CHAPTER 2
LITHOSPHERIC ARCHITECTURE AND TECTONIC EVOLUTION OF THE CANADIAN APPALACHIANS AND ASSOCIATED ATLANTIC MARGIN

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ABSTRACT

The Canadian Appalachians are a segment of the long and narrow Appalachian–Caledonian mountain belt, one of the Earth's classic orogens. The Canadian segment comprises a complex tectonic collage of mainly Early Paleozoic suprasubduction zone oceanic and ribbon-shaped microcontinental terranes, which were accreted to Laurentia as a result of Early to Late Paleozoic closure of the Iapetus and Rheic oceans. During this period, the Appalachians were an accretionary orogen, with the Laurentian margin progressively expanding oceanwards due to terrane accretion. The accreted suprasubduction terranes represent proto fore arc, extensional arc, and back-arc rocks, whereas the microcontinents either had Laurentian (Dashwoods) or Gondwanan (Ganderia, Avalonia, and Meguma) provenance. Tectonic addition of juvenile crust is less important than the accretion of already existing continental material. The structures associated with the accretion-related events are relatively well preserved in the Canadian Appalachians compared to other parts of the orogen due to its position in the hinterland of the Alleghanian collision, which led to formation of Pangea. Laurentia was the upper plate during the Alleghanian collision, but the suture between Laurentia and Gondwana is now hidden somewhere beneath the margins of the Atlantic Ocean.

Crustal thickness is least beneath the central part of the Canadian Appalachians, which is mainly underlain by Ganderian lower crust. Its relative thinness probably reflects a combination of resetting of the Moho as a result of large-scale melting of the lowermost crust and/or transformation into eclogite, possibly aided by the thinned nature of the Ganderian crust, achieved during several phases of back-arc rifting prior to accretion. Successive accretion of the oceanic and continental terranes to one another and/or to Laurentia caused several punctuated collisional events between the Late Cambrian and Late Carboniferous. The Taconic orogenic cycle comprises three distinct tectonic events between the Late Cambrian and Late Ordovician (495–450 Ma). The most important event in this cycle was the Early–Middle Ordovician (475–459 Ma) collision between the Humber margin of Laurentia and Dashwoods with the Notre Dame arc built upon it, forming the first phase of composite Laurentia. The Taconic orogenic cycle finished with accretion to composite Laurentia between 460 and 450 Ma of the peri-Gondwanan Popelogan–Victoria arc, which had been built on the leading edge of Ganderia.

In the periphery of Gondwana, Early Ordovician closure of the back-arc basin that separated the Penobscot arc and Ganderia caused the Early Ordovician Penobscot orogeny. Silurian accretion of the Gander margin to Laurentia caused the Salinic orogeny and was a result of the closure of the Tetagouche–Exploits back-arc basin. The Early Devonian Acadian orogeny was due to closure of the Acadian seaway between composite Laurentia (Ganderia) and Avalonia. Subsequent accretion of Meguma to composite Laurentia (Avalonia) caused the Middle Devonian to Early Carboniferous Neoacadian orogeny, although it is unsure whether Avalonia and Meguma were separated by true oceanic lithosphere. Orogenesis was terminated by the Carboniferous–Permian Alleghanian orogeny that welded Laurentia and Gondwana together into Pangea. The Canadian Appalachians underwent rifting during the Mesozoic, which led to diachronous opening of the Atlantic Ocean from south to north. Rifting was largely asymmetric and led to non-volcanic margins in Atlantic Canada and exhumation of serpentinized mantle onto the seafloor prior to seafloor spreading.

RÉSUMÉ

Les Appalaches canadiennes sont un segment de la chaîne de montagnes longue et étroite Appalaches–Calédonides, l'un des orogènes classiques de la planète Terre. Le segment canadien est constitué d'un collage tectonique complexe de terranes de zone océanique de supra-subduction et de terranes micro-continentaux en ruban du début du Paléozoïque principalement, qui ont été accrétés au continent laurentien par la fermeture des océans Iapétus et Rhéique du début jusqu'à la fin du Paléozoïque. Durant cette période, les Appalaches ont été un orogène d'accrétion, la marge laurentienne s'avançant progressivement vers l'océan par accrétion. Alors que les terranes de supra-subduction sont constitués de roches des zones de proto-avant-arc, d'arc d'extension, et d'arrière-arc, et celles des microcontinents provenaient ou de Laurentie (Dahwoods) ou du Gondwanie (Ganderia, Avalonia, et Meguma). L'adjonction de nouvelle croûte est moins importante que l'accrétion de matériau continental existant. Les structures associées aux événements accrétionnaires sont relativement bien préservées dans les Appalaches canadiennes par rapports aux autres segments de l'orogène étant donné sa position en arrière-pays durant la collision alléghénienne, laquelle a mené à la formation de la Pangée. La Laurentie constituait la plaque supérieure durant la collision alléghénienne, mais la suture entre les continents de Laurentie et de Gondwana est maintenant cachée quelque part sous les marges de l'océan Atlantique.

C'est sous la partie centrale des Appalaches canadiennes que la croûte est la plus mince, et là, elle repose principalement sur la croûte inférieure ganderienne. Sa minceur relative est probablement le reflet d'une combinaison de reconfiguration de la Moho suite à une fusion à grande échelle de la croûte inférieure et/ou une transformation en éclogite, peut-être facilitée par la minceur de la croûte ganderienne, cela s'étant produit en plusieurs phases de distension d'arrière-arc précédant l'accrétion. Les accrétions successives de terranes océaniques et continentaux, entre eux et avec le continent Laurentie, ont constitué plusieurs événements collisionnels distincts depuis la fin du Cambrien jusqu'à la fin du Carbonifère. Le cycle orogénique taconique est constitué de trois événements tectoniques distincts, depuis la fin du Cambrien jusqu'à la fin de l'Ordovicien (495–450 Ma). L'événement le plus important de ce cycle est la collision survenue, au début de l'Ordovicien moyen (475–459 Ma), entre la marge de Humber et les terranes de Laurentie et Dashwoods, l'arc insulaire de Notre Dame au-dessus formant la première phase de l'assemblage de Laurentie. Le cycle orogénique taconique s'est terminé par l'accrétion de l'assemblage de Laurentie, entre 460 Ma jusqu'à 450 Ma, avec l'arc péri-gondwanienne de Popelogan–Victoria qui s'était édifiée sur la marge frontale de Ganderia.

En périphérie du Gondwana, la fermeture, au début de l'Ordovicien, du bassin d'arrière-arc qui séparait l'arc de Penobscot et Ganderia, a provoqué l'orogenèse de Penobscot au début de l'Ordovicien. L'accrétion de la marge de Gander à la Laurentie au Silurien a provoqué l'orogenèse de Salinic, résultat de la fermeture du bassin d'arrière-arc de Tetagouche–Expoits. L'orogenèse acadienne au début du Dévonien a été provoquée par la fermeture du bras de mer acadien entre l'assemblage de Laurentie (Ganderia) et Avalonia. L'accrétion subséquente de Meguma à l'assemblage de Laurentie (Avalonia) a provoqué l'orogenèse néoacadienne entre le Dévonien moyen et le début du Carbonifère, bien qu'il ne soit pas certain qu'Avalonia ait été séparée de Meguma par une authentique lithosphère océanique. L'orogenèse s'est terminée par l'orogène alléghanienne au Carbonifère–Permien qui a soudé ensemble la Laurentie et le Gondwana formant ainsi la Pangée. Les Appalaches canadiennes ont commencé à se distendre au Mésozoïque, ce qui a provoqué l'ouverture diachronique de l'océan Atlantique du sud au nord. Cette distension, largement asymétrique, a donné des marges non-volcaniques au Canada atlantique et mené à l'exhumation de matériau mantélique serpentinisé sur le fond marin avant l'expansion des fonds océaniques.

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1.0 INTRODUCTION

The Appalachian orogen of eastern North America (Fig. 1) is one of the Earth's classic mountain belts. It was the birthplace of many fundamental geological concepts, including geosynclinal theory (Kay 1951), and 100 years later, the Wilson Cycle and application of plate tectonic principles to orogenic systems (Wilson 1966). In spite of that long history of investigation, many fundamental questions about the orogen have been resolved only in recent years, in large part as a result of studies directly and indirectly linked to the LITHOPROBE East study area of the national LITHOPROBE project (Clowes 2010). During the LITHOPROBE East project and in subsequent years, our view of the orogen changed from relatively simple "accordion-style" tectonic models (e.g. Wilson 1966; Williams 1964, 1979) to complex scenarios involving multi-stage rifting to form a number of separate seaways and oceans, a variety of independent continental and oceanic fragments, and significant orogen-parallel transport of these elements (e.g. van Staal et al. 2009).

This chapter focuses on the Canadian part of the Appalachian orogen, including the adjacent offshore continental margin (Fig. 2). However, the Canadian Appalachians cannot be viewed in isolation because in the Paleozoic they were located in the centre of the orogen, which today continues to the south through New England to the southern Appalachians in the southeastern United States and beyond. Prior to the Mesozoic opening of the Atlantic Ocean, the Appalachians also continued to the northeast into the British and Scandinavian Caledonides (Fig. 1). It is now known that this long and relatively narrow mountain belt was formed principally by the Cambrian (515 Ma) through Permian (275 Ma) closure of not just the Iapetus Ocean and related seaways but also the Rheic Ocean, culminating in Late Paleozoic Alleghanian continental collisions, which formed the supercontinent Pangea (e.g. Wilson 1966; van Staal et al. 1998, 2009; Nance et al. 2008; Hibbard et al. 2010). The northern Appalachians, and the Canadian segment especially, largely escaped the penetrative effects of Alleghanian collision-related deformation, metamorphism, and magmatism because the collision zone was situated mainly offshore from the present coastline. As a consequence, the evidence for earlier accretion-related orogenesis is relatively well preserved in the Canadian segment compared to elsewhere in the Appalachians. Hence, the intrepretations summarized here have been instrumental in providing better understanding of the entire Caledonian–Appalachian orogen, not just the Canadian part (e.g. van Staal et al. 1998, 2009; Hibbard et al. 2006, 2007, 2010).

The LITHOPROBE East project was a major contribution to this enhanced understanding by generating geophysical data to provide insight into crustal architecture in the third dimension, as well as through supporting geoscience projects that provided a wealth of geological information of all types. Deep seismic reflection profiling (approximately 6000 km) and refraction surveys were carried out in the Gulf of St. Lawrence, in the vicinity of the coastline of Newfoundland and on the island itself, both as a direct part of the LITHOPROBE East transect and as contributions to it from the Frontier

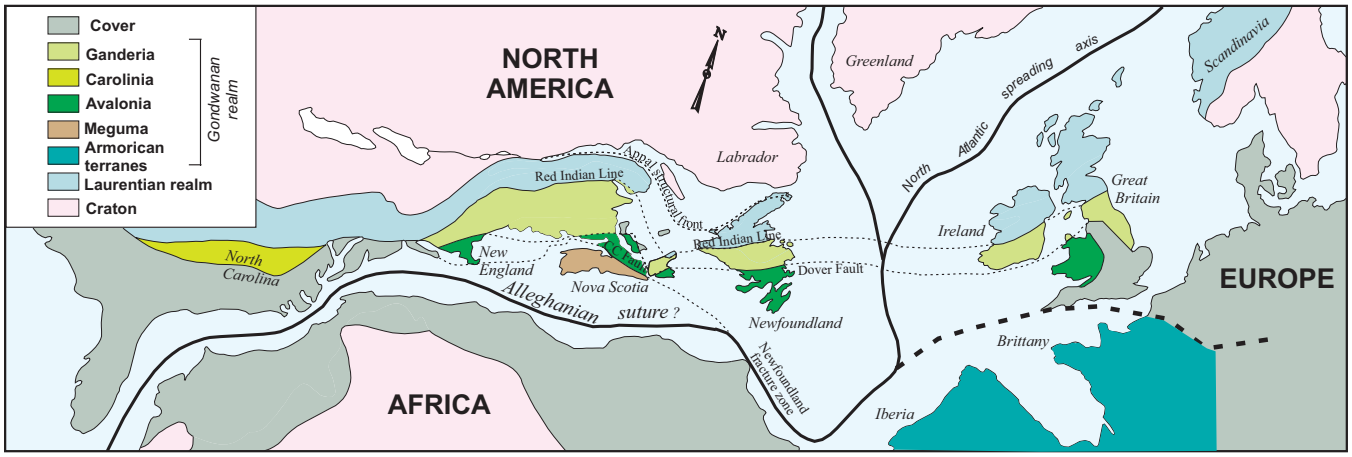


Figure 1. Regional distribution of the main lithotectonic elements of the Appalachian–Caledonian mountain belt in an Early Mesozoic restoration of the North Atlantic region (modified from Williams 1984). Note that the Atlantic Ocean is thought to have opened close to or along the inferred Alleghanian suture between Africa and North America. The dashed line approximately outlines the boundary between composite Laurentia and the Gondwanan elements in Europe (Armorican terranes). Abbreviation CC fault: Cobequid–Chedabucto fault system.

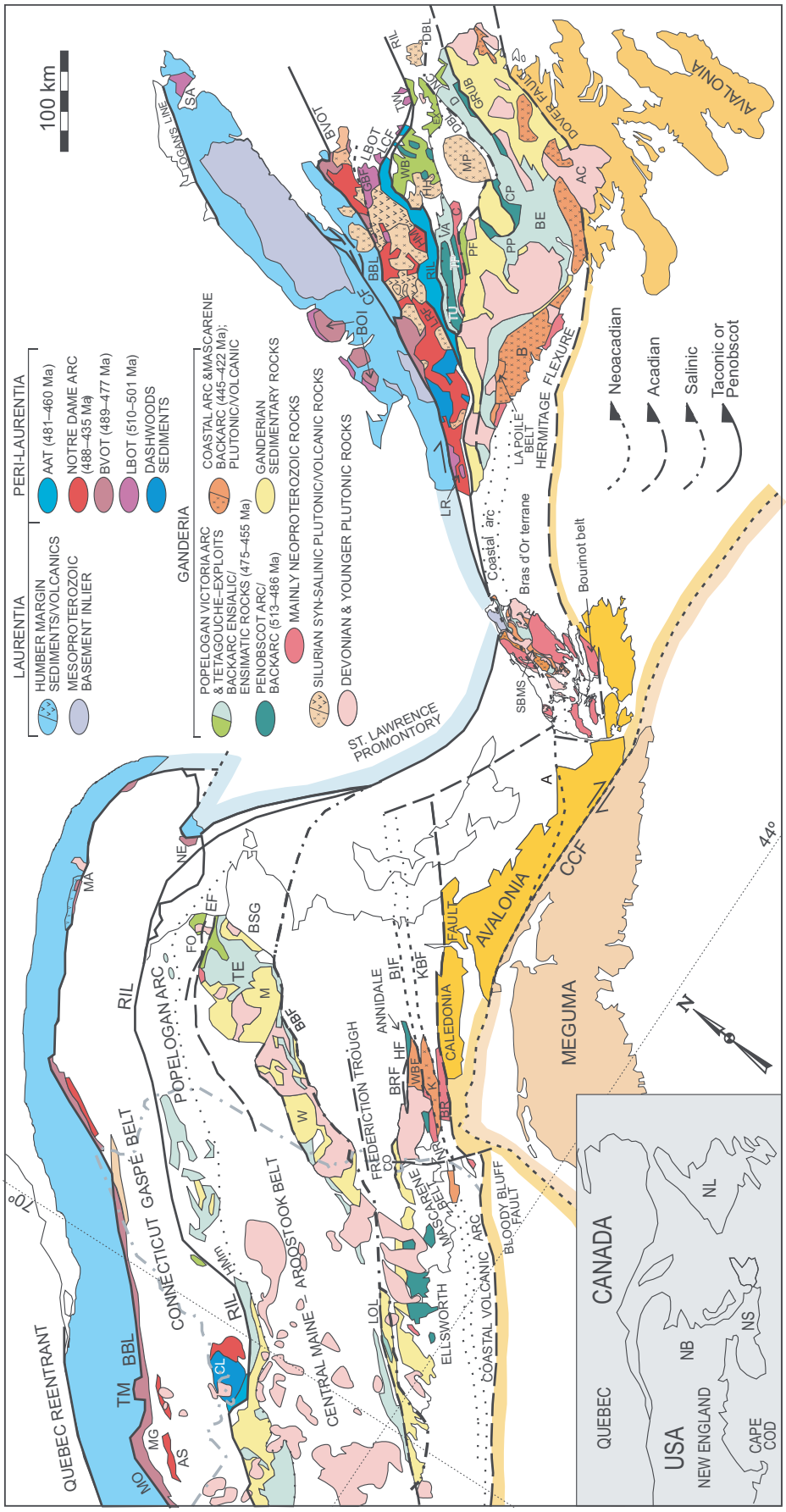


Figure 2. Tectonic map of the Canadian and part of the New England Appalachians with the distribution of the Early Paleozoic lithotectonic elements (coloured) and magmatic belts discussed in the text. The major sutures are shown. Figure modified from van Staal et al. (2009). Abbreviations A: Arisaig Group; AAT: Annieopsquotch accretionary tract; AC: Ackley granite;AN: Annidale belt; AS: Ascott Complex; B: Burgeo batholith; BB; Badger belt; BBF: Bamford Brook fault; BBL: Baie Verte Brompton line; BE: Baie d'Espoir Group; BIF: Belleisle fault; BO: Bocabec pluton; BOIC: Bay of Island Complex; BVOT: Baie Verte oceanic tract; BR: Brookville terrane; BRF: Basswood Ridge fault; BSG: Bathurst Supergroup; C: Cripple Back–Valentine Lake plutons; CC: Coastal Complex; CCF: Cobequid-Chedabucto fault; CF: Cabot fault; CL: Chain Lakes Massif; CO: Cookson Group; CP: Coy Pond Complex; D: Davidsville Group; DBL: Dog Bay line; EF: Elmtree fault; ESZ: Exploits Subzone; EX: Exploits Group; FO: Fournier Group; GBF: Green Bay fault; GF: Guadeloupe fault; GRUB: Gander River ultrabasic belt; GZ: Gander Zone; HF: Hollow fault; HH: Hodges Hill Pluton; HM: Hungry Mountain thrust; HMm: Hurricane Mountain mélange; HZ: Humber Zone; K: Kingston belt; KBF: Kennebacasis fault; LBOT: Lushs Bight oceanic tract; LOL: Liberty-Orrington line; LR: Long Range ultramafic-mafic complex; M: Miramichi Group; MA: Mont Albert Complex; MG: Magog Group; MO: Mount Orford ophiolite Complex; MP: Mount Peyton pluton; NC: Noggin Cove Formation; NE: Neckwick Formation; NDSZ: Notre Dame Subzone; NR: New River belt; PF: Pine Falls Formation;; PP: Pipestone Pond Complex; PT: Pointe aux Trembles Formation; RBF: Rocky Brook-Millstream fault system; RF: Restigouche fault; RIL: Red Indian line; SA: St. Anthony Complex; SBMS: Sarach Brook Metamorphic Suite; TE: Tetagouche Group; TM: Thetford Mines Complex; TP: Tally Pond Group; TU: Tulks Group; TW: Twillingate trondhjemite; U: Utopia granite; VA: Victoria arc; VRF: Victoria River fault;; W: Woodstock Group; WB: Wild Bight Group; WBF: Wheaton Brook fault; WF: Weedon Formation.

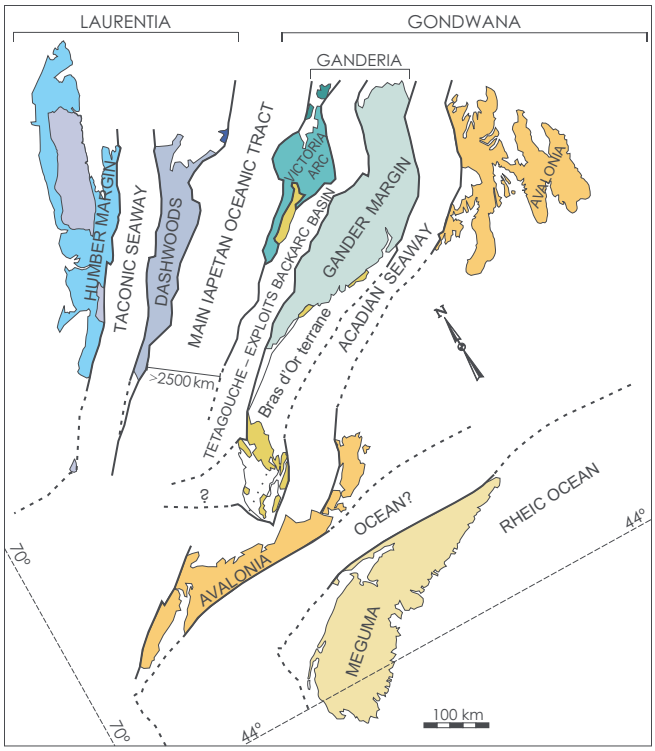


Figure 3. Lithotectonic elements of the Canadian Appalachians schematically expanded to show oceans and seaways that were present outboard of the Humber margin of Laurentia.

Geoscience Program of the Geological Survey of Canada (Marillier et al. 1989, 1994; Stockmal et al. 1987, 1990; Hall et al. 1998; Jackson et al. 1998). The seismic reflection database includes two onshore seismic reflection surveys in Newfoundland, which were done to investigate the continuity of features seen in offshore seismic transects and to extend and link these features to known surface structures (Quinlan et al. 1992; van der Velden et al. 2004).

The geophysical findings of LITHOPROBE East combined with the continuously expanding geological database and evolving ideas have shed new light on the tectonic evolution of the Canadian portion of the northern Appalachians. The purpose of this chapter is to summarize the current understanding of the components, three-dimensional structure, and tectonic evolution of the Canadian Appalachians.

2.0 LITHOTECTONIC COMPONENTS OF THE CANADIAN APPALACHIANS – AGE AND MAKE-UP OF THE MARGINS, ARCS, AND ACCRETED TERRANES

Based on contrasts in lithology, stratigraphic sequence, fauna, structure, metamorphism, geophysical signatures, magmatism, and metallogenic character of Lower Paleozoic and older rocks, Williams (1979) divided the Canadian Appalachians into tectonostratigraphic zones (Humber, Dunnage, Gander, Avalon, and Meguma) and subsequently subzones (e.g. Williams et al. 1988; Williams 1995), separated from one another by major faults. The zonal terminology was useful as a first-order subdivision of the rocks exposed at the surface, but became progressively more difficult to retain when it was realized that some zones comprise distinct tectonic elements, each with a separate Paleozoic geological evolution and separated from adjacent terranes by major faults (e.g. Barr and Raeside 1989; Barr and White 1996; van Staal et al. 1998, 2007; Lissenberg et al. 2005a; Valverde-Vaquero et al. 2006a; Zagorevski et al. 2006, 2007b, 2010; Lin et al. 2007). In addition, seismic reflection data suggest that some of these zones have no roots at depth and instead appear to represent relatively thin allochthonous slices (e.g. van der Velden et al. 2004). Hence, instead of zones, here we use terminology that reflects tectonic setting (Fig. 2), following precedents set by Hibbard et al. (2006, 2007, 2010) and references therein. This approach also facilitates placing the extensive geophysical data set used later on to constrain the third dimension of the Appalachian lithosphere into a tectonic context.

The former Humber zone mainly comprises the remnants of Laurentia's Appalachian margin during the Cambrian and Ordovician and hence is referred to herein as the Humber margin. The former Gander, Avalon, and Meguma zones represent the remnants of three different Gondwana-derived microcontinents and are referred to herein as Ganderia, Avalonia, and Meguma, respectively (Fig. 3, 4). The Dunnage Zone, which is situated between the Humber and Gander zones, mainly contains the remnants of various continental and oceanic arc (*sensu lato*) terranes that prior to their episodic accretion to Laurentia occurred as ribbons in the Iapetus Ocean, either on the Laurentian side (peri-Laurentian tectonic elements) or on the Ganderian side (peri-Ganderian tectonic elements) (Figs. 2, 3, 4). As a result of these accretions, ultimately including Avalonia and Meguma, Laurentia progressively expanded eastwards to form what is referred to herein as composite Laurentia. We begin by summarizing the distinguishing features of these tectonic elements as they are currently understood, moving from northwest to southeast across the orogen. Post-orogenic rifting in the Mesozoic opened the present-day Atlantic Ocean more or less along or close to the Late Paleozoic Alleghanian suture (Fig. 1), and thus the

original eastern (Gondwanan) margin is no longer part of the orogen. Today, all of the Canadian Appalachian orogen consists of lithotectonic elements that were already part of composite Laurentia prior to its Carboniferous–Permian Alleghanian collision with Gondwana (Africa).

