795-798, 1990.

Ebinger, C. J., Geometric and kinematic development of border faults and accommodation zones, Kivu-Rusizi rift, Africa, *Tectonics*, *8*, 117-133, 1989a.

Ebinger, C. J., Tectonic development of the western branch of the East African rift system, *Geol. Soc. Am. Bull.*, *101*, 885-903, 1989b.

Froelich, A. J., and D. Gottfried, An overview of early Mesozoic intrusive rocks in the Culpeper basin, Virginia and Maryland, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 58-63, 1988.

Froelich, A. J., and P. E. Olsen, Newark Supergroup, a revision of the Newark Group in eastern North America, *U.S. Geol. Surv. Bull. 1537A*, A55-A58, 1984.

Froelich, A. J., and G. R. Robinson, Jr. (Eds.), Studies of the Early Mesozoic Basins of the Eastern United States, *U.S. Geol. Surv. Bull. 1776*, 423 pp., 1988.

Gibbs, A. D., Structural evolution of extensional basin margins, *J. Geol. Soc. London*, *141*, 609-620, 1984.

Gibson, J. R., J. J. Walsh, and J. Watterson, J., Modelling of bed contours and cross-sections adjacent to planar normal faults, *J. Struct. Geol.*, *11*, 317-328, 1989.

Goodwin, B. K., and K. M. Farrell, Geology of the Richmond basin, in Geology and Coal Resources of the Richmond Triassic Basin, *Va. Div. Min. Res. Open File Rep.*, 115 pp., 1979.

Hubert, J. F., and M. F. Forlenza, Sedimentology of braided river deposits in Upper Triassic Wolfville redbeds, southern shore of Cobequid Bay, Nova Scotia, Canada, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 231-247, Elsevier, New York, 1988.

Hubert, J. F., P. E. Feshbach-Meriney, and M. A. Smith, The Triassic-Jurassic Hartford rift basin, Connecticut and Massachusetts: Evolution, sandstone diagenesis, and hydrocarbon history, *AAPG Bull.*, *76*, 1710-1734, 1992.

Hutchinson, D. R., A. J. Golmshtok, L. P. Zonenshain, T. C. Moore, C. A. Scholz, and K. D. Klitgord, Depositional and tectonic framework of the rift basins of Lake Baikal from multichannel seismic data, *Geology*, *20*, 589-592, 1992.

Jackson, J., and D. McKenzie, The geometrical evolution of normal fault systems, *J. Struct. Geol.*, *5*, 471-482, 1983.

Kelley, V. E., Tectonics, middle Rio Grande rift, New Mexico, in *Rio Grande Rift—Tectonics and Magmatism*, edited by R. E. Riecker, pp. 57-70, AGU, Washington, D. C., 1979.

Keppie, J. D. (Compiler), Geological Map of Nova Scotia, scale 1:500,000, Nova Scotia Dep. of Mines and Energy, Halifax, 1979.

Keppie, J. D., The Minas geofracture, in *Major Structural Zones and Faults of the Northern Appalachians*, edited by P. St. Julien and J. Beland, *Geol. Assoc. Can. Spec. Pap. 24*, pp. 1-34, 1982.

Kilembe, E. A., and B. R. Rosendahl, Structure and stratigraphy of the Rukwa rift, *Tectonophysics*, *209*, 143-158, 1991.

Klitgord, K. D., D. R. Hutchinson, and H. Schouten, U.S. Atlantic continental margin; Structural and tectonic framework, in *The Geology of North America*, vol. I-2, *The Atlantic Continental Margin—U.S.*, edited by R. E. Sheridan and J. A. Grow, pp. 19-55, Geological Society of America, Boulder, Colo., 1988.

Lai, S. F., J. F. Ferguson, C. L. V. Aiken, and D. G. Ziegler, A test of gravity and magnetic inversion in defining the structure of the Sanford basin, North Carolina, as an example of a Triassic basin, *Soc. Expl. Geophys., Expanded Abstracts with Biographies, Technical Programs*, 207-210, 1985.

Larsen, P.-H., Relay structures in a Lower Permian basement-involved extension system, East Greenland, *J. Struct. Geol.*, *10*, 3-8, 1988.

Leavy, B. D., A. J. Froelich, and E. C. Abram, Bedrock map and geotechnical properties of rocks of the Culpeper Basin and vicinity, Virginia and Maryland, scale 1:125,000, *U.S. Geol. Surv. Map I-1313-C*, 1983.

Lindholm, R. C., Triassic-Jurassic faulting in eastern North America—A model based on pre-Triassic structures, *Geology*, *6*, 365-368, 1978.

Logatchev, N. A., and N. A. Florensov, The Baikal system of rift valleys, *Tectonophysics*, *45*, 1-13, 1978.

Lucas, M., J. Hull, and W. Manspeizer, A foreland-type fold and related structures of the Newark rift basin, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 307-332, Elsevier, New York, 1988.
Lyttle, P. T., and J. B. Epstein, Geologic map of the Newark 1°x2° Quadrangle, New Jersey, Pennsylvania and New York, scale 1:250,000, *U.S. Geol. Surv. Misc. Invest. Ser. Map I-1715*, 1987.
Machette, M. N., S. F. Personius, A. R. Nelson, D. P. Schwartz, and W. R. Lund, The Wasatch fault zone, Utah—Segmentation and history of Holocene earthquakes, *J. Struct. Geol.*, *13*, 137-149, 1991.
Manspeizer, W., Early Mesozoic basins of the central Atlantic passive margin, in Geology of Passive Continental Margins: History, Structure and Sedimentologic Record (With Special Emphasis on the Atlantic Margin), *AAPG Educ. Course Note Ser., 19*, pp. 4-1 to 4-60, 1981.
Manspeizer, W. (Ed.), *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, 998 pp., Elsevier, New York, 1988.
Manspeizer, W., J. de Boer, J. K. Costain, A. J. Froelich, C. Çoruh, P. E. Olsen, G. J. McHone, J. H. Puffer, and D. C. Prowell, Post-Paleozoic activity, in *The Geology of North America*, vol. F-2, *The Appalachian-Oachita Orogen in the United States*, edited by R. D. Hatcher, Jr., W. A. Thomas, and G. W. Viele, pp. 319-374, Geological Society of America, Boulder, Colo., 1989.
Marrett, R., and R. W. Allmendinger, Estimates of strain due to brittle faulting: sampling of fault populations, *J. Struct. Geol.*, *13*, 735-738, 1991.
Marrett, R., and R. W. Allmendinger, Amount of extension on “small” faults: An example from the Viking graben, *Geology*, *20*, 47-50, 1992.
McBride, J. H., K. D. Nelson, and L. D. Brown, Evidence and implications of an extensive early Mesozoic rift basin and basalt/diabase sequence beneath the southeast Coastal Plain, *Geol. Soc. Am. Bull.*, *101*, 512-520, 1989.
McDonald, N. G., and P. M. LeTourneau, Revised paleogeographic model for Early Jurassic deposits, Connecticut valley: Regional easterly paleoslopes and internal drainage in an asymmetrical extensional basin, *Geol. Soc. Am. Abstr. Programs*, *20*, 54, 1990.
McHone, J. G., Tectonic and paleostress patterns of Mesozoic intrusions in eastern North America, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 607-620, Elsevier, New York, 1988.
McLaughlin, D. B., Newly recognized folding in the Triassic of Pennsylvania, *Pa. Acad. Sci. Proc.*, *37*, 156-159, 1963.
Meyertons, C. T., Triassic formations of the Danville Basin: *Va. Div. Min. Res. Rep. Invest.*, *6*, 65 pp., 1963.
Mickus, K. L., C. L. V. Aiken, W. J. Peeples, and D. G. Ziegler, A gravity and magnetic study of the Triassic Richmond basin, Virginia, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 401-421, Elsevier, New York, 1988.
Morley, C. K., R. A. Nelson, T. L. Patton, and S. G. Munn, Transfer zones in the East African rift system and their relevance to hydrocarbon exploration in rifts, *AAPG Bull.*, *74*, 1234-1253, 1990.
Muraoka, H., and H. Kamata, Displacement distribution along minor fault traces, *J. Struct. Geol.*, *5*, 483-495, 1983.
Nelson, R. A., T. L. Patton, and C. K. Morley, Rift-segment interaction and its relation to hydrocarbon exploration in continental rift systems: *AAPG Bull.*, *76*, 1153-1169, 1992.
Olsen, P. E., Fossil great lakes of the Newark Supergroup in New Jersey, in *Field Studies of New Jersey Geology and Guide to Field Trips*, edited by W. Manspeizer, 52nd Ann. Mtg. New York State Geol. Assoc., pp. 352-398, Rutgers University, Newark, 1980a.
Olsen, P. E., The latest Triassic and Early Jurassic formations of the Newark basin (eastern North America, Newark Supergroup): Stratigraphy, structure, and correlation, *N.J. Acad. Sci. Bull.*, *25*, 25-51, 1980.
Olsen, P. E., Continuity of strata in the Newark and Hartford basins of the Newark Supergroup, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 6-18, 1988.
Olsen, P. E., and R. W. Schlische, Transtensional arm of the early Mesozoic Fundy rift basin: Penecontemporaneous faulting and sedimentation, *Geology*, *18*, 695-698, 1990.
Olsen, P. E., A. J. Froelich, D. L. Daniels, J. P. Smoot, and P. J. W. Gore, Rift basins of early Mesozoic age, in *Geology of the Carolinas: Carolina Geological Society 50th Anniversary Volume*, edited by J. W. Horton and V. A. Zullo, pp. 142-170, University of Tennessee Press, Knoxville, 1990.
Olsen, P. E., R. W. Schlische, and P. J. W. Gore (Eds.), *Tectonic, Depositional, and Paleoecological History of Early Mesozoic Rift Basins of Eastern North America, International Geological Congress Field Trip T-351*, 174 pp., AGU, Washington, D. C., 1989.
Parker, R. A., H. F. Houghton, and R. C. McDowell, Stratigraphic framework and distribution of early Mesozoic rocks of the northern Newark basin, New Jersey and New York, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 31-39, 1988.
Peacock, D. C. P., and D. J. Sanderson, Displacements, segment linkage and relay ramps in normal fault zones, *J. Struct. Geol.*, *13*, 721-733, 1991.
Plint, A. G., and H. W. van de Poll, Structural and sedimentary history of the Quaco Head area, southern New Brunswick, *Can. J. Earth Sci.*, *21*, 753-761, 1984.
Puffer, J. H., and A. R. Philpotts, Eastern North American quartz tholeiites: geochemistry and petrology, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 579-605, Elsevier, New York, 1988.
Puffer, J. H., and P. C. Ragland (Eds.), Eastern North American Mesozoic Magmatism, *Geol. Soc. Am. Spec. Pap. 268*, 365 pp., 1992.
Ramsay, J. G., and M. I. Huber, *The Techniques of Modern Structural Geology; Volume 2: Folds and Fractures*, 700 pp., Academic, San Diego, Calif., 1987.
Randazzo, A. F., and R. E. Copeland, The geology of the northern portion of the Wadesboro Triassic basin, North Carolina, *Southeastern Geol.*, *17*, 115-138, 1976.
Ratcliffe, N. M., and W. C. Burton, Fault reactivation models for the origin of the Newark basin and studies related to U.S. eastern seismicity, in Proceedings of the Second U.S. Geological Survey Workshop on the Early Mesozoic Basins of the Eastern United States, edited by G. R. Robinson, Jr. and A. J. Froelich, *U.S. Geol. Surv. Circ. 946*, pp. 36-45, 1985.
Ratcliffe, N. M.,and W. C. Burton, Structural analysis of the Furlong fault and the relationship of mineralization to faulting and diabase intrusion, Newark basin, Pennsylvania, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 176-193, 1988.
Ratcliffe, N. M., W. C. Burton, R. M. D’Angelo, and J. K. Costain, Low-angle extensional faulting, reactivated mylonites, and seismic reflection geometry of the Newark basin margin in eastern Pennsylvania, *Geology*, *14*, 766-770, 1986.
Reinemund, J. A., Geology of the Deep River coal field, North Carolina, *U.S. Geol. Surv. Prof. Pap. 246*, 159 pp., 1955.
Reynolds, D. J., P. E. Olsen, M. S. Steckler, and C. F. Burgess, Structural framework of the Newark basin, *Eos Trans. AGU*, *71*, 1605-1606, 1990.
Reynolds, D. J., and R. W. Schlische, Comparative studies of continental rift systems, *Geol. Soc. America Abstr. Programs*, *21*, 61, 1989.
Rodgers, J. (Compiler), Bedrock geological map of Connecticut, scale 1:125,000, Connecticut Geol. Nat. History Surv., Hartford, 1985.
Root, S. I., Structure and hydrocarbon potential of the Gettysburg basin, Pennsylvania and Maryland, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 353-367, Elsevier, New York, 1988.
Root, S. I., Basement control of structure in the Gettysburg rift basin, Pennsylvania and Maryland, *Tectonophysics*, *166*, 281-292, 1989.

