Extensional development of the Fundy rift basin, southeastern Canada

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The Fundy rift basin of Nova Scotia and New Brunswick, Canada, is part of the Eastern North American rift system that formed during the breakup of Pangaea. Integrated seismic-reflection, field, digital-elevation and aeromagnetic data indicate that the Fundy rift basin underwent two phases of deformation: syn-rift extension followed by post-rift basin inversion. Inversion significantly modified the geometries of the basin and its rift-related structures. In this paper, we remove the effects of inversion to examine the basin’s extensional development. The basin consists of three structural subbasins: the Fundy and Chignecto subbasins are bounded by low-angle, NE-striking faults; the Minas subbasin is bounded by E- to ENE-striking faults that are steeply dipping at the surface and gently dipping at depth. Together, these linked faults form the border–fault system of the Fundy rift basin. Most major faults within the border–fault system originated as Palaeozoic contractional structures. All syn-rift units imaged on seismic profiles thicken towards the border–fault system, reflecting extensional movement from Middle Triassic (and possibly Permain) through Early Jurassic time. Intra-rift unconformities, observed on seismic profiles and in the field, indicate that uplift and erosion occurred, at least locally, during rifting. Based on seismic data alone, the displacement direction of the hanging wall of the border–fault system of the Fundy rift basin ranged from SW to SE during rifting. Field data (i.e. NE-striking igneous dykes, sediment-filled fissures and normal faults) indicate NW–SE extension during Early Jurassic time, supporting a SE-displacement direction. With a SE-displacement direction, the NE-striking border–fault zones of the Fundy and Chignecto subbasins had predominantly normal dip slip during rifting, whereas the E-striking border–fault zone of the Minas subbasin had oblique slip with left-lateral and normal components. Sequential restorations of seismic-reflection profiles (coupled with projections from onshore geology) show that the Fundy rift basin underwent 10–20 km of extension, most of which was accommodated by the border–fault system, and was considerably wider and deeper prior to basin inversion. Post-rift deformation tilted the eastern side of the basin to the northwest/north, producing significant uplift and erosion. Copyright © 2009 John Wiley & Sons, Ltd.

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1. INTRODUCTION

During Mesozoic time, a massive rift zone developed within the Pangaeanean supercontinent (inset, Figure 1) (e.g. Olsen 1997). The breakup of Pangaea splintered this rift zone into fragments, each now separated and preserved on the passive margins of eastern North America, northwestern Africa and Europe. The fragment on the North American margin, called the Eastern North American rift system, extends from northern Florida to the eastern Grand Banks of Canada (e.g. Manspeizer and Cousminer 1988; Olsen et al. 1989; Schlische 1993, 2003; Withjack et al. 1998; Withjack and Schlische 2005) (Figure 1). It is one of the world’s largest rift systems and occupies one of the world’s oldest intact passive margins. Furthermore, the Eastern North American rift system includes part of the

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Central Atlantic Magmatic Province (CAMP) (Marzolli et al. 1999; Hames et al. 2003), a short-lived (<1 m.y.; Olsen et al. 1997, 2002; Whiteside et al. 2007), one of the largest mass-extinction events (Sepkoski 1993).

Based on the age range of preserved syn-rift strata, Withjack and Schlische (2005) divided the Eastern North American rift system into three geographic segments (Figure 1). Rifting was underway throughout the entire rift system (i.e. the southern, central and northern segments) by Late Triassic time (e.g. Olsen 1997). The cessation of rifting and the onset of seafloor spreading was diachronous, occurring first in the southern segment (latest Triassic/earliest Jurassic), then in the central segment (Early to Middle Jurassic) and finally in the northern segment (Early Cretaceous) (Withjack et al. 1998; Withjack and Schlische 2005). After rifting, many of the rift basins in the southern and central segments of the Eastern North American rift system were inverted (i.e. underwent shortening), which significantly modified the geometries of the original rift-related structures (Withjack et al. 1998; Schlische 2003; Schlische et al. 2003; Withjack and Schlische 2005; and references therein).
The Fundy rift basin is the northernmost basin in the central segment of the Eastern North American rift system (Figure 1). Nearly a century of research has provided valuable information about its complex tectonic history involving Palaeozoic shortening, early Mesozoic rifting and post-rift inversion (e.g. Powers 1916; Klein 1960, 1962; Keppie 1982; Plint and van de Poll 1984; Nadon and Middleton 1984; Olsen et al. 1989; Olsen and Schlische 1990; Schlische and Ackermann 1995; Withjack et al. 1995; Wade et al. 1996; Baum et al. 2008). In this study, we focus on the extensional development of the Fundy rift basin. We have: (1) synthesized the results of previous studies; (2) re-interpreted 2D seismic data from the Bay of Fundy, including some newly processed lines; (3) interpreted new 2D seismic data from onshore Nova Scotia; (4) mapped, in detail, cliff and tidal-flat exposures from onshore Nova Scotia, some of which were not well exposed during previous studies and (5) incorporated new high-resolution DEM data (Nova Scotia Department of Natural Resources 2004; Canadian Council on Geomatics 2008) and aeromagnetic data (King 2005a, b; Oneschuk and Dumont 2005a, b) from the Bay of Fundy region. Furthermore, we have restored several key cross sections from the Fundy rift basin to the time of rifting. These sequential restorations show how the geometry of the Fundy rift basin changed through time during rifting.

