Diachronous Rifting, Drifting, and Inversion on the Passive Margin of Central Eastern North America: An Analog for Other Passive Margins

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ABSTRACT

Integration of new data with existing information indicates that the tectonic development of the passive margin of eastern North America between the Carolina Trough and Scotian Basin was considerably more complex than the classic two-stage, rift-drift model. First, the transition from rifting to drifting was diachronous. In the southeastern United States, the rift-drift transition occurred after the Late Triassic synrift deposition and before eastern North America magmatism in the earliest Jurassic (~200 Ma). In maritime Canada, the rift-drift transition occurred after eastern North America magmatic activity and synrift deposition in the Early Jurassic and before postrift deposition in the early Middle Jurassic (~185 Ma). Second, the deformational regime changed substantially after rifting on both the southern and northern segments of the margin. Generally, northwest-southeast postrift shortening replaced northwest-southeast synrift extension. Northeast-striking reverse faults formed, and many of the rift-basin boundary faults had reverse displacements. In the southeastern United States, the change in the deformational regime occurred in the Late Triassic–Early Jurassic during the rift-drift transition. Simultaneously, diabase sills and dikes, many striking nearly perpendicular to the trend of the rift basins, intruded the continental crust; and a massive wedge of volcanic or volcaniclastic rocks developed near the continent–ocean boundary. In maritime Canada, the change in the deformational regime occurred during or after the Early Jurassic and before or during the Early Cretaceous; that is, during the rift-drift transition or early stages of sea-floor spreading.

INTRODUCTION

The geologic history of the passive margin of central eastern North America (i.e., from the Carolina Trough to the Scotian Basin) resembles that of many other passive margins. Indeed, it shares much of its geologic history with the petroliferous passive margins of the Grand Banks, northwest Africa, and western Europe. Like most passive margins, it developed in two stages: rifting and drifting (Figure 1). According to most workers, rifting occurred during the Middle to Late Triassic and continued into the Early Jurassic (e.g., Manspeizer and Cousminer, 1988; Olsen et al., 1989). Drifting, associated with the separation of North America and Africa and the creation of sea-floor-spreading centers in the Atlantic Ocean, began during the late Early to early Middle Jurassic and continues today (e.g., Klitgord and Schouten, 1986; Benson and Doyle, 1988; Klitgord et al., 1988; Welsink et al., 1989). Although Klitgord et al. (1988) acknowledged that the transition from rifting to drifting probably was not a perfectly synchronous event from the Carolina Trough to the Scotian Basin, they believed that it occurred during a short time span between the late Early Jurassic and early Middle Jurassic. To the south in the Blake Plateau region, the transition from rifting to drifting appears to have occurred slightly later in the Middle Jurassic (e.g., Klitgord et al., 1988). To the north in the Grand Banks region, the transition occurred much later in the Early Cretaceous (e.g., Srivastava and Tapscott, 1986). During the past 5–10 yr, new information has emerged about the passive margin of central eastern North America. Improved radiometric-dating...
techniques, geochemical correlations, and cyclostratigraphy have considerably constrained the timing and duration of the eastern North America magmatic activity (e.g., Sutter, 1988; Olsen et al., 1989; Dunning and Hodych, 1990; Hodych and Dunning, 1992; Ragland et al., 1992; Olsen, 1997). Field and seismic studies have documented the presence of inversion structures in the Fundy rift basin of southeastern Canada (Withjack et al., 1995). Deep seismic-reflection data have clearly imaged wedges of seaward-dipping reflections, inferred to be volcanic or volcanioclastic rocks, beneath the Baltimore Canyon and Carolina troughs (Austin et al., 1990; Sheridan et al., 1993; Oh et al., 1995). In this paper, we integrate this new information with existing geological and geophysical data from onshore and offshore eastern North America. The integration of these data provides new insights into the development of the passive margin of central eastern North America and other passive margins. Our work shows that the transition from rifting to drifting was diachronous for central eastern North America. Rifting ended and drifting began significantly earlier in the south (~200 Ma) than in the north (~185 Ma).

Our work also shows that, on both the southern and northern segments of the passive margin of central eastern North America, the tectonic regime changed substantially during the transition from rifting to drifting, or during the early stages of sea-floor spreading. Normal faulting ceased, reverse faults formed, and rift-basin boundary faults had reverse displacements.

**GEOLOGIC BACKGROUND**

**Paleozoic Orogenic Activity**

Orogenic activity associated with subduction, accretion, and collision occurred throughout eastern North America during much of the Paleozoic. The final collision, the late Paleozoic Alleghanian–Variscan orogeny, welded the North American and African continents and created the Pangean supercontinent (e.g., Rast, 1988; Rankin, 1994). Numerous gently to moderately dipping, basement-involved thrust faults formed during the various Paleozoic orogenies. For example, moderately dipping, large-amplitude reflectors in the Gulf of Maine
and Bay of Fundy are interpreted as thrust-fault zones associated with an Alleghanian–Variscan suture zone (Figures 2, 3A) (Brown, 1986; Hutchinson et al., 1988; Keen et al., 1991a; Withjack et al., 1995). Deep-seated, southeast-dipping reflectors beneath the continental shelf of offshore Virginia are interpreted as thrust-fault sequences associated with an Ordovician suture zone (Figure 3B) (Sheridan et al., 1993).

