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Time-Transgressive Salinic and Acadian Orogenesis, Magmatism and Old Red Sandstone Sedimentation in Newfoundland

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SUMMARY
We propose an intimate relationship between Silurian terrestrial red bed sedimentation (Old Red Sandstone), slab breakoff-related magmatism and deformation in the Newfoundland Appalachians. Red bed sedimentation started during the Early Silurian, and records the progressive rise of the Salinic mountains in the tectonic hinterland of the orogen. The red beds were mainly deposited in molasse-style foreland basins in front of an east-propagating terminal Salinic deformation front. New U–Pb zircon dating of volcanic rocks interlayered with the Silurian red beds in key structural locations yielded ages ranging between 425 and 418 Ma, which, combined with the existing geochronological database, suggests that the sedimentary rocks are progressively younger from west to east and overstep the accreted Gondwana-derived terranes. We propose that deposition of the red beds is a good proxy for the time of cratonization of the accreted terranes. Eastward migration of the Salinic deformation front was accompanied by eastward-widening of a slab-breakoff-related asthenospheric window. The latter is interpreted to have formed due to a combination of progressive steepening of the down-going plate following entrance of the leading edge of the Gander margin and its eduction. Gander margin eduction (reversed subduction) is proposed to have been instigated by the trench migration of the Acadian coastal arc built upon the trailing edge of the Gander margin, which developed contemporaneously with the Salinic collision. The resultant thinning of the lithosphere beneath the Salinic orogen, built upon the leading edge of the Gander margin immediately prior to the onset of the Early Devonian Acadian orogeny, set the stage for generation of the widespread bloom of Acadian magmatism.

SOMMAIRE
Nous proposons qu'il y a eu une relation intime entre la sédimentation des couches rouges continentales au Silurien (vieux-grès-rouges), un magmatisme lié à une rupture de segments de croûte, et la déformation appalachienne à Terre-Neuve. La sédimentation des couches rouges qui a débuté au début du Silurien témoigne du soulèvement progressif des monts saliniques de l'arrière-pays tectonique de l'orogène. Les couches rouges se sont déposées sous forme de molasses dans des bassins d'avant-pays, à l'avant du front de déformation salinique terminale qui se déployait vers l'est. De nouvelles datations U-Pb sur zircon de roches volcaniques interstratifiées avec des couches rouges siluriennes en des lieux structurels stratégiques montrent des âges qui varient entre 425 Ma et 418 Ma, ce qui, combiné aux bases de données géochronologiques existantes permet de penser que les roches sédimentaires sont progressivement plus jeunes d'ouest en est, et qu'elles surplombent les terranes accrétés du Gondwana. Nous suggérons que les couches rouges sont de bons indicateurs temporels de la cratonisation des terranes accrétés. La migration vers l'est du front de la déformation salinique a été accompagnée par un élargissement vers l'est d'une fenêtre asthénosphérique liée à une rupture de la croûte. Cette dernière aurait été provoquée par la combinaison de l'enfoncement progressif de la plaque qui a suivi l'entrée du bord d'attaque de la marge de Gander, et son éduction. Nous proposons que l'éduction (l'inverse de la subduction) de la marge de Gander a été provoquée par la migration de la fosse tectonique côtière acadienne, induite par la migration du bord d'attaque de la marge de Gander, contemporaine de la collision salinique.

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L'amincissement de la lithosphère sous l'orogène salinique qui en a résulté, et qui s'est déployé au bord d'attaque de la marge de Gander juste avant l'enclenchement de l'orogénie acadienne au début du Dévonien, a préparé le terrain du déploiement à grande échelle du magmatisme acadien.

INTRODUCTION

Beginning with his benchmark 1967 publication on the Silurian rocks of Newfoundland (Williams 1967) the Silurian Period was of special interest to Harold (Hank) Williams. His highly insightful stratigraphic and structural studies of the superbly exposed Silurian rocks in the Notre Dame Bay of north-central Newfoundland (Williams 1972, 1993a), led him to conclude that the Exploits subzone must have been internally divided into two parts separated by a significantly wide Middle Ordovician to Silurian marine seaway that is now sutured along the Dog Bay Line (Williams et al. 1993). Williams considered the closure of this vestige of the Iapetus Ocean to be the tectonic cause of the Silurian deformation, although the responsible tectonic processes (i.e. subduction vs. strike-slip and transpression) were never specified. Williams' work was contrary to the established pre-1985 dogma, which divided the pre-Alleghanian deformation in the northern Appalachians into only two orogenic phases: the Ordovician Taconic and Devonian Acadian orogenies (e.g. Rodgers 1970). The intervening Silurian period was thought to be characterized by tectonic quiescence (e.g. Colman-Sadd 1982; Bradley 1983). Concurrent with Williams' work, evidence for latest Ordovician to Silurian, post-Taconic to pre-Acadian orogenesis was progressively growing in the Northern Appalachians (e.g. van Staal 1987, 1994; van der Pluijm and van Staal 1988; Dunning et al. 1990; van Staal et al. 1990; West et al. 1992; Cawood et al. 1994; Hibbard 1994; van Staal and de Roo 1995). Following the discovery by Boucot (1962) of stratigraphic evidence for Silurian tectonism (Salinic disturbance) in central Maine, the latest Ordovician to Silurian tectonism was referred to as the Salinic orogeny by Dunning et al. (1990) and Malo and Bourque (1993).

When the Salinic exactly ended and the Acadian began at any given location in the Appalachian orogen is a contentious issue and depends on whether one considers the definition of orogeny from a purely absolute age perspective or from a tectonic-dynamic point of view (see discussion in van Staal et al. 2008 and below). Defining the end of the Salinic orogeny and the start of the Acadian orogeny from an absolute age perspective has been problematic, at least in part due to the changes in the mean absolute ages of the series and stage boundaries of the Silurian and Devonian (e.g. 408 ± 12 Ma [Palmer 1983] vs. 419.2 ± 3.2 Ma [ICS 2013]). This led to confusion and grouping of unrelated events (e.g. Dunning et al. 1990; Lin et al. 1994; O'Brien 2003; van Staal and Zagorevski 2012). In addition, deformation associated with both Acadian (e.g. Bradley et al. 2000) and Salinic orogenies (see below) was diachronous and consequently appears to overlap in time and/or be continuous, which led some workers to group them together (e.g. O'Brien 2003; Murphy and Keppie 2005). However, overprinting relationships and tight age constraints on the structures generated by the Salinic and Acadian orogenies consistently show that the Acadian deformation postdated Salinic structures in any given area where both are present (van Staal and de Roo 1995; van Staal et al. 2009; Wilson and Kamo 2012). This suggests that phases of Salinic and Acadian orogenic events may have overlapped in time but never in space at any time (van Staal et al. 2009). In addition, a number of different lines of evidence indicate that they constitute two kinematically and dynamically distinct events that followed each other closely in time (e.g. Dunning et al. 1990; West et al. 1992; van der Pluijm et al. 1993; Hibbard 1994; Holdsworth 1994; Cawood et al. 1994; Burgess et al. 1995; Kerr et al. 1995; D'Lemos et al. 1997; Schofield and D'Lemos 2000; O'Brien 2003; van Staal et al. 2003; McNicoll et al. 2006; Zagorevski et al. 2007; van Staal et al. 2009; Wilson and Kamo 2012). As such we espouse a tectonic-based definition of orogeny or mountain building in this manuscript. With this we mean a series of kinematically and dynamically closely related tectonic processes that led to a distinct phase of mountain building.

Understanding of the tectonic significance and absolute age of the Salinic and Acadian orogenies is particularly pertinent to the origin and affiliation of Wenlockian to earliest Devonian bimodal magmatic suites (Fig. 1). The magmatic axes of these suites become younger towards the east-southeast and they are commonly spatially associated with deposition of syn-tectonic terrestrial sedimentary rocks, such as the Botwood Group in central Newfoundland (Fig. 1). The affiliation of these magmatic suites to either Salinic or Acadian orogenesis is not always clear and sometimes problematic. The standard model for late syn- to post-tectonic bimodal magmatism invokes orogen-wide extension (Turner et al. 1992); however, Salinic and Acadian bimodal magmatism was synchronous with regional shortening and uplift throughout the central orogenic core of the Newfoundland Appalachians (e.g. Dunning et al. 1990). In this contribution, we utilize new and existing geochronological and geochemical databases to integrate the magmatic and tectonic evolution of the Newfoundland Appalachians and explore the possible tectonic settings and causes of this Late Silurian magmatism, terrestrial sedimentation and associated deformation.

MIDDLE ORDOVICIAN–LATE SILURIAN GEOLOGICAL ARCHITECTURE OF THE CENTRAL NEWFOUNDLAND APPALACHIANS

The central Newfoundland Appalachians (Fig. 1) are underlain by Paleozoic and older rocks of the Dunnage and Gander tectonostratigraphic zones (Williams 1978). Tectonostratigraphic zones and subzones were principally defined on their pre-Silurian geological characteristics, although it was later shown that geological contrasts locally continued into the Silurian (Williams et al. 1988, 1993; Reusch and van Staal 2012). The Dunnage zone mainly comprises Cambro-Ordovician arc terranes and associated slivers of supra-subduction zone ophiolites, which were divided into the Notre Dame subzone (peri-Laurentian realm) and the Exploits subzone (peri-Gondwanan realm), separated by the Red Indian Line (Williams et al. 1988; Williams 1993b;

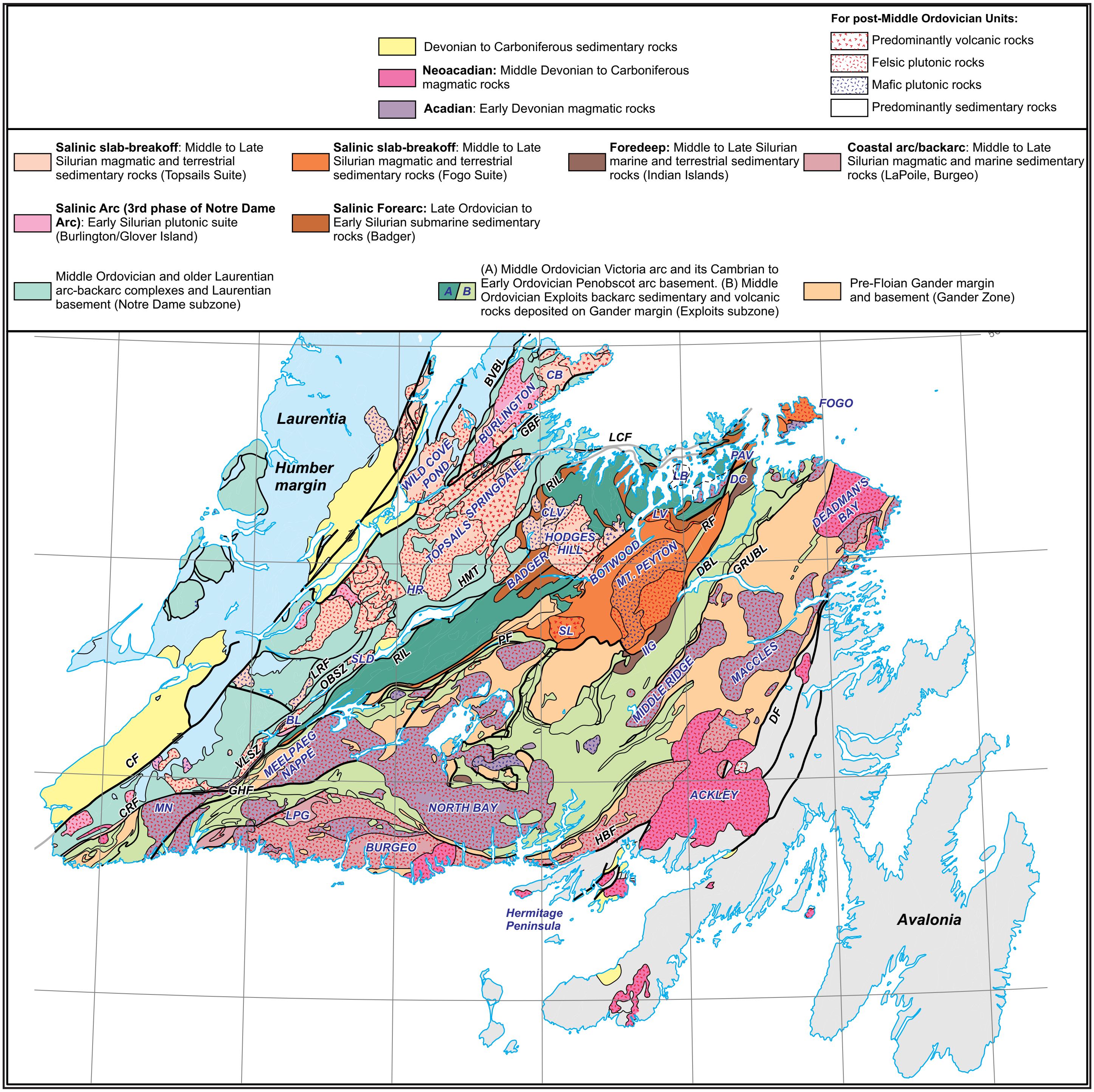

Figure 1. Geology of Newfoundland (modified from Hibbard et al. 2006). BL: Boogie Lake pluton; BVBL: Baie Verte-Brompton Line; CB: Cape Brule porphyry; CF: Cabot Fault; CLV: Charles Lake volcanic unit; CRF: Cape Ray Fault; DBL: Dog Bay Line; DC: Duder Complex; DF: Dover Fault; GBF: Green Bay Fault; GHF: Gunflap Hills Fault; GRUBL: Gander River Ultrabasic Belt Line; HBF: Hermitage Bay Fault; HMT: Hungry Mountain Thrust; HR: Harry's River; IIG: Indian Islands Group; LB: Loon Bay pluton; LCF: Lobster Cove Fault; LPG: La Poile Group; LRF: Lloyds River Fault; LV: Lawrenceton volcanic unit; MN: Meelpaeg Nappe; OBSZ: Otter Brook shear zone; PAV: Port Albert volcanic unit; PF: Pine Falls Formation; RF: Reach Fault; SL: Stony Lake volcanic unit; RIL: Red Indian Line; SLD: Star Lake Dam; VLSZ: Victoria Lake Shear Zone.

Hibbard et al. 2006, 2007). The Red Indian Line (Fig. 1) represents the fundamental Iapetus suture zone along which these two unrelated realms were juxtaposed during the Sandbian (ca. 455 Ma; van Staal et al. 1998; Zagorevski et al. 2008).