Rosendahl, B. R., Architecture of continental rifts with special reference to East Africa, *Ann. Rev. Earth Planet. Sci.*, *15*, 445-503, 1987.

Rosendahl, B. R., Continental rifts: Structural traits, in *Encyclopedia of Geophysics*, edited by D. E. James, pp. 104-126, Van Nostrand Reinold, New York, 1990.

Rosendahl, B. R., E. Kilembe, and K. Kaczmarick, Comparison of the Tanganyika, Malawi, Rukwa, and Turkana rift zones from analyses of seismic reflection data, *Tectonophysics*, *213*, 235-256, 1992.

Sander, S. D., and B. R. Rosendahl, The geometry of rifting in Lake Tanganyika, *J. Afr. Earth Sci.*, *8*, 323-354, 1989.

Sanders, J. E., Late Triassic tectonic history of northeastern United States, *Am. J. Sci.*, *261*, 501-524, 1963.

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

Schlische, R. W., Half-graben basin filling models: New constraints on continental extensional basin development, *Basin Res.*, *3*, 123-141, 1991.

Schlische, R. W., Structural and stratigraphic development of the Newark extensional basin, eastern North America: Evidence for the growth of the basin and its bounding structures, *Geol. Soc. Am. Bull.*, *104*, 1246-1263, 1992.

Schlische, R. W., and P. E. Olsen, Structural evolution of the Newark basin, in *Geology of the Central Newark Basin, Field Guide and Proceedings*, edited by J. M. Husch and M. J. Hozik, pp. 44-65, Fifth Ann. Mtg. Geol. Assoc. N.J., Rider College, Lawrenceville, N.J., 1988.

Schlische, R. W., and P. E. Olsen, Quantitative filling model for continental extensional basins with applications to the early Mesozoic rifts of eastern North America, *J. Geol.*, *98*, 135-155, 1990.

Schwartz, D. P., and K. J. Coppersmith, Fault behavior and characteristic earthquakes: examples from the Wasatch and San Andreas fault zones, *J. Geophys. Res.*, *89*, 5681-5698, 1984.

Scott, D. L., M. A. Etheridge, and B. R. Rosendahl, Oblique slip deformation in extensional terrains: A case study of the Lakes Tanganyika and Malawi rift zones: *Tectonics*, *11*, 998-1012, 1992.

Sheridan, R. E., and J. A. Grow (Eds.), *The Geology of North America*, vol. I-2, *The Atlantic Continental Margin—U.S.*, 610 pp., Geological Society of America, Boulder, Colo., 1988.

Smoot, J. P., The closed basin hypothesis and the use of working models in facies analysis of the Newark Supergroup, in *Proceedings of the Second U.S. Geological Survey Workshop on the Early Mesozoic Basins of the Eastern United States*, edited by G. R. Robinson, Jr. and A. J. Froelich, *U.S. Geol. Surv. Circ. 946*, pp. 4-10, 1985.

Smoot, J. P., Sedimentary facies and depositional environments of early Mesozoic Newark Supergroup basins, eastern North America, *Palaeogeogr., Palaeoclimatol., Paleoecol.*, *84*, 369-423, 1991.

Sumner, J. R., Geophysical investigation of the structural framework of the Newark-Gettysburg Triassic basin, Pennsylvania, *Geol. Soc. Am. Bull.*, *88*, 935-942, 1977.

Sutter, J. F., Innovative approaches to dating igneous events in the early Mesozoic basins, eastern North America, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 194-200, 1988.

Swanson, M. T., Preexisting fault control for Mesozoic basin formation in eastern North America, *Geology*, *14*, 419-422, 1986.

Tankard, A. J., and H. R. Balkwill, Extensional tectonics and stratigraphy of the North Atlantic margins: Introduction, in *Extensional Tectonics and Stratigraphy of the North Atlantic Margins*, edited by A. J. Tankard and H. R. Balkwill, *AAPG Memoir 46*, pp. 1-6, 1989.

Taylor, K. G., and R. Ressetar, Depositional and structural histories of the Richmond and Taylorsville Triassic rift basins [abstract], *AAPG Bull.*, *69*, 1449, 1985.

Thayer, P. A., Stratigraphy and geology of Dan River Triassic basin, North Carolina, *Southeastern Geol.*, *12*, 1-31, 1970.

Unger, J. D., A simple technique for analysis and migration of seismic reflection profiles from the Mesozoic basins of eastern North America, in Studies of the Early Mesozoic Basins of the Eastern United States, edited by A. J. Froelich and G. R. Robinson, Jr., *U.S. Geol. Surv. Bull. 1776*, pp. 229-235, 1988.

Van Houten, F. B., Triassic-Liassic deposits of Morocco and eastern North America: Comparison, *AAPG Bull.*, *61*, 79-94, 1977.

Venkatakrishnan, R., and R. Lutz, A kinematic model for the evolution of the Richmond Triassic basin, in *Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins*, edited by W. Manspeizer, pp. 445-462, Elsevier, New York, 1988.

