TECTONICS, VOL. 14, NO. 2, PAGES 390-405, APRIL 1995

Tectonic evolution of the Fundy rift basin, Canada: Evidence of extension and shortening during passive margin development

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Abstract. Seismic and field data show that the Fundy rift basin of southeastern Canada experienced two distinct episodes of deformation during Mesozoic time. The first episode, during Middle Triassic to Early Jurassic time, was extensional. Rifting associated with NW-SE extension reactivated NE trending Paleozoic compressional structures as normal faults, forming the northwestern boundary faults of the Fundy basin. Displacements on the low-angle boundary faults locally exceeded 10 km. Rifting also reactivated east trending Paleozoic compressional structures as oblique-slip faults with normal and sinistral strike-slip components, forming the northern boundary faults of the Fundy basin. Several kilometers of sediments and lava flows filled the basin during rifting. The second deformational episode occurred during or after Early Jurassic time and probably before or during Early Cretaceous time. Inversion associated with NW-SE shortening occurred along all faulted margins of the Fundy basin. The northwestern boundary faults experienced several kilometers of reverse displacement, broad anticlines developed within their hanging walls, and the Fundy basin acquired its synclinal form. The northern boundary faults of the Fundy basin became oblique-slip faults with reverse and dextral strike-slip components. Gentle synclines, tight anticlines, and faults with reverse separation deformed the synrift strata near the northern margin of the Fundy basin. Neither collision nor subduction zones existed near the Fundy basin during Mesozoic time. Hence we believe that tectonic processes associated with seafloor spreading (e.g., incipient ridge push forces, continental resistance to plate motion) produced the shortening in the Fundy basin. Shortening occurred during the transition from rifting to drifting as North America separated from northern Africa and/or during the early stages of drifting as the seafloor-spreading centers of the North Atlantic propagated northward.

Paper number 94TC03087.
0278-7407/95/94TC-03087$10.00

Introduction

Most tectonic models of passive margin development, including those for the continental margin of eastern North America, have two distinct stages: rifting and drifting [e.g., *Dewey and Bird,* 1970; *Falvey,* 1974; *McKenzie,* 1978; *Bally,* 1979, 1981; *Lister et al.,* 1986; *Bond and Kominz,* 1988; *Keen and Beaumont,* 1990]. Rifting, accommodated by pure and/or simple shear, extends and attenuates the continental lithosphere, eventually triggering continental breakup and the onset of seafloor spreading. Drifting occurs when oceanic crust forms and the conjugate continental margins separate and thermally subside. For the continental margin of the eastern United States and southeastern Canada, the rifting stage began during Middle to Late Triassic time and, at least for the northern part of this continental margin, continued into Early Jurassic time [*Manspeizer and Cousminer,* 1988; *Olsen et al.,* 1989]. The drifting stage associated with the separation of North America and northern Africa and the creation of seafloor spreading centers in the North Atlantic Ocean began in late Early Jurassic to early Middle Jurassic time and continues today [*Klitgord and Schouten,* 1986; *Klitgord et al.,* 1988; *Tankard and Welsink,* 1989; *MacLean and Wade,* 1992].

In this paper, we provide additional information about the development of the continental margin of eastern North America by documenting the tectonic evolution of the Fundy basin, the westernmost rift basin on the passive margin of southeastern Canada. The seismic coverage within the Bay of Fundy is extensive, and the outcrops along its shores are spectacularly exposed. We have used these seismic and field data together to define the structural geometries within the Fundy basin and to determine how these structural geometries have evolved through time. Also, we have related the tectonic evolution of the Fundy basin to the development of the passive margin of southeastern Canada. Our work indicates that not all passive margins have a simple two-stage evolution. The western edge of the passive margin of southeastern Canada experienced three stages of development: rifting, shortening during the rift-drift transition and/or during the early phases of drifting, and relative tectonic quiescence during the later phases of drifting.

Fundy Basin

Numerous rift basins of early Mesozoic age exist on the continental margin of eastern North America (Figures 1a and 2). The Fundy basin of New Brunswick and Nova Scotia, Canada, contains several kilometers of nonmarine sedimentary rocks and tholeiitic basalt flows of Middle Triassic to Early Jurassic age [*Powers,* 1916; *Klein,* 1962; *Keppie,* 1979; *Olsen et al.*, 1989; *Olsen and Schlische,* 1990; *Withjack et al.*, 1991, 1992] (Figure 1b). These synrift strata overlie mildly to intensely deformed Paleozoic and Precambrian rocks. The oldest exposed synrift units, the Wolfville and Blomidon Formations, are composed of clastic sedimentary rocks of Middle to Late Triassic age. The Early Jurassic age North Mountain Basalt (202 ± 2 Ma [*Hodych and Dunning,* 1992]) overlies these formations. The youngest synrift unit, the McCoy Brook Formation, is composed primarily of clastic sedimentary rocks of Early Jurassic age.

The Fundy basin has three structural components: the Chignecto, Fundy, and Minas subbasins [*Olsen and Schlische,* 1990] (Figure 1c). Field observations indicate that SE dipping normal faults bound the NE trending Chignecto and Fundy subbasins on the northwest, whereas a series of normal, strike-slip, and oblique-slip faults bound the east trending Minas subbasin on the north [*Powers,* 1916; *Keppie,* 1982; *Plint and van de Poll,* 1984; *Nadon and Middleton,* 1985; *Olsen et al.*, 1989; *Olsen and Schlische,* 1990]. Many of the faults that bound the Chignecto, Fundy, and Minas subbasins are reactivated Paleozoic compressional structures [*Keppie,* 1982; *Plint and van de Poll,* 1984; *Brown,* 1986; *Olsen and Schlische,* 1990]. The extensional reactivation of Paleozoic compressional structures with differing orientations produced the contrasting structural styles along the faulted margins of the three subbasins [*Olsen and Schlische,* 1990]. NW-SE extension during early Mesozoic time reactivated preexisting

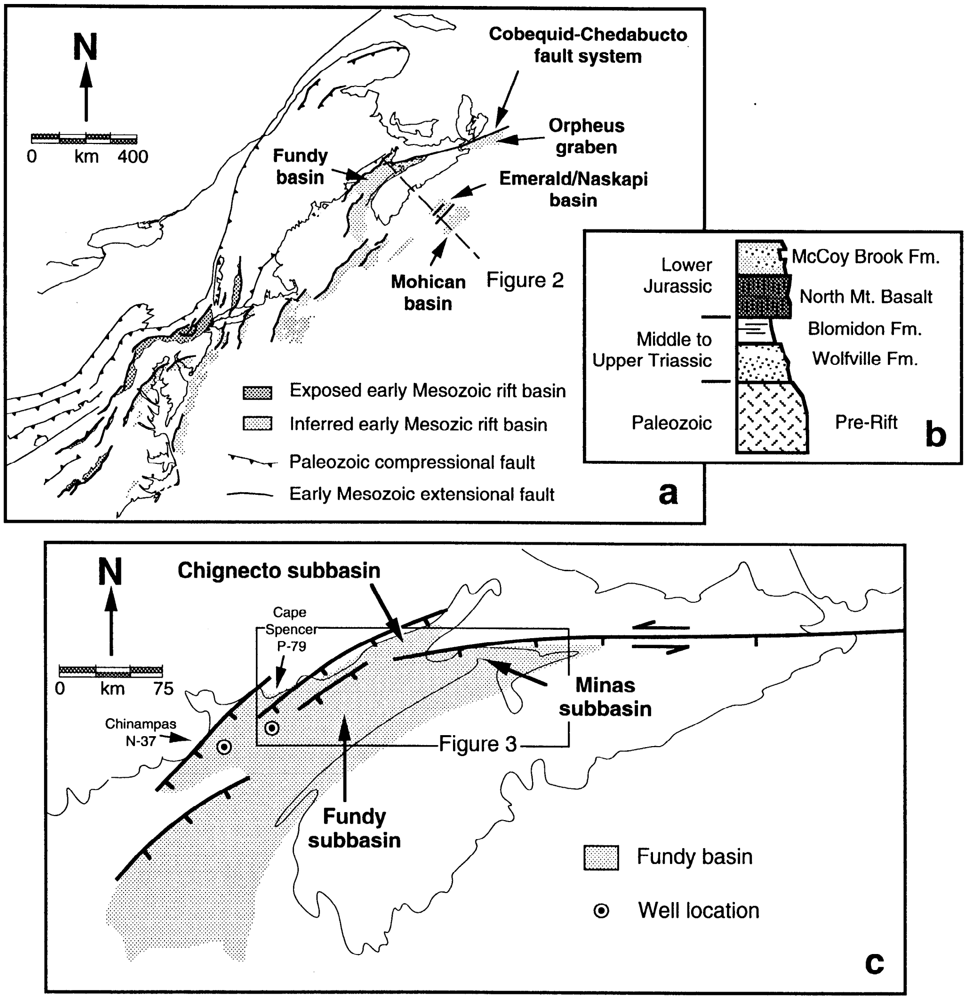

Figure 1. (a) Major Paleozoic compressional structures and early Mesozoic rift basins of eastern North America [after *Olsen et al.*, 1989]. Dashed line shows location of section shown in Figure 2. (b) Stratigraphic column for Fundy basin showing synrift formations. (c) Three structural components of Fundy basin: Chignecto, Fundy, and Minas subbasins. During early Mesozoic time, SE dipping normal faults bounded the Chignecto and Fundy subbasins on the northwest, and south dipping faults with normal and/or sinistral strike-slip components bounded the Fundy and Minas subbasins on the north [after *Olsen and Schlische,* 1990].