2. OVERVIEW OF FUNDY RIFT BASIN

The Fundy rift basin has three structural components: the Chignecto, Fundy and Minas subbasins (Olsen and Schlische 1990; Figure 2b). The Chignecto and Fundy subbasins are bounded on the northwest by the NE-striking Chignecto and Fundy border–fault zones, respectively; the Fundy and Minas subbasins are bounded on the north by the E- to ENE-striking, Cobequid/Chedabucto border–fault zone (also known as the Minas fault zone) (Powers 1916; Keppie 1982; Nadon and Middleton 1984; Plint and van de Poll 1984; Olsen and Schlische 1990; Withjack et al. 1995; Wade et al. 1996). Much of the Cobequid/Chedabucto border–fault zone is exposed in the eastern Minas subbasin and is visible on shaded-relief maps of the region (Figure 3a). Together, the Chignecto, Fundy and Cobequid/Chedabucto border–fault zones form a composite border–fault system consisting of linked NE- and E-striking border–fault zones. Most faults within this border–fault system originated as Palaeozoic contractional structures (Keppie 1982; Plint and van de Poll 1984; Brown 1986; Withjack et al. 1995).

In the Fundy basin, rifting began by Middle Triassic time or possibly earlier in Permian time (Olsen et al. 1989; Olsen et al. 2000; Olsen and Et-Touhami 2008). During rifting, the basin subsided and filled with several kilometres of nonmarine sedimentary rocks and basalt flows associated with CAMP (e.g. Kontak 2008). These syn-rift rocks overlie mildly to intensely deformed Carboniferous and older sedimentary, igneous and metamorphic rocks. Olsen et al. (2000) and Olsen and Et-Touhami (2008) identified four tectonostratigraphic sequences in the Fundy rift basin and used these sequences (rather than purely lithostratigraphic criteria) to subdivide the syn-rift section (Figure 2d). The syn-rift section, based on this subdivision, consists of the following units, from oldest to youngest: (1) the Honeycomb Point Formation, not well dated but possibly as old as Permian and comprised predominantly of alluvial-fan and aeolian deposits; (2) the Middle to Upper Triassic Wolfville Formation, comprised predominantly of fluvial deposits with minor aeolian beds; (3) lacustrine, fluvial and aeolian rocks of the Upper Triassic/Lower Jurassic Blomidon Formation; (4) the Lower Jurassic North Mountain Basalt, dated at 202 ± 2 Ma (Hodych and Dunning 1992) and (5) the Lower Jurassic fluvio-lacustrine McCoy Brook Formation, which also contains minor aeolian and talus-slope facies. Seismic, well and outcrop data indicate that these syn-rift units generally thicken towards the NE-striking Fundy and Chignecto border–fault zones (Figure 3c) and have considerable thickness variations within the Cobequid/Chedabucto border–fault zone of the Minas subbasin (Figure 2c). Rifting ceased in Fundy rift basin in the Early to Middle Jurassic when continental breakup occurred and seafloor spreading began (Klitgord and Schouten 1986; Klitgord et al. 1988; Tankard and Welsink 1989; Withjack et al. 1998; Withjack and Schlische 2005).