2.1 Humber Margin

The Humber margin represents the eastern (present coordinates) margin of Laurentia in the Early Paleozoic and underlies most of western Newfoundland and southern Quebec (Figs. 1, 2). It contains the remnants of deformed upper Neoproterozoic to Ordovician sedimentary and minor volcanic rocks deposited on Mesoproterozoic crystalline basement of the Grenville structural province of Laurentia (e.g. Heaman et al. 2002). The cover rocks are related to rifting, passive margin development, and formation of a foreland basin (Williams and Hiscott 1987; Knight et al. 1991; Waldron et al. 1998; Lavoie et al. 2003). Mesoproterozoic basement is mainly exposed in western Newfoundland in a series of structural inliers, some of which, such as the Steel Mountain Inlier, have only a Pinwarian (ca. 1.5 Ga) orogenic history and lack a Grenvillian magmatic or metamorphic overprint, suggesting the presence of strike-slip-generated allochthonous basement terranes along the Humber margin (Brem 2007; Brem et al. 2007). The Humber margin between Newfoundland and Quebec has a sinuous geometry (Figs. 1, 2), which is generally interpreted to reflect a promontory-reentrant pair known as the St. Lawrence promontory and Quebec reentrant, formed during rifting (Thomas 1977, 2006). The Blair River Inlier of northern Cape Breton Island currently occupies the tip of the St. Lawrence promontory (Figs. 1, 2). Like the Steel Mountain Inlier, it appears to have a Pinwarian connection, but unlike the Steel Mountain Inlier, it experienced subsequent Grenvillian as well as Lower Paleozoic thermal overprints (Miller et al. 1996; Miller and Barr 2000, 2004). The western boundary of the Humber margin approximately coincides with the Appalachian structural front, known as Logan's line (Figs. 1, 2), although autochthonous little- or undeformed Cambrian–Ordovician platformal rocks extend much further west (Hibbard et al. 2006). The eastern exposed limit of the Humber margin is defined by a complex and long-lived, steeply dipping fault zone marked by mélange and numerous discontinuous ophiolitic fragments of various sizes and known as the Baie Verte–Brompton line (Fig. 2; Williams and St. Julien 1982). The Humber margin extends further to the east at depth (e.g. Hall et al. 1998; van der Velden et al. 2004).

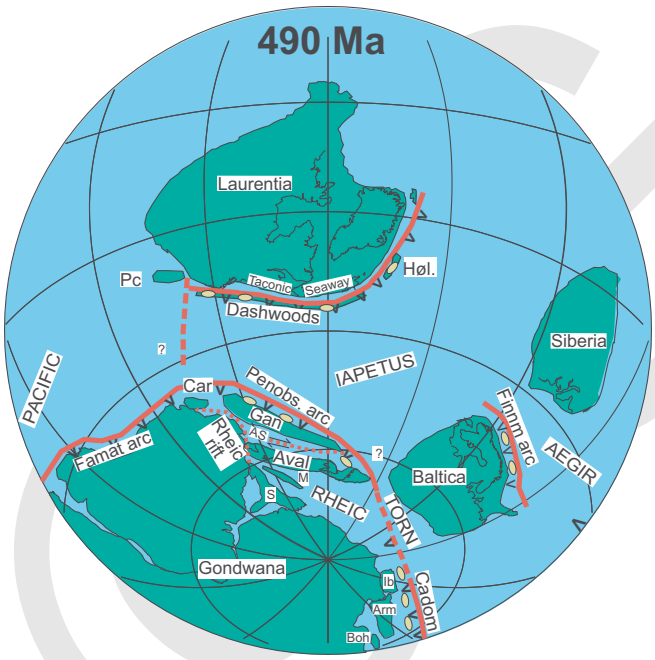


Figure 4. Distribution of the main continents, oceans, subduction zones, arc–back-arc systems, and microcontinents at the end of the Late Cambrian. Figure is modified from van Staal and Hatcher (2010). Abreviations Aegir: Aegir Sea; Arm: Armorican terrane; Aval: Avalonia; AS: Acadian seaway; Boh: Bohemia; Cadom: Cadomia; Car: Carolinia; Famat arc: Famatinian arc–back-arc system; Finnm arc: Finnmark arc–back-arc system; Gan: Ganderia; Hol: Holanda terrane; Ib: Iberia; M: Meguma; Pc: Precordillera; Penobs. Arc: Penobscot arc–back-arc system; S: Suwanee terrane. TORN: Tornquist ocean.

The Humber margin was initiated as a rift, characterized by tholeiitic mafic magmatism and siliciclastic sediments deposited in fault-bounded grabens, by at least 615 Ma (Kamo et al. 1989). Rift-related magmatism, associated with opening of the Iapetus Ocean and Taconic Seaway, lasted intermittently until ca. 550 Ma (Figs. 3, 4). Shortly thereafter (540–530 Ma), rift-related magmatic activity was followed by deposition of a Lower Cambrian transgressive sequence (Fig. 5), generally interpreted to represent a rift-drift transition (Williams and Hiscott 1987; Lavoie et al. 2003). Waldron and van Staal (2001) and Hibbard et al. (2007) related this event to separation of microcontinental ribbons (e.g. Dashwoods) from Laurentia, implying that the Iapetus Ocean had opened earlier (ca. 570 Ma) as a separate event, because paleomagnetic and other tectonic evidence suggest that the Iapetus Ocean had already achieved a significant width by 540 Ma (Cawood et al.

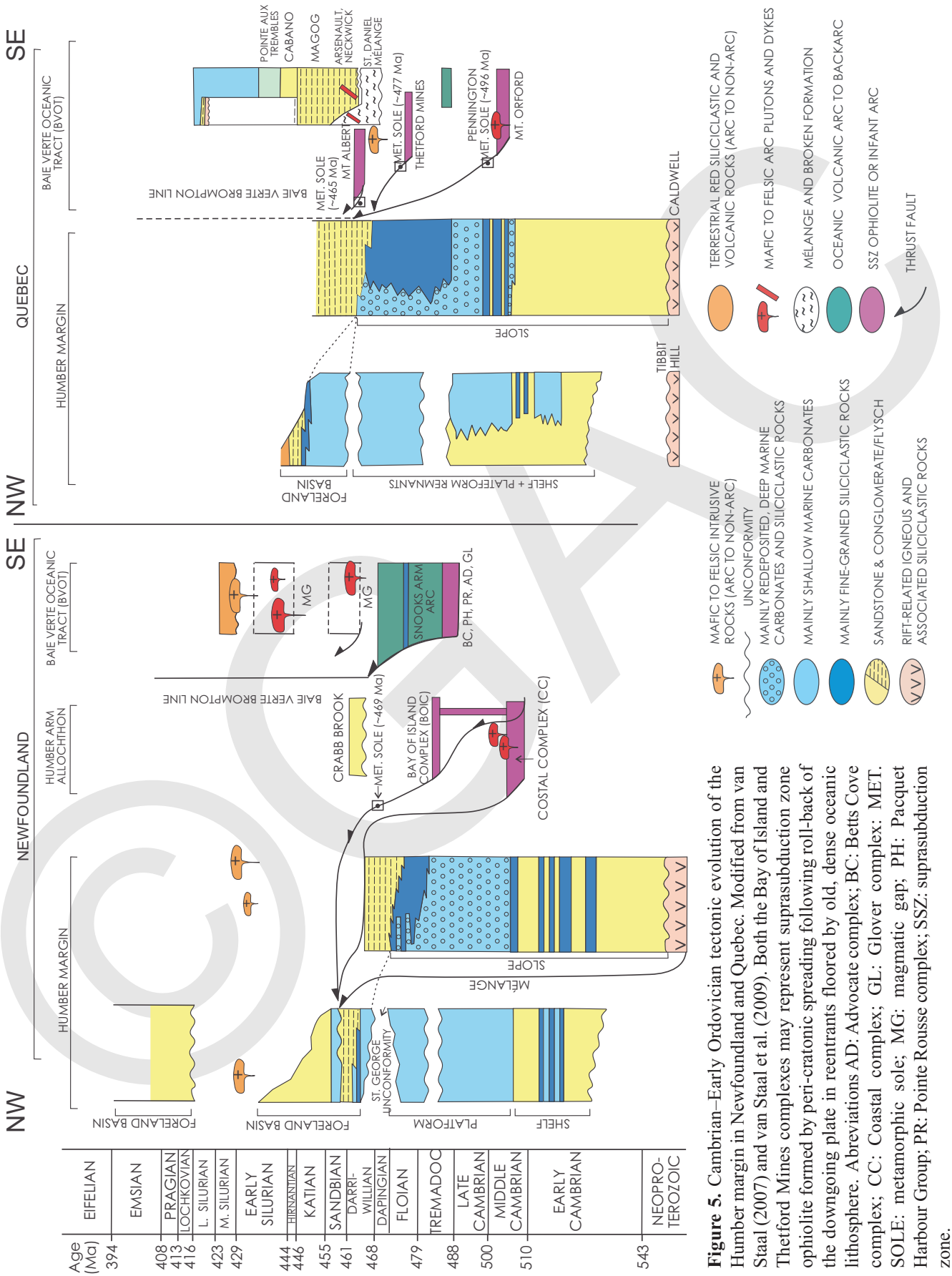


Figure 5. Cambrian–Early Ordovician tectonic evolution of the Humber margin in Newfoundland and Quebec. Modified from van Staal (2007) and van Staal et al. (2009). Both the Bay of Island and Thetford Mines complexes may represent suprasubduction zone ophiolite formed by peri-cratonic spreading following roll-back of the downgoing plate in reentrants floored by old, dense oceanic lithosphere. Abreviations AD: Advocate complex; BC: Betts Cove complex; CC: Coastal complex; GL: Glover complex: MET. SOLE: metamorphic sole; MG: magmatic gap; PH: Pacquet Harbour Group; PR: Pointe Rousse complex; SSZ: suprasubduction zone.

2001). Tholeiitic gabbroic dykes in the Grenvillian Blair River Inlier of Cape Breton Island with U-Pb zircon ages of 581±6 and 576±6 Ma, as well as scattered units of similar age preserved along the Humber margin in Quebec, likely reflect the initial rifting and opening of the Iapetus Ocean (e.g. Miller and Barr 2004).

A Lower Cambrian siliciclastic shelf with local carbonate deposits was formed concurrently with the early stages of seafloor spreading in the Taconic Seaway (Figs 3, 4, 5), which remained a relatively narrow arm of Iapetus, separating the Humber margin from the Dashwoods ribbon (see below). The clastic shelf was followed by a carbonate platform at ca. 510 Ma (James et al. 1989), which remained in place until its tectonic destruction during the Taconic orogeny (Figs. 3, 4). The slope of the passive margin was characterized by clastic sediments, rudaceous carbonate rocks, and pelagic shales. The shallow-marine platform segment of the passive margin is poorly preserved in Quebec (Fig. 5), where mainly siliciclastic rocks deposited on the slope and rise have been preserved (Lavoie et al. 2003).

Between 480 and 470 Ma, the passive Humber margin (Fig. 5) was progressively converted into a convergent margin (Knight et al. 1991). The earliest evidence of tectonism on the slope segment of the Humber margin may date back to ca. 480 Ma in Newfoundland. This early tectonism was probably responsible for initiating transgression and seismically generated mass flows on the continental slope. Progressive loading of the margin created a foreland migrating peripheral bulge (Jacobi 1981; Knight et al. 1991). Evidence for a peripheral bulge is provided by diachronous uplift and local karst erosion (Jacobi 1981), and formation of a marine foreland basin immediately after its passage (Fig. 5). Tectonic loading of the outboard part of the passive margin and formation of a marine foreland basin appear to have started slightly later, at ca. 468 Ma, and may have lasted longer in the Quebec reentrant (Fig. 4, Hiscott 1978; Malo et al. 2001; Lavoie et al. 2003). By 455 Ma, all peri-Laurentian tectonic elements (discussed below) had been fully assembled with the Humber margin (van Staal et al. 2007) into the first phase of composite Laurentia. Thus the Humber margin by this time had finished being an ocean-facing Atlantic margin, because Laurentia's leading edge then lay further outboard and represented the ocean-facing margin of the accreted tectonic elements.

2.2 Peri-Laurentian Tectonic Elements

A collage of spatially associated Early Paleozoic terranes, which include the Lushs Bight and Baie Verte oceanic tracts, Dashwoods microcontinent, and Annieopsquotch accretionary tract, lies tectonically on or immediately adjacent to the Humber margin (Figs. 2, 3, 4, 6). These terranes correspond to the Notre Dame subzone of the Dunnage zone of Williams et al. (1988) and are best exposed and/or preserved in Newfoundland. The descriptions given below therefore rely heavily on relationships established here.

Paleomagnetic data and faunal provinciality suggest that these terranes originated close to Laurentia, albeit in different tectonic settings. They were assembled together piecemeal into one composite terrane during the Taconic orogenic cycle (see below), which was finished by at least 450 Ma. However, most of these terranes were already assembled together by 467 Ma (Lissenberg et al. 2005a) and formed the basement on which the voluminous second phase (467–459 Ma) of the Notre Dame arc was built (van Staal et al. 2007).

2.2.1 Lushs Bight Oceanic Tract

The Lushs Bight oceanic tract (LBOT) corresponds largely to the Twillingate subzone of Williams (1995). It mainly consists of Middle to Upper Cambrian (510–501 Ma) predominantly mafic volcanic and plutonic rocks (Figs. 2, 6) (Elliot et al. 1991; Szybinski 1995; Swinden et al. 1997), which are well exposed in north-central Newfoundland (Fig. 2). Correlative rocks occur as isolated structural outliers in western, northern, and southern Newfoundland, and may also occur in the eastern townships of Quebec (Fig. 2) (Huot et al. 2002; van Staal et al. 2007). The LBOT in Newfoundland mainly comprises an association of pillow basalt, sheeted diabase dykes, gabbro, and rare ultramafic rocks (Kean et al. 1995), which indicate oceanic, ophiolitic affinity. Consanguineous intrusive bodies of juvenile trondhjemite (e.g. Twillingate trondhjemite, Williams and Payne 1975; Fryer et al. 1992) and diorite and the abundance of boninite and primitive island arc tholeiite (Swinden 1996; Swinden et al. 1997) suggest that the LBOT represents suprasubduction-zone oceanic lithosphere, probably an infant arc terrane (van Staal et al. 1998, 2007). Geochemical evidence suggests that subducted Laurentia-derived sediment was involved in the petrogenesis of the LBOT (Swinden et al. 1997; van Staal et al. 2007), consistent with formation near the Laurentian margin. Field relationships, structures, and isotopic evidence from crustally contaminated cross-cutting diabase and granitoid dykes suggest that the LBOT was deformed and emplaced onto Laurentia-derived continental crust (Dashwoods, see below) between 500

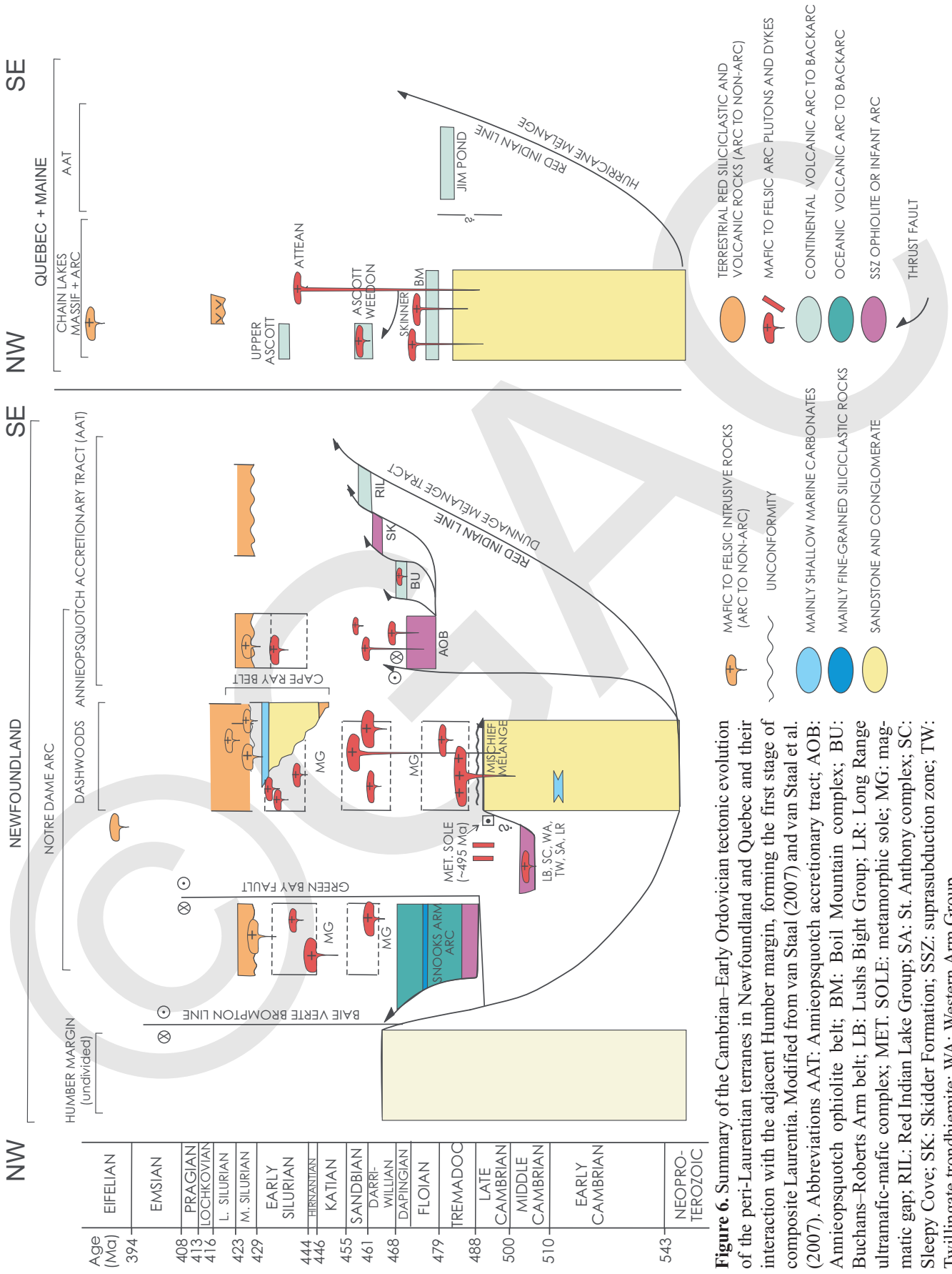


Figure 6. Summary of the Cambrian–Early Ordovician tectonic evolution of the peri-Laurentian terranes in Newfoundland and Quebec and their interaction with the adjacent Humber margin, forming the first stage of composite Laurentia. Modified from van Staal (2007) and van Staal et al. (2007). Abbreviations AAT: Annieopsquotch accretionary tract; AOB: Annieopsquotch ophiolite belt; BM: Boil Mountain complex; BU: Buchans–Roberts Arm belt; LB: Lushs Bight Group; LR: Long Range ultramafic-mafic complex; MET. SOLE: metamorphic sole; MG: magmatic gap; RIL: Red Indian Lake Group; SA: St. Anthony complex; SC: Sleepy Cove; SK: Skidder Formation; SSZ: suprasubduction zone; TW: Twillingate trondhjemite; WA: Western Arm Group.

and 490 Ma (Szybinski 1995; Swinden et al. 1997; van Staal et al. 2009).

2.2.2 Baie Verte Oceanic Tract

The Baie Verte oceanic tract (BVOT) mainly occurs as a narrow fault-bounded wedge along the Baie Verte–Brompton line (Figs. 2, 5) in both Newfoundland and Quebec. It stretches from the north coast of the Baie Verte peninsula to south of Grand Lake in west-central Newfoundland and from Gaspe into the eastern townships and beyond into New England (Figs. 1, 2). Like the LBOT, it comprises mainly suprasubduction-zone ophiolitic rocks of mafic to ultramafic composition, commonly with boninite. However, it is significantly younger than the LBOT (Fig. 2) and underwent a markedly different geological history (see below). Four of its ophiolite bodies in Newfoundland yielded identical ages of ca. 490 Ma (Dunning and Krogh 1985; Cawood et al. 1996; Skulski et al. 2010), whereas ophiolites in southern Quebec are ca. 10 my younger (Whitehead et al. 2000), probably due to the influence of the Quebec reentrant (van Staal et al. 2007; Zagorevski and van Staal 2011) During the Lower Ordovician (480–467 Ma), ophiolite of the BVOT in Newfoundland formed basement to a volcanic-sedimentary complex characterized by an association of E-MORB and calc-alkaline intermediate to felsic volcanic rocks (Skulski et al. 2010), which has been referred to as the Snooks Arm arc (Bédard et al. 2000) to emphasize its upper plate setting, although it does not represent a typical arc setting.

2.2.3 Dashwoods Microcontinent

Dashwoods (Figs. 2, 3, 4, 6) is a ribbon-shaped peri-Laurentian microcontinent (Waldron and van Staal 2001). Its crustal basement is not exposed, but has been detected at depth based on isotopic evidence (Swinden et al. 1997; Whalen et al. 1997a) and inherited zircon grains in Lower Ordovician (490–476 Ma) arc plutons of the Notre Dame arc (e.g. Dubé et al. 1996; van Staal et al. 2007). Today, the Dashwoods microcontinent is 25–50 km wide and approximately 400 km long in Newfoundland. It is separated from the BVOT to the northwest by the Geen Bay–Little Grand Lake fault system (Brem et al. 2007; van Staal et al. 2007;) and from the Annieopsquotch accretionary tract to the east (Figs. 2, 6) by the Lloyd's River–Hungry Mountain–Lobster Cove fault system (Lissenberg and van Staal 2006; Zagorevski et al. 2008, 2009; van Staal et al. 2009).

Dashwoods was interpreted as a peri-Laurentian microcontinent rather than the leading edge of Laurentia on the basis of its Early Paleozoic geological evolution, which is markedly different from that of the Humber margin (Waldron and van Staal 2001; van Staal et al. 2007), from which it is separated by oceanic rocks of the BVOT. The LBOT was obducted above the Dashwoods microcontinent; hence, they were assembled together into a composite terrane by at least 493 Ma (see below), before formation of the BVOT. This new composite terrane formed the basement to the 490–433 Ma phases of the Notre Dame arc (Whalen et al. 2006; van Staal et al. 2007). The oldest known rocks in the Dashwoods microcontinent are Late Neoproterozoic–Cambrian, strongly metamorphosed and locally migmatized siliciclastic rocks, which have been correlated traditionally with rocks deposited on the Humber margin during late Neoproterozoic–Cambrian rifting and subsequent construction of a passive margin there (Waldron et al. 1998; Waldron and van Staal 2001 and references therein).

Based on isotopic studies (Tremblay et al. 1994), an equivalent peri-Laurentian continental ribbon terrane along strike also has been detected in the subsurface of southern Quebec; it is exposed mainly in northern Maine in the Chain Lakes massif (Gerbi et al. 2006a). The presence of Lower to Middle Ordovician arc plutons in the Chain Lakes massif (Gerbi et al. 2006b) makes unlikely the model of de Souza and Tremblay (2010) that the root of the Quebec ophiolites lies east of the Chain Lakes massif. Based on similarity in timing of the rift-drift transition along the length of the Humber margin, a peri-Laurentian continental ribbon terrane has been postulated to have been present along the whole length of the Appalachian orogen (Hibbard et al. 2007; Allen et al. 2010).

2.2.4 Annieopsquotch Accretionary Tract

The Annieopsquotch accretionary tract (AAT) is sandwiched between the Lloyd's River–Hungry Mountain–Lobster Cove fault system and the Red Indian line (Figs. 2, 6) (Colman-Sadd et al. 1992b; van Staal et al. 1998; Lissenberg et al. 2005a; Zagorevski et al. 2007b, 2009). The AAT comprises a collage of tectonic slices of 480–473 Ma suprasubduction zone ophiolite (e.g. Annieopsquotch ophiolite belt, Lissenberg et al. 2005b) as well as slices with arc and back-arc rocks (e.g. Buchans-Robert Arm belt, Swinden et al. 1997; Zagorevski et al. 2006) that formed outboard of Dashwoods (Zagorevski et al. 2006, 2009). Ultramafic rocks are rare in the AAT (Lissenberg et al. 2006), probably because accretion to Dashwoods involved under-

thrusting and structural underplating rather than obduction (Lissenberg and van Staal 2006; Zagorevski et al. 2009). The ultramafic parts of the accreted ophiolite slices were probably delaminated from the crustal underplate and remained attached to the down-going slab. The AAT is not exposed in the eastern townships of Quebec, but correlative elements occur in northern Maine in the Boil Mountain and Jim Pond complexes (Figs 2, 6) (Gerbi et al. 2006b) and remnants have also been preserved as clasts in Ordovician mélange and conglomerate in Gaspé (Dupuis et al. 2009).

2.2.5 Notre Dame Magmatic Arc

The Notre Dame magmatic arc in Newfoundland existed intermittently from ca. 488–435 Ma and is represented by three major magmatic pulses separated by two significant gaps, which correlate with collisional events (Fig. 6, van Staal et al. 2007). The three phases of the Notre Dame arc are related to kinematically and/or dynamically distinct tectonic events (discussed later). They have been grouped under the same name because they formed more or less in the same location. The first phase is represented by granodiorite to diorite plutons and associated volcanic rocks, which range in age from ca. 488 to 476 Ma (Szybinski 1995; van Staal et al. 2007). Isotopic evidence and inherited zircons (Whalen et al. 1987, 1997a, b; Dubé et al. 1996) suggest that these plutons came up through continental crust. Following an 8 to 10 Ma gap in magmatic activity (MG in Fig. 6), the second phase of the Notre Dame arc peaked from 467 to 459 Ma with a major flare-up of tonalite (van Staal et al. 2007). The third phase is represented by calc-alkaline gabbro to granodiorite. Related volcanic rocks lie unconformably on rocks of the second phase. The third phase lasted from 445 to 435 Ma (Whalen 1989) and was followed by emplacement of a 433–429 Ma mixed arc/non-arc-like bimodal magmatic suite inferred to have been related to slab-breakoff (Whalen et al. 1996, 2006). This phase is related to a later orogenic cycle (Salinic, see below). Equivalents of the Notre Dame arc in Quebec are poorly preserved, probably because many of them are buried beneath the Silurian–Devonian cover sequences of the Gaspé belt (Fig. 1). Volcanic and plutonic rocks of the Ascot–Weedon continental arc belt (Tremblay 1992) are mainly correlatives of the second phase of the Notre Dame arc. The third phase is represented by 435–432 Ma andesite of the Pointe aux Trembles Formation and associated calc-alkaline intrusive rocks (Fig. 5) in Matapedia, Quebec, near the border with New Brunswick (David and Gariepy 1990), and coeval calc-alkaline volcanic rocks in the Chaleurs Group in northern New Brunswick (Wilson et al. 2008).

2.3 Ganderia

Ganderia (Figs. 1, 2, 3, 7) comprises several distinct, but tectonically related Paleozoic terranes, which may initially have formed part of a single, Gondwana-derived Early Paleozoic microcontinent. The Ganderian terranes are linked by a common long-lived Proterozoic history that is well preserved in southern New Brunswick, central Cape Breton Island, and locally in southern Newfoundland (Raeside and Barr 1990; Johnson and McLeod 1996; White and Barr 1996; Barr et al. 1998; Johnson et al. 2009), and by a distinctive Late Neoproterozoic–Early Paleozoic tectonic evolution (van Staal et al. 1996, 1998, 2009; Barr et al. 2003a; Rogers et al. 2006; Lin et al. 2007; Fyffe et al. 2009) that is absent in Avalonia and Meguma. Faunal, paleomagnetic, and zircon provenance studies suggest that Ganderia represents a continental fragment that rifted from the Amazonian margin (Fig. 4) of Gondwana between 505 and 495 Ma (van Staal et al. 1996, 2009; Schultz et al. 2008). Its trailing edge became the site of a passive margin (Gander margin, van Staal 1994), whereas its leading edge saw development of two successive arc–back-arc systems (see below) separated by the 485–478 Ma Penobscot orogenic event. Remnants of these peri-Ganderian arc–back-arc tectonic elements form most of the Exploits subzone of Williams et al. (1988), whereas the Gander margin corresponds to the bulk of the original Gander zone of Williams (1979) (Fig. 7).

2.3.1 Penobscot and Popelogan–Victoria Arcs

The Penobscot and Popelogan–Victoria arcs (Figs. 2, 3, 7) were peri-Ganderian arc–back-arc systems that were active at 515–485 Ma and 475–455 Ma, respectively, and formed at or near Ganderia's leading edge (van Staal et al. 1996; Ayuso and Schultz 2003; Ayuso et al. 2003; Schultz and Ayuso 2003; Wilson 2003; Valverde-Vaquero et al. 2006a; Zagorevski et al. 2007c, 2010). In central and southern Newfoundland, these rocks are sandwiched between the Red Indian line in the west and the GRUB line–Day Cove fault system (Figs.1, 7; van der Velden et al. 2004) to the east. The Penobscot and Popelogan–Victoria arcs are also well exposed in central Newfoundland (van Staal et al. 1998; Rogers et al. 2006; Valverde-Vaquero et al. 2006a; Zagorevski et al. 2007c, 2010), southern New Brunswick (Fyffe et al. 2009; Johnson et al. 2009), northern new Brunswick (Rogers et al. 2003 a,b; van Staal et al. 2003;), and Maine (van

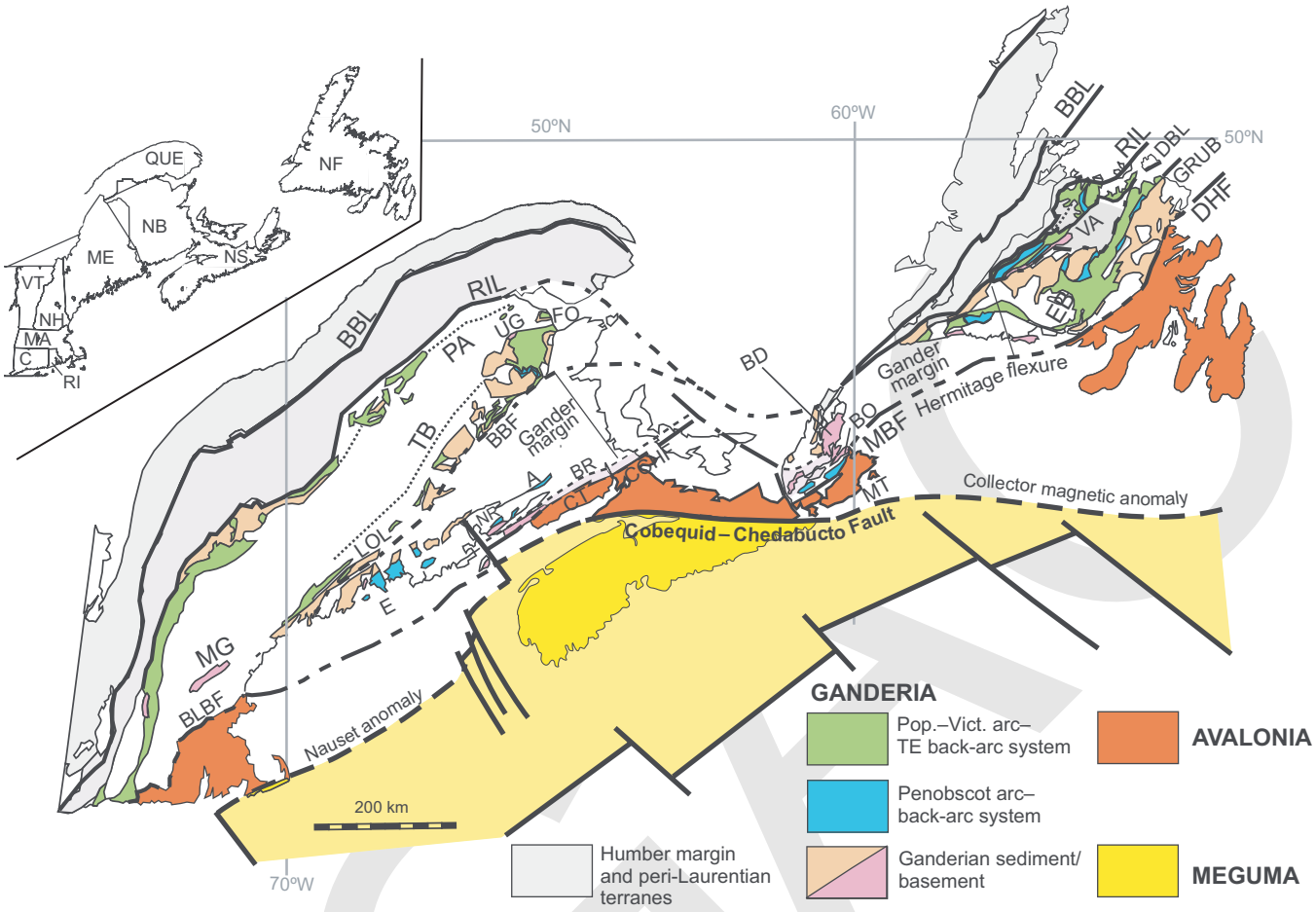
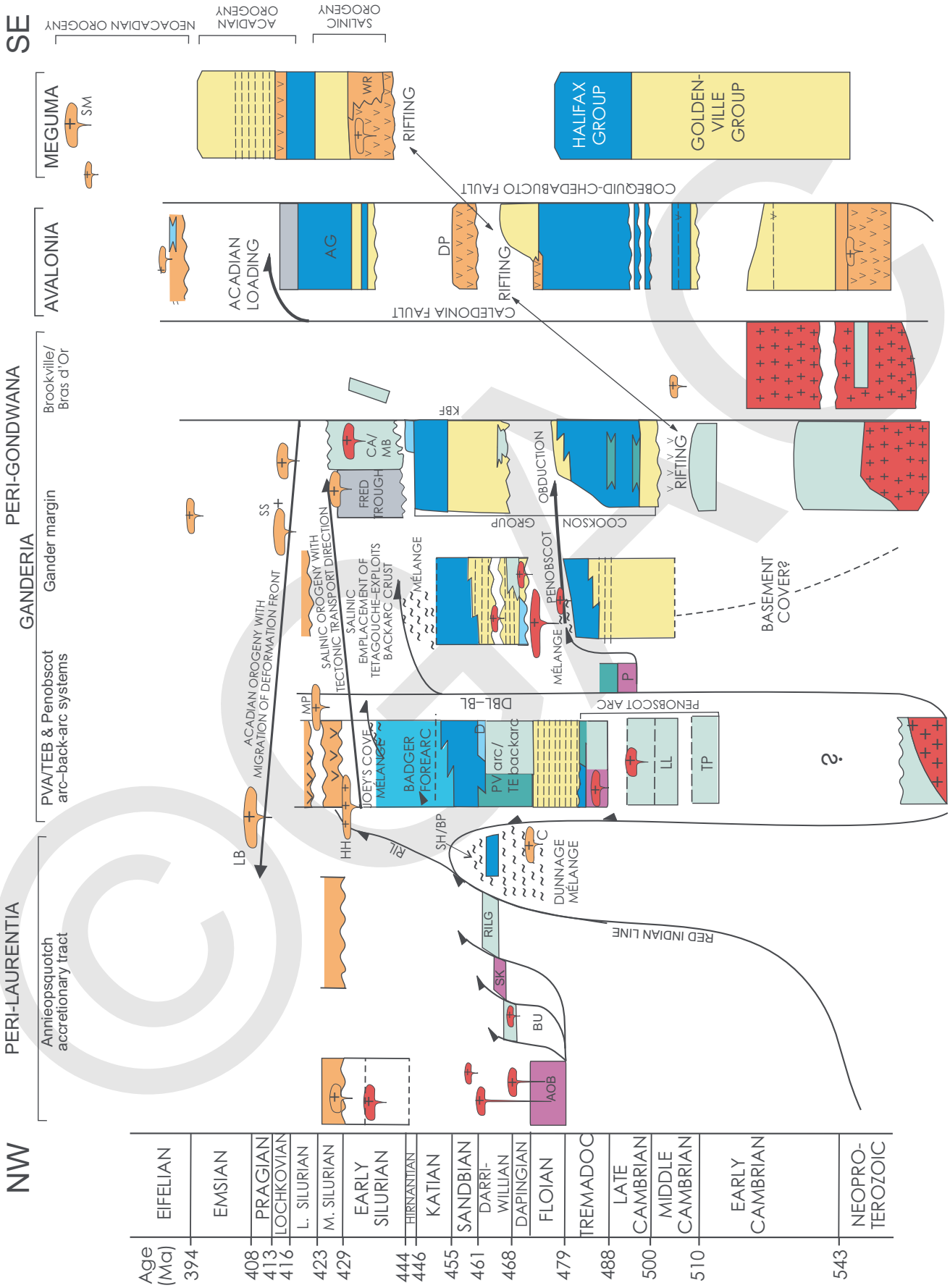


Figure 7. Map showing the extent of Ganderia, Avalonia, and Meguma onshore and offshore in the northern Appalachians, after Hibbard et al. (2006), Hutchinson et al. (1988), Keen et al. (1991a,b), Pe-Piper and Loncarevic (1989), Pe-Piper and Jansa (1999), and White (2010). The regional distribution of the Popelogan–Victoria (Pop.–Vict.) arc – Tetagouche–Exploits (TE) back arc and Penobscot arc–back-arc systems are outlined in green and blue respectively. Post-Ordovician cover sequences and plutons have been left white in Ganderia. Abreviations A: Annidale belt; BBF: Bamford Brook fault; BBL: Baie Verte–Brompton line; BD: Bras d'Or terrane; BLBF: Bloody Bluff fault; BO:Bourinot belt; BR: Brookville; CCHF: Caledonia–Clover Hill fault; CT: Caledonia terrane; E: Ellsworth belt; DBL: Dog Bay line; DHF: Dover–Hermitage Bay fault; EB: Exploits back-arc basin; FO: Fournier ophiolite; GRUB: Gander River ultrabasic belt; LOL: Liberty–Orrington line; MBF: MacIntosh Brook fault; MG: Massabesic gneiss; MT: Mira terrane; NR: New River belt; PA: Popelogan arc; RIL: Red Indian line; TB: approximate trend of Tetagouche back-arc rocks; UG: ca. 550 Ma Upsalquitch gabbro; VA: Victoria arc.

Staal et al. 1998; Schultz and Ayuso 2003; van Staal 2007) (Figs. 1, 5). Both arcs are represented by mafic to felsic volcanic rocks, associated subvolcanic intrusions and minor sedimentary rocks generally of low metamorphic grade. They were largely built on late Neoproterozoic–Early Cambrian (620–530 Ma) arc basement, in both Newfoundland and New Brunswick (Figs. 7, 8), comprising subduction-related plutonic and volcanic rocks (van Staal et al. 1996, 2003; Barr et al. 2003a; Rogers et al. 2006; Zagorevski et al. 2007c, 2010; Johnson et al. 2009). The two Early Paleozoic arc systems, although spatially closely associated and largely built on the same basement, are traditionally named differently, because they were separated by a short-lived orogenic phase (Penobscot orogeny, see below)

2.3.2 Gander Margin

The Gander margin (Figs. 2, 3, 7) largely corresponds to the Gander Zone of Williams (1979). It was originally defined on the basis of a distinctive sequence of Lower Cambrian to Lower Ordovician arenite, siltstone and shale, generally considered to represent the outboard part (outer shelf to slope) of a passive margin (van Staal 1994). This sedimentary sequence can be traced from northeastern Newfoundland into New Brunswick and Maine and has been called various names: Gander Group and Spruce Brook Formation in eastern and cen-


NW
SE
PERI-LAURENTIA
Annieopsquotch accretionary tract
PERI-GONDWANA
GANDERIA
PVA/TEB & Penobscot arc–back-arc systems
Gander margin
Brookville/ Bras d'Or
AVALONIA
MEGUMA
NEOACADIAN OROGENY
ACADIAN OROGENY
SALINIC OROGENY
Age (Ma)
EIFELIAN
EMSIAN
PRAGIAN
LOCHKOVIAN
L. SILURIAN
M. SILURIAN
EARLY SILURIAN
HIRNANTIAN
KATIAN
SANDBIAN
DARRI-WILLIAN
DAPINGIAN
FLOIAN
TREMADOC
LATE CAMBRIAN
MIDDLE CAMBRIAN
EARLY CAMBRIAN
NEOPRO-TEROZOIC
394
408
413
416
423
429
444
446
455
461
468
479
488
500
510
543
ACADIAN OROGENY WITH MIGRATION OF DEFORMATION FRONT
SALINIC OROGENY WITH TECTONIC TRANSPORT DIRECTION
SALINIC EMPLACEMENT OF TETAGOUCHE–EXPLOITS BACKARC CRUST
ACADIAN LOADING
CALEDONIA FAULT
COBEQUID-CHEDABUCTO FAULT
RIFTING
HALIFAX GROUP
GOLDEN-VILLE GROUP
WR
AG
DP
SM
SS
LB
MP
HH
RIL
SH/BP
KBF
FRED TROUGH
CA/ MB
GROUP
COOKSON
OBDUCTION
MÉLANGE
PENOBSCOT
BASEMENT COVER?
DBL–BL
PENOBSCOT ARC
JOEY'S COVE MÉLANGE
BADGER FOREARC
D
PV arc/ TE backarc
P
LL
TP
?
DUNNAGE MÉLANGE
C
RILG
SK
BU
AOB
RED INDIAN LINE

tral Newfoundland, respectively, and Miramichi, Woodstock, and Cookson groups in northern, eastern, and southern New Brunswick and Maine, respectively. In Maine it also includes the Grand Pitch Formation and units in the Benner Hill sequence of Tucker et al. (2001). These sequences are separated from Avalonia by the Dover–Hermitage Bay–Caledonia fault system (Figs. 1, 2, 7). Detrital zircon studies have shown that Ganderian sandstones of the Miramichi, Woodstock, and Cookson groups, situated on both sides of the Bamford Brook–Liberty Orrington line in New Brunswick, are approximately coeval and have similar provenance (Fyffe et al. 2009). Detrital zircon data combined with field relationships and sparse fossils indicate an Early Cambrian to Early Ordovician age (520–480 Ma); the Upper Cambrian–Lower Ordovician top of the sequence is mainly represented by dark grey or black shale (van Staal and Fyffe 1995; Tucker et al. 2001; van Staal et al. 2003). Field relationships and detrital zircon and titanite data indicate a similar age range in Newfoundland (O'Neill 1991; Colman-Sadd et al. 1992a). This distinctive sedimentary sequence is considered the geological calling card of Ganderia.

The Gander margin was deposited on Neoproterozoic–Lower Cambrian subduction-related igneous and sedimentary rocks, themselves built on an older passive margin (see below). Formation of the Gander margin was coeval with construction of the Penobscot arc on Ganderia's leading edge, which faced the Iapetus Ocean. Hence, the Gander margin developed on Ganderia's trailing edge facing Gondwana and/or on the passive side of the Penobscot back-arc basin (van Staal 1994; Zagorevski et al. 2007c, 2010). Rifting was coeval with Penobscot arc activity at Ganderia's leading edge (Fig. 4). Ganderia's break-up and departure from Gondwana thus may have started as a back-arc rift induced and/or aided by slab roll-back. The Cambrian arenites of the Ellsworth belt (Figs. 2, 7) have detrital zircon populations very similar to those in coeval arenites deposited elsewhere in the Gander margin (Fyffe et al. 2009). Hence, the Gander margin sequences probably formed a nearly continuous blanket of siliciclastic sediment that covered most of Ganderia's trailing edge. A short period of loading and deformation of the Gander margin took place during the Early Ordovician (486–478 Ma) Penobscot orogeny (Colman-Sadd et al. 1992a; Zagorevski et al. 2007c, 2010; Johnson et al. 2009) (Fig. 8), as described below. The Gander margin was resurrected after the construction of the Popelogan–Victoria arc and opening of the Tetagouche–Exploits back-arc basin at ca. 475 Ma, largely on the ruins of the collapsed Penobscot orogen (van Staal 1994). The second phase of margin development lasted until the Early Silurian (Fig. 8). Subsequent tectonic loading associated with the Salinic orogeny (described below) created a foreland basin (Fredericton trough) (Figs. 2, 8), which was deformed together with the underlying Middle–Upper Ordovician arenite and shale during terminal Salinic orogenesis at the end of the Silurian (West et al. 1992; van Staal 1994).

Gander margin basement mainly comprises Neoproterozoic to lower Cambrian (620–515 Ma) subduction-related metavolcanic, metaplutonic, and metasedimentary rocks, which are exposed in the New River terrane in New Brunswick (Johnson and McLeod 1996; Johnson 2001; Barr et al. 2003a) and the Mabou and Cheticamp areas in the Aspy terrane of Cape Breton Island (Lin et al. 2007), as well as in the adjacent

Figure 8 Legend.

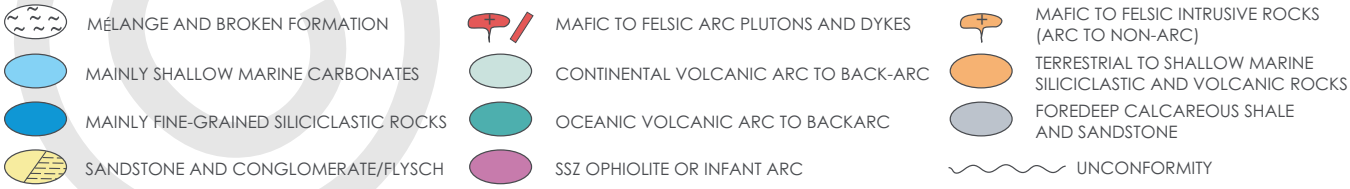


Figure 8 (opposite). Summary of the stratigraphic and tectonic relationships of the various tectonic elements present in the Canadian Appalachians. Peri-Laurentian elements west of the Red Indian line facilitate connection with Figure 6. Elements west of the Gander margin and Dog Bay Line are mainly based on relationships established in Newfoundland. Elements of the Gander margin, Avalonia, and Meguma are principally based on relationships in maritime Canada and Maine. Abreviations AG: Arisaig Group; AOB: Annieopsquotch ophiolite belt; BU; Buchans Group; C: Coaker porphyry; CA/MB: Coastal arc/ Mascarene back-arc basin; DBL–BBL: Dog Bay line–Bamford Brook fault; DP: Dunn Point volcanics; Fred trough: Fredericton trough; HH: Hodges Hill pluton; KBF: Kennebacesis fault; LB: Loon Bay pluton; LL: Long Lake volcanics; MP: Mount Peyton pluton; RILG: Red Indian Lake Group; P: Penobscot back-arc oceanic lithosphere; PV: Popelogan–Victoria arc; RIL: Red Indian line; SH/BP: Sops Head/Boones Point Complex (mélange); SK: Skidder ophiolite; SM: South Mountain batholith; SS: St. Stephen gabbro; TE: Tetagouche–Exploits back-arc basin; TP: Tally Pond volcanics; WR: White Rock Group.

Brookville and Bras d'Or terranes (see below). It also includes Neoproterozoic rocks of the Hermitage flexure (Fig. 7) in southern Newfoundland (O'Brien et al. 1996), generally inferred to represent basement to the Paleozoic sedimentary rocks of the Gander margin on the basis of their spatial association, isotopic signatures (Kerr et al. 1995), and complex Early Paleozoic orogenic overprint (Dunning and O'Brien 1989; O'Brien et al. 1991; Valverde-Vaquero et al. 2006b). Although a basement-cover relationship has not been identified in Newfoundland, such a relationship is preserved in southern New Brunswick where Cambrian arenite and quartz-rich conglomerate of the Matthews Lake Formation disconformably overlie Lower Cambrian and Neoproterozoic igneous rocks of the New River terrane (Johnson and McLeod 1996). The detrital zircon content of the Matthews Lake Formation is similar to that of the Miramichi, Woodstock, and Cookson groups, suggesting that they all form part of the same cover sequence (Fyffe et al. 2009). Arc rocks of similar age occur also in Avalonia, but the Ganderian rocks are generally isotopically more evolved (e.g. Kerr et al. 1995; Samson et al. 2000). In addition, Avalonia lacks Cambrian arc magmatism and has a more protracted and complex Neoproterozoic tectonothermal history (e.g. Barr and White 1996; Hibbard et al. 2006). Hence the boundary between Ganderia and Avalonia can be mapped with some confidence (Barr et al. 2003a; van Staal 2005; Hibbard et al. 2006; Fyffe et al. 2009; van Staal et al. 2009).

Faunal provinciality in Ganderia changed from Gondwanan in the Cambrian and Early Ordovician to mixed Gondwana–Laurentian and endemic faunas (Celtic province) in the Middle Ordovician, to Laurentian marginal and midcontinent faunas during the Late Ordovician (e.g. Nowlan et al. 1997; Fortey and Cocks 2003; Harper et al 2009), reflecting progressive transfer of Ganderia from Gondwana to Laurentia (van Staal et al. 1998).