The paleogeographic and tectonic evolution of the Exploits subzone and adjacent Gander zone (Fig. 1) are intimately related because most of the pre-Middle Ordovician rocks of

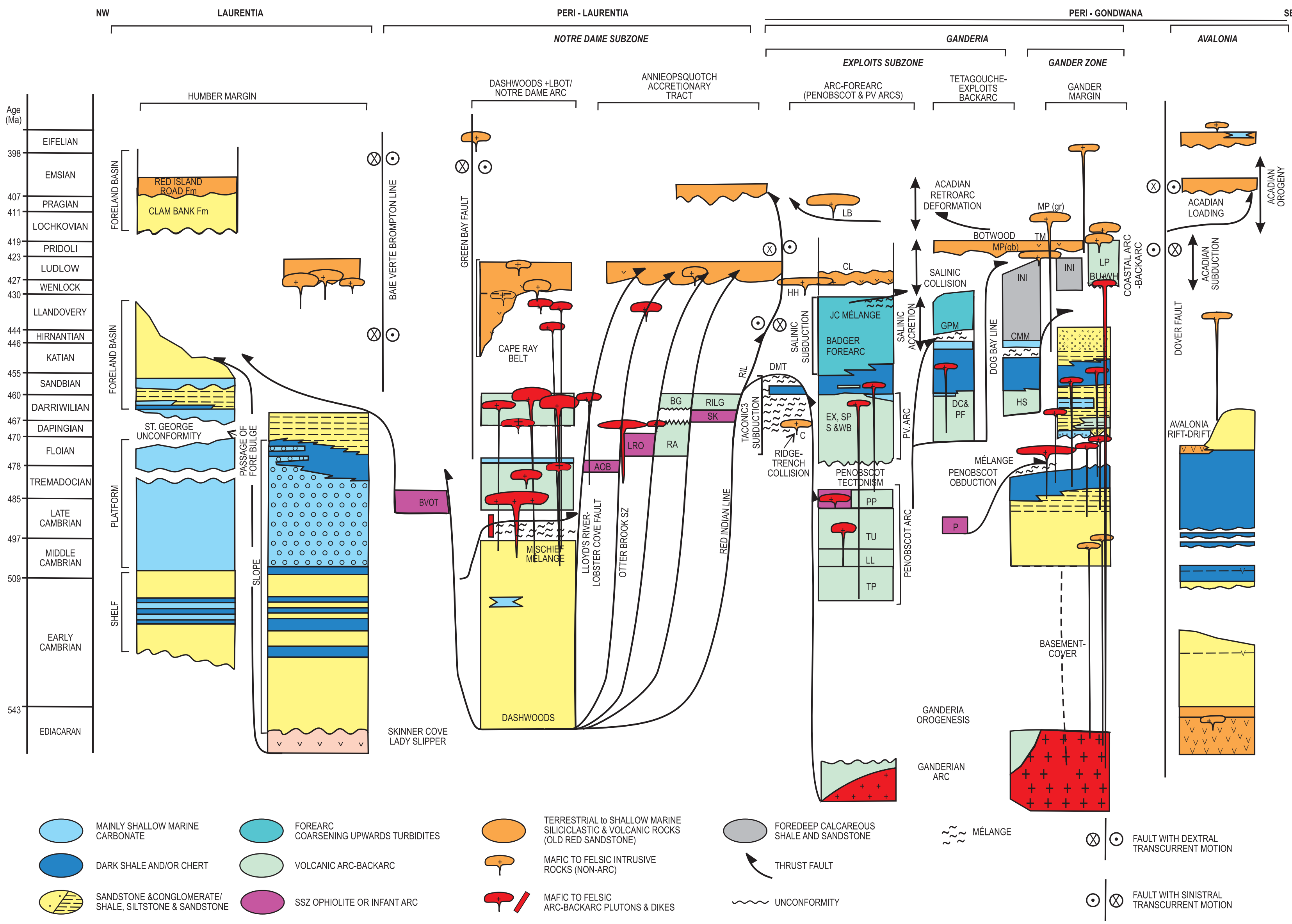

Figure 2. Tectono-stratigraphic columns through the various belts of central Newfoundland discussed in this paper. AOB: Annieopsquotch ophiolite belt; BG: Buchans Group; BVOT: Baie Verte oceanic tract; BU: Burgeo Intrusive Suite; C: Coaker Porphyry; CL: Charles Lake volcanic unit; CMM: Carmanville mélange; DC: Duder Complex; DMT: Dunnage Mélange tract; EX: Exploits Group; GPM: Garden Point mélange; HH: Hodges Hill complex; HS: Hamilton Sound Group; INI: Indian Islands Group; JC: Joey's Cove; LB: Loon Bay pluton; LBOT – Lushs Bight oceanic tract, LL: Long Lake Group; LP: La Poile Group; LRO: Lloyds River ophiolite complex; MPgb: gabbro phase of Mount Peyton pluton; MPgr: late granite phase of Mount Peyton pluton; P: Pipestone Pond Complex; PF: Pine Falls Formation; PP: Pats Pond Group; PV: Popelogan-Victoria; RA: Robert's Arm Group; RIL: Red Indian Line; RILG: Red Indian Lake group; S: Summerford Group; SK: Skidder Formation; SP: Sutherlands Pond group; SSZ – suprasubduction zone; TM: Ten Mile Formation; TP: Tally Pond Group; TU: Tulks Group; WB: Wild Bight Group; WH: Western Head granite.

the Exploits subzone appear to be built on the Gander zone basement (Fig. 2; van Staal et al. 1998; Rogers et al. 2006; Zagorevski et al. 2010, 2012). The latter rocks together with the rocks underlying the Gander zone constitute Ganderia (van Staal et al. 1998, 2012). The Gander zone mainly comprises a Middle Cambrian to Late Ordovician passive margin (Gander margin) deposited on Neoproterozoic–Early Cambrian arc basement (Fig. 2; Dunning and O'Brien 1989; van Staal 1994; Fyffe et al. 2009; van Staal and Barr 2012). The late Early Cambrian–Early Ordovician part of the Gander margin represented the passive margin of the Penobscot backarc basin, which was bordered on its active side by the 515–485 Ma Penobscot arc (Exploits subzone) (Figs. 1, 2). Passive Gander margin sedimentation was interrupted for a brief period (≤ 5 m.y.) between 483 and 478 Ma as a result of inversion of the backarc basin and obduction of the Penobscot backarc ophiolites (Zagorevski et al. 2010). Passive margin sedimentation was re-established again by ca. 475 Ma (e.g. Colman-Sadd et al. 1992); the upper part of the Gander margin tectonically associated with the younger (475–458 Ma) Popelogan-Victoria arc, which also rifted, opening the Tetagouche-Exploits backarc basin (TEB) (van Staal 1994; van Staal et al. 1998). Contrasts in the Ordovician–Silurian geology led to an additional subdivision of the Exploits subzone by Williams et al. (1993), and introduction of the Dog Bay Line (Fig. 1). The western part of the Exploits subzone, situated between the Red Indian and Dog Bay lines (Fig. 1), is mainly underlain by volcanic and associated sedimentary rocks of the Cambrian–Middle Ordovician Penobscot and Popelogan-Victoria arcs (Zagorevski et al. 2010, 2012), Upper Ordovician to Upper Silurian clastic sedimentary rocks of the marine Badger and terrestrial Botwood groups (Figs. 1, 2; Williams et al. 1993). The marine Badger Group basin was correlated with rocks of the Salinic Matapedia basin in northern New Brunswick and mainly interpreted as a southeast-facing Salinic forearc basin that overstepped onto accreted backarc basin blocks (van Staal 1994; van Staal et al. 1998, 2008, 2009; Wilson et al. 2004). This correlation was recently reinforced by the discovery of Mediterranean Faunal Province conodonts in correlatives of the Badger Group deposited above the Middle to Late Ordovician Duder Complex (Figs. 1, 2), which have not been identified elsewhere in North America, except in coeval rocks of the Matapedia basin (Dickson et al. 2007). The Duder Complex (Williams et al. 1993) is fault-bounded and has been interpreted as an allochthonous slice that originated in the Tetagouche-Exploits backarc basin (Zagorevski et al. 2010). It was accreted to the Badger forearc system during the Late Ordovician–Early Silurian (Fig. 2). Accretion was manifested by formation of the Garden Point mélange (Williams et al. 1993) and deposition of Badger Group clastic rocks above it (Figs. 1, 2). The mélange contains rare limestone blocks (Williams et al. 1993) that we correlate with Katian–Hirnantian limestone that overlies the Duder Complex and was included into the Badger Group by Dickson et al. (2007).

The eastern part of the Exploits subzone, situated east or south of the Dog Bay Line, comprises Middle Ordovician to Lower Silurian clastic sedimentary rocks (e.g. Davidsville, Baie d'Espoir and Red Cross groups) with minor volcanic rocks (mainly felsic) and calcareous sedimentary rocks near the base of the succession and coarse clastic rocks near the top (O'Neill 1991; Colman-Sadd et al. 1992; Williams et al. 1993; Valverde-Vaquero et al. 2006). This sequence disconformably overlies Cambrian–Tremadocian sandstone and shale of the Gander Group; the late Tremadocian hiatus marks obduction of the Penobscot ophiolites onto the Gander margin (Fig. 2; Williams and Piasecki 1990; Colman-Sadd et al. 1992; Zagorevski et al. 2010). This part of the Exploits subzone mainly preserves a continuation of sedimentation on the Cambrian–Ordovician passive Gander margin following opening of the Tetagouche-Exploits backarc basin shortly after 475 Ma (van Staal 1994). The Middle to Late Ordovician Gander margin rocks in turn are overlain by Silurian marine calcareous sedimentary rocks (Donovan et al. 1997; Boyce and Dickson 2006; McNicoll et al. 2006) of the Indian Island Group (Figs. 1, 2). Parts of the Indian Island Group have been correlated with lithologically similar rocks in the Fredericton trough of central New Brunswick, particularly the Flume Ridge Formation (van Staal et al. 2009). The Fredericton trough was interpreted as a Salinic-related marine foredeep formed on the down-going Gander margin while it was being overridden by the Brunswick subduction complex (van Staal 1994; van Staal et al. 1998, 2008, 2009). The overall depositional history of the Indian Island Group, however, is poorly understood at present, mainly because of its poor preservation in Newfoundland (Fig. 1).

The Middle Ordovician rocks of the Popelogan-Victoria arc and the Gander margin are structurally separated by a narrow belt of highly tectonized, polyphase deformed volcanic and sedimentary rocks, including the 469–453 Ma Duder Complex (Currie 1997; Dickson et al. 2007; Zagorevski et al. 2010) and Hamilton Sound Group (Johnston et al. 1994) in northeastern Newfoundland (Figs. 1, 2), and the intensely deformed MORB-like rocks of the Pine Falls Formation (Valverde-Vaquero et al. 2006) in central Newfoundland (Fig. 1). This relatively narrow belt of highly tectonized rocks situated between the Reach fault and the Dog Bay Line (Fig. 1) as defined by Williams et al. (1993) has been interpreted as remnants of the Middle Ordovician Tetagouche-Exploits backarc basin (TEB), which were progressively accreted to the Badger forearc. The tectonic setting, style of deformation and metamorphism of these rocks suggest that they are correlatives of the Brunswick subduction complex (van Staal et al. 2008). The TEB corresponds to the oceanic seaway delineated by Williams et al. (1993), which separated the Popelogan-Victoria arc and the Middle Ordovician rocks of the Gander margin (van Staal et al. 1998; Zagorevski et al. 2010). The Dog Bay Line thus represents the terminal suture along which the TEB closed (cf. Dickson 2006; Dickson et al. 2007).

EXISTING CONSTRAINTS ON THE DURATION AND DISTRIBUTION OF SALINIC COLLISION-RELATED DEFORMATION AND METAMORPHISM IN NEWFOUNDLAND

Robust age constraints on the duration of Salinic convergence and collision-related deformation are relatively sparse in Newfoundland. This is unfortunate because there are indications that collision-related deformation is diachronous from west to east (see below). The collision of the first peri-Gondwanan elements with the composite Laurentian margin marks the end of the strictly Laurentian realm Taconic orogenesis (Fig. 2) and the start of the Salinic orogenic cycle, which involves peri-Gondwanan elements as well. Thus, in central Newfoundland, the absolute age constraint on the start of the Salinic cycle is ca. 455 Ma (van Staal et al. 2007, 2009; Zagorevski et al. 2008).

West of the Red Indian Line

The shutdown of Salinic arc magmatism and initiation of non-arc magmatism at ca. 433 Ma (Whalen et al. 2006) provides the best minimum age for the onset of Salinic collision between the composite Laurentian margin and the Gander margin in Newfoundland. High-grade metamorphic rocks of the Ordovician Notre Dame arc east of the Baie Verte Line, but west of the Lloyds River-Hungry Mountain-Lobster Cove thrust system (Figs. 1, 2) have yielded monazite, titanite and rutile ages ranging between 433 and 430 Ma (Lissenberg et al. 2005; van Staal et al. 2007) supporting rapid Salinic-related exhumation. Salinic deformation and high-grade metamorphism in the Laurentian hinterland immediately west of the Baie Verte Line (Fig. 1) was already occurring by ca. 434 Ma and was followed by rapid cooling of high-grade metamorphic rocks between 430 and 425 Ma (Cawood et al. 1994; Lin et al. 2013).

Salinic deformation reactivated or overprinted Ordovician Taconic structures in the Annieopsquotch accretionary tract (Fig. 2), between the Lloyds River-Hungry Mountain-Lobster Cove thrust system and the Red Indian Line. Salinic slab break-off-related plutonic rocks show both cross-cutting (Fig. 3A) and syn-tectonic relationships (Fig. 3B, C) along reactivated Ordovician thrust faults. Syn-tectonic intrusive rocks were emplaced into the Otter Brook Shear Zone and Red Indian Line at 427 and 432 Ma respectively (Fig. 3C, D; Zagorevski et al. 2007). More importantly, Wenlockian red beds and interlayered felsic volcanic rocks (ca. 429 Ma: Chandler et al. 1987; Dunning et al. 1990) constrain the age of thrusting because they are locally preserved in the hanging wall (Fig. 3E) and footwall of southeast-directed sinistral-oblique thrust faults (Fig. 3F, G), yet contain material of the eroded hanging wall (Zagorevski et al. 2007). This foreland fold and thrust belt style relationship between magmatism, deformation and molasse-style sedimentation (for definition see Mitchell and Reading 1978) was also established in the correlative rocks along strike in north-central Newfoundland (Fig. 3H; Dean and Strong 1977; van Staal et al. 2009), and thus represents a regional tectonic style. It emphasizes the relationship between red terrestrial molasse-like sedimentation (Old Red Sandstone facies of the British Caledonides, Friend et al. 2000) and Salinic deformation.

East of the Red Indian Line (Badger Belt)

Evidence for high-level Llandoverian–Wenlockian Salinic folding and thrusting is well preserved in Upper Ordovician–Lower Silurian turbiditic rocks of the Badger Group (Fig. 1), immediately southeast of the Red Indian Line (O'Brien 2003). Thrusts breaching the sea floor overrode their own olistostromes and have been dated by fossils in their matrix as Early Silurian (Fig. 4A; Jacobi and Schweikert 1976; Arnott 1983; van der Pluijm 1986; Reusch 1987). Elsewhere this deformation produced inclined or recumbent F_1 folds and associated bedding-parallel S_1 cleavage in the Badger Group and underlying older rocks (e.g. Elliott et al. 1991). Structural transport direction varied, but generally was to the southeast or south taking into account the reorientations caused by younger deformation events (van der Pluijm 1986; Lafrance and Williams 1992; Kusky and Kidd 1996; O'Brien 2003). This D_1 phase of the Salinic deformation predated eruption and intrusion of the main phases of the Hodges Hill-Charles Lake magmatic suite (Figs. 1, 2), which yielded ages between 435 and 429 Ma (Kusky and Kidd 1996; Dickson 2000; O'Brien 2003, pers. comm.), and overlaps with the age constraints established by Zagorevski et al. (2007) nearby, along and immediately to the west of the Red Indian Line. On the western margin of the Hodges Hill Batholith, Badger Group sedimentary rocks are deformed and contact metamorphosed to cordierite schist and higher metamorphic grade (Fig. 4B). In contrast, the Twin Lakes phase of the Hodges Hill Batholith and the Dawes Pond Granodiorite do not display penetrative fabric development (Fig. 4C, D).