After rifting, the Fundy rift basin underwent post-depositional inversion (Withjack et al. 1995) during which the hanging wall of the linked border–fault system moved ~NNE relative to the footwall (Baum et al. 2008). This oblique inversion significantly altered the geometry of the Fundy rift basin by producing numerous folds and left-lateral strike-slip faults, most of which are subparallel to the adjacent border–fault zones (Figure 3b,c). The most
Figure 2. Geology of the Fundy basin. a: Shaded-relief map showing topography and main geographic features in the area around the Fundy basin. Based on data from the Canadian Council on Geomatics (2008). b: Shaded-relief map showing the approximate extent of the Fundy rift basin (grey), its principal boundary faults (tick marks indicate dip direction), and its three subbasins. Modified from Olsen and Schlische (1990). c: Variations in syn-rift stratigraphic thickness. Locations shown in (b). Modified from Olsen and Schlische (1990). d: Stratigraphy of the Fundy basin region, illustrating pre-rift, syn-rift and post-rift units, major unconformities and syn-rift formations and their ages as defined by Olsen et al. (2000) and Olsen and Et-Touhami (2008). The stratigraphic subdivisions, which differ from previous subdivisions, reflect tectonostratigraphic rather than lithostratigraphic criteria.
Figure 3. a: Map of northern Fundy rift basin showing junction of three structural subbasins (Fundy, Minas and Chignecto), geographic features, seismic coverage and location of cross-section G1. Shaded-relief map based on data from Canadian Council on Geomatics (2008). b, c: Geologic map and cross sections through northern Fundy rift basin based on seismic, outcrop and well data (Withjack et al. 1995; Wade et al. 1996), seafloor topography and subcrop information (Swift and Lyall 1968) and aeromagnetic data (King 2005a, b; Oineschuk and Dumont 2005a, b). Seismic sections are displayed without vertical exaggeration assuming a velocity of 3.5 km s⁻¹.
prominent of these inversion-related folds is a gentle, SW-plunging, basin-scale syncline (S1) whose axial trace is subparallel to the Fundy and Cobequid/Chedabucto border–fault zones (Figure 3b). This syncline produces the hook-shaped ridge of North Mountain Basalt in the Cape Blomidon region (Figure 3a).

3. SEISMIC EVIDENCE FOR EXTENSIONAL HISTORY OF THE FUNDY RIFT BASIN

3.1. Seismic data

More than 1200 km of seismic-reflection profiles, acquired by industry (Chevron and Mobil) from 1980 to 1982, cover the offshore portion of the Fundy rift basin (Figure 3a) (Withjack et al. 1995; Wade et al. 1996; Baum et al. 2008). Reprocessing of several key profiles has suppressed multiples and improved imaging of subsurface geometries. Additionally, recently acquired seismic-reflection profiles (Consolidated Beacon Resources Ltd) provided coverage for the onshore Minas subbasin (Baum et al. 2008). Projections from onshore geology (Keppie 1979; Donahoe and Wallace 1982; Withjack et al. 1995; Wade et al. 1996), offshore subcrops (Swift and Lyall 1968), industry well data (Mobil Oil Canada Ltd 1975; Chevron Canada Resources Ltd 1984) and high-resolution aeromagnetic data from the Fundy region (King 2005a, b; Oneschuk and Dumont 2005a, b) constrain our interpretation of these seismic data.

3.2. Seismic interpretation

The two main seismic packages in the Fundy rift basin are a lower, pre-rift package (Carboniferous and older) and an upper, syn-rift package (possibly Permian, definitely Middle Triassic into Early Jurassic) (Withjack et al. 1995; Wade et al. 1996). Reflections in the pre-rift package are discontinuous and show considerable variability in dip direction, whereas the syn-rift package is characterized by relatively continuous, subparallel reflections. The North Mountain Basalt generates a distinctive series of closely spaced, large-amplitude events within the syn-rift package of the Fundy subbasin.

The NE-striking Fundy and Chignecto border–fault zones are composed of a series of low-angle (10–30°), planar to slightly listric, SE-dipping faults (Figures 3c and Figure 4) that likely are reactivated thrust faults (Plint and van de Poll 1984; Brown 1986; Withjack et al. 1995). The E-striking, S-dipping Cobequid/Chedabucto border–fault zone that bounds the Minas subbasin is composed of a series of high-angle, poorly imaged, upper fault segments above a gently dipping lower segment (Figure 3c and Figure 5a,b). The syn-rift strata consistently thicken towards the NE- and the E-striking border–fault zones, indicating that all border–fault zones were active during syn-rift deposition (i.e. possibly Permian time, definitely from Middle Triassic into Early Jurassic time) (Figure 3c). For example, seismic-line BF-20 shows that syn-rift strata thicken and dip gently towards the NE-striking Chignecto border–fault zone (Figure 4a,b). Thickening towards the border–fault zone is most pronounced in the oldest syn-rift units. A restoration of seismic-line 82–28 to the time of rifting (which removes the post-rift folding related to inversion) shows that syn-rift strata thicken towards an E-striking fault within the Cobequid/Chedabucto border–fault zone (Figure 5c).