### Middle Triassic to Early Jurassic Rifting

Rift basins developed in eastern North America from the Middle Triassic to Early Jurassic during the breakup of the Pangean supercontinent (e.g., Manspeizer and Cousminer, 1988; Olsen et al., 1989; Schlische, 1993) (Figures 2–4). Strata within the exposed northern rift basins are Middle Triassic to Early Jurassic in age, whereas strata within the exposed southern rift basins (i.e., the basins south of the Culpeper basin) are exclusively Late Triassic in age (Olsen et al., 1989; Olsen, 1997) (Figure 5). Most rift basins are asymmetric, bounded on one side by a normal fault or a series of normal faults. Most boundary faults strike northeast-southwest. They dip either seaward or landward and have displacements that locally exceed 10 km (e.g., Hutchinson and Klitgord, 1988; Schlische, 1993; Withjack et al., 1995) (Figures 3, 4). The presence of conglomeratic facies near the boundary faults and the thickening of stratigraphic packages toward the faults indicate that deposition and movement on the boundary faults were coeval (e.g., Hutchinson and Klitgord, 1988; Schlische, 1992, 1993; Withjack et al., 1995) (Figure 4A–D).

Generally, the attitudes of the boundary faults reflect the crustal fabric produced during the Paleozoic orogenies (Lindholm, 1978; Ratcliffe and Burton, 1985; Swanson, 1986; Olsen and Schlische, 1990). For example, the normal faults bounding the Richmond rift basin of southern Virginia (Figures 3B, 4E) (Bobyarichick and Glover, 1979) and the Fundy rift basin of New Brunswick and Nova Scotia (Figures 3A, 4A) (Brown, 1986; Withjack et al., 1995) are reactivated thrust-fault zones. Structural analyses suggest that the maximum compressive stress was subvertical, and the directions of the maximum (\(S_{\text{max}}\)) and minimum (\(S_{\text{min}}\)) horizontal compressive stresses were approximately northeast-southwest and northwest-southeast, respectively, during rifting throughout eastern North America (e.g., Ratcliffe and Burton, 1985; Venkatakrishnan and Lutz, 1988; Olsen et al., 1989; Olsen and Schlische, 1990; de Boer, 1992; Schlische, 1993). Under this stress regime, northeast-striking Paleozoic structures were reactivated as normal faults, east-northeast–striking structures became oblique-slip faults with normal and sinistral strike-slip components of displacement, and north-northeast-striking structures became oblique-slip faults with normal and dextral strike-slip components of displacement.

### Jurassic Continental Breakup

Magnetic-anomaly patterns provide little information about the early sea-floor-spreading history of the North Atlantic Ocean because the oldest dated magnetic anomaly, chron M-25, formed at about 155 Ma (Klitgord and Schouten, 1986) (Figures 2, 5). Extrapolations based on Late Jurassic sea-floor–spreading rates suggest that the breakup of North America and Africa occurred during the Middle Jurassic at about 175 Ma (Klitgord and Schouten, 1986).

Seismic and well data, however, suggest that breakup occurred before 175 Ma. An unconformity, termed the postrift unconformity (PRU), separates deformed synrift rocks from relatively undeformed postrift strata on the continental margin of eastern North America (Grow et al., 1983) (Figures 3, 4C). Presumably, the separation of North America and Africa began after the deposition of the youngest synrift strata below the PRU and before the deposition of the oldest postrift strata above the PRU. In the Georges Bank and Scotian basins, the oldest postrift sedimentary rocks are late Early Jurassic to early Middle Jurassic in age (e.g., Klitgord et al., 1988; Welsink et al., 1989). Thus, in the northeastern United States and maritime Canada, drifting began after the deposition of the Lower Jurassic synrift strata and before postrift deposition in the late Early Jurassic to early Middle Jurassic (Figure 5). Farther to the south in the Baltimore Canyon and Carolina troughs, offshore wells have penetrated postrift sedimentary rocks of Late Jurassic age (e.g., Poag and Valentine, 1988). Seismic data show that a thick sequence of postrift strata underlies these Upper Jurassic rocks (e.g., Klitgord et al., 1988; Poag, 1991). The age of these postrift rocks is unknown. Although Poag (1991) proposed an Aalenian age for the oldest postrift rocks in the Baltimore Canyon and Carolina troughs, he acknowledged that this age is equivocal. Reflections from the oldest postrift strata cannot be directly correlated with dated reflections from the Georges Bank Basin. Thus, the timing of breakup is poorly constrained in the southeastern United States; breakup occurred sometime after the deposition of the Upper Triassic strata within the exposed rift basins and before the deposition of the Upper Jurassic postrift strata encountered in the offshore wells.
Figure 2—Major Paleozoic compressional structures and early Mesozoic rift basins of eastern North America and key tectonic features of the eastern North Atlantic Ocean (Benson and Doyle, 1988; Klitgord et al., 1988; Manspeizer and Cousminer, 1988; Costain and Çoruh, 1989; Olsen et al., 1989; Tankard and Welsink, 1989; MacLean and Wade, 1992; Sheridan et al., 1993; Rankin, 1994). Thick dashed lines and squares with notation show location of transects in Figure 3. Thin double lines and circles with notation show location of sections in Figure 4.
Eastern North America Magmatic Activity