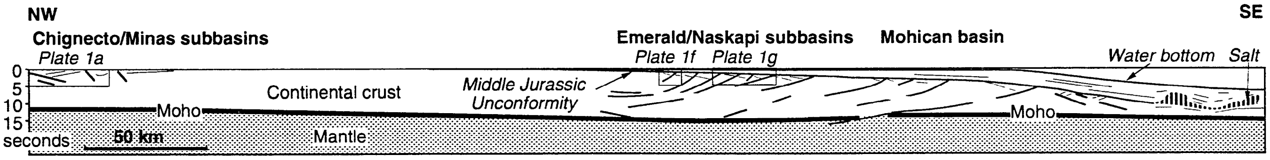

Figure 2. NW-SE section across the continental margin of southeastern Canada based on seismic line 82-29 (this paper) and deep seismic reflection profiles 88-1 and 88-1A [*Keen et al.*, 1991b]. Section location is shown in Figure 1a.

NE trending structures along the northwest margins of the Chignecto and Fundy subbasins as normal faults. Preexisting east trending structures along the northern margin of the Minas subbasin (i.e., the Cobequid-Chedabucto fault system or Minas geofracture [*Keppie*, 1982]) became oblique-slip fault zones with normal and sinistral strike-slip components of displacement.

Seismic Data, Character, and Interpretation

More than 1000 km of seismic reflection data cover the junction of the Chignecto, Fundy, and Minas subbasins (Figure 3). These industry data, acquired from 1980 to 1982, were processed using standard methods including signature deconvolution, predictive deconvolution, velocity analysis, stacking, and finite difference migration. The 1980 and 1981 seismic data are 36-fold and have a 5-s record length. The 1982 data are 48-fold and have a 6-s record length. Interval velocities, derived from stacking velocities, range from about 2 km/s directly below the water bottom to about 5 km/s near the bottom of the seismic sections. Peg leg multiples originating from the water bottom and the North Mountain Basalt commonly obscure reflection geometries.

Two major packages of reflections exist on the seismic lines (Plates 1a to 1e). Reflections in the upper package generally are closely spaced, continuous, and subparallel, and dip toward the north or northwest. Reflections in the lower package generally are widely spaced and discontinuous, and dip in several directions. On the northern ends of many seismic lines, reflections in the lower package are subparallel and dip toward the south or southeast. Projections of outcrop data from western Nova Scotia [*Powers*, 1916; *Klein*, 1962; *Olsen et al.*, 1989] and southeastern New Brunswick [*Powers*, 1916; *Rast and Grant*, 1973; *Ruitenberg and McCutcheon*, 1982; *Plint and van de Poll*, 1984; *Nadon and Middleton*, 1985] and ties with the Cape Spencer P-79 and Chinampas N-37 wells within the Bay of Fundy [*Pe-Piper et al.*, 1992] (Figure 1c) indicate that (1) synrift strata of Middle Triassic to Early Jurassic age probably produce the upper package of reflections, and (2) prerift strata and structures of Precambrian to Paleozoic age probably create the lower package of reflections. The North Mountain Basalt generates a

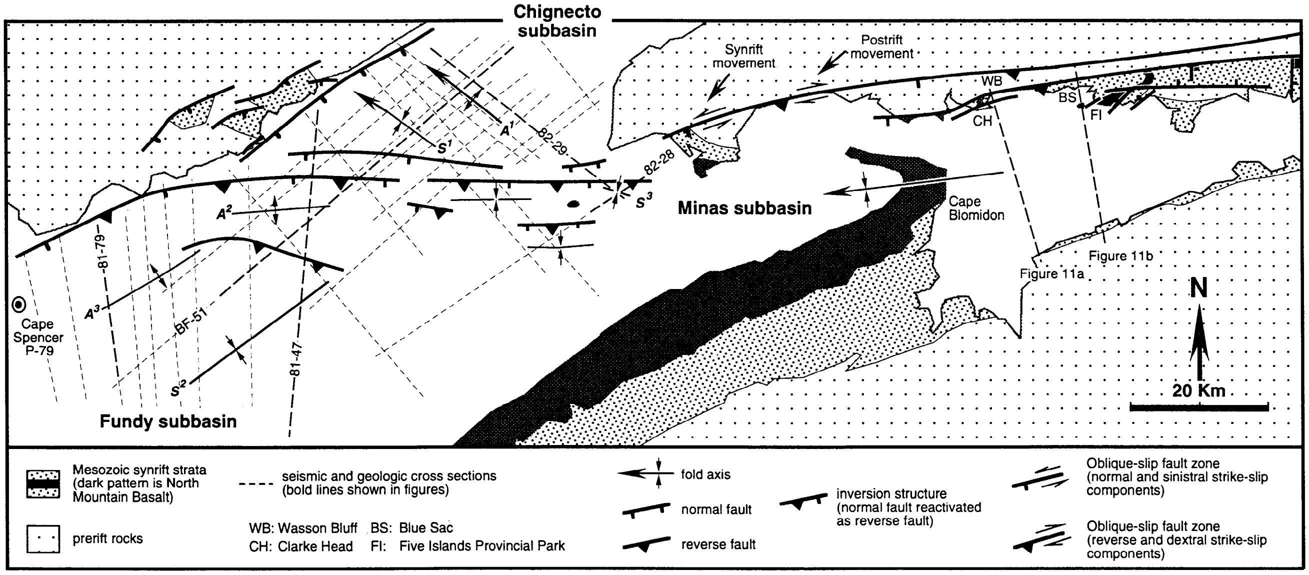

Figure 3. Map of junction of Chignecto, Fundy, and Minas subbasins showing distribution of strata and structures of early Mesozoic age, seismic coverage, and location of geologic cross sections across the Minas subbasin shown in Figure 11. Onshore geology is from *Keppie* [1979], *Donohoe and Wallace* [1982], *Plint and van de Poll* [1984], *Nadon and Middleton* [1985], *Olsen et al.* [1989], and *Olsen and Schlische* [1990]. Offshore geology is based on seismic interpretation. Map location is shown in Figure 1c.