East of the Dog Bay Line (Gander Margin)

Evidence for Salinic deformation in rocks of the Gander margin mainly comes from syn-tectonic intrusive, migmatitic and metamorphic rocks. Metamorphism accompanying the high-strain D_3 deformation in this part of Newfoundland was dated at ca. 417 Ma both in the northeastern (Holdsworth 1994; D'Lemos et al. 1997; Schofield and D'Lemos 2000) and southwestern parts of the Gander margin (Valverde-Vaquero et al. 2000) over a strike length of more than 300 km, indicating that the preceding deformation, low to high grade metamorphism and migmatization mainly formed during Salinic orogenesis. Migmatitic rocks associated with D_{1-2} were dated with U–Pb zircon, monazite and titanite methods between 425 and 423 Ma (Dunning et al. 1990; Schofield and D'Lemos 2000), whereas Ar-dating of metamorphic muscovite yielded ages as old as 429 Ma (O'Neill and Colman-Sadd 1993). Structural data and pattern of metamorphism suggest that the D_{1-2} structures had a vergence to the southeast or south (O'Neill 1991; Piasecki 1992). The Salinic deformation in the Gander margin thus overlaps with the age constraints further to the west, but lasts at least until 423 Ma and possibly until 417 Ma if one considers the similar, sinistral shear kinematics associated with D_3 structures (Holdsworth 1994). This suggests that the final increments

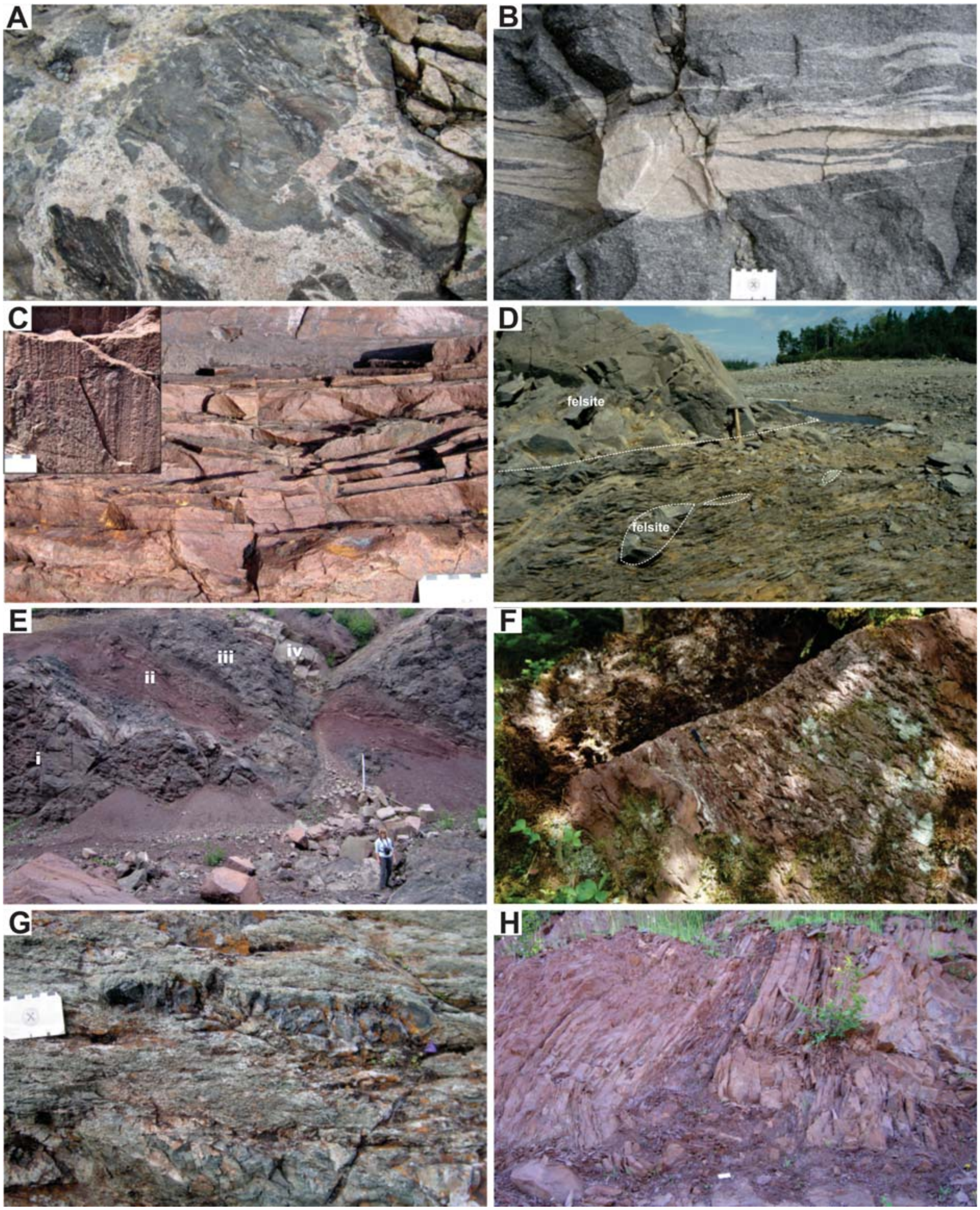

Figure 3. Representative photographs of Silurian rocks along and west of the Red Indian Line. A. Boogie Lake granodiorite ($429^{+7}/_{-3}$ Ma: Dunning et al. 1990) contains xenoliths of Ordovician mylonite of the Otter Brook shear zone. B. Syn-tectonic sheets of Silurian granite (ca. 426 Ma U/Pb titanite: Zagorevski et al. 2007) in the Otter Brook shear zone, Star Lake dam. C. Strongly foliated and lineated (inset) Silurian Topsails Igneous Suite syenite in the Hungry Mountain Thrust, Buchans area. D. Syn-tectonic ca. 432 Ma felsite intruded into mélange of the Victoria River delta fault (Thurlow et al. 1992; Rogers et al. 2005; Zagorevski et al. 2007), a subsidiary thrust fault of the Red Indian Line. E. Tilted and normally faulted sequence of Silurian (Chandler et al. 1987) terrestrial columnar jointed andesite (i), siltstone and bentonite (ii), conglomerate (iii) and columnar jointed rhyolite ignimbrite (iv) at Hind's Lake dam. These rocks unconformably overlie Middle Ordovician mid-crustal plutons of the Notre Dame arc (see also Whalen et al. 1987). F. Cleaved Silurian red bed sandstone and conglomerate in the footwall syncline to the Otter Pond shear zone. G. Foliated Silurian peralkaline rhyolite tuff in the footwall of the Hungry Mountain Thrust, Buchans area. H. Folded red bed sandstone in a footwall syncline of the Lobster Cove Fault, Pilley's Island.

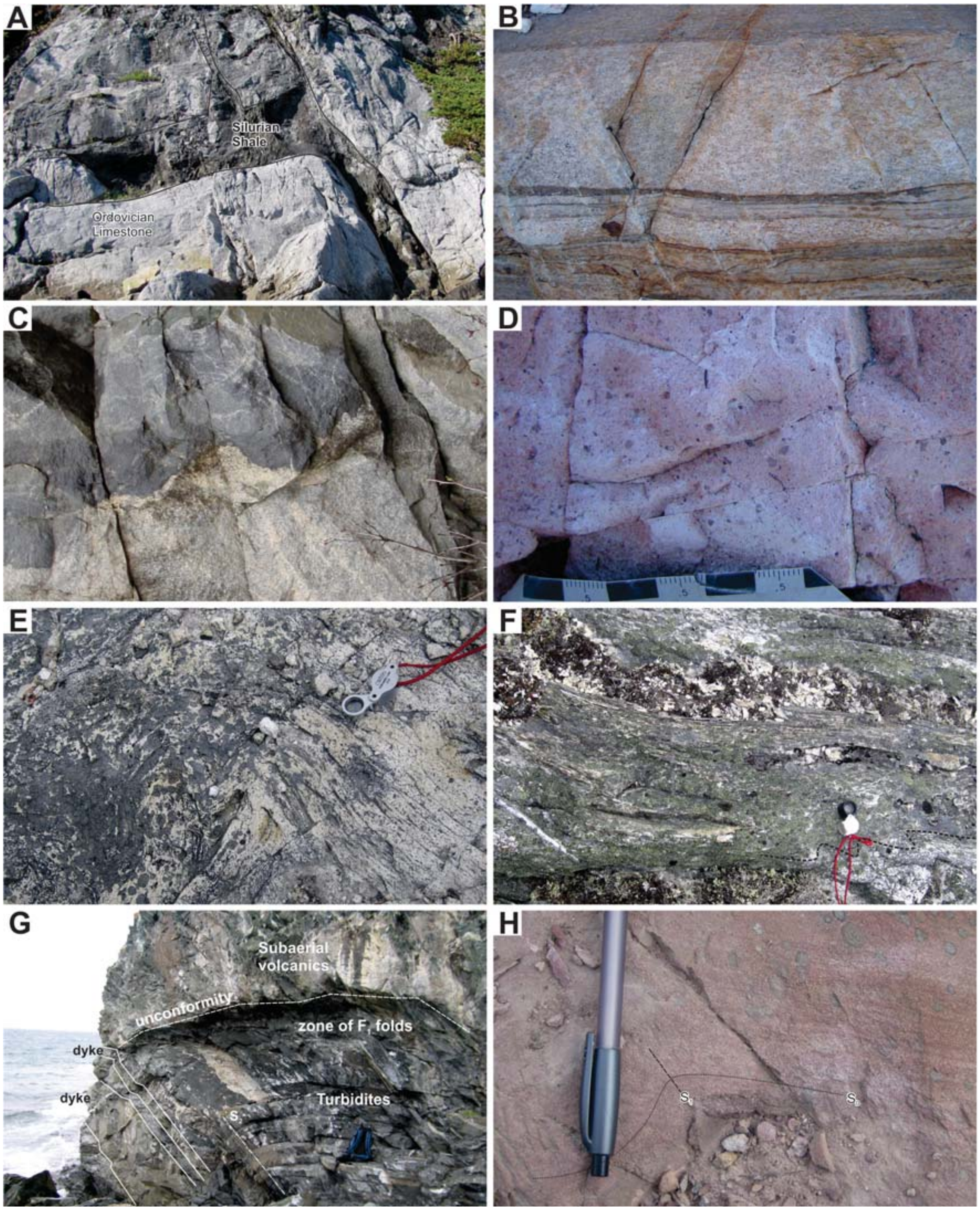

Figure 4. Representative photographs of Silurian rocks along and east of the Red Indian Line. A.Darriwilian limestone blocks in a matrix of Silurian shale of the Joey's Cove mélange at Rogers Cove, New World Island (see Reusch 1987) B. Contact metamorphosed Badger Group turbidites adjacent to the Dawes Pond Granodiorite and Hodges Hill Batholith. Note bedding subparallel to foliation. Nearby outcrops of the Roberts Arm Group preserved evidence for polyphase deformation. C. Twin Lakes phase of the Hodges Hill Batholith displaying scalloped contact between mafic and felsic phases, suggesting they are co-magmatic. D. Pristine hypabyssal quartz-phyric rhyolite of the Dawes Pond Granodiorite. E. Foliated Stony Lake rhyolite deformed into symmetrical folds in hinge of anticline. Folds are accompanied by weak axial planar cleavage. Hand lens for scale. F. Strongly foliated ca. 423 Ma Stony Lake rhyolite. Foliation is tectonic and folded by younger mesoscopic folds containing a weak axial planar cleavage (lower right corner). Hand lens for scale. G. Unconformity between Salinic (F_1) deformed Badger Group turbidites and subaerial volcanic rocks of the Port Albert Formation of the Botwood Group. Turbidites below the unconformity contain a zone of F_1 folds (Karlstrom et al. 1982) and are overlain by a thin conglomerate layer that locally contains pebbles with pre-entrainment cleavage. Both units are folded by NW-verging Acadian(?) structures (F_3 of Lafrance and Williams 1992). H. Folded and cleaved red beds of Wigwam Formation near the contact with the overlying Stony Lake Formation felsic volcanic rocks.

of Salinic collision-related deformation and metamorphism are younger in the east, whereas the oldest ages are preserved in the west. If correct, there is a west to east diachroneity in the timing of Salinic deformation.

SILURIAN MAGMATISM AND SEDIMENTATION RELATED TO SALINIC OROGENESIS

Arc magmatism associated with Salinic subduction of the TEB lasted from 446 to 435 Ma in Newfoundland (Whalen et al. 2006; Brem et al. 2007). Field, petrological and isotopic evidence suggest that the younger Silurian non-arc magmatism (433– 424 Ma) was related to break-off of the oceanic slab attached to the down-going Gander margin (Whalen et al. 2006). This oceanic slab, parts of which are preserved in the Brunswick subduction complex, New Brunswick, and the Red Cross Group, Newfoundland, was similar to the lithosphere in the Sea of Japan, characterized by discontinuous spreading centres, felsic volcanic centres and fragmented arc basement (e.g. Zagorevski et al. 2010 and references therein). The slab break-off of this weak backarc crust occurred during deformation associated with the Salinic collision (e.g. Dunning et al. 1990; Zagorevski et al. 2007). The bulk of the slab break-off-related igneous rocks (Topsails Igneous Suite: Whalen et al. 1987, 2006) define a relatively narrow (75–100 km wide) belt, which is mainly situated between the Baie Verte and Red Indian lines (Fig 1). This suggests that the mantle window created by the slab breakoff was also relatively narrow.

A younger, Ludlovian–early Lochkovian syn-to post-collisional suite of magmatic rocks (424–419 Ma) occurs mainly between the Red Indian and the Dog Bay lines (Fig. 1; Williams et al. 1993; van Staal et al. 2009). This magmatic suite straddles the as yet poorly delineated Dog Bay Line (Dickson et al. 2007), which forms part of the Salinic suture zone along which the active and passive sides of the TEB were juxtaposed following subduction of the intervening backarc lithosphere. In contrast to the slightly older syn-collisional magmatic suite, the Late Silurian–Early Devonian magmatic rocks at least in part post-date Salinic deformation, but consistently predate Acadian deformation both in Newfoundland and Maritime Canada (van Staal et al. 2009). Evidence presented below indicates this magmatism overlapped with the last gasp of Salinic deformation. A similar tectonic scenario was recognized in New Brunswick and Maine (van Staal et al. 2003, 2009; Wilson et al. 2008) where the Bamford Brook-Liberty Line is a correlative of the Dog Bay Line suture zone (Ludman et al. 1993; van Staal et al. 2009; Reusch and van Staal 2012).

In Newfoundland, igneous rocks belonging to the Late Silurian–Early Devonian suite of magmatic rocks (Fig. 1) comprise the ca. 424–422 Ma phases of the Mt. Peyton and Fogo Island batholiths (Dunning 1992, 1994; Aydin 1995; Sandeman and Malpas 1995; Currie 2003) and coeval volcanic eruptive and associated epizonal intrusive dikes and sills such as the Stony Lake volcanic complex (Fig. 4E, F; ca. 423 Ma: Dunning et al. 1990) and the Port Albert composite dikes (ca. 422 Ma: Elliott et al. 1991). This Late Silurian–Early Devonian magmatic suite is herein informally referred to as the Fogo suite. Detailed petrological investigations of the Fogo suite indicate that it comprises both mantle and crust-derived melts (Strong and Dupuy 1982; Sandeman and Malpas 1995; Currie 2003; Kerr 2013). Such a petrogenesis is consistent with formation of the Fogo suite above an asthenospheric window created by slab break-off (Whalen et al. 2006; van Staal et al. 2009).