Versfelt, J., and B. R. Rosendahl, Relationships between pre-rift structure and rift architecture in Lakes Tanganyka and Malawi, East Africa, *Nature*, *337*, 354-357, 1989.

Walsh, J. J., and J. Watterson, Distributions of cumulative displacement and seismic slip on a single normal fault surface, *J. Struct. Geol.*, *9*, 1039-1046, 1987.

Walsh, J. J., and J. Watterson, Analysis of the relationship between displacements and dimensions of faults, *J. Struct. Geol.*, *10*, 239-247, 1988.

Walsh, J. J., and J. Watterson, Geometric and kinematic coherence and scale effects in normal fault systems, in *The Geometry of Normal Faults*, edited by A. M. Roberts, G. Yielding, and B. Freeman, *Geol. Soc. Spec. Publ. 56*, pp. 193-203, 1991.

Watterson, J., Fault dimensions, displacements and growth, *Pure Appl. Geophys.*, *124*, 365-373, 1986.

Wenk, W. J., Seismic refraction model of fault offset along basalt horizons in the Hartford rift basin, Connecticut and Massachusetts, *Northeastern Geol.*, *6*, 168-173, 1984a.

Wenk, W. J., Seismic refraction model of depth of basement in the Hartford rift basin, Connecticut and Massachusetts, *Northeastern Geol.*, *6*, 196-202, 1984b.

Wernicke, B., and B. C. Burchfiel, Modes of extensional tectonics, *J. Struct. Geol.*, *4*, 105-115, 1982.

Wheeler, G., Triassic fault-line deflections and associated warping, *J. Geol.*, *47*, 337-370, 1939.

Wheeler, W. H., The Livingstons Mountains border fault system, Lake Nyasa (Malawi), East Africa: A case study of an oblique slip rift basin border fault from onshore and sub-surface perspectives, M.S. thesis, Duke Univ., Durham, N.C., 1989.

Wise, D. U., Dip domain method applied to the Mesozoic Connecticut Valley rift basins, *Tectonics*, *11*, 1357-1368, 1992.

Withjack, M. O., and D. J. Drickman Pollock, Synthetic seismic reflection profiles of rift-related structures: *AAPG Bull.*, *68*, 1160-1178, 1984.

Withjack, M. O., M. H. Link, and P. E. Olsen, Structure, stratigraphy, and climate of the Mesozoic Chignecto subbasin, Bay of Fundy, Canada [abstract], *AAPG Bull.*, *75*, 695, 1991.

Woodcock, N. H., and M. Fischer, Strike-slip duplexes, *J. Struct. Geol.*, *8*, 725-735, 1986.

Zablocki, F. S., A gravity study in the Deep River-Wadesboro Triassic basin of North Carolina, M.S. thesis, University of North Carolina, Chapel Hill, 1959.

Zen, E. (Ed.), Bedrock Geologic Map of Massachusetts, scale 1:250,000, U.S. Geol. Surv., Washington, D. C., 1983.

Zhang, P., D. B. Slemmons, and F. Mao, Geometric pattern, rupture termination and fault segmentation of the Dixie Valley-Pleasant Valley active normal fault system, Nevada, U.S.A., *J. Struct. Geol.*, *13*, 165-176, 1991.

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R. W. Schlische, Department of Geological Sciences, Rutgers University, Busch Campus, Piscataway, N.J. 08855-1179.

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TABLE 1. Features of exposed Mesozoic rift basins in eastern North America

Feature	FUNDY: Fundy-Chignecto	FUNDY: Minas	Connecticut Valley	Pomperaug	Newark	Narrow Neck	Gettysburg	Culpeper	Richmond	Dan River-Danville	Deep River
Basin trend	048°	093°	010°	012°	050°	080°	042°	020° in N 043° in S	016°	043° in N 066° in S	034° in N 059° in S
Maximum length	356 km	143 km	155 km	11 km	190 km	47 km	115 km	142 km	53 km	165 km	235 km
Maximum width	77 km	31 km	31 km	4 km	50 km	7 km	35 km	26 km	15 km	10 km	25 km
Estimated max. depth	3.5 km	1.0 km	5.5 km (3,4)	1.5 km	7.0 km	4.0 km	5.0 km	5.0 km	2.8 km (5)	1.8 km	2.6 km
General dip direction	NW	Variable	ESE	ESE	NW	N	NW	WNW	WNW	NW	SE
Type of border fault system (BFS)[a]	Dip slip	Left-oblique slip	Mostly dip slip	Mostly dip slip	Mostly dip slip	Left-oblique slip	Mostly dip slip	Mostly dip slip	Mostly dip slip	Mostly dip slip	Mostly dip slip
Strike of BFS	051°	072°	011°	011°	047°	082°	026°	015° in N 032° in S	019°	035° in N 060° in S	023° in N 049° in S
Dip of BFS	40°SE (6,7)	Steep (6,7)	35-40° WNW (4,8)	?	70°SE in NE; 30°SE in SW (9)	Steep (?)	60-70°SE (10)	60° (11)	Hfz[d]: 42°E (12) Steep (13)	44°SE (14)	40°NW (15)
Reactivated BFS[b]	Yes (6,7,16)	Yes (6,7,17)	Yes	?	Yes (18)	?	Yes (19)	Yes	Yes (12)	Yes	?
Listric BFS	Yes (6,20)	No	Inferred yes (4,21,22)	?	No; perhaps kinked (23)	?	Inferred yes (10)	Yes (11)	Inferred yes (24)	?	Yes (15)
Relay ramps	Yes	No	No	No	Yes	No	No	Yes	Yes	No	Yes
Rider blocks	Yes	No	Yes	?	Yes	?	Yes	No ?	Yes	Yes	Yes
Transverse folds	Yes (20)	No	Yes (25,26)	No	Yes (23,25)	No	Yes (27)	Yes (28)	No ?	Yes (29)	Yes (15,30)
Longitudinal geometry	?	Complex	2 synclinal subbasins	Single syncline	Faulted syncline	Complex	Single syncline	Single syncline	Single syncline	Multiple synclinal subbasins	4 synclinal subbasins
Ramping margin	Overlap	Overlap	Overlap; antithetic faults	Overlap	Overlap; minor synthetic and antithetic faults	Overlap	Overlap; synthetic and antithetic faults	Overlap; antithetic faults	Overlap	Overlap	Overlap; synthetic faults
Mean strike of intrabasinal faults	031° (mostly synthetic)	055°	029° (synthetic); 334°	064°; 344°	030° (mostly synthetic); 283°	037°; 315°	047° (synthetic and antithetic); 330°	041° (synthetic and antithetic)	000° (synthetic and antithetic)	072° (synthetic)	057° (synthetic)
Mean trend of dikes	060°	——	037°	——	031°; 319°	056°	020°	013°	339°	348°	348°
Age of deposits (1,2)	Ladinian-Pleinsbachian	Anisian-Pleinsbachian	L. Carnian-Pleinsbachian	L. Carnian-Hettangian	M. Carnian-Sinemurian	?	M. Carnian-Hettangian	L. Carnian-Hettangian	E. Carnian-M. Carnian	M. Carnian-E. Norian	E. Carnian-L. Carnian
Strat. succession[c]	F-L-LS-B-LS	F-L-LS-B-LS	F-B-LS-F	F-B-LS	F-LD-LS-B-LS	F	F-L	F-L-F-B	F-LD-LS-F	F-LD-LS-F	F-L-F
Onlap	?	Tranverse (2,31,32)	?	?	Transverse (23,32); Longitudinal	?	?	?	Transverse (2,32)	Longitudinal	?

SOURCES: 1, Cornet and Olsen, 1985; 2, Olsen et al., 1989; 3, Wenk, 1984b; 4, Wise, 1992; 5, Mickus et al., 1988; 6, Olsen and Schlische, 1990; 7, Withjack et al., 1991; 8, Chandler, 1978; 9, Ratcliffe and Burton, 1985; 10, Root, 1988, 1989; 11, Manspeizer et al., 1989; 12, Venkatakrishnan and Lutz, 1988; 13, Goodwin and Farrell, 1979; 14, C.R. Halladay, 1985, as cited by Olsen et al., 1990; 15, Olsen et al., 1990; 16, Plint and van de Poll, 1984; 17, Keppie, 1982; 18, Ratcliffe et al., 1986; 19, Root, 1989; 20, Schlische, 1990; 21, Zen, 1983; 22, de Boer and Clifford, 1988; 23, Schlische, 1992; 24, Taylor and Ressetar, 1985; 25, Wheeler, 1939; 26, Withjack and Drickman Pollock, 1984; 27, Berg, 1980; 28, Leavy et al., 1983; 29, Thayer, 1970; 30, Reinemund, 1955; 31, Hubert and Forlenza, 1988; 32, Schlische and Olsen, 1990.

[a]During Triassic-Early Jurassic rifting.

[b]Mostly on the basis of Swanson [1986 and references therein]. More recent citations given in individual entries.