The intrusion of diabase dikes and sills and the extrusion of basalt flows occurred throughout eastern North America during the Early Jurassic (e.g., King, 1971; Olsen et al., 1989, 1996; McHone, 1996). Basalt flows are intercalated with synrift strata in the Culpeper, Gettysburg, Newark, Hartford, Deerfield, and Fundy rift basins, but are not present in the exposed southern rift basins (Olsen et al., 1989). Instead, basalt flows in the southeastern United States overlie the rift basins (Dillon et al., 1983; Behrendt, 1985, 1986; Costain and Ėoruh, 1989; McBride et al., 1989) (Figures 4F; 6). These postrift basalts are relatively flat-lying and have great lateral extent.

Until recently, the timing of the eastern North America (ENA) magmatic activity was poorly constrained because many of the K-Ar and $^{40}$Ar/$^{39}$Ar ages reflected postcrystallization alteration involving argon loss or gain (e.g., Sutter, 1988). Recent studies using improved radiometric-dating techniques, however, have better constrained the timing. In the Culpeper, Gettysburg, Newark, and Fundy rift basins, most ENA magmatic activity occurred at 201 ±2 Ma (Sutter, 1988; Dunning and Hodych, 1990; Hodych and Dunning, 1992) (Figure 5). Geochemical correlations and cyclostratigraphy demonstrate that the duration of the magmatic activity, at least in the northeastern United States and southeastern Canada, was extremely short, about 600,000 yr (Olsen et al., 1989, 1996).

Lanphere (1983) radiometrically dated postrift basalts from South Carolina at 184 ±3.3 Ma. Recent studies, however, suggest that this age is too young, reflecting postcrystallization alteration. Ragland et al. (1992), after reviewing all available geologic, paleomagnetic, and geochemical information, concluded that the most probable age for the ENA magmatic activity in the southeastern United States (basalt flows, dikes, and sills) is 200 ±5 Ma, consistent with that of the ENA magmatic activity in the northeastern United States and maritime Canada.

A massive wedge, presumably composed of volcanic or volcanioclastic rocks, is present along the edge of the passive margin of the eastern United States (Hinz, 1981; Benson and Doyle, 1988; Klitgord et al., 1988; Austin et al., 1990; Holbrook and Kelemen, 1993; Sheridan et al., 1993; Kelemen and Holbrook, 1995; Oh et al., 1995) (Figures 3B, C; 6). The wedge lies near the continent–ocean boundary and formed during the transition from rifting to drifting (Hinz, 1981; Benson and Doyle, 1988; Austin et al., 1990). A similar wedge is not observed on the passive margin of southeastern Canada (Keen and Potter, 1995). Work by Kelemen and Holbrook (1995) suggested that the formation of this volcanic/volcanioclastic wedge requires active asthenospheric upwelling. The exact age of the wedge is unknown and, in fact, may vary along the margin (e.g., Oh et al., 1995). Beneath the Georges Bank Basin, the seaward-dipping reflectors...
Figure 4—Sections through rift basins of eastern North America. Vertical lines with wide spacing mark contact between synrift strata of early Mesozoic age (shaded) and prerift rocks of Precambrian–Paleozoic age (unshaded). Thick black lines are fault surfaces. Arrows show Mesozoic motions. Vertical axes of seismic lines are in two-way traveltime. Section locations are shown in Figure 2. (A) Line drawing of time-migrated seismic line 82-29 through the Chignecto and Minas subbasins of the Fundy rift basin of New Brunswick and Nova Scotia (after Withjack et al., 1995). (B) Line drawing of northern segment of time-migrated seismic line 81-47 through the Fundy rift basin of New Brunswick and Nova Scotia (after Withjack et al., 1995). Vertical lines with close spacing denote reflection from Lower Jurassic North Mountain Basalt. (C) Line drawing of segment of seismic line 3630-1/2-85 through the Emerald/Naskapi rift basin of offshore Nova Scotia (Tankard and Welsink, 1989). (D) Line drawing of seismic line NB-1 through the Newark rift basin (Costain and Çoruh, 1989). (E) Cross section through the Richmond rift basin of Virginia (after Shaler and Woodworth, 1899). (F) Line drawing of seismic line VT-5 through the Jedburg rift basin of South Carolina (Costain and Çoruh, 1989). (G) Line drawing of segment of seismic line SC10 from onshore South Carolina (Hamilton et al., 1983).
within the wedge underlie a relatively flat-lying Middle Jurassic sequence (Schlee and Klitgord, 1988). Thus, the wedge beneath the Georges Bank Basin formed before the Middle Jurassic deposition of these postrift strata. Beneath the Baltimore Canyon Trough, the wedge overlaps rift basins (Benson and Doyle, 1988) and underlies the postrift unconformity (Sheridan et al., 1993) (Figure 3B). Thus, the wedge beneath the Baltimore Canyon Trough formed after the deposition of the Upper Triassic to Lower Jurassic synrift strata and before the deposition of the thick package of postrift strata beneath the Upper Jurassic rocks encountered in the offshore wells. Beneath the Carolina Trough, the wedge appears to underlie the postrift basalts (Austin et al., 1990; Oh et al., 1995) (Figure 3C). If so, then the wedge beneath the Carolina Trough formed before the eruption of the postrift basalts (~200 Ma).