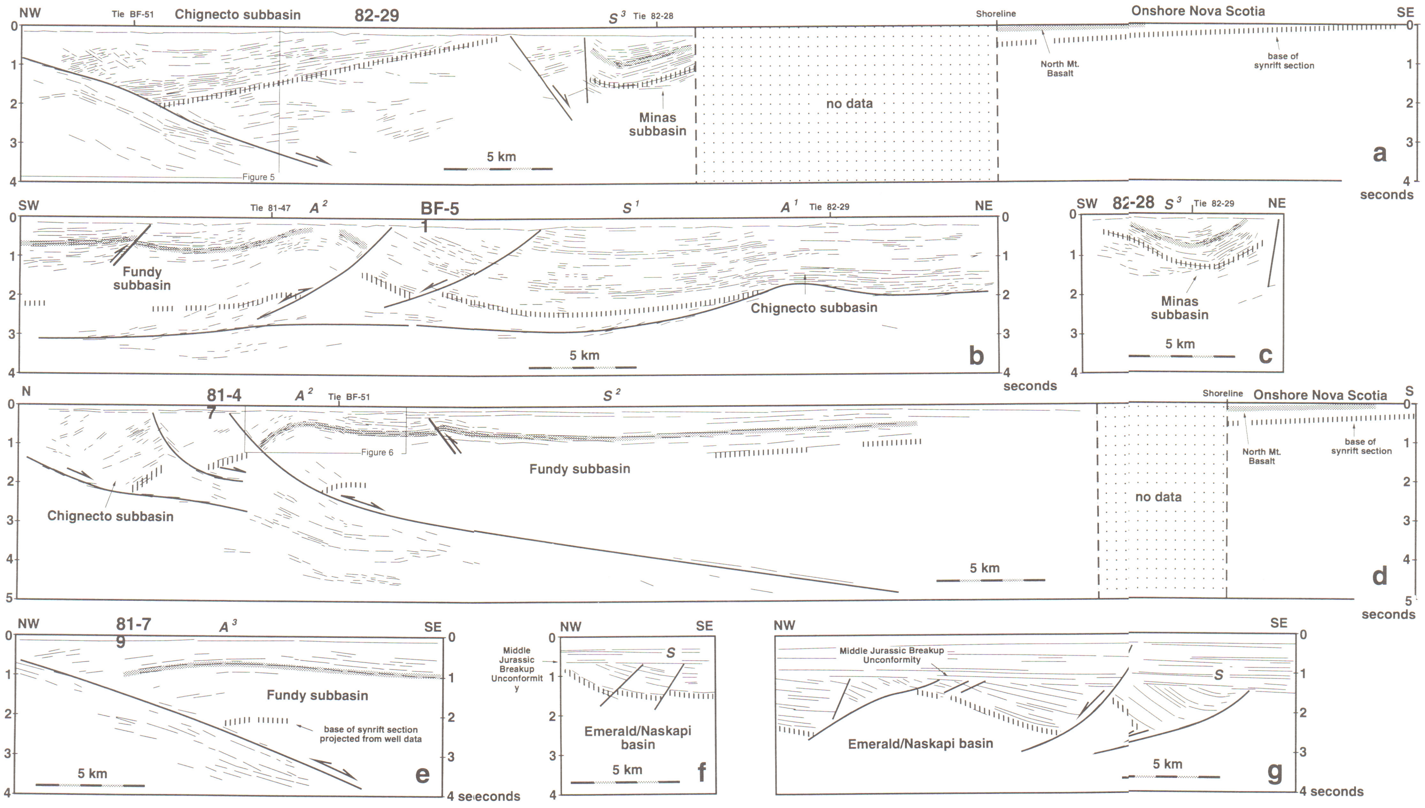

Plate 1. Line drawings of time-migrated seismic lines from the junction of the Chignecto, Fundy, and Minas subbasins and from the Emerald/Naskapi basin offshore Nova Scotia. Locations of lines from the Chignecto, Fundy, and Minas subbasins are shown in Figure 3. Onshore geology from western Nova Scotia [*Olsen et al.*, 1989] is converted to two-way travel time assuming a velocity of 3.5 km/s. (a) Line 82-29 through the Chignecto and Minas subbasins and onshore continuation. (b) Northeastern part of line BF-51 through the Chignecto and Fundy subbasins. (c) Northeastern part of line 82-28 through the Minas subbasin. (d) Northern part of line 81-47 through the Fundy subbasin and onshore continuation. (e) Northwestern part of line 81-79 through the Fundy subbasin. (f) and (g) Segments of seismic line 3630-1/2-85 through the Emerald/Naskapi basin [*Tankard and Welsink*, 1989]. A and S indicate anticlines and synclines, respectively. Numbered anticlines and synclines are described in text. Thin solid lines are fault surfaces, thick shaded lines show reflections from North Mountain Basalt (Early Jurassic), and striped lines show contact between synrift strata (Middle Triassic to Early Jurassic) and Precambrian/Paleozoic prerift rocks.

distinctive series of closely spaced, large-amplitude events within the upper reflection package in the Minas and Fundy subbasins. Thrust-fault zones of late Paleozoic age probably produce the south to southeast dipping events within the lower reflection package on the northern ends of the seismic lines. *Keen et al.* [1991a] and *Hutchinson et al.* [1988] have observed similar south to southeast dipping reflections in the southern Bay of Fundy and Gulf of Maine, respectively. They have interpreted these reflections as thrust-fault zones associated with the Alleghenian/Variscan front.

Deformation in the Southern Chignecto Subbasin

The seismic data show that the southern Chignecto subbasin is more than 4 km deep and 25 km wide (Plate 1a and Figures 3 and 4). A NE striking, SE dipping normal fault with up to 10 km of displacement bounds the subbasin on the northwest. The boundary fault is gently dipping (about 20°) and relatively planar, although its surface has undulations. The boundary fault overlies and parallels interpreted thrust-fault zones and is probably a reactivated Paleozoic compressional structure. Strata within the Chignecto subbasin dip and thicken toward the boundary fault, indicating that normal faulting occurred during the deposition of the synrift strata during Middle Triassic to Early Jurassic time.

Two major NW plunging folds exist in the hanging wall of the Chignecto boundary fault. One fold (A^1), a gentle anticline about 10 km wide, directly overlies a major undulation on the fault surface (Figure 3 and Plate 1b). Crestal thinning of the synrift strata shows that the anticline developed, at least in part, during the deposition of the synrift strata. Similar folds exist in the hanging walls of the boundary faults of other early Mesozoic rift basins of eastern North America. These folds commonly overlie fault surface irregularities and result from fault bend folding and/or along-strike variations of fault displacement [*Wheeler,* 1939; *Withjack and Drickman Pollock,* 1984; *Schlische,* 1992, 1993]. A second fold (S^1), an asymmetric syncline about 15 km wide, exists near the extreme southwestern end of the Chignecto subbasin (Figure 3 and Plate 1b). Beds in the gentle northern limb (shared with the previously described anticline) dip less than 10° SW, whereas beds in the steep southern limb dip more than 30° NE. The Chignecto boundary fault beneath the steep southern limb is not affected by the folding. The synrift strata consistently thicken toward the southwest across the syncline, showing that the steep southern limb formed after the deposition of the synrift strata. Two different processes could have produced the steep southern limb of the syncline: (1) detachment folding above the Chignecto boundary fault associated with NE-SW shortening, or (2) folding caused by differential displacement on the Chignecto boundary fault. In the latter case, either the northern end of the boundary fault experienced several kilometers of additional normal displacement or the southern end of the boundary fault experienced several kilometers of reverse displacement.

Deformation in the Northern Fundy Subbasin

The northern Fundy subbasin is more than 4 km deep and 70 km wide. East striking, moderately dipping faults with up to several kilometers of normal separation bound the subbasin on the north, whereas NE striking, gently dipping faults with 5 to 10 km of normal separation bound the subbasin on the northwest (Figure 3 and Plates 1b, 1d, and 1e). Like the boundary fault of the Chignecto subbasin, many of the boundary faults of the Fundy subbasin overlie and parallel interpreted thrust faults and are probably reactivated Paleozoic compressional structures. Strata within the subbasin thicken toward the boundary faults, indicating that normal faulting occurred during the Middle Triassic to Early Jurassic deposition of the synrift strata (Plate 1d and Figures 5 and 6a). Data from the Cape Spencer P-79 and Chinampas N-37 wells

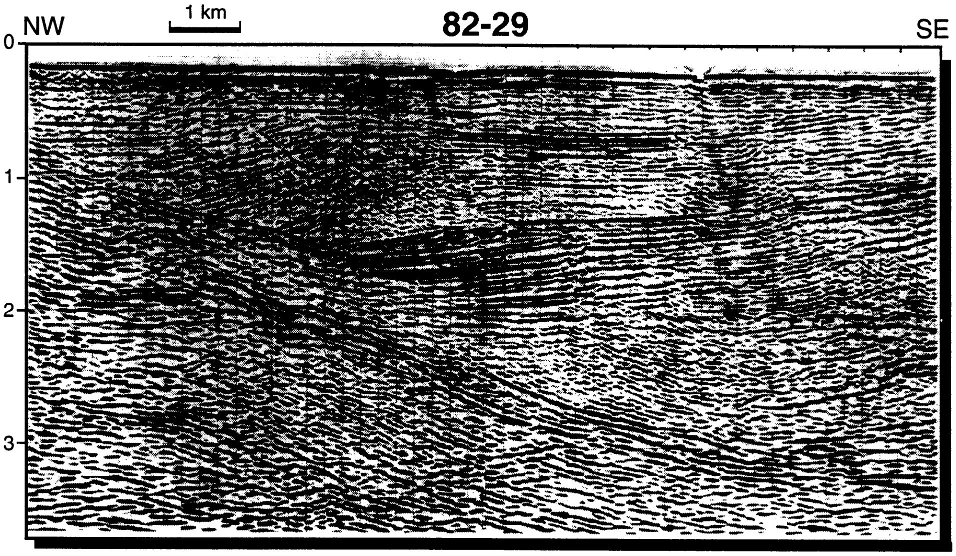

Figure 4. Close-up view of northwestern end of seismic line 82-29 (time-migrated) showing faulted margin of Chignecto subbasin. Location shown in Plate 1a. SE dipping events are fault-surface reflections. NW dipping events are reflections from synrift strata of Middle Triassic to Early Jurassic age. Thickening of synrift strata toward the boundary fault indicates that faulting and deposition were coeval.

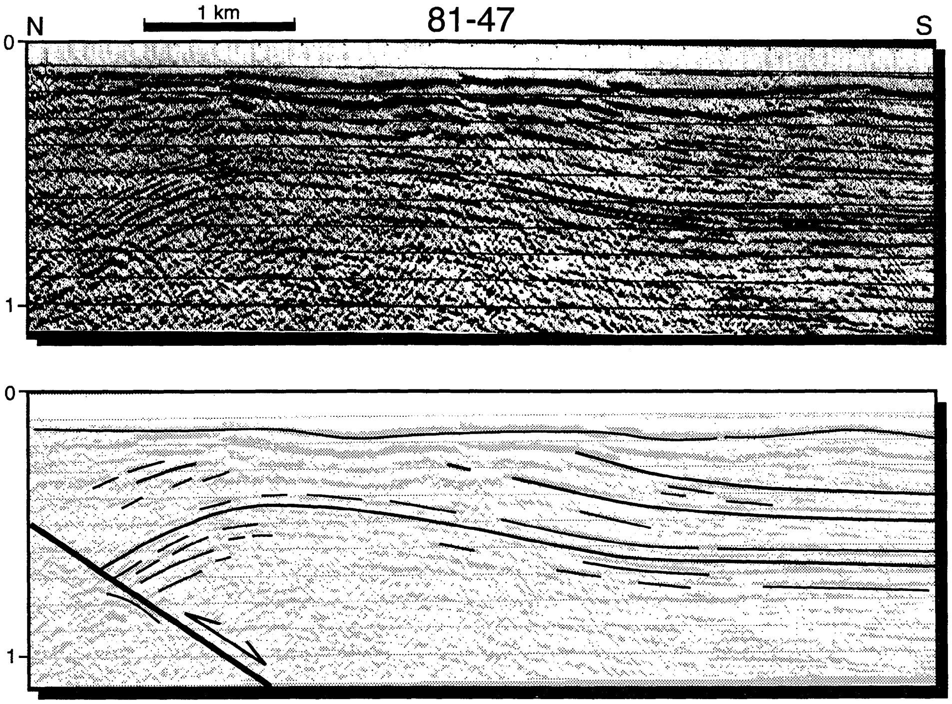

Figure 5. Close-up view of near northern end of seismic line 81-47 (time-migrated). Location shown in Plate 1d. Large-amplitude events at about 0.5 s are reflections from North Mountain Basalt. Overlying events are reflections from synrift strata of Early Jurassic age. Lack of thinning of the synrift strata across the anticline (A^2) indicates that the structure formed after Early Jurassic deposition. See Figure 6a.

and field observations from western Nova Scotia confirm our observation that the synrift strata thicken toward the Fundy boundary faults [*Olsen and Schlische,* 1990; *Pe-Piper et al.,* 1992].