[c]Abbreviations are B, basalt flows and interbedded lacustrine strata; F, fluvial strata; L, lacustrine strata; LD, mostly deep-water lacustrine strata; LS, mostly shallow-water lacustrine strata; modified from Olsen et al. [1989] and Schlische [1991].

[d]Hylas fault zone.

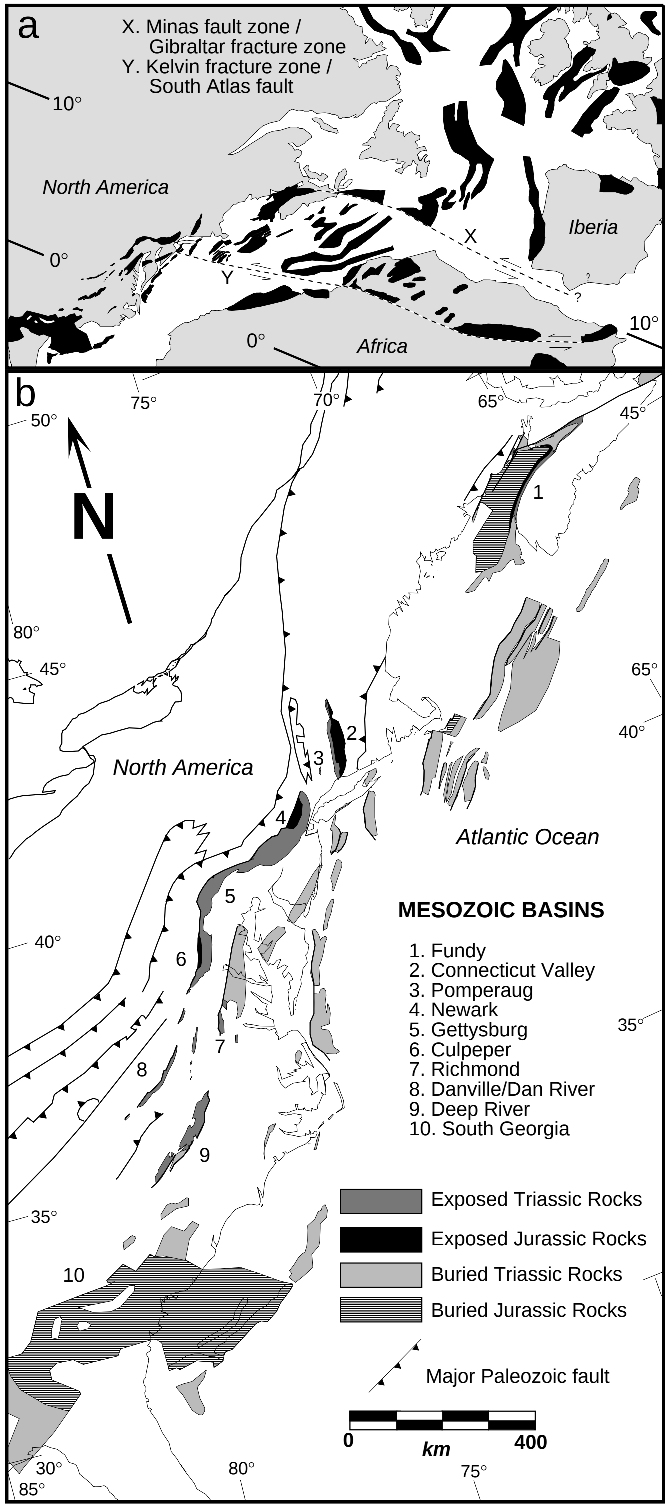

Fig. 1. (a) Rift basins (black) of the circum-North Atlantic shown on a predrift reconstruction. Modified from Tankard and Balkwill [1989] and Olsen and Schlische [1990]. (b) Mesozoic basins of eastern North America. Note the general parallelism between the basins and Paleozoic structures. The Jurassic rocks of the South Georgia basin are postrift deposits [McBride et al., 1989] and thus extend beyond the boundaries of some of the synrift basins. Modified from Froelich and Olsen [1984], Klitgord et al. [1988], and Olsen et al. [1990], and Schlische [1990].

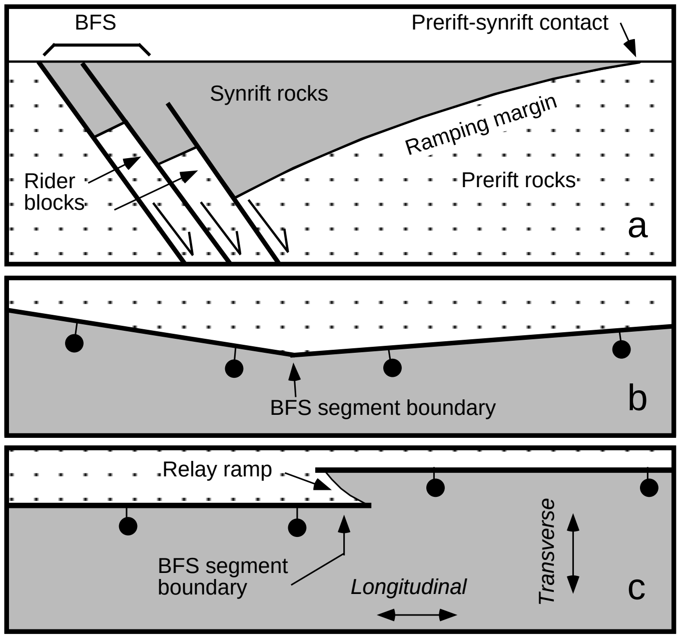

Fig. 2. Nomenclature of rift basin structures as seen in cross section (a) and map views (b and c). BFS refers to the border fault system. See text for discussion.

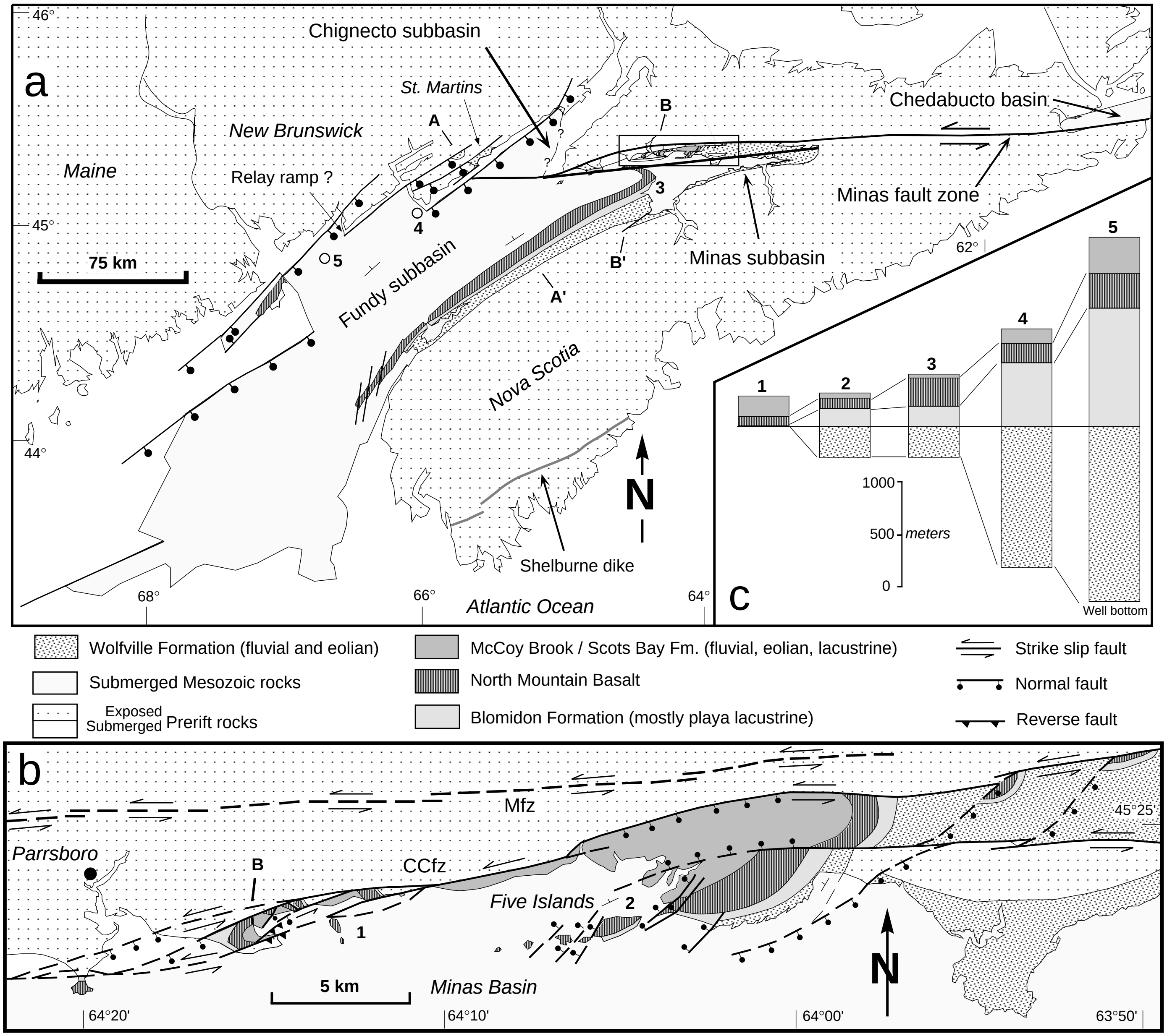

Fig. 3. (a) Geologic map of the Fundy basin. Rectangle shows area detailed in Figure 3b. Open circles are wells. Geometry of Chignecto subbasin is schematic. (b) Geologic map of Five Islands region, illustrating the structure of the northern margin of the Minas subbasin. Note the presence of extensional strike-slip duplexes. The geometry of dashed faults is in part schematic. Mfz is the Minas fault zone. (c) Stratigraphic sections from several localities within the Fundy basin, locations of which are shown by the numbers in Figures 3a and 3b. Modified from Keppie [1979], Donohoe and Wallace [1982], and Olsen and Schlische [1990].

