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RAPID TRANSITION FROM THE SALINIC TO ACADIAN OROGENIC CYCLES IN THE NORTHERN APPALACHIAN OROGEN: EVIDENCE FROM NORTHERN NEW BRUNSWICK, CANADA

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ABSTRACT. Geochronological data from volcanic rocks aid in reconstructing the Silurian–Devonian evolution of the northern Appalachians of New Brunswick in the context of Salinic (Silurian) and Acadian (Devonian) orogenesis. Late Silurian to Early Devonian sedimentation, volcanism and deformation in northern New Brunswick is complex, and characterized by transgressive-regressive sedimentary cycles, local disconformities and angular unconformities, and two stages (*ca.* 422–419 Ma, and *ca.* 417–407 Ma) of typically bimodal, within-plate magmatism. These events overlapped the end of the Salinic orogenic cycle, related to the *ca.* 430 Ma collision of the Ganderian passive margin with composite Laurentia, and the beginning of the Acadian cycle, which is associated with subsequent (*ca.* 424–422 Ma) collision and underthrusting of Avalonian crust beneath composite Laurentia (now including the Gander margin).

Following Ganderia–Laurentia collision, the earlier (Pridolian) stage of within-plate magmatism is interpreted to result from *ca.* 425 Ma detachment of the subducting Tetagouche backarc lithospheric slab and is linked to the Salinic cycle, along with local, pre-Acadian (Salinic C) deformation of Ludlow–Pridoli rocks. Break-off of the Tetagouche slab also led to rapid uplift and subsequent extensional collapse of the Salinic metamorphic core, resulting in latest Silurian-earliest Devonian D_3 vertical shortening and flat belt formation in the Brunswick subduction complex.

Precise ID-TIMS U-Pb (zircon) dating of volcanic rocks reveals the existence of a 2.2 myr hiatus in the stratigraphic record, corresponding to an interpreted latest Silurian–earliest Devonian erosional surface between volcanic rocks identified with the Salinic cycle (Dickie Cove Group and lower part of Tobique Group) and those associated with the Acadian cycle (Dalhousie Group). The disconformity coincides with a widespread marine regression recorded by roughly coeval deposition of shallow-water to intertidal redbeds across the northern Appalachians. Marine regression is closely followed by deposition of Lochkovian deep marine sedimentary rocks (turbidites) in the Seboomook foredeep. This regressive-transgressive sequence immediately precedes arrival of the northwest-migrating Acadian deformation front, and is interpreted as evidence of a migrating Acadian forebulge-foredeep system associated with flexural loading of Ganderian crust by the Acadian orogenic wedge situated farther southeast in southern New Brunswick and coastal Maine. Early Devonian volcanism and Acadian retro-arc foreland deformation are linked to flat-slab convergence of Avalonia after underthrusting of buoyant Avalonian continental crust began *ca.* 418 Ma. It is proposed that the latest Silurian–earliest Devonian disconformity and marine regression mark the beginning of the Acadian orogenic cycle, implying that, in northern New Brunswick, the effects of waning, late Silurian (Salinic) deformation temporally overlapped the early, far-field effects of Acadian orogenesis.

Key words: Salinic orogenic cycle, Acadian orogenic cycle, Ganderia, regressive-transgressive sequence, within-plate magmatism, U-Pb geochronology, slab breakoff, retro-arc foreland, flat subduction

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INTRODUCTION

In the Ganderian core of the northern Appalachians (fig. 1), separation of structures formed during the predominantly Silurian Salinic orogenic cycle from those that formed during the latest Silurian—Early Devonian Acadian orogenic cycle can be problematic if these structures are defined solely on the basis of their absolute ages (van Staal and others, 2008, 2009; Wilson and Kamo, 2012). Analysis of Ganderia's tectonic evolution (van Staal and others, 2009, 2014) has revealed that while its leading segment was in the waning stages of Salinic collision with composite Laurentia between 423 and 418 Ma, subduction beneath its trailing segment in southern New Brunswick, coastal Maine and southern Newfoundland had nearly closed the Acadian seaway (fig. 1). Closure of this oceanic tract culminated in accretion of Avalonia to composite Laurentia *ca.* 424 to 422 Ma, and southeast to northwest diachronous deformation until the Middle to Late Devonian (Donohoe and Pajari, 1973; Bradley and others, 2000). The onset of the Acadian orogeny is historically defined by late Silurian (422–420 Ma) compressive deformation in southernmost New Brunswick, which is mainly manifested by inversion of the Mascarene backarc basin (Robinson and others, 1998; van Staal and others, 2009, 2014) and coincident termination of Coastal Arc magmatism (Barr and others, 2002; Piñán Llamas and Hepburn, 2013) (fig. 1). In northern New Brunswick, the end of the Salinic cycle and onset of Acadian deformation in the hinterland of the Acadian collision zone have not been well constrained, although Acadian deformation was known to have started after deposition of Lower Devonian turbidites (St. Peter and Boucot, 1981; Bradley and Tucker, 2002). However, new U-Pb (zircon) ID-TIMS ages for upper Silurian—Lower Devonian volcanic rocks and their integration with a well-defined sedimentary history establishes tight controls on the nature and relationships of the Salinic and Acadian orogenic cycles. Our results have major implications for the tectonic interpretations of correlative magmatic and sedimentary rocks elsewhere in the northern Appalachians. In this contribution we present five new U-Pb (zircon) ages that improve our understanding of the late Silurian–Early Devonian evolution of the northern Appalachians, establish important constraints on the tectonic setting and genesis of Silurian–Devonian volcanic rocks, and provide a guide for separating the effects of the Salinic and Acadian orogenies. In particular, we explore the significance of a late Pridolian–early Lochkovian disconformity and coeval marine regression in defining the end of the Salinic orogenic cycle and the beginning of the Acadian cycle.

REGIONAL AND TECTONIC SETTING: CAMBRIAN TO EARLY SILURIAN

The focus of this paper is the critical late Silurian–Early Devonian interval in the northern Appalachians. In this section, the depositional and structural events that preceded this interval are summarized to establish the background for ensuing discussions. To help clarify the regional nomenclature of the diverse tectonostratigraphic elements of northern New Brunswick, relevant data pertaining to Lower Paleozoic inliers and Middle Paleozoic successor basin rocks are presented in table 1. Note that, as formal Early/Late Silurian and Llandoverian do not exist in recent International Commission on Stratigraphy timescales (for example, Cohen and others, 2013), the informal early/late and upper/lower Silurian are used throughout. Additionally, in most cases, the more familiar epoch names for the Silurian (Llandovery, Wenlock, *et cetera*) are employed as opposed to more recently introduced age names, unless reference to a more specific time interval is warranted. Epoch and age divisions are both illustrated on a stratigraphic correlation chart for northern New Brunswick (fig. 2).

Upper Silurian to Lower Devonian volcanic rocks in northern and western New Brunswick underlie the eastern part of a broad Middle Paleozoic successor basin known as the Matapedia Cover Sequence (Fyffe and Fricker, 1987; van Staal and de

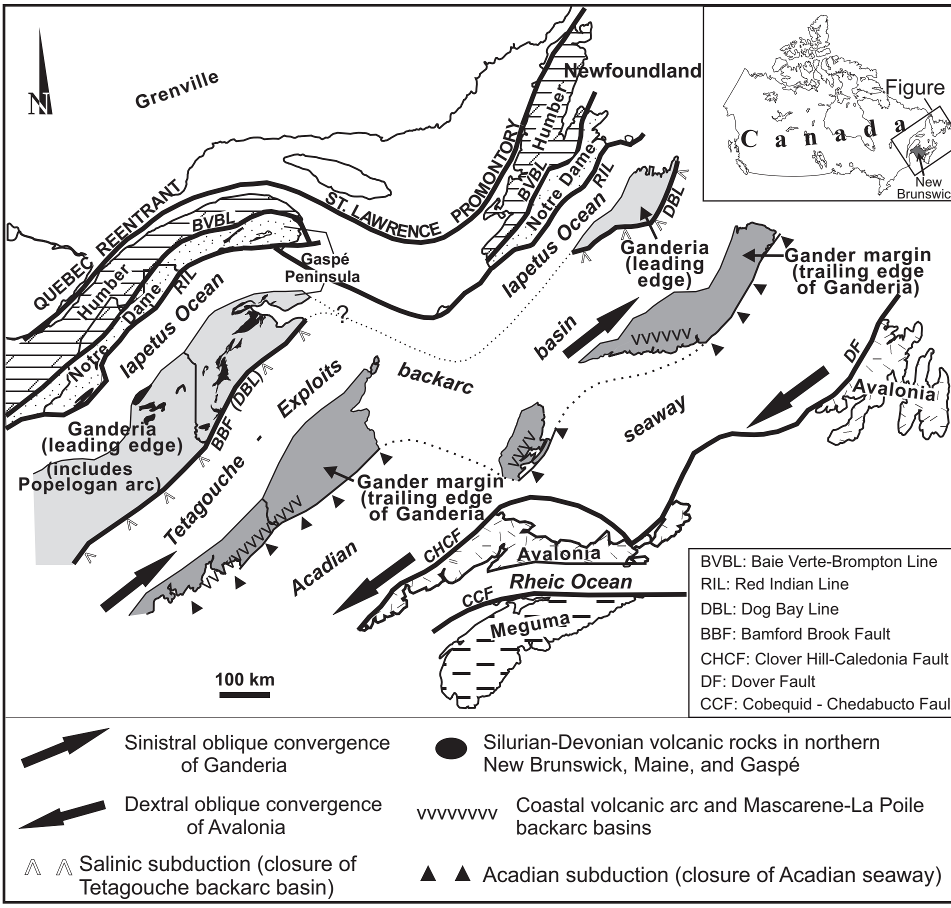

Fig. 1. Schematic illustration of the main tectonic elements involved in the accretionary history of the northern Appalachians (modified from van Staal and Barr, 2012). Also shown is the distribution of Silurian–Devonian volcanic rocks, and the location of the Salinic and Acadian subduction zones (subduction in direction of arrows). No scale is implied by the amount of separation shown between respective terranes. The nature and timing of the various phases of oblique convergence are at present still poorly constrained.

Roo, 1995), which extends from the eastern Gaspé Peninsula of Quebec, southwest through northern New Brunswick into northern and central Maine. Rocks of the Matapedia Cover Sequence (MCS) largely obscure remnants of the *ca.* 478 to 455 Ma Popelogan Arc that are presumed to lie at depth between known inliers in northern Maine and northern New Brunswick. In New Brunswick, mafic and minor felsic volcanic rocks of the Popelogan Arc are exposed in the Popelogan inlier south of Campbellton (Wilson, 2003) (figs. 3 and 4) and in a series of very small inliers southwest of the Popelogan inlier. The Popelogan Arc formed on Ganderian continental crust (van Staal and others, 2016) above a subduction zone that dipped to the southeast (present coordinates) beneath Ganderia during Early Paleozoic closure of the Iapetus Ocean (van Staal, 1994; van Staal and others, 1998, 2003, 2016). Ganderia is a Late Neoproterozoic–Early Cambrian arc terrane that rifted from the Amazonian margin of Gondwana (van Staal and others, 1996, 2012) and forms the basement of most of New Brunswick (fig. 1). The Popelogan Arc was extensional throughout much

Table 1

Summary of Early and Middle Paleozoic tectonostratigraphic elements of Ganderia in northern New Brunswick

Ganderian inliers

Inlier	Depositional setting	Upper-rank units	Age	Deformation
Miramichi	Ganderian passive margin*	Miramichi Group	Cambrian-Early Ordovician	Salinic A,B,C, Acadian B (D1 to D4)
	Tetagouche ensialic backarc*	Bathurst Supergroup	Middle to Late Ordovician	Salinic A,B,C, Acadian B (D1 to D4)
	Tetagouche ensimatic backarc*	Fournier Supergroup	Middle to Late Ordovician	Salinic A,B,C, Acadian B (D1 to D4)
Elmtree	Tetagouche ensimatic backarc*	Fournier Supergroup	Middle to Late Ordovician	Salinic A,B,C, Acadian B (D1 to D4)
Popelogan	Popelogan arc	Balmoral Group	Middle to Late Ordovician	Weak Salinic B & Acadian B

Matapedia Cover Sequence

Geographic Division	Depositional setting	Upper-rank units	Age	Deformation
Aroostook-Percé Anticlinorium	Matapedia forearc	Grog Brook, Matapedia, and Quinn Point groups	Late Ordovician-Early Silurian	Salinic B, Acadian B
Chaleur Bay Synclinorium: Chaleur zone		Dickie Cove & Petit Rocher groups	Late Silurian	Salinic C, Acadian A & B
	Acadian retroarc foreland	Dalhousie Group	Early Devonian	Acadian B
Chaleur Bay Synclinorium: Tobique zone		Petit Rocher Group; lower Tobique Group	Late Silurian	Salinic C, Acadian A & B
	Acadian retroarc foreland	upper Tobique Group	Early Devonian	Acadian B
Connecticut Valley-Gaspé Synclinorium	Acadian retroarc foreland	Fortin & Seboomook groups	Early Devonian	Acadian B

* Units amalgamated in the Brunswick Subduction Complex.