Although most folds in the Fundy rift basin formed after rifting during post-depositional basin inversion, some folds developed during rifting. For example, anticline A5 within the Chignecto subbasin plunges northwest and trends nearly orthogonally to the NE-striking border–fault zone (Figure 3b). The anticline is up to 10 km wide, and its amplitude decreases away from the border–fault zone. It occurs above a structural high on the border–fault surface and shows some subtle crestal thinning of syn-rift strata, indicating that folding occurred, at least partially, during deposition (Withjack et al. 1995; Withjack and Schlische 2005). Similar folds exist in the hanging walls of the border faults of other rift basins and result from along-strike variations in fault displacement during rifting (Wheeler 1939; Withjack and Drickman-Pollock 1984; Schlische 1995; Withjack et al. 2002).

Seismic profiles through the Chignecto subbasin, where the syn-rift units are best imaged, show that an angular, rift-onset unconformity separates the syn-rift and pre-rift packages. For example, an angular unconformity,
separating SE-dipping syn-rift strata from subhorizontal pre-rift strata, is well imaged on seismic-line BF-20 (Figure 4a,b). A subtle, angular syn-rift unconformity, separating the oldest syn-rift strata from overlying syn-rift strata, is also imaged on BF-20 (Figure 4a,b). It is unclear whether the syn-rift units below this syn-rift unconformity are associated with the Honeycomb Point Formation or the Wolfville Formation (or their time-equivalents), both of which crop out northwest of the seismic line (Nadon and Middleton 1984, 1985; Plint and van de Poll 1984; Olsen and Et-Touhami 2008) (Figure 3b). The seismic profiles through the Chignecto subbasin also show that a series of high-amplitude reflections exist within the deeper part of the syn-rift section near the border–fault zone (Figure 4a,b). Thus, the seismic characteristics of the syn-rift units (and presumably their lithologic compositions) change with depth and with proximity to the border–fault zone (Withjack et al. 1995; Wade et al. 1996), indicating that the border–fault zone was active during deposition of these syn-rift units.

Figure 4. a: Part of seismic-line BF-20 (uninterpreted and displayed with no vertical exaggeration assuming a velocity of 3.5 km s$^{-1}$). See Figure 3 for location. b: Same part of seismic-line BF-20 (interpreted), showing Chignecto border–fault zone and syn-rift strata. c: Sketch of seismic-line BF-20 today and restored through time. Restoration is schematic with assumptions of line-length conservation and 2 km of uplift and erosion after rifting (Wade et al. 1996). Any footwall relief that existed during rifting is not shown. Restorations show that the Chignecto border–fault zone accommodated substantial extension during rifting, and its hanging wall subsided considerably during rifting.
3.3. Restorations of the seismic profiles

Restorations of key seismic profiles from the Fundy rift basin (Figures 4c and 5c,d) show that the NE-striking, gently dipping, border–fault zones of the Chignecto and Fundy subbasin accommodated a significant amount of extension during rifting (~10–20 km). During rifting, the hanging walls of all border–fault zones (i.e. the NE- and E-striking zones) subsided relative to the footwalls. Subsidence, although greatest near the border–fault zones, was widespread, producing a very wide, deep rift basin (Figure 5d). Thus, the Fundy rift basin was considerably wider during and immediately after rifting than it is today, with the southeastern/southern side of the basin now eroded away. Shortening after rifting produced inversion-related structures near the border–fault zones (e.g. anticlines and synclines). Post-rift deformation also tilted the eastern side of the basin to the northwest/north and produced regional uplift and erosion. Thus, post-rift deformation substantially modified the original geometry and dimensions of the Fundy rift basin.

4. FIELD EVIDENCE FOR EXTENSIONAL HISTORY OF FUNDY RIFT BASIN

Extensive cliff and tidal-flat exposures on the northern margin of the eastern Minas subbasin (Figure 6) permit detailed study of the small-scale structures within the Cobequid/Chedabucto border–fault zone. This border–fault zone is an approximately 10-km-wide zone composed of ENE- and NE-striking, steeply dipping faults (Figure 6b). Onshore seismic data show that these high-angle faults overlie a gentle, south-dipping fault (Figure 6c) (Baum et al., 1996).
Figure 6. a: Shaded-relief map of the central Minas subbasin. Compiled from images available from Nova Scotia Bureau of Mineral Resources (2004). b: Geologic map of same region as (a). Modified from Keppie (1979) and Donahoe and Wallace (1982). c: Cross section of central Minas subbasin (see (b) for location). T is motion towards viewer; A is away.
A series of inversion-related folds are present between and adjacent to the faults within the Cobequid/Chedabucto border–fault zone. Hook-shaped ridges of basalt are the topographic expression of some of these folds (e.g. at Clarke Head and to the northeast of the Five Islands region; Figure 6a).