2.3.3 Brookville–Bras d'Or Terrane

The Brookville–Bras d'Or terrane (Figs. 3, 7) has been interpreted to be a remnant of the oldest exposed Neoproterozoic to Mesoproterozoic (?) basement of the Ganderian microcontinent on which the Gander margin was constructed in the early Paleozoic (Raeside and Barr 1990), although some authors persist in linking it to Avalonia, mainly because of its similarity in age (e.g. Keppie et al. 1998, 2000). Brookville–Bras d'Or terrane is preserved in fault-bounded slivers in both southern New Brunswick (Brookville) and central Cape Breton Island (Bras d'Or), sandwiched between the Gander margin to the northwest and Avalonia to the southeast (Figs. 2, 3) (Barr and White 1996; Barr et al. 1998). The Neoproterozoic rocks of the Hermitage flexure in southern Newfoundland (Figs. 2, 3) (O'Brien et al. 1996) are likely related to the Brookville–Bras d'Or terrane on the basis of their lithologies, isotopic signatures, and tectonic evolution (Barr et al. 1998). Also related on the basis of similar evidence are the Neoproterozoic rocks of the New River terrane in southern New Brunswick (Johnson and McLeod 1996; Johnson and Barr 2004; Johnson et al. 2009) and isolated Neoproterozoic rocks in northern New Brunswick (van Staal et al. 1996), the Mabou and Cheticamp areas of western Cape Breton Island (Lin et al. 2007) and central Newfoundland (Rogers et al. 2006), although these areas typically lack the clastic-carbonate sequences preserved in the Brookville–Bras d'Or terrane. These clastic-carbonate sequences are termed the Green Head Group in Brookville terrane and George River Metamorphic Suite in Bras d'Or terrane, and their maximum age remains uncertain. Based on locally preserved stromatolites, Hofmann (1974) suggested that the Green Head Group is Neohelikian (Mesoproterozoic), consistent with the ca. 1230 Ma age of the youngest detrital zircon grain yet recorded from the Ashburn Formation of the Green Head Group (Barr et al. 2003b). In contrast, the other formation in the Green Head Group, Martinon, yielded detrital zircon grains as young as 602 ± 8 Ma, interpreted to represent the maximum depositional age of the formation (Fyffe et al. 2009). However, the relationship between the Martinon and Ashburn formations is debated; Wardle (1978) and White (1996) suggested that they are gradational and to some extent lateral facies equivalents, whereas other workers (e.g. Alcock 1938; Leavitt 1963; Nance 1987; Currie 1991) interpreted the Martinon to unconformably overlie the Ashburn Formation. In Cape Breton Island, the presence of zircon grains as young as 644 Ma led Keppie et al. (1998) to suggest that as a maximum age for the George River Metamorphic Suite. The age of these passive-margin sequences is especially important because the presence of abundant carbonate rocks requires a low-latitude location during sediment accumulation. As Ganderia is interpreted to have formed on part of the Amazonian margin of Gondwana during their deposition, Amazonia could not yet have reached the intermediate or high latitudes where it was situated during the Cambrian to Middle Ordovician (see below).

In both Brookville and Bras d'Or terrane, the mainly low-metamorphic-grade clastic-carbonate sequences are in tectonic contact with gneissic rocks that have been interpreted to have a maximum depositional age of 640 Ma (Bevier et al. 1990). Crystallization ages of 605 ± 3 Ma and 565 Ma have been reported for orthogneissic units, and metamorphic ages of 564 ± 6 Ma and 550–540 Ma have been reported for Brookville and Bras d'Or, respectively (Bevier et al. 1990; Dallmeyer et al. 1990; Barr et al. 1995; Keppie et al. 1998; White et al. 2003). The original relationship between the high- and low-grade metamorphic units remains problematic but it is likely that the low pressure-high temperature metamorphism that characterizes the gneissic rocks (Jamieson 1984; Raeside and Barr 1990; White and Barr 1996) originated in a back-arc setting associated with subduction at the former passive margin or was the product of ridge subduction (Fyffe et al. 2009). Both the low- and high-grade metamorphic units are intruded by Neoproterozoic–lower Cambrian subduction-related granitoid rocks; comagmatic volcanic rocks are present locally (Raeside and Barr 1990; White and Barr 1996). These rocks show strong magmatic, isotopic, and tectonic links to coeval igneous rocks exposed sporadically in other parts of Ganderia (Barr et al. 1998; Lin et al. 2007), consistent with the inferred presence of a Silurian and/or Upper Ordovician sedimentary overstep sequence in Cape Breton Island (Lin 1992; Chen et al. 1995). However, in contrast to other parts of Ganderia, the Brookville–Bras d'Or terrane lacks evidence for a penetrative Siluirian–Devonian tectonothermal overprint except in the boundary zone with the Gander margin (e.g. Reynolds et al. 1989; Barr et al. 1995; Lin 2001). The lack of overprint has been atrributed to the different tectonic setting of the Brookville–Bras d'Or terrane with respect to the Gander margin during the Silurian and Devonian. For example it may have been translated a significant distance along the Gander margin as result of strike-slip deformation (Lin et al. 2007, see also below).

Although the Gander margin may have formed at the edge of the Iapetus Ocean, the earlier Proterozoic (possibly Mesoproterozoic) passive margin on Brookville–Bras d'Or terrane predates the Iapetus Ocean. Middle Cambrian to Lower Ordovician volcanic and sedimentary rocks of the Bourinot belt in central Cape Breton Island form part of the Brookville–Bras d'Or terrane (White et al. 1994), as do similar assemblages preserved locally in southern New Brunswick (Landing et al. 2008). They are likely correlatives of sedimentary rocks deposited elsewhere on the Gander margin. Particularly, the association of Upper Cambrian–Lower Ordovician dark grey to black shale and sandstone in both sedimentary sequences is similar, although the Bourinot belt and equivalent units in New Brunswick are more fossiliferous than typical Gander margin rocks elsewhere (Landing et al. 2008). The volcanic – plutonic rocks of the Bourinot belt have middle Cambrian ages (509–505 Ma) and compositions indicative of a non-orogenic, within-plate tectonic setting (White et al. 1994), and were probably related to rifting of Ganderia from Gondwana (van Staal et al. 1998). Consanguineous plutonic rocks occur in several places elsewhere in Brookville–Bras d'Or terrane, and also in southern Newfoundland (Dunning and O'Brien 1989; Valverde-Vaquero et al. 2006b).

2.4 Avalonia

Avalonia (Figs. 3, 7) is a collage of fault-bounded Neoproterozoic, partly juvenile arc-related volcanic-sedimentary belts or terranes and associated plutonic rocks that experienced a complicated and long-lived Neoproterozoic tectonic history before deposition of an overstepping Cambrian–Early Ordovician platformal sedimentary succession (e.g. Kerr et al. 1995; Landing 1996, 2004; O'Brien et al. 1996). Avalonia has been traced from the British Caledonides (e.g. Gibbons and Horak 1996), to Rhode Island (Thompson et al. 2010a,b) and in Canada comprises eastern Newfoundland, the Mira terrane of southeastern Cape Breton Island, the Antigonish and Cobequid highlands of northern mainland Nova Scotia, and the Caledonia terrane of southern New Brunswick (Barr and Kerr 1997; Barr et al. 1998; Hibbard et al. 2006). The parts of Avalonia in eastern North America are commonly termed West Avalonia, to distinguish them from those of the UK, where they are termed East Avalonia. Although the two areas are generally considered to have been part of the same microcontinent (Cocks and Fortey 2009), that connection is equivocal (Schofield et al. 2010; Waldron et al. 2010; Horak and Evans 2011). Palaeomagnetic data indicate that Avalonia resided at a high southerly latitude near Gondwana (Fig. 4) at the time of the Cambrian–Ordovician boundary (van der Voo and Johnson 1985; Johnson and van der Voo 1986; MacNiocaill 2000; Hamilton and Murphy 2004; Thompson et al. 2010a), following a more intermediate latitude position during the Late Neoproteozoic (ca. 580 Ma; McNamara et al. 2001). Fossils also show strong links to Gondwana (e.g. Fortey and Cocks 2003), but previously proposed con-

nections to northwest Africa seem inconsistent with a wide range of geological arguments (e.g. Landing 1996). This led Murphy et al. (2002) to propose an alternative position further west along the Gondwanan margin, between Northwest Africa and the northeastern margin of Amazonia (Fig. 4), in proximity to but not connected yet with Ganderia (van Staal et al. 1996; Rogers et al. 2006; van Staal and Hatcher 2010). Such a position is supported by faunal (Cocks and Fortey 2009), paleomagnetic (Thompson et al. 2010a), and detrital zircon (Pollock et al. 2009; Satkoski et al. 2010) evidence. Proximity of Ganderia and Avalonia during the Cambrian and early Ordovician is also consistent with similarities in their sedimentary and faunal record (Fortey and Cocks 2003; Landing 2004). However, unlike Ganderia, Avalonia was not involved in Appalachian tectonic events until at least the Late Silurian. Hence, the Early Paleozoic tectonic evolution of these two microcontinents is fundamentally different, as is the nature of their underlying crusts and their Proterozoic histories (e.g. Barr and White 1996; Samson et al. 2000; Potter et al. 2008a,b).

Although the Neoproterozoic stratigraphy of Avalonia in the Canadian Appalachians has been widely documented (e.g. Barr et al. 1996; O'Brien et al. 1996; Barr and White 1999; Murphy et al. 1999), new data about the ages of units and their petrological characteristics continue to provide insights into Avalonia's complexity (e.g. Sparkes et al. 2005; Escarraga et al. 2010; Thompson et al. 2010b). It is now clear that different parts of Avalonia are characterized by different assemblages of volcanic-sedimentary-plutonic units, with significant hiatuses in activity, and hence depictions of Neoproterozoic Avalonia in a single composite stratigraphic column (e.g. Nance et al. 2008) are oversimplified. Avalonia's margin is a predominatly siliciclastic platformal succession dominated by shelf-facies shale and arenite during deposition of which water depth generally increased upsection (e.g. Landing 1996). The succession is punctuated by several disconformities, reflecting relative uplift and subsidence, and is generally preserved in scattered, isolated pockets, although correlations throughout Avalonia have been proposed (e.g. Landing 1996). Cambrian volcanic rocks are rare, and are well documented only in the Cape St. Mary's area in Newfoundland, where they are of Middle Cambrian age (Greenough and Papezik 1985). Other volcanic rocks previously included in Avalonia are now considered part of Ganderia (e.g. Bourinot belt and correlative units – see above).

Avalonia has been interpreted to have finally separated from Gondwana during the Early Ordovician at 480–475 Ma (Prigmore et al. 1997; van Staal et al. 1998; Cocks and Fortey 2009; Pollock et al. 2009). The rift-drift event is characterized by deposition of a transgressive arenitic cover sequence both in Gondwana (Armorican quartzite) and parts of Avalonia (e.g. Stiperstones and Bell Island quartzites in England and Newfoundland, respectively). However, other workers have suggested an earlier departure of Avalonia from Gondwana, at ca. 550 Ma, based on widespread evidence for extension (bimodal volcanic rocks and rift-related fluvial clastic successions), a pervasive late Neoproterozoic ^{18}O depletion (Potter et al. 2008a,b), and evidence for widespread resetting of detrital muscovite ages at 550 Ma (Reynolds et al. 2009), as well as faunal evidence (Landing 1996). In any case, the ca. 550 Ma extension in Avalonia was a prelude to widespread submergence,in some places by the latest Ediacaran (e.g. at the Global Stratotype for the Ediacaran–Cambrian boundary at Fortune Head, Newfoundland; Myrow and Hiscott 1993) or earliest Cambrian (in Mira and Caledonia terranes; Barr et al. 1996; Tanoli and Pickerill 1988).

Little is known concerning the tectonic history of Avalonia while moving northward during the Middle to Late Ordovician. Middle to Upper Ordovician shelf-facies sandstone and siltstone were identified in core samples (King et al. 1986) and an unconformity (break-up?) separating lower Ordovician rocks from Middle Ordovician to Silurian rocks has been recognized on seismic profiles on the Avalon platform, in the offshore east of Newfoundland (Durling et al. 1987). Within-plate magmatism occurred sporadically in parts of Avalonia during its northerly drift towards Laurentia. Recently recognized voluminous gabbro and A-type granite and syenite forming the core of the Antigonish Highlands of northern mainland Nova Scotia have yielded a U-Pb (zircon) age of 469.4 ± 0.5 Ma (Escarraga et al. 2010), somewhat older than the ca. 460 Ma Dunn Point volcanic rocks in the northern Antigonish Highlands (Hamilton and Murphy 2004). These rocks could be related to prolonged rifting between Avalonia and Meguma (see below), but the full extent and significance of these rocks in the Avalonian story has yet to be determined. The Dunn Point volcanic rocks yielded a southerly paleolatitude of approximtately 41º (Johnson and van der Voo 1990) compared to a latitude of approximately 65º for Avalonian rocks in southeastern New England at 490 Ma (Thompson et al. 2010a), which demonstrates that Avalonia had left Gondwana and had

become an Iapetan microcontinent by the Middle Ordovician. The lowermost Silurian (ca. 440 Ma) Cape St. Mary's sills in southeastern Newfoundland indicate that Avalonia had reached intermediate latitude of approximately 32ºS by the earliest Silurian, still about 1000 km away from the composite Laurentian margin (Hodych and Buchan 1998). Avalonia was moving north with a latitudinal component of drift of approximately 9.0 cm/a until about 460 Ma, after which it slowed down to 5 cm/a (Hamilton and Murphy 2004; Thompson et al. 2010a). Since Avalonia and Ganderia were probably situated on the same microplate (van Staal and Hatcher 2010), these rates are also a proxy for Ganderia's drift history, which is consistent with the existing paleomagnetic data (van Staal et al. 1998 and references therein). The change in drift-rate approximately coincides with the start of organized sea-floor spreading in the Tetagouche–Exploits back-arc basin between 475 and 465 Ma (van Staal et al. 2003; Valverde-Vaquero et al. 2006a). If these two events are indeed related, the slowing down of Avalonia's northerly drift, and by implication also that of the proximal Gander margin, was compensated by a half-spreading rate of approximately 2 cm/a in the Tetagouche–Exploits back-arc basin, which separated the Gander margin from Popelogan–Victoria arc at Ganderia's leading edge.

Subsidence analysis of shelf-facies siliciclastic rocks of the Silurian Arisaig Group in the Antigonish Highlands of Nova Scotia indicates that part of Avalonia's margin was still a passive margin during most of the Silurian (Waldron et al. 1996; Hamilton and Murphy 2004). Increased subsidence during the latest Silurian (420–418 Ma) has been interpreted to indicate development of a foreland basin due to tectonic loading of Avalonia (Waldron et al. 1996), probably related to its accretion to composite Laurentia, which was the upper plate (see below).

2.5 Meguma

Meguma (Figs. 3, 7) represents the most outboard terrane preserved in the Canadian Appalachians and is exposed on land only in southern Nova Scotia (Fig. 1). However, its regional extent is much larger and its rocks have been traced offshore by geophysical and well data (Fig. 7) from the southernmost part of the Grand Banks southeast of Newfoundland across the Scotian shelf, and through the Gulf of Maine to southernmost Cape Cod (Hutchinson et al. 1988; Keen et al. 1991a,b; Pe-Piper and Jansa 1999). Meguma is characterized by unique stratigraphy, including a thick (>12 km) Early Cambrian to Early Ordovician (537–475 Ma) turbiditic clastic succession divided into a lower metasandstone-dominated Goldenville Group and an upper pelite-dominated Halifax Group (White 2010). In the western and northwestern part of the Meguma terrane, the Halifax Group is unconformably overlain by Lower Silurian volcanic and sedimentary rocks of the White Rock Formation and equivalent units (Blaise et al. 1991) deposited in a shallow marine (rift) environment (Lane 1975). This age difference indicates a hiatus of 30–40 my across the unconformity. The overlying Early Devonian Torbrook Formation was deposited in shelf conditions (Jensen 1975). Meguma then experienced intense, punctuated orogenesis during the late Early Devonian to Middle Carboniferous (ca. 395–320 Ma, Culshaw and Reynolds 1997; Hicks et al. 1999), during which the mid-Devonian to early Carboniferous South Mountain Batholith and satellite plutons were emplaced. Detrital zircon ages from the lower part of the Goldenville Group suggest derivation from local sources in the Avalonian/Pan-African orogen on the margins of Early Cambrian Gondwana. Samples from the top of the group show a broader distribution (Krogh and Keppie 1990; Waldron et al. 2009), including ages back to Archean. The trend is consistent with deposition in a rift, in which uplifted rift-flanks were the main source of the early basin fill, whereas subsequent thermal subsidence of rift margins allowed for more widespread sediment sourcing. Meguma was probably located in a rift between Gondwana and Avalonia (Fig. 4), along which Avalonia separated in the Early Ordovician (Waldron et al. 2009).

Fossil evidence suggests that during the Late Silurian, Meguma was close to Avalonia and/or Baltica and had separated from Gondwana (Bouyx et al. 1997). Combined with the occurrence of late Early Devonian to Middle Carboniferous orogenesis, it is likely that Meguma was a separate microcontinent during the Silurian and Devonian (see further below). Subsequently, the Meguma terrane, as well as by-then adjacent Avalonia and Ganderia, were covered by mainly non-marine Carboniferous sedimentary cover sequences (see below).

3.0 THE THIRD DIMENSION

3.1 Crustal Structure, Composition, and Seismic Characteristics

In contrast to the complex distribution of lithotectonic elements identified in the near-surface geology, as described above, only three zones with distinctly differ-

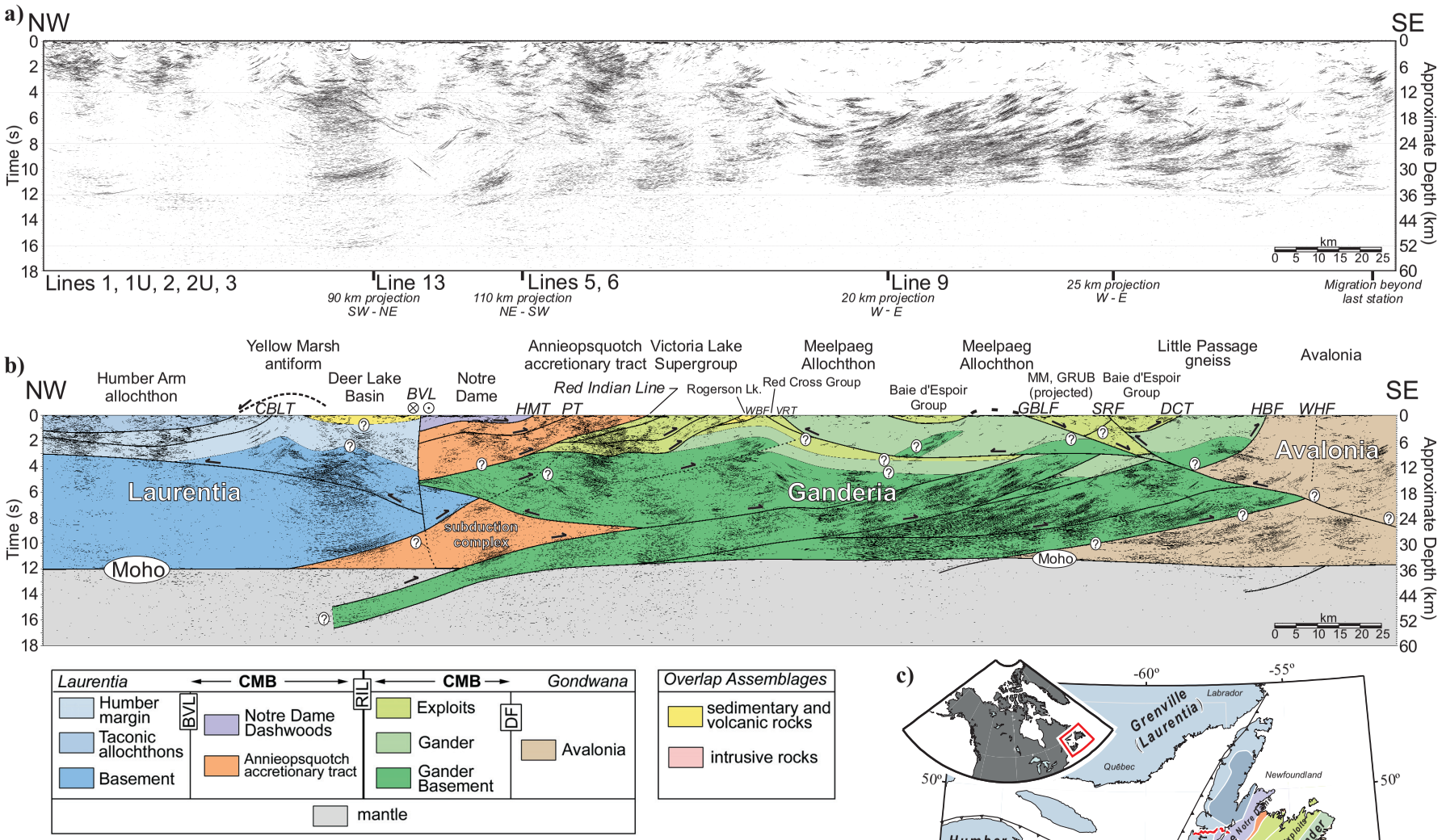


Figure 9. a) Composite migrated seismic reflection section across central Newfoundland; profile location shown in (c). Approximate depth and 1:1 format based on an average velocity of 6 km/s. Short bars with text at lower boundary indicate offsets associated with the composite. **b)** Interpreted seismic section with major features identified. Abbreviations BVL: Baie Verte line; CBLT: Corner Brook Lake thrust; CMB: Central Mobile belt; DCT: Day Cove thrust; DF: Dover fault; GBLF: Great Burnt Lake fault; GRUB: Gander River ultrabasic belt; HBF: Hermitage Bay fault; HMT: Hungry Mountain thrust; MM: Mount McCormack window; PT: Powerline thrust; RIL: Red Indian line; SRF: Salmon River fault; VRT: Victoria River thrust; WBF: Wood Brook fault; WHF: White Horse fault. **c)** Location map for seismic section shown in (a) and (b). Figure adapted from van der Velden et al. (2004).

ent lower crustal reflection fabrics were originally recognized in Newfoundland (Keen et al. 1986) and beneath the Gulf of St.Lawrence: the Grenville, central, and Avalon lower crustal blocks (LCBs) (Marillier et al. 1989). Most of the central block was susbsequently reinterpreted as Ganderian lower crust by van der Velden et al. (2004). The positions of the Mohorovičić (Moho) discontinuities of the three LCBs are relatively well defined (e.g. Fig. 9) because reflection and refraction Mohos are generally coincident in Newfoundland (Hall et al. 1998). Hughes et al. (1994) and Jackson et al. (1998) constructed velocity structural models of Newfoundland's crust that are characterized by three major layers: upper, middle, and lower crust (Fig. 10). The boundaries between these layers vary across and along strike of the orogen (Fig. 11; Chian et al. 1998), particularly between different tectonic elements. Across Newfoundland, the upper crust is situated between depths of 0 and 15 km, the middle crust between 5 and 25 km, and the lower crust between 15 and 45 km. The upper crust generally becomes thicker, mainly at the expense of the middle crust, towards the southeast.

The Grenville LCB underlies the Humber margin and extends beneath the Laurentian craton (Grenville province) to the west, well beyond the Appalachian structural front (Marillier et al. 1989; Hall et al. 1998). Its moderately strong lower crustal reflectivity, which generally dips southeast towards the core of the orogen, was interpreted, therefore, to be mainly an inherited Grenvillian fabric with little or no Appalachian structural overprint. The thick crust of the Grenville province is characterized by a distinctive pattern of negative gravity anomalies (Hall et al. 1998). Upper crustal velocities are 5.7 to 6.2 km/s and probably correspond mainly to a bulk composition dominated by quartzofeldspathic gneiss and schist with felsic to intermediate compositions. Middle crustal velocities are 6.2 to 6.45 km/s and lower crustal velocities are in the range of 6.7 to 7.0 km/s, progressively reflecting a higher percentage of intermediate to mafic (diorite-gabbro) metamorphic rocks (mainly granulite) in the lower crust. Within the Grenville province, the Moho is relatively deep at 13 to 14 s two-way reflection traveltime, which correspond to a depth of 40 to 45 km (Hall et al. 1998; Jackson et al.

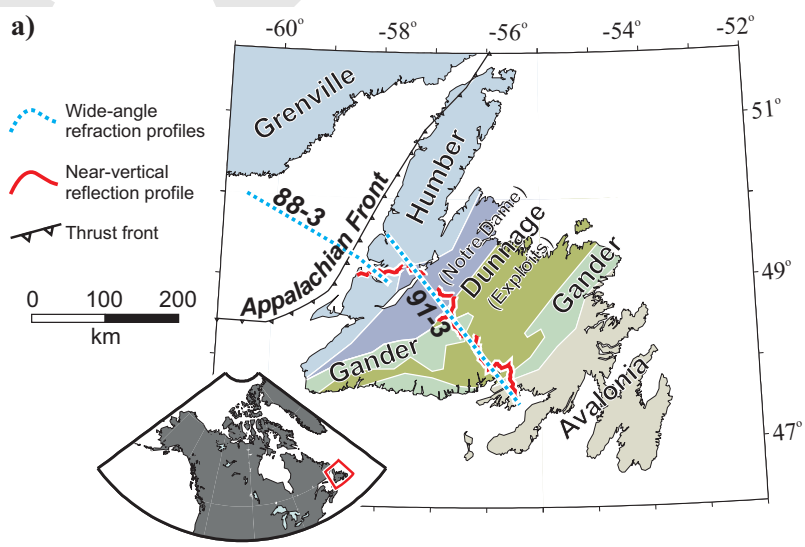
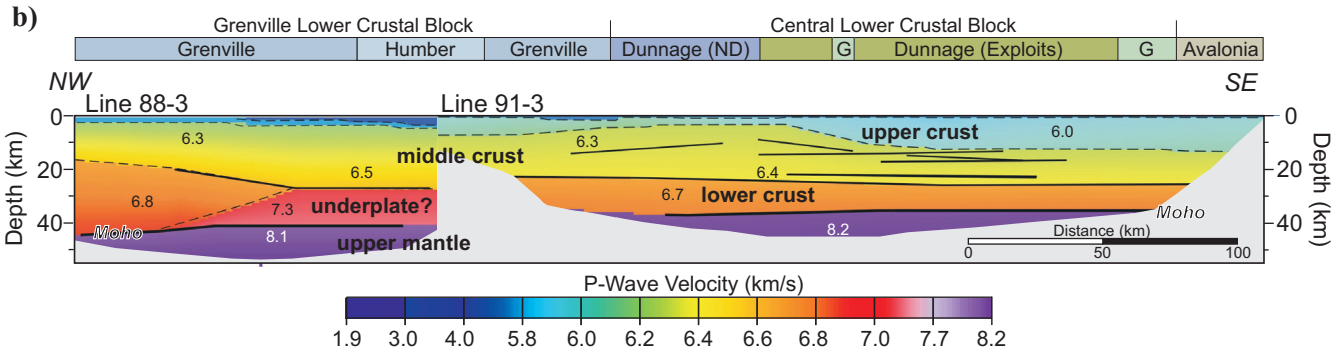


Figure 10. a) Location map for refraction/wide-angle reflection lines 88-3 and 91-3. **b)** Interpreted velocity model from offshore line 88-3 and onshore line 91-3. Bar above model identifies the tectonostratigraphic zones and subzones and lower crustal blocks. Abbreviations G: Gander zone; ND: Notre Dame subzone. Solid black lines indicate locations from which wide-angle reflections were identified; dashed lines show prominent changes in velocity. The Moho is well defined by wide-angle reflections and a velocity discontinity. An approximately 100 km wide zone with a lower crustal velocity of 7.2 to 7.3 km/s at the leading edge of line 88-3 may represent a thick mafic-ultramafic underplate emplaced during opening of the Iapetus Ocean or the Taconic Seaway. Figure adapted from Hammer et al. (2010) based on results from Jackson et al. (1998) (Line 88-3) and Hughes et al. (1994) (Line 91-3).



1998). Near the Appalachian structural front, the Moho becomes shallower, gravity anomalies less negative, the lower crust more reflective, and seismic velocity increases to 7.2 km/s (Hall et al. 1998). The high-velocity part of the Grenville LCB (Fig. 10) defines a westward-tapering sheet-like body, situated in the lowermost crust, the top of which truncates the older southeast-dipping Grenvillian reflection fabric (Jackson et al. 1998). The high-velocity sheet has been interpreted as an underplate or thick lopolith of gabbroic to ultramafic composition, either representing mantle-derived material associated with opening of Iapetus or a younger, synorogenic Appalachian event (Hall et al. 1998). We prefer the former interpretation (see below). The extent of unreworked Grenville LCB beneath the Humber margin and beyond is contentious (Jackson et al. 1998). Reprocessed reflection data suggest that the upper parts of the Grenville LCB extend east of the Baie Verte–Brompton line beneath Dashwoods, but definitely not further than the Red Indian line (van der Velden et al. 2004) (Fig. 9). These results are consistent with geological evidence (described above) that links Dashwoods to Laurentia, with the Red Indian line being the fundamental Iapetan suture between peri-Laurentian and peri-Gondwanan terranes. An eastward-tapering Grenville LCB also has been observed to extend well east of the Baie Verte–Brompton line in New England below equivalents of Dashwoods and the Notre Dame arc (Hughes et al. 1993). Along the Maine–Quebec transect, Grenvillian middle-lower crust has been traced to the eastern edge of the Chain Lakes massif (Figs. 1, 2), near the boundary with Ganderia (Spencer et al. 1989;

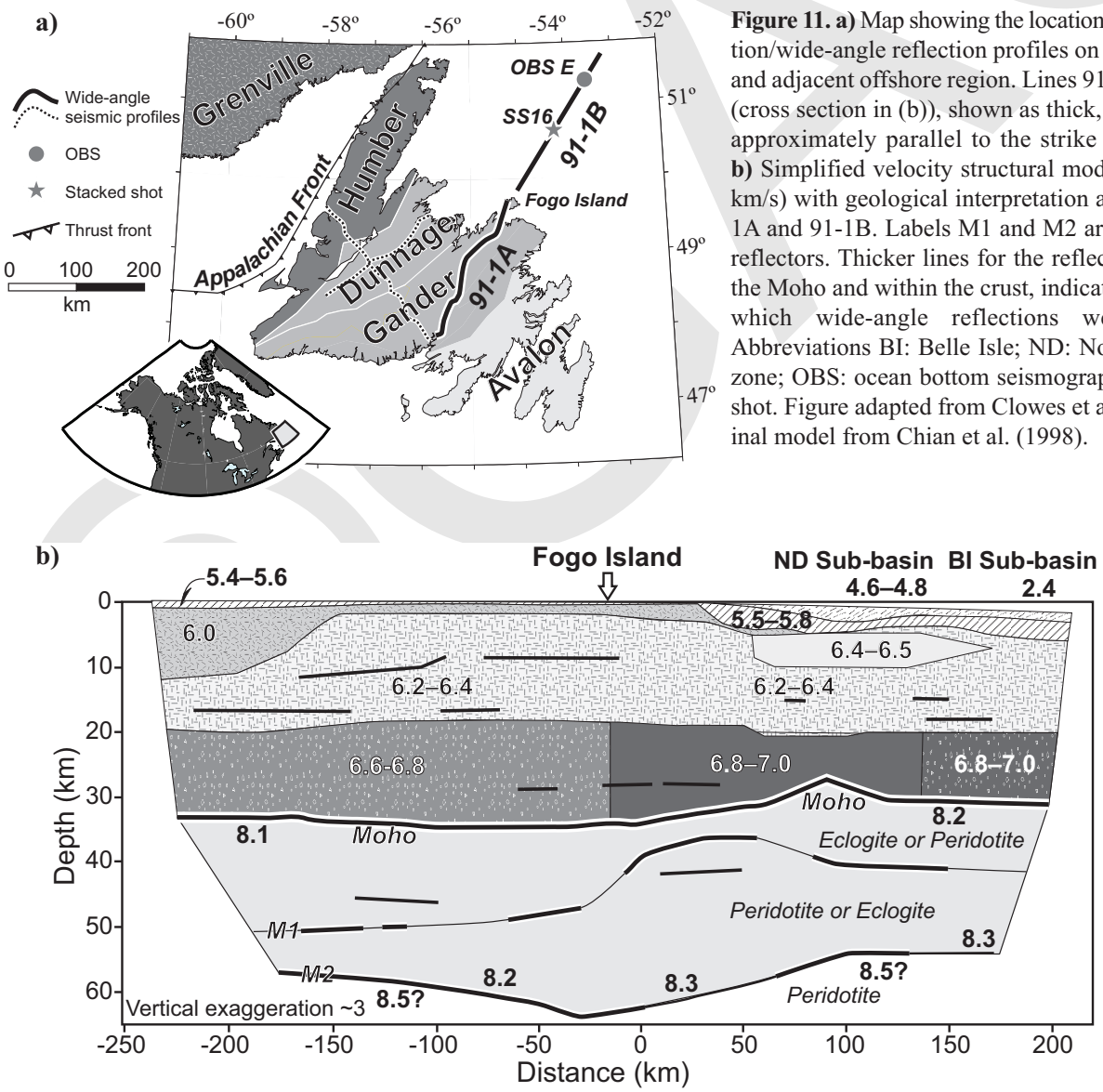


Figure 11. a) Map showing the location of some refraction/wide-angle reflection profiles on Newfoundland and adjacent offshore region. Lines 91-1A and 91-1B (cross section in (b)), shown as thick, solid lines, are approximately parallel to the strike of the orogen. **b)** Simplified velocity structural model (numbers in km/s) with geological interpretation along Lines 91-1A and 91-1B. Labels M1 and M2 are upper mantle reflectors. Thicker lines for the reflectors, including the Moho and within the crust, indicate regions from which wide-angle reflections were observed. Abbreviations BI: Belle Isle; ND: Notre Dame subzone; OBS: ocean bottom seismograph; SS: stacked shot. Figure adapted from Clowes et al. (2010); original model from Chian et al. (1998).

Stewart et al. 1993). This boundary thus corresponds with the position of the Red Indian line in this segment of the orogen, because the Chain Lakes massif is interpreted to be equivalent to Dashwoods (Waldron and van Staal 2001; Gerbi et al. 2006a,b). The Canadian LCBs and their overlying middle-upper crustal rocks thus represent a crustal architecture that may be characteristic of the whole northern Appalachians.

The central LCB of Marillier et al. (1989) was interpreted by van der Velden et al. (2004), based on a combination of different geological arguments, to represent Ganderian lower crust (GLC). It is characterized by a strongly reflective, shallow (11–12 s reflection time) Moho (Fig. 9), which is situated at a depth of 32–36 km, consistent with refraction data (Fig. 10, Chian et al. 1998). GLC reflections have a marked westerly dip and locally project into the mantle to a depth of at least 16 s reflection time (~52 km) beneath the Humber margin, which was interpreted to represent a fossil subduction zone related to the Silurian Salinic orogenic cycle (van der Velden et al. 2004), thus supporting the geological evidence for Salinic subduction polarity (van Staal 1994; van Staal et al. 1998, 2008; Valverde-Vaquero et al. 2006a). The GLC reflections were interpreted to represent compositional layering within the infrastructure of Ganderia, probably due to sheets of Neoproterozoic to Early Cambrian felsic to mafic arc-related igneous rocks.

The nature of the postulated compositional layering is unknown since Ganderian lower crust is nowhere exposed. It may be an original feature formed as a result of sill-injection, but regardless of its origin, the layering was probably modified and enhanced during shearing and transposition that were imposed as a result of non-coaxial deformation accompanying Ganderia's entrance and descent in the Salinic subduction channel at 445–423 Ma. Structural modification, at least part of it, may also have occurred during the collision of Avalonia with composite Laurentia and its partial subduction beneath its leading edge, which at that stage was represented by Ganderia (van Staal 2007). The underthrusted segment of Avalonia's LCB locally shows strong reflections not unlike those of Ganderia (van der Velden et al. 2004). It is important to note here that the Latest Silurian to Middle Devonian (421–380 Ma) Acadian orogenesis started immediately after the end of the Salinic orogeny (ca. 423 Ma) in Newfoundland and New Brunswick (van Staal et al. 2008, 2009). Salinic and Acadian orogenesis thus was probably a continuous event at the level of the lower-middle crust.

A Silurian and/or Devonian age for the reflection fabrics is consistent with the observed marked thinning of the Ganderian crust to 30 km along the axis of the orogen to the northeast (Fig. 11) below a deep Carboniferous basin (Chian et al. 1998; Hall et al. 1998;), which is situated offshore north of north-central Newfoundland. Thinning also affected its underlying upper mantle layer and thus affected the whole lithosphere. Spatial coincidence between lithospheric thinning and basin position suggests that the along-axis variation in crustal thickness is related to a phase of Carboniferous extension, thus implying that the GLC reflections and mantle reflections are older.

The upper crust in central Newfoundland, disregarding the low velocities observed in areas underlain by Devonian/Carboniferous or younger sedimentary basins, generally is characterized by velocities in the range of 5.9–6.1 km/s , which correspond to a bulk composition dominated by quartzofeldspathic, felsic to intermediate igneous and metamorphic rocks (Hughes et al. 1994). Indeed, upper-middle crustal Ganderian rocks typically comprise granitic to granodioritic plutons, associated migmatite, and quartzofeldspathic metamorphic rocks, where exposed (Fyffe et al. 1988a; Burgess et al. 1995; Valverde-Vaquero et al. 2000; Lin et al. 2007)

Avalonia has been imaged mainly on offshore transects (Keen et al. 1986; Marillier et al. 1989) and is crossed for only a short distance in one on-land transect (Fig. 9) in Newfoundland (Quinlan et al. 1992). Avalonia's LCB is characterized by a generally less reflective, thicker crust, with the Moho situated at 12–14 s reflection time or a depth of 36–40 km (Hall et al. 1998; van der Velden et al. 2004). In northeastern Newfoundland, the Avalonia–Ganderia boundary is the Dover fault, a subvertical structure that cuts the entire lithosphere and appears to have accommodated significant Silurian (sinistral) and Devonian (dextral) strike-slip movements (Holdsworth 1994). The Moho shows an abrupt step across the projected offshore position of the Dover fault and lower crustal reflections are truncated at the approximate position of the fault (Keen et al. 1986). Both relationships suggest a pre-Carboniferous age for the reflection fabrics. A step in the Moho is absent crossing the Avalonia–Ganderia boundary in southeastern Newfoundland. Here, Avalonia appears to have thrust beneath Ganderia along a listric northwest-dipping fault (Fig, 9; van der Velden et al. 2004). Avalonia's LCB reflections continue into the mantle to approximately14 s, which may be a remnant of the west-dipping A-subduction zone inferred to have been responsible for the Acadian orogenic cycle (see

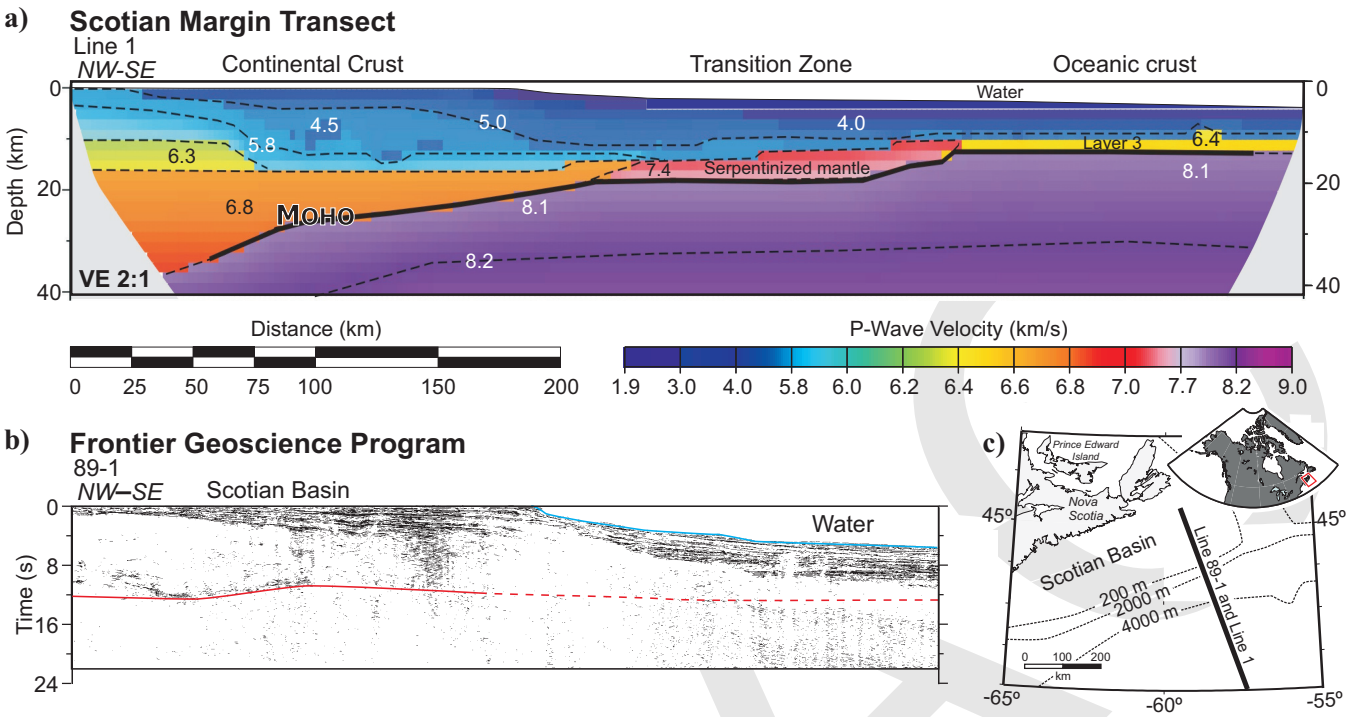


Figure 12. Seismic results from the Atlantic margin offshore Nova Scotia. **a)** Seismic velocity model for line 1 (location shown in (c)); numbers are P-wave velocities in km/s. The Moho is well located by wide-angle reflections and a velocity discontinuity. Rapid crustal thinning occurs in a 180 km wide transition zone. A 5 km thick lower layer with a velocity of 7.2 to 7.6 km/s in this transition zone was interpreted by Funck et al. (2004) as serpentinized mantle exhumed during rifting. **b)** Migrated seismic reflection section along line 89-1. At the northwest end, crustal thickness is about 36 km (12 s two-way traveltime; solid red line), which is consistent with the refraction data. Darker reflectivity below the water represents sediments. The section is plotted in traveltime and because water and sediments have low seismic velocities, traveltimes are longer; hence the geometry of this section is different from the velocity model, which is plotted in depth. **c)** Map showing the location of line 1 and line 89-1. Figure adapted from Hammer et al. (2010) based on results from Funck et al (2004) and Keen and Potter (1995a).

below). Upper crustal velocities in Avalonia are overall low, ~5.7 km/s, which is consistent with the predominance of sedimentary and volcanic rocks of subgreenschist facies exposed on surface.

Meguma (Fig. 7) has been imaged by offshore transects along the east and south coasts of Nova Scotia (Marillier et al. 1989; Keen et al. 1991a,b), in the Bay of Fundy (Keen et al. 1991a; Wade et al. 1996), and in the Gulf of Maine (Hutchinson et al. 1988; Keen et al. 1991a). Meguma is underlain, at least in part, by a very reflective LCB, characterized by horizontal to shallowly south-dipping relections, named the Sable block by Keen et al. (1991a). It is uncertain whether Meguma is completely allochthonous with respect to the Sable LCB or forms its suprastructure. Keen et al. (1991a) preferred a link between the Sable LCB and Meguma, which implied that Avalonia had been thrust over Meguma in that area. However, farther north in the Bay of Fundy, subsequent workers (e.g. Wade et al. 1996) generally adopted a version of the alternative model of Keen et al. (1991a) where the Sable LCB represents modified Avalonian basement and both Meguma and its unknown basement were thrust over Avalonia and its modified equivalent. This interpretation is consistent with the "Meguma over Avalonia" situation inferred along the Cobequid–Chedabucto fault system onshore and its eastward extension (Marillier et al. 1989).

The Sable LCB continues beneath the Gulf of Maine, based on correlations between the U.S. Geological Survey (Hutchinson et al. 1988) and Geological Survey of Canada (Keen et al. 1991a) seismic reflection lines. The boundary between the Avalon and Sable LCBs lies just north of the Nauset anomaly (Fig. 7) of Hutchinson et al. (1988), and is marked by a north-dipping break in reflectors, similar to the north-dipping Avalon–Meguma boundary identified by Keen et al. (1991a) as zone B. The Nauset anomaly crosses Cape Cod and hence Meguma is inferred to be present in the subsurface of easternmost Massachussets (Fig. 7). The reflection Moho of the Sable LCB varies in reflection time between 9 and 12 s. The highly reflective Sable LCB continues for a short distance east of the on-land

exposed part of Meguma (Fig. 7) beneath the continental shelf of Nova Scotia's Atlantic margin on line 88-1 of Keen et al. (1991b). Further east, the lower crust of Nova Scotia's margin (Meguma) is less reflective. The prominent reflections on this part of line 88-1 have a dominantly landward (northwest)-dip and cut both the lower crust and upper mantle near the eastern edge of the continental shelf and near the continent–ocean transition zone (Keen et al. 1991b). These reflectors may image deeply seated structures that initially had formed during the Alleghanian collision. However, the absence of such landward-dipping reflectors on line 89-1 (Fig. 12), further to the north, is inconsistent with such an interpretation (Keen and Potter 1995a). Regardless of whether the structures imaged by these reflectors formed during the Alleghanian collision or not, they probably accommodated low-angle normal movements during asymmetric Mesozoic extension (see section 7.0) associated with opening of the Atlantic Ocean (Keen and Potter 1995a). A velocity model developed from a refraction survey along line 89-1 (Fig. 12) indicates that such normal faulting had exhumed a 5 km thick slab of partially serpentinized ultramafic continental mantle (7.2–7.6 km/s) onto the seafloor (Funck et al. 2004).

3.2 Lithospheric Mantle Structure and Composition

Velocities of the uppermost mantle range between 8.1 and 8.3 km/s based on seismic refraction data (Figs. 10, 11) but little is known about the composition of the mantle beneath the Appalachians. West-dipping reflections projecting into the mantle have been observed below Ganderia, Avalonia, and Meguma. All three zones of mantle reflections have been interpreted as remnants of west-dipping A-subduction zones associated with the sequential accretion of these three continental terranes (Keen et al. 1991a,b; Wade et al. 1996; van der Velden et al. 2004; van Staal et al. 2009). If correct, this interpretation implies that significant volumes of crust have been incorporated in the upper mantle and acquired bulk physical properties (velocity and density) resembling those of the surrounding mantle. Parts of the subducted and thickened lower crust may have been transformed into eclogite. A thin 1–2 km thick reflector recognized at a depth of 40 to 50 km beneath Ganderia's LCB (Fig. 11) was previously proposed to represent eclogitic subducted oceanic crust (Chian et al. 1998). Segments of the lower crust may also have lost most or all their feldspar and quartz by extensive partial melting and metamorphism, leaving a bulk mantle-like residue and hence resetting the geophysical Moho, consistent with indications that the Ganderian lower crustal reflection fabric is truncated by the reflection Moho (van der Velden et al. 2004).

A mantle discontinuity associated with a deep mantle reflector, which is associated with an abrupt rise in velocity from 8.1 to 8.5 km/s, has been observed at approximately 60 km depth below Ganderia's LCB (Fig. 11). Chian et al. (1998) suggested that this discontinuity may mark the boundary between unmodified mantle and mantle that comprises a significant volume of reworked and/or eclogitic crust that now seismologically resembles normal mantle.

3.3 Implications for Lithospheric Structure and Orogenic Processes

Both seismic reflection and refraction data indicate that the geophysical Moho in Newfoundland is at a shallower depth beneath its orogenic core (central LCB) than it is beneath the Grenville LCB overlain mainly by the Humber margin in the west or beneath the Avalonian LCB in the east. Crustal thickness in the central LCB varies between 30 and 34 km (Chian et al. 1998), whereas the Humber margin and Avalonia have crustal thicknesses of approximately 45 and 40 km, respectively (Figs. 10, 11) (Hall et al. 1998). Within the the central LCB of Newfoundland, crustal thickness increases along the trend of the orogen (Fig. 11) from northeast (~30 km) to southwest (~34 km; Chian et al. 1998; Hall et al. 1998), which coincides with an increase in metamorphic grade of rocks exposed on surface. The most favoured explanation as to why the core of the Appalachian orogen has a relatively shallow geophysical Moho is that it was partly reset during Early Devonian (Acadian) orogenesis (see above). This process may have been accomplished through metamorphism (e.g. eclogite facies) of a predominantly mafic lower crust and/or large-scale partial melting of thickened lowermost crust (van der Velden et al. 2004; Cook et al. 2010), which produced a mafic restite with mantle-like geophysical properties. The latter process is consistent with the large bloom of mafic to felsic magmatism throughout Ganderia (but not elsewhere) that accompanied Early Devonian orogenesis (see below). Other proposed models such as whole-scale lithospheric delamination (Nelson 1992) or a vice model (Ellis et al. 1998) have been rejected or relegated to a possible contributing role, respectively, on the basis of a combination of geological and geophysical arguments (see discussion in van der Velden et al. 2004). Nevertheless, the anisotropy of the fossilized olivine fast

axis in the upper mantle of the southern New England Appalachians has an orientation perpendicular to the orogen, suggesting that some form of lithospheric mantle thinning has taken place, at least there, after Late Paleozoic collisional orogenesis (Levin et al. 2000).