See text for discussion of the various stages of Salinic and Acadian deformation. Intense polyphase deformation (D_1 to D_4 of van Staal and de Roo, 1995) is present in the Miramichi and Elmtree inliers, which consist of rocks incorporated in the Brunswick subduction complex; this contrasts sharply with weak deformation in the Popelogan inlier, which occupied the upper plate during development of the accretionary wedge. In the Matapedia Cover Sequence, early Silurian rocks are affected by the *ca.* 430–425 Ma Salinic B deformation, whereas the effects of Salinic C and Acadian A events are locally observed in late Silurian and younger rocks. Acadian B is the main phase of compressive Acadian deformation and affects all rocks in all areas.

of its history; arc rifting (in progress by at least *ca.* 470 Ma) led to opening of the Tetagouche backarc basin and culminated in seafloor spreading (van Staal and others, 2003, 2016), separating Ganderia into an inboard, leading segment that includes the ensialic Popelogan Arc, and an outboard, trailing segment referred to as the Gander (passive) margin (fig. 1). Late Ordovician (~455 Ma) collision of the Popelogan Arc (Ganderia's leading segment) with composite Laurentia (van Staal, 1994; van Staal and others, 2003, 2009) along the Red Indian Line (fig. 1) marked the closure of the main tract of Iapetus and the end of the Taconic orogenic cycle (fig. 2). Subsequent convergence of the Gander margin was achieved by Late Ordovician to mid-Silurian

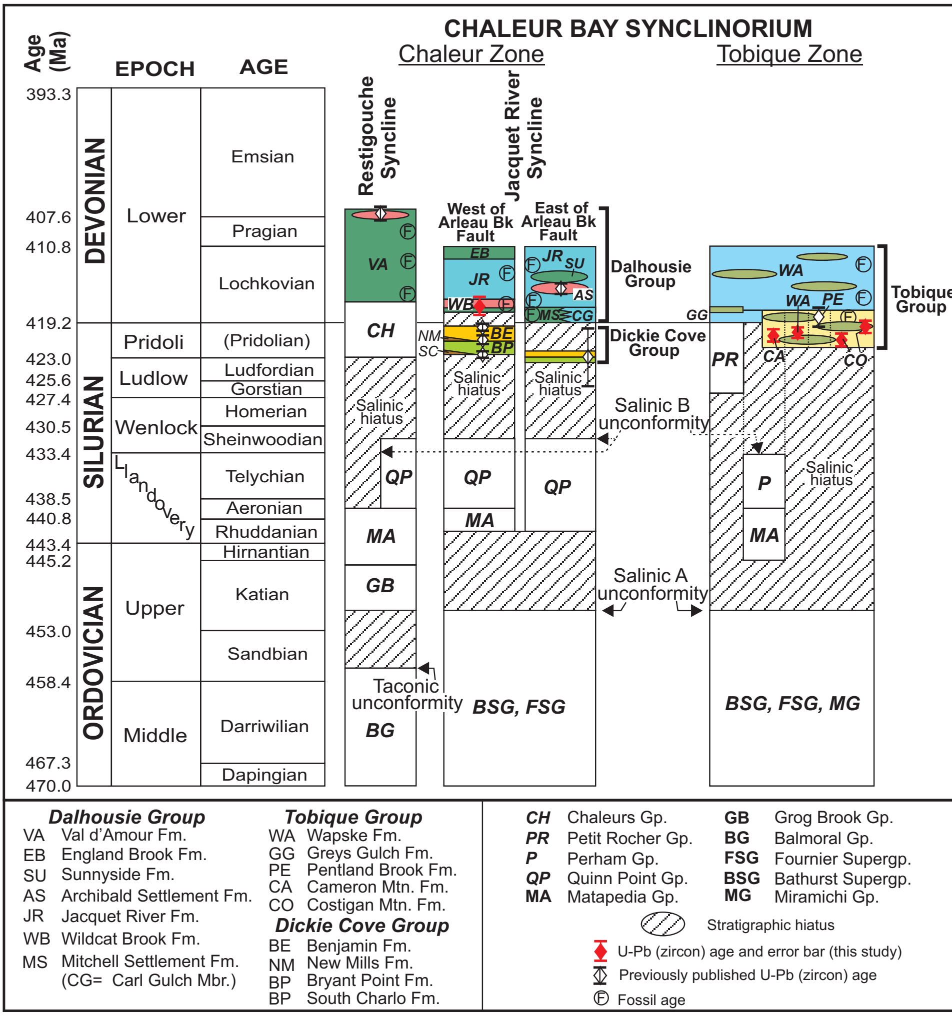

Fig. 2. Stratigraphic columns showing the age and correlation of upper Silurian and Lower Devonian volcanic rocks of the Chaleur Bay Synclinorium. Timescale is that of Cohen and others (2013). See figure 3 for references to previously published radiometric data. The Salinic orogenic cycle starts *ca.* 450 Ma, shortly after termination of the Taconic cycle *ca.* 455 Ma due to the docking of the Popelogan arc to composite Laurentia (Taconic unconformity). The Salinic A unconformity is defined as the erosional surface above exhumed rocks of the Brunswick subduction complex (Bathurst and Fournier Supergroups and Miramichi Group), regardless of the age of overlying rocks. The Salinic B unconformity separates lower Silurian from upper Silurian rocks and marks uplift and deformation associated with collision of the Gander margin with composite Laurentia. See figure 7 and text (Discussion) for additional detail on late Silurian–Early Devonian stratigraphy, Salinic C deformation, and the Acadian A disconformity/marine regression.

(*ca.* 450–430 Ma) northwesterly subduction of Tetagouche backarc basin lithosphere beneath the active Laurentian margin, leading to formation of the Brunswick subduction complex (van Staal, 1994; van Staal and others, 2003, 2009). The Brunswick subduction complex encompasses rocks in the Miramichi and Elmtree inliers, including the Miramichi Group, Bathurst and Fournier Supergroups (table 1). Thrusts, folds, and associated foliations that developed in the subduction complex during this period constitute the D_1 deformation seen in Lower Paleozoic rocks. B subduction of Tetagouche

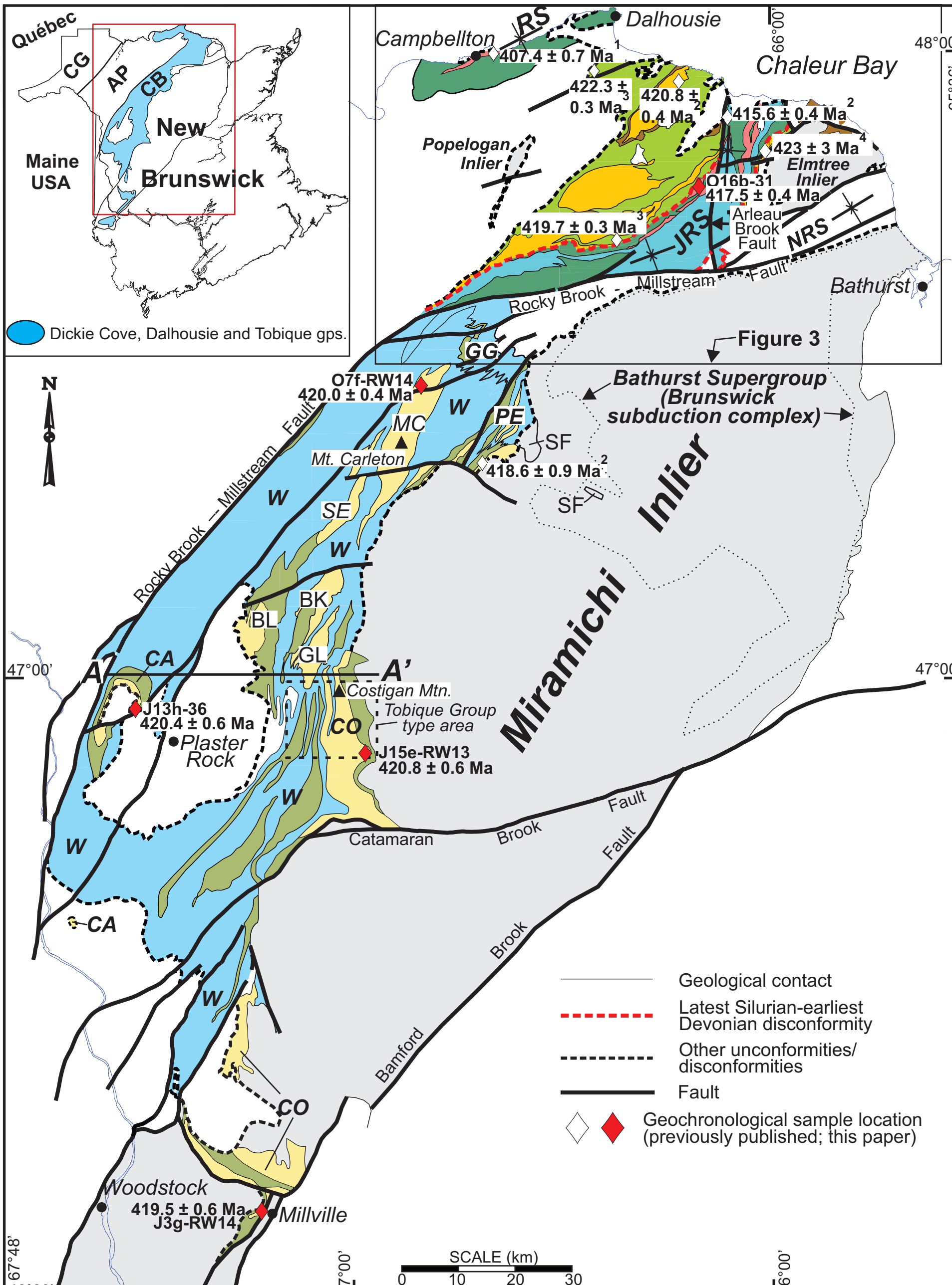

Fig. 3. (continued).

backarc oceanic lithosphere ended with *ca.* 430 Ma collision of the Gander margin with composite Laurentia (Salinic orogeny), which resulted in D_2 deformation in the Brunswick subduction complex and Salinic B orogenesis (Wilson and Kamo, 2012) in the MCS (table 1). Ganderia–Laurentia collision is also recorded by inversion of the Fredericton

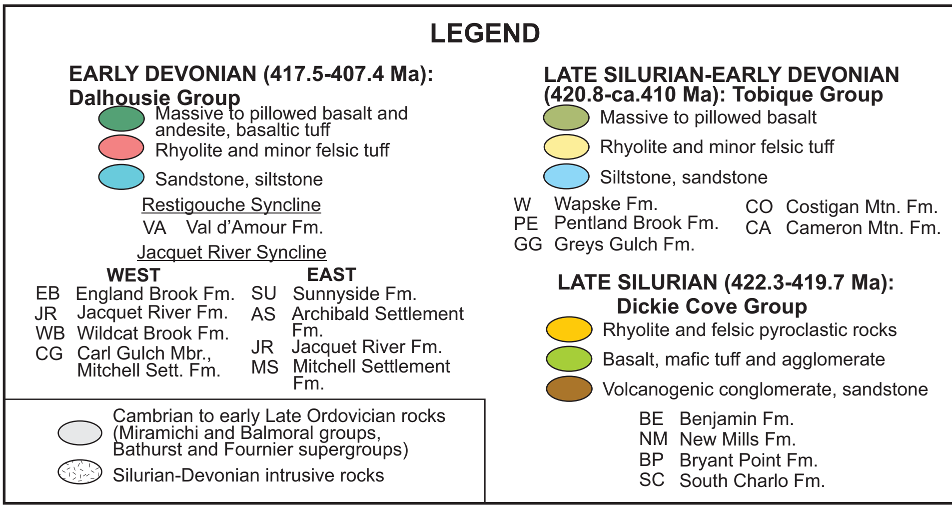

Fig. 3. Simplified geology of the Chaleur Bay Synclinorium, showing the locations of samples collected for CA-ID-TIMS dating (including those published previously). Numerical superscripts on published ages refer to data sources: 1 – Wilson and others (2005); 2 – Wilson and Kamo (2008); 3 – Wilson and Kamo (2012); 4 – Walker and others (1993). The Chaleur zone is located north of the Rocky Brook–Millstream Fault and the Tobique zone is south of the fault. Inliers of Cambrian to Upper Ordovician rocks (Miramichi Group, Bathurst and Fournier Supergroups, and Balmoral Group) are shown in light gray. Uncolored areas are underlain by miscellaneous sedimentary cover rocks ranging in age from Late Ordovician to Early Carboniferous. Abbreviations: RS = Restigouche Syncline; JRS = Jacquet River Syncline; NRS = Nigadoo River Syncline; SF = Simpsons Field Formation. Volcanic edifices of the Wapske Formation in the central Tobique basin: MC = Mount Carleton; SE = Serpentine Mountain; BK = Black Mountains; BL = Blue Mountains; GL = Gulquac Mountains. The inset at upper left illustrates the three subdivisions of the Matapedia Cover Sequence: CB = Chaleur Bay Synclinorium; AP = Aroostook – Percé Anticlinorium; CG = Connecticut Valley – Gaspé Synclinorium. Cross-section A-A' is shown in figure 6.

foredeep basin *ca.* 426 to 422 Ma (West and others, 1992; Park and Whitehead 2003; van Staal and Barr, 2012).