4.1. Five Islands Region

The syn-rift units in the Five Islands region (Figure 7) are the Blomidon Formation, North Mountain Basalt and McCoy Brook Formation. In the Blomidon cliffs region between Red Head and Old Wife Point, a series of NE-striking, moderate-angle faults cut the syn-rift rocks (Figures 7, 8). The faults exhibit normal separation and have dip-slip slickenlines. These normal faults likely formed during rifting. Furthermore, they formed during and/or after the emplacement of the North Mountain Basalt because (1) they offset the lower contact of the North Mountain Basalt and (2) growth beds are not present in the faulted Blomidon Formation. Unlike the NE-striking faults at Old Wife Point that show predominantly left-lateral slickenlines (commonly overprinting dip-slip slickenlines), the normal faults in the Blomidon cliffs region show no evidence of reactivation during basin inversion.

4.2. Blue Sac

The Lower Jurassic McCoy Brook Formation is present on both sides of the E-striking, S-dipping Blue Sac fault (Figure 7a). Bedding dips are steep (and locally overturned) near the fault. These steep dips developed after deposition during inversion (Withjack et al. 1995; Baum et al. 2008). On the north side of the fault, the McCoy Brook Formation consists of sandstone, mudstone and minor conglomerate. On the south side, the McCoy Brook Formation consists of basalt–clast breccia (see Figure 10c) adjacent to the fault and sandstone and mudstone elsewhere.

The basalt–clast breccia is composed of mostly angular, clast-supported cobbles and boulders of basalt. The variety of clasts suggests that multiple flow units contributed to this facies. The large size of the clasts and their angularity suggest that transport was minimal and that they accumulated at the base of a cliff. Because these deposits are located adjacent to a fault, the cliff was most likely a fault scarp (Tanner and Hubert 1991). Thus, the basalt–clast breccia accumulated as a direct consequence of faulting and is a growth deposit (Figure 9). After the eruption and widespread emplacement of the North Mountain Basalt, activity on the Blue Sac fault dropped the basalt to the south, forming a fault scarp (Figure 9b). In response, basalt–clast breccia accumulated on the downthrown side directly adjacent to the fault, whereas sandstone and mudstone accumulated farther to the south. Sandstone and mudstone of the McCoy Brook Formation eventually covered the North Mountain Basalt, ending the accumulation of the basalt–clast breccia (Figure 9c). During basin inversion, the south side of the Blue Sac fault moved upward, folding the McCoy Brook Formation next to the fault (Figure 9d). Slickenlines on minor faults with the same attitude as the Blue Sac fault indicate that the Blue Sac fault had both reverse and left-lateral strike-slip components of displacement during inversion (Baum et al. 2008).

4.3. Wasson Bluff

Wasson Bluff (Figure 10) is located on the north limb of an ENE-plunging syncline expressed as the hook-shaped ridge of North Mountain Basalt to the north of Clarke Head (Figure 6a). Consequently, most strata at Wasson Bluff dip to the south (Figure 10a). Most faults at Wasson Bluff are steeply dipping (>70°). Faults 1, 3, 4 and 6 have gently raking slickenlines, and kinematic indicators show that their last movement was predominantly left-lateral strike-slip. Most folding and some faulting at Wasson Bluff occurred after deposition during basin inversion (Withjack et al. 1995). As discussed below, however, some deformation at Wasson Bluff occurred during rifting.

At Wasson Bluff, the Blomidon Formation unconformably overlies Carboniferous pre-rift rocks (Figure 10a). The Wolfville Formation is absent. The Blomidon Formation varies considerably in thickness and facies (Olsen and Schlische 1990). It is only a few metres thick on the south side of fault 2, and consists of sandstone and
conglomerate. It is considerably thicker on the south side of fault 7 and at nearby Clarke Head (Figure 2c), where it consists mostly of mudstone and sandstone. These thickness variations suggest different subsidence rates of the fault blocks at Wasson Bluff during the deposition of the Blomidon Formation. Basalt–clast breccia within the McCoy Brook Formation occurs at three locations (Olsen et al. 1989; Olsen and Schlische 1990; Tanner and Hubert 1991): on the south side of fault 3 (Figure 10c), on the east-southeast side of fault 5, and on the south side of fault 6.
As at Blue Sac, these are growth deposits, indicating that the adjacent faults were active during deposition of the McCoy Brook Formation (Olsen and Schlische 1990). Further evidence of faulting during deposition includes the thickening of the strata of the McCoy Brook Formation towards fault 5 (Figure 10b; Olsen et al. 1989; Olsen and Schlische 1990). Evidence of this growth is reflected by the divergence of bedding strike in the hanging wall of fault 5: the basal McCoy Brook Formation and its contact with the underlying North Mountain Basalt strike WNW, whereas younger strata (expressed by ridges of sandstone in the tidal flat) strike ENE (Figure 10a).