Hanging-wall anticlines parallel the northern and northwestern boundary faults of the Fundy subbasin. Along the northern margin, the anticline (A^2) is east trending, symmetric, and about 3 km wide (Plates 1b and 1d and Figures 3 and 5). Along the northwestern margin, the anticline (A^3) is NE trending, symmetric, and more than 10 km wide (Figure 3 and Plate 1e). The synrift strata consistently thicken toward the northwest across both anticlines, showing that folding occurred after the deposition of the synrift strata (Figure 6a). Anticlines in the hanging walls of normal faults are either rollover folds associated with fault bends [e.g., *Xiao and Suppe,* 1992) or inversion structures produced by the reactivation of normal faults as reverse faults [e.g. *Harding,* 1985; *Mitra,* 1993]. In rollover anticlines, the hanging-wall strata deposited during normal faulting are thinnest near the anticlinal crests. In inversion structures, however, the

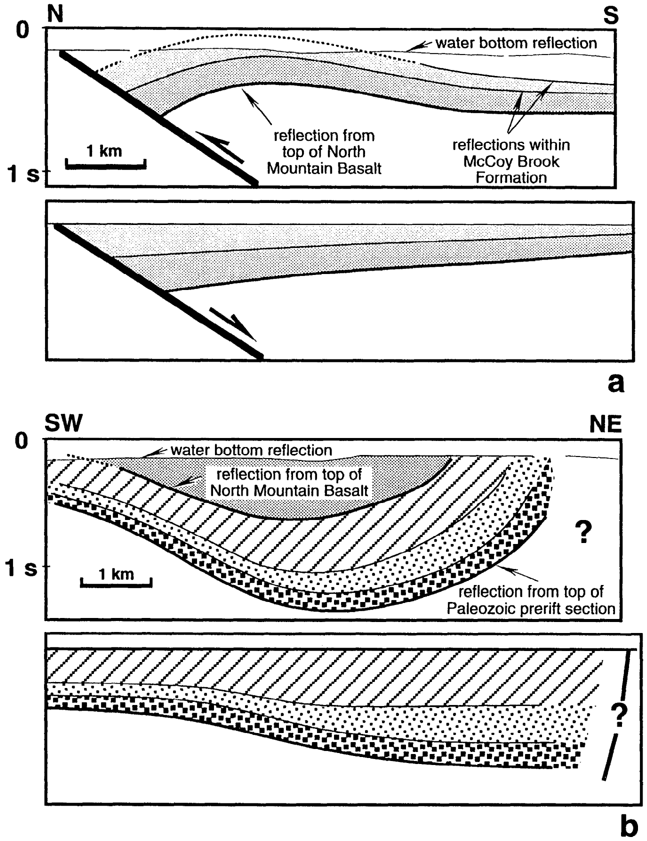

Figure 6. (a) Geometry of synrift strata near northern end of seismic line 81-47 (top) today and (bottom) during deposition. (b) Geometry of synrift strata on northeastern end of seismic line 82-28 (top) today and (bottom) during deposition. Geometries are displayed with no vertical exaggeration assuming a velocity of 3.5 km/s.

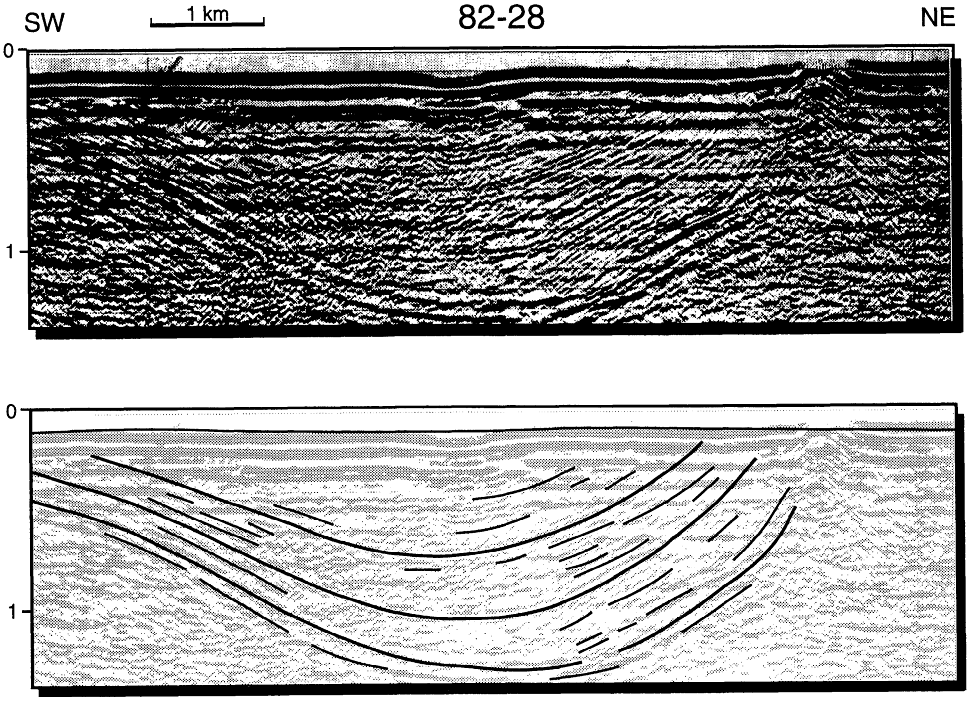

Figure 7. Close-up view of northeastern end of seismic line 82-28 (time-migrated) showing asymmetric syncline (S^3). Most horizontal events are water bottom multiples. Most dipping events are reflections from synrift strata of Middle Triassic to Early Jurassic age. Thickening of synrift strata toward the northeast indicates that faulting occurred during deposition. Lack of thinning of synrift strata on the northern limb of the syncline shows that folding occurred after synrift deposition. See Figure 6b.

hanging-wall strata deposited during normal faulting thicken toward the master normal faults across the anticlinal crests. Consequently, the anticlines in the Fundy subbasin appear to be inversion structures produced by the reactivation of the Fundy boundary faults as reverse faults. A minor, east trending fault with reverse separation also deforms the synrift strata in the northern Fundy subbasin (Figure 3 and Plates 1b and 1d). Thicknesses are constant across the structure, showing that reverse faulting occurred after synrift deposition.

A broad syncline (S^2), about 50 km wide, warps the synrift strata and basement of the Fundy subbasin (Figure 3 and Plate 1d). Our seismic data, together with published seismic lines from the southern Fundy subbasin [*Swift and Lyall,* 1968; *Keen et al.,* 1991a], suggest that this SW plunging syncline extends from the southern to the northern ends of the Fundy subbasin. Synrift strata consistently thicken toward the northwest across the syncline. Apparently, the southeast limb of the syncline reflects the half-graben geometry of the Fundy subbasin during the Middle Triassic to Early Jurassic deposition of the synrift strata. The northwest limb, however, formed later during the development of the anticlines near the northwestern boundary faults of the Fundy subbasin.

Deformation in the Western Minas Subbasin

The western Minas subbasin is asymmetric, bounded on the north by several east striking, south dipping faults or fault zones (Figure 3 and Plates 1a and 1c). In the west, these faults merge with the northern boundary faults of the Fundy subbasin. Generally, the strata thicken toward the north, indicating that some of the Minas boundary faults were active during the Middle Triassic to Early Jurassic deposition of the synrift strata (Figure 6b and 7).

Several asymmetric synclines (e.g., S^3) deform the synrift strata along the northern margin of the Minas subbasin (Plates 1a and 1c and Figures 3 and 7). In the steep northern limbs of these synclines, beds dip more than 40°. Beds in the gentle southern limbs dip about 10°. A broad, east trending syncline exists in the hanging wall of the southernmost boundary fault of the Minas subbasin. This syncline is the eastward continuation of the Fundy syncline and the westward continuation of a west plunging syncline that affects the synrift strata exposed at Cape Blomidon, Nova Scotia (Figure 3). The lack of thinning of the synrift strata on the northern limbs of any of the synclines shows that folding occurred after synrift deposition (Figure 6b).