The Fogo suite has a close spatial association with the Botwood belt (Williams et al. 1993, 1995), which is largely underlain by the terrestrial Botwood Group (Williams 1972). The Botwood Group is a critical unit to constrain the duration of Salinic collision-related deformation, because it disconformably overlies the Early Salinic, syn-tectonic marine Badger Group sedimentary rocks (Fig. 4G; van der Pluijm 1986; Reusch 1987; van der Pluijm et al. 1993; Zagorevski et al. 2008; van Staal et al. 2009; Waldron et al. 2012). In the type locality near the towns of Botwood and Laurenceton, the Botwood Group comprises the Lawrenceton[1] and the Wigwam formations. The Lawrenceton Formation is dominated by mafic and felsic volcanic rocks at its base, whereas the overlying Wigwam Formation comprises red sandstone (Fig. 4H) and minor felsic volcanic rocks. Other formation names have been used elsewhere (Williams et al. 1993; Pollock et al. 2007). Members of the Fogo suite either intruded (Elliott et al. 1991) or were inferred to unconformably overlie folded and cleaved sandstone of the presumed Middle to Upper Silurian Wigwam Formation of the Botwood Group (Fig. 4H; Anderson and Williams 1970; Colman-Sadd and Russell 1982) in the southwestern part of the Botwood belt (Fig. 1). The Fogo suite thus represents a potential marker separating the Salinic- and Acadian-related deformation. Additional evidence presented herein suggests that the slab break-off-related magmatic front was diachronous in Newfoundland and migrated from west (Topsails Suite) to east (Fogo suite). This relationship suggests that the asthenospheric window beneath the Salinic orogen was progressively widening before the onset of the Acadian orogeny.

TECTONIC SETTING AND START OF THE ACADIAN OROGENY

Early Devonian Acadian orogenesis affected nearly the entire width of the northern Appalachian orogen, including Avalonia (Fig. 1), with the exception of Meguma (Williams 1993b; van Staal and Barr 2012). Middle–Late Devonian orogenesis in Meguma has been named Neoacadian (van Staal 2005; van Staal and Barr 2012). Acadian-related deformation and metamorphism were most complex and penetrative in Ganderia, but generally less strong or non-penetrative in Avalonia and west of the Red Indian Line. The presence of Early Devonian Acadian deformation in Avalonia but not

[1] Williams (1972) defined this formation as ‘Lawrenceton’ rather than ‘Laurenceton’ in the type area of the Botwood belt.

Meguma suggests a first order relationship between orogenesis and accretion of Avalonia to the composite Laurentian margin (van Staal 2005).

During the Llandoverian, Ganderia was accreted to and became the leading edge of Laurentia (see previous section). The trailing edge (i.e. southeastern in present coordinates) of Ganderia (van Staal 2005) is characterized by development of a short-lived Silurian southeast-facing arc-backarc system (Figs. 1, 2), referred to as the Coastal arc-Mascarene backarc in southern New Brunswick, Nova Scotia (e.g. Barr et al. 2002; Lin et al. 2007) and Maine (Bradley 1983; Robinson et al. 1998). The Coastal arc-backarc system continues into southern Newfoundland (Burgeo pluton–La Poile basin: Figs. 1, 2; O'Brien et al. 1991; Kerr et al. 1995; Lin et al. 2007), but is truncated further to the east by the subvertical Dover-Hermitage Bay fault, which represents the on-land boundary between Avalonia and Ganderia in Newfoundland (Fig. 1, Williams 1993b). The polarity of the arc is inferred from: 1) the inboard position of the Mascarene backarc relative to the Coastal arc suggesting that Ganderia was facing an open Acadian seaway to the southeast; 2) preservation of Early Devonian high pressure–low temperature forearc rocks immediately to the southeast of the Coastal arc in New Brunswick (White et al. 2006); 3) lack or scarcity of Silurian–Early Devonian magmatic rocks in Avalonia; 4) seismic reflectors in Avalonia dip towards the northwest (van der Velden et al. 2004); and 5) sedimentary evidence for Early Devonian tectonic loading of Avalonia (Waldron et al. 1996). These lines of evidence suggest that the Acadian seaway closed by north-westwards dipping subduction during the Silurian–Early Devonian (van Staal et al. 2009, 2012; van Staal and Barr 2012).

The Mascarene and La Poile backarc and/or intra-arc basins in southern New Brunswick and Maine, and southern Newfoundland respectively were inverted starting at ca. 420 Ma (O'Brien et al. 1991; Robinson et al. 1998; van Staal et al. 2009; van Staal and Barr 2012; Piñán-Llamas and Hepburn 2013). Coeval inversion of these basins is considered a proxy for the start of the Acadian collision between composite Laurentia and Avalonia, since most other informative rocks such as those deposited in the arc-trench gap (White et al. 2006) are generally not preserved. The general absence of Acadian forearc-related rocks suggest that forearc block subduction, subduction erosion (van Staal et al. 2009) and/or uplift and erosion of the forearc block prevailed during collision. Subduction erosion and/or forearc block subduction may have resulted in underplating of forearc rocks beneath Ganderia, further towards the Laurentian hinterland.

Subsidence analysis of the remnants of a Silurian passive margin built upon Avalonia revealed a significant phase of tectonic loading starting at the Silurian–Devonian boundary (~419 Ma) in Maritime Canada and formation of a marine foreland basin (Waldron et al. 1996). This tectonic loading is broadly coeval with termination of Coastal arc magmatism and inversion of the Acadian backarc basin(s), suggesting a tectonic relationship.

Late Early to Late Devonian clastic red sediments that were deposited on southern Newfoundland Avalonia immediately adjacent to the Dover-Hermitage Bay fault contain metamorphic and magmatic detritus sourced from the exhuming Acadian orogen in adjacent Ganderia (Williams and O'Brien 1995). These sediments are interpreted to represent the remnants of a terrestrial (molasse) foreland basin built on Avalonia during the Acadian collision. The combination of the absence of Early Devonian magmatism, evidence of tectonic loading and molasse sedimentation on Avalonia support our interpretation that it was situated on the lower plate during the Acadian collision.

Following the initiation of Acadian deformation at ca. 420 Ma near the Avalonian–composite Laurentia suture, the Acadian deformation is progressively younger to the northwest (Donohoe and Pajari 1974; Bradley et al. 2000) and predominantly has a northwest-directed vergence. Acadian deformation started between 408 and 406 Ma in central Maine and New Brunswick (Bradley and Tucker 2002; Wilson et al. 2004). Hence, it can be easily separated from the significantly older Salinic structures there. The Acadian retro-arc deformation front reached the border between Maine and Quebec by the end of the Early Devonian (~ 394 Ma, Bradley et al. 2000). A similar process may have also taken place in Newfoundland (see below). The combined geological evidence indicates that during the Early Silurian (443–425 Ma) Ganderia occupied a unique tectonic setting (van Staal et al. 2009), which has few modern analogues today. Ganderia was the lower plate at its leading Salinic edge and in an upper plate setting at its trailing Acadian edge (Figs. 2, 5). Such a setting is present in western Mindanao (Fig. 5; Sajona et al. 2000 and references therein) and a similar setting occurred in parts of Indonesia during the Late Cretaceous closure of the Meso-Tethys Ocean (Hall 2012). We argue below that this tectonic setting was ultimately critical for the generation of the large Acadian magmatic suite in Ganderia and formation of the immediately preceding, but problematic late-syn- to post-Salinic to pre-Acadian (423–417 Ma) magmatic suite, which occurs in the part of composite Laurentia underlain by the leading edge of Ganderia (east of the Red Indian Line in Fig. 1). If correct, this indicates there is a dynamic linkage between the Salinic and Acadian orogenies.

U–Pb GEOCHRONOLOGY

To better constrain the age of Salinic orogenesis, magmatism and sedimentation and to evaluate the sense of diachroneity, we sampled three felsic volcanic rocks in central Newfoundland. These include syn-tectonic rhyolite from the Topsails Suite along the Hungry Mountain thrust west of the Red Indian Line, and two samples from the Botwood belt east of the Red Indian Line. Botwood belt samples comprise Lawrenceton Formation rhyolite near Laurenceton and Port Albert Formation rhyolite tuff near Port Albert Peninsula.

Analytical Techniques

Heavy mineral concentrates were prepared from the samples using standard mineral separation techniques, including crushing, grinding, Wilfley™ table, and heavy liquid separation. Final sepa-

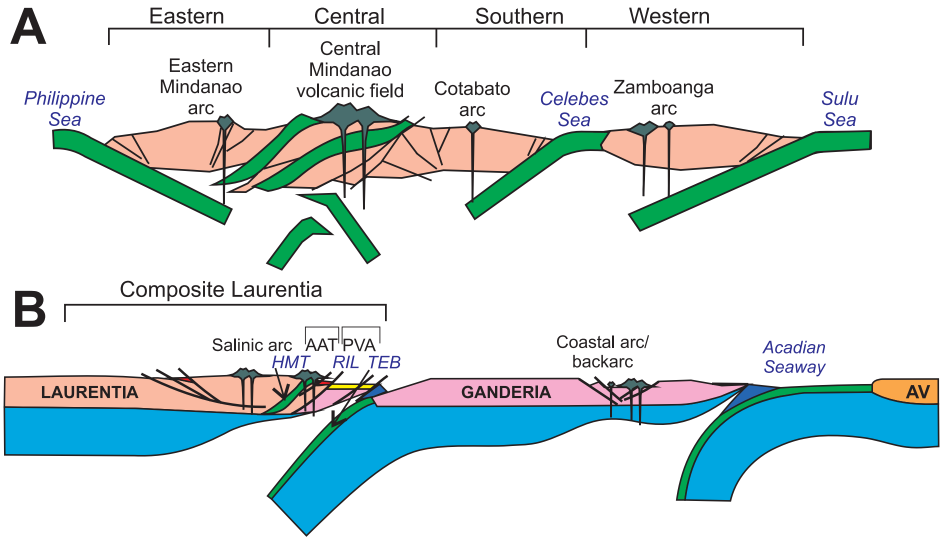

Figure 5. Comparison of the tectonic setting of (A) Mindanao (from Sajona et al. 2000) and (B) Ganderia during the Early Silurian. Subduction zones and accreted oceanic lithosphere outlined in green. The Popelogan-Victoria arc (PVA), the leading edge of Ganderia, had accreted to the Annieopsquotch accretionary tract (AAT) at ca. 455 Ma along the Red Indian Line (RIL). Subduction had stepped-back in the Tetagouche-Exploits backarc basin (TEB), which was largely floored by oceanic lithosphere. Late Ordovician–Early Silurian closure of the TEB was the principal cause of the Salinic orogenic cycle, formation of the Salinic arc and accretion of the remaining part of Ganderia to composite Laurentia. Silurian closure of the Acadian seaway between Ganderia and Avalonia (AV) may have been initiated as a result of the onset of collision between Ganderia and composite Laurentia. HMT: Hungry Mountain thrust. See text and Figure 8 for more details.

ration of the zircon grains was done by magnetic susceptibility using a Frantz™ isodynamic separator and hand-picking with a binocular microscope. SHRIMP II (Sensitive High Resolution Ion MicroProbe) analyses were conducted at the Geological Survey of Canada using analytical procedures described by Stern (1997), with standards and U–Pb calibration methods following Stern and Amelin (2003). Zircon from the samples and fragments of the GSC laboratory zircon standard (z6266 zircon, with $^{206}Pb/^{238}U$ age = 559 Ma) and a secondary zircon standard (Temora 2) were cast in an epoxy grain mount (GSC mount # 387), polished with diamond compound to reveal the grain centres, and photographed in transmitted light. Internal features of the zircon grains (such as zoning, structures, alteration, etc.) were characterized in back-scattered electron (BSE) and cathodoluminescence (CL) modes utilizing a Zeiss Evo 50 scanning electron microscope (SEM). Mount surfaces were evaporatively coated with 10 nm of high purity Au. Analyses were conducted using an $^{16}O^{-}$ primary beam, projected onto the zircon grains with an elliptical spot (13 µm x 16 µm in size). The count rates of ten masses including background were sequentially measured over 6 scans with a single electron multiplier and a pulse counting system with deadtime of 23 ns. Off-line data processing was accomplished using customized in-house software. The SHRIMP analytical data are presented in Table 1, where the 1σ external errors of $^{206}Pb/^{238}U$ ratios reported in the data table incorporate a 1.0% error in calibrating the standard zircon (Stern and Amelin 2003). No fractionation correction was applied to the Pb-isotope data; common Pb correction used the Pb composition of the surface blank (Stern 1997). Isoplot v. 3.00 (Ludwig 2003) was used to generate concordia plots, cumulative probability plots, and calculate weighted means. The error ellipses on the concordia diagrams and the weighted mean errors are reported at 2σ.

Harry's River Rhyolite

The Harry's River volcanic sequence (Figs. 3G, 6A, B) occurs in the immediate footwall of the Hungry Mountain Thrust (HMT; Fig. 1). This thrust represents the backstop of the Annieopsquotch accretionary tract (Fig. 2) and is a fundamental structure between it and the igneous rocks of the adjacent Notre Dame arc (van der Velden et al. 2004; Lissenberg et al. 2005; Zagorevski et al. 2009; van Staal et al. 2009). The Harry's River volcanic sequence was previously included in the Ordovician Buchans Group (Thurlow and Swanson 1987). The physical, structural and geochemical characteristics and its association with numerous, co-magmatic amygdaloidal mafic sills led us to suspect it formed part of the Silurian Topsails Suite and related rocks in the Springdale Group (429–426 Ma: Whalen et al. 1987, 2006). The rhyolite is macroscopically undeformed (Fig. 6A, B); however, the associated volcaniclastic rocks exhibit a very low metamorphic grade cleavage defined by chlorite and sericite (Fig. 3G). This cleavage is parallel to the foliation in the amphibolite- to greenschist-facies mylonite that characterizes the HMT. The occurrence of the cleavage and its parallelism with the HMT suggests that this deformation was associated with this fault.

A sample of red rhyolite (RAX06A058; z9554) was collected along Harry's River, north of the town of Buchans. The volcanic rock forms part of a sequence of brick-red flow-banded rhyolite, beige columnar-jointed rhyolite and orange, beige and light green tuff. The red rhyolite is in part extrusive and brecciated (Fig. 6A), although portions of it appear to be intrusive and display mutual cross-cutting relationships with mafic sills. The rhyolite is compositionally heterogeneous and contains reddish-grey zones that appear to be intermediate in composition. These zones contain brick-red felsic (Fig. 6B) and green mafic amoeboid inclusions, suggesting that the intermediate zones may represent hybridized mafic-felsic magma, similar to hybridized magmas common in the adjacent Topsails Igneous Suite (Whalen and Currie 1984; Whalen 1989).

The rhyolite sample yielded a

Table 1: U–Pb SHRIMP analytical data

Spot name	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	^{204}Pb (ppb)	^{204}Pb/^{206}Pb	± ^{204}Pb/^{206}Pb	f(206)^{204}	^{208*}Pb/^{206*}Pb	± ^{208}Pb/^{206}Pb	^{207*}Pb/^{235}U	± ^{207}Pb/^{235}U	^{206*}Pb/^{238}U	± ^{206}Pb/^{238}U	Corr Coeff	^{207*}Pb/^{206*}Pb	± ^{207}Pb/^{206}Pb	Apparent Ages (Ma) ^{206}Pb/^{238}U	± ^{206}Pb/^{238}U	^{207}Pb/^{206}Pb	± ^{207}Pb/^{206}Pb	Disc. (%)
Harry's River rhyolite: RAX06A058 (z9554); UTM NAD83 zone 21, 506590E - 5411251N																						
9554-5.1	423	218	0.533	31	1	0.000056	0.000045	0.0010	0.1819	0.0121	0.533	0.012	0.0689	0.0008	0.577	0.0561	0.0011	429	5	457	43	6.1
9554-9.1	412	215	0.539	30	2	0.000096	0.000057	0.0017	0.1806	0.0118	0.519	0.014	0.0684	0.0009	0.566	0.0550	0.0013	426	5	412	52	-3.4
9554-11.1	570	331	0.600	42	3	0.000087	0.000049	0.0015	0.1862	0.0049	0.518	0.017	0.0685	0.0008	0.455	0.0548	0.0016	427	5	402	68	-6.2
9554-19.1	588	281	0.493	42	2	0.000049	0.000039	0.0009	0.1543	0.0043	0.517	0.017	0.0681	0.0008	0.464	0.0551	0.0017	425	5	417	68	-1.9
9554-32.1	488	239	0.506	35	0	0.000010	0.000010	0.0002	0.1648	0.0068	0.534	0.011	0.0681	0.0009	0.712	0.0569	0.0008	425	5	487	31	12.8
9554-34.1	591	336	0.587	42	0	0.000010	0.000010	0.0002	0.1771	0.0094	0.525	0.010	0.0677	0.0008	0.719	0.0563	0.0007	422	5	463	29	8.8
9554-51.1	587	316	0.556	42	3	0.000082	0.000050	0.0014	0.1787	0.0086	0.509	0.015	0.0678	0.0009	0.550	0.0545	0.0013	423	5	393	55	-7.6
9554-52.1	260	127	0.505	18	3	0.000169	0.000182	0.0029	0.1587	0.0112	0.557	0.045	0.0670	0.0011	0.319	0.0603	0.0047	418	7	613	177	31.7
9554-53.1	289	143	0.510	21	1	0.000044	0.000061	0.0008	0.1716	0.0064	0.536	0.015	0.0684	0.0009	0.557	0.0568	0.0013	426	5	485	53	12.1
9554-50.1	427	224	0.541	30	3	0.000134	0.000060	0.0023	0.1561	0.0054	0.517	0.019	0.0674	0.0008	0.424	0.0557	0.0019	420	5	440	77	4.4
9554-47.1	555	304	0.566	40	4	0.000109	0.000054	0.0019	0.1802	0.0104	0.517	0.013	0.0685	0.0009	0.597	0.0547	0.0011	427	5	399	46	-7.1
9554-44.1	483	268	0.574	35	1	0.000039	0.000066	0.0007	0.1908	0.0067	0.532	0.018	0.0684	0.0009	0.486	0.0564	0.0017	426	5	469	66	9.2
9554-42.1	521	294	0.583	38	4	0.000119	0.000105	0.0021	0.1753	0.0103	0.528	0.024	0.0689	0.0009	0.393	0.0556	0.0023	430	5	436	95	1.5
9554-38.1	486	262	0.556	35	1	0.000041	0.000065	0.0007	0.1724	0.0058	0.515	0.014	0.0686	0.0008	0.520	0.0544	0.0013	428	5	387	53	-10.6
9554-37.1	421	221	0.543	30	4	0.000169	0.000082	0.0029	0.1510	0.0067	0.500	0.023	0.0682	0.0008	0.377	0.0532	0.0023	426	5	337	102	-26.2
9554-56.1	455	245	0.555	32	2	0.000065	0.000066	0.0011	0.1693	0.0109	0.521	0.020	0.0675	0.0008	0.408	0.0560	0.0020	421	5	451	82	6.6
9554-55.1	352	184	0.539	25	4	0.000203	0.000133	0.0035	0.1669	0.0080	0.502	0.023	0.0676	0.0008	0.365	0.0539	0.0023	421	5	368	99	-14.4
9554-54.1	381	188	0.510	27	3	0.000128	0.000110	0.0022	0.1574	0.0067	0.510	0.022	0.0682	0.0008	0.383	0.0542	0.0021	425	5	380	91	-11.9
9554-60.1	542	306	0.583	39	5	0.000154	0.000071	0.0027	0.1886	0.0061	0.491	0.015	0.0671	0.0009	0.545	0.0531	0.0014	419	5	331	59	-26.5
9554-59.1	506	284	0.579	37	4	0.000123	0.000052	0.0021	0.1847	0.0086	0.534	0.012	0.0689	0.0008	0.571	0.0562	0.0011	430	5	459	43	6.4
9554-61.1	359	181	0.521	25	10	0.000489	0.000211	0.0085	0.1559	0.0146	0.487	0.035	0.0672	0.0008	0.289	0.0526	0.0036	420	5	311	164	-35.0
9554-62.1	528	292	0.571	39	3	0.000111	0.000054	0.0019	0.1735	0.0052	0.539	0.013	0.0692	0.0008	0.562	0.0564	0.0012	432	5	469	46	8.0
9554-63.1	379	200	0.547	28	2	0.000079	0.000076	0.0014	0.1764	0.0112	0.539	0.025	0.0688	0.0008	0.358	0.0569	0.0025	429	5	487	98	11.9
9554-68.1	443	231	0.540	32	2	0.000066	0.000090	0.0011	0.1733	0.0062	0.517	0.022	0.0677	0.0008	0.384	0.0553	0.0021	422	5	426	89	0.9
9554-67.1	496	262	0.545	36	3	0.000113	0.000063	0.0020	0.1769	0.0074	0.517	0.016	0.0689	0.0011	0.615	0.0544	0.0013	430	7	387	56	-11.0
Grain mount IP494; spot size 17µm x 23µm; # of scans = 6; Error in ^{206}Pb/^{238}U calibration = 1.0%																						
RAX05-905 (z8772); UTM NAD83 zone 21, 679379E - 5491148N																						
8772-1.1	225	134	0.617	16	7	0.000553	0.000161	0.00959	0.2003	0.0076	0.507	0.027	0.0672	0.0009	0.368	0.0546	0.0027	420	5	397	115	-5.6
8772-2.1	180	95	0.545	13	7	0.000666	0.000249	0.01153	0.1718	0.0117	0.533	0.040	0.0672	0.0010	0.314	0.0576	0.0041	419	6	513	164	18.3
8772-18.1	263	117	0.459	19	4	0.000294	0.000111	0.0051	0.1571	0.0054	0.539	0.026	0.0674	0.0009	0.386	0.0580	0.0026	420	5	531	102	20.9
8772-19.1	94	29	0.319	6	5	0.000865	0.000305	0.01499	0.1128	0.0157	0.584	0.052	0.0675	0.0009	0.278	0.0628	0.0054	421	6	702	195	40.0
8772-20.1	165	70	0.439	12	4	0.000399	0.000177	0.00692	0.1552	0.0092	0.584	0.031	0.0671	0.0009	0.357	0.0632	0.0031	418	5	715	109	41.5
8772-4.1	131	48	0.381	9	7	0.000945	0.000393	0.01637	0.1158	0.0178	0.548	0.064	0.0677	0.0010	0.247	0.0587	0.0067	422	6	557	270	24.2
8772-5.1	96	37	0.398	7	6	0.001004	0.000349	0.0174	0.1488	0.0168	0.600	0.054	0.0667	0.0010	0.280	0.0653	0.0057	416	6	782	196	46.8
8772-6.1	82	29	0.362	6	6	0.001233	0.000423	0.02137	0.1121	0.0205	0.573	0.067	0.0677	0.0013	0.288	0.0614	0.0069	422	8	652	261	35.2
8772-21.1	92	30	0.343	6	8	0.001409	0.000396	0.02442	0.1124	0.0201	0.496	0.064	0.0670	0.0010	0.236	0.0536	0.0068	418	6	355	316	-17.9
8772-22.1	88	33	0.385	6	9	0.001733	0.000450	0.03004	0.1118	0.0191	0.448	0.067	0.0654	0.0011	0.231	0.0497	0.0073	408	6	182	310	-124.1
8772-24.1	132	63	0.492	9	10	0.001303	0.000253	0.02259	0.1366	0.0116	0.467	0.041	0.0683	0.0009	0.276	0.0496	0.0042	426	6	175	187	-144.0
8772-25.1	277	170	0.634	20	11	0.000653	0.000176	0.01132	0.1878	0.0086	0.510	0.029	0.0677	0.0008	0.328	0.0546	0.0029	422	5	395	125	-6.8
8772-26.1	88	25	0.293	6	6	0.001229	0.000554	0.0213	0.1022	0.0215	0.519	0.083	0.0667	0.0011	0.222	0.0565	0.0089	416	6	471	391	11.6
8772-27.1	106	52	0.509	8	3	0.000563	0.000300	0.00976	0.1633	0.0141	0.580	0.048	0.0672	0.0011	0.314	0.0626	0.0050	419	7	696	180	39.8
8772-11.1	236	116	0.509	16	12	0.000897	0.000346	0.01555	0.1548	0.0136	0.515	0.053	0.0669	0.0009	0.254	0.0559	0.0056	417	6	448	240	6.9
8772-12.1	105	28	0.273	7	11	0.001792	0.001146	0.03105	0.0785	0.0435	0.492	0.168	0.0658	0.0016	0.196	0.0543	0.0183	411	10	382	1017	-7.5
8772-28.1	253	117	0.479	18	12	0.000783	0.000248	0.01357	0.1505	0.0104	0.509	0.038	0.0669	0.0009	0.292	0.0551	0.0040	418	5	416	171	-0.3
8772-29.1	77	20	0.270	5	9	0.001929	0.000509	0.03344	0.0719	0.0202	0.460	0.077	0.0660	0.0011	0.227	0.0505	0.0083	412	7	220	340	-87.4
8772-30.1	98	44	0.467	7	11	0.001819	0.000398	0.03152	0.1482	0.0164	0.535	0.066	0.0667	0.0013	0.273	0.0582	0.0070	416	8	536	286	22.3
8772-14.1	88	29	0.346	6	8	0.001457	0.000414	0.02525	0.1176	0.0167	0.569	0.074	0.0671	0.0012	0.259	0.0616	0.0078	418	7	659	296	36.5
8772-15.1	333	190	0.589	24	14	0.000710	0.000151	0.0123	0.1771	0.0072	0.519	0.025	0.0677	0.0008	0.371	0.0556	0.0025	422	5	438	104	3.5
8772-16.1	109	51	0.484	7	18	0.002803	0.000721	0.04858	0.1311	0.0284	0.427	0.107	0.0652	0.0013	0.200	0.0475	0.0118	407	8	76	728	-435.4
RAX05-909 (z8776); UTM NAD83 zone 21, 629015E - 5450810N																						
8776-110.1	249	88	0.364	16	21	0.001456	0.000458	0.02524	0.1081	0.0190	0.465	0.071	0.0664	0.0011	0.227	0.0508	0.0076	415	6	231	312	-79.8
8776-53.1	169	70	0.431	11	19	0.001928	0.000297	0.03341	0.1214	0.0123	0.468	0.051	0.0665	0.0008	0.237	0.0511	0.0055	415	5	243	243	-70.4
8776-3.1	215	96	0.460	15	8	0.000640	0.000224	0.0111	0.1407	0.0097	0.524	0.036	0.0665	0.0010	0.335	0.0572	0.0038	415	6	498	152	16.6
8776-54.1	272	106	0.403	18	19	0.001186	0.000199	0.02056	0.1301	0.0100	0.510	0.039	0.0665	0.0012	0.352	0.0557	0.0040	415	7	438	168	5.3
8776-9.1	112	42	0.386	8	11	0.001593	0.000441	0.02762	0.1223	0.0174	0.519	0.071	0.0669	0.0010	0.233	0.0563	0.0075	417	6	463	327	9.9
8776-23.1	182	81	0.459	13	14	0.001294	0.000413	0.02242	0.1425	0.0173	0.522	0.063	0.0671	0.0011	0.252	0.0565	0.0066	419	6	471	282	11.1

continued

Table 1: U–Pb SHRIMP analytical data *(Continued)*

Spot name	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	^{204}Pb (ppb)	^{204}Pb/^{206}Pb	± ^{204}Pb/^{206}Pb	f(206)^{204}	^{208*}Pb/^{206*}Pb	± ^{208}Pb/^{206}Pb	^{207*}Pb/^{235}U	± ^{207}Pb/^{235}U	^{206*}Pb/^{238}U	± ^{206}Pb/^{238}U	Corr Coeff	^{207*}Pb/^{206*}Pb	± ^{207}Pb/^{206}Pb	Apparent Ages (Ma) ^{206}Pb/^{238}U	± ^{206}Pb/^{238}U	^{207}Pb/^{206}Pb	± ^{207}Pb/^{206}Pb	Disc. (%)
RAX05-909 (z8776); UTM NAD83 zone 21, 629015E - 5450810N *(Continued)*																						
8776-119.1	390	184	0.489	27	25	0.001105	0.000178	0.01914	0.1385	0.0100	0.480	0.030	0.0673	0.0008	0.307	0.0518	0.0031	420	5	275	141	-52.6
8776-118.1	154	46	0.312	10	23	0.002468	0.000558	0.04277	0.1069	0.0237	0.504	0.085	0.0673	0.0012	0.225	0.0543	0.0090	420	7	384	384	-9.5
8776-28.1	330	149	0.467	23	18	0.000931	0.000198	0.01614	0.1414	0.0091	0.491	0.035	0.0674	0.0008	0.281	0.0529	0.0037	420	5	325	167	-29.2
8776-7.1	155	51	0.342	10	7	0.000761	0.000190	0.01318	0.0981	0.0082	0.545	0.031	0.0677	0.0008	0.324	0.0583	0.0032	422	5	542	125	22.0
8776-115.1	274	116	0.435	19	15	0.000921	0.000188	0.01596	0.1428	0.0090	0.559	0.031	0.0677	0.0008	0.335	0.0599	0.0031	422	5	600	118	29.6
8776-26.1	316	116	0.380	21	20	0.001094	0.000168	0.01896	0.1049	0.0070	0.482	0.028	0.0678	0.0009	0.343	0.0516	0.0028	423	5	269	129	-57.1
8776-45.1	171	56	0.340	12	18	0.001793	0.000279	0.03108	0.1026	0.0115	0.503	0.046	0.0678	0.0010	0.283	0.0539	0.0048	423	6	365	213	-15.9
8776-30.1	565	411	0.752	43	16	0.000477	0.000082	0.00827	0.2353	0.0044	0.527	0.020	0.0678	0.0007	0.398	0.0563	0.0019	423	4	465	78	9.1
8776-62.1	231	95	0.427	16	18	0.001319	0.000518	0.02287	0.1288	0.0199	0.527	0.081	0.0679	0.0011	0.226	0.0563	0.0085	424	7	465	375	8.9
8776-60.1	324	103	0.329	22	11	0.000595	0.000144	0.01031	0.1096	0.0062	0.572	0.025	0.0680	0.0008	0.386	0.0610	0.0024	424	5	639	88	33.6
8776-38.1	144	50	0.356	10	14	0.001639	0.000290	0.0284	0.1109	0.0146	0.559	0.047	0.0681	0.0010	0.302	0.0596	0.0048	425	6	588	186	27.8
8776-117.1	177	58	0.337	12	18	0.001726	0.000322	0.02992	0.1178	0.0132	0.562	0.051	0.0682	0.0009	0.264	0.0598	0.0053	425	5	597	203	28.8
8776-44.1	254	130	0.529	18	21	0.001354	0.000227	0.02346	0.1476	0.0101	0.493	0.037	0.0686	0.0010	0.314	0.0522	0.0037	428	6	293	172	-45.8
8776-113.1	139	71	0.526	11	20	0.002225	0.000734	0.03857	0.1521	0.0296	0.557	0.120	0.0727	0.0015	0.216	0.0556	0.0118	452	9	437	437	-3.4
8776-109.1	114	48	0.436	8	22	0.002928	0.000472	0.05075	0.1062	0.0190	0.462	0.080	0.0727	0.0012	0.215	0.0461	0.0078	452	7	7	473	-6480.1
8776-8.1	120	48	0.415	9	14	0.001805	0.000317	0.03128	0.1035	0.0128	0.422	0.053	0.0729	0.0011	0.242	0.0420	0.0052	453	7	0	0	0.0
8776-20.1	70	29	0.425	5	12	0.002619	0.000491	0.0454	0.1447	0.0245	0.585	0.085	0.0734	0.0015	0.264	0.0578	0.0082	457	9	521	344	12.2
8776-116.1	226	111	0.507	17	21	0.001417	0.000219	0.02455	0.1544	0.0094	0.520	0.039	0.0741	0.0010	0.303	0.0508	0.0037	461	6	233	175	-97.5
8776-56.1	48	33	0.707	4	21	0.006055	0.001061	0.10494	0.1938	0.0440	0.511	0.208	0.0746	0.0023	0.200	0.0497	0.0199	464	14	180	1364	-157.1
8776-108.1	178	69	0.400	13	25	0.002111	0.000301	0.03659	0.1095	0.0122	0.558	0.052	0.0749	0.0010	0.269	0.0541	0.0049	465	6	374	220	-24.4
8776-40.1	204	85	0.430	16	15	0.001158	0.000219	0.02007	0.1413	0.0103	0.578	0.039	0.0750	0.0009	0.299	0.0559	0.0037	466	5	448	152	-4.1
8776-46.1	194	115	0.616	16	19	0.001519	0.000237	0.02632	0.1893	0.0116	0.591	0.043	0.0751	0.0010	0.296	0.0571	0.0040	467	6	494	163	5.6
8776-33.1	227	100	0.454	18	9	0.000592	0.000375	0.01027	0.1473	0.0148	0.610	0.064	0.0751	0.0010	0.252	0.0589	0.0060	467	6	564	239	17.2
8776-57.1	204	60	0.302	16	12	0.000857	0.000226	0.01484	0.1017	0.0094	0.660	0.041	0.0768	0.0009	0.310	0.0624	0.0037	477	5	687	133	30.6
8776-111.1	427	108	0.261	71	15	0.000250	0.000064	0.00433	0.0796	0.0028	1.753	0.035	0.1686	0.0018	0.633	0.0754	0.0012	1005	10	1079	32	6.9
8776-61.1	699	104	0.153	364	19	0.000064	0.000017	0.00111	0.0426	0.0008	11.916	0.140	0.4985	0.0054	0.961	0.1734	0.0006	2607	23	2590	6	-0.7