The stratigraphic framework and evolution of the MCS in northern New Brunswick has been described by Wilson and others (2004). It is convenient to regard the MCS as consisting of three domains, namely, from northwest to southeast, the Connecticut Valley–Gaspé Synclinorium, Aroostook–Percé Anticlinorium, and Chaleur Bay Synclinorium (Rodgers, 1970) (fig. 3 inset and table 1). The oldest rocks are preserved in the Aroostook–Percé Anticlinorium, where Late Ordovician to *ca.* mid-Wenlock sedimentation occurred in a forearc setting with respect to sparse arc magmatism related to subduction of Tetagouche backarc oceanic lithosphere (Wilson and others, 2008; van Staal and others, 2009; Wilson and Kamo, 2012). The forearc basin comprises a shallowing-upward sequence of Upper Ordovician to lower Silurian siliciclastic and calcareous turbidites that culminated with deposition of upper Llandovery–Wenlock shallow marine carbonates (Wilson and others, 2004). The carbonate rocks immediately underlie a widespread erosional surface (Salinic B unconformity of Wilson and Kamo, 2012) that marks a Wenlock−Ludlow hiatus of 5 to 8 myr duration (fig. 2). The Salinic B unconformity records a major phase of deformation, exhumation and erosion of lower Silurian rocks that is attributed to the Gander margin−composite Laurentia collision and attempted subduction of buoyant Gander margin continental crust (van Staal, 1994; van Staal and others, 2003, 2009). Shutdown of Salinic arc magmatism roughly coincides with the 429.2 ± 0.5 Ma age of felsic tuff in the upper part of the forearc sequence (Wilson and others, 2008), and the onset of collision-related deformation and metamorphism, which continued until at

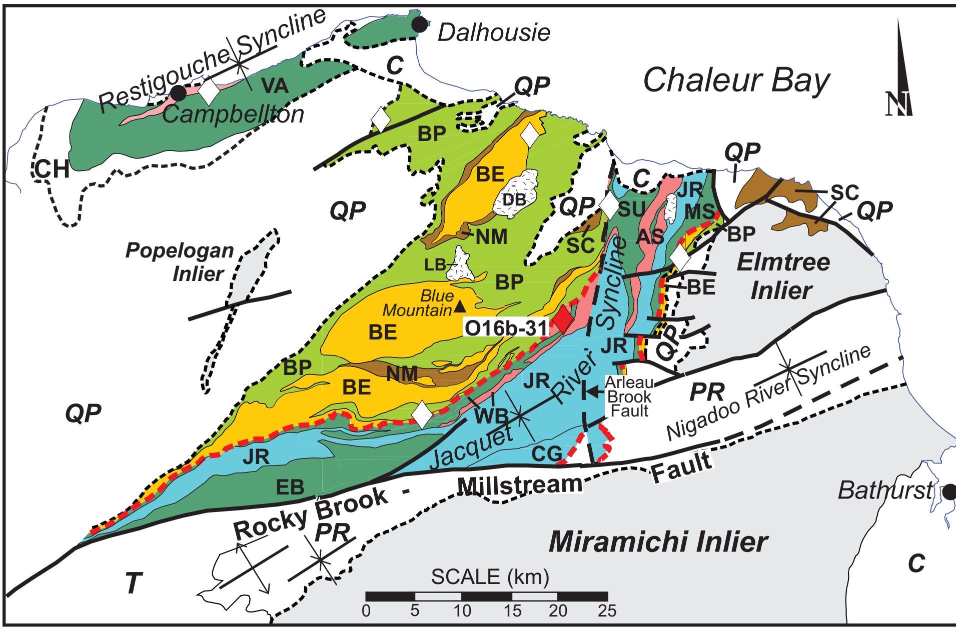

Fig. 4. Distribution of constituent formations of the Dickie Cove and Dalhousie Groups in the Chaleur zone of the Chaleur Bay Synclinorium (see fig. 3 for location). Legend as in figure 3. Abbreviations: C = Carboniferous rocks; CH = Chaleurs Group; PR = Petit Rocher Group; QP = Quinn Point Group; T = Tobique Group. LB (Landry Brook Quartz Monzonite) and DB (Dickie Brook Quartz Monzonite) are composite mafic-felsic plutons that are coeval with Dickie Cove volcanism and have yielded ages of 419.6 ± 0.2 Ma and 418.1 ± 1.3 Ma, respectively (Pilote and others, 2012).

least Pridolian time (van Staal and others, 2008; Wilson and Kamo, 2012). On the western flank of the Aroostook−Percé Anticlinorium, the Connecticut Valley−Gaspé Synclinorium is underlain mainly by Lower Devonian rocks assigned to the Fortin Group in New Brunswick and Quebec, and the Seboomook Group in northern Maine. On the east side of the Aroostook−Percé Anticlinorium, rocks of the Chaleur Bay Synclinorium disconformably overlie (or are locally faulted against) those of the Aroostook–Percé Anticlinorium; they also rest unconformably on Cambrian to lower Silurian rocks within and peripheral to the Miramichi and Elmtree inliers (figs. 2–4; table 1). Upper Silurian and Lower Devonian rocks of the Chaleur Bay Synclinorium are described in detail in the following section.

Silurian-Devonian stratigraphy of the Chaleur Bay Synclinorium

The Chaleur Bay Synclinorium displays much greater stratigraphic and structural complexity than the other two domains in the MCS, manifested in transgressive-regressive sedimentary cycles, unconformities, at least two phases of deformation, and widespread volcanism. The Chaleur Bay Synclinorium is divided into two parts by the Rocky Brook–Millstream Fault – the Chaleur zone to the north and the Tobique zone to the south (figs. 3 and 4; table 1). The Chaleur zone is underlain by upper Silurian (Ludlow–Pridoli) rocks of the Dickie Cove and Petit Rocher Groups, Pridoli to Lower Devonian rocks of the Chaleurs Group, and Lower Devonian rocks of the Dalhousie Group (figs. 3 and 4). The Tobique zone is underlain almost entirely by Pridoli to Lower Devonian rocks of the Tobique Group; however, the northern part of the zone (adjacent to the Rocky Brook–Millstream Fault) is underlain, in part, by the Petit

Rocher Group, representing a southwesterly extension of Petit Rocher stratigraphy exposed in the Nigadoo River Syncline north of the fault (figs. 2–4). The volcanic rocks discussed herein belong to the Dickie Cove and Dalhousie Groups north of the Rocky Brook–Millstream Fault, and the Tobique Group south of the fault. Volcanic rocks dominate the Dickie Cove Group, the lower part of the Tobique Group, and the Dalhousie Group in the Restigouche Syncline, whereas sedimentary rocks are dominant and volcanic rocks subordinate in the upper part of the Tobique Group and the Dalhousie Group in the Jacquet River Syncline (figs. 2–4).

North of Rocky Brook–Millstream Fault

The oldest rocks that post-date the Salinic B unconformity in the Chaleur Bay Synclinorium are assigned to the Petit Rocher Group, which comprises sedimentary rocks exposed in the Nigadoo River Syncline (figs. 3 and 4). The Petit Rocher Group records a transgressive sequence of mainly coarse-grained, shallow water to terrestrial sedimentary rocks (Simpsons Field Formation), overlain by limestone and associated calcareous sedimentary rocks (LaPlante Formation), and marine siliciclastic rocks (Free Grant Formation). Late Silurian exhumation and erosion of Ordovician and lower Silurian rocks exposed within and peripheral to the Miramichi and Elmtree inliers is reflected in the clast composition of Simpsons Field conglomerates deposited along the inlier margins. No rocks younger than the latest Silurian Free Grant Formation are preserved in the Nigadoo River Syncline, but in the western extension of the Petit Rocher Group south of the Rocky Brook – Millstream Fault, the Free Grant Formation is conformably overlain by red siltstone of the Greys Gulch Formation at the base of the Tobique Group (see below). Ludlow−Pridoli fossils in the Petit Rocher Group indicate that deposition of these sedimentary rocks was, in part, coeval with eruption of subaerial volcanic rocks of the Dickie Cove Group farther west (figs. 2 and 4).

The Dickie Cove Group is mainly exposed on the west limb of the Jacquet River Syncline and comprises, from oldest to youngest, the South Charlo, Bryant Point, New Mills and Benjamin Formations (figs. 2 and 4). The Dickie Cove Group disconformably overlies lower Silurian rocks of the Quinn Point Group (fig. 2), and in the type area, its age is tightly constrained between 422.3 ± 0.3 (rhyolite interbedded with basalt at the base of the Bryant Point Formation) and 419.7 ± 0.3 Ma (rhyolite at the top of the Benjamin Formation) (Wilson and Kamo, 2008, 2012). On the east limb of the Jacquet River Syncline, the Benjamin Formation has yielded an age of 423 ± 3 Ma (Walker and others, 1993), suggesting that these volcanic rocks may be somewhat older than those on the west limb, although the error bars overlap (fig. 2).

The Dalhousie Group is exposed in two areas − the Jacquet River Syncline and the Restigouche Syncline (figs. 3 and 4) − that show marked contrasts in lithology, depositional setting, and volcanic geochemistry (Wilson and others, 2005; Wilson and Kamo, 2008). Furthermore, stratigraphy in the western part of the Jacquet River Syncline (west of the Arleau Brook Fault) differs from that east of the fault (figs. 2 and 4). East of the Arleau Brook Fault, the stratigraphic succession comprises, in ascending order, the Mitchell Settlement Formation (massive subaerial basalt flows interbedded with subordinate red siltstone), Jacquet River Formation (marine sedimentary rocks), and mainly subaerial volcanic rocks of the Archibald Settlement Formation (flow-layered rhyolite and local felsic pyroclastic rocks) and Sunnyside Formation (massive to well-bedded mafic ash and lapilli tuff, and minor pillow basalt). Volcanic rocks of the Archibald Settlement and Sunnyside Formations form thick lenticular masses that are overlain and underlain by sedimentary rocks of the Jacquet River Formation (figs. 2 and 4). West of the Arleau Brook Fault, the Dalhousie Group consists of the Wildcat Brook Formation (subaerial basalt, mafic pyroclastic rocks, and flow-layered rhyolite), Jacquet River Formation (marine sedimentary rocks), and England Brook Formation (massive subaerial basalt, and minor mafic tuff and terrestrial sedimentary rocks), in ascending stratigraphic order (figs. 2 and 4). A thick unit of red siltstone-mudstone underlying the Jacquet River Formation adjacent to the Rocky Brook–Millstream Fault is correlated with red siltstone east of the Arleau Brook Fault and assigned member status (Carl Gulch Member; figs. 2 and 4) in the Mitchell Settlement Formation. The lithology and sedimentology of the Carl Gulch Member is consistent with very shallow water, estuarine and/or intertidal deposition. Hence, the Dalhousie Group in the Jacquet River Syncline records an Early Devonian marine transgression that followed a short-lived, early Lochkovian regression represented by the Mitchell Settlement redbeds.

On the northwest limb of the Jacquet River Syncline, the contact between the Dickie Cove Group and overlying Dalhousie Group is marked by thin, discontinuous or sporadically exposed units of polymictic conglomerate and red mudstone (in the west), and limestone and fossiliferous siltstone (farther northeast); these beds were interpreted by Wilson and Kamo (2008) to mark a disconformity between the Benjamin Formation (Dickie Cove Group) and the Wildcat Brook Formation (basal Dalhousie Group; figs. 2–4). In all parts of the Jacquet River Syncline, brachiopod and spore assemblages indicate that the Dalhousie Group is mainly early to mid Lochkovian, but probably ranges into the late Lochkovian (Greiner, 1970; Skinner, 1974; McGregor, 1989a, 1989b, 1992; Irrinki, 1990; Wilson and Kamo, 2008) (fig. 2). Rhyolite in the Archibald Settlement Formation has yielded a U-Pb (zircon) age of 415.6 ± 0.4 Ma (Wilson and Kamo, 2008).

In the Restigouche Syncline, all Dalhousie Group strata are assigned to the Val d'Amour Formation (Wilson and others, 2004, 2005) (fig. 2). The Val d'Amour comprises a very thick (*ca.* 5.7 km) section of subaerial effusive and pyroclastic volcanic rocks, and shows an upward lithological and chemostratigraphic transition from dominantly basalt, to dominantly intermediate rocks, to rhyolite (Wilson and others, 2005). In most of the unit, sedimentary rocks are rare, and consist of thin intervals of terrestrial sandstone, siltstone and conglomerate. However, the easternmost exposed part of the unit at Dalhousie includes two sequences of very fossiliferous calcareous siltstone and calcarenite (Clarke, 1909; Alcock, 1935), as well as pillow basalt and related hyaloclastite, together indicating that shallow marine conditions prevailed in that area. Nevertheless, the Val d'Amour Formation exhibits no evidence of the Lochkovian transgression that occurred in the Jacquet River Syncline. Palynological data show that the Val d'Amour Formation ranges from *ca.* mid-Lochkovian to Pragian or early Emsian, which agrees well with a U-Pb (zircon) age of 407.4 ± 0.7 Ma from rhyolite at the top of the unit (Wilson and others, 2004, 2005). In the Restigouche Syncline, therefore, the Dalhousie Group spans a greater age range and encompasses rocks that are significantly younger than in the Jacquet River Syncline (fig. 2).

South of Rocky Brook–Millstream Fault

In the Tobique zone, the Tobique Group is in fault contact with lower Silurian and older rocks of the MCS to the west, whereas to the south it disconformably overlies those rocks. To the east, the Tobique Group unconformably overlies Cambro-Ordovician rocks of the Miramichi inlier (fig. 3); this contact is locally intruded by the 419.0 ± 0.5 Ma, hypabyssal (in part) Redstone Mountain Granite (Wilson and Kamo, 2016). The Tobique Group was originally defined in the area east of Plaster Rock (fig. 3), where it comprises the Costigan Mountain Formation and overlying Wapske Formation (St. Peter, 1978). Here, the Costigan Mountain Formation consists of mixed subaerial and subaqueous felsic and mafic volcanic rocks, and subordinate marine sedimentary rocks, whereas the Wapske Formation comprises interbedded marine sedimentary and mafic volcanic rocks (St. Peter, 1978, 1979, 1981). Lower Devonian subaerial volcanic rocks that underlie a northwest-southeast-trending enclave within the southern part of the Miramichi inlier in the Millville area (fig. 3) are also considered part of the Costigan Mountain Formation (St. Peter, 1982; Venugopal, 1982). Local names have been introduced for other, typically subaerial volcanic-dominated units in the lower part of the Tobique Group, including the Pentland Brook Formation (Wilson and Burden, 2006) east of Mount Carleton, and the Cameron Mountain Formation (Wilson, 1990) west of Plaster Rock (figs. 2 and 3). Wilson and Kamo (2008) report a U-Pb (zircon) age of 418.6 ± 0.9 Ma from rhyolite at the base of the Pentland Brook Formation, although they provisionally assigned this rhyolite to the Dickie Cove Group.

In the northern part of the Tobique zone, the oldest unit in the Tobique Group is the Greys Gulch Formation (Wilson, 2005), which conformably and gradationally overlies the Free Grant Formation, the uppermost unit of the transgressive, Ludlow–Pridoli Petit Rocher Group (figs. 2–4). The lithology and sedimentology of the Greys Gulch Formation is identical to that of shallow-water to terrestrial redbeds in the Mitchell Settlement Formation (Dalhousie Group; see above). Each represents a latest Pridolian to earliest Lochkovian marine regression, and they are interpreted as an originally continuous unit that was transected and offset by about 24 km of lateral dextral movement on the Rocky Brook−Millstream Fault during and/or after the Early Devonian (Wilson, 2007).