CHANGE IN DEFORMATIONAL REGIME AND STRESS STATE IN THE SOUTHEASTERN UNITED STATES BEFORE ENA MAGMATISM

The following geologic evidence suggests that the deformational regime and stress state changed significantly in the southeastern United States before the ENA magmatic activity in the earliest Jurassic.

(1) No sedimentary rocks of Early Jurassic age are present within any onshore rift basin south of the Culpeper basin (Olsen et al., 1989; Olsen, 1997) (Figure 5). Synrift sedimentary rocks in the southern rift basins are exclusively of Late Triassic age. Either deposition ceased in the southern rift basins by the Early Jurassic, or later erosion removed the Lower Jurassic synrift strata. Burial history reconstructions and thermal models based on analyses of fluid inclusions, vitrinite reflectance, and apatite-fission tracts indicate that, at least within the Taylorsville basin, synrift deposition ceased prior to the Jurassic (Malinconico, 1996; Tseng et al., 1996a, b).

(2) An angular unconformity separates the synrift and postrift strata in the subsurface of South Carolina (Behrendt, 1986; Costain and Çoruh, 1989). The postrift sequence includes a series of flat-lying basalt flows with great lateral extent (Figures 4E, 6). Locally, a thin package of sedimentary rocks separates the basalts from the underlying postrift unconformity (Dillon et al., 1983; Behrendt, 1986; Costain and Çoruh, 1989). Apparently, rifting ceased and erosion occurred in the southeastern United States before the eruption of the postrift basalts (~200 Ma) and the deposition of the postrift sedimentary rocks beneath them.

(3) The Cooke fault is a northeast-striking, basement-involved reverse fault in South Carolina (Figure 4G). Seismic data show that it had about 140 m of reverse displacement before the eruption of the postrift basalts (Behrendt et al., 1981; Hamilton et al., 1983).

(4) Numerous northeast-striking, basement-involved reverse faults and associated fault-propagation folds are present in the Richmond rift basin of southern Virginia (Shaler and Woodworth, 1899; Venkatakrishnan and Lutz, 1988) (Figure 4D). The constant thickness of the synrift strata across these
structures indicates that they developed after the Late Triassic deposition of the synrift strata. Venkatakrishnan and Lutz (1988) reported that north-northwest-striking dikes clearly cut these structures; thus, the reverse faults and associated fault-propagation folds in the Richmond rift basin formed before dike emplacement (~200 Ma).

(5) Most dikes strike northwest-southeast in Georgia, South Carolina, and North Carolina, and north-northwest–south-southeast in southern Virginia, roughly orthogonal to the trend of the rift basins (Figures 6, 7B). Many of these dikes cut across the boundary faults of the southern rift basins without exhibiting any offset. Unless the
fault-dike intersections, activity on the boundary faults of the southern rift basins ceased before dike emplacement (~200 Ma). If dike trends reflect the stress state during their injection (e.g., Anderson, 1951; Zoback et al., 1993), then $S_{\text{hmax}}$ and $S_{\text{hmin}}$ were approximately northwest-southeast and northeast-southwest, respectively, during dike emplacement (~200 Ma).

These geologic data consistently indicate that the tectonic regime changed significantly in the southeastern United States after the deposition of the Upper Triassic synrift strata and before the ENA magmatic activity during the earliest Jurassic. Under this new tectonic regime, rifting and its concomitant normal faulting ceased, and northeast-striking reverse faults formed. The minimum compressive stress became subvertical, and $S_{\text{hmax}}$ and $S_{\text{hmin}}$ were approximately northwest-southeast and northeast-southwest, respectively.

CONTINUED RIFTING IN THE NORTHEASTERN UNITED STATES AND SOUTHEASTERN CANADA DURING AND AFTER ENA MAGMATISM

The northern rift basins, unlike their southern counterparts, remained active during and after the ENA magmatic activity. Generally, the Lower Jurassic stratal packages within the northern rift basins, including the interbedded ENA basalts, thicken toward the boundary faults. For example, the Lower Jurassic North Mountain Basalt (202 ±2 Ma) (Hodych and Dunning, 1992) and the overlying Lower Jurassic McCoy Brook Formation thicken toward the northern and northwestern boundary faults of the Fundy rift basin (Olsen and Schlische, 1990; Withjack et al., 1995). Fault-controlled thickness variations such as these show that rifting continued in the northern rift basins during and after the ENA magmatic activity. In fact, accumulation rates for strata of Early Jurassic age are markedly higher than those for strata of Late Triassic age, suggesting that normal faulting and basin subsidence accelerated during this period (Schlische and Olsen, 1990; Schlische and Anders, 1996). Conglomerates are abundant in the Lower Jurassic synrift strata near the boundary faults of the northern rift basins (e.g., Olsen et al., 1989; Schlische, 1992; Withjack et al., 1995). The presence of these coarse-grained rocks near the boundary faults provides additional evidence that rifting continued in the northern rift basins during and after the ENA magmatic activity. Most dikes in the northeastern United States and southeastern Canada strike northeast-southwest, subparallel to the trend of the rift basins (e.g., McHone, 1988) (Figures 6, 7A). If these dike trends reflect the stress state during their injection, then $S_{\text{hmax}}$ and $S_{\text{hmin}}$ were approximately northeast-southwest and northwest-southeast, respectively, during dike emplacement. These orientations of $S_{\text{hmax}}$ and $S_{\text{hmin}}$ are indistinguishable from those during Late Triassic rifting.

CHANGE IN DEFORMATIONAL REGIME AND STRESS STATE IN NORTHEASTERN UNITED STATES AND SOUTHEASTERN CANADA

The oldest postrift sedimentary units in the Georges Bank and Scotian basins are late Early Jurassic to early Middle Jurassic in age (Kligord et al., 1988; Tankard and Welsink, 1989) Thus, rifting ceased and drifting began in the northeastern United States and southeastern Canada before the deposition of the postrift sedimentary rocks in the late Early Jurassic to early Middle Jurassic.

After rifting, the deformational regime and stress state changed in the northeastern United States and southeastern Canada. Postrift deformation within the Newark, Hartford, and Deerfield rift basins includes reverse faults, strike-slip faults, and folds (e.g., de Boer and Clifton, 1988; Lucas et al., 1988; Wise, 1992). Most structural analyses of the postrift deformation in these basins indicate that the intermediate compressive stress was subvertical, and $S_{\text{hmax}}$ and $S_{\text{hmin}}$ were approximately north-south and east-west, respectively (Lomando and Engelder, 1984; de Boer and Clifton, 1988; Lucas et al., 1988; Wise, 1992). Unfortunately, the age of this postrift deformation is poorly constrained—it may have occurred during the Jurassic, the Cretaceous, or even the Tertiary.

Recent studies have documented the presence of compressional structures in the Fundy rift basin (Withjack et al., 1995). The Fundy basin has the key features of an inverted rift basin (e.g., Glennie and Boegner, 1981; Bally, 1984; Cooper et al., 1989; Eisenstadt and Withjack, 1995). The basin’s boundary faults had several kilometers of reverse displacement after rifting, and small- to large-scale folds, reverse faults, and normal faults reactivated as reverse faults affected the synrift strata within the basin (Withjack et al., 1995). Seismic data indicate that the northeast-striking boundary fault of the Fundy basin experienced more than 4 km of reverse displacement after synrift deposition. In response, the Fundy basin rose more than 1.5 km relative to its northwestern margin; a wide, northeast-striking anticline developed along the northwestern margin of the basin, and the basin acquired a broad synclinal geometry (Figure 4B). Structural analyses (Withjack et al., 1995) suggest that during inversion, the minimum compressive stress was subvertical, and $S_{\text{hmax}}$ and $S_{\text{hmin}}$ were
Figure 7—Geologic maps of (A) the Connecticut Valley basin (Massachusetts and Connecticut) and (B) the Deep River basin (North Carolina and South Carolina). Near the Connecticut Valley basin, earliest Jurassic diabase dikes are northeast striking and are subparallel to the regional trend of the basin. Near the Deep River basin, earliest Jurassic diabase dikes are north and northwest striking and are subperpendicular to the regional trend of the basin. Early Jurassic lava flows and sedimentary rocks are present in the Connecticut Valley basin, but are absent in the Deep River basin. Maps are simplified from Schlische (1993).
approximately northwest-southeast and northeast-southwest, respectively.