Notes (see Stern 1997):
Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number. Multiple analyses in an individual spot are labelled as x-y.z.z
Uncertainties reported at 1σ (absolute) and are calculated by numerical propagation of all known sources of error
f206^{204} refers to mole fraction of total ^{206}Pb that is due to common Pb, calculated using the ^{204}Pb-method; common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)
* refers to radiogenic Pb (corrected for common Pb)
Discordance relative to origin = 100 * ((207/206 age –206/238 age)/(^{207}Pb/^{206}Pb age))
Calibration standard 6266; U = 910 ppm; Age = 559 Ma; ^{206}Pb/^{238}U = 0.09059
Th/U calibration: F = 0.03900*UO + 0.85600
Grain mount IP387; spot size 13μm x 16μm; # of scans = 6; Error in ^{206}Pb/^{238}U calibration = 1.0%

moderate number of zircon grains. Most are quite magnetic. The euhedral zircon grains are predominantly stubby prismatic crystals in the 100 μm to 200 μm size range (Fig. 6C inset). Most of the grains contain numerous inclusions and fractures. BSE–SEM imaging reveals faint growth or sector zoning in many of the grains. SHRIMP II analyses comprise a single age population with no evidence of inheritance (Fig. 6C). A concordia age for these analyses is calculated to 425.3 ± 2.0 Ma (MSWD of concordance and equivalence = 0.72; probability = 0.98; n = 25). Taking into account the error associated with the zircon standards analyzed on the SHRIMP grain mount, the crystallization age of the Harry's River rhyolite is interpreted to be 425 ± 4 Ma.

Fogo Suite Felsic Volcanic Rocks, Lawrenceton Formation of the Botwood Group

Lawrenceton Formation is exposed in the core of an anticline near Laurenceton (Fig. 1). Felsic volcanic rocks are prominently exposed along a ridge crest that crosses the Burnt Arm Pond road, where they are dominated by felsic pyroclastic breccia. Lithic clasts comprise flow-banded, K-feldspar-porphyritic rhyolite, aphyric rhyolite and flattened pumice in a fine-grained crystal-rich matrix (Fig. 6D). Outcrops are pervasively fractured by steep joints. A cleavage is lacking. Crude size distribution of clasts suggests that the felsic volcanic rocks are sub-horizontal. The breccia forms part of unit 4 of the Lawrenceton Formation as mapped by Colman-Sadd (1994), which overlies a cleaved polymict conglomerate (unit 3 of the Lawrenceton Formation of Colman-Sadd 1994) containing rounded to angular clasts of basalt, rhyolite and red sandstone. The rocks of unit 3 have a steeply dipping layering and a pronounced cleavage. The contact between units 3 and 4 is not exposed on Burnt Arm Pond road, but may be nonconformable because of the apparent differences in structure, a relationship also suggested as an option by Colman-Sadd (1994).

Sample RAX05-909 was collected from monomict, trachytic, felsic pyroclastic breccia (Fig. 6D) with ovoid spherulites (?) and banded perlite. The

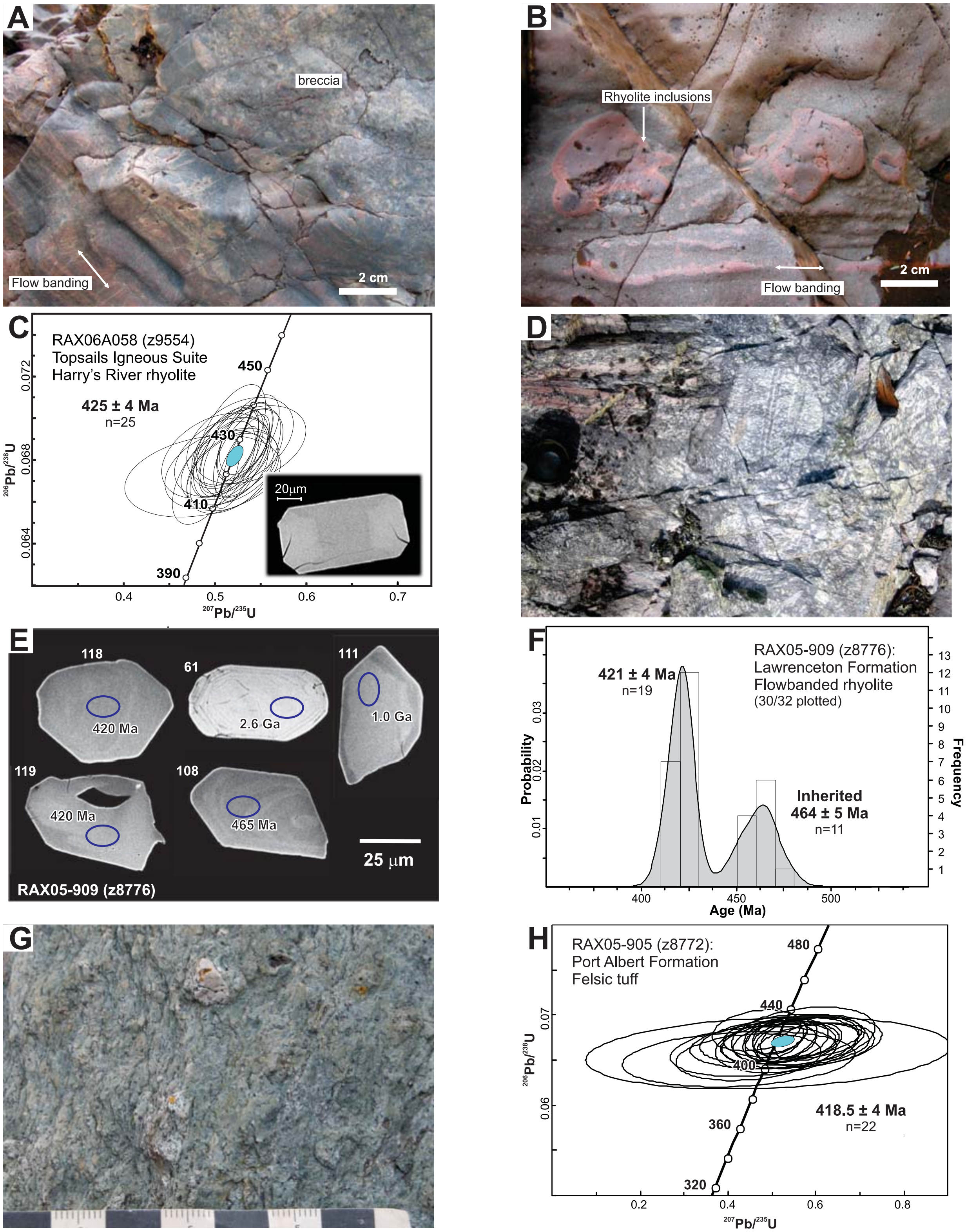

Figure 6. A, B: Representative photographs of the Harry's River rhyolite (Topsails Suite) below the Hungry Mountain Thrust. A. Flow-banded rhyolite transitions to volcanic breccia supporting its extrusive nature. Nearby outcrops of the rhyolite are columnar-jointed. B. Hybrid zone in rhyolite at the dated locality. Hybrid zone is characterized by rhyolite inclusions in intermediate flow. Amoeboid inclusions of basalt occur throughout this outcrop. C. U–Pb Concordia diagram of the Harry's River rhyolite. D. Representative photograph of Lawrenceton Formation monomictic rhyolite breccia. E. Representative SEM images of igneous and inherited zircon grains. F. Probability plot for zircon analyses of the Lawrenceton Formation rhyolite. G. Representative photograph of Port Albert Formation rhyolitic lapilli tuff. H. U–Pb Concordia diagram of Port Albert Formation rhyolitic lapilli tuff.

rhyolite sample yielded zircon grains ranging from <50 μm to >200 μm in size including: equant multifaceted crystals, euhedral stubby prisms to prismatic crystals, subhedral grains with rounded facets, and rounded grains (Fig. 6E, Appendix 1). Many zircon grains contain inclusions or fractures and some are quite altered.

BSE–SEM imaging reveals weak growth zoning in some of the zircon crystals. SHRIMP II analyses of zircon grains from the rhyolite resulted in several age populations (Fig. 6F, Table 1). The oldest analyses include grains of 2.6 Ga and 1.0 Ga, which are interpreted to represent inherited components (not plotted). An age population of 464 ± 5 Ma (weighted average of $^{206}Pb/^{238}U$ ages; MSWD = 1.4, probability = 0.16, n = 11) is also interpreted to be inherited in origin. Analyses that are interpreted as inherited were obtained from probable inherited cores as well as entirely inherited zircon grains with no evidence of younger overgrowths. A weighted average $^{206}Pb/^{238}U$ age of the youngest dominant age population is calculated to be 421.0 ± 2.5 Ma (MSWD = 0.45, probability = 0.98, n = 19). Taking into account the associated error on the standard zircon grains analyzed on the SHRIMP grain mount, the crystallization age of the rhyolite is interpreted to be 421 ± 4 Ma.

Fogo Suite Rhyolite Lapilli Tuff, Port Albert Formation of the Botwood Group

Port Albert Peninsula locally exposes the contact between Badger and Botwood groups. On the northwest limb of the Farewell syncline, 1 km northeast of the Port Albert Harbour, rocks of the Badger Group are overlain by volcanic and sedimentary rocks of the Port Albert Formation (Karlstrom et al. 1982; Williams et al. 1993), a correlative of the Lawrenceton Formation (Fig. 4G). The base of the Port Albert Formation is characterized by a thin coarse sandy bed with a conglomeratic base (Williams 1993a; Williams et al. 1993). The latter contains pebbles and cobbles of cleaved Badger Group, suggesting that the deposition of the Port Albert Formation postdates early Salinic deformation of the underlying Badger Group (Reusch 1987; van der Pluijm et al. 1993), which mimics the unconformable relationships observed between the Badger and Botwood groups on the nearby Change Islands (van der Pluijm et al. 1993).

Above the unconformity, the Port Albert Formation is characterized by vesicular mafic flows and felsic tuff. Sample RAX05-905 was collected from the outcrops stratigraphically above the Badger–Botwood unconformity (i.e. topographically above the outcrop shown in Fig. 4G). The sample comprises an aphyric, rhyolitic lapilli tuff (Fig. 6G). The sample yielded abundant euhedral zircon grains ranging in size from 75 to 150 μm (Appendix 1). Zircon morphologies include stubby prismatic to slightly elongate crystals and most of the zircon grains contain numerous inclusions and fractures. SHRIMP II analyses from the rhyolite produced a single age population (Fig. 6H; Table 1). A weighted average $^{206}Pb/^{238}U$ age of all of these analyses is calculated to be 418.5 ± 2.5 Ma (MSWD = 0.52, probability = 0.96, n = 22). Taking into account the error on the standard zircon grains analyzed on the SHRIMP grain mount, the crystallization age of the rhyolite is interpreted to be 418.5 ± 4 Ma, which overlaps in error with the 422 ± 2 Ma Port Albert felsite of Elliott et al. (1991), suggesting they are consanguineous.

DISCUSSION

Timing of Salinic Collision-Related Deformation Northwest of the Red Indian Line

The oldest Salinic tectonism in the peri-Laurentian realm (Notre Dame subzone of Williams et al. 1988) (Figs. 1, 2) is locally recorded by the Notre Dame Arc, which underwent complex exhumation in Late Ordovician (van Staal et al. 2007). In southwestern Newfoundland, a thick sequence of conglomerate, sandstone and terrestrial ignimbrite of the Windsor Point Group (ca. 453 Ma: Dubé et al. 1996) unconformably overlies Taconic deformed Ordovician Notre Dame arc plutons (ca. 488 to 459 Ma: van Staal et al. 2007). Similar evidence for Salinic exhumation is preserved in northwestern Newfoundland, where 442 Ma and younger terrestrial red beds unconformably overlie Upper Ordovician–Lower Silurian plutons (e.g. ca. 446 Ma Burlington granodiorite: Skulski et al. 2010; Fig. 1) of the southeast-facing Salinic arc (Whalen et al. 2006). In addition, Salinic exhumation of the Notre Dame arc in Late Ordovician (Katian to Llandoverian) is supported by peri-Laurentian provenance of Late Ordovician sedimentary rocks to the east of the Red Indian Line in the Badger Group (van Staal et al. 2009; Waldron et al. 2012; see following). Deformation, uplift and erosion continued into the Wenlockian, when extrusive components of the syn-collisional Topsails Suite and Springdale Group were unconformably deposited above the exhumed Notre Dame arc (Whalen et al. 2006). The Topsails Suite and Springdale Group volcanic and interlayered terrestrial red sandstone (Old Red Sandstone) (Fig. 3E) and consanguineous plutons form an overstep sequence on all peri-Laurentian terranes and can thus be utilized as markers of Salinic deformation. The age of the oldest phases of the Topsails Suite indicate that the Salinic collision-related deformation had started by at least 433 Ma (Whalen et al. 2006). These relationships provide strong evidence for uplift, exhumation, progressive localization of deformation into high level ductile-brittle structures and cratonization of the composite Laurentian margin (van Staal et al. 2009).