Just north of and along strike from the Tobique Group type area (fig. 3), large felsic or composite felsic-mafic volcanic piles are enveloped by dark grey siltstone of the Wapske Formation. These volcanic rocks include the Gulquac Mountains, Black Mountains and Blue Mountains suites of Wilson (1992), and the Serpentine Mountain and Mount Carleton volcanic edifices farther north (fig. 3). The Gulquac Mountains, Black Mountains and Mount Carleton volcanic suites were emplaced subaqueously; they are typically associated with hyaloclastic breccias and display a range of textural features consistent with eruption and rapid quenching in water (Wilson, 1992). Just west of these volcanic suites, the Blue Mountains–Serpentine Mountain belt comprises subaerial felsic flows and pyroclastic rocks, whereas associated mafic rocks include massive to pillowed basalt flows indicating at least local subaqueous emplacement. Although volcanic rocks assigned to the Wapske Formation are abundant in this part of the Tobique zone, the Wapske Formation as a whole is dominated by marine sedimentary rocks, many of which display sedimentary structures typical of turbidite deposits (St. Peter, 1978; Skinner, 1982; Wilson, 1990; Han and Pickerill, 1994). The age of the Wapske is defined by brachiopod and spore ages that span the Lochkovian (Helmstaedt, 1971; St. Peter, 1978, 1979, 1982; Skinner, 1982; McGregor, 1989a; Wilson, 1990; Boucot and Wilson, 1994; Wilson and Burden, 2006) (fig. 2). Volcanic rocks intercalated with these sedimentary rocks (aside from the large volcanic piles discussed above) are mainly basalt, whereas felsic volcanic rocks are comparatively rare; none of the latter has as yet been dated. In summary, the Tobique Group stratigraphy defines an overall transgressive sequence consisting of mainly subaerial volcanic and local terrestrial sedimentary rocks, overlain by marine sedimentary and minor volcanic rocks.

Fine-grained sedimentary rocks (mudstone and siltstone) have accommodated nearly all the superimposed strain in the Tobique Group, and are cast into close to tight, northeast-trending folds having a well-developed axial planar cleavage. Volcanic rocks are mesoscopically weakly deformed or undeformed (St. Peter, 1978; Wilson, 1990), although they are folded at the macroscale. Volcanic rocks underlie the region's higher peaks, the flanks of which are commonly covered in talus, obscuring contacts with the sedimentary rocks that underlie the lower elevations and complicating the interpretation of structural-stratigraphic relationships between the two rock types.

GEOCHRONOLOGY

Analytical Methods

Zircon grains were pretreated by chemical abrasion to penetratively remove zones of alteration and radiation damaged zones (Mattinson, 2005) by placing them in a muffle furnace at ~1000 °C for ~24 to 48 hours to anneal radiation damage, followed by a single, partial dissolution step in 50 percent HF in Teflon dissolution vessels at 200 °C for approximately 17 hours. Each zircon grain was cleaned in HNO_3, and transferred to a Teflon bomb (Krogh, 1973). A mixed ^{205}Pb-$^{233-235}$U spike was added to the Teflon dissolution capsules during sample loading (EARTHTIME community tracer, to facilitate inter-laboratory comparisons, see www.earth-time.org). Zircon was dissolved using ~0.10 ml concentrated HF acid and ~0.02 ml 7N HNO_3 at 200 °C for 3 to 5 days, dried to a precipitate and re-dissolved in ~0.15 ml of 3N HCl. U and Pb were isolated from the zircon solutions using anion exchange columns, dried down with dilute phosphoric acid, deposited onto outgassed rhenium filaments with silica gel (Gerstenberger and Haase, 1997), and analyzed with a VG354 mass spectrometer using a Daly detector in pulse counting mode. Corrections to the ^{206}Pb-^{238}U and ^{207}Pb/^{206}Pb ages for initial ^{230}Th disequilibrium in the zircon data have been made assuming a Th/U ratio in the magma of 4.2. Laboratory procedural blanks are routinely at the 0.5 pg and 0.1 pg level for Pb and U, respectively, but common Pb can be introduced during sample preparation. All common Pb was assigned to procedural Pb blank. Dead time of the counting system for Pb was 16 ns and 14 ns for U. The mass discrimination correction for the Daly detector is constant at 0.05 percent per atomic mass unit. Amplifier gains and Daly characteristics were monitored using the SRM 982 Pb standard. Thermal mass discrimination correction for Pb is 0.10 percent per atomic mass unit. U fractionation was measured internally and corrected for each measurement cycle. Decay constants are those of Jaffey and others (1971), and age calculations were done using an in-house program by D.W. Davis. All age errors quoted in the text and table 2, and error ellipses in the concordia diagrams (fig. 5) are given at the 95 percent confidence interval. Plotting and age calculations were done using Isoplot 3.00 (Ludwig, 2003).

J13h-36

The rock sample, a pink aphyric rhyolite from near the base of the Cameron Mountain Formation (Tobique Group; fig. 3), yielded abundant euhedral, doubly terminated, prismatic zircon crystals. Two single zircon grains and one, two-grain fraction, gave overlapping, concordant data that have a weighted mean ^{206}Pb/^{238}U age of 421.1 ± 1.4 Ma (2σ). The MSWD of 3 indicates some scatter in the data. The ^{206}Pb/^{238}U ages of the two single crystals show strong overlap and have a mean age of 420.44 ± 0.56 Ma (MSWD = 0.23). One interpretation is that z1 is slightly older due to a minor inherited component, and the more precise age from the two single crystals is more indicative of the true age of crystallization. Our preferred interpretation, although based on only two analyses, is that 420.44 ± 0.56 Ma represents the best estimate for zircon crystallization and formation of the rhyolite (fig. 5A; table 2).

O16b-31

The sample is a reddish maroon, sparsely feldspar-phyric rhyolite from the Wildcat Brook Formation near the base of the Dalhousie Group, just above the interpreted disconformity with the underlying Benjamin Formation, and a short distance west of the Arleau Brook Fault (fig. 4). The age of this formation establishes the time of onset of Early Devonian volcanism in the Chaleur zone.

Zircon crystals recovered from this rock sample form a uniform population of euhedral stubby prisms that were severely etched during the chemical abrasion

TABLE 2

U-Pb CA-ID-TIMS zircon data for rhyolite samples from the Tobique and Dalhousie groups, northern New Brunswick

#	No. grains.	Th/U	Pb_c (pg)	$^{206}Pb/^{204}Pb$ meas.	$^{207}Pb/^{235}U$	2σ	$^{206}Pb/^{238}U$	2σ	Error Corr.	$^{206}Pb/^{238}U$ Age(Ma)	2σ	$^{207}Pb/^{235}U$ Age(Ma)	2σ
J13h-36 (Lat. 46-56-38.6N, Long. 67-31-00.0W), Cameron Mountain Formation													
Z1	1	0.48	1.7	3729	0.5148	0.0021	0.06756	0.00006	0.604	**421.42**	**0.38**	421.7	1.4
Z2	1	0.49	1.1	8518	0.5145	0.0025	0.06741	0.00012	0.569	**420.55**	**0.71**	421.5	1.7
Z3	1	0.52	0.5	10152	0.5126	0.0016	0.06737	0.00015	0.816	**420.27**	**0.91**	420.2	1.1
O16b-31 (Lat. 47-46-59.9N, Long. 66-10-04.3W), Wildcat Brook Formation													
Z1	2	0.80	1.6	3052	0.5123	0.0024	0.06724	0.00015	0.625	**419.52**	**0.92**	420.0	1.6
Z2	2	0.73	0.3	6429	0.5116	0.0020	0.06694	0.00014	0.669	**417.69**	**0.86**	419.6	1.3
Z3	2	0.79	0.5	2055	0.5067	0.0055	0.06695	0.00019	0.477	**417.74**	**1.14**	416.2	3.7
Z4	2	0.80	0.7	2161	0.5047	0.0029	0.06690	0.00008	0.565	**417.46**	**0.50**	414.9	2.0
Z5	2	0.83	1.1	1538	0.5104	0.0045	0.06686	0.00025	0.559	**417.22**	**1.49**	418.7	3.0
J15e-RW13 (Lat. 46-52-33.8N, Long. 66-57-42.0W), Costigan Mountain Formation													
Z1	2	0.66	0.3	3167	0.5224	0.0025	0.06846	0.00009	0.564	**426.86**	**0.57**	426.7	1.7
Z2	3	0.65	0.9	2646	0.5161	0.0027	0.06774	0.00008	0.564	**422.55**	**0.48**	422.6	1.8
Z3	1	0.56	0.4	2860	0.5144	0.0036	0.06756	0.00023	0.612	**421.46**	**1.42**	421.4	2.4
Z4	1	0.58	0.4	5753	0.5140	0.0017	0.06743	0.00010	0.658	**420.62**	**0.59**	421.1	1.1
J3g-RW14 (Lat. 46-08-47.0N, Long. 67-12-49.5W), Costigan Mountain Formation													
Z1	1	0.48	0.6	7888	0.6286	0.0022	0.07907	0.00009	0.614	**490.55**	**0.51**	495.2	1.4
Z2	2	0.54	1.8	679	0.5434	0.0101	0.07039	0.00016	0.633	**438.52**	**0.98**	440.7	6.7
Z3	1	0.52	1.0	1067	0.5309	0.0059	0.07020	0.00017	0.445	**437.34**	**0.99**	432.4	3.9
Z4	1	0.54	1.6	787	0.5291	0.0079	0.06922	0.00011	0.719	**431.48**	**0.65**	431.2	5.2
Z5	1	0.52	0.8	2717	0.5192	0.0025	0.06801	0.00013	0.483	**424.15**	**0.80**	424.6	1.7
Z6	1	0.59	0.6	2871	0.5135	0.0024	0.06718	0.00013	0.485	**419.13**	**0.80**	420.8	1.6
Z7	1	0.52	0.4	6275	0.5121	0.0014	0.06729	0.00013	0.516	**419.78**	**0.79**	419.9	1.0

TABLE 2
(continued)

#	No. grains.	$\frac{Th}{U}$	Pb_c (pg)	$\frac{^{206}Pb}{^{204}Pb}$ meas.	$\frac{^{207}Pb}{^{235}U}$	2σ	$\frac{^{206}Pb}{^{238}U}$	2σ	Error Corr.	$\frac{^{206}Pb}{^{238}U}$ Age(Ma)	2σ	$\frac{^{207}Pb}{^{235}U}$ Age(Ma)	2σ
O7f-RW14 (Lat. 47-27-39.4N, Long. 66-50-10.3W), Wapske Formation													
Z1	1	1.01	1.9	530	0.5163	0.0121	0.06783	0.00022	0.568	**423.06**	**1.32**	422.7	8.1
Z2	1	0.98	1.9	531	0.5151	0.0116	0.06747	0.00020	0.604	**420.91**	**1.18**	421.9	7.8
Z3	1	0.99	1.0	1669	0.5120	0.0038	0.06735	0.00011	0.499	**420.18**	**0.67**	419.8	2.6
Z4	1	0.93	2.1	873	0.5113	0.0069	0.06730	0.00011	0.630	**419.88**	**0.68**	419.3	4.6
Z5	2	0.83	2.5	1134	0.5116	0.0055	0.06726	0.00018	0.485	**419.61**	**1.09**	419.5	3.7
Z6	1	1.14	1.6	547	0.5137	0.0113	0.06721	0.00021	0.529	**419.35**	**1.24**	421.0	7.6

Notes: Chemical abrasion of zircon grains modified from Mattinson (2005); all analyses used EARTHTIME community spike ET535. Th/U calculated from radiogenic $^{208}Pb/^{206}Pb$ ratio and $^{207}Pb/^{206}Pb$ age.

Correction for ^{230}Th disequilibrium in $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ assuming Th/U of 4.2 in the magma. Pb_C is total common Pb assuming isotopic composition of laboratory blank. (laboratory Pb isotopic composition $^{206}Pb/^{204}Pb$: 18.221; $^{207}Pb/^{204}Pb$: 15.612; $^{208}Pb/^{204}Pb$: 39.360; 2σ errors of 1 %).

$^{206}Pb/^{204}Pb$ and Pb/U corrected for fractionation and common Pb in spike; Pb/U ratios also corrected fro blank. Error Corr. is 'correlation coefficients' of X-Y errors on the concordia plot.