Summary of Seismic Data

The seismic data show that the Chignecto, Fundy, and Minas subbasins experienced two distinct episodes of deformation during Mesozoic time. The first episode, during Middle Triassic to Early Jurassic time, was extensional and synsedimentary. Extension reactivated Paleozoic compressional structures as normal faults forming the boundary faults of the Chignecto, Fundy, and Minas subbasins. The second episode occurred after the Middle Triassic to Early Jurassic deposition of the synrift strata. Anticlines, synclines, and

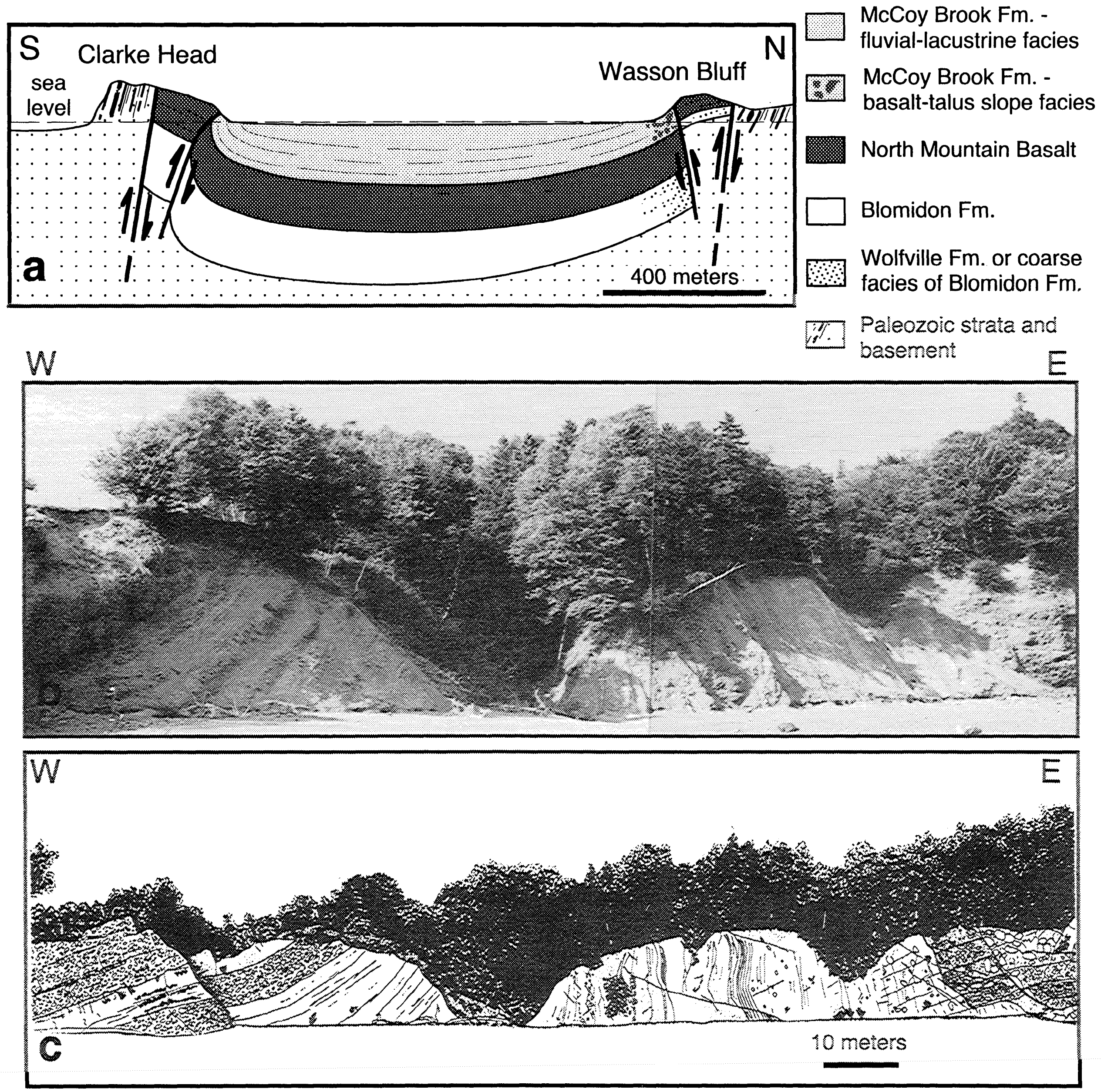

Figure 8. (a) Cross section from Wasson Bluff to Clarke Head, Nova Scotia. A syncline, bounded and cut by numerous faults, deforms the synrift strata. Many faults with reverse separation probably have strike-slip as well as reverse components of displacement. Some faults with reverse separation are actually rotated normal faults. (b) and (c) Photograph and sketch of deformed beds of McCoy Brook Formation at Wasson Bluff [after *Olsen et al.*, 1989]. Eastern beds are vertical to overturned.

faults with reverse separation developed along all faulted margins of the subbasins, and the Fundy and Minas subbasins acquired their synclinal geometries.

Field Data From the Eastern Minas Subbasin

Previous field studies have shown that, during Middle Triassic to Early Jurassic time, the northern margin of the Minas subbasin was a divergent strike-slip fault zone associated with NW-SE extension [*Keppie*, 1982; *Olsen et al.*, 1989; *Olsen and Schlische*, 1990]. Structures within the fault zone include NE trending normal faults and east trending sinistral strike-slip faults [*Olsen et al.*, 1989; *Olsen and Schlische*, 1990]. Evidence of coeval deposition and extensional deformation includes fault-controlled variations in thickness and coarseness of the synrift strata, basalt talus slope deposits within the McCoy Brook Formation, and NE trending neptunian and clastic dikes [*Olsen et al*, 1989; *Olsen and Schlische*, 1990; *Tanner and Hubert*, 1991; *Schlische and Ackermann*, 1995].

Our field studies suggest that a second episode of deformation, one involving horizontal shortening rather than

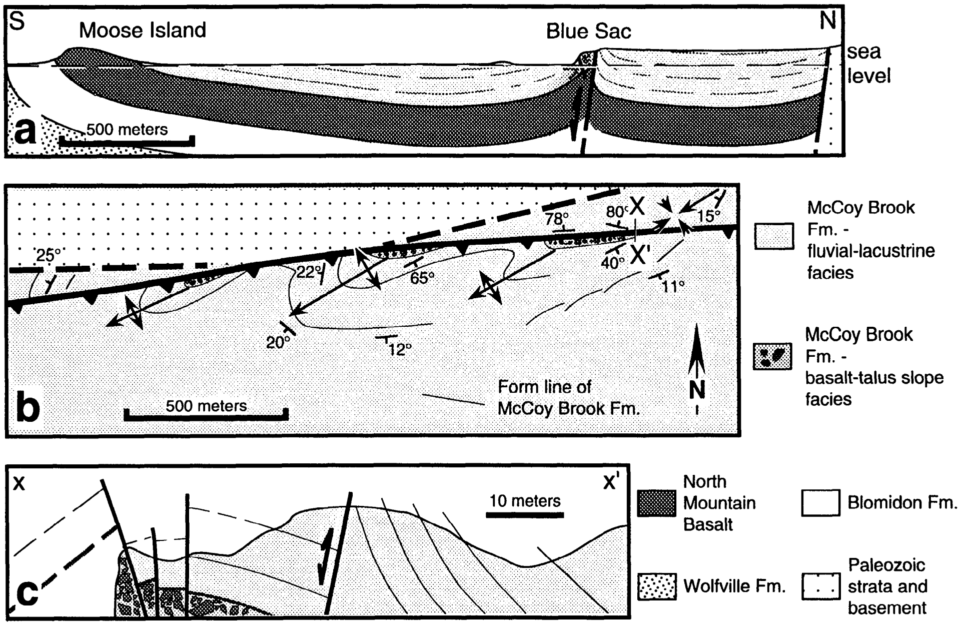

Figure 9. (a) Cross section at Blue Sac, Nova Scotia, with central antiform and surrounding synclines. (b) Map of area near Blue Sac showing NE trending, en echelon anticlines and steeply dipping (about 80°), ENE trending fault with reverse separation. Basalt talus slope deposits occur only in hanging wall, suggesting that the hanging wall was downthrown during deposition and upthrown after deposition. (c) Sketch of outcrop at Blue Sac (cross section X-X' in Figure 9b) showing high-angle fault with reverse separation and steeply dipping beds of McCoy Brook Formation. Minor faults on the southern end of the outcrop, and possibly the high-angle fault with reverse separation, have strike-slip as well as dip-slip components of displacement.

extension, affected the northern margin of the Minas subbasin after the Middle Triassic to Early Jurassic deposition of the synrift strata. The three locations with the clearest evidence of shortening are Wasson Bluff-Clarke Head, Blue Sac, and Five Islands Provincial Park, Nova Scotia (Figure 3).

1. A syncline deforms the synrift strata between Wasson Bluff and Clarke Head (Figure 8a). At Wasson Bluff on the northern limb of the syncline, the entire synrift sequence dips about 30° SE. Many of the normal faults that cut the synrift strata were also affected by this tilting, becoming either low-angle normal faults or high-angle reverse faults. Locally at Wasson Bluff, beds of the McCoy Brook Formation are steeply dipping to overturned (Figures 8b and 8c) and cut by high-angle, NE trending faults with reverse separation. At Clarke Head on the southern limb of the syncline, a NE trending, SE dipping fault with reverse separation emplaces older, gently dipping North Mountain Basalt over younger, steeply dipping and locally overturned beds of the McCoy Brook Formation (Figure 8a).