The along-strike variation in crustal thickness is generally ascribed to domains of enhanced crustal thickening associated with promontory–promontory collision in southwestern Newfoundland and Cape Breton Island compared to less collisional convergence and crustal shortening in the adjacent embayments (Hall et al. 1998; Lin et al. 1994). The west-dipping mantle reflectors beneath the central LCB line-up with major lithotectonic block boundaries and have been interpreted as fossil A-subduction zones that were responsible for the accretion of Ganderia and Avalonia, respectively (van der Velden et al. 2004; see below). Hence, the crust of at least some of the accreted blocks appears to be still attached to parts of their original lithospheric mantle. Reinterpretation of the reprocessed seismic reflection data suggests that both the penetrative west-dipping lower crustal Ganderian (Salinic) and Avalonian (Acadian) fabrics were overprinted by major west-directed thrusts that root in the middle crust such that two crustal-scale wedge-like geometries were created (Fig. 9). These thrusts probably formed during the later stages of Acadian orogenesis, consistent with truncation of the most western thrusts by the Baie Verte line, suggesting that they formed before the Carboniferous (van der Velden et al. 2004).

4.0 STYLES OF RIFTING AND NATURE OF THE MARGINS

The various lithotectonic elements in the Canadian Appalachians have preserved evidence for several rifting events. First were the rifting events responsible for terrane dispersal (Dashwoods, Ganderia, Avalonia, and Meguma), including opening of the Iapetus and Rheic oceans, opening of the Taconic seaway, and opening of several marginal basins in both the peri-Laurentian and peri-Gondwanan realms (Fig. 4). A very important protracted phase of rifting was the break-up of Rodinia between 750 and 570 Ma and isolation of Laurentia as a separate continent following opening of the Iapetus Ocean (Cawood et al. 2001). Laurentia had a near-equatorial position at the terminal Neoproterozoic (Hodych et al. 2004). While Iapetus had just started to open, a new large continental mass, Gondwana, was being assembled at intermediate to high southerly latitudes as a result of late Neoproterozoic to Early Cambrian collisions, suggesting a tectonic link between these two events (van Staal and Hatcher 2010). The western arm of the Iapetus Ocean separated the south-facing proto-Appalachian margin of Laurentia from Gondwana (North Africa, Amazonia, and Rio de la Plata), while the northeastern arm separated Laurentia's east-facing proto-Caledonian margin (Greenland) from Baltica (Meert and Torsvik 2003). Baltica at this time was an isolated continental plate. By the end of the Cambrian (ca. 490 Ma), the western arm of Iapetus, which separated Laurentia's proto-Appalachian margin from the opposite Gondwanan margin (Fig. 4), had achieved a significant width, locally in the order of 4000 to 5000 km (Johnson et al. 1991; Trench et al. 1992; MacNiocaill and Smethurst 1994; Thompson et al. 2010a).

The style of rifting accompanying the above events, i.e. whether the rifts were symmetrical or asymmetrical (Lister et al. 1991), is generally poorly known because comparative studies of the conjugate margins of the rifts, now located somewhere outside the Appalachians, commonly cannot be made. In addition, the rift basins were generally severely tectonized during subsequent collisional orogenesis. An exception is the Humber margin where significant progress has been made in interpreting changes in styles of rifting (Thomas 2006; Allen et al. 2010), mainly by comparing the sedimentary–volcanic architecture of the syn- and post-rift sequences along the length of the margin. Lower plate settings are generally characterized by larger and earlier subsidence whereas a relatively abrupt transition from normal thickness continental crust to oceanic crust is an indication that the margin occupied the upper plate (Buck et al. 1988). Using these principles, their work suggests that the Humber margin, which together with the Dashwoods microcontinent, formed as a result of Early Cambrian (ca. 540 Ma) opening of the Taconic Seaway (Figs. 3, 13), mainly occupied an upper plate setting in Newfoundland and southern Quebec and a lower plate setting in northern Quebec (Gaspé), the lower and upper plate segments separated by transform faults. Although sedimentary rocks of the Dashwoods microcontinent are highly tectonized, the abundance of immature feldspathic psammite and lack of platform deposits is most consistent with a lower plate setting. Final rifting and the start of seafloor spreading in the Taconic Seaway were preceded by eruption of basalt, mainly between 565 and 550 Ma on the Humber margin. The mafic magmas generally have transitional to alkaline, ocean island basalt compositions, which has been related to the location of a superplume centred on the Sutton Mountains in southern Quebec (Kumarapeli 1993; Puffer 2002;

Hodych and Cox 2007). Such a superplume may have been responsible for the long-lived (615–550 Ma) Ediacaran magmatism observed in the Humber margin, which accompanied both opening of the Iapetus Ocean at 580–570 Ma and subsequently the Taconic Seaway at 565 to 550 Ma (Figs. 3, 4, 13). The presence of voluminous basalt indicates that the Humber margin was generally a volcanic rifted margin in Quebec and Newfoundland. This is consistent with the absence of evidence for exhumation of the mantle during rifting, as is typical of non-volcanic margins, and the presence of a lower high-velocity layer approximately 12 km thick on seismic line 88-3 (Fig. 10) at Newfoundland's west coast (Hall et al. 1998; Jackson et al. 1998), which we interpret as a mafic underplate. However, exhumed late Ediacaran mantle and lower crust (ca. 558 Ma) have been recognized in the Birchy Complex of the Baie Verte Peninsula of northern Newfoundland and western Ireland (van Staal et al. 2010), suggesting that the rifted margin became progressively more non-volcanic, and hence a sedimentary passive margin, to the north.

Evidence for the time of rifting of Ganderia from the Amazonian margin of Gondwana is best preserved in the Ellsworth belt of coastal Maine (Schultz et al. 2008). Here, rifting led to exhumation of mantle tectonite and eruption of 509–505 Ma bimodal magma (Fig. 8). In addition, the Gander margin was covered shortly thereafter by a widespread blanket of arenite and black shale, which probably represent a transgressive sequence related to the drift of Ganderia away from Gondwana. Exhumation of mantle suggests that the volume of syn-rift magmatism was initially low. Coeval, small volumes of rift-related magmatism of similar age and extending up to 495 Ma occur regionally and can be traced into the lower part of the Cookson Group (Fig. 8) in southern New Brunswick and Maine (Fyffe et al. 1988b; Tucker et al. 2001), Cape Breton Island (e.g. Bourinot belt and related plutons of White et al. 1994), and the Hermitage flexure of southern Newfoundland (Dunning and O'Brien 1989; O'Brien et al. 1991). Rifting was coeval with Penobscot arc activity at Ganderia's leading edge (Fig. 14) (Rogers et al. 2006; McNicoll et al. 2010; Zagorevski et al. 2010), suggesting that rifting and hence opening of the Rheic Ocean started as a back-arc basin event (van Staal et al. 1998, 2009).

Opening of the Rheic Ocean behind Avalonia apparently took place slightly later than rifting of Ganderia (see above). Considering that Avalonia is generally thought to have been situated further east, between West Africa and Amazonia, the opening of the Rheic Ocean

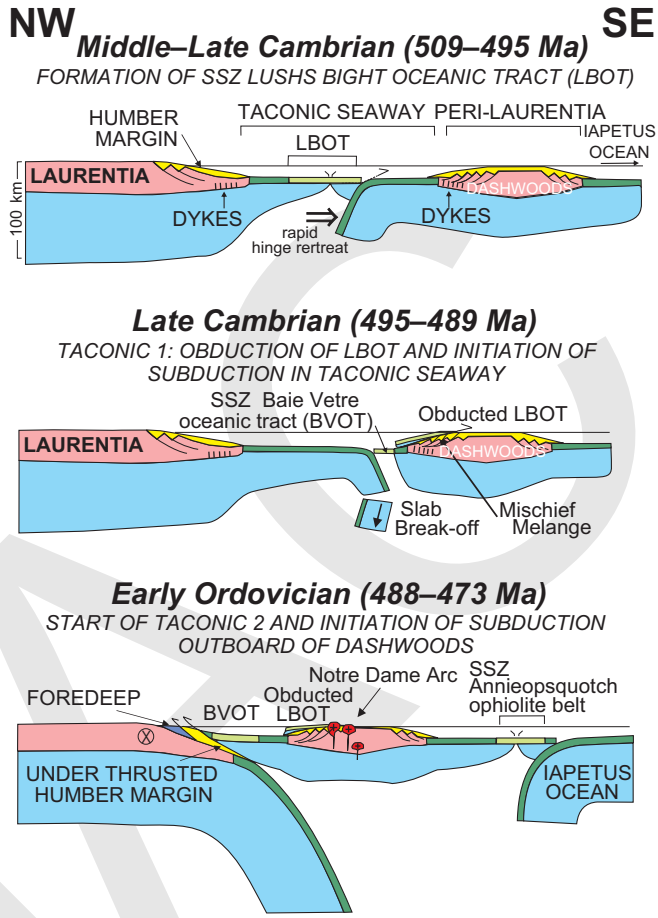


Figure 13. Middle Cambrian–Early Ordovician tectonic evolution of the peri-Laurentian realm. Dashwoods had rifted-off Laurentia at ca. 550 Ma, opening the Taconic seaway. Subduction initiation in the Taconic Seaway (Zagorevski and van Staal, in press) at ca. 515 Ma formed the Lushs Bight oceanic tract (LBOT). Obduction of LBOT onto Dashwoods comprises Taconic I. Initiation of west-directed subduction outboard of Dashwoods, forming the Annieopsquotch ophiolite belt, took place at ca. 480 Ma after Dashwoods had started to collide with promontories in the Humber margin, which slowed down convergence. Figure is modified from van Staal et al. (2007, 2009). Abbreviations BVOT: Baie Verte oceanic tract; SSZ: suprasubduction zone.

was apparently diachronous, becoming younger eastwards (Fig. 4). An alternative model suggests that rifting began earlier (ca. 550 Ma) and Avalonia migrated from a South American Gondwanan location towards a near West African location prior to its cross-ocean migration towards Laurentia (Samson et al. 2009; Satkoski et al 2010).

Meguma experienced two rifting events (see above). The first event was recorded by the Lower Cambrian (ca. 537 Ma and older) – Lower Ordovician (ca. 485 Ma) Goldenville and Halifax groups (Fig. 8), and was accom-

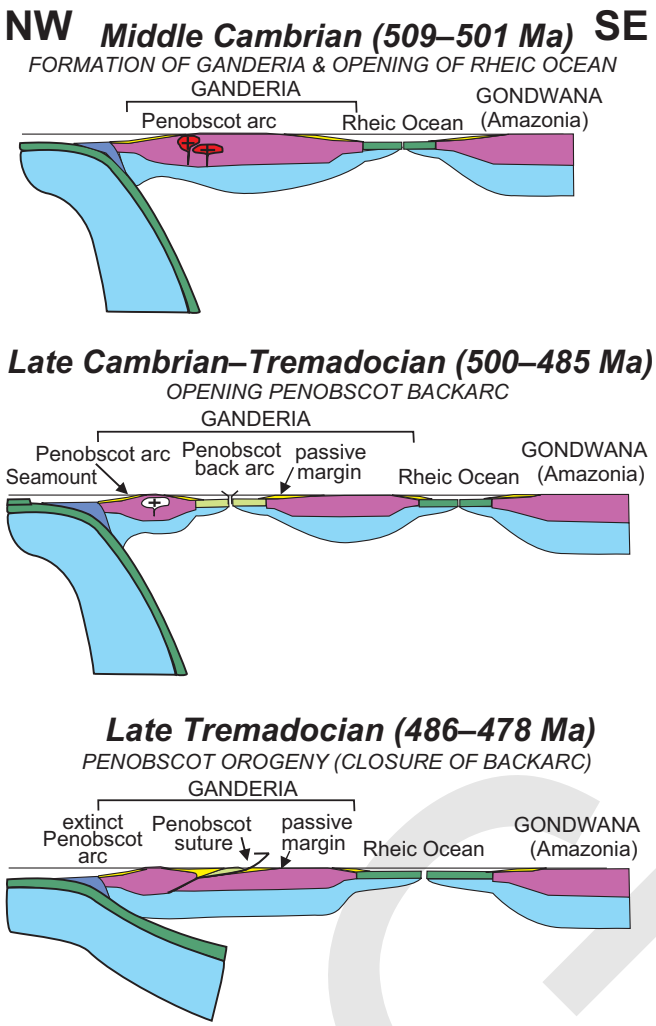


Figure 14. Formation of Ganderia and Middle Cambrian–Early Ordovician tectonic evolution of the Penobscot arc–back-arc system, culminating in the short-lived Penobscot orogeny (486–478 Ma) due to closure of the Penobscot back-arc basin. Back-arc closure is inferred to have been caused by shallowing of the downgoing Iapetan slab due to arrival of an oceanic plateau or seamount at the trench. Figure is modified from Zagorevski et al. (2010).

panied by emplacement of mafic sills and dykes on the presumed Avalonian margin of the rift (White and Barr 2004; Waldron et al. 2009). This rift has been ascribed to protracted rifting during departure of Avalonia, beginning in the late Ediacaran, and hence is more compatible with the alternative early-rifting model for Avalonian separation from Gondwana noted above. This model implies that part of Meguma remained attached to Avalonia, developing a widening rift basin in which the Goldenville and Halifax groups were deposited. However, the nature of sediments in the Goldenville and Halifax groups requires a large continental source, presumably Gondwana, and hence Meguma must also have stayed in close proximity to Gondwana through at least the Cambrian and Early Ordovician (Waldron et al. 2009; White and Barr 2010). A second rifting event is represented by Lower Silurian rift-related bimodal volcanic and sedimentary rocks (442–438 Ma) of the White Rock Formation (Schenk 1997; Keppie and Krogh 2000; MacDonald et al. 2002), which locally has a thickness of up to 6 km and may mark the onset of rifting and departure of Meguma from Avalonia (Fig. 8), at which time Meguma may have become an independent microcontinent in the Rheic Ocean (van Staal 2007). The volcanic rocks in the White Rock Formation are coeval with the Cape St. Mary's sills in Avalonia of southeastern Newfoundland (Greenough et al. 1993), which suggests that this rifting event may have been the culmination of a protracted period of rifting and separation between Meguma and Avalonia. If this interpretation is correct, then Meguma was situated on the same plate as Avalonia until at least the Early Silurian.

Other rifting events are represented by the formation of suprasubduction zone ophiolites and intra-arc and back-arc basins. Suprasubduction zone ophiolites are principally recognized by their abundance of boninite and/or island arc tholeiite, as described above. They occur in the peri-Laurentian Lushs Bight and Baie Verte oceanic tracts (Bédard et al. 1998, 2000) and the Annieopsquotch accretionary tract (Lissenberg et al. 2005b). The ophiolites formed over a ca. 20 my period between 510 and 480 Ma as a result of proto-forearc oceanfloor spreading during subduction initiation events (e.g. Stern and Bloomer 1992) and/or syn-collision roll-back of trapped old oceanic lithosphere in embayments near the Humber and/or Dashwoods margins (Cawood and Suhr 1992; Schroetter et al. 2003; van Staal et al. 2007). The rift architecture of the Newfoundland ophiolite massifs is generally not well known, although the volume of sheeted dykes varies significantly between ophiolites, suggesting that the spreading rate probably also varied. Based on a combination of evidence including the orientation of the synmagmatic normal faults and exhumation of mantle in oceanic core complexes, asymmetric rifting accompanied by slow-spreading was proposed for the Thetford Mines ophiolite in southern Quebec (Schroetter et al. 2003; Tremblay et al. 2009). Formation of intra-arc and back-arc basins took place both in the peri-Laurentian (Zagorevski et al. 2006, 2008) and peri-Gondwanan realms (Ganderia) (van Staal et al. 1991, 1998, 2003, 2009; Zagorevski et al. 2010) during and before the Appalachian orogenic cycle. These rifts are generally recognized on the basis of a combi-

nation of evidence, including volcanic geochemistry and stratigraphy. A characteristic of back-arc basin formation is that it leads to significant thinning of the underlying lithosphere, hence progressively reducing the interaction of melts with crust during the rifting. The latter can be tested using geochemistry.

Ganderia has preserved evidence for the formation of two separate back-arc basins (Figs. 7, 8), both of which led to ocean-floor spreading: the Penobscot back-arc basin that formed between 496 and 485 Ma, and the Tetagouche–Exploits back-arc basin that formed between 473 and 455 Ma. Opening of the Penobscot back-arc basin was either coeval with, or immediately postdated, the rifting of Ganderia from Gondwana (Fig. 14). Opening of the Tetagouche–Exploits back-arc basin started as an intra-arc rift, characterized first by multiple phases of extension-related, voluminous explosive subaqeous to subaerial calc-alkaline felsic volcanism, each phase shortly thereafter (2–5 Ma) followed by widespread mafic dyking and eruption of subaerial to marine tholeiitic basalt (e.g. Rogers and van Staal 2003; Rogers et al. 2003a). Over time the basalts changed in composition from tholeiitic and MORB-like (including ophiolite) to enriched, alkaline basalts while the associated sediments progressively became more pelagic, culminating in chert and black shale, which confirm establishment of oceanic conditions in the back-arc basin (van Staal et al. 2003). The eastern side of the back-arc basin became a passive margin, charcterized by sandstone and shale and little or no volcanism (van Staal and Fyffe 1995; Valverde-Vaquero et al. 2006a; Fyffe et al. 2009).

5.0 AGE AND STYLE OF OROGENY – FROM ACCRETION TO COLLISION

Accretion-related orogenesis in the Appalachians started in the Cambrian (or indeed in the Neoproterozoic if we consider the pre-Appalachian tectonic events recorded in Ganderia and Avalonia) and continued through the Carboniferous. Orogenesis was terminated by the Laurentia–Gondwana collision, which was finished in the Permian. During that time, the Appalachians changed from an accretionary orogen, as the Laurentian margin expanded as a result of the addition of a variety of tectonic elements and microcontinents, to a collisional orogen within Pangea.

Before the introduction of the plate tectonic paradigm, tectonic events were grouped into separate collisonal events or orogenies on the basis of age. This approach does not work well in the Appalachians where coeval but unrelated orogenesis took place on both sides of the Iapetus Ocean (see discussions by van Staal 1994 and van Staal and Hatcher 2010). Rather than creating a completely new nomenclature, the old orogenic classification has been retained, but restricted and modified to clarify which tectonic events are spatially and kinematically related.

5.1 Taconic Orogenic Cycle

The Taconic orogenic cycle was redefined by van Staal et al. (2007) to mainly comprise subduction and deformation related to accretion of oceanic and continental arcs in the peri-Laurentian realm between the Late Cambrian and Late Ordovician (500–450 Ma). Its orogenic cycle was terminated by arrival of the leading edge of Ganderia (Popelogan–Victoria arc) at the Laurentian margin between 460 and 450 Ma. Its full spectrum of critical rock assemblages and structures are best preserved and defined in western and central Newfoundland, augmented by data from Quebec and adjacent New England (Figs. 2, 5). As redefined, the Taconic orogenic cycle comprises three distinct tectonic events designated Taconic 1, 2, and 3.

5.1.1 Taconic 1

Taconic 1 comprises the initiation of subduction, formation of the Lushs Bight oceanic tract (LBOT), and its obduction onto the nearby Dashwoods microcontinent (Figs. 6, 13; Waldron and van Staal 2001; van Staal et al. 2007). The polarity and position of the LBOT, i.e. outboard or inboard of Dashwoods, is poorly known but we have adopted the new ideas of Zagorevski and van Staal (2011) (Fig. 13). Intraoceanic detachment of the LBOT lithosphere and its obduction onto Dashwoods started in the Middle to Late Cambrian, between 500 and 493 Ma (Szybinski 1995; Swinden et al. 1997), consistent with the ca. 495 Ma U-Pb zircon age (Jamieson 1988) of the metamorphic sole of the St. Anthony's Complex (Fig. 2). Structures other than the St. Anthony's metamorphic sole (thin sheets of rocks derived from the downgoing plate and accreted to the hot mantle of the detached upperplate ophiolite during or shortly after initiation of subduction) potentially associated with these events are Middle to Late Cambrian (510–493 Ma) mylonites in the Twillingate area (Williams and Payne 1975), chlorite-rich shear zones in Lushs Bight (Szybinski 1995) and mélanges in Dashwoods (Fox and van Berkel 1988; Hall and van Staal 1999). Upper Cambrian (500–488 Ma) crust-contaminated mafic dykes and plutons of the Notre Dame arc cut the shear zones and mélanges (Szybinski 1995; Swinden et al.

1997; Whalen et al. 1997b; van Staal et al. 2007). The mylonites and chlorite-rich shear zones accommodated a component of dextral transcurrent shear (Szybinski 1995), which has been related to dextral oblique convergence (Dewey 2002). Structures and metamorphism that could be related to the Taconic 1 have not been positively identified in Quebec, although some workers (e.g. Huot et al. 2002) argued for late Cambrian tectonic interaction between Laurentian continental crust and Mt. Orford (LBOT) oceanic lithosphere that could correspond with Taconic 1. The Cambrian age of amphibolite of the metamorphic sole to the ophiolitic Belvidere Complex (505–490 Ma; Laird et al. 1993) in northern Vermont near the border with Quebec, possibly a correlative of the Mt. Orford ophiolite, suggests that intraoceanic detachment had started nearly coevally with intraoceanic decoupling of LBOT lithosphere in Newfoundland.

5.1.2 Taconic 2

Taconic 2 was due to closure of the Taconic Seaway and culminated in prolonged collision of Dashwoods with the Humber margin (Figs. 6, 13, 15). The Taconic 2 orogenic cycle started with initiation of east-directed subduction in the oceanic Taconic Seaway at ca. 490 Ma. Subduction thus was directed beneath the then-composite Dashwoods-LBOT lithosphere (Figs. 6, 8, 13). This subduction zone is necessary to explain formation of the Upper Cambrian–Lower Ordovician (490–477 Ma) volcanic and plutonic rocks of the first phase of the ensialic Notre Dame arc, the suprasubduction zone oceanic lithosphere of the Baie Verte oceanic tract (BVOT) of similar age, and the obduction of the latter onto the Humber margin. The BVOT is compositionally similar to the LBOT with an abundance of boninite (Swinden et al. 1997; Bédard et al. 1998, 2000) and is for identical reasons interpreted to have formed during subduction initiation. While the Dashwoods and BVOT were invaded by continental arc magmas between 490 and 476 Ma, the ca. 485 Ma Bay of Islands Complex formed by transtensional rifting (Cawood and Suhr 1992; Kurth et al. 1998; Suhr and Edwards 2000; Dewey 2002) due to roll back of the down going slab in a second-order reentrant in the Laurentian margin characterized by trapped old, and hence dense, oceanic lithosphere. This phase of roll back became important after the slowing down of Taconic 2 subduction following arrival of Humber margin promontories at the trench (van Staal et al. 2007). The ca. 480 Ma Thetford Mines and younger Mt. Albert ophiolite complexes in Quebec (Dunning and Pedersen 1988; Whitehead et al. 2000; Malo et al. 2008) probably formed in a similar manner as the Bay of Islands Complex with the nearby ca. 505 Ma Mt. Orford and Belvidere ophiolites being remnants of the oceanic lithosphere in which they formed (Huot et al. 2002; Tremblay et al. 2009).