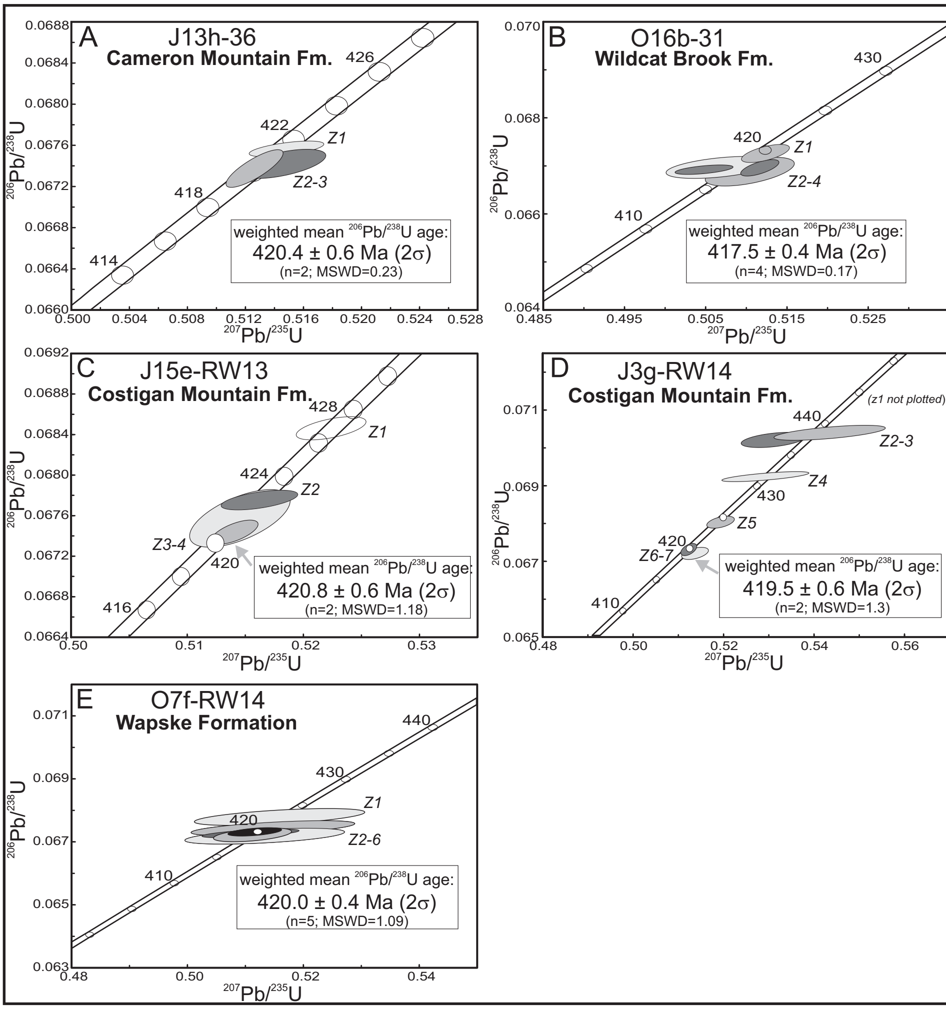

Fig. 5. U-Pb concordia diagrams showing data for single, chemically abraded zircon grains from felsic volcanic rocks of the Tobique and Dalhousie Groups. (A) Cameron Mountain Formation; (B) Wildcat Brook Formation; (C) Costigan Mountain Formation (type area); (D) Costigan Mountain Formation (Millville area); and (E) Wapske Formation. See figure 2 for stratigraphic position and figures 3 and 4 for locations.

procedure, which left a residue of small translucent fragments of zircon. Five fractions of multiple small fragments gave overlapping U-Pb data. The four youngest have a weighted mean $^{206}Pb/^{238}U$ age of 417.53 ± 0.39 Ma (2σ, MSWD = 0.17). The fifth datum is slightly older with a $^{206}Pb/^{238}U$ age of 419.5 ± 0.9 Ma, which overlaps with the mean of the other four data, but may have contained an inherited fragment, and was therefore excluded from the mean age to avoid biasing the age upwards. Our preferred interpretation is that the zircon crystallized at 417.53 ± 0.39 Ma, which represents the best indication of the age of the rhyolite (fig. 5B; table 2).

J15e-RW13

The dated sample is a massive, pink-buff, sparsely feldspar-phyric rhyolite from the base of the Costigan Mountain Formation (Tobique Group), central New Brunswick (fig. 3). Abundant zircon crystals are euhedral stubby prismatic grains. Four concordant data were obtained from two multi- and two single grains. Data for Z1 and Z2 (two grains each) indicate that one or both grains in each fraction are inherited (or contain an inherited component) and have $^{206}Pb/^{238}U$ ages of 426.86 ± 0.57 Ma and 422.55 ± 0.48 Ma, respectively. The first is significantly older than all other grains analyzed and the second is resolvably older than the youngest analyzed grain. Concordant data for the two youngest data (Z3-4) have a weighted mean $^{206}Pb/^{238}U$ age of 420.75 ± 0.55 Ma (2σ; MSWD = 1.18), which is considered the best age estimate for crystallization of the rhyolite (fig. 5C; table 2). This is consistent with the 419.0 ± 0.5 Ma age of the nearby Redstone Mountain Granite (Wilson and Kamo, 2016), which intrudes the contact between the Tobique Group and much older rocks of the Miramichi inlier.

J3g-RW14

A sample of reddish maroon aphyric rhyolite was collected from an outcrop of Costigan Mountain Formation on Route 104 just west of Millville (fig. 3). The zircon population comprised euhedral, multi-facetted, equant, and four-sided short to elongate prismatic grains. Seven U-Pb analyses produced concordant data with discrete ages that range from 490.55 ± 0.51 Ma to 419.78 ± 0.79 Ma. The two youngest, obtained from the four-sided prismatic grains, gave concordant, overlapping data that have a weighted mean $^{206}Pb/^{238}U$ age of 419.46 ± 0.56 Ma (Z6-7), which is considered a maximum age for deposition of this rhyolite unit (fig. 5D; table 2).

O7f-RW14

The dated sample, collected about 12 km north of Mount Carleton (fig. 3), is a fragmental felsic rock interpreted as a rhyolite hyaloclastite, a common rock type in the Wapske Formation (Wilson, 1992). The rock comprises pinkish orange angular clasts of aphyric rhyolite in a greyish green groundmass of fine-grained, altered felsic glass. Analyzed zircon grains were euhedral, four-sided, short to elongate prismatic crystals. U-Pb data for five zircon analyses (Z2-6) are concordant and overlap, and have a weighted mean $^{206}Pb/^{238}U$ age of 420.00 ± 0.38 Ma (2σ, MSWD = 1.09). One analysis (Z1) yielded an older $^{206}Pb/^{238}U$ age of 423.06 ± 1.32 Ma; this grain is considered a xenocryst and is excluded from the mean. The crystallization age of the zircon grains is interpreted to be 420.00 ± 0.38 Ma, which is considered the best estimate for the time of deposition of the rhyolite (fig. 5E; table 2).

DISCUSSION OF RESULTS

New geochronological data reported herein demand a reassessment of existing correlations between the Tobique and Dalhousie Groups, have tectonic implications for the large-scale structure of the Tobique zone, and allow better separation of the effects of the Salinic and Acadian orogenies in the Ganderian core of the northern Appalachians.

Stratigraphic Implications

Rhyolite from the northern part of the Mount Carleton volcanic edifice yields a Pridolian age of 420.2 ± 0.5 Ma, essentially the same as rhyolite from the Costigan Mountain, Cameron Mountain and Pentland Brook Formations in the lower part of the Tobique Group. This age is also within error of a Laser Ablation–Inductively Coupled Plasma Mass Spectrometry age of 422.4 ± 2.3 Ma reported by Sánchez Mora (ms, 2014) for Wapske felsic volcanic rocks from the same area as sample O7f-RW14. If the volcanic edifices lying between the Tobique Group type area and Mount Carleton (fig. 3) can be assumed to be roughly coeval with the Mount Carleton rhyolite, volcanic rocks formerly interpreted to be interbedded with Wapske sedimentary rocks instead mainly underlie them, and represent fold repetitions of the volcanic-dominated basal

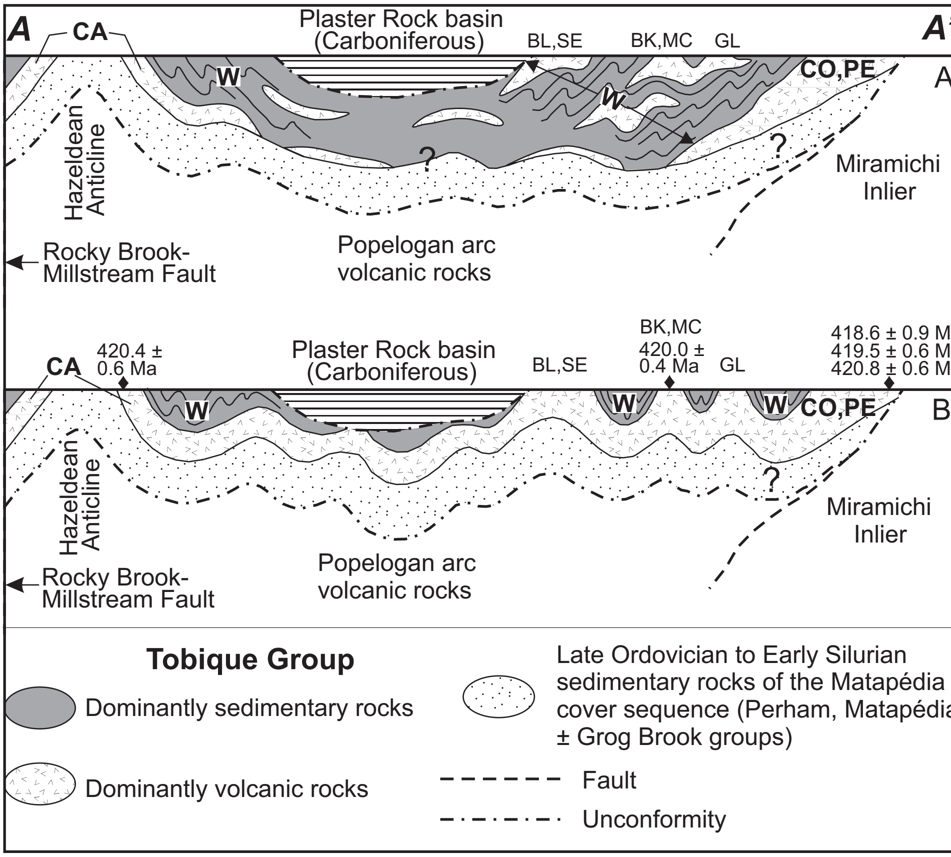

Fig. 6. Schematic rendering of the geometry of the Tobique zone, projected onto west-east cross-section A-A' across the central part of the Tobique basin (see fig. 3 for location). A. Previous interpretation of large-scale structure; and B. revised interpretation based on newly acquired geochronological data. No vertical scale is implied. CA = Cameron Mountain Formation; CO = Costigan Mountain Formation; PE = Pentland Brook Formation; W = Wapske Formation. Volcanic edifices assigned to the Wapske Formation (fig. 3): MC = Mount Carleton; SE = Serpentine Mountain; BK = Black Mountains; BL = Blue Mountains; GL = Gulquac Mountains.

stratigraphy (fig. 6). Thus, the bulk of Tobique volcanism occurred between *ca.* 421 and 419 Ma and correlates more closely with Pridolian volcanic rocks of the Dickie Cove Group than with those in the Dalhousie Group, which spans the same age range as Lochkovian sedimentary and subordinate volcanic rocks in the upper part of the Wapske Formation (fig. 2). Instead of an overall west-younging trend from Costigan Mountain in the east to Blue Mountain farther west (fig. 3), volcanic rocks exposed along the margins of the Tobique basin are interpreted to crop out in the central part of the basin in the hinges of non-cylindrical Acadian anticlines (figs. 3 and 6). As mentioned previously, the near complete lack of exposure of contacts between sedimentary and volcanic rock units has obscured field evidence of their structural and stratigraphic relationships.

Along the western limb of the Jacquet River Syncline, rhyolite from the Wildcat Brook Formation near the base of the Dalhousie Group has yielded an age of 417.5 ± 0.4 Ma, whereas rhyolite at the top of the Dickie Cove Group, roughly along strike but underlying a thin unit of conglomerate-siltstone-limestone, has been dated at 419.7 ±

0.3 Ma (Wilson and Kamo, 2012; fig. 4). An erosional surface interpreted by Wilson and Kamo (2008, 2012), is therefore confirmed by a hiatus of 2.2 myr (fig. 2). No clear evidence for an equivalent disconformity between correlative rocks in the Tobique Group (for example, between the Costigan Mountain and Wapske Formations) has been recognized, although a cryptic disconformity may coincide with one of several discontinuous bands of locally quartz pebble-rich conglomerate observed in the lower Tobique section.

Northern Appalachian Silurian–Devonian Correlations

Numerous parallels exist between the Silurian–Devonian sections in northern New Brunswick, northern Maine, and north-central Newfoundland, allowing tectonic inferences to be made where the New Brunswick data are incomplete or equivocal. In all areas, late Silurian sedimentation and volcanism immediately postdate a several myr stratigraphic hiatus (Salinic B unconformity of Wilson and Kamo, 2012) (fig. 2) that marks uplift and deformation associated with the Salinic collision of the Gander margin and composite Laurentia. In Newfoundland, this unconformity separates Salinic-deformed Upper Ordovician to Llandovery (≥433 Ma) marine sedimentary rocks of the Badger Group from the overlying, upper Silurian to lowermost Devonian, largely terrestrial Botwood Group (Williams and others, 1993, 1995; Pollock and others, 2007; van Staal and others, 2014, and references therein). In New Brunswick, the unconformity occurs between locally Salinic-deformed Llandovery rocks of the Quinn Point, Matapedia and Perham Groups, and overlying Ludlow–Pridoli rocks of the Dickie Cove, Chaleurs and Tobique Groups (fig. 2). Upper Silurian rocks were generally less strongly deformed or remained undeformed during Salinic orogenesis. This is most obvious, both in Newfoundland and New Brunswick, where the unconformity is demonstrably angular (van der Pluijm and others, 1993; Wilson and Kamo, 2012). The terrestrial Botwood Group, along with spatially related, Ludlow–Pridoli volcanic rocks of the Stony Lake and other volcanic complexes (Dunning and others, 1990; Elliott and others, 1991; van Staal and others, 2014) can be correlated with volcanic and subordinate sedimentary rocks of the Dickie Cove and Tobique Groups, and terrestrial sedimentary rocks in the lower part of the Petit Rocher Group. The Botwood Group includes upper Silurian to lowermost Devonian red beds (Dickson, 1994; Boyce and Ash, 1994) that likely correlate with terrestrial conglomerate and sandstone in the upper Silurian Petit Rocher and Dickie Cove Groups, and/or red siltstone and mudstone in the lowermost Devonian Greys Gulch Formation (Tobique Group) and Mitchell Settlement Formation (Dalhousie Group). The age of a possible unconformity present within the Botwood Group in Newfoundland (Williams and others, 1995; van Staal and others, 2014; Donaldson and others, 2015) has not been clearly defined, but may coincide with pre-Acadian deformation of the Petit Rocher Group in the Nigadoo River Syncline (van Staal and de Roo, 1995).