None of the compressional structures within the Fundy rift basin show evidence of growth. Thus, inversion occurred after the deposition of the Lower Jurassic strata within the basin. Geological relationships from the Orpheus graben, the eastern offshore continuation of the Fundy rift basin, provide additional constraints on the timing of inversion (Withjack et al., 1995). The Cobequid-Chedabucto fault system bounds the Fundy basin and the Orpheus graben on the north (Figure 2). If the faults at the western end of this fault system (i.e., the faults bounding the Fundy basin) had reverse displacements during inversion, then the faults at the eastern end (i.e., the faults bounding the Orpheus graben) probably had similar movements. Except for regional subsidence and minor salt movement, structural activity in the Orpheus graben ceased during the Early Cretaceous (Tankard and Welsink, 1989; Wade and MacLean, 1990; MacLean and Wade, 1992); hence, any inversion in the Orpheus graben and, by inference, in the Fundy basin occurred before or during the Early Cretaceous.

**DISCUSSION**

Our work indicates that the tectonic regime in eastern North America changed substantially after rifting. During rifting, the maximum compressive stress was subvertical, and $s_{\text{max}}$ and $s_{\text{min}}$ were approximately northeast-southwest and northwest-southwest, respectively. After rifting in the southeastern United States and maritime Canada, the minimum compressive stress became subvertical, and $s_{\text{max}}$ and $s_{\text{min}}$ were approximately northwest-southeast and northeast-southwest, respectively. In response to this stress reorientation, rifting and its concomitant northeast-striking normal faulting ceased, northeast-striking reverse faults formed, and the rift-basin boundary faults had reverse displacements. In the southeastern United States, reverse faulting began after the deposition of the Upper Triassic synrift strata and before the ENA magmatic activity (Figure 8B). In southeastern Canada, reverse faulting/inversion began after the ENA magmatic activity, specifically after the deposition of the Lower Jurassic synrift strata and before or during the Early Cretaceous (Figure 8A).

The formation of the volcanic/volcaniclastic wedge beneath the Carolina Trough and, by inference, the rift-drift transition in the southeastern United States occurred after the deposition of the Upper Triassic synrift strata and before the eruption of the postrift basalts (Austin et al., 1990; Oh et al., 1995) (Figure 8B). Thus, the rift-drift transition and the onset of reverse faulting were approximately coeval in the southeastern United States. In the northeastern United States and maritime Canada, the rift-drift transition occurred after the deposition of the Lower Jurassic synrift strata and before postrift deposition in the late Early Jurassic to early Middle Jurassic (Klitgord et al., 1988; Welsink et al., 1989) (Figure 8A). Thus, the rift-drift transition in the northeastern United States and southeastern Canada occurred immediately before or during the onset of reverse faulting and inversion in maritime Canada.

Based on these observations, we conclude that the onset of reverse faulting/inversion on the passive margin of central eastern North America occurred during the rift-drift transition or early
stages of sea-floor spreading. We also conclude that the rift-drift transition was diachronous; rifting ended and drifting began earlier in the southeastern United States and later in the northeastern United States and southeastern Canada. Subsidence analyses by Dunbar and Sawyer (1989) support this latter conclusion. Their work suggests that sea-floor spreading in the North Atlantic Ocean occurred first in the south between the Blake Spur and Delaware Bay fracture zones (Figure 2). Sea-floor spreading occurred later in the north between the Delaware Bay and Pico fracture zones. As the North Atlantic Ocean continued to develop, sea-floor-spreading centers propagated northward (e.g., Srivastava and Tapscott, 1986; Srivastava and Verhoef, 1992; Chalmers and Laursen, 1995). Drifting began between Newfoundland and Iberia in the Early Cretaceous, between Labrador and western Greenland in the Late Cretaceous or Paleocene, and between eastern Greenland and northwest Europe in the early Eocene. The diachronous rift-drift transition along the passive margin of central eastern North America appears to have been the first of many stages in the episodic development of the North Atlantic Ocean.

No collision or subduction zones existed near eastern North America during the early Mesozoic. Thus, the cause of the postrift shortening/inversion on the passive margin of central eastern North America is enigmatic. We propose that the change in the tectonic regime during the rift-drift transition or early stages of sea-floor spreading resulted, at least in part, from incipient ridge-push forces and an initial continental resistance to plate motion (Figure 9). Previous workers have suggested that these mechanisms produced shortening and inversion on other passive margins during the early stages of sea-floor spreading (Dewey, 1988; Bott, 1992; Boldreel and Andersen, 1993). During the early stages of drifting, distant plate-tectonic forces (e.g., slab pull) produced divergent lithospheric displacements and widespread extension in eastern North America (Figure 9A). The active asthenospheric upwelling identified by Kelemen and Holbrook (1995) initially had little influence on lithospheric displacements. By the final stages of rifting, the lithosphere had thinned substantially (Figure 9B). Magma associated with the active asthenospheric upwelling intruded the attenuated lithosphere and erupted at the surface, forming a massive volcanic/volcaniclastic wedge (Kelemen and Holbrook, 1995). Gravitational-body forces and traction forces associated with the hot, low-density asthenospheric upwelling (Bott, 1992) increased significantly. In response, lithospheric displacements near the upwelling exceeded those far from the upwelling, causing shortening (inversion) in the intervening region. During the early stages of drifting, lithospheric displacements far from the upwelling increased, eventually equaling displacements near the upwelling (Figure 9C). Most shortening/inversion ceased. Because the volume of lithosphere displaced laterally equaled the volume of upwelling asthenosphere, the asthenospheric upwelling became passive (Kelemen and Holbrook, 1995). Magmatism and sea-floor spreading dissipated the thermal anomaly in the upper mantle, after which the formation of the volcanic/volcaniclastic wedge ceased, and normal oceanic crust developed along the Mid-Atlantic Ridge.