Loading of the Humber margin by oceanic lithosphere was well under way in Newfoundland by at least 475 Ma (Fig. 5; Knight at al, 1991; Waldron and van Staal 2001) and possibly slightly later in Quebec (ca. 468 Ma, Hiscott 1978; Malo et al. 2001; Lavoie et al. 2003). Loading predated the age of the Bay of Islands Complex metamorphic sole (469 ± 5 Ma, Dallmeyer and Williams 1975). If the latter age is correct and not tainted by excess argon, it supports the argument presented above that the Bay of Islands Complex was generated in a second- or third-order reentrant (Cawood and Suhr 1992) and hence that its obduction took place slightly later than obduction of BVOT lithosphere on nearby promontories (van Staal et al. 2007). The evidence discussed above and other data (e.g. Schroetter et al. 2006; Tremblay et al. 2009) suggest that most of the Taconic Seaway was closed before 468 Ma in both Newfoundland and Quebec. Hence the 462–459 Ma, voluminous second phase of the Notre Dame arc and accompanying deformation and metamorphism (Fig. 6) cannot be explained simply as a result of subduction of Taconic Seaway oceanic lithosphere. First, arc magmatism in Dashwoods was shut off by at least 476 Ma, and did not flare up seriously until ca. 464 Ma (van Staal et al. 2007), with the magmatic gap corresponding to the Dashwoods–Laurentia collision (Waldron and van Staal 2001). Second, rocks of the Notre Dame arc and internal parts of the Humber margin in both Quebec and Newfoundland were shortened and metamorphosed between 470 and 460 Ma (Whitehead et al. 1996, 2000; Castonguay et al. 2001; Pehrsson et al. 2003; Lissenberg et al. 2005a; Gerbi et al. 2006a), also consistent with collision. Late syn-collisional timing for the second phase of the Notre Dame arc is consistent with mutually crosscutting relationships between tonalite plutons and the main phase of deformation and associated amphibolite-facies metamorphism in the Dashwoods subzone (Fig. 6; van Staal et al. 2007). In addition, the tonalites include a relatively minor, but significant, component of non-arc-like rocks, lacking Th-enrichment with respect to Nb, whereas high La/Yb ratios suggest that garnet was locally present in the source area of both the arc and non-arc tonalites (Whalen et al. 1997a). Combined, these relationships have been explained by magma generation

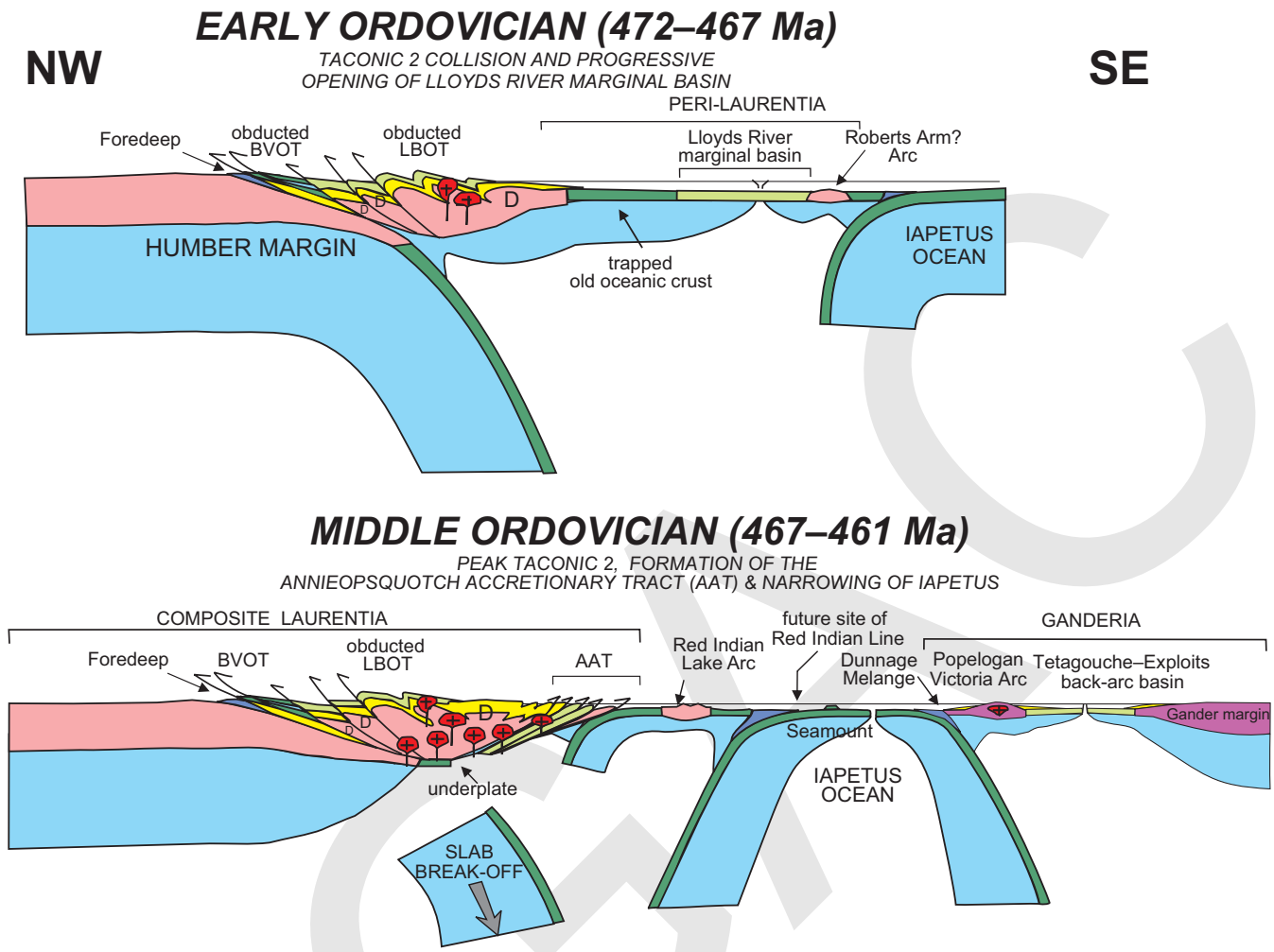


Figure 15. Early–Middle Ordovician tectonic evolution of the peri-Laurentian realm showing prolonged Taconic 2 collision between Dashwoods and Humber margin of Laurentia; modified from van Staal et al. (2007, 2009). Rapid closure of the main Iapetan tract in the Middle Ordovician coincided with bivergent subduction beneath the leading edge of Ganderia (Popelogan–Victoria arc) and peri-Laurentia (Red Indian Lake arc). Late Ordovician (ca. 455 Ma) collision between the two arcs comprises Taconic 3; it closed the main Iapetan tract and initiated Salinic subduction in the trailing Tetagouche–Exploits back-arc basin. Abbreviations AAT: Annieopsquotch accretionary tract; BVOT: Baie Verte oceanic tract; D: Dashwoods crust; LBOT: Lushs Bight oceanic tract.

following break-off of the Laurentian slab (Fig. 15) beneath a collision-thickened Notre Dame arc (van Staal et al. 2007). Ensialic calc-alkaline arc magmatism in the Ascott Complex in the eastern townships of Quebec also overlaps with the age of peak Taconic 2 deformation and metamorphism here (Whitehead et al. 1996; Castonguay et al. 2001), suggesting a similar syn-collisional setting as in Newfoundland.

Collision between the Humber margin and Dashwoods microcontinent probably started at promontories at ca. 480 Ma and slowed down convergence between these two areas. Transfer of convergence into Iapetus, outboard of Dashwoods, is inferred to be the cause of initiation of a new, west-dipping subduction zone (Fig. 13) immediately to the east of Dashwoods (van Staal et al. 2007). It produced the ca. 481–473 Ma Annieopsquotch suprasubduction zone ophiolite belt (AOB; Lissenberg et al. 2004, 2005b) and other arc to back-arc-like magmatism until the Late Ordovician (Zagorevski et al. 2006, 2008). Approximately 10 my after its initiation (ca. 470 Ma), the AOB, which was by then positioned in a back-arc setting (Lloyd's River back arc), due to roll back of the west-dipping Iapetan slab (Fig. 15), was being underthrust and underplated beneath the retroarc side of the Notre Dame arc, together with a crustal flake containing the Buchans–Roberts Arm arc (Figs. 7, 15; Lissenberg et al. 2005a; Zagorevski et al. 2006, 2008). West-directed underthrusting led to the closure of the

Lloyds River back-arc basin, which probably started out as east-directed retro-arc overthrusting of the Notre Dame arc induced by the expanding Dashwoods–Laurentia hard collision further to the west (Figs. 7, 15). Subduction of relatively young oceanic lithosphere of the Lloyd's River back arc may have been sustained by trapped old dense oceanic lithosphere in this basin along strike (Zagorevski and van Staal 2011). The underplating-related accretion initiated the Annieopsquotch Accretionary tract (AAT).

5.1.3 Taconic 3

The main Iapetan tract that was situated between the AAT and the peri-Gondwanan Popelogan–Victoria arc (PVA, Figs. 3, 4, 7, 8, 15), the leading edge of Ganderia, had a width of at least 2500 kilometers at 475 Ma (van Staal et al. 1998) but was from then on being closed by two outward-dipping subduction zones (Fig. 15), accelerating convergence between the two opposing arc systems (Figs.7, 8, 15). This convergence terminated in a Moluccan Sea-style arc–arc collision in the Sandbian–Katian (455–450 Ma), which sutured along the Red Indian line (Figs. 2, 3, 7). Laurentia-derived detritus in upper Caradoc (ca. 450 Ma) sediments of the Badger Group that conformably overlies the Victoria arc in Newfoundland (Fig. 7) confirms that the two arcs were welded together by this time (McNicoll et al. 2001; Waldron et al. 2012). Arc–arc collision also terminated the Taconic orogenic cycle, leaving Iapetan oceanic lithosphere only in the Tetagouche–Exploits back-arc basin and in the Acadian Seaway between Ganderia and Avalonia (Figs. 3, 4).

Deformation associated with Taconic 3 arc–arc convergence and collision is represented by mélanges (Figs. 2, 5, 15) along the Red Indian line (van Staal et al. 1998; McConnell et al. 2002), sinistral oblique reverse faulting, folding, and thrusting and associated metamorphism in the AAT (Lissenberg and van Staal 2006; Zagorevski et al. 2007a) and to a much lesser extent in the Popelogan–Victoria arc, which after the onset of collision became the downgoing plate. Metamorphism locally achieved amphibolite-facies conditions in the AAT, but is generally of greenschist facies or lower grade (Lissenberg et al. 2005a; Zagorevski et al. 2007a, 2009). Regional assemblages with pumpellyite, prehnite, epidote, and actinolite are common in the collision zone in Newfoundland (Franks 1974, Zagorevski et al. 2008) and Maine (Richter and Roy 1974), suggesting that tectonic burial did not exceed depths in the order of 2–5 kb (Liou et al. 1985). Metamorphism in the exposed part of the Popelogan arc in New Brunswick never exceeded zeolite-facies conditions (Wilson 2003).

5.2 Penobscot Orogenic Cycle

The Penobscot orogenic cycle started with subduction beneath the leading edge of Ganderia (Figs. 4, 14) and development of an arc (Penobscot arc) at ca. 515 Ma. It culminated with Penobscot orogenesis between 486 and 478 Ma in both Newfoundland and New Brunswick (van Staal 1994; Johnson et al. 2009; Zagorevski et al. 2010). The Penobscot orogenic cycle chronologically overlaps with the early phases of the Taconic 2 cycle in the peri-Laurentian realm (Fig. 5), but is restricted to Ganderian rocks (Neuman 1967; Colman-Sadd et al. 1992a) while Ganderia was still attached to or situated in the periphery of Gondwana at high southerly latitudes (Liss et al. 1994; van Staal et al. 1998; Schultz et al. 2008) on the opposite side of the Iapetus Ocean from the Laurentian margin. Given that the two opposing margins were more than 3000 km apart, no direct tectonic link existed between these two orogenic systems, despite usage to the contrary by some workers (e.g. Pinet and Tremblay 1995).

The nature of Penobscot orogenesis is best understood in central Newfoundland (Colman-Sadd et al. 1992a), but is supported by relationships preserved in New Brunswick (Johnson et al. 2009). In central Newfoundland, an early Arenig (ca. 474 Ma) granite stitches an Upper Cambrian (ca. 494 Ma) suprasubduction zone ophiolite belt (Fig. 2, Coy Pond and Pipestone Pond complexes) that was obducted onto quartz arenite of the Gander margin (Colman-Sadd et al. 1992a) between 486 and 475 Ma (van Staal 1994). The ophiolites were interpreted to have formed in a back arc (Jenner and Swinden 1993) related to coeval Cambrian–Early Ordovician volcanic arc rocks farther to the west in the Victoria Lake Supergroup and Wild Bight and Exploits groups (Figs. 2, 7, 8, 14) (Zagorevski et al. 2007b, 2010). This relationship suggests formation of the upper plate rocks (referred to as the Penobscot arc–back-arc complex) above an east-dipping subduction zone (Rogers et al. 2006; Zagorevski et al. 2007b, 2010). An overall westward younging of the Cambrian volcanic arc rocks in the Victoria Lake Supergroup (Fig. 2; Rogers et al. 2006; Zagorevski et al. 2007b, 2010) suggests west-directed arc-trench migration over time as a result of roll back of an east-dipping slab. The cause of ophiolite obduction, i.e. closure of the Penobscot back-arc basin, is speculative, although the Penobscot arc shut off for at least 4–6 Ma (486–478 Ma) until its resurrec-

tion as the Victoria arc at ca. 475 Ma (van Staal 1994; O'Brien et al. 1997). Shut-off of the Penobscot arc and contraction in its back arc were possibly due to the arrival of a large seamount at the trench (Fig. 14), which caused shallowing of the subduction angle (flat-slab subduction), which in turn increased the traction between the two plates. Jacobi and Wasowski (1985) recognized seamount-related basalt (Summerford seamount) adjacent to the Red Indian line on New World Island in northern Newfoundland (Fig. 2) and as large blocks in the Dunnage mélange. These basalts are in part at least as old as Early Ordovician (Kay 1967) and may be a remnant of this postulated seamount–trench collision (van Staal et al. 1998, 2010; Zagorevski et al. 2010).

Deformation associated with closure of the Penobscot back arc and ophiolite obduction in central Newfoundland is largely restricted to high-level thrusting and associated black shale mélange, which contains blocks of ultramafic rock and Gander Group quartzite (e.g. Williams and Piasecki 1990), folding, and foliation formation. Differences in structural intensity and complexity between the Ordovician sedimentary rocks below (Gander Group) and above (Davidsville–Baie d'Espoir groups) the obducted ophiolites, combined with the presence of pre-entrainment foliations in the ultramafic blocks in the mélange and conglomerates (e.g. Kennedy 1975; Currie 1992), suggest that Penobscot obduction may have caused penetrative deformation and low-grade metamorphism that was subsequently obscured by later deformation.

5.3 Salinic Orogenic Cycle

The prelude to Salinic orogenesis started after accretion of the leading edge of Ganderia, represented by the Popelogan–Victoria arc, to Laurentia during Taconic 3, when the west-dipping subduction zone beneath composite Laurentia stepped back into the Tetagouche–Exploits back-arc basin (Figs. 8, 15, 16a; van Staal 1994; van Staal et al. 1998, 2003 Valverde-Vaquero et al. 2006a). Subsequent convergence between the trailing Gander margin and composite Laurentia was sinistral-oblique. Evidence for Salinic subduction of the back-arc basin lithosphere includes 447–430 Ma blueschist, high-pressure greenschist, ophiolite, and mélange in the east-verging Brunswick subduction complex in northern New Brunswick (van Staal et al. 1990, 2003 2008; Currie et al. 2003). In northern and central Newfoundland, accretion-related Salinic structures include mélange (e.g. Lower Silurian (435–430 Ma) Joey's Cove mélange), folds, and thrust faults (Williams et al. 1988; Elliot et al. 1991; Lafrance and Williams 1992; Currie 1995; Lee and Williams 1995; O'Brien 2003; Zagorevski et al. 2007a; van Staal et al. 2009) (Fig. 8). The upwards-shallowing Badger Group and equivalents in New Brunswick and Maine were deposited in the arc–trench gap (Figs. 8, 16a; Pickering et al. 1988; Ludman et al. 1993; van Staal and de Roo 1995; Kusky and Kidd 1996; Wilson et al. 2004; Zagorevski et al. 2008; Reusch and van Staal 2012) of the west-dipping subduction zone (van Staal 1994; van Staal et al. 2008, 2009). Magmatic rocks related to this phase of subduction are represented by the third, ca. 445–435 Ma phase of the Notre Dame arc (Figs. 6, 16a), the volcanic component of which lies unconformably on the products of the second phase. This last phase of the Notre Dame arc was not voluminous (Whalen et al. 2006), which is not surprising because the back-arc basin was probably not wider than 1000 km (van Staal 1994; van Staal et al. 1998) and hence subduction of its oceanic lithosphere was short-lived. Nevertheless, in additon to Newfoundland, volcanic arc rocks related to this phase occur also in Quebec (David and Gariepy 1990), New Brunswick, and Maine (Wilson et al. 2008).

Lower Silurian sediments on the lower and upper plates of this subduction system in Newfoundland remained distinct (Figs. 2, 7, 8, 16a) and had different source areas (Williams et al. 1993; Pollock et al. 2007) until ca. 423 Ma, when the basin had closed along the Dog Bay line (Figs. 2, 7) and was lifted above sea level (Williams et al. 1993). Terrestrial sediments were deposited on both sides of the Dog Bay line after basin closure (Currie 1995; Pollock et al. 2007). Geochronological constraints on the age of structures indicate that closure was approximately coeval in New Brunswick (Bamford Brook fault system) and Maine (Codyville–Liberty–Orrington fault; West et al. 1992, 2003; Ludman et al. 1993; Tucker et al. 2001; van Staal et al. 2008) (Figs. 2, 7, 8). The Fredericton and Indian Island belts immediately southeast of the Salinic sutures represent marine foredeep sequences formed by tectonic loading of the back-arc basin's passive (Gander) margin by foreland-directed thrust sheets (Figs. 7, 8, 16a; van Staal and de Roo 1995; Tucker et al. 2001, p. 225; Park and Whitehead 2003). Lithological similarities between parts of the Lower Silurian forearc and foredeep sediments of the Central Maine belt in the USA and Fredericton belt, respectively (Tucker et al. 2001; Reusch and van Staal 2012) suggest that sediments locally could spill over the forearc basin rim and be transported across the accretionary prism into the trench (Dickinson and Seely 1979). Cross-cutting plutons indi-

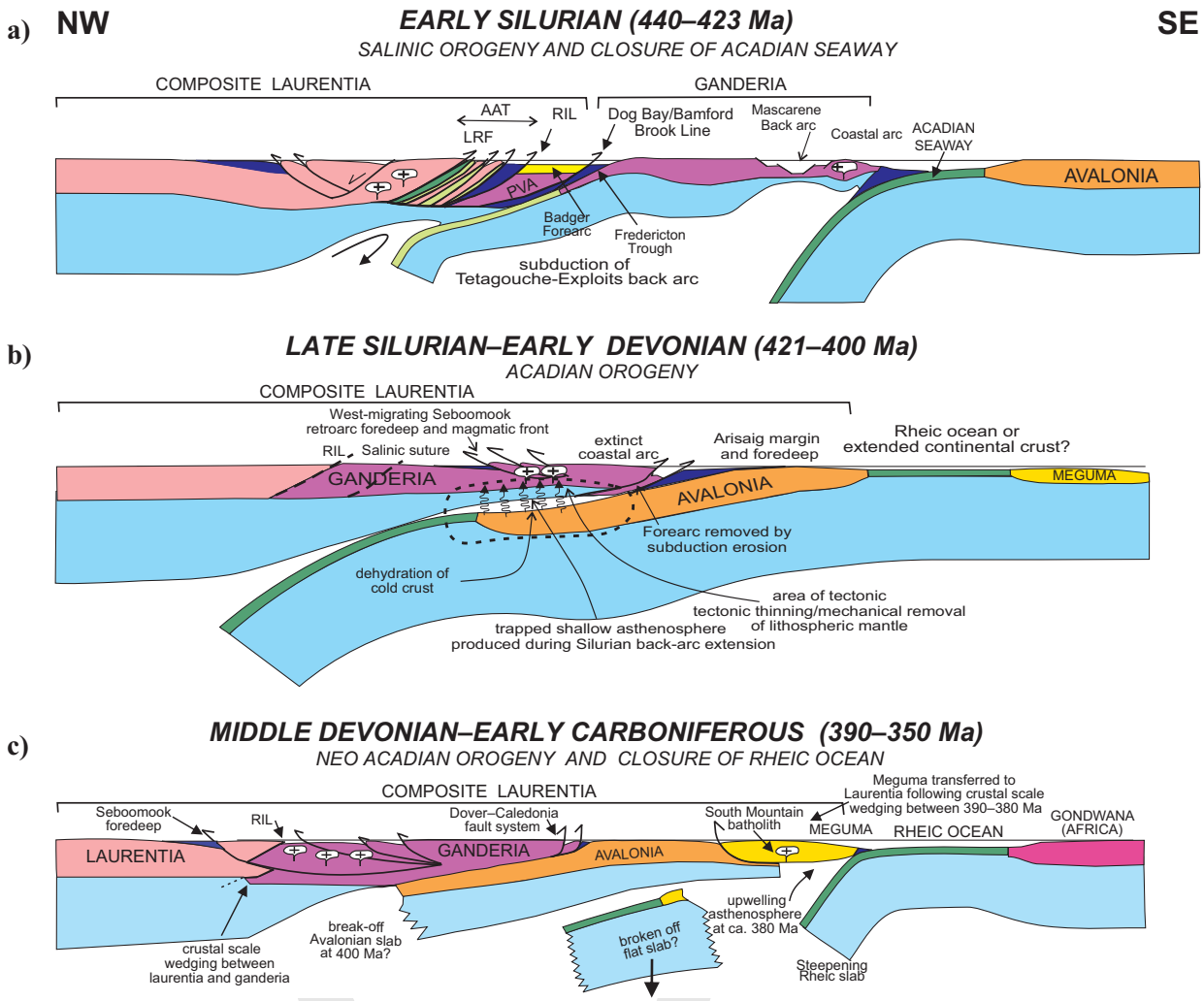


Figure 16. Silurian–Carboniferous tectonic evolution of the northern Appalachian orogen due to sequential arrival of peri-Gondwanan microcontinents and progressive growth of composite Laurentia; modified from van Staal et al. (2009). **a)** Late Ordovician–Early Silurian closure of the Tetagouche–Exploits back-arc basin, which led to Salinic collision between the Gander margin and composite Laurentia, coeval with subduction of the Acadian seaway beneath the trailing edge of Ganderia. **b)** Late Silurian–Early Devonian Acadian collision between Avalonia and composite Laurentia following closure of the Acadian seaway. Note that pockets of subduction-modified mantle were trapped above the progressively shallowing Avalonian plate, which locally led to arc-like magmatism. Also note that most of the forearc lithosphere to the coastal arc was subducted and probably underplated beneath Ganderia. Partial melting of this underplated material may also have led to arc-like magmatism. **c)** Late Early Devonian–Late Devonian Neoacadian protracted collision of Meguma with composite Laurentia. Convergence between these two elements could be due to flat-slab subduction of the Rheic Ocean, putting the upper plate containing Avalonia and Meguma in compression, or due to closure of an oceanic seaway between these two terranes. It is uncertain whether Avalonia and Meguma were separated by a Rheic oceanic seaway or connected by highly extended continental and/or transitional crust. The later stages of collision were coeval with progressive steepening of the subducting Rheic slab outboard of Meguma. The latter may have been achieved following break-off of the segment of the slab that was originally situated beneath and/or attached to Meguma. Seismic data suggest that most of the Meguma crust was delaminated from its underlying lithospheric mantle and wedged above the leading edge of Laurentia (Avalonia). Steepening of the downgoing Rheic oceanic slab and inflowing of asthenosphere into the mantle wedge generated subduction-related magmas in the leading edge of composite Laurentia, which in Meguma assimilated a large volume of sedimentary rock (Tate and Clarke 1995; Clarke et al. 1997, 2004). The component of strike-slip movement of terranes such as Meguma and Avalonia with respect to composite Laurentia is not shown in these cross-sections because their absolute amount of translation is at present unknown, although it probably was significant. Devonian–Early Carboniferous subduction of the Rheic Ocean beneath the leading edge of composite Laurentia culminated in the Alleghanian collision between Africa (lower plate) and composite Laurentia between 340 and 330 Ma. Abbreviations AAT: Annieopsquotch accretionary tract; LRF: Lloyd's River fault; PVA: Popelogan–Victoria arc; RIL: Red Indian line.

cate that structural inversion of the Fredericton belt in New Brunswick and Maine took place between 426 and 422 Ma (West et al. 1992; Tucker et al. 2001), which represents the terminal phase of the Salinic collision between Laurentia and Ganderia.

An inferred angular unconformity between the penetratively folded, steeply dipping Lower Silurian Botwood Group and the overlying, gently dipping ca. 423 Ma Stony Lake volcanics (Anderson and Williams 1970; Dunning et al. 1990a) west of the Dog Bay line supports other evidence indicating that the Salinic orogeny was approximately coeval in Newfoundland and New England and restricted to the Late Ordovician–Silurian (445–423 Ma). Deformation and metamorphism were particularly intense in Gander margin rocks in Newfoundland, Cape Breton Island (Lin 2001), New Brunswick, and southern Maine. Deformation locally was accompanied by generation of migmatite in eastern Newfoundland (Colman-Sadd et al. 1992a; D'Lemos et al. 1997; Schofield and D'Lemos 2000) and central New Brunswick (Fyffe et al. 1988a). These parts of the Appalachians generally also underwent intense orogenesis during the subsequent Acadian orogeny, obscuring relationships (Figs. 8, 16b). However, as described above, the deep crustal architecture of the Salinic orogen and the remnants of its west-dipping subduction channel have been imaged on reprocessed seismic lines in Newfoundland (Fig. 9).

Salinic orogenesis was not restricted to the Laurentia–Ganderia collision zone, because rocks were deformed as far west as the Humber margin (e.g. Cawood et al. 1994; Castonguay and Tremblay 2003). It also caused rapid exhumation of deep levels of the Notre Dame arc in Newfoundland (Pehrsson et al. 2003; van Staal et al. 2007).