Equally striking parallels exist between the upper Silurian to Lower Devonian sections in New Brunswick and northern Maine. On the flanks of the Chesuncook Dome, Munsungan Anticlinorium, and Lobster Mountain Anticlinorium, Cambro-Ordovician rocks are unconformably overlain by upper Silurian transgressive sequences (Hall, 1970; Marvinney, 1984; Schoonmaker and others, 2011) that resemble the Petit Rocher Group in New Brunswick (fig. 7). For example, volcanic rocks of the Spider Lake and Carpenter Pond Formations likely correlate with the Dickie Cove Group, and basal conglomerates in the Ripogenus Formation, Spider Lake Formation and East Branch Group correlate with the South Charlo and Simpsons Field Formations in the Dickie Cove and Petit Rocher Groups, respectively. Notably, Pridolian to earliest Devonian disconformities, possibly correlative with the unconformity within the Botwood Group in Newfoundland and late Silurian deformation in the Petit Rocher Group, are reported within the Ripogenus Formation and East Branch Group

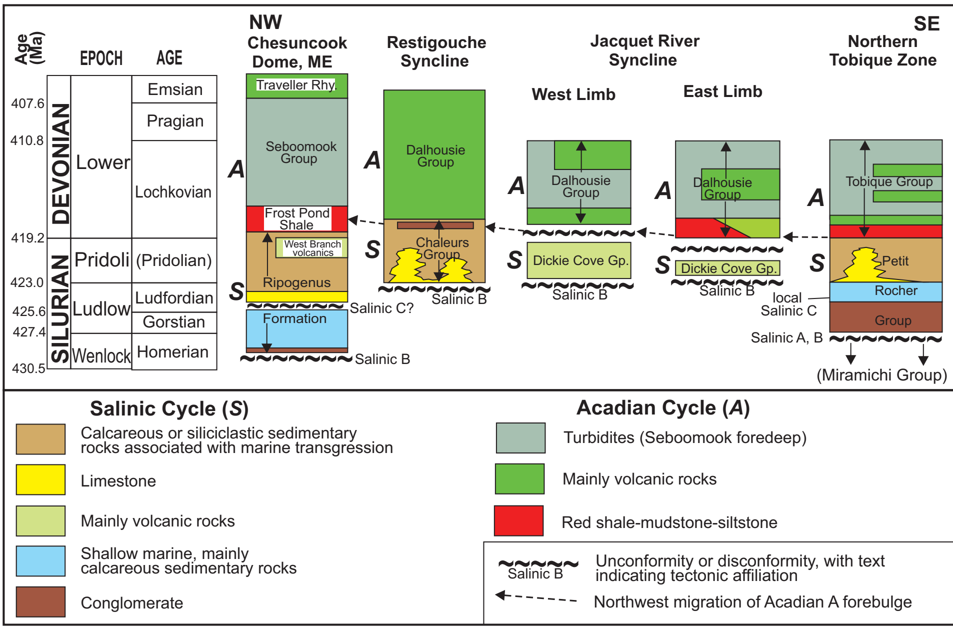

Fig. 7. Generalized stratigraphic columns for the late Silurian and Early Devonian in selected areas of northern New Brunswick and northern Maine, linking depositional and tectonic events to Salinic and Acadian orogenesis. It is proposed that latest Silurian–earliest Devonian disconformities and a coeval marine regression, represented mainly by deposition of red shale/siltstone, defines the transition between the Salinic and Acadian orogenic cycles. Geology of the Chesuncook Dome is modified from Schoonmaker and others (2011).

(Hall, 1970; Schoonmaker and others, 2011) (fig. 7). Upper Silurian units in Maine are overlain, at least locally, by Greys Gulch/Carl Gulch-like redbeds such as the Frost Pond Shale and Capens Formation (Boucot, 1969; Schoonmaker and others, 2011) (fig. 7), signifying a short-lived marine regression near the Silurian–Devonian boundary. Griscom (ms, 1976) inferred an estuarine environment for the Frost Pond Shale based on its red, oxidized colour, although Schoonmaker and others (2011) point out that it lacks fossils and primary structures that would confirm this interpretation. The Greys Gulch and Carl Gulch redbeds in New Brunswick similarly lack fossils and, in most cases, sedimentary structures; however, locally preserved primary structures such as wave- and current-ripple laminations, and the presence of local interbedded subaerial volcanic rocks, support a nearshore environment of deposition. In northern Maine, redbeds predate deposition of Lochkovian to early Emsian marine sedimentary rocks (flysch) of the Seboomook Group and Carrabassett Formation (Hall, 1970; Burroughs and Marvinney, 1981) and equivalent units that accumulated within the Seboomook foredeep of Bradley and others (2000) (fig. 7).

Tectonic Setting of Silurian–Devonian Volcanism

Terrestrial to submarine volcanic rocks account for a large proportion of the upper Silurian–Lower Devonian sequences of northern New Brunswick and adjacent Maine and Quebec (Dostal and others, 1989, 1993; Hon and others, 1992; Keppie and Dostal, 1994; Wilson and others, 2005; Schoonmaker and others, 2011; Wilson and Kamo, 2012). These volcanic rocks are tectonically important for several reasons: 1. they erupted immediately after the main Wenlock–Ludlow (430–423 Ma) phase of

Salinic collision-related deformation in underlying and adjacent early Silurian and older rocks (van Staal and de Roo, 1995; Wilson and Kamo, 2012); 2. they are coeval, at least in part, with deformation and exhumation of Salinic metamorphic tectonites (blueschist and greenschist to upper amphibolite facies) in the Brunswick subduction complex in New Brunswick (Fyffe and others, 1988; van Staal and others, 2003, 2008) and in south-central Maine (West and others, 1992, 1995, 2003; Tucker and others, 2001; Gerbi and West, 2007); and 3. they preceded and accompanied a late Pridolian–Lochkovian regressive-transgressive sequence in the Acadian retro-arc foreland basin (Malo and Bourque, 1993; Bradley and others, 2000; Bradley and Tucker, 2002; Wilson and others, 2004).

Upper Silurian–Lower Devonian bimodal volcanic rocks in northern New Brunswick, Maine and the Gaspé Peninsula of Quebec all display geochemical characteristics of within-plate magmatism, although they also commonly show trace element features such as negative Nb-Ta anomalies suggesting derivation from a subduction-modified or continental source (Laurent and Bélanger, 1984; Dostal and others, 1989, 1993; Hon and others, 1992; Wilson and others, 2005). Andesitic to dacitic compositions more typical of arc-type volcanism are common in some of the younger (mid to late Early Devonian) volcanic rocks, including the Edmunds Hill Andesite (Dockendorff Group) in Maine (Hon and others, 1992), the Val d'Amour Formation (Dalhousie Group) in the Restigouche Syncline (Wilson and others, 2005), and correlative rocks in the southern Gaspé Peninsula (Dostal and others, 1993). Nevertheless, all Silurian and Devonian volcanic rocks in northern New Brunswick show a close overall geochemical similarity; subtle compositional differences are explained by source heterogeneity and variations in fractionation history (Dostal and others, 1989, 2016). Hence, the transition from early stage (mainly Pridolian) to late stage (mainly Lochkovian) volcanism, which we will link to the Salinic and Acadian orogenic cycles, respectively (see below), is not reflected in volcanic rock geochemistry.

Dostal and others (1989, 1993), Hon and others (1992), and Keppie and Dostal (1994) explain all Silurian–Devonian bimodal, within-plate magmatism in the northern Appalachians as resulting from development of a continental rift within the Quebec re-entrant during orogen-parallel transcurrent movements in an overall transpressive system. The postulated rift has been variously attributed to sinistral transpression (Dostal and others, 1989; Hon and others, 1992), dextral transpression (Dostal and others, 1993), or coincident with a switch from sinistral to dextral transpression (Keppie and Dostal, 1994); transpression in each case is associated with oblique convergence of Gondwanan terranes with the irregular Laurentian margin (fig. 1). However, this model does not easily explain the somewhat uneven, along-strike distribution of these rocks from northern Newfoundland to Maine, while crossing at least one promontory and a re-entrant.

In contrast to the transpressional rift model, van Staal and others (2009, 2014) explain Pridolian to earliest Lochkovian (422.3–418.6 Ma) volcanism as a result of detachment of the subducting Tetagouche oceanic backarc slab (fig. 8A), whereas later (417.5–407.4 Ma) Lochkovian volcanism (discussed below) is associated with shallow underthrusting of Avalonia (flat slab stage) following closure of the Acadian seaway and collision of Avalonia and composite Laurentia (fig. 8B). First, in the late Silurian, *ca.* 425 Ma slab break-off resulted in juxtaposition of hot mantle against the overlying lithosphere of the thickened collision zone along the southeastern edge of composite Laurentia, leading to intraplate volcanism. The lag times between collision and break-off of the relatively young and hot Tetagouche backarc slab, and between slab break-off and within-plate volcanism, are only a few myr in each case. This compares well with lag times reported for Newfoundland (Whalen and others, 2006), the Alps (van Blanckenburg and Davies, 1995), and Baja California (Hildebrand and

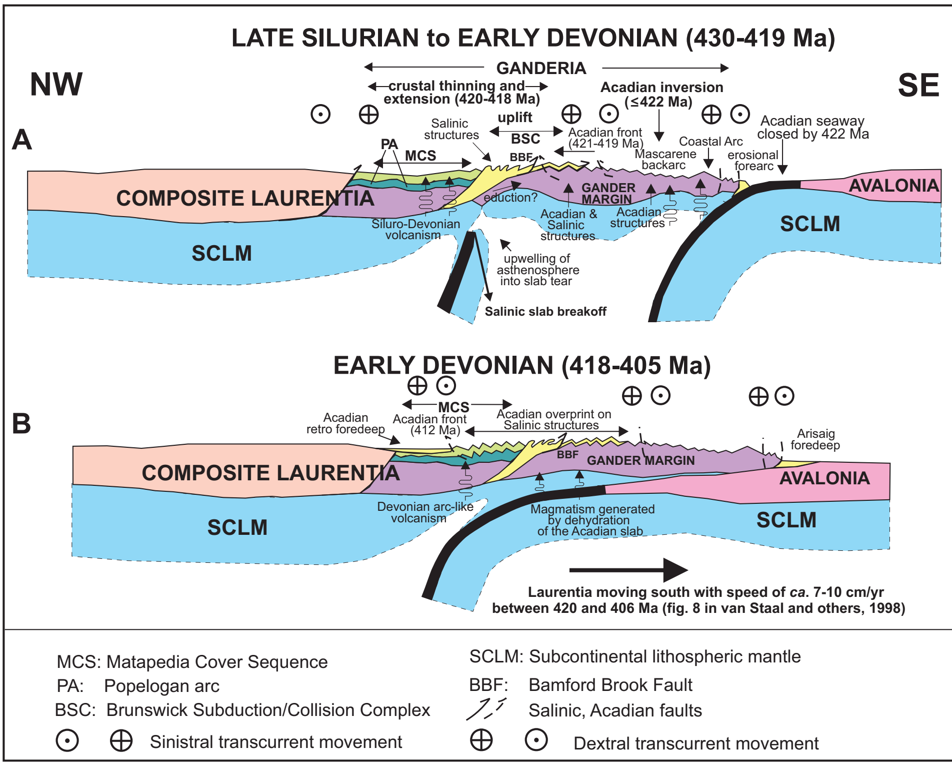

Fig. 8. (A) Late Silurian tectonic evolution of Ganderia, illustrating overlap between the waning stages of Salinic orogenesis in the northwestern part of Ganderia, the onset of Acadian orogenesis in the southeastern part of Ganderia's trailing edge, and inception of Acadian retroarc deformation, which migrated progressively northwestward over time. The Salinic orogenic cycle, which involved exclusively Ganderian elements, is associated with Late Ordovician to Llandovery (≤450 Ma) sinistral oblique convergence and accretion of backarc elements in the Brunswick subduction complex until the *ca.* 430 Ma onset of collision between Ganderia and composite Laurentia. The final stage of Salinic orogenesis is associated with break-off of the Salinic (Tetagouche) slab *ca.* 425 Ma, and consequent rapid uplift (including possible eduction of underthrusted Ganderian crust) and extensional collapse of the Salinic orogenic core. (B) The Acadian orogenic cycle begins with the onset of Pridolian–Lochkovian collision between Avalonia and composite Laurentia (now including all of Ganderia), which led to shut-off of the Coastal Arc. Rapid southerly advance of the composite Laurentian plate during the Early Devonian (≤ 418 Ma) was coeval with shallow (flat) underthrusting of Avalonia, which led to progressive expansion of retroarc deformation and a northwest-migrating retroarc foreland basin. Modified after van Staal and others (2009, 2014).

Whalen, 2014), although variation can be expected based on the age and physical properties of the subducting slab. Based on a recent Nd-isotopic study of Silurian and Devonian basalts in northern New Brunswick, Dostal and others (2016) suggested that the lack of a juvenile asthenospheric mantle signature in the petrogenesis of these rocks suggested melting of an old subcontinental lithospheric mantle (SCLM) in a transpressive rift setting rather than as a result of slab break-off. In general, it is difficult to explain how the old SCLM presumably underlying this part of New Brunswick survived the juvenile contributions from melts related to the Ordovician Popelogan Arc and Tetagouche backarc cycles (van Staal and others, 2016), including formation of oceanic backarc lithosphere preserved in the Fournier Supergroup (Winchester and others, 1992; table 1), without significant modification. Regardless, the rifting and slab break-off scenarios are not mutually incompatible, and the Nd isotope evidence can be accommodated within the break-off model. For example, von Blanckenburg and

Davies (1995) showed that the presence of an asthenospheric melt component is also absent in the type area of slab break-off magmatism in the Alps, and that the presence or absence of syn-collision asthenospheric melts in the collision zone is a function of the depth of slab detachment. Relatively deep (>50 km) break-off of the slab may prevent asthenospheric melting, which is significant in view of the estimate of Dostal and others (1993) that Silurian and Devonian basalts were generated at depths of 80 km to 60 km, respectively. The depth of slab detachment in turn is dependent on the width and age of the subducting oceanic basin as well the nature of passive margin subduction − that is, gradual versus abrupt (van Hunen and Allen, 2011; Sizova and others, 2012). Hence, the spatial and temporal overlap between Salinic collision-related deformation, rapid uplift, and Pridolian magmatism in central and northern New Brunswick (van Staal and others, 2003, 2008; Wilson and Kamo, 2012) supports a relationship to slab break-off, although an additional component of superimposed regional extension may have been present as well (see below).

It is suggested that, in the area underlain by the Silurian–Devonian volcanic belts discussed herein, upwelling asthenospheric mantle provided a source of heat to initiate partial melting in the overlying SCLM − the mantle source inferred for these volcanic rocks by Dostal and others (1989, 1993, 2016), Hon and others (1992), and Keppie and Dostal (1994). This is not unusual, as the source of collision-related magmatism in the eastern Anatolian plateau was also shown to be the SCLM (Pearce and others, 1990). Local convection-induced advective heating of the SCLM shortly after slab detachment in the Anatolian−Iranian plateau may have promoted local “dripping” of the SCLM into the underlying asthenosphere, producing syncollision magmatism over wide and in some cases isolated areas (Kaislaniemi and others, 2014) rather than in a strictly linear pattern as would be predicted by a simple break-off scenario.

Eruption of the younger (*ca.* 417–407 Ma) volcanic rocks, as well as Lochkovian to Emsian Acadian deformation in northern New Brunswick, has been attributed to a hinterland-migrating and progressively flattening Acadian slab (van Staal and others, 2009; van Staal and Barr, 2012). Paleomagnetic evidence (van Staal and others, 1998) shows that underthrusting of progressively more buoyant Avalonian continental crust (fig. 8B) was coeval with rapid southerly advance (*ca.* 7–10 cm/yr) of the Laurentian plate during the Early Devonian, suggesting that this process may have led to swift overthrusting of the subducting hinge and hence flattening of the Acadian slab beneath composite Laurentia after 418 Ma (Murphy and others, 1999; van Staal and others, 2009, 2014). It also must have led to expulsion of the intervening mantle wedge towards the hinterland (fig. 8B). Pockets of asthenosphere trapped above the flattening Acadian slab underwent partial melting as the slab dehydrated (Humphreys, 2009), which may account for local manifestations of Early Devonian arc-like volcanism like that in the Val d’Amour Formation in New Brunswick and Edmunds Hill Andesite in Maine. Many of these rocks, especially in the Val d’Amour Formation and Edmunds Hill Andesite (data reported by Wilson and others (2005) and Hon and others (1992), respectively) have elevated La/Yb and Sr/Y ratios (>10 in each case), and Yb <2 ppm, like “adakitic” volcanic rocks in Chile that have been associated with shallow subduction and/or subduction erosion (Kay and others, 1991, 2005; Haschke and others, 2002). This is significant in view of the evidence for subduction erosion (that is, absence of forearc sedimentary rocks) along the Avalonia−Ganderia suture in southern New Brunswick (van Staal and others, 2009).

Schoonmaker and others (2011) discussed the “pros and cons” of four tectonic models for Silurian – Devonian volcanism in Maine and New Brunswick, and several points were raised in opposition to the flat-slab convergence scenario (their Model 2) of van Staal and others (2009, 2014) compared to the classic Laramide and Andean flat-slab models: 1) the overall orogen-parallel linearity of Silurian–Devonian volcanic rocks within the boundaries of Ganderia's leading segment (that is, between the Red Indian Line and Dog Bay Line; fig. 1), despite the somewhat uneven distribution within this belt; 2) contrasting geochemical trends compared to the Laramide–Andean model; 3) sedimentation and volcanism associated with subsidence rather than regional uplift; 4) (closely related to 3): marine rather than terrestrial sedimentation; and 5) a lack of pyroclastic rocks and volatile-rich magmas, at least in northern Maine. However, arguments 1 and 2 of Schoonmaker and others (2011) are equivocal and mainly depend on mantle conditions and source complexities, whereas arguments 3 to 5 are, at least in part, not applicable. For example, development of a Cordilleran retro-arc marine foreland basin was coeval with flat subduction of the Farallon plate (DeCelles, 2004). In general, many factors, primarily subduction velocity and topographic growth of the associated mountain belt, may control the degree of subsidence in retro-arc foreland basins, such that a wide range of subsidence patterns may develop (Sinclair and Naylor, 2012). Second, although pyroclastic rocks are uncommon in the older volcanic sequences (Dickie Cove and Tobique Groups), they make up a significant proportion (*ca.* 30%) of the Dalhousie Group. Most importantly, Schoonmaker and others (2011) assume that all Pridolian–Lochkovian volcanism in central and northern New Brunswick and adjacent Maine is associated with a single petrogenetic process and geodynamic setting, and do not take into account the effects of late Salinic (*ca.* 422–419 Ma) tectonic processes associated with Tetagouche slab detachment immediately preceding the proposed ≤418 Ma flattening of the Avalonian slab. We suggest that the first-order, orogen-parallel linearity of the magmatic belt, and the difference in geochemical trends, are at least partly a consequence of the mantle structure and source complexity imposed by these two, temporally closely-spaced events.

In view of the ample evidence for oblique convergence between Avalonia, Ganderia and composite Laurentia, we agree that local extension of Ganderian crust (now part of composite Laurentia) may have been caused by transcurrent motions, producing volcanism in pull-apart rift basins along releasing bends (Dostal and others, 1989, 1993; Keppie and Dostal, 1994). However, this process is expected to be only a local phenomenon in an overall transpressive regime, and the nearly continuous and irregular distribution of upper Silurian–Lower Devonian magmatic rocks from Newfoundland into Maine does not readily support this model, as the normal faults associated with the formation of pull-apart basins have not been identified.

Another potential cause of extension can be inferred from the late Silurian–Early Devonian plate configuration of the Gander margin, which occupied a unique tectonic setting during this time: the northern, leading edge of the Gander margin formed the lower plate with respect to its convergence with composite Laurentia, whereas its southern, trailing edge formed the upper plate with respect to converging Avalonia (figs. 8A and 8B). Evidence for steepening (Regard and others, 2008) of the downgoing Tetagouche slab at the leading edge of the Gander margin during the latest Silurian, waning stages of the Salinic collision (van Staal and others, 2014) implies a slight temporal overlap between the final stages of the Salinic cycle and the start of Acadian collision between Avalonia and the Gander margin in southern Newfoundland, New Brunswick and coastal Maine. Rollback and/or steepening of the oceanic part of the Acadian slab during the last (*ca.* 424–421 Ma) stages of closure of the Acadian seaway along Ganderia's trailing southeastern edge (van Staal and others, 2009) (fig. 8A) prior to (≤418 Ma) slab flattening, may have induced rapid eduction of the underthrusted Gander margin in the Salinic collision zone, in addition to its progressive steepening. Either process or a combination of both (van Staal and others, 2014) could have subjected the overriding composite Laurentian plate to a short-lived period of Pridolian extension (Royden, 1993) (fig. 8A). If correct, such extension may have been responsible for, or at least contributed to: 1. orogenic collapse and formation of the Pridolian–early Lochkovian D_3 flat belts in the Salinic metamorphic core (Brunswick subduction complex) in northern New Brunswick (de Roo and van Staal, 1994; van Staal and de Roo, 1995; van Staal and Wilson, 2014); and 2. the regional onset of Pridolian–earliest Lochkovian volcanism.

The Silurian-Devonian Record in Relation to Salinic and Acadian Orogenesis

Salinic cycle.—Diverse sedimentary, deformational, and magmatic events that have been linked to either the Salinic or Acadian orogenic cycles are summarized in figure 9. Following the mid-Silurian Ganderia–Laurentia collision *ca.* 430 Ma and the onset of the main (*ca.* 430–425 Ma) phase of Salinic B orogenesis, break-off of the downgoing Tetagouche backarc slab is inferred to have started *ca.* 425 Ma (fig. 9). Slab break-off – a Salinic event – explains rapid late Silurian uplift of the overriding, composite Laurentian plate, as well as upwelling of asthenosphere in the created slab window, resulting in initial eruption at 422.3 ± 0.3 Ma of non-arc volcanic rocks in the MCS. Furthermore, uplift following slab break-off was directly responsible for the orogenic collapse recorded by D_3 vertical shortening and widespread formation of flat belts in the Brunswick subduction complex (de Roo and van Staal, 1994; van Staal and de Roo, 1995), hence that deformation is also assigned to the Salinic cycle (figs. 7 and 9). D_3 deformation is typically best developed in areas of relatively high-grade metamorphism associated with late Silurian–Early Devonian granite plutonism. For example, de Roo and van Staal (1994) have shown that, in the western part of the Miramichi inlier, D_3 occurred shortly after or during intrusion of the Mount Elizabeth Granite (U-Pb monazite age of 418 ± 1 Ma; Bevier and Whalen, 1990), based on flattening of the S_2 foliation around elongated cordierite porhyroblasts, which had also overgrown S_2, in its contact aureole. Flattening of S_2 around the cordierite porphyroblasts occurred on the limbs of tight to isoclinal recumbent F_3 folds of S_2 (van Staal and Wilson, 2014), indicating that the flattening of S_2 was a D_3 rather than D_2 phenomenon, and implying that D_3 took place near the Silurian–Devonian boundary. It is significant that D_3 structures are absent in the adjacent and/or overlying Lower Devonian rocks.

Late Salinic tectonism in the MCS is manifested mainly in deformation of the Ludlow–Pridoli Petit Rocher Group in the Nigadoo River Syncline (van Staal and de Roo, 1995); this event likely correlates with an unconformity reported within the Ripogenus Formation in northern Maine (Schoonmaker and others, 2011) (fig. 7). Similarly, northwesterly bedding trends in the Simpsons Field Formation (lower part of the Petit Rocher Group) near the margin of the Miramichi inlier east of Mount Carleton (fig. 3) are sharply discordant to those in overlying rocks of the Pentland Brook Formation (Tobique Group). A nearby outlier of Simpsons Field rocks within the Bathurst Supergroup of the Miramichi inlier (fig. 3) also displays a northwesterly trend. Because Pentland Brook volcanism (418.6 ± 0.9 Ma; Wilson and Kamo, 2008) has been linked (above) to the Salinic cycle, it follows that deformation in underlying rocks is also a consequence of Salinic tectonism; this event is referred to as the Salinic C deformation (figs. 7 and 9). A similar scenario is observed in Newfoundland, where the terrestrial Botwood Group post-dated the main (*ca.* 433–425 Ma) phase of Salinic deformation, but field evidence for temporal overlap between pre-Acadian deformation, sedimentation, and intrusion/eruption of upper Silurian–Lower Devonian magmatic rocks suggests that the Botwood Group was deposited late syn-tectonically during the waning stages of the Salinic orogeny (van Staal and others, 2014, and references therein).

Acadian cycle.—Acadian tectonism in the MCS cannot be discussed without reference to events occurring at the active southern edge of the Gander margin in southern

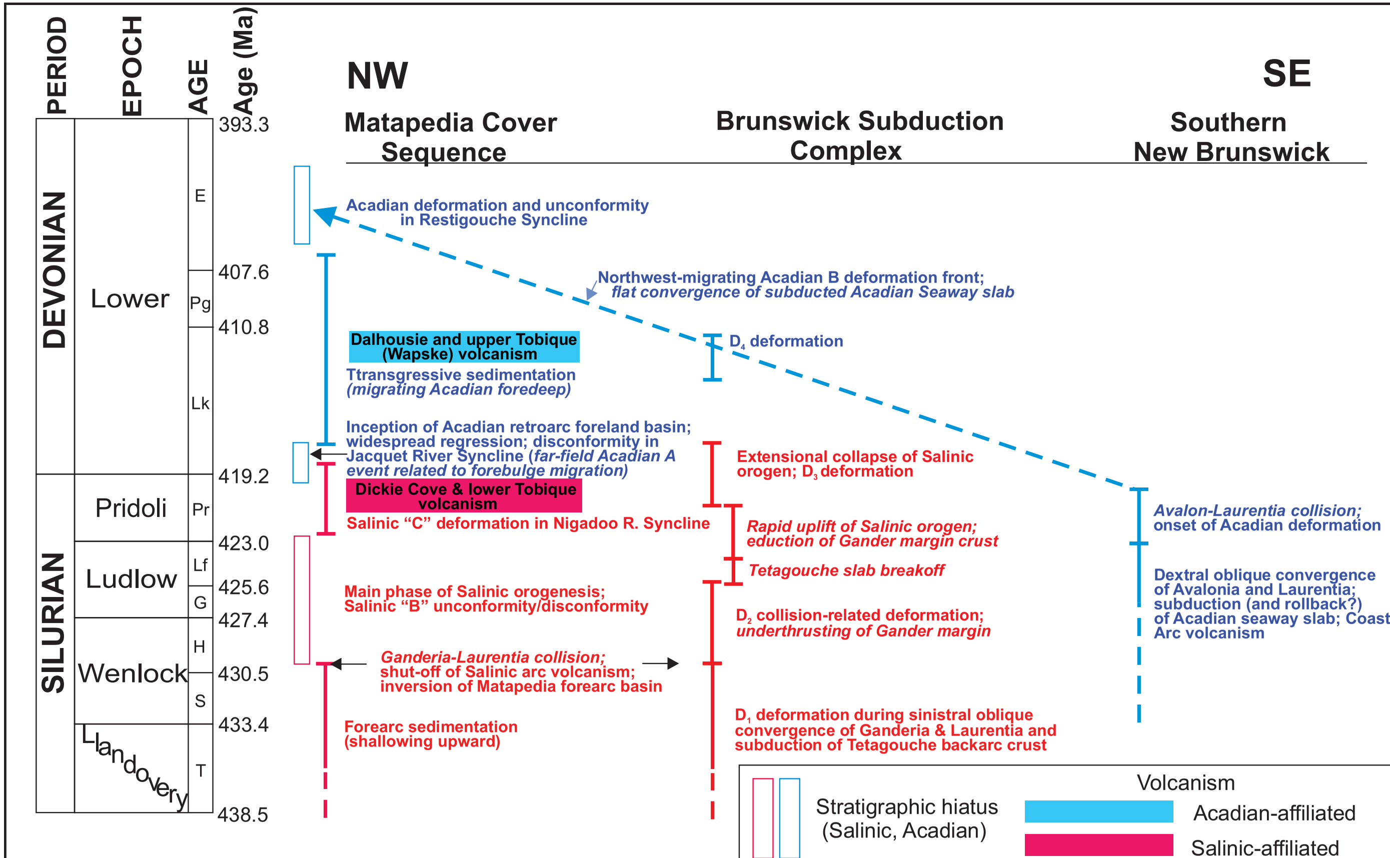

Fig. 9. Summary chart of tectonic events identified with the Salinic and Acadian orogenic cycles in the northern Appalachians. Observations are in plain font and interpretations in italic font. The latest Silurian–earliest Devonian disconformity between the Dickie Cove and Dalhousie Groups in the Jacquet River Syncline is interpreted to mark the break between the respective orogenic cycles. In general, overlap between the two cycles is temporal but not spatial; that is, the waning stages of Salinic deformation at the leading (northwestern) segment of Ganderia was coeval with the early stages of Acadian deformation at Ganderia's trailing (southeastern)

New Brunswick, where late Silurian closure of the Acadian seaway led to collision of Avalonia and composite Laurentia *ca.* 424 to 422 Ma (figs. 8B and 9). In Maine, the youngest age obtained from volcanic rocks of the Coastal Volcanic Belt is 423 ± 1 Ma (Van Wagoner and others, 2002). In southwestern New Brunswick, Acadian cleavage in the Silurian Mascarene Group (which was unaffected by Salinic deformation) is truncated by intrusive rocks that have yielded ages of 421 to 422 Ma (West and others, 1992; Fyffe and others, 1999; McLaughlin and others, 2003), and probably represent the last manifestation of Coastal Arc−Mascarene backarc-related magmatism. Acadian deformation in this area, therefore, temporally overlapped the latter stages of Salinic orogenesis in northern New Brunswick.