Sediment-filled fissures at Wasson Bluff provide evidence that constrains the timing and direction of extension (Schlische and Ackermann 1995). These fissures are common near the top of the North Mountain Basalt and occur along joints in the columnar basalt (Figure 10d). These fissures, which contain sedimentary material similar to the

Figure 8. Photograph and interpreted line drawing of normal faults (based on separation and slickenlines) exposed in the Blomidon cliffs between Red Head and Old Wife Point (see Figure 7 for location). Dashed line is contact between Blomidon Formation and North Mountain Basalt and its projected location in eroded part of cliff. Note the baked zone (light) in the uppermost Blomidon Formation.
basal McCoy Brook Formation, have a wide variety of attitudes (reflecting the polygonal cooling joints), but the NE-striking fissures are widest (Figure 10d). These features indicate NW–SE extension after the cooling of the North Mountain Basalt and during deposition of the basal McCoy Brook Formation.

4.4. Carrs Brook

Our geologic map of Carrs Brook (Figure 11a), based on the presence of three distinctive structural and lithologic domains, differs from the geologic maps of Donahoe and Wallace (1982) and Olsen et al. (2000). Domain 1 consists of NE-striking and moderately to steeply NW-dipping pre-rift strata of the Riversdale Group. Domain 2 consists of E- to ESE-striking and moderately to steeply S- to SSW-dipping syn-rift strata of the lower Wolfville Formation. Domain 3 consists of N- to NNW-striking and gently W- to WSW-dipping syn-rift strata. Previously mapped as Wolfville Formation by Donahoe and Wallace (1982), Olsen et al. (2000) assigned the strata in this domain to the lower Blomidon Formation. We concur with this interpretation because the conglomerates in domain 2 (clast-
supported, calcite-cemented, composed of well-rounded clasts) differ distinctly from those in domain 3 (very poorly sorted, matrix-supported, composed of very angular clasts).

Like Donahoe and Wallace (1982) and Olsen et al. (2000), we interpret the ENE-striking contact between the Riversdale Group and the Wolfville Formation as a fault with normal separation. Bedding in both units steepens towards the fault contact and rotates into parallelism with it. The folds in both units adjacent to the fault contact are likely fault-propagation folds. Olsen et al. (2000) interpreted the WNW-striking contact between the Riversdale Group and the Wolfville Formation as an angular unconformity. However, as with the ENE-striking contact, bedding steepens and rotates into parallelism with the contact, suggesting that this contact is a splay of the main
Figure 11. Geology of the Carrs Brook region (see Figure 6b for location). a: Geologic map and cross sections showing fault zone (and associated fault-propagation folds) between the pre-rift Riversdale Group and the Wolfville Formation, the inferred angular unconformity between the same units at depth, and the angular unconformity between the Blomidon Formation and older units. Based, in part, on Baum (2002). b: Photograph viewed to the west of moderate to steeply S-dipping bedding in the Wolfville Formation immediately south of main fault zone. Cliff is ~25 m high. c: Schematic geologic evolution of structures similar to those along cross-section E–E'. The border fault that controlled deposition of the Wolfville and Blomidon formations was probably located several kilometres to the north or northwest of this area. See text for further discussion.
Neither the Blomidon/Wolfville contact nor the Blomidon/Riversdale contact is exposed at Carrs Brook. The dip of bedding in the Blomidon Formation differs from that in the adjacent Wolfville Formation and Riversdale Group, but does not change substantially near the contact (Figure 11a). This suggests that the Blomidon/Wolfville/Riversdale contact is an angular unconformity not a fault. The relationship between this unconformity and the faults at Carrs Brook is unclear. We infer, however, that the unconformity overlaps the faults because bedding dips in the Wolfville Formation and the Riversdale Group change significantly near the faults, whereas those in the Blomidon Formation do not change near the faults (Figure 11a). We cannot exclude the possibility that relatively minor movement on these faults later offset the unconformity (a separate minor fault with left-lateral separation does offset the unconformity).