REVISED MODEL FOR THE TECTONIC EVOLUTION OF THE PASSIVE MARGIN OF EASTERN NORTH AMERICA

Based on these observations and interpretations, we propose a revised model for the tectonic development of the passive margin of central eastern North America (Figure 10).

Middle to Late Triassic

Following Paleozoic deformation, numerous rift basins developed throughout eastern North America during the Middle to Late Triassic (Figure 10A). Many of the normal faults bounding the rift basins were reactivated Paleozoic structures. Similar rift basins developed on the conjugate continental margin of Africa (e.g., Van Houten, 1977; Lee and Burgess, 1978; Laville and Petit, 1984; Beauchamp, 1988; Medina, 1988; Laville and Piqué, 1992).

Late Triassic to Early Jurassic (Shortly Before to After ENA Magmatic Activity)

In the southeastern United States, rifting ceased before the ENA magmatic activity during the early Early Jurassic (~200 Ma). During the subsequent rift-drift transition, the southern rift basins were eroded, northeast-striking reverse faults formed, and postrift deposition occurred locally (Figures 8B, 10B). A massive volcanic/volcaniclastic wedge formed near the continent–ocean boundary. ENA magmatism led to the emplacement of diabase sills and northwest-striking dikes, and the eruption of postrift basalt flows. In the northeastern United States and southeastern Canada, rifting continued before, during, and after the ENA magmatic activity. ENA magmatism led to the emplacement of diabase
sills and northeast-striking dikes, and the eruption of synrift basalt flows (Figures 8A, 10B).

**Early Jurassic (After ENA Magmatic Activity) to Early Cretaceous**

In the southeastern United States, full-fledged drifting began shortly after the ENA magmatism. The North American and African continents separated, and sea-floor-spreading centers developed in the embryonic North Atlantic Ocean. The continental margin of the southeastern United States subsided, and postrift strata progressively onlapped the erosional surface (i.e., either the postrift unconformity or the top of the postrift basalt).

In the northeastern United States and southeastern Canada, drifting ceased by the early Middle Jurassic (~185 Ma). The northern rift basins were eroded, and a massive volcanic/volcaniclastic wedge formed near the continent–ocean boundary of the northeastern United States during the rift-drift transition. Inversion occurred in southeastern Canada before or during the Early Cretaceous, during the rift-drift transition or during the early stages of sea-floor spreading. Inversion structures, similar to those in the Fundy rift basin, developed on the conjugate continental
Figure 10—Sketches showing evolution of central eastern North America in map and cross-sectional views. (A) Middle to Late Triassic. Northeast-striking rift basins developed, subsiding and filling with sediments. (B) Late Triassic to Early Jurassic, shortly before to after eastern North America (ENA) magmatic activity. In the southeastern United States, rifting ceased, and northeast-striking reverse faults and associated folds developed. ENA magmatic activity led to the emplacement of northwest-striking dikes. The northern rift basins continued to develop, their boundary faults remained active, and the ENA magmatic activity favored the emplacement of northeast-striking diabase dikes. (C) Early Jurassic (after ENA magmatic activity) to Early Cretaceous. Full-fledged drifting began between the southeastern United States and western Africa. In the northeastern United States and southeastern Canada, rifting ceased before the late Early Jurassic to early Middle Jurassic. Inversion occurred in southeastern Canada before or during the Early Cretaceous. See text for details.
Our work suggests that the passive margin of eastern North America experienced shortening and inversion during the rift-drift transition or the early stages of sea-floor spreading. Other passive margins have had similar tectonic activity during the early stages of their development. For example, Laville and Piqué (1992) showed that inversion structures developed in several Moroccan rift basins during the Middle Jurassic as North America separated from northern Africa. Boldreel and Andersen (1993) reported that compressional structures formed in the Faeroe-Rockall area during the late Paleocene-early Eocene, during the initiation of sea-floor spreading between Greenland and the Faeroe-Rockall Plateau. Other passive margins of exploration and production interest (i.e., those of eastern Canada, western Europe, eastern South America, Africa, India, and Australia) may have had shortening and inversion during the early stages of their development.