2. Two broad, ENE trending synclines warp the synrift strata near Blue Sac (Figure 9a). The intervening antiform consists of a series of tight, NE trending anticlines and synclines and an ENE trending fault with at least several tens of meters of reverse separation (Figures 9a, 9b, and 9c). This high-angle fault commonly emplaces basalt talus slope deposits of the McCoy Brook Formation over younger, steeply dipping beds of the same formation. The basalt talus slope deposits occur exclusively in the hanging wall of the fault (Figures 9b), suggesting that the hanging wall was downthrown during deposition and upthrown after deposition.

3. A NE trending, NW dipping fault zone exposed within Five Islands Provincial Park has predominantly normal, down-to-the-NW displacement (Figures 10a, 10b, and 10c). Beds of the Blomidon Formation and North Mountain Basalt are gently dipping on the southeastern edge of the fault zone. On the northwestern edge, however, beds of the North Mountain Basalt and McCoy Brook Formation are steeply dipping. Within the fault zone, high-angle faults with reverse separation emplace North Mountain Basalt over McCoy Brook Formation. Slicken lines indicate that many of these faults have strike-slip as well as reverse components of displacement. East of the fault zone, the North Mountain Basalt and Blomidon Formation, cut by numerous NE trending normal faults, are warped into a series of gentle, NE trending anticlines and synclines. In contrast, several NE trending clastic dikes, more than 10 m long, in the Blomidon Formation attest to NW-SE extension during deposition.

Many of the faults with reverse separation at Wasson Bluff-Clarke Head, Blue Sac, and Five Islands Provincial Park are associated with tight folds and steeply dipping to overturned beds. These faults probably have a reverse component of displacement (i.e., they are not simply rotated normal faults or strike-slip faults with an apparent reverse component of displacement). The abundance of these faults, the tight folds, and the steeply dipping beds suggests that shortening affected the northern margin of the Minas subbasin. The lack of thinning of the synrift strata on the upthrown sides of these faults or across the crests of the anticlines indicates that the shortening occurred after the deposition of the McCoy Brook Formation (i.e., during or after Early Jurassic time). The

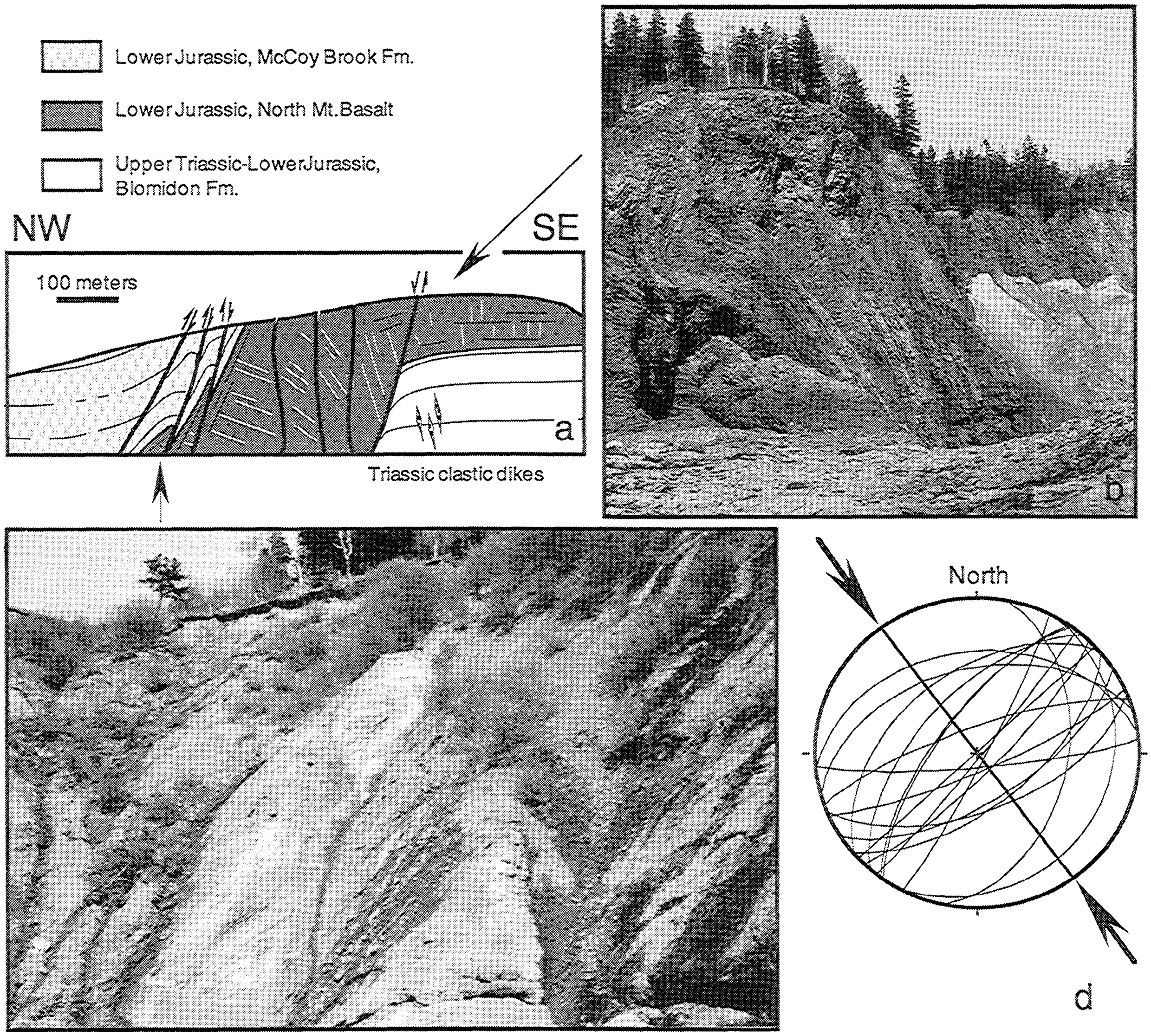

Figure 10. (a) Sketch of fault zone within Five Islands Provincial Park, Nova Scotia. Thin solid lines show bedding, and thin open lines represent cooling joints in the North Mountain Basalt. (b) Beds of Blomidon Formation and North Mountain Basalt are gently dipping on the southeastern edge of the fault zone. (c) On the northwestern edge, beds of the North Mountain Basalt and McCoy Brook Formation are steeply dipping. High-angle faults with reverse separation emplace North Mountain Basalt over McCoy Brook Formation. Many of these faults have strike-slip as well as reverse components of displacement. (d) Equal-area projection showing attitudes of faults with reverse separation and associated strike-slip faults from the northern margin of the Minas subbasin. The indicated direction of postdepositional shortening is perpendicular to the average strike of the faults.

prevalence of east to NE trending fold axes and faults with reverse separation (Figure 10d) suggests that the direction of shortening was approximately NW-SE.

Regional thickness variations also indicate that the Minas subbasin experienced two distinct episodes of deformation. The Wolfville and Blomidon Formations and the North Mountain Basalt are thicker near the center of the Minas subbasin than near the northern margin [*Powers*, 1916; *Klein*, 1962; *Keppie*, 1979; *Olsen et al.*, 1989; *Olsen and Schlische*, 1990]. Hence the center of the Minas subbasin subsided more than the northern margin during the deposition of the synrift strata. Today, however, the tops or projected tops of the Wolfville and Blomidon Formations and the North Mountain Basalt are higher near the center of the subbasin than near the northern margin (Figure 11). Thus the Minas subbasin was elevated relative to its northern margin after the deposition of the synrift strata.

Tectonic Evolution of the Fundy Basin

Both seismic and field data show that the Fundy basin experienced two distinct episodes of deformation during Mesozoic time. The first deformational episode, during Middle Triassic to Early Jurassic time, was extensional and

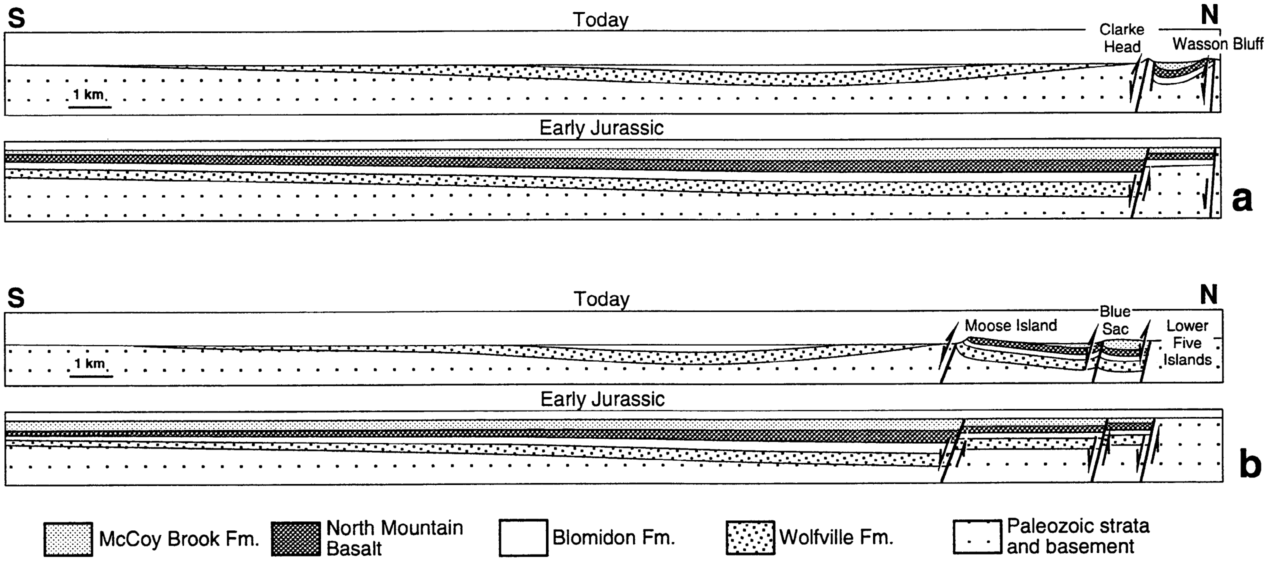

Figure 11. Geologic cross sections through the Minas subbasin. (a) Cross section from Wasson Bluff-Clarke Head to the southern margin (top) today and (bottom) during Early Jurassic deposition. (b) Cross section from Lower Five Islands to the southern margin (top) today and (bottom) during Early Jurassic deposition. Faults may have strike-slip as well as dip-slip components of displacement. Cross sections based on work by *Powers* [1916], *Donohoe and Wallace* [1982], *Olsen et al.* [1989], and *Olsen and Schlische* [1990].

synsedimentary (Figure 12a). Regional NW-SE extension associated with rifting reactivated NE trending Paleozoic compressional structures as normal faults, forming the northwestern boundary faults of the Chignecto and Fundy subbasins. Displacements on the low-angle boundary faults locally exceeded 10 km, and more than 3 km of sediments and extrusives filled the subsiding subbasins near the boundary faults. Extension also reactivated east trending Paleozoic compressional structures along the northern margins of the Fundy and Minas subbasins as oblique-slip fault zones with normal and/or sinistral strike-slip components of displacement [*Olsen and Schlische,* 1990]. More than 1 km of sediments and extrusives filled the Minas subbasin during rifting.

The second deformational episode, affecting all faulted margins of the Fundy basin, occurred after the deposition of the youngest synrift strata (i.e., during or after Early Jurassic time). A NE trending anticline developed along the northwestern margin of the Fundy subbasin; NE and east trending anticlines, synclines, and faults with reverse separation formed along the northern margins of the Fundy and Minas subbasins; the Fundy and Minas subbasins acquired their synclinal geometries; the Minas subbasin rose relative to its northern margin; and a NW plunging syncline with a steep southern limb developed at the southwestern end of the Chignecto subbasin. The simplest interpretation for these postdepositional structures is that widespread shortening, oriented approximately NW-SE, produced them (Figure 12b). With this interpretation, the NE trending anticline along the northwestern margin of the Fundy subbasin and the anticlines, synclines, and faults with reverse separation along the northern margins of the Fundy and Minas subbasins are inversion structures that developed when the boundary faults experienced reverse displacement. Basin inversion associated with shortening elevated the Minas subbasin relative to its northern margin. The steep southern limb of the syncline at the southwestern end of the Chignecto subbasin developed because the northwestern boundary fault of the Fundy subbasin experienced more reverse displacement than the Chignecto boundary fault during inversion. To accommodate this differential displacement, the strata at the southwestern end of the Chignecto subbasin were dragged up the boundary fault and tilted toward the northeast.

Because Quaternary strata overlie the synrift beds in the Fundy basin [*Swift and Lyall,* 1968], the timing of shortening is uncertain. Information from other early Mesozoic rift basins on the passive margin of southeastern Canada, however, can constrain the timing. The Cobequid-Chedabucto fault system bounds the Minas subbasin and its eastern offshore continuation, the Orpheus graben, on the north (Figure 1a). If the faults at the western end of this fault system (i.e., the faults bounding the Minas subbasin) had reverse displacements during inversion, then the faults at the eastern end (i.e., the faults bounding the Orpheus graben) probably experienced similar movements. Previous studies have shown that, except for regional subsidence and minor salt movement, structural activity in the Orpheus graben ceased during Early Cretaceous time [*Tankard and Welsink,* 1989; *Wade and MacLean,* 1990; *MacLean and Wade,* 1992]. Thus any shortening/inversion in the Orpheus graben and, by inference, in the Fundy basin occurred before or during Early Cretaceous time. Virtually all deformation in the Emerald/Naskapi basin, an early Mesozoic rift basin on the Scotian shelf (Figure 1a), occurred before or during the development of the early Middle Jurassic (Aalenian) "breakup" unconformity that separates the synrift and postrift strata [*Welsink et al.,* 1989]. *Welsink et al.* [1989] believe that some structures in this basin are not

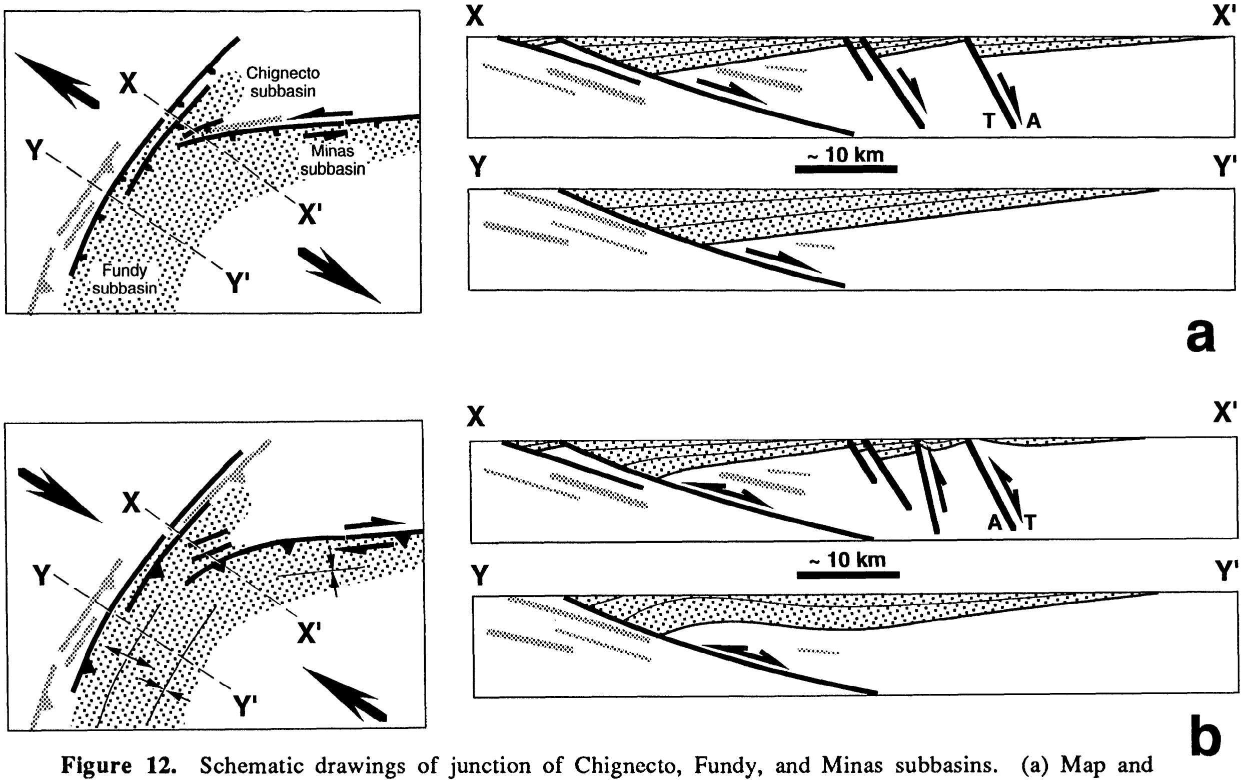

Figure 12. Schematic drawings of junction of Chignecto, Fundy, and Minas subbasins. (a) Map and cross-sectional views during rifting from Middle Triassic to Early Jurassic time. The regional extension direction was approximately NW-SE. (b) Map and cross-sectional views during shortening/inversion during or after Early Jurassic time and before or during Early Cretaceous time. The regional shortening direction was approximately NW-SE.

extensional in origin. Instead, they propose that movement of prerift salt caused folding and created steep stratal dips (Plates 1f and 1g). Alternatively, horizontal shortening may have produced these anomalous structures. If so, then shortening would have occurred in the Emerald/Naskapi basin and, by inference, in the Fundy basin before or during the development of the early Middle Jurassic (Aalenian) unconformity. Hence direct evidence from the Fundy basin shows that shortening occurred during or after Early Jurassic time. Indirect evidence from other early Mesozoic rift basins on the passive margin of southeastern Canada suggests that shortening probably occurred before or during Early Cretaceous time and possibly before or during early Middle Jurassic time.

Displacements During Inversion

The folded synrift beds at the extreme southwestern end of the Chignecto subbasin are about 1.5 km higher than their counterparts within the Chignecto subbasin (Plate 1b). If as discussed previously, differential reverse displacement on the Chignecto and Fundy boundary faults produced this folding, then the vertical component of dip slip on the Fundy boundary fault was about 1.5 km greater than that on the Chignecto boundary fault during inversion. Assuming a 20° dip for the boundary faults, the difference in horizontal dip slip on the Chignecto and Fundy boundary faults was about 4 km during inversion.