Using cross-section E–E' (Figure 11a) as a guide, we have schematically reconstructed the geologic history of the Carrs Brook region (Figure 11c). Following deposition (Figure 11c-i) and initial deformation (Figure 11c-ii) of the Carboniferous Riversdale Group, the main fault propagated upward through the Riversdale Group and produced a set of fault-propagation folds related to faulting with a reverse component of slip (Figure 11c-iii). Following uplift and erosion, the Wolfville Formation was deposited on top of the eroded Riversdale Group and the main fault (Figure 11c-iv). The rift-onset unconformity, although not exposed at Carrs Brook, crops out at Tennycape, Nova Scotia (Figure 6b; see Withjack et al. 2002, figure 16; Olsen and Et-Touhami 2008, figure 50; Leleu et al. 2009), where gently NNW-dipping strata of the Wolfville Formation progressively onlap subvertical Carboniferous strata. Following deposition of the Wolfville Formation (as there is little or no evidence of growth beds in the Wolfville Formation adjacent to the main fault; Figure 11b), the main fault was reactivated with a normal component of slip and propagated upward and folded the Wolfville Formation (Figure 11c-v). Following another period of erosion and/or nondeposition, the lower Blomidon Formation was deposited (Figure 11c-vi) and subsequently tilted (Figure 11c-vii). The tilting may have occurred during rifting and/or basin inversion. Exposures of an angular unconformity between the Wolfville and lower Blomidon formations on the south side of Pinnacle Island (Olsen et al. 2000, figure 4b; Olsen and Et-Touhami 2008, figure 48) also suggest a period of uplift and erosion during this time interval.

5. DISCUSSION

The seismic data from the Fundy rift basin provide information about the geometries of large-scale structural and stratigraphic features within the basin (e.g. the geometries of major faults and folds, regional unconformities and growth beds). In contrast, field data provide information about small-scale structural and stratigraphic features (i.e. slip indicators on fault surfaces, fracture orientations and depositional patterns and facies). Thus, the seismic and field data are complementary and together provide a more complete picture of the extensional development of the Fundy rift basin.

5.1. Timing

The seismic data show that all border–fault zones of the Fundy rift basin (i.e. the Chignecto, Fundy and Cobequid/Chedabucto) were active during rifting (Figure 12). Specifically, the hanging walls of the linked border–fault zones subsided relative to the footwalls, producing a very wide, deep rift basin to the south/southeast of the border–fault zones (Figure 5d). Subsidence was greatest adjacent to the NE-striking Chignecto and Fundy border–fault zones, especially during the early stages of rifting. Projections from onshore geology and industry well data indicate that rifting began by Middle Triassic time (and possibly during Permian time) and continued into Early Jurassic time. Both seismic and field data show that an angular unconformity (rift-onset unconformity; e.g. Olsen et al. 1989; Withjack et al. 2002; Leleu et al. 2009) separates the Carboniferous and older pre-rift rocks from the syn-rift rocks throughout the Fundy rift basin. The variable thickness of the Blomidon and McCoy Brook formations at Wasson Bluff as well as the presence of basalt–clast breccia in the McCoy Brook Formation at Blue Sac and Wasson Bluff verify that rifting was active during Late Triassic and Early Jurassic time.
Figure 12. Evolution of Fundy rift basin. Left column shows timing of deposition and extrusion. Column on right shows timing of tectonic events and evidence for extension direction during rifting and displacement direction of hanging wall of composite border-fault system of Fundy rift basin during rifting and inversion. Note that the time-scale (adapted from Palmer and Geissman 1999; Olsen and Et-Touhami 2008) is not linear, and that the extent of the basin during rifting is larger than the present-day extent (see Figures 4 and 5). Displacement direction during basin inversion is from Baum et al. (2008).
Although the Cobequid/Chedabucto border–fault zone was active during rifting, field data show that not all faults within this zone were active at the same time. For example, a fault to the north of the Carrs Brook outcrops likely was active during deposition of the Wolfville Formation, leading to the widespread deposition of the Wolfville Formation at Carrs Brook and to the south (Figure 11c-iv). The main mapped fault at Carrs Brook became active later, folding the overlying Wolfville strata as it propagated upward (Figure 11c-v). The absence of the Wolfville Formation at Wasson Bluff suggests that a fault to the south of Wasson Bluff was active during and/or after the deposition of the Wolfville Formation. Later, faults north of and within the Wasson Bluff region became active, leading to the widespread deposition of the McCoy Brook Formation. Intra-rift angular unconformities (e.g. at Carrs Brook and Pinnacle Island) likely reflect local episodes of uplift and erosion associated with deformation within the Cobequid/Chedabucto border–fault zone. Olsen and Et-Touhami (2008) reported that the stratigraphic interval in which these unconformities are present is not exposed on the south shore of the Minas subbasin away from the Cobequid/Chedabucto border–fault zone. Thus, it is unclear whether these intra-rift unconformities are present beyond the Cobequid/Chedabucto border–fault zone. On the south shore of the Minas subbasin, the strata above and below this stratigraphic interval have similar dip magnitudes and directions, indicating that an unconformity, if present, would be subtle. A widespread, subtle, intra-rift angular unconformity is imaged on seismic lines through the Chignecto subbasin (Figure 4).