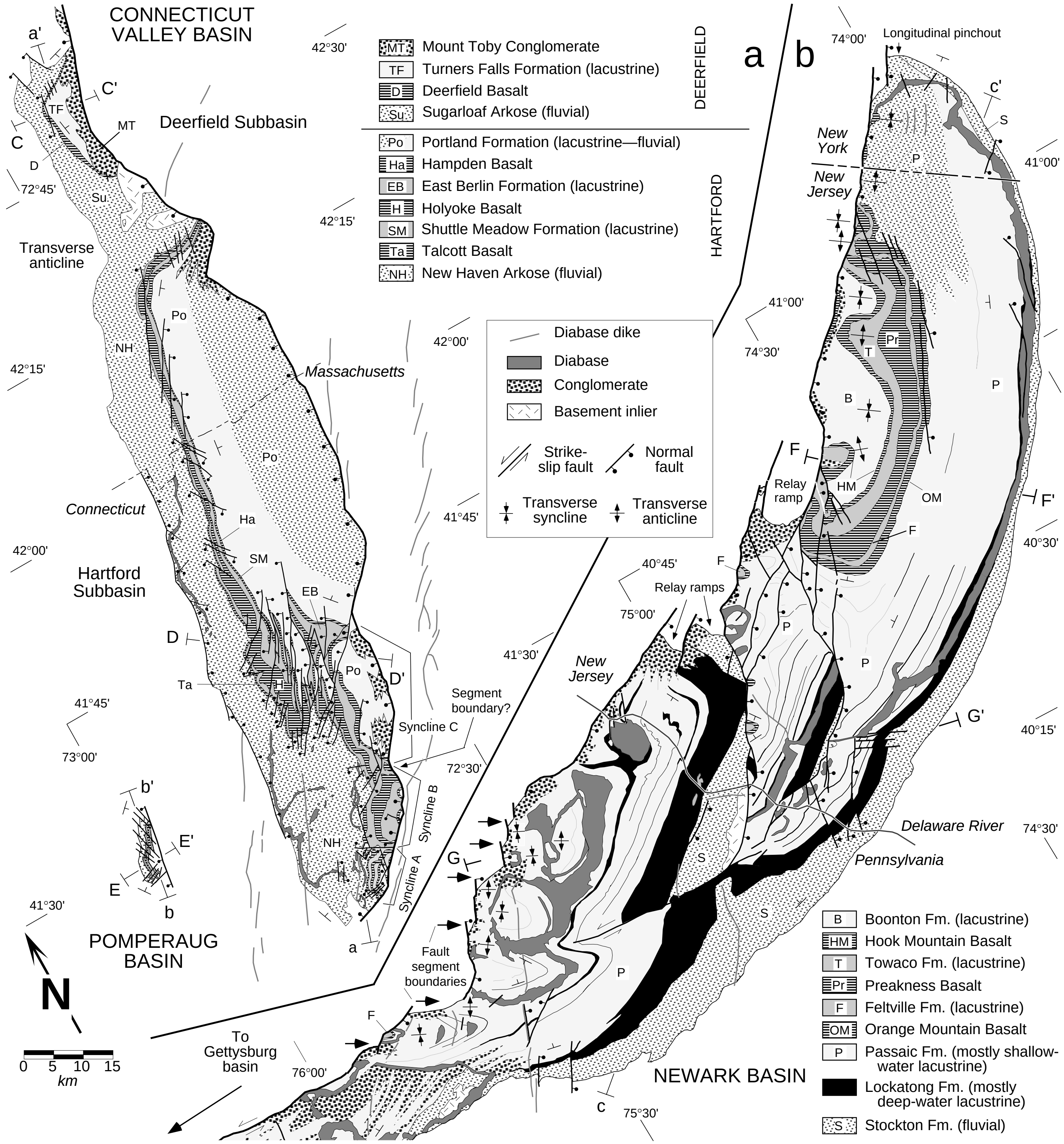

Fig. 4. (a) Geologic map of the Connecticut Valley and Pomperaug basins. Strata in the Pomperaug basin are assigned the same formations as in the Hartford basin. Compiled from Zen [1983] and Rodgers [1985]. (b) Geologic map of the Newark basin. Thin black lines are distinctive clusters of deeper-water lacustrine strata within the Passaic Formation; thin gray lines depict form lines of bedding. Modified from Schlische [1992] based on Berg [1980], Olsen [1980b], Ratcliffe et al. [1986], Lyttle and Epstein [1987], Parker et al. [1988], Ratcliffe and Burton [1988], and Schlische and Olsen [1988].

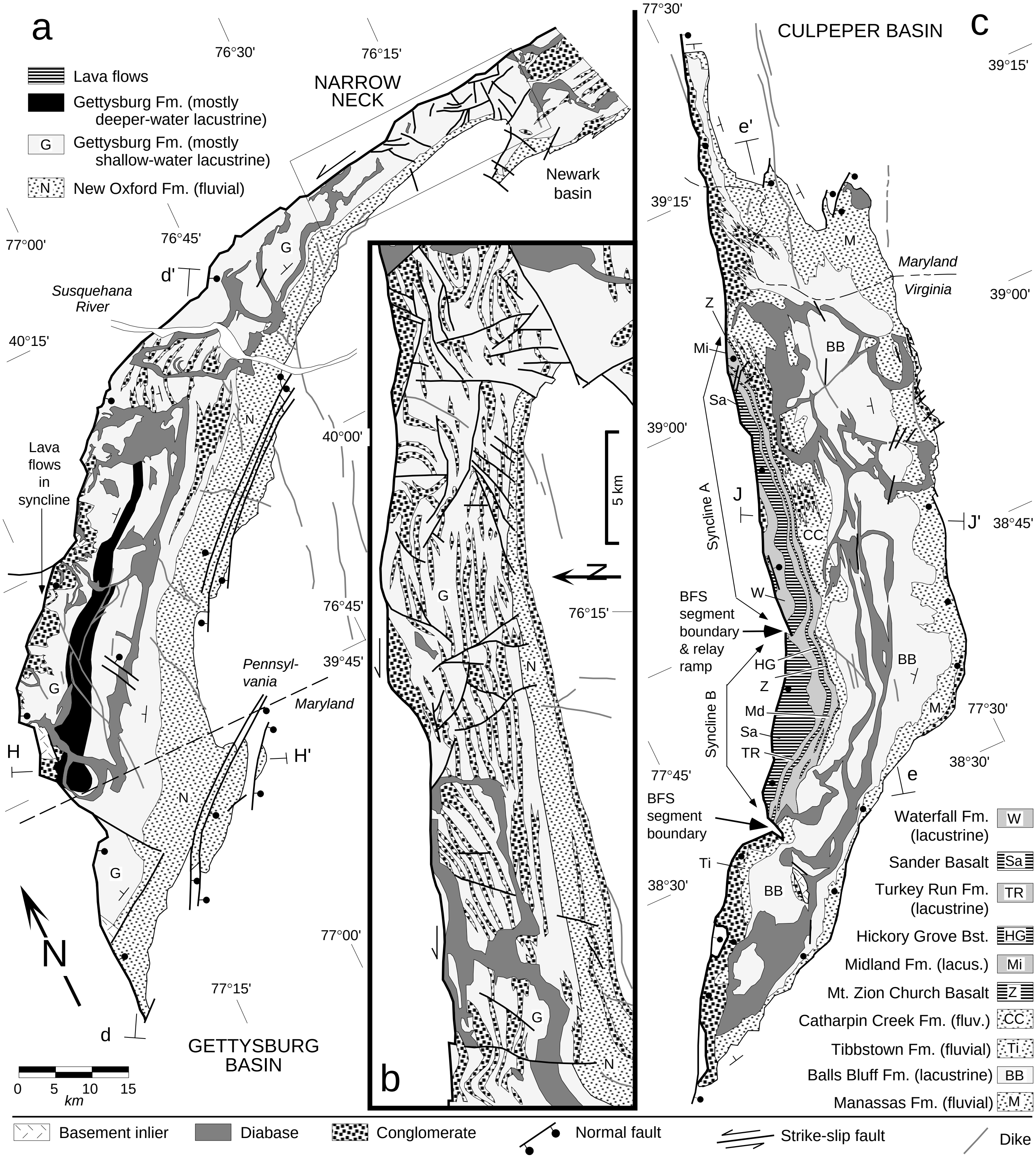

Fig. 5. (a) Geologic map of the Gettysburg basin. Rectangle shows area enlarged in (b). Compiled from Berg [1980] and Root [1988]. (b) Geologic map of the Narrow Neck between the Newark and Gettysburg basin. Modified from Berg [1980]. (c) Geologic map of the Culpeper basin. Modified from Leavy et al. [1983].