The ca. 425 Ma age of the Harry's River rhyolite confirms our hypothesis that it forms a late phase of the Silurian Topsails Suite. Thus, Topsails Suite volcanic and clastic sedimentary rocks occur in both the hanging and footwall of the HMT. Topsails Suite sills, dikes and plutons that are consanguineous with the dated rhyolite cut across the HMT and related faults (Figs. 3C, 7), suggesting that magmatism was contemporaneous with latest faulting. These relationships indicate that the Middle Ordovician HMT was reactivated at ca. 425 Ma during the Salinic collision-related deformation. This result provides tighter age constraints on the age of the Salinic collision-related deformation in the Annieopsquotch accretionary tract (Zagorevski et al. 2007), and the area straddling the Red Indian Line (Fig. 1), which may have slightly outlasted

Figure 7. Flow-banded (inset) Topsails Suite rhyolitic dike (correlative of Harry's River rhyolite) cuts across foliation and a minor thrust fault, Buchans area. The minor thrust is a subsidiary structure to the Hungry Mountain and Airport thrusts (Thurlow et al. 1992).

Salinic deformation further to the west (see above). In addition, this age also dates the Silurian terrestrial red beds (Old Red Sandstone), which are invariably interlayered with the volcanic rocks (Fig. 3E) in this part of Newfoundland. They confirm that this part of composite Laurentia was now fully uplifted above sea level and cratonized (van Staal et al. 2009).

Similar relationships are observed along the Otter Brook shear zone (Fig. 2), a correlative of the HMT to the southwest. The Otter Brook shear zone is principally an Ordovician structure (Lissenberg et al. 2005) that was reactivated during the Silurian (Zagorevski et al. 2007). Silurian red beds of the Topsails Suite occur both in the hanging wall and footwall of the Otter Brook shear zone (Zagorevski et al. 2007). The latest motion of the Otter Brook Shear zone is constrained by a ca. 426 Ma age (titanite) on Silurian granite that has been deformed at amphibolite facies in the hanging wall and sub-greenschist facies in the footwall. Similar relationships are described in the Notre Dame Bay area (Dean and Strong 1977); however, absolute age constraints are lacking there.

Timing of Salinic Collision-Related Deformation Southeast of the Red Indian Line

Nature of the Contact between Badger and Botwood Groups

East or southeast of the Red Indian Line, Ordovician phases of the extinct Popelogan-Victoria arc are overlain by marine sedimentary rocks of the Katian–Llandoverian Badger Group and terrestrial volcano-sedimentary rocks of the younger Botwood Group. The nature of the contact between the Badger and Botwood groups has been a matter of contentious debate and both conformable and unconformable relationships have been proposed (Fig. 4G; Williams 1993a; van der Pluijm et al. 1993; O'Brien 2003; Dickson 2006; McNicoll et al. 2006). Our new age dates of the Botwood Group supplement sparse existing ages pertaining to the nature of the contact between the Badger and Botwood groups.

Botwood Group felsic volcanic rocks in the Lawrenceton (421 ± 4 Ma) and Port Albert formations (418.5 ± 4 Ma) yielded within error ages that constrain the age of Botwood Group magmatism to the Ludlovian–Pridolian and possibly the lowermost part of the Lochkovian (ICS 2013). These ages are consistent with the inferred fossil age range for the Botwood Group following reclassification of the Silurian units in central Newfoundland (Dickson 1994; Boyce and Ash 1994). The Botwood Group volcanic rocks overlap within error the U–Pb zircon ages of the spatially associated igneous rocks of the Fogo suite (ca. 422 Ma Fogo Batholith: Aydin 1995; ca. 424 Ma Mount Peyton Suite: Dunning 1992, 1994; ca. 422 Ma Port Albert composite dikes: Elliott et al. 1991). This supports the proposed consanguineous relationship between these volcanic and plutonic rocks.

Our new age constraints on the Botwood Group have major implications for the stratigraphy and deformation history west of the Dog Bay Line. They indicate that the contact between the Badger and the Botwood groups is everywhere unconformable. Badger Group has been constrained as no younger than Telychian (> 433 Ma: O'Brien 2003), whereas Botwood Group and consanguineous rocks are younger than 426 Ma. This suggests that the contact between Badger and Botwood groups is characterized by a hiatus of 11–7 m.y. (433–422 Ma) based on the most recent time-stratigraphic table (ICS 2013). This interpretation contrasts with Williams (1993a), but is consistent with our observations and those of other workers (van der Pluijm et al. 1993; O'Brien 2003; Dickson 2006; McNicoll et al. 2006). The Badger–Botwood unconformity likely corresponds to the main phase of collision-related Salinic deformation (D_1 of Elliott et al. 1991; Lafrance and Williams 1992; van der Pluijm et al. 1993; Kusky and Kidd 1996 and O'Brien 2003) and covers a similar time-span determined for the Salinic collision-related deformation in correlative rocks in central New Brunswick (van Staal et al. 2003; Wilson and Kamo 2012).

Fogo Suite and Botwood Group Deformation

The Siluro–Devonian deformation is complex and polyphase (e.g. Lafrance and Williams 1992; Piasecki 1992; Holdsworth 1994; O'Brien 2003; van Staal et al. 2009).The evidence presented above indicates that Fogo suite magmatism (Fig. 1) and deposition of the Botwood Group postdated the main phase of collision-related Salinic deformation. Deformation that is preserved in the Fogo suite and Botwood Group thus must represent either (1) the latest stages of the Salinic orogeny or (2) the start of the Acadian Orogeny. In general, Fogo suite plutonic rocks do not display penetrative deformation; however, the Fogo Island batholith is tilted, exposing the full section of a bimodal magma chamber from layered gabbro to evolved granite and high-level hypabyssal rocks (Fig. 1; Kerr 2013). Associated Botwood Group rocks are commonly folded and cleaved suggesting that strain may be heterogeneous and preferentially partitioned into relatively incompetent rocks such as the Botwood Group sandstone and siltstone rather than into competent igneous rocks. However, as the Mt Peyton gabbro (424 ± 2 Ma: Dunning 1992) was interpreted to intrude into deformed Botwood Group, the deformation preceded and/or was coeval with magmatism. Alternatively, there may be a cryptic unconformity within the Botwood Group.

In the southwestern part of the Botwood belt, an unconformable relationship was inferred between the Wigwam (older) and Stony Lake (younger) formations of the Botwood Group. Macroscopically undeformed, subhorizontally overlying felsic volcanic rocks of the 423 $^{+3}/_{-2}$ Ma Stony Lake Formation have been interpreted to unconformably overlie folded and cleaved red sandstone of the Wigwam Formation (Anderson and Williams 1970; Colman-Sadd and Russell 1982; Williams et al. 1995; Pollock et al. 2007). An unconformable relationship was also inferred between the 423 ± 3.5 Ma Patch Valley rhyolite and the Botwood Group (Rogers and van Staal 2005; McNicoll et al. 2008). The Stony Lake Formation contains several volcanic units (Colman-Sadd and Russell 1982) and our field investigations confirm that they do indeed locally appear fresh and undeformed; however, they are folded and cleaved elsewhere (Fig. 4E, F). The marked contrast in the degree of deformation from one locality to the other suggests (1) that the heterogeneous development of cleavage may be rheologically controlled and the deformation of the Botwood Group everywhere postdated the Fogo suite or (2) that the deposition of the Botwood Group and the Fogo suite magmatism partially overlapped the last gasp of Salinic deformation. The evidence for either case is ambiguous with the existing dataset.

Scenario 1 is simple and consistent with the unconformable relationship between the Salinic-deformed Badger and Botwood groups. In this scenario, the deformation postdated the Fogo suite and the strain was heterogeneous. The coplanar nature of the Acadian structures in the deformed Indian Island belt rocks (McNicoll et al. 2006) with those in the adjacent Botwood belt could suggest that the main folding and cleavage forming event in the Botwood belt is Acadian. This is consistent with the age constraints for the main phase of Acadian folding and cleavage formation in the adjacent Indian Island belt (415–410 Ma: McNicoll et al. 2006) and age constraints on Acadian (F_3 dextral transpression of Lafrance and Williams 1992) deformation in the Notre Dame Bay area (≥ 408 Ma: Elliott et al. 1991; O'Brien 2003). However, as some phases of the Fogo suite cut deformed and cleaved Badger and Botwood groups rocks (McNicoll et al. 2006; Dickson et al. 2007), there is a requirement for deformation following the initiation of Botwood deposition but prior to intrusion of parts of the ca. 424–422 Ma Fogo suite. This indicates that at least some of the deformation in the Botwood Group is pre-Acadian. This deformation may equate with the south-verging F_2 folds of Lafrance and Williams (1992).

Scenario 2 implies that the major structures in the Botwood belt were late Salinic and that folding and cleavage formation must have taken place over a very short time span. This is consistent with the observation that some Fogo suite phases cut folded and/or cleaved Badger and Botwood groups rocks (McNicoll et al. 2006; Dickson et al. 2007; Kerr 2013) and are deformed by the same deformation elsewhere. The coplanar orientation of some of the Botwood belt structures and the Acadian structures in the adjacent Indian Islands belt (McNicoll et al. 2006) may be coincidental, as the strain axes related to Salinic and Acadian convergence may have been similar locally. Hence, these structures may be unrelated and/or Acadian deformation reactivated and/or reoriented Salinic structures (see Lafrance and Williams 1992).

The second scenario requires that the Botwood Group red beds were deposited syn-tectonically near the late-Salinic deformation front, similar to what happened slightly earlier farther to the west (see above), and that there is evidently west to east diachroneity in the timing of Salinic tectonism, magmatism and red bed deposition (e.g. O'Brien 2003; van Staal et al. 2009). These relationships mirror those established in central New Brunswick (van Staal et al. 2003, 2009; Wilson et al. 2004, 2008; Wilson and Kamo 2012) and, weighing all combined lines of evidence, favour the second model. This scenario is complex but not anomalous in the northern Appalachians (e.g. see Bradley and Tucker 2002) and other dissected shallow level orogenic tracts preserved elsewhere (e.g. White 1982). Our new interpretations thus indicate that the Salinic orogeny may have locally lasted until the end of the Silurian (~ 419 Ma) or even slightly younger farther east, was diachronous from west to east and spatially and temporally overlapped bimodal magmatism and terrestrial red bed sedimentation.

Separation of the Salinic and Acadian Structures

Acadian ductile deformation in Newfoundland started at the leading edge of composite Laurentia (originally the trailing edge of Ganderia) at ca. 420 Ma with the inversion of the La Poile basin in Newfoundland (O'Brien et al. 1991) but was progressively younger to the northwest. It took place between 415 and 410 Ma near the Dog Bay Line (McNicoll et al. 2006) and had reached the Red Indian Line probably slightly before 408 Ma (Elliott et al. 1991), because locally it also affects Emsian terrestrial sandstone occurring immediately west of it (Chandler 1982).

Separation of Salinic and Acadian structures on the basis of age is therefore possible in central Newfoundland (e.g. Zagorevski et al. 2007), but is problematic in the area where the Acadian started near the leading edge of composite Laurentia. For example, deep-seated Acadian metamorphic nappes started to form shortly after ca. 418 Ma (Valverde-Vaquero et al. 2000, 2006; van der Velden et al. 2004) immediately northwest of the Coastal arc-back arc system and thus may have overlapped in time with the last gasps of Salinic-related deformation further to the west or north. However, the fact that the La Poile basin (Coastal arc-back arc system) was an extensional intra-arc or backarc basin until 420 Ma, suggests that the Silurian Salinic deformation front never crossed the full width of the Gander margin and left its trailing (eastern) edge partially untouched. Separation of Salinic and Acadian structures in this part of Newfoundland can thus be achieved on the basis of geological arguments and kinematic analysis.

Salinic Orogenesis and Deposition of the Old Red Sandstone

One of the important implications of our new data is that the final phase of Salinic deformation, slab break-off-related magmatism and sedimentation of clastic terrestrial molasse rocks became progressively younger from west to east. The terrestrial sedimentary rocks thus form a time-transgressive overstep sequence that links together the sequentially accreted terranes to composite Laurentia from west to east (van Staal et al. 2009), as well as dating when these terranes became fully emergent and hence cratonized. These clastic rocks were probably deposited in molasse-like foreland basins, sourced from the nearby, progressively rising Salinic mountains, situated west of an east-migrating late Salinic continental deformation front. The red beds, carefully dated using their interlayered felsic volcanic rocks, form part of the Old Red Sandstone facies as defined in the British Caledonides (Friend et al. 2000), which had started by at least 442 Ma in western Newfoundland. This is significantly older than the oldest known Old Red Sandstone lithologies in the British Isles, which are Wenlockian and/or younger (Friend et al. 2000). The Old Red Sandstone overstepped the Red Indian Line and the western part of the marine Badger belt by ca. 429 Ma (the age of the subaerial Charles Lake volcanic suite of Dickson (2000) which occurs immediately east of the Red Indian Line) followed by deposition of the Ludlovian–early Lochkovian red beds of the Botwood Group farther to the east. The Upper Silurian red beds overstep the Dog Bay Line tectonic zone and lie disconformably on Llandoverian–Ludlovian marine sedimentary rocks of the Indian Islands Group and underlying Ordovician rocks deposited on the Gander margin (Williams et al. 1993; Currie 1997; Dickson et al. 2007). Hence, Ganderia was fully emerged and incorporated into composite Laurentia by the end of the Silurian. The next overstep phase of the Old Red Sandstone took place when it was deposited onto Avalonia. This phase probably took place late during the Early Devonian (Williams and O'Brien 1995). Like their counterparts in the British Isles the Old Red Sandstone was consistently deformed by Early Devonian Acadian deformation (Soper and Woodcock 2003; van Staal and Zagorevski 2012) and thus represents an important tectonic marker.

Our interpretation also implies that the regional sub-Botwood unconformity represents a change from Katian–Ludlovian marine deposition in the arc-trench gap and foredeep during subduction and early collision (van Staal et al. 1998, 2009; Zagorevski et al. 2008; Waldron et al. 2012) to a late syn-collisional molasse foreland basin setting during the Late Silurian. This change followed progressive Salinic deformation and uplift of the marine Badger basin and the adjacent Late Ordovician–Ludlovian marine Indian Island foredeep (Boyce and Dickson 2006; Dickson et al. 2007; van Staal et al. 2009; van Staal and Zagorevski 2012). Red siliciclastic sedimentary rocks of the Big Indian Pond and Ten Mile Lake formations overlie the marine sedimentary rocks of the Indian Island Group (Williams 1993a; Currie 1995; Dickson 2006). The latter are younger than ca. 430 Ma and older than ca. 411 Ma based on U–Pb detrital zircon data (Pollock et al. 2007) and intrusive relationships (McNicoll et al. 2006; Dickson et al. 2007). Fossil evidence suggests these red beds are younger than the Ludlovian. Hence, they are obvious correlative units of the Botwood Group. The Old Red Sandstone, including the Botwood Group, thus probably represents a time-transgressive, terrestrial overstep sequence as was previously suggested by Pollock et al. (2007) and van Staal et al. (2009). As predicted by our model, the red bed–molasse deposition on Avalonia is even younger and started during the Early Devonian (Williams and O'Brien 1995) as a result of its Acadian collision with Laurentia.