Shortening and inversion can significantly affect the hydrocarbon potential of rift basins and passive margins. Some attributes of inversion enhance hydrocarbon potential. For example, inverted rift basins have both extensional structures (e.g., tilted fault blocks) and compressional structures (e.g., broad hanging-wall anticlines) to trap hydrocarbons. Many attributes of inversion, however, reduce hydrocarbon potential. Generally, inverted rift basins experience significant uplift and erosion. This erosion can remove reservoir and source rocks, and suspend hydrocarbon maturation and generation. Topographic relief associated with uplift and erosion can increase rates of groundwater flow into the subsurface (Tseng et al., 1996a, b), leading to diagenetic reduction of reservoir porosity and permeability and degradation of hydrocarbon accumulations. Secondary deformation associated with shortening/inversion can destroy seal integrity, causing leakage from hydrocarbon traps on surface-penetrating faults. Commonly, the magnitude of shortening and uplift is poorly constrained (Eisenstadt and Withjack, 1995). Consequently, in many inverted rift basins, it is difficult to accurately define subsidence histories and conduct basin-modeling studies. Analyses of more than 100 rift basins by Macgregor (1995) showed that rift basins with little inversion have dispersed hydrocarbon distributions and high rates of exploration success. Rift basins with significant inversion, however, have concentrated hydrocarbon distributions (commonly in one field) and much lower success rates. Macgregor's (1995) analyses also indicated that inverted rift basins have low rates of exploration success unless the compressional structures formed before or during hydrocarbon generation and migration.

If the rift-drift transition is diachronous, as on the passive margin of central eastern North America, then adjacent segments of the passive margin can have very different tectonic and subsidence histories and, consequently, petroleum potentials. For example, the strata within the Taylorsville rift basin of the southeastern United States experienced their greatest burial depths and temperatures about 200 Ma (Figure 11). During the subsequent inversion, these strata were uplifted more than 2 km and eroded until the Early Cretaceous (Tseng et al., 1996a). Strata within the Fundy rift basin of maritime Canada, however, experienced their greatest burial depths and temperatures later, about 185 Ma (Figure 11). During the subsequent inversion, they were uplifted more than 1.5 km and eroded (Withjack et al., 1995). The maturation and generation of any potential source rocks in the Taylorsville basin would have ceased at 200 Ma, whereas the maturation and generation of any potential source rocks in the Fundy basin would have continued until 185 Ma.

**CONCLUSIONS**

Recent studies have constrained the age of igneous activity and identified new structural styles on the passive margin of central eastern North America, probably during the Middle Jurassic (Laville, 1988; Laville and Piqué, 1992).
America. The integration of this new information with existing geological and geophysical data suggests that the tectonic development of the passive margin of central eastern North America was considerably more complex than the classic two-stage, rift-drift model. The passive margin of central eastern North America had three stages of development: rifting, shortening/inversion during the rift-drift transition or during the early phases of drifting, and relative tectonic quiescence during the later phases of drifting.

The transition from rifting to drifting was diachronous on the passive margin of central eastern North America. In the southeastern United States, the rift-drift transition occurred after synrift deposition in the Late Triassic and before eastern North American (ENA) magmatic activity in the early Early Jurassic (~200 Ma). In the northeastern United States and maritime Canada, the rift-drift transition occurred after ENA magmatic activity and synrift deposition in the Early Jurassic and before postrift deposition in the early Middle Jurassic (~185 Ma). On both the southern and northern segments of the passive margin of eastern North America, the tectonic regime changed substantially after rifting. During rifting, the maximum compressive stress was subvertical, and $S_{\text{max}}$ and $S_{\text{min}}$ trended approximately northeast-southwest and northwest-southeast, respectively. After rifting in the southeastern United States and maritime Canada, the minimum compressive stress was subvertical, and $S_{\text{max}}$ and $S_{\text{min}}$ trended approximately northwest-southeast and northeast-southwest, respectively. In response to this stress reorientation, rifting and its concomitant northeast-striking normal faulting ceased, northeast-striking reverse faults formed, and the rift-basin boundary faults had reverse displacements. In the southeastern United States, the change in the tectonic regime occurred in the Late Triassic–Early Jurassic during the rift-drift transition. A massive volcanic or volcanoclastic wedge developed simultaneously near the continent–ocean boundary. In maritime Canada, the change in the tectonic regime occurred during or after the Early Jurassic and before or during the Early Cretaceous (i.e., during the rift-drift transition or the early stages of sea-floor spreading).

The recognition of shortening/inversion during early passive-margin development is critical to assess the hydrocarbon potential of passive margins. Uplift and erosion associated with shortening/inversion can remove reservoir and source rocks, suspend hydrocarbon maturation and generation, and increase rates of groundwater flow, causing diagenetic alteration of reservoir rocks and degradation of hydrocarbon accumulations. Secondary deformation associated with shortening/inversion can destroy seal integrity, allowing leakage of hydrocarbons to the surface. If the rift-drift transition is diachronous, as on the passive margin of central eastern North America, adjacent segments of the passive margin can have very different tectonic and subsidence histories and petroleum potentials.

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