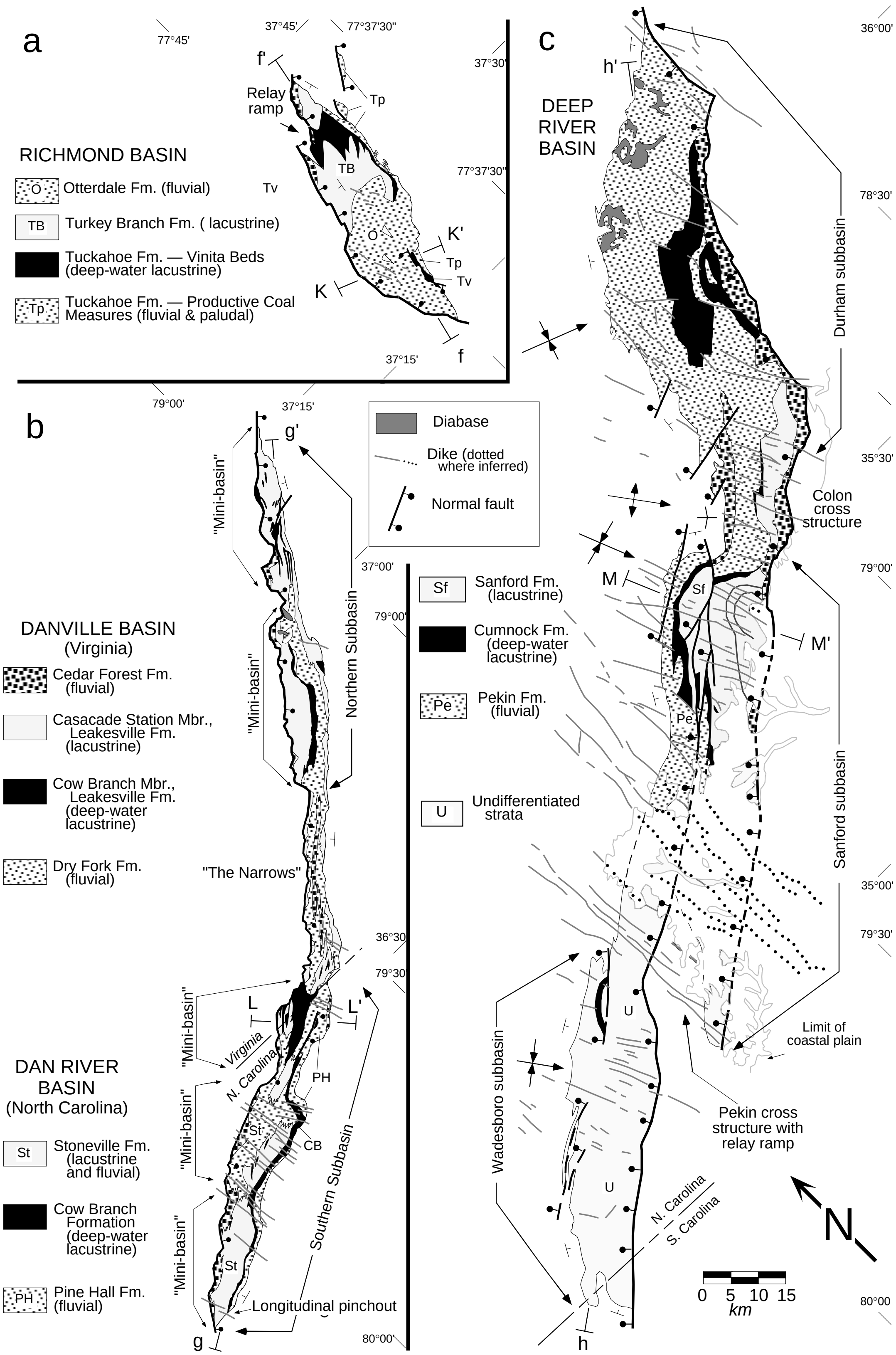

Fig. 6. (a) Geologic map of the Richmond basin and associated outlier basins. Modified from Olsen et al. [1989]. (b) Geologic map of the Danville-Dan River basin. Compiled from Meyertons [1963] and Thayer [1970]. (c) Geologic map of the Deep River basin. Compiled from Bain and Harvey [1977] and Brown et al. [1985].

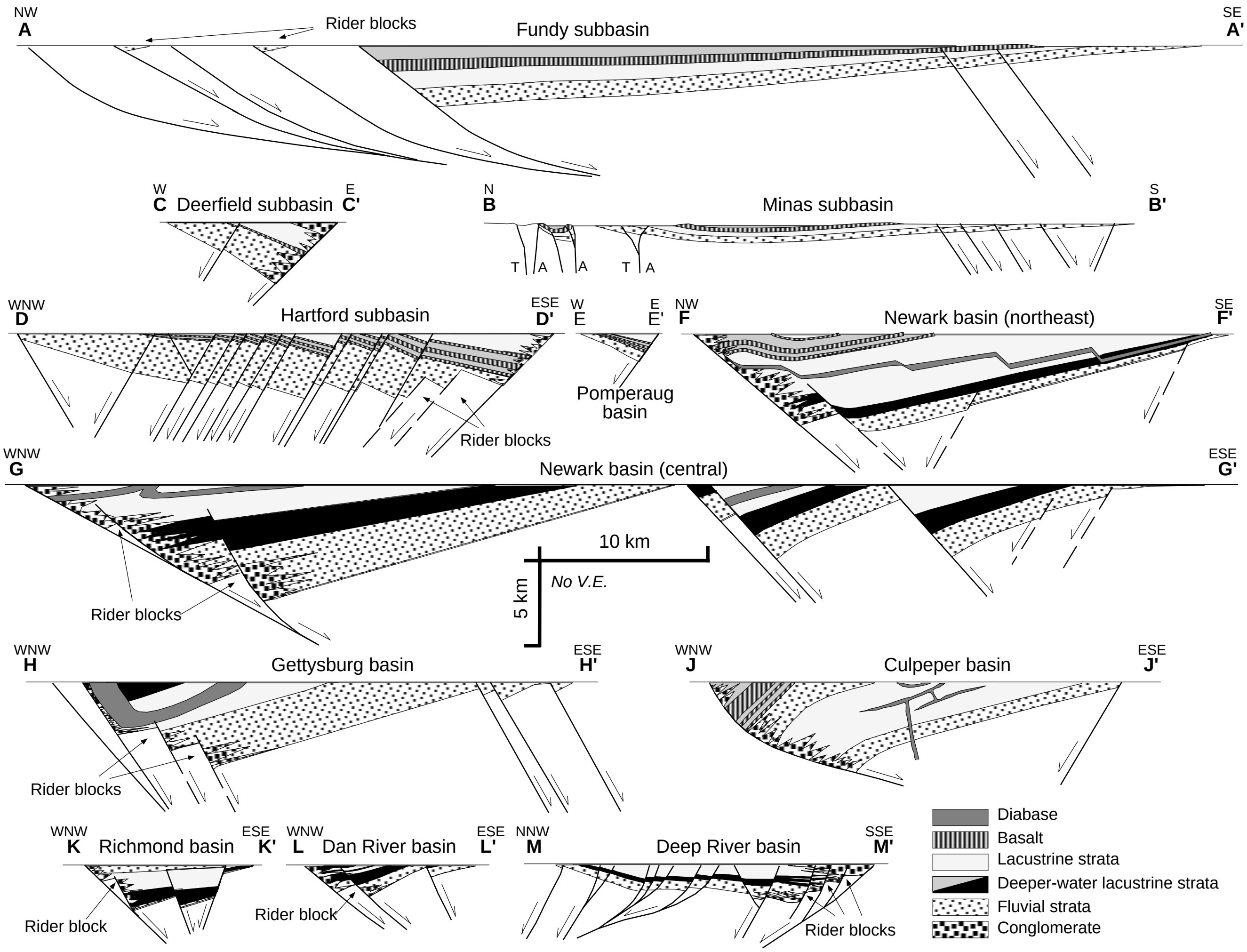

Fig. 7. Same-scale cross sections of exposed basins in eastern North America. Locations of cross sections shown in Figures 3-6. The geometry of the basins at depth is, in part, conjectural. Note the general half-graben geometry of most basins, the presence of rider blocks, and the mostly synthetic intrabasinal faults. Fundy basin section A-A' based partly on seismic reflection data [after Olsen and Schlische, 1990]; Newark basin sections modified from Schlische [1992]; Gettysburg basin section based partly on Root [1988]; Culpeper basin section based on seismic reflection profiles [Manspeizer et al., 1989]; Richmond basin section based partly on well and seismic data [B. Cornet, personal communication, 1990]; Dan River-Danville and Deep River sections modified from Olsen et al. [1990]; Deep River section based on proprietary seismic data [P.E. Olsen, personal communication, 1990].