Tectonic Model for East-Migrating Bimodal Magmatism

The widespread bloom of late Early Silurian to Early Devonian (433–419 Ma) bimodal magmatism, concentrated in the central orogenic core of the Canadian Appalachians (i.e. central mobile belt of Williams 1964) has been interpreted to be generated during break-off of the northwest-dipping Salinic slab and formation of an asthenospheric window beneath the Salinic collision zone (Whalen et al. 2006; van Staal et al. 2009). The slab break-off model is consistent with the observed spatial and temporal overlap between Salinic collision-related deformation and magmatism and concurrent rapid uplift (van Staal et al. 2007), as well as sedimentation in Silurian continental (red beds) and marine (Indian Island Group) foreland basins. The latter phenomenon is primarily a consequence of the isostatic rebound and unloading of the upper plate following break-off of the subducting slab (Regard et al. 2008 and references therein) and continuing deformation; however, this model raises the question of why the slab break-off-related processes are diachronous from west to east (Fig. 2). That is, they were retreating towards the foreland (Gander margin), over a period of 6 to 10 m.y. Alternative models capable of producing asthenospheric windows beneath the Salinic orogen, such as retreating delamination (progressive peeling-off of the lithospheric mantle) are not considered viable. This is because the latter process requires more time than is observed and also predicts significant uplift and at least locally, exhumation of deeply buried crust in the collision zone (e.g. see Krystopowicz and Currie 2013), either of which is rare in the core of the Canadian Appalachians (van Staal and de Roo 1995; van Staal and Barr 2012). Hence, an asthenospheric window created by processes such as rollback or eduction (reversed subduction and coherent exhumation of a portion of the subducted lithosphere during collision, Andersen et al. 1991; Duretz et al. 2012) following break-off of the down-going slab is thought to be more realistic.

Field data, numerical modelling and laboratory experiments provide some guidelines and constraints on the possible tectonic processes. The rheology of the lower crust, the characteristics of the subducting oceanic lithosphere (e.g. width and thermal structure) and of the attached passive margin (e.g. width and buoyancy) are important factors that impose a control on the nature of the slab break-off, degree of exhumation of deeply buried rocks and how the orogen evolved after the onset of collision (Sizova et al. 2012). The original architecture of the down-going Gander (passive) margin is poorly known, although seismic data (van der Velden et al. 2004) suggest it was wide and gradual rather than short and abrupt, which would enhance its ability to subduct (slab pull). On the other hand, the Tetagouche-Exploits backarc basin was relatively narrow (≤ 800 km), young (10–20 m.y.) and contained numerous fragmented slivers of felsic crust (van Staal 1994; van Staal et al. 2003, 2012), which would lower the overall slab pull and subduction velocity. A low slab pull results in relatively shallow subduction of the leading edge of the continental block, breakoff shortly after the onset of collision and modest exhumation (van Hunen and Allen 2011; Sizova et al. 2012), which would be consistent with what is observed in the Salinic orogen exposed in central Newfoundland. Salinic subduction-related high pressure metamorphic rocks, such as blueschist, are confined to narrow zones and are very rare. Where they occur they apparently returned by extrusion along a narrow subduction channel (van Staal et al. 2008). With a few exceptions (e.g. Cawood et al. 1994), deeply buried Salinic crustal metamorphic rocks in general are rarely exposed. The main evidence for deep subduction of Gander margin continental crust comes from shallow mantle reflectors, which were interpreted to image a fossil west-dipping Salinic A-subduction zone beneath central Newfoundland (van der Velden et al. 2004).

Both numerical modelling and laboratory experiments (Regard et al. 2008) generally indicate that after the onset of collision, the down-going plate progressively steepens at depth and may even roll back or reverse its dip above the break-off scar. Such a process could progressively open an asthenospheric window beneath the foreland (Fig. 8) and hence, may explain the observed west–east diachroneity of collision-related magmatism and potentially also a reversal in structural vergence from southeast- to northwest-directed tectonic transport (O'Brien 2003). However, the latter could also represent a change from Salinic to Acadian-related thrusting. More careful dating of structures is needed to test these options. Another process that could progressively open an asthenospheric window beneath the Salinic foreland is eduction of the leading edge of the Gander margin following break-off of the oceanic part of the slab (Fig. 8C). Modelling has shown that eduction is to a large extent dependent on the viscosity of the subduction channel and asthenosphere and generally leads to progressive flattening of the down-going slab above the break-off scar (Duretz et al. 2012) rather than widening the asthenospheric window. However, the mechanism controlling eduction may have been different in the Salinic collision zone, because of the unique tectonic setting of the Gander margin with the Acadian seaway subducting beneath its trailing edge concurrent with slab detachment along its leading edge. The former process may have imposed an additional extensional force due to slab roll-back in the Acadian seaway (van Staal et al. 2009, 2012), which may have triggered trenchward motion of Ganderia and hence, rapid eduction of the subducted part of the Gander margin at its leading edge (Fig. 8). Continued magmatism in the Mascarene-La Poile backarc basin until ca. 421 Ma indicates that the Coastal arc was extensional throughout its life span and attests to the viability of this process. The proposed tectonic model is attractive because any of slab break-off, eduction, roll-back and/or slab-dip reversal may have led to removal and/or significant thinning of the lithospheric mantle beneath the Gander margin at the end of the Silurian, immediately before the onset of the Acadian orogeny (Fig. 8C). Such a process could explain the generation of the Acadian magmatic suite exclusively in Ganderia and not elsewhere. In addition, both processes will lead to uplift and partial collapse of the Salinic orogen above the retreating and/or educting Gander margin slab. This could explain continuous shortening near the deformation front and the evidence for significant late Salinic crustal thinning in the core of the collision zone, both in central New Brunswick (D_3 of van Staal and de Roo 1995) and Newfoundland (D_3 of Zagorevski et al. 2007). The time of D_3 crustal thinning

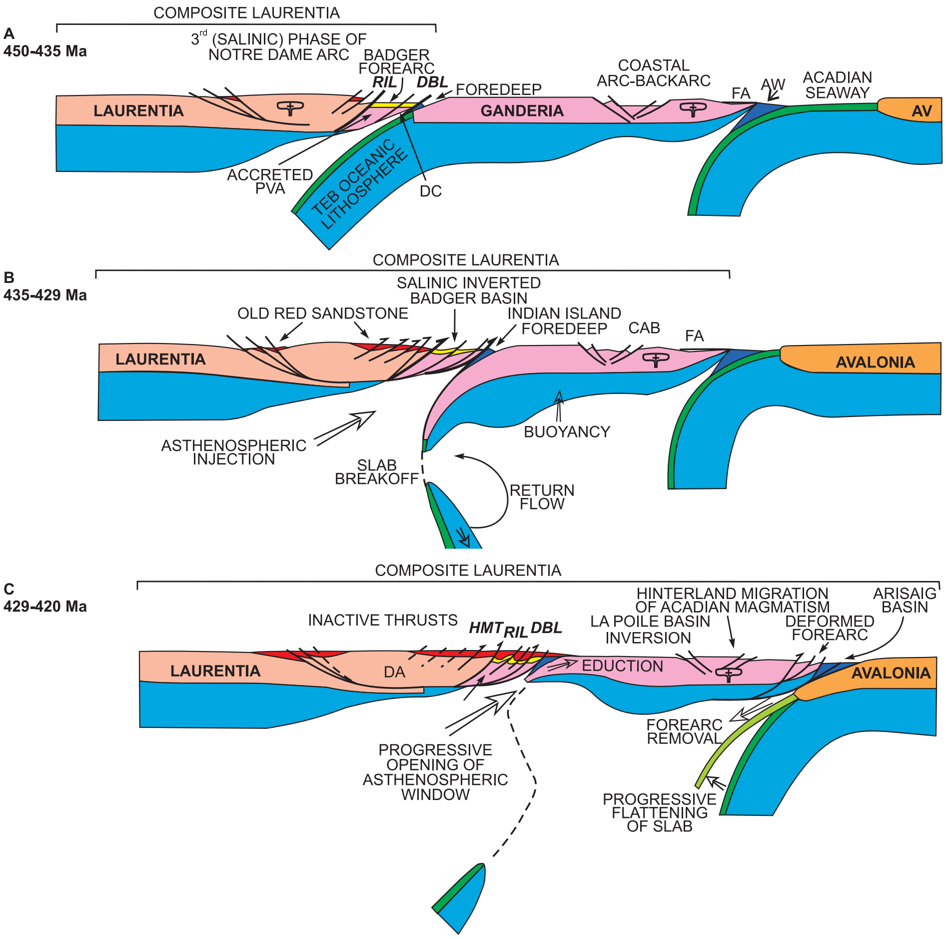

Figure 8. Tectonic model for Salinic and Acadian subduction and orogenesis. A. Late Ordovician–Early Silurian closure of the Tetagouche-Exploits backarc basin (TEB) and start of the Salinic collision of composite Laurentia with leading edge of Gander margin (see also Fig. 5). Waning Salinic-related arc magmatism took place to the west (3rd phase of Notre Dame arc of van Staal et al. 2007) of the Red Indian Line (RIL). Salinic collision was coeval with and hence, may have been the cause of initiation of subduction in the Acadian seaway beneath the trailing edge of Gander margin, forming the Coastal arc–backarc system (CAB). Red bed molasse sedimentation (shown in red) initiated in basins that formed during early Salinic thrusting and uplift in the hinterland of orogen. Marine sedimentation took place in the Badger forearc basin. The Badger forearc progressively grew eastwards due to accretion of buoyant blocks such as the Duder Complex (DC), which formed in the TEB. A marine foredeep accumulated Badger-like clastic sedimentary rocks (e.g. Outflow Formation, Currie 1997) and sedimentary rocks of the younger Indian Island Group on the leading edge of the Gander margin of Ganderia. B. Progressive steepening of the Gander margin slab during and/or following tearing (break-off) of the TEB oceanic slab. Red bed sedimentation reached the Red Indian Line (RIL), while the adjacent marine Badger forearc basin was inverted. DBL: Dog Bay Line; C. Eduction of the Gander margin and eastward expansion of an asthenospheric window beneath the Salinic orogen. Red bed sedimentation had now overstepped onto all parts Ganderia. Start of Acadian collision at ca. 420 Ma at the trailing edge of the Gander margin (now the leading edge of composite Laurentia) due to the entrance of Avalonia at the trench. Progressive flattening of the down-going slab following the start of collision produced a hinterland-migrating retro-arc deformation zone and terminated in expulsion of the mantle wedge and extinction of Acadian magmatism during the Emsian. AW: accretionary wedge; DA: Dashwoods; FA: forearc; PVA: Popelogan-Victoria arc.

in central New Brunswick is constrained by circumstantial evidence to the Early Devonian (van Staal and de Roo 1995; van Staal et al. 2008), but in central Newfoundland took place between 426 and 417 Ma.

The Tectonic Setting of Early Devonian Acadian Magmatism

The two tectonic processes, slab steepening and eduction, potentially associated with break-off of the Salinic slab discussed above, would have created and progressively widened an asthenospheric window beneath the foreland of the Salinic orogen. These processes would not only have allowed Salinic slab break-off magmatism to migrate towards the Salinic foreland, but they would also have made this area a potentially magmatically fertile, hot and hence, weak (Hyndman et al. 2005) backarc region of the progressively growing Acadian orogen during the Early Devonian. The Acadian collision had started at ca. 420 Ma along Ganderia's trailing edge. Continued underthrusting of Avalonia's leading edge (A-subduction) beneath this region could have caused melting and generation of arc-like mafic melts in the mantle wedge trapped above a progressively flattening and dehydrating Avalonian slab (Fig. 5), which in turn may have partially melted the overlying crustal rocks. In particular, underplated forearc sedimentary rocks of the Coastal arc may have been a fertile source for the large bloom of Acadian granitoid plutons in Ganderia (van Staal et al. 2009, see above).

The resultant voluminous bimodal Acadian suite has mixed arc to non-arc geochemical and isotopic signatures (van Staal et al. 2009). The western extent of the Acadian magmatic suite closely mirrors the position of the Acadian structural–metamorphic front in central Newfoundland (van Staal and Zagorevski 2012). This front is defined by the leading edge of the west-directed Acadian Meelpaeg metamorphic nappe (van der Velden et al. 2004; Valverde-Vaquero et al. 2006), which locally overrode unmetamorphosed grey Emsian (408–394 Ma) sandstone and mudstone (Chandler 1982) for a short distance, as shown by local overturning and folding, beyond which these sedimentary rocks generally remained undeformed and lie subhorizontally (van Staal et al. 2005). The most spectacular evidence of this structural juxtaposition is shown by tightly folded and cleaved Emsian sandstone of the Billiards Brook formation of Chorlton (1980). This unit is situated structurally directly beneath the Meelpaeg nappe in the Cape Ray-Gunflap Hills fault zone along the upper reaches of the La Poile River in southwestern Newfoundland (Lin et al. 1994). Here the Meelpaeg nappe overrode and buried the Red Indian Line, whereas farther northeast it coincides (or nearly so) with this structure (van Staal et al. 2005; Valverde-Vaquero et al. 2006). Acadian deformation locally occurred farther west and even affected parts of the Humber margin (Fig. 1), but deformation is typically heterogeneous and not associated with penetrative metamorphism (van Staal and Zagorevski 2012). This suggests there is a close relationship between migration of the Acadian front and the Acadian magmatism, a relationship well established earlier in Maine by Bradley and Tucker (2002).

CONCLUSIONS

The red beds of the Botwood Group are Upper Silurian (mainly Pridolian), but may have extended into the Lower Devonian (Lochkovian). They form part of the Old Red Sandstone facies and were deposited unconformably on early Salinic-deformed (D_1) Katian–Llandoverian marine turbidite sequences of the Badger Group. This Middle to Late Silurian unconformity marks the main phase of the Salinic collision. The red beds overstepped the deformation zone associated with the Dog Bay Line during the Late Silurian, heralding the emergence of the Gander margin above sea level. Deposition of the Botwood Group overlapped with bimodal magmatism, related to break-off of the west-dipping slab of Tetagouche-Exploits backarc oceanic lithosphere attached to the Gander margin (van Staal et al. 2009), and the final phase of Salinic deformation.

Our new data, integrated with the existing geochronological database pertinent to the tectonic evolution of Silurian and Early Devonian rocks in Newfoundland, suggest that the deformation front associated with the final phase of Salinic tectonism, associated deposition of the Old Red Sandstone and slab break-off magmatism were time-transgressive from west to east. They date the progressive growth of the Salinic mountains and the time the sequentially accreted terranes became fully emerged.

Steepening and rollback of the downgoing slab possibly accompanied by a slab dip reversal of its broken-off tip (Regard et al. 2008) and eduction of the Gander margin (Fig. 8), progressively opened an asthenospheric window beneath the Salinic foreland, which led to time-transgressive uplift, sedimentation of Old Red Sandstone molasse, and magmatism near an east-migrating deformation front. Eduction of the Gander margin may have been triggered by slab rollback of the northwest-dipping oceanic slab of the Acadian seaway attached to Avalonia, which was subducting beneath the trailing edge of Ganderia, while the Salinic orogeny was taking place at its leading edge. The cause of the continuity of contraction near the deformation front for a short period after slab detachment and the onset of eduction in northeast Newfoundland may be due to compressive stresses generated in the upper plate following slab detachment and progressive uplift of the overlying Salinic collisional wedge (e.g. Price and Audley-Charles 1987; Buiter et al. 2002). Thinning and heating of the lithosphere beneath the leading edge of the Salinic mountain belt set the stage for the Acadian orogenesis which was superimposed on the Salinic-deformed rocks shortly thereafter during the Early Devonian. Fluids released during dehydration of the down-going Avalonian slab caused large-scale melting in the extensive mantle wedge trapped between it and the overlying, thin edge of composite Laurentia. Ponding of mafic arc-like melt bodies in the lower crust may have produced voluminous felsic magmatism. Progressive flattening of the Avalonian slab over time (Fig. 8C) led to a hinterland-propagating (northwest) retro-arc foreland fold and thrust belt, expulsion and cooling of the trapped mantle wedge and hence termination of syn-Acadian magmatism.

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