5.2. Extension direction

The seismic data show that all of the border–fault zones of the Fundy rift basin had a dip–slip (normal) component during rifting. The seismic data, however, cannot indicate whether these border–fault zones also had a strike-slip component during rifting. Consequently, the possible displacement direction of the hanging wall of the NE-striking Fundy border–fault zone ranges (minimally) from SSW to ENE (Figure 12); for the E-striking Cobequid/Chedabucto border–fault zone, the possible displacement direction ranges (minimally) from WSW to ESE. Growth strata indicate that activity on both border–fault zones was synchronous. Thus, the displacement direction for the hanging wall of the composite border–fault system of the Fundy rift basin lies within the overlapping ranges for the two fault zones, that is SSW to ESE (Figure 12). The seismic data cannot further constrain the range. The field data, however, provide some constraints. The NE-strike of the Shelburne dyke in southern Nova Scotia indicates that the regional extension direction was NW–SE during CAMP magmatic activity in the earliest Jurassic (Hodych and Hayatsu 1988). The NE-striking normal faults that cut the Blomidon cliffs in the Five Island region indicate that the extension direction was NW–SE during or after the eruption of the North Mountain Basalt in the earliest Jurassic. The sediment-filled fissures at Wasson Bluff indicate that the extension direction was NW–SE during the Early Jurassic deposition of the basal McCoy Brook Formation. Thus, the seismic and field data together indicate that the hanging wall of the linked border–fault system of the Fundy rift basin moved SE relative to the footwall during Early Jurassic time in response to NW–SE extension (Figure 12). It is possible, however, that conditions differed during the earlier phases of rifting. With a SE-displacement direction, the NE-striking border–fault zones of the Fundy and Chignecto subbasins had predominantly normal dip slip during rifting, whereas the E-striking border–fault zone of the Minas subbasin had oblique slip with left-lateral and normal components.

6. CONCLUSIONS

- The Fundy rift basin has undergone two phases of deformation: syn-depositional extension and post-depositional shortening (basin inversion).
- Syn-rift units imaged on seismic-reflection profiles thicken towards the composite border–fault system of the Fundy rift basin, which consists of the NE-striking Fundy and Chignecto border–fault zones and the E- to ENE-striking Cobequid/Chedabucto border–fault zone. Thus, syn-depositional faulting occurred from the Middle Triassic (and possibly Permian) to the Early Jurassic.
Field evidence of growth faulting includes talus-slope deposits that accumulated adjacent to fault scarps and marked changes in thickness among different fault blocks comprising the Cobequid/Chedabucto border–fault zone.

Intra-rift unconformities indicate that uplift and erosion occurred, at least locally, during rifting.

The hanging wall of the composite border–fault system of the Fundy rift basin moved down and towards the southeast during the later stages of rifting (i.e. Early Jurassic time). With this displacement direction, the NE-striking border–fault zones of the Fundy and Chignecto subbasins had predominantly normal dip slip during rifting, whereas the E-striking border–fault zone of the Minas subbasin had oblique slip with left-lateral and normal components. The movement direction during the early stages of rifting (i.e. Middle Triassic to Late Triassic time) is poorly constrained (ranging from SSW to ESE) by the available seismic and field data.

Sequential restorations show that the Fundy rift basin underwent 10–20 km of extension during rifting, most of which was accommodated by the border–fault system; the basin was significantly wider and deeper than it is today. Post-depositional shortening produced inversion-related structures near the border–fault zones of the Fundy rift basin. Post-rift deformation also tilted the eastern side of the basin to the northwest/north and produced regional uplift and erosion. Thus, the post-depositional shortening substantially modified the original geometry and dimensions of the Fundy rift basin.

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