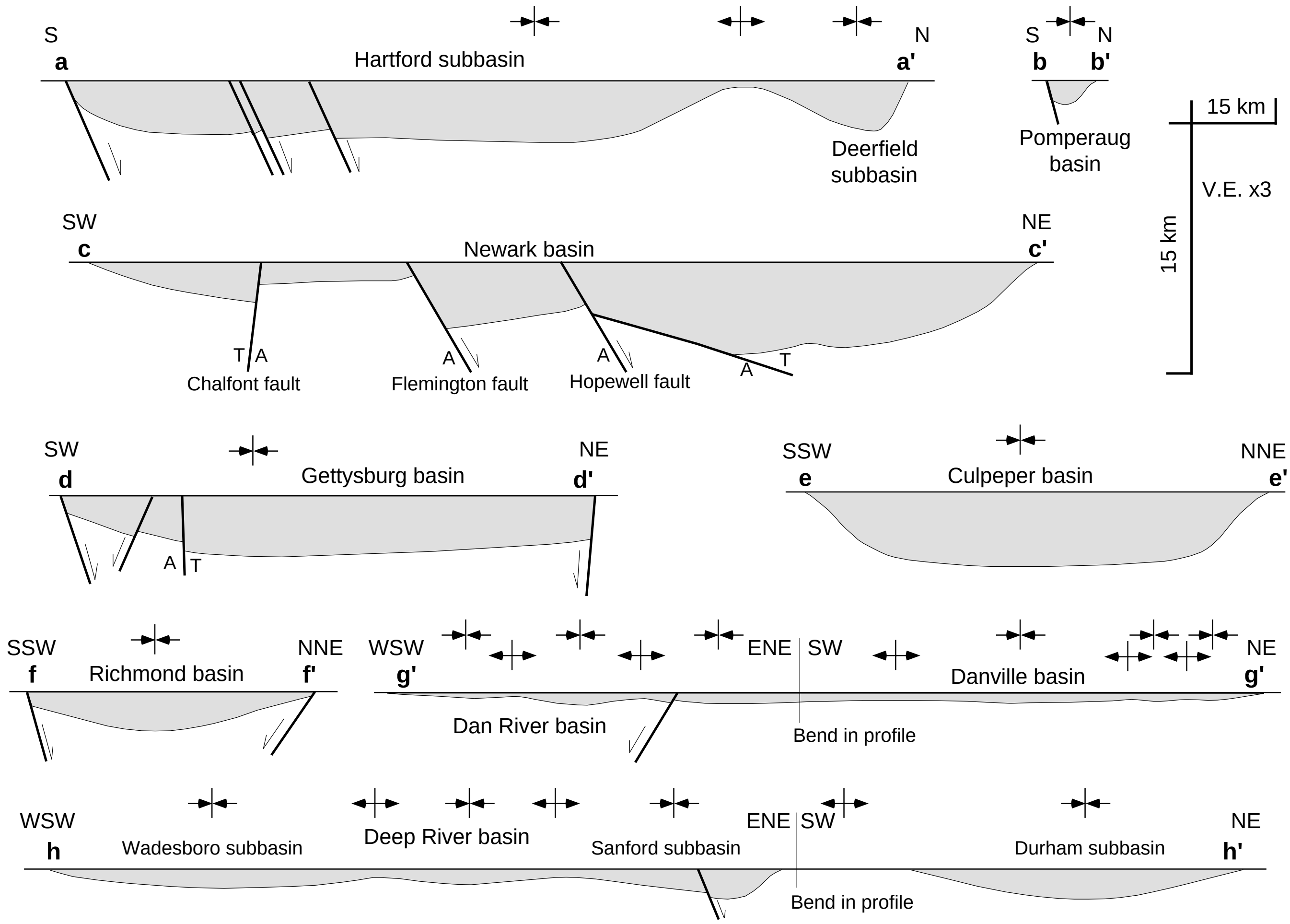

Fig. 8. Diagrammatic longitudinal cross sections of selected half graben (for locations, see Figures 4-6). Although complicated by intrabasinal faults, basins consist of either a single elongate syncline or multiple synclines separated by transverse anticlines.

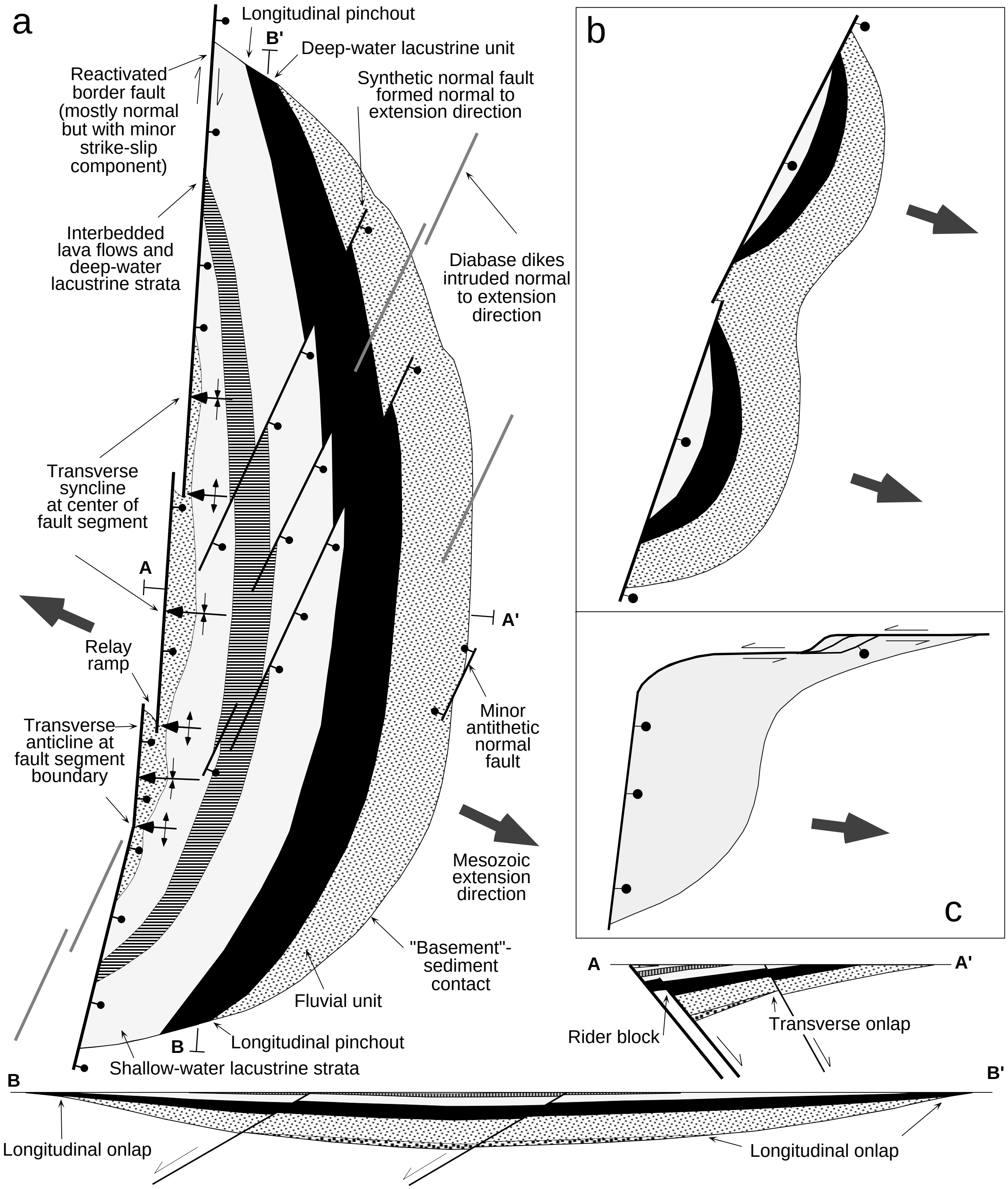

Fig. 9. (a) Geologic map and cross sections of an idealized, dip-slip-dominated Mesozoic rift basin. Note that although the trace of the prerift-synrift contact defines a large syncline, it is not affected by the transverse folds adjacent to the border fault system. (b) Geologic map of idealized Mesozoic basin containing multiple subbasins. (c) Geologic map of basin with both dip-slip and strike-slip-dominated margins.