5.4 Acadian Orogenic Cycle

The predominantly Early Devonian Acadian orogenic cycle was both kinematically and chronologically distinct from the Salinic orogenic cycle (van Staal 2005; van Staal et al. 2009). It was related to Silurian (442–421 Ma) convergence during closure of the narrow Acadian Seaway (Figs. 3, 4, 16a,b) that separated Avalonia and Ganderia (van Staal 2005, 2007; Valverde-Vaquero et al. 2006a) and a protracted collision between Avalonia and composite Laurentia from 421 to 390 Ma. At the onset of collision (ca. 421 Ma), Ganderia was already fully assembled into composite Laurentia as a result of the immediately preceding Salinic orogeny. The Acadian orogeny was named after the original Acadian settlements in eastern Maine, New Brunswick, and Nova Scotia, where evidence for post-Early Silurian, pre-Late Devonian deformation (Fig. 6) is well preserved, as was recognized very early on during geological investigations in this part of the Appalachians (see Robinson et al. 1998 for a more complete review).

Evidence of Acadian subduction is best preserved in the Kingston belt in southern New Brunswick (Fig. 2), which was built on Neoproterozoic–Early Cambrian basement of Ganderia's trailing edge (Barr et al. 2002; Fyffe et al. 2009). The Kingston belt comprises both Early Silurian arc volcanic and intrusive rocks (442–435 Ma; Doig et al. 1990; Barr et al. 2002) and a sliver of Silurian (435–416 Ma) high-pressure low-temperature forearc rocks (White et al. 2006). The Kingston belt continues to the south into the Coastal Volcanic belt of Maine (Gates 1969) and to the northeast through correlative rocks in Cape Breton Island to the south coast of Newfoundland (Barr and Jamieson 1991; Lin et al. 2007). In Cape Breton Island, the volcanic arc rocks form the Sarach Brook Metamorphic Suite, Money Point Group, and equivalent units (Fig. 2), as well as related plutons with ages of ca. 442–427 Ma (Dunning et al. 1990b; Barr and Jamieson 1991; Keppie et al. 1992; Price et al. 1999; Lin et al. 2007). In southern Newfoundland, the arc is represented by the ca. 429 Ma Burgeo granite (Dunning et al. 1990a; Kerr et al. 1995) and the 429–422 Ma La Poile Group (O'Brien et al. 1991), which has been interpreted as an intra-arc basin (van Staal 2007; van Staal et al. 2009). Further to the east, this arc is truncated by the Dover–Hermitage Bay fault (Figs. 1, 7).

Because it largely follows the outline of the Atlantic coast in Maine, this arc has been referred to historically as the Coastal arc (e.g. Bradley 1983), and that name is retained here. The arc rocks of the Kingston belt have been intruded by numerous mafic dykes, locally sheeted, suggesting that the arc was extensional (e.g. Nance and Dallmeyer 1993; McLeod et al. 2001; Barr et al. 2002). Silurian mafic dyke swarms are also ubiquitous in the adjacent Mascarene basin where voluminous and largely consanguineous bimodal volcanic rocks generally have mixed compositions ranging from within-plate to arc environments (McLeod 1997), suggesting a back-arc basin setting (Fyffe et al. 1999; Miller and Fyffe 2002), with respect to the arc rocks of the Kingston belt. Hence, subduction was to the northwest beneath composite Laurentia (Ganderia), and Avalonia was situated on the lower plate. Deposition and magmatism in the Mascarene basin largely overlaps with that in the Sarach

Brook Metamorphic Suite and Money Point Group in Cape Breton Island and the La Poile Group in southern Newfoundland.

The onset of collision between Avalonia and composite Laurentia is best constrained by the inversion of both the Mascarene and La Poile basins at ca. 421 Ma (O'Brien et al. 1991; Fyffe et al. 1999), which accompanied inboard migration of the Coastal arc (McLaughlin et al. 2003), and Late Silurian (420–418 Ma) loading of Avalonia as recorded in the Arisaig Group in the Antigonish Highlands (see above and Waldron et al. 1996). After the onset of collision, Acadian deformation was mainly characterized by prolonged deformation and magmatism in the soft, back-arc region. The retro-arc deformation and magmatic fronts progressively migrated towards the west (Bradley et al. 2000; Bradley and Tucker 2002) over time (Figs. 8, 16b), the migration of the retro-arc Seboomook foredeep tracking the location of the deformation front.

5.5 Neoacadian Orogenic Cycle

The term Neoacadian orogeny was introduced by Robinson et al. (1998) to cover Middle Devonian to Early Carboniferous deformation and metamorphism in southern New England. It was later changed to Quaboagian by Robinson et al. (2007) and the Famennian event by Hibbard et al. (2010). These tectonic events correlate chronologically with dextral-oblique docking and protracted collision of Meguma with composite Laurentia, which started during the latest Early Devonian (ca. 395 Ma) and lasted into the Early Carboniferous (ca. 340 Ma) (Hicks et al. 1999; Keppie et al. 2002; Reynolds et al. 2004b). Terminal collision of Meguma with composite Laurentia was interpreted as the principal cause of the main, Late Devonian phase of Neoacadian/Quaboagian orogenesis throughout the Northern Appalachians (van Staal 2007; van Staal et al. 2009). However, Hibbard et al. (2010) expressed caution about using this nomenclature on the basis of doubts that all events in this period in New England and Atlantic Canada truly represent an orogenic event that was kinematically separate from the subsequent Alleghanian-related orogenic events (see below).

Accretion of Meguma was dextral oblique and mainly accommodated by the Cobequid–Chedabucto fault system on land in Nova Scotia (Figs. 1, 2, 7). Convergence between composite Laurentia (Avalonia) and Meguma continued into the Early Carboniferous and locally included a component of northwest-directed thrusting (Pe-Piper and Piper 2002). The mode of accretion of Meguma to composite Laurentia (Avalonia) is not well constrained at present. Geophysical evidence (see above) and xenolith studies in Upper Devonian dykes (Greenough et al. 1999) suggest that a wedge of Avalonian crust was structurally thrust beneath Meguma in the northern mainland area of Nova Scotia. In contrast, upper mantle reflectors have been interpreted to suggest that a northwest-dipping subduction zone was present beneath Nova Scotia (Keen et al. 1991a). Murphy et al. (1999) postulated that the dip of that subduction zone was very shallow (flat slab) due to interaction with a rising plume. Whatever the cause of a flat-slab setting, it explains the scarcity or absence of Early Devonian or younger arc magmatism in the leading edge of the composite Laurentian (Avalonian) upper plate. However, that absence could also be a result of highly oblique convergence of Meguma along a system of major orogen-parallel strike-slip faults (e.g. Barr et al. 2010) and/or the absence of sufficient wide oceanic lithosphere between these two terranes to generate arc magmatism. The lithospheric configuration of Meguma partly underlain by Avalonian crust suggests wedging of Meguma by the leading (Avalonian) edge of Laurentia (Fig. 16c). Such a process explains accretion and transfer of Meguma to the upper (Laurentian) plate and the continuation of northwest-directed thrusting along the Cobequid–Chedabucto fault system into the Carboniferous (Waldron et al. 1989; Pe-Piper and Piper 2002). However, whether Avalonia and Meguma were truly separated by Rheic oceanic lithosphere rather than thinned continental and/or transitional crust (Fig. 16b) is poorly known at present. Normal steep subduction under composite Laurentia and re-establishment of an asthenospheric wedge above the downgoing slab capable of producing mafic to intermediate arc magmas was probably renewed after 395 Ma (Fig. 16c). Interaction (assimilation and/or melting) of such magmas (e.g. Tate and Clarke 1995) with the thick sedimentary pile of the Goldenville and Halifx group and/or underlying Neoproterozoic sedimentary rocks probably gave rise to the 380–370 Ma S-type granites such as the South Mountain batholith (Clarke et al. 1997, 2004; Reynolds et al. 2004a; Collins and Richards 2008). Continuing Late Devonian–Early Carboniferous subduction produced younger 363–357 Ma magmatism further southeast in Meguma (Moran et al. 2007), suggesting a component of slab roll-back. Regardless, arc magmatism beneath Meguma implies that it was not a promontory of Africa, but a microcontinent in the Rheic Ocean and hence its accretion was not directly related to the subse-

quent Alleghanian collision between composite Laurentia (by then including Meguma) and Gondwana.

5.6 Alleghanian Orogenic Cycle

The Alleghanian cycle encompasses all events related to final closure of the Rheic Ocean (Fig. 16c) and collision/amalgamation of composite Laurentia with the main mass of Gondwana, which led to formation of the supercontinent Pangea. Much of the Alleghanian collision zone in the northern Appalachian orogen is now submerged beneath the Atlantic margin (Figs. 1, 7) and most of the available constraints are derived from rare syncollision Alleghanian plutons (e.g. Pe-Piper et al. 2010), paleomagnetic and paleontologic data (McKerrow et al. 2000; Torsvik and Cocks 2004), dating of Alleghanian shear zones and associated metamorphism in the leading edge of composite Laurentia (Waldron et al. 1989; Culshaw and Liesa 1997; Culshaw and Reynolds 1997), and events preserved in the hinterland of the orogen, particularly from upper Devonian–Permian rocks of the Maritimes Basin (e.g. van de Poll et al. 1995). The first three lines of constraint suggest that the Alleghanian collision had started before 320 Ma and probably took place during the late Mississippian.

The Maritimes Basin shows a complex tectonic evolution. Basin formation started in the Late Devonian and was associated with protracted dextral strike-slip motion along the already-established terrane-bounding faults in composite Laurentia, which suggests that composite Laurentia formed the upper plate and Gondwana the lower plate during the final stages of Rheic Ocean closure. Similar polarity has been proposed for the southern Appalachians (e.g. Hatcher 1987). The Maritimes Basin appears to have formed mainly by extension associated with a releasing stepover beneath the Gulf of St. Lawrence (Bradley 1982; Hibbard and Waldron 2009). Early subsidence was also associated with a thick underplate of mafic magmatic rocks in the centre of the basin (Marillier and Verhoef 1989). In the first stages of basin formation, predominantly nonmarine sediments comprising red beds and lacustrine shales, and locally, bimodal volcanic rocks were deposited. Subsequent basin evolution was punctuated by several marine transgressions (incursions of the Windsor Sea) and basin inversions, the first of which took place during the Early Carboniferous (ca. 345–340 Ma) and which included folding and thrusting (Reynolds et al. 2004b; Park and St. Peter 2005; White et al. 2006). The main phase of shortening that affected the basin more regionally took place at the end of the Visean (335–325 Ma). In addition, orogenesis and formation of the Alleghanian clastic wedge in the Southern Appalachians had started by at least 335–330 Ma (Hatcher 2002; Hibbard et al. 2010 and references therein). Combined, all lines of evidence suggest that the dextral-oblique Alleghanian Laurentia–Gondwana collision had started by at least 340 Ma in the Northern Appalachians. Collisional orogenesis continued into the Middle Permian (ca. 260 Ma) and comprised continuous dextral movements, which may have been due to protracted southwest-directed motion and clockwise rotation of Gondwana (Africa) along the Laurentian continent, progressively closing any remaining Rheic oceanic lithosphere in a zipper-like fashion (Hatcher 2002).

6.0 CRUSTAL GROWTH

Most of Laurentia's growth during the Appalachian orogenic cycle was due to transfer to Laurentia of already existing peri-Laurentian (Dashwoods) or peri-Gondwanan (Ganderia, Avalonia, and Meguma) continental crust and associated continent-derived sediment. However, significant volumes of new crust also were added due to accretion of oceanic suprasubduction zone terranes, particularly oceanic proto-forearc (suprasubduction zone ophiolite), arc, and back-arc basin rocks, both by obduction (e.g. Baie Verte oceanic tract) and underthrusting [e.g the Brunswick subduction complex (van Staal et al. 2003, 2008) and Annieopsquotch accretionary tract (Zagorevski et al. 2009)]. In addition, large volumes of mantle-derived arc and back-arc magmatic rocks (e.g. Notre Dame arc, Penobscot arc–back-arc; Popelogan–Victoria arc, Tetagouche–Exploits back-arc basin, Coastal volcanic arc) were added. However, the poor preservation of forearc terranes in general suggests that growth was at least in part compensated by removal of crust during subduction erosion, which led to return of tracts of forearc lithosphere back into the mantle, during the various tectonic events.

7.0 NATURE AND RECOGNITION OF PALEO-SUTURE ZONES

The various lithotectonic elements (Fig. 3) described above were accreted piecemeal to Laurentia, either as separate ribbons or as composite terranes that had amalgamated outboard in the Iapetus or Rheic oceans (van Staal et al. 2009; Zagorevski et al. 2009). As a result, composite Laurentia grew in size over time (see above). These various lithotectonic elements are separated by sutures, all of which are marked by major ductile and/or brittle faults (Fig. 3). These faults were reactivated mul-

tiple times during the life-span of the Appalachian orogen and in part even during opening of the Atlantic Ocean (e.g. Stewart et al. 1993). In general, arc-trench gap rocks are poorly preserved near the sutures. Forearc-foredeep basin pairs (van Staal et al. 2009), forearc mélanges (van Staal et al. 1998, 2010; Schroetter et al. 2006) and rare exhumed low-temperature high-pressure metamorphic rocks (e.g. White et al. 2001, 2006; van Staal et al. 2008; Zagorevski et al. 2008, 2009) are evidence indicating that trench and forearc terranes are at least partially preserved in some places. However, the paucity of such evidence indicates that most such rocks were either strongly overprinted by subsequent tectonism and metamorphism, so that their original tectonic setting is unclear, or were largely removed by erosion and/or structural processes such as subduction erosion (van Staal et al. 2009).

The suture between the LBOT and Dashwoods is poorly preserved, but is now marked by curviplanar faults juxtaposing Cambrian juvenile oceanic lithosphere and unrelated ensialic Ordovician arc rocks in Notre Dame Bay, Newfoundland (Fig. 2), and a poorly preserved and subsequently metamorphosed mélange belt in southwestern Newfoundland (as described above). The LBOT is now everywhere underlain by Dashwoods Laurentian crust, and hence the suture between these two terranes is probably, overall, a shallowly dipping structure. The Baie Verte–Brompton line, sometimes also loosely referred to as the Taconic suture, separates the Humber margin and the BVOT with its trailing Notre Dame arc/Dashwoods terrane (Fig. 2). This fault zone, which in many places was remobilized during subsequent episodes of deformation, is marked by a string of discontinuous, commonly dismembered ophiolite bodies and associated mélange along its entire length (e.g. Williams and St. Julien 1982). This fault is difficult to image geophysically (Keen et al. 1986; Spencer et al. 1989), but recent seismic interpretations in Newfoundland suggest that it is subvertical and cuts through the lower crust of the Humber margin to a depth of approximately 30 km (10 s) (Fig. 9; van der Velden et al. 2004). If correct, the imaged fault is relatively young and cannot represent the original suture at depth, because the Humber margin crust dips shallowly and extends further east at depth (Keen et al. 1986; Spencer et al. 1989; Hall et al. 1998; van der Velden et al. 2004).

The suture between the Notre Dame arc/Dashwoods terrane and ophiolite belts of the Annieopsquotch accretionary tract was imaged in central Newfoundland as a shallowly to moderately west-dipping fault zone that continues into the mantle and coincides on surface with the Lloyd's River fault and Hungry Mountain–Lobster Cove fault system (Fig. 9). The latter fault system accommodated sinistral oblique reverse movements during the Middle Ordovician (470–460 Ma) (Lissenberg and van Staal 2006), but was reactivated during the Silurian Salinic cycle, again as a sinistral-oblique reverse fault (Zagorevski et al. 2007b, 2009; van Staal et al. 2009).

The Red Indian line (Williams et al. 1988) separates the eastern limit of the Annieopsquotch accretionary tract from the Popelogan–Victoria arc (Figs. 2, 7, 8) and is the most fundamental Late Ordovician suture that separates Laurentia-derived from Gondwana-derived elements. The main tract of the Iapetus Ocean (Figs. 4, 15) was closed along this line. Other than narrow, discontinuous belts of forearc mélange (e.g. Dunnage–Sops Head–Hurricane mélanges) and slivers of tectonized oceanic basalt and gabbro (Fig. 6), the fault is locally inconspicuous at surface and was reactivated during the Silurian (Zagorevski et al. 2007b, 2009). Nevertheless, the fault trace is well defined in Newfoundland on the basis of a large multidisciplinary data set (Williams et al. 1988; van Staal et al. 1998, 2009; Zagorevski et al. 2008) and was seismically imaged as a west-dipping thrust fault (Fig. 9) that continues at least to the base of the lower crust (van der Velden et al. 2004). The Red Indian line is not exposed in northern New Brunswick and adjacent Quebec, but its approximate trace has been outlined mainly using geochemistry (Dupuis et al. 2009). The Red Indian line resurfaces again in Maine, where it is marked by the Hurricane mélange (Figs. 2, 6), which is a correlative of the Dunnage mélange in Newfoundland (Reusch and van Staal 2012).

The Dog Bay–Bamford Brook–Codyville–Liberty line (Ludman et al. 1993; Williams et al. 1993; van Staal et al. 2008, 2009; Reusch and van Staal 2012) is the Salinic, mid-Silurian (430–423 Ma) suture along which the Gander margin was accreted to composite Laurentia as a result of the closure of the wide Tetagouche–Exploits back-arc basin (Figs. 2, 7, 8, 16a). The fault zone is marked locally by mélange and fault rocks such as phyllonite and cataclasite along which rocks with different Silurian and older magmatic, sedimentary, structural, and metamorphic histories were juxtaposed (Ludman et al. 1993; van Staal et al. 1998, 2008, 2009; Pollock et al. 2007; Reusch and van Staal 2012). The suture zone was imaged at depth in central Newfoundland as a major west-dipping fault zone, characterized by strong reflections that continue beyond the

Moho into the mantle to a depth of approximately 17 seconds (van der Velden et al. 2004). Seismic profiling in southeastern Maine imaged this fault zone as a crust-penetrating, shallowly to moderately west-dipping structure (Ludman et al. 1990; Ludman 1991).

The boundary between Ganderia and Avalonia is represented by the Dover and Hermitage Bay faults in Newfoundland, the MacIntosh Brook fault in Cape Breton Island, and the Caledonia–Clover Hill fault in New Brunswick (Figs. 2, 7). Geophysical data, based both on seismic (e.g. Keen et al. 1986; Marillier et al. 1989) and potential field data (e.g. Loncarevic et al. 1989; King and Barr 2004) suggest that these faults are generally steep and appear to have accommodated a component of syn- to post-accretion strike-slip deformation (e.g. Holdsworth 1994; Park et al. 1994). Metamorphic tectonites and/or other structures associated with accretion/collision of Avalonia appear to be rarely preserved on surface. They occur only locally as slivers in the fault zones (e.g. White et al. 2006). For this reason, van Staal et al. (2009) inferred that the docking of Avalonia was accompanied by major subduction erosion of its forearc.

In northern mainland Nova Scotia, the boundary between Meguma and composite Laurentia (Avalonia) is the curviplanar Cobequid–Chedabucto fault zone (Figs. 2, 7). The fault zone has approximately an easterly strike with a steep southerly dip and has accommodated Middle to Late Devonian dextral transpression (Mawer and White 1987). The fault continues offshore to the east along the Collector magnetic anomaly (Fig. 7), where it achieves a more moderate southerly dip and is accompanied by development of the Mesozoic Orpheus half-graben in its hanging wall (Marillier et al. 1989; Keen and Potter 1995a). Another important Mesozoic half-graben has developed in its hanging wall in the Bay of Fundy (Wade et al. 1996). Further to the south, the boundary between Meguma and composite Laurentia (Avalonia) approximately follows the Nauset anomaly and dips moderately to the west (Hutchinson et al. 1988; Keen et al. 1991a). The Devonian terrane boundary between Meguma and composite Laurentia (Avalonia) was thus reactivated, at least in part, as a normal fault during Mesozoic rifting.

The Alleghanian suture between Meguma and Gondwana (Africa) may exist under the continental margin east of Nova Scotia, but if so, it has not been recognized on seismic profiles (e.g. Fig. 12) across the area (Keen et al. 1991b; Keen and Potter 1995a). In general, the Mesozoic rifting-related structures (see sections 3.1 and 8.0) probably strongly masked the Alleghanian suture zone (e.g. Tucholke et al. 2007; Redfern et al. 2011), which may have served as the locus for separation when Gondwana finally rifted away again (Fig.1) during the Jurassic–Cretaceous opening of the Atlantic Ocean (see section 8.0).

8.0 NATURE OF THE APPALACHIAN ATLANTIC MARGIN IN CANADA

The protracted expansion of eastern Laurentia into composite Laurentia during the Paleozoic culminated in the formation of the supercontinent Pangea during the Permian, but the break-up of Pangaea began soon after its formation, which led to development of narrow Permian–Triassic rift basins; their location and orientation were largely controlled by the inherited Appalachian–Caledonian structures (Redfern et al. 2011). Rifting in Atlantic Canada started in the Triassic by reactivating older compressional structures associated with accretion and wedging of Meguma (see section 3.1) and the Alleghanian collision (Fig. 1), which led to formation of half grabens such as the Bay of Fundy basin (Wade et al. 1996). Rifting in the Bay of Fundy basin was accompanied by eruption of plateau basalts (North Mountain Basalt; Kontak 2008). Rifting culminated in the opening of the Atlantic Ocean, which was a diachronous event, becoming progressively younger to the north. Opening took place at ca. 175 Ma along the Scotian margin at the latitude of Nova Scotia (Klitgord and Schouten 1986), but significantly later (ca. 112 Ma) opposite the Grand Banks of Newfoundland (Tucholke et al. 2007 and references therein), north of the Newfoundland fracture zone (Fig. 1). These two stages in opening of the Atlantic Ocean separated North America from Africa and the Galician margin of Iberia, respectively. The crustal structure of the continental margins of Nova Scotia and Newfoundland has been studied extensively using seismic data (e.g. Keen et al. 1991b; Keen and Potter, 1995 a,b). The southernmost part of the Nova Scotia margin is an extension of the volcanic margin that is present along most of the eastern seaboard of the United States (Keen and Potter 1995b). However, the Atlantic margin of the remainder of Nova Scotia and Newfoundland is non-volvanic, characterized by prolonged rifting and basin formation, slow-spreading, lack of syn-rift melt generation, and exhumation of continental serpentinized mantle (Fig. 12) beneath highly thinned continental crust and/or onto the ocean floor (Funck et al. 2004; Tucholke et al. 2007). Along Flemish Cap, part of the extended continental shelf about 500 km east of Newfoundland,

Hopper et al. (2007) showed that the continental crust at its leading edge was abruptly thinned from 30 km to 3 km over a distance of 80 km, suggesting that it occupied the upper plate, whereas the conjugate Galician margin was the lower plate. The exhumation of serpentinized mantle requires embrittlement of the entire crust to allow sufficient convection of seawater down through the crust into the underlying mantle to transform it into serpentinite (Reston 2007). The first phase of crustal separation represented by mantle exhumation thus took place before seafloor spreading, which in turn led to anomalously thin (3–5 km thick) oceanic crust, typical of slow-spreading ridges with a low supply of magma.

9.0 CONCLUSIONS

The Canadian Appalachians represent a segment of a long-lived Paleozoic accretionary orogen, growth of which culminated in the Late Paleozoic with the Laurentia–Gondwana collision. Our understanding of the development of this orogen has grown exponentially in recent years as a result of geophysical and supporting geoscience data acquired as part of LITHOPROBE East (e.g. Quinlan 1998) and subsequent studies. The Early Paleozoic Laurentian margin progressively expanded oceanward over time by the punctuated accretion of intra-oceanic suprasubduction zone terranes and four microcontinents (Dashwoods, Ganderia, Avalonia, and Meguma). Normal oceanic lithosphere was rarely, if ever, accreted and mostly lost as a result of subduction. Accretion of the microcontinents was the main cause of the classic Ordovician Taconic (Dashwoods), Late Ordovician to Late Silurian Salinic (Ganderia), Late Silurian to Early Devonian Acadian (Avalonia), and late Early Devonian to Early Carboniferous Neoacadian (Meguma) orogenies. Orogenesis was terminated by the Late Carboniferous to Permian Alleghanian collision of Laurentia with Gondwana to form Pangaea. The preserved part of the Canadian Appalachians largely escaped the penetrative effects of this collision; hence evidence of the earlier terrane accretion-related orogenic events is better preserved in Canada than elsewhere in the Appalachians. Ultimately, the Appalachian orogen was dismembered by the Mesozoic break-up of Pangaea, which led to the formation of the present-day Atlantic Ocean. However, this mainly affected the Alleghanian collision zone. Most of the pre-Alleghanian tectonic architecture of its hinterland remained intact to constitute the easternmost part of North America.

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