The strike-slip component on the Fundy boundary fault during inversion was

$$s_f = h_f / \tan \lambda \qquad (1)$$

where h_f is the horizontal dip slip on the Fundy boundary fault and λ is the angle between the trace of the Fundy boundary fault and the horizontal projection of the displacement direction (Figure 13a). Assuming that the Chignecto boundary fault experienced no displacement during inversion, then the horizontal dip-slip and strike-slip components on the Minas boundary fault zone were

$$\begin{aligned} h_m &= h_f \sin(\lambda - \gamma) / \sin \lambda, \\ s_m &= h_f \cos(\lambda - \gamma) / \sin \lambda, \end{aligned} \qquad (2)$$

respectively, where γ is the angle between the trace of the Fundy and Minas boundary faults (Figure 13a).

Our work suggests that λ was between 30° and 120° during inversion. If $\lambda > 90°$, and assuming that $h_f = 4$ km and $\gamma = 30°$, then the strike-slip component on the Fundy boundary fault was sinistral and less than 2.5 km during inversion. The horizontal dip slip on the Minas fault zone was between 3.5 and 5 km, and the strike slip was dextral and less than 2 km (Figure 13b). If $\lambda < 90°$, then the strike-slip component on the Fundy boundary fault was dextral and less than 7 km during inversion. The horizontal dip slip on the Minas fault zone was less than 3.5 km, and the strike slip was dextral and between 2 and 8 km (Figure 13b).

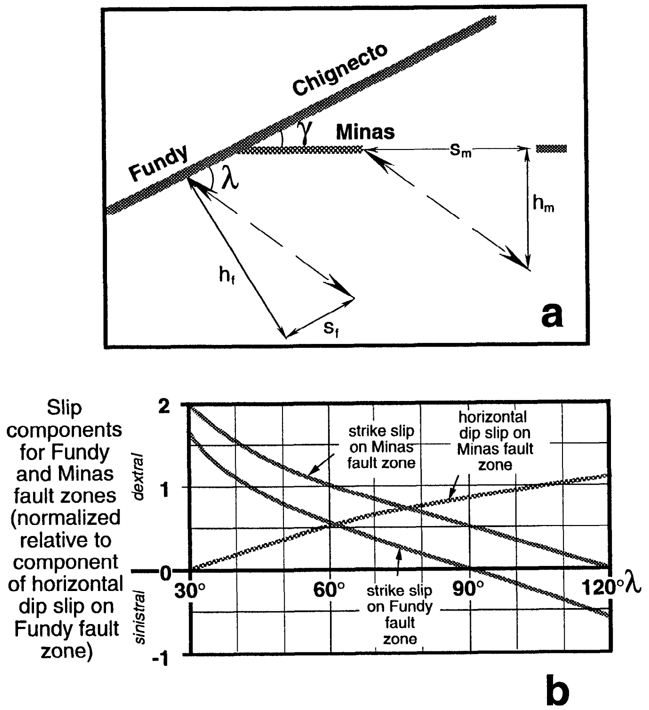

Figure 13. (a) Map view showing horizontal dip-slip and strike-slip components for Fundy boundary fault (h_f and s_f) and Minas boundary fault zone (h_m and s_m). λ is the angle between the trace of the Fundy boundary fault and the horizontal projection of the displacement, and γ is the angle between the traces of the Fundy and Minas boundary faults. (b) Graph showing slip components for Fundy and Minas boundary fault zones as a function of λ. All components are normalized relative to h_f. The angle between the traces of the Fundy and Minas boundary faults, γ, is 30°.

Discussion

Our work demonstrates that deformational styles in the Fundy basin, the westernmost rift basin on the passive margin of southeastern Canada, changed considerably through time. Extensional structures developed during rifting from Middle Triassic to Early Jurassic time. Compressional/inversion structures formed after rifting, during or after Early Jurassic time and probably before or during Early Cretaceous time. As discussed previously, other rift basins on the passive margin of southeastern Canada may have experienced this shortening/inversion. Conclusive evidence, however, is lacking because salt movement has obscured the early deformational history of many rift basins (e.g., the Orpheus graben), and because mild-to-moderate inversion structures commonly resemble extensional and/or salt structures [*Eisenstadt and Withjack,* 1995]. Rift basins also formed on the conjugate continental margin of Morocco during Late Triassic to Early Jurassic time [*Van Houten,* 1977; *Lee and Burgess,* 1978; *Laville and Petit,* 1984; *Beauchamp,* 1988; *Laville,* 1988; *Medina,* 1988; *Laville and Piqué,* 1992]. Recent field studies have shown that compressional structures, similar to those in the Fundy basin, developed in several Moroccan rift basins probably during mid-Jurassic time [*Laville,* 1988; *Laville and Piqué,* 1992].

No collision or subduction zones existed near the passive margin of southeastern Canada during Mesozoic time. Hence we believe that seafloor-spreading processes produced the shortening in the Fundy basin. Several authors have proposed that incipient ridge push forces and/or continental resistance to plate motion can produce shortening on passive margins during the early stages of seafloor spreading [*Dewey,* 1988; *Bott,* 1992]. Two seafloor-spreading events affected the continental margin of southeastern Canada from Early Jurassic to Early Cretaceous time [*Klitgord et al.,* 1988; *Ziegler,* 1989; *Keen et al.,* 1990; *Wade and MacLean,* 1990; *Grant and McAlpine,* 1990]. The first seafloor-spreading event beginning in late Early Jurassic to early Middle Jurassic time separated North America from northern Africa and formed the passive margin of Nova Scotia and the transform margin along the southern edge of the Grand Banks. In Early Cretaceous time, a second seafloor-spreading event separated North America from southern Europe, forming the passive margin along the eastern edge of the Grand Banks. Considering the uncertainty in timing and the proximity of the Fundy basin to both spreading systems, either or both of these seafloor-spreading events could have produced the shortening in the Fundy basin. Consequently, shortening could have occurred during the rift-drift transition as North America separated from northern Africa and/or during the early stages of drifting as the North Atlantic seafloor-spreading centers propagated northward. Several factors favor shortening during the rift-drift transition. Numerical simulations [*Bott,* 1992] and passive margin studies [*Dewey,* 1988] indicate that structural styles can change abruptly from extensional to contractional during this transition. Also, shortening during late Early Jurassic/early Middle Jurassic time can explain the anomalous structures in the Emerald/Naskapi basin that formed prior to the development of the early Middle Jurassic breakup unconformity.

Conclusions

Seismic and field data suggest that the Chignecto, Fundy, and Minas subbasins experienced two distinct episodes of deformation during Mesozoic time. The first deformational episode, during Middle Triassic to Early Jurassic time, was extensional and synsedimentary. Regional extension oriented NW-SE reactivated NE trending Paleozoic compressional structures as normal faults, forming the northwestern boundary faults of the Chignecto and Fundy subbasins. Locally, displacements on the low-angle boundary faults exceeded 10 km. NW-SE extension also reactivated east trending Paleozoic compressional structures along the northern margins of the Fundy and Minas subbasins as oblique-slip fault zones with normal and sinistral strike-slip components.

During the second deformational episode, shortening oriented approximately NW-SE inverted the Chignecto, Fundy, and Minas subbasins. This compressional episode occurred during or after Early Jurassic time and probably before or during Early Cretaceous time. During inversion, the northwestern boundary fault of the Fundy subbasin experienced several kilometers of reverse displacement. In response, an anticline developed along its northwestern margin, the subbasin acquired a synclinal form, and the strata near the southwestern end of the Chignecto subbasin were dragged up the boundary fault and tilted toward the northeast. The east trending fault zones along the northern margins of the Fundy and Minas subbasins became oblique-slip fault zones with reverse and dextral strike-slip components. In response, synclines, anticlines, and faults with reverse separation developed along the northern margins of the Fundy and Minas subbasins.

Our work suggests that not all passive margins have a simple two-stage evolution of rifting and drifting. The western edge of the passive margin of southeastern Canada experienced three stages of development: rifting, shortening during the rift-drift transition and/or during the early phases of drifting, and relative tectonic quiescence during the later phases of drifting. No collision or subduction zones existed near the passive margin of southeastern Canada during Mesozoic time. Hence the seafloor-spreading process itself probably produced the shortening in the Chignecto, Fundy, and Minas subbasins.

Acknowledgments. We thank Mobil Research and Development Corporation for permission to publish this work. Partial support for Paul Olsen came from a grant from the National Science Foundation (EAR 89-16726) to Paul Olsen and Dennis Kent. Field work by Roy Schlische was partially supported by a grant from the Rutgers University Research Council. ShayMaria Silvestri and Rolf Ackermann provided field assistance, and Dave Brown introduced us to the seismic data from the Fundy rift basin and provided many stimulating discussions. Gloria Eisenstadt, Alexander Gates, Peter Hennings, Warren Manspeizer, Kris Meisling, Jan Nowicki, Eric Peterson, Colleen Fitzgerald Reimer, Joana Vizgirda, and an anonymous reviewer critically evaluated the manuscript.

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(Received May 4, 1994;
revised November 8, 1994;
accepted November 18, 1994)