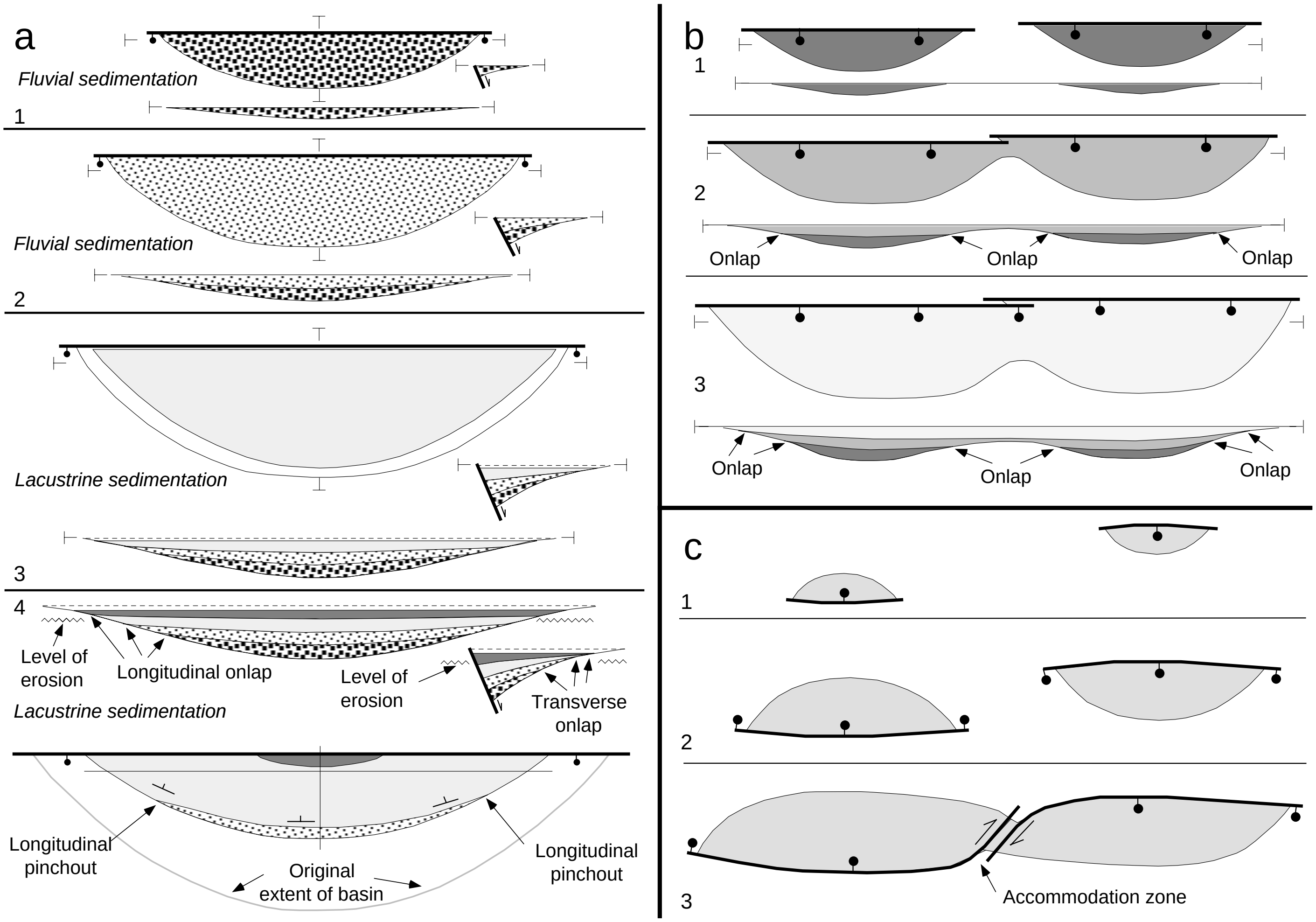

Fig. 10. (a) Model for the growth of a rift basin bounded by a single border fault system. The basin increases in depth, width, and length through time such that depositional surface area progressively increases, resulting in onlap geometries and a fluvial-lacustrine transition as the basin becomes sediment starved. Stage 4 shows a map view of the basin following erosional truncation. The oldest unit does not crop out at the surface, and the second oldest unit pinches out longitudinally. (b) Model for the growth of a basin complex containing two border fault segments. Early isolated basins merge as the fault tips propagate along strike. (c) Model for the growth and linkage of two basins bounded by oppositely dipping border fault segments; an accommodation zone forms in the zone of overlap.

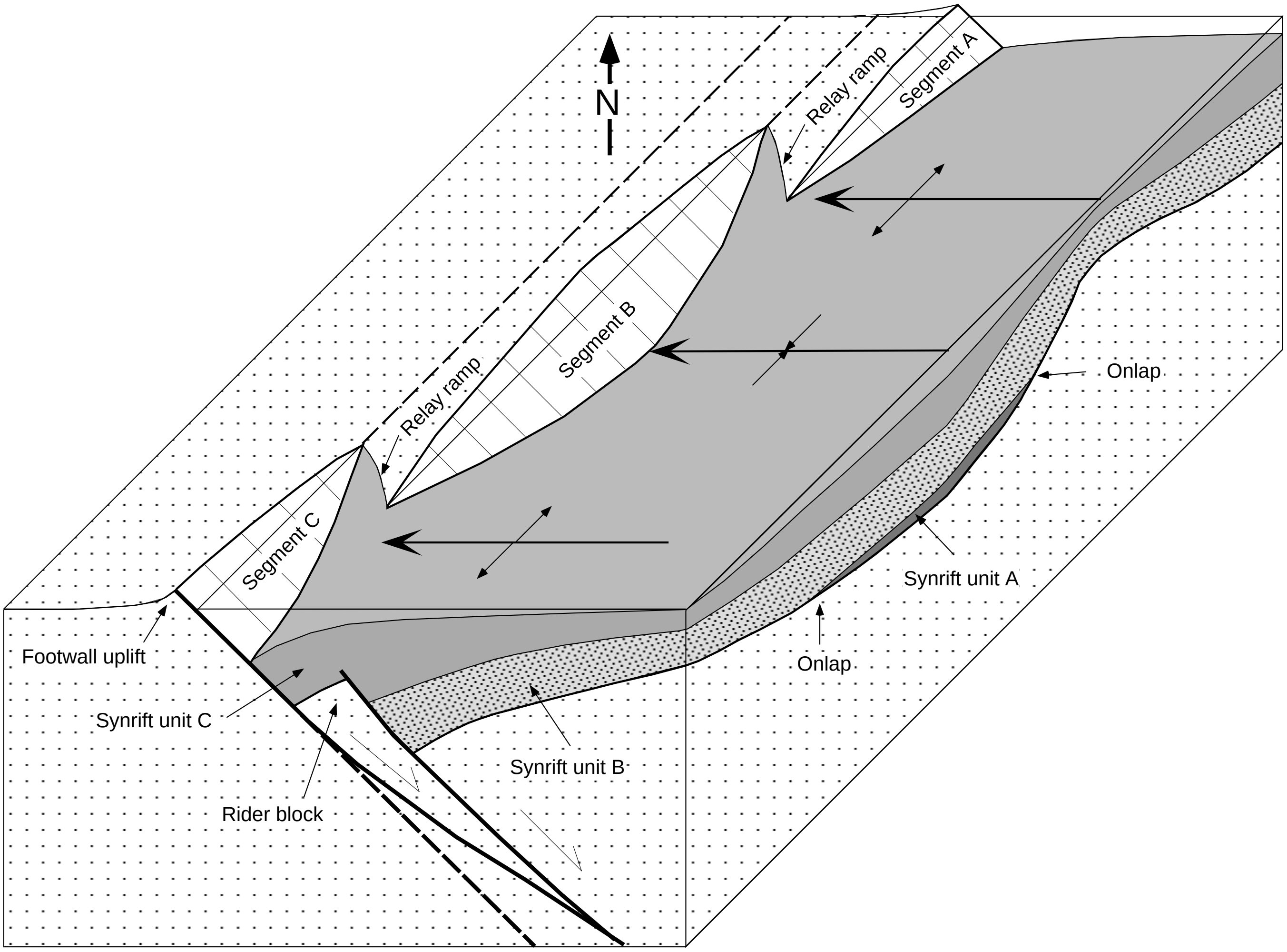

Fig. 11. Block diagram illustrating the geometric relationships among border fault segmentation, relay ramps, transverse folds, and rider blocks. Synrift unit A forms a restricted wedge, suggesting that fault segment B lengthened through time. Synrift unit B is absent from the hanging wall block of fault segment C, suggesting that segment C is younger than segment B. Segments B and C may merge at depth, forming a kinematically linked fault system, although this is not required. Faults (dashed) may extend from the ends of segments B and C if they involve only partial reactivation of preexisting structures. Partly on the basis of Ramsay and Huber [1987], Peacock and Sanderson [1991], and Schlische [1992].