Late Salinic tectonism predates and, in part, temporally overlaps a late Pridolian–early Lochkovian disconformity and coeval marine regression (fig. 9), evidence for which is found in various locations in the northern Appalachians. In the MCS, it is manifested in several ways (figs. 7 and 9): 1) the 419.7 to 417.5 Ma hiatus between volcanic rocks of the Dickie Cove and Dalhousie Groups on the west limb of the Jacquet River Syncline, and a coeval or slightly older disconformity east of the Arleau Brook Fault on the east limb of the Jacquet River Syncline (fig. 2); 2) an earliest Devonian marine regression recorded by redbeds in the Greys Gulch and Mitchell Settlement Formations; minor redbeds are also reported in the upper Silurian–Lower Devonian Tracy Brook Formation in northwest New Brunswick (Carroll, 2003); and 3) limestone conglomerate in the upper part of the upper Silurian–Lower Devonian Indian Point Formation; the conglomerate consists exclusively of rounded clasts of Upper Ordovician to Llandovery calcilutite of the Matapedia Group (table 1), indicating uplift and erosion of those rocks near the Silurian–Devonian boundary (Wilson and others, 2004). At least locally, for example in the Jacquet River Syncline, the disconformity is interpreted to be diachronous from southeast to northwest.

The late Pridolian−early Lochkovian marine regression predates deposition of Lower Devonian turbidites, which in north-central New Brunswick are Lochkovian (Wapske and Jacquet River Formations), but in Maine, extreme northwestern New Brunswick and Quebec (Seboomook and Fortin Groups) extend into the early Emsian (St. Peter and Boucot, 1981; Bradley and Tucker, 2002) (figs. 2 and 9), again implying northwest-directed diachroneity. This suggests that the complete sequence records the passage of a forebulge-foredeep system in a retro-arc foreland setting (Catuneanu, 2004; DeCelles, 2012), preceding the northwesterly migration of the Acadian deformation front situated at that time farther southeast in New Brunswick and Maine (figs. 7 and 9) (compare Bradley and others, 2000). Late Pridolian−early Lochkovian disconformities were previously considered a manifestation of late Salinic rather than early Acadian tectonism in this part of Ganderia (van Staal and others, 2008; Wilson and Kamo, 2012), because the main phase of Acadian compressive deformation did not begin in northern New Brunswick until several myr later (see below). However, northwest migration of Acadian deformation was likely related to traction-coupling of the overriding Laurentian plate with the shallow Avalonian slab, suggesting that the forebulge-foredeep precursor of the deformation front should similarly be attributed to the Acadian rather than the Salinic cycle; hence, the latest Silurian−earliest

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Fig. 9 (continued). segment (van Staal and others, 2009, 2014) (fig. 1). However, Salinic-affiliated volcanism in the Tobique Group locally persists into the early stages of the latest Silurian−earliest Devonian marine regression and disconformity in the Matapedia Cover Sequence. D_1 through D_4 refer to the successive generations of folding and cleavage/schistosity observed in the Brunswick subduction complex (Miramichi inlier); D_4 structures in the Brunswick subduction complex correspond to the main phase of Acadian folding and cleavage in rocks of the Matapedia Cover Sequence. Abbreviations: T = Telychian; S = Sheinwoodian; H = Homerian; G = Gorstian; Lf = Ludfordian; Pr = Pridolian; Lk = Lochkovian; Pg = Pragian; E = Emsian.

Devonian disconformity and related marine regression are referred to as the Acadian A event (figs. 7 and 9).

Based on the Lochkovian to Emsian positions of the retro-arc Acadian deformation front in northern Maine (Bradley and others, 2000), and a *ca.* 405 to 400 Ma (mid-Emsian) age of Acadian deformation in the Restigouche Syncline of northern New Brunswick (Wilson and others, 2004, 2005), it is estimated that Acadian folding and penetrative cleavage (Acadian B deformation) in the New Brunswick part of the MCS occurred between 415 and 395 Ma (fig. 9). This is consistent with a recently acquired $^{40}Ar/^{39}Ar$ (muscovite-sericite) age of 414.7 $\pm$ 1.3 Ma for Acadian cleavage in the Tobique Group (D. Sanchez Mora, ms, 2014). Similarly, Acadian deformation in Newfoundland occurred between 415 and 410 Ma immediately east of the Dog Bay Line (McNicoll and others, 2006), which is correlated with the Bamford Brook Fault on the eastern margin of the Miramichi inlier in New Brunswick (van Staal and others, 2009; Reusch and van Staal, 2012) (fig. 1). Acadian deformation *ca.* 415 to 395 Ma is also consistent with the post-410 Ma age (Currie and others, 2003) assigned to Acadian structures and metamorphism that overprint Salinic structures in the adjacent Brunswick subduction complex (de Roo and van Staal, 1994; van Staal and de Roo, 1995) (figs. 2 and 8A).

Interpreting the latest Pridolian−earliest Lochkovian disconformity and marine regression as Acadian-related is accompanied by some degree of ambiguity, and depends on how one defines the end of one orogenic cycle and beginning of another: on one hand, late Silurian to earliest Devonian deformation in the Brunswick subduction complex (D_3) and Petit Rocher Group can be linked to Salinic tectonism, whereas the nearly contemporaneous passage of a retro-arc forebulge can be interpreted as a far-field effect of early Acadian deformation occurring much farther to the southeast (fig. 8B). Regardless of which tectonic process was the main cause of the disconformity, the complexities and diachronous nature of tectonism associated with Salinic and Acadian orogenesis underscore the case for defining orogenies on the basis of the nature and progression of tectonic and kinematic processes rather than on absolute time alone. Some of this complexity is reflected in the fact that not all parts of the MCS conform to the late Silurian−Early Devonian model presented here. In the Restigouche Syncline, though there is evidence for an earliest Devonian marine regression in the upper part of the Chaleurs Group (fig. 7), it was not followed by flysch sedimentation in a foredeep basin, but by eruption of almost 6 km of subaerial volcanic rocks (Wilson and others, 2005) (figs. 2 and 7). Indeed, except for the Pridolian to early Lochkovian period of Chaleurs Group sedimentation, the entire Restigouche region (the "Squaw Cap block" of Wilson and others, 2004), comprising rocks as old as Late Ordovician, occupied a high-level, low-strain position following Gander margin−Laurentia collision *ca.* 430 Ma, as shown by weak Acadian deformation and low regional thermal maturation (Wilson and others, 2004, 2005; Bertrand and Malo, 2005).

CONCLUSIONS

The record of late Silurian to Early Devonian sedimentation, volcanism and deformation in the northern Appalachians is complex, and characterized by transgressive-regressive sedimentary cycles, local disconformities and angular unconformities, and long-lived (*ca.* 422–407 Ma) within-plate magmatism. Precise ID-TIMS U-Pb (zircon) dating of volcanic rocks in the Chaleur Bay Synclinorium demonstrate that: 1) a *ca.* 2 myr latest Silurian–earliest Devonian (419.7–417.5 Ma) hiatus exists between the Dickie Cove Group and the overlying Dalhousie Group in the Jacquet River Syncline; and 2) volcanic rocks in the lower part of the Tobique Group range in age from *ca.* 421 to 419 Ma, and therefore correlate more closely with volcanic rocks of the late Silurian (Pridolian) Dickie Cove Group than those of the Dalhousie Group, which was previously regarded as wholly coeval with the Tobique Group. Instead, the Dalhousie Group is now correlated only with sedimentary and subordinate volcanic rocks in the upper part of the Wapske Formation. The newly acquired age of 420.2 ± 0.5 Ma for rhyolite in the northern part of the Mount Carleton volcanic edifice demands a re-interpretation of the large-scale structure of the Tobique zone; this and other volcanic suites in the central part of the Tobique basin should be regarded as structural repetitions of the basal, volcanic-dominated section represented by Pridolian volcanic rocks of the Costigan Mountain, Cameron Mountain and Pentland Brook Formations.

Geochronological data from volcanic rocks aid in reconstructing the Silurian–Devonian evolution of northern New Brunswick in the context of Salinic (Silurian) and Acadian (Devonian) orogenesis. The main (Salinic B) phase of Salinic orogenesis in the MCS followed the *ca.* 430 Ma collision of the Gander margin with composite Laurentia, resulting in widespread disconformities and angular unconformities in the northern Appalachians. Late Silurian extension and transgression in the Gander margin−composite Laurentia collision zone may be related to progressive steepening of the Salinic (Tetagouche) slab before and/or during slab detachment (Regard and others, 2008), possibly enhanced by eduction of the underthrusted Gander margin following rollback/steepening of the Acadian seaway slab prior to Avalonia−Laurentia collision (van Staal and others, 2014) (fig. 8A). We interpret the earlier (Pridolian) stage of bimodal, within-plate magmatism as a product of Salinic slab detachment, which probably started *ca.* 425 Ma. Pre-Acadian (Salinic C) deformation of Ludlow–Pridoli rocks of the Petit Rocher Group is contemporaneous with this early-stage volcanism, and possibly with late Salinic (D_3) deformation in the Brunswick subduction complex, and is therefore associated with the Salinic orogenic cycle.

The latest Silurian−earliest Devonian disconformity in the Jacquet River Syncline coincides with a widespread marine regression recorded by deposition of shallow-water to intertidal redbeds (Greys Gulch Formation and correlative units elsewhere in the northern Appalachians). Post-regression marine sedimentary rocks in the Dalhousie Group and the upper part of the Tobique Group are Lochkovian in age, but in northern Maine, deposition of correlative rocks of the Seboomook Group persisted into the early Emsian. These largely turbiditic rocks reflect a Lochkovian marine transgression preceding the northwest-migrating Acadian deformation front (Bradley and others, 2000). Both the regressive and transgressive cycles show evidence of southeast-to-northwest diachronism (fig. 7), and probably record the passage of an Acadian forebulge-foredeep pair in the Acadian retro-arc foreland basin, associated with flexural loading of Ganderian crust by the Acadian orogenic wedge following the collision of Avalonia and composite Laurentia *ca.* 424 to 422 Ma. It is proposed that, in the MCS, the disconformity and evidence of marine regression near the Silurian–Devonian boundary, defined here as the Acadian A event, mark the transition from the Salinic to the Acadian orogenic cycles. If correct, the effects of waning, latest Silurian–earliest Devonian Salinic deformation in northern New Brunswick slightly overlapped the far-field effects of Acadian orogenesis occurring farther to the southeast.

Early Devonian (Lochkovian to Emsian) volcanism and Acadian retro-arc foreland (Acadian B) deformation is linked to shallow (flat) underthrusting of buoyant Avalonian lithosphere, which began *ca.* 418 Ma (Whalen and others, 2006; van Staal and others, 2009, 2014) (fig. 8B). The flat-slab stage is temporally linked to rapid southerly motion of the composite Laurentian plate (van Staal and others, 1998). The initial manifestation of Acadian orogenesis was restricted to the area near the Avalonia–Ganderia suture in southern New Brunswick and adjacent Maine, but progressively migrated to the northwest, where deformation in the MCS is estimated to have mainly occurred between 415 and 395 Ma. For example, it is constrained to the mid-Emsian

(405–400 Ma) in the Restigouche Syncline (Wilson and others, 2004, 2005) (fig. 9) and to the mid-LochKovian (*ca.* 415 Ma) in the northern part of the Tobique zone (Sánchez Mora, ms, 2014).

Whatever the cause or combination of causes leading to Silurian–Devonian volcanism, it is clear that local extension and volcanism preceded the compressional phase of Acadian folding and cleavage formation, as all units hosting Silurian and Devonian volcanic rocks were deformed during Acadian orogenesis. Extension at this time also explains the evidence for late Pridolian (D_3) crustal thinning in the Salinic metamorphic core, coeval with basin subsidence towards the hinterland, and deposition of some of the volcanic rocks in actively subsiding marine basins. The model presented herein cannot be considered complete and definitive without additional data, but serves to underscore the complex interactions that occurred between key elements of the Appalachian collage at a critical period in the evolution of the northern Appalachian orogen.

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