Research Article

Formation of Anorthositic Rocks within the Blair River Inlier of Northern Cape Breton Island, Nova Scotia (Canada)

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Rocks from the Blair River inlier of Northern Cape Breton Island (Nova Scotia, Canada) have been correlated with either the Grenville basement of eastern Laurentia or the accreted Avalon terrane. Additional zircon U-Pb dates of spatially associated anorthositic dykes (425.1 ± 2.2 Ma) and a metagabbro (423.8 ± 2.5 Ma) from the Fox Back Ridge intrusion of the Blair River inlier reveal Late Silurian emplacement ages. Their contemporaneity suggests that they may be members of a larger intrusive complex. The anorthositic rocks have high Eu/Eu* values (>2.5), and bulk compositions are similar to the mineral compositions of labradorite (An85–70) and andesine (An80–90). The metagabbro is compositionally similar to alkali basalt and does not seem to have been affected by crustal contamination (Nb/U > 24; Th/NbPM ≤ 1.1) although it was metamorphosed. The high Tb/YbN (1.8-1.9) ratios suggest that the parental magma of the metagabbro was derived from a garnet-bearing peridotite. Fractional crystallization and mass balance calculations indicate that the anorthositic rocks can be derived by mineral accumulation from a mafic parental magma similar to the metagabbro of this study. The Late Silurian ages suggest that the rocks were emplaced into the Avalon terrane after the closure of the Iapetus Ocean but before Early Devonian (415–410 Ma) sinistral transpression.

1. Introduction

The eastern margin of North America records a complex middle Proterozoic to middle Paleozoic geological history. The most complete record is preserved in a belt of basement inliers with overlying cover sedimentary sequences that extend discontinuously from Alabama to Newfoundland along the western flank of the Appalachian orogen [1–3]. The inliers are composed of middle to late Proterozoic rocks that were either formed or deformed during the Grenvillian orogeny and dated at 1200–980 Ma [4]. These inliers include rocks of easternmost Laurentia that are elsewhere typically included in the Appalachian orogen [5]. They are distinct from the outboard Appalachian terranes (i.e., Avalonia, Meguma, Carolinia) composed of Neoproterozoic and Paleo-
At present, the basement rocks of the Blair River inlier in Northern Cape Breton Island have different interpretations. Brown [14], Barr and Raeside [13], and Miller and Barr [15] correlate the inlier with Grenvillian (Laurentian) basement of the Humber Zone, whereas Neale and Kennedy [9] correlated them with the Gander terrane (peri-Gondwana) and Keppie [10] and Keppie et al. [16] interpreted them as the basement rocks of the Avalon terrane (peri-Gondwana), thereby providing a source for the common 1-1.3 Ga detrital zircons in Avalonia (Figure 1). A critical factor in the correlation with Grenvillian basement is the presence of anorthosite bodies, which are restricted to the area of the Blair River inlier. In order to provide a better database for the comparison with anorthosite of the Grenville Province, we have investigated these Cape Breton rocks.

The Blair River inlier is located at the northernmost tip of Cape Breton Island, Nova Scotia, and is comprised of Mesoproterozoic to Carboniferous rocks. It is bounded by the Gulf of St. Lawrence to the west and north and separated from the adjacent Aspy terrane by the Wilkie Brook fault to the east and the Red River fault to the south. The Mesoproterozoic (~1.2 Ga to ~1.0 Ga) rocks consist of composite orthogneiss complexes (i.e., Sailor Brook gneiss, Otter Brook gneiss, and Polletts Cove River gneiss) that are intruded by comparatively less deformed plutonic rocks. The plutonic rocks include gabbro, syenite, anorthosite, granite, and granodiorite [15, 17]. Single-grain analyses of abraded zircon from the anorthosite of the Blair River inlier (Red River anorthosite) indicate that the minimum crystallization U-Pb age was 1095 ± 3 Ma although there is evidence suggesting that the zircons experienced multiple stages of lead loss [15]. The uniqueness of the Blair River inlier relative to the peri-Gondwanan terranes (i.e., Avalon, Gander, and Meguma) in the northern Appalachians suggests that it may represent a fragment of Laurentian basement and that the Red River anorthosite is interpreted to be correlative with Proterozoic massif-type anorthosite found in the Grenville Province [15, 17, 18]. However, recent zircon U-Pb geochronology on the Red River anorthosite indicates that it was emplaced during the Late Silurian (421 ± 3 Ma) and contains Meso- to Neoproterozoic (865 ± 18 Ma to 1044 ± 20 Ma)-inherited zircons [16]. An evaluation of the whole-rock Pb isotopes of the Blair River inlier rocks, including the Red River anorthosite, indicates that it was emplaced during the Late Silurian (421 ± 3 Ma) and contains Mesozoic to Neoproterozoic (865 ± 18 Ma to 1044 ± 20 Ma)-inherited zircons [16]. The absence of ophiolites between the Blair River inlier and the neighbouring Aspy terrane, which are present in southern Newfoundland, argues against a Laurentia-Avalon suture zone [16, 19].

In this paper, we present new zircon U-Pb geochronology and whole-rock geochemistry of metagabbro from the Fox...
2. Geological Background

The Blair River inlier includes several anorthosite bodies, the largest of which is the Red River anorthosite suite (Figure 2). This intrusion is about 11 km long and up to 1 km wide. It occurs within a NNE-trending belt of quartz-feldspathic gneiss, which is about 30 km long and 2-8 km wide [15]. The belt is bounded by mylonite zones, and thus, any direct correlation with other units in the area is tenuous. The Red River anorthosite is representative of the Cape Breton anorthosites [20]. It is composed of anorthosites passing in places into interlayered anorthositic gabbro and pyroxene-rich rocks [15, 21]. The anorthosite grades from massive and equigranular to foliated. Its texture ranges from protoclastic and porphyroclastic to recrystallized granoblastic. The dominant mineral is plagioclase (An$_{35-50}$) with grain size ranging from 1 to 5 mm [22]. The anorthositic rocks typically contain minor amphibole with rare pyroxene relics in the core. Both the anorthosite and the surrounding gneisses of the belt have experienced polyphase deformation, which has transposed the contacts into parallel features. The rocks within and bounding the Blair River inlier were metamorphosed to amphibolite facies along shear zones. Deformation was accompanied by partial melting and mylonitization in some cases. In between these shear zones, anorthosites, monzodiorite-gabbro, and granitic and syenitic plugs are undeformed.

A suite of eight samples was collected from the Blair River inlier. Two samples from the main body of the Red River anorthosite, three samples from an anorthositic dyke that intrudes the Fox Back Ridge diorite/gabbro intrusion, and three host metagabbroic rocks of the Fox Back Ridge intrusion [15] were collected. The metagabbro unit was originally mapped together as orthogneiss and paragneiss with minor amphibolite- and kyanite-bearing horizons by Smith and MacDonald [23] and assigned to the Helikian (850 Ma to 1600 Ma) Poletts Cove River Group by Barr et al. [24]. However, Miller [22] reexamined the area and identified the region west of the Red River anorthosite as a distinct intrusive unit and named it the Fox Back Ridge intrusion of probable Paleozoic age [17]. The dated samples were collected along the Cabot Trail highway immediately west of
the Red River anorthosite (Figure 3). Sample CBCH-1 is an amphibole-rich metagabbro, and CBGR-1 is an anorthositic dyke that cross-cuts the Fox Back Ridge intrusion.

3. Petrography

3.1. Anorthosite. Sample CBGR-1 (46°48′31.57″N, 60°41′44.53″W) was collected from an undeformed anorthositic dyke that intruded the host Fox Back Ridge intrusion. It is comprised of 90-95 vol.% plagioclase with minor (~5 vol.%) pyroxene and accessory amounts (<~1 vol.%) of Fe-Ti oxide minerals and apatite. The plagioclase crystals tend to be blocky rather than tabular and display polysynthetic twinning under cross-polars. Generally, the crystals have similar grain size but some, based on the twinning angle, appear to be bent (Figures 4(a) and 4(b)). The pyroxene crystals are small, anhedral, and interstitial to the plagioclase. Although both orthopyroxene and clinopyroxene are identified in the anorthosite, most pyroxene crystals are altered to uralite suggesting that it was likely clinopyroxene [22]. Secondary muscovite is replacing some plagioclases, and a narrow carbonate vein cross-cuts a portion of the thin section.

3.2. Metagabbro. The metagabbro (CBCH-1; 46°48′31.57″N, 60°41′44.53″W) from the Fox Back Ridge intrusion is coarse to medium grained and is comprised primarily of replacement minerals of plagioclase (45-50 vol.%) and clinopyroxene (40-45 vol.%), with abundant apatite (~5 vol.%), and minor amounts (<5 vol.%). Secondary oxide minerals (Figures 4(c) and 4(d)). The plagioclase is mostly altered to saussurite, but there is a minority of unaltered crystals that retain the euhedral to subhedral texture and polysynthetic twinning. Similar to plagioclase, there is a minority of small, subrounded, unaltered clinopyroxene crystals but most crystals are altered to uralite. The original rock texture is difficult to discern as there is a weak foliation of the uralite and saussurite, but it appears that the rock may have been ophitic. Apatite is relatively abundant and occurs as either small euhedral crystals or larger tabular crystals. The oxide minerals, including hematite, appear to be exclusively secondary as they are commonly found within or spatially associated with uralite. Short and narrow quartz veins are present but comprised <2 vol.%

4. Methods

4.1. Zircon U-Pb Geochronology. Zircon separation from samples CBCH-1 and CBGR-1 was processed using magnetic separation and heavy-liquid techniques at the Yu-Neng Rock and Mineral Separation Company (Lanfang, Hebei, China). Cathodoluminescence (CL) images were taken at the Institute of Earth Sciences, Academia Sinica, Taipei, for selecting suitable positions for U-Pb analyses and to examine individual crystal internal structures. Zircon U-Pb isotopic analyses were performed by laser ablation-inductively coupled plasma mass spectrometry (LA-ICP-MS) at the Department of Geosciences, National Taiwan University, Taipei, using an Agilent 7500S Q-ICP-MS and a Photon Machines Analyte G2 193 nm laser ablation system following the method of Chiu et al. [25]. A spot size of 35 μm with a laser repetition rate of 5 Hz was used, and the laser energy density was 3.83 to 5.33 J/cm². Calibration was performed by using the zircon standard GJ-1 (608.5 ± 0.4 Ma; [26]) and 91500
(1065 Ma; [27]). The Plešovec (337.1 ± 0.4 Ma; [28]) zircon standard was also used for data quality control. Measured U-Th-Pb isotope ratios were calculated using the GLITTER 4.4.4 software [29], and the relative standard deviations of reference values for GJ-1 were set at 2%. The common lead was corrected using the common lead correction function of Andersen [30, 31], and the weighted mean U-Pb ages and concordia plots were created using Isoplot v. 4.1 [32]. Only results within 10% discordance (\(1 - \left(\frac{^{206}\text{Pb}^{238}\text{U}}{^{207}\text{Pb}^{235}\text{U}}\right)_{\text{uncorrected age}}\) \times 100) were used for the data interpretations.

4.2. Major and Trace Elemental Geochemistry. Whole-rock major and trace elements were analyzed at the Activation Laboratories Ltd. in Ancaster, Ontario, Canada, using lithium metaborate-tetraborate fusion. Major elements were determined by an inductively coupled plasma-optical emission spectrometer, whereas trace element analyses were performed by an inductively coupled plasma-mass spectrometer. Replicate analyses of the reference standard rocks indicate that 1 sigma error is between 2 and 8% of the values cited. The detailed information on the major and trace element analyses is available at the Activation Laboratories web site (https://www.actlabs.com).

5. Results

5.1. Geochronology. The CL images show that some crystals have bright cores with darker rims, whereas others are entirely bright or entirely dark. The crystals have a variety of habits that include euhedral prismatic, subhedral round to subround, and anhedral fragmented (Fig. DR1). Individual zircon grains of the anorthosite sample (CBGR-1) range in size from ~25 μm to ~100 μm in length/width and typically show oscillatory zonation (Fig. 4). Ninety-five of 97 analyses of zircons are within 10% discordance and yielded a total range of 206Pb/238U ages from 401 ± 9 Ma (1σ) to 794 ± 18 Ma (1σ). Four populations are identified based on the clustering of the data (Table DR1). The oldest three zircons yielded ages of 501 ± 11 Ma (1σ), 657 ± 15 Ma (1σ), and 794 ± 18 Ma (1σ). The second oldest group consists of 13 middle to late Ordovician zircons. This group appears to cluster into two subgroups: (1) 463 ± 10 Ma (1σ) to 449 ± 10 Ma (1σ) and (2) 442 ± 11 Ma (1σ) to 443 ± 10 Ma (1σ). The largest group (76) of zircons range in age from 415 ± 9 Ma (1σ) to 437 ± 10 Ma (1σ), whereas the youngest group is comprised of two Early Devonian zircons (401 ± 9 Ma (1σ) and 406 ± 9 Ma (1σ)). The oldest zircons and the Middle to Late Ordovician zircons are interpreted to represent inheritance, whereas the youngest zircons may be related to the effects of regional deformation during the Early Devonian. Consequently, we interpret the crystallization age to be based on the Late Silurian zircons which yielded a weighted-mean 206Pb/238U age of 425 ± 2.2 Ma (2σ, Figure 5(a)).

Individual zircon grains of the metagabbro (CBCH-1) range in size from ~50 μm to ~225 μm (Fig. DR2), typically show oscillatory zonation, and have fragmented, anhedral to subhedral textures. The zircons from this sample tend to have uniform CL brightness compared to those in the anorthosite (CBGR-1). Seventy-seven of 82 analyses are within 10% discordance and yielded a total range of 206Pb/238U ages from 404 ± 10 Ma (1σ) to 468 ± 12 Ma (1σ). Three distinct zircon age populations are identified based on age gaps (Table DR2). The largest (65 zircons) group ranges from
The Late Silurian (5.5-4.2 Ma) anorthosites of the Red River are peraluminous (Al/Ca+Na+K (mol.) SI > 1) and have relatively high volatile content as loss on ignition (LOI) is between 1.4 and 2.3 wt%. The quantities of transition elements for the anorthosites are low (<30 ppm) across all elements with Zn having the highest concentrations (50 to 100 ppm). The primitive mantle-normalized incompatible element patterns show enrichment of the large ion lithophile elements (Cs, Rb, Ba, and Sr), and Th and U relative to the rare earth elements and the remaining high-field-strength elements (Figure 7(a)). The chondrite-normalized rare earth element patterns show enrichment of the light rare earth elements with relatively flat heavy rare earth elements. The prominent Eu-anomaly (Eu/Eu* > 2.5) is typical of anorthositic rocks and is a direct consequence of their feldspar-rich nature (Figure 7(b)).

The gabbroic rocks of this study (BR-2-1, BR-2-4, and CBCH-1) are compositionally similar to alkali basalt as they have low SiO2 (~47 wt%), Al2O3 ~24 wt%, Na2O ~5 wt%, and K2O ~3 wt%) but K2O, TiO2, FeO, MgO, and TiO2 are higher and reflect the presence of oxide and mafic silicate minerals in the mode (Figure 6(b)). In contrast, the second group (BR-3-1, BR-3-2, CBGR-1) from the anorthositic dykes are somewhat similar to andesine (SiO2 ~56 wt%, Al2O3 ~24 wt%, CaO ~5 wt%, Na2O ~6.2 wt%, and K2O ~3 wt%) but K2O, TiO2, FeO, MgO, and TiO2 are higher and reflect the presence of oxide minerals, mafic silicates (biotite, K-feldspar, and mica (Figure 6(b)). Including data from Miller et al. (2000), the anorthositic rocks have a relatively restricted SiO2 (wt%) content but the Fe/(Fe + Mg) ratio is variable (~0.3 to ~0.8) (Figure 6(c)). Both types of anorthosite are peraluminous (Al/Ca+Na+K (mol.) SI > 1) and have relatively high volatile content as loss on ignition (LOI) is between 1.4 and 2.3 wt%. The quantities of transition elements for the anorthosites are low (<30 ppm) across all elements with Zn having the highest concentrations (50 to 100 ppm). The primitive mantle-normalized incompatible element patterns show enrichment of the large ion lithophile elements (Cs, Rb, Ba, and Sr), and Th and U relative to the rare earth elements and the remaining high-field-strength elements (Figure 7(a)). The chondrite-normalized rare earth element patterns show enrichment of the light rare earth elements with relatively flat heavy rare earth elements. The prominent Eu-anomaly (Eu/Eu* > 2.5) is typical of anorthositic rocks and is a direct consequence of their feldspar-rich nature (Figure 7(b)).

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Table 1: Whole-rock geochemical data of the Blair River inlier rocks.

(a)

| Sample | BR-1-1 | BR-1-2 | BR-2-1 | BR-2-4 | CBCH-1 | BR-3-1 | BR-3-2 | CBGR-1 |
|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| Lat.   | 46°48'31.57"N | 46°41'44.53"W | An | An | MG | MG | MG | An | An | An |
| Lon.   | 60°41'18.17"W | 60°41'57.08"W | An | An | MG | MG | MG | An | An | An |
| Rock   | An | An | MG | MG | An | An | An | MG | MG | MG |
| SiO₂ (wt.%) | 53.79 | 53.66 | 46.61 | 41.26 | 46.45 | 56.41 | 56.03 | 56.15 |
| TiO₂   | 0.20 | 0.14 | 1.95 | 1.97 | 1.79 | 0.14 | 0.13 | 0.22 |
| Al₂O₃  | 26.22 | 26.33 | 17.19 | 18.26 | 18.05 | 22.62 | 24.33 | 23.57 |
| Fe₂O₃  | 2.30 | 2.16 | 11.56 | 11.75 | 11.14 | 2.85 | 1.84 | 1.80 |
| MnO    | 0.03 | 0.03 | 0.16 | 0.13 | 0.15 | 0.06 | 0.03 | 0.03 |
| MgO    | 1.63 | 0.93 | 4.85 | 4.75 | 4.72 | 1.23 | 0.56 | 0.60 |
| CaO    | 8.76 | 9.74 | 7.68 | 8.27 | 6.92 | 5.46 | 4.61 | 5.46 |
| Na₂O   | 5.07 | 4.75 | 3.04 | 3.84 | 3.51 | 6.22 | 6.23 | 6.32 |
| K₂O    | 0.53 | 0.68 | 1.82 | 2.07 | 1.62 | 2.13 | 3.21 | 3.12 |
| P₂O₅   | 0.03 | 0.05 | 1.58 | 1.60 | 1.28 | 0.07 | 0.32 | 0.10 |
| LOI    | 1.82 | 1.41 | 3.44 | 6.40 | 4.01 | 2.27 | 1.86 | 3.04 |
| Total  | 100.38 | 99.88 | 99.88 | 100.29 | 99.65 | 99.45 | 99.15 | 100.40 |
| Mg#    | 58.4 | 46.0 | 45.4 | 44.5 | 45.6 | 46.1 | 37.6 | 39.8 |
| Sc (ppm)| 3 | 2 | 24 | 22 | 27 | 3 | 1 | 18 |
| V      | 20 | 20 | 310 | 303 | 267 | 24 | 16 | 15 |
| Cr     | b.d. | b.d. | b.d. | b.d. | b.d. | 40 | b.d. | 18 |
| Co     | 7 | 5 | 28 | 23 | 29 | 4 | 2 | 4 |
| Ni     | b.d. | b.d. | b.d. | b.d. | 5 | b.d. | b.d. | 6 |
| Cu     | b.d. | b.d. | 90 | 100 | 115 | b.d. | b.d. | 73 |
| Zn     | 17 | 19 | 22 | 23 | 24 | 18 | 18 | 19 |
| Rb     | 14 | 14 | 31 | 38 | 33 | 42 | 64 | 72 |
| Sr     | 970 | 963 | 1162 | 659 | 1113 | 869 | 848 | 963 |
| Y      | 2 | 2.9 | 34.5 | 34.8 | 33.5 | 5.7 | 4.1 | 2.9 |
| Zr     | 9 | 9 | 197 | 215 | 241 | 17 | 11 | 14 |
| Nb     | b.d. | 0.2 | 47.7 | 49.2 | 68.8 | 2.5 | 1.7 | 2.3 |
| Cs     | 2.8 | 1.7 | 1.2 | 1.5 | 1.9 | 1.8 | 2.5 | 3.5 |
| Ba     | 205 | 201 | 1441 | 1176 | 1482 | 987 | 1217 | 1602 |
| La     | 2.5 | 3.2 | 96.1 | 91.7 | 99.9 | 7.7 | 7.9 | 6.6 |
| Ce     | 5.00 | 6.19 | 173 | 172 | 181 | 14.2 | 14.4 | 11.6 |
| Pr     | 0.61 | 0.73 | 18.3 | 18.5 | 20.1 | 1.64 | 1.53 | 1.4 |
| Nd     | 2.54 | 2.96 | 67.4 | 69.2 | 69.8 | 6.34 | 5.69 | 5.0 |
| Sm     | 0.48 | 0.66 | 11.1 | 11.4 | 11.2 | 1.25 | 1.01 | 0.89 |
| Eu     | 0.64 | 0.71 | 3.25 | 3.35 | 3.51 | 1.01 | 1.04 | 1.45 |
| Gd     | 0.51 | 0.50 | 8.56 | 8.83 | 10.5 | 1.12 | 0.91 | 0.83 |
| Tb     | 0.05 | 0.09 | 1.14 | 1.18 | 1.28 | 0.16 | 0.11 | 0.10 |
| Dy     | 0.35 | 0.52 | 6.29 | 6.42 | 6.58 | 0.98 | 0.68 | 0.51 |
| Ho     | 0.06 | 0.09 | 1.18 | 1.18 | 1.26 | 0.20 | 0.13 | 0.09 |
| Er     | 0.19 | 0.28 | 3.21 | 3.31 | 3.37 | 0.56 | 0.35 | 0.25 |
| Tm     | 0.03 | 0.04 | 0.42 | 0.45 | 0.48 | 0.08 | 0.05 | 0.03 |
| Yb     | 0.18 | 0.26 | 2.74 | 2.87 | 3.02 | 0.56 | 0.34 | 0.21 |
| Lu     | 0.02 | 0.05 | 0.44 | 0.46 | 0.47 | 0.09 | 0.06 | 0.03 |
| Hf     | 0.2 | 0.2 | 4.2 | 4.2 | 5.7 | 0.5 | 0.4 | 0.4 |

Lithosphere
Table 1: Continued.

| Sample    | BR-1-1 | BR-1-2 | BR-2-1 | BR-2-4 | CBCH-1 | BR-3-1 | BR-3-2 | CBGR-1 |
|-----------|--------|--------|--------|--------|--------|--------|--------|--------|
| Ta        | 0.02   | 0.03   | 3.21   | 3.21   | 3.17   | 0.29   | 0.38   | 0.22   |
| Th        | b.d.   | 0.24   | 5.43   | 6.85   | 8.33   | 1.33   | 3.07   | 1.53   |
| U         | 0.02   | 0.17   | 1.93   | 1.88   | 2.12   | 0.42   | 0.89   | 0.70   |
| Eu/Eu*    | 4.0    | 3.6    | 0.98   | 0.98   | 0.97   | 2.6    | 3.3    | 5.1    |
| (La/Yb)N  | 9.8    | 8.9    | 25.2   | 22.9   | 23.7   | 9.8    | 16.7   | 22.6   |

(b)

| Sample    | BIR-1 | BIR-1 | AGV-2 | AGV-2 | BCR-2 | BCR-2 | BHVO-2 | BHVO-2 |
|-----------|-------|-------|-------|-------|-------|-------|--------|--------|
| ROCK      | r.v.  | m.v.  | r.v.  | m.v.  | r.v.  | m.v.  | r.v.   | m.v.   |
| SiO2 (wt.%) | 47.96 | 48.07 | 59.30 | 60.41 | 54.10 | 54.46 |        |        |
| TiO2      | 0.96  | 0.96  | 1.05  | 1.05  | 2.26  | 2.27  |        |        |
| Al2O3     | 15.50 | 15.59 | 16.91 | 17.31 | 13.50 | 13.62 |        |        |
| Fe2O3t    | 11.30 | 11.16 | 6.69  | 6.83  | 13.80 | 13.80 |        |        |
| MnO       | 0.175 | 0.17  | 0.10  | 0.10  | 0.20  | 0.20  |        |        |
| MgO       | 9.70  | 9.65  | 1.79  | 1.79  | 3.59  | 3.63  |        |        |
| CaO       | 13.30 | 13.30 | 5.20  | 5.22  | 7.12  | 7.16  |        |        |
| Na2O      | 1.82  | 1.85  | 4.19  | 4.38  | 3.16  | 3.24  |        |        |
| K2O       | 0.03  | 0.03  | 2.88  | 2.93  | 1.79  | 1.82  |        |        |
| P2O5      | 0.02  | 0.02  | 0.48  | 0.48  | 0.35  | 0.36  |        |        |
| LOI       | -0.24 | 0.00  |       |       |       |       |        |        |
| Total Mg# |       |       |       |       |       |       |        |        |
| Sc (ppm)  | 13.1  | 9.8   | 33.5  | 37.9  | 31.8  | 32.6  |        |        |
| V         | 119   | 96.1  | 418   | 418   | 318   | 323   |        |        |
| Cr        | 16.2  | 12.1  | 15.9  | 15.9  | 287   | 275   |        |        |
| Co        | 15.5  | 13.4  | 37.3  | 37.4  | 44.9  | 43.6  |        |        |
| Ni        | 18.9  | 183   | 120   | 105   |       |       |        |        |
| Cu        | 51.5  | 49.5  | 19.7  | 28.4  | 129   | 127   |        |        |
| Zn        | 86.7  | 73.1  | 130   | 143   | 134   | 110   |        |        |
| Ga        | 20.4  | 18.5  | 22.1  | 23.1  | 21.4  | 22.4  |        |        |
| Rb        | 67.8  | 61.3  | 46.0  | 49.8  | 9.3   | 10.4  |        |        |
| Sr        | 660   | 560   | 337   | 339   | 394   | 390   |        |        |
| Y         | 19.1  | 17.3  | 36.1  | 36.5  | 25.9  | 26.5  |        |        |
| Zr        | 232   | 202   | 187   | 187   | 171   | 170   |        |        |
| Nb        | 14.1  | 12.1  | 12.4  | 12.6  | 18.1  | 18.0  |        |        |
| Cs        | 1.17  | 1.2   | 1.16  | 1.4   | 0.1   | 0.1   |        |        |
| Ba        | 1134  | 1091  | 684   | 752   | 131   | 142   |        |        |
| La        | 38.2  | 34.9  | 25.1  | 26.7  | 15.2  | 16.1  |        |        |
| Ce        | 69.4  | 62.8  | 53.1  | 55.6  | 37.5  | 39.2  |        |        |
| Pr        | 8.17  | 7.3   | 6.83  | 7.1   | 5.3   | 5.5   |        |        |
| Nd        | 30.5  | 26.1  | 28.3  | 28.4  | 24.3  | 24.1  |        |        |
| Sm        | 5.51  | 4.7   | 6.55  | 6.5   | 6.0   | 6.0   |        |        |
| Eu        | 1.55  | 1.5   | 1.99  | 2.1   | 2.0   | 2.0   |        |        |
| Gd        | 4.68  | 4.4   | 6.81  | 6.9   | 6.2   | 6.2   |        |        |
| Tb        | 0.65  | 0.58  | 1.08  | 1.09  | 0.94  | 0.97  |        |        |
| Dy        | 3.55  | 3.0   | 6.42  | 6.4   | 5.3   | 5.3   |        |        |
| Ho        | 0.68  | 0.59  | 1.31  | 1.32  | 1.00  | 1.00  |        |        |
| Er        | 1.83  | 1.59  | 3.67  | 3.67  | 2.51  | 2.53  |        |        |
or magmatic system and either directly or indirectly related. There are significant local and regional lithotectonic implications given that the anorthositic rocks are Late Silurian rather than Mesoproterozoic as previously interpreted.

Locally, the zircon U-Pb ages are similar to titanite, rutile, and hornblende ages reported for the dioritic rocks of the Fox Back Ridge intrusion (423 ± 3 Ma, U-Pb titanite; 417 ± 6 Ma, 40Ar/39Ar hornblende) and Red Ravine syenite (425 ± 2 Ma, U-Pb titanite) suggesting that they may be members of a larger igneous complex (Table 2). Moreover, in western Cape Breton, there are a number of Silurian plutonic rocks (MacLean Brook = 442.4 ± 2.0 Ma; Lavis Brook = 440.0 ± 1.6 Ma; Gillanders Mountain = 436.7 ± 2.6 Ma, 428.6 ± 1.9 Ma; Sammys Barren = 435 ± 7/−3 Ma; Gills Brook = 436.4 ± 1.5 Ma; Lake Ainslie = 428.5 ± 1.7 Ma; Grand Falaise = 427.9 ± 2.1 Ma; Leonard MacLeod Brook = 420.9 ± 2.8 Ma; and Glasgow Brook pluton = 416.0 ± 1.9 Ma) that have 206Pb/238U ages from 424.2 ± 2.0 Ma to 416.0 ± 1.9 Ma [17, 19, 33, 35]. In addition, middle to late Silurian (435 ± 7/−3 Ma to 417.6 ± 6 Ma) postmetamorphic cooling ages are reported for a variety of rocks in the Blair River inlier [17].

Regionally, middle to late Silurian magmatic rocks are common in the Gander terrane of Newfoundland and New Brunswick (432 ± 2 Ma to 419.5 ± 0.6 Ma), and Silurian (443 ± 4 Ma to 424 ± 2/−1 Ma) magmatic rocks (Notre Dame arc, Attean pluton, Quimby Formation) were also emplaced into the Laurentian margin crust [36–42]. Silurian magmatism is attributed to the Salinic orogeny (445 Ma to 422 Ma) which corresponds to the subduction and accretion of the leading (western) edge of Gander terrane to Laurentia. It is thought that the Salinic orogeny produced older subduction-related plutons and younger postcollisional (back-arc) plutons via slab detachment after the cessation of subduction [40–43]. The Salinic arc magmatic rocks are located in northern and central New Brunswick and western and central Newfoundland proximal to and within the Laurentian margin, but there are a number of postcollisional plutons, batholiths, and intrusions that were emplaced closer to and along the boundary region of the Gander terrane and the Avalon terrane [38, 42, 44–46].

### Table 1: Continued.

| Sample     | BIR-1 | BIR-1 | AGV-2 | AGV-2 | BCR-2 | BCR-2 | BHVO-2 | BHVO-2 |
|------------|-------|-------|-------|-------|-------|-------|--------|--------|
| Tm         | 0.26  | 0.23  | 0.53  | 0.54  | 0.54  | 0.33  | 0.34   |
| Yb         | 1.65  | 1.42  | 3.39  | 3.41  | 1.99  | 2.02  |
| Lu         | 0.25  | 0.22  | 0.51  | 0.52  | 0.52  | 0.28  | 0.28   |
| Hf         | 5.14  | 4.66  | 4.97  | 5.10  | 4.47  | 4.55  |
| Ta         | 0.87  | 0.75  | 0.79  | 0.81  | 1.15  | 1.18  |
| Th         | 6.17  | 5.23  | 5.83  | 5.83  | 1.22  | 1.18  |
| U          | 1.89  | 1.76  | 1.68  | 1.81  | 0.41  | 0.44  |

Eu/Eu* (La/Yb)_N

*From Keppie et al. [16]. Lat.: latitude; Lon.: longitude; An: anorthosite; MG: meta-gabbro; t: total Fe expressed as Fe_2O_3. LOI: loss on ignition. Eu/Eu* = [2 * Eu_{n+1}/(Sm_{n+1} + Gd_{n+1})] normalized to C1 chondrite [96]. Mg# = (Mg^{2+}/(Mg^{2+} + Fe^{3+})) * 100. N: normalized to the chondritic value [96]. b.d.: below detection; m.v.: measured value; r.v.: recommended value. The measured standard reference material trace element values are based on an average of three samples for AGV-2 and two samples each for BCR-2 and BHVO-2.

Silurian 206Pb/238U ages of intrusive rocks (mafic to felsic) are reported from the Avalon terrane of Newfoundland (Pass Island = 423 ± 4 Ma), Gander terrane of southern New Brunswick (Welsford = 442 ± 1.0 Ma; Utopia = 428 ± 1.0 Ma, 425.5 ± 2.1 Ma, 420.4 ± 2.4 Ma, and 420 ± 3.5 Ma; and Bocabec = 422.1 ± 1.3 Ma; Jake Lee Mountain = 418 ± 2 Ma), and coastal Maine (Mooselorn = 421.1 ± 0.8 Ma; Penobscot = 419 ± 2.0 Ma; Spruce Head = 421 ± 1.0 Ma; Cadillac Mountain = 424 ± 2.0 Ma and 419 ± 2.0 Ma; Sedgewick = 419.5 ± 1.0 Ma; and Cranberry Island = 424 ± 1.0 Ma). Early Ludlovian (427.4 Ma to 425.6 Ma) K-bentonite beds were identified at Arisaig (Avalon terrane, Nova Scotia) and are thought to be regionally correlative with ash beds from Gotland (Sweden) and those of the British Isles [47]. Similar ages rocks are found within the Meguma terrane of Western Nova Scotia (Brenton pluton = 439 ± 4/3 Ma; White Rock formation = 438 + 3/2 Ma; and Mavillette gabbro = 426 ± 2 Ma) as well [44–46, 48–58].

Plutonism in the region continued from the Late Silurian to Early Devonian and may be related to either the early stages of the Acadian orogeny (420 Ma to 410 Ma) or postcollisional crustal relaxation. The plutons are found across the trailing (eastern) edge regions of the Gander terrane and the boundary region between the Gander and Avalon terranes [38, 45, 46, 48, 53, 59–63]. Subsequent to the Avalon-Laurentia collision (435 Ma to 430 Ma), the Meguma terrane was thrust over Avalonia (~400 Ma) and both terranes were intruded by middle to late Devonian mafic and silicic plutonic rocks [33, 45, 60, 64–67]. Consequently, the precise terrane correlation of the Blair River inlier rocks is difficult to discern based on just their ages as they are more likely to be related to either the Avalon terrane or the Gander terrane than Laurentia. On the basis of whole-rock Nd isotope model ages and Pb isotopes of rocks from the Blair River inlier, including ~580 Ma mafic dykes, Keppie et al. [16] suggests that they have more in common with Avalonian crust, specifically Oaxacian crust (a proxy for Amazonia), than Laurentia [9, 10, 15, 16, 33, 46, 48, 68]. Nevertheless, the new age dates indicate that they are contemporaneous with the period of postcollisional magmatism that developed along the boundary between the Gander terrane and Avalon terrane.
An examination of the inherited zircons from the anorthositic rocks may provide additional support for the specific terrane correlation as they indicate the age of the country rock through which the parental magma interacted with during emplacement. The majority of inherited zircons have ages of ~440 Ma and ~465 Ma with individual ages of ~500 Ma and ~660 Ma and a slightly discordant age of ~790 Ma (Tables DR1 and DR2). Cambrian-Ordovician ages are common in Avalonian sedimentary and metasedimentary rocks but are also consistent with plutonic rocks in the Avalon and Gander terranes [34, 67, 69, 70]. The oldest concordant inherited zircon from the anorthosite dyke has a $^{206}\text{Pb}/^{238}\text{U}$ age of 657 ± 15 Ma (1σ) which is more indicative of the Avalon terrane than the Gander terrane [59]. Zircons
with ~650 Ma ages are identified in the Arisaig Group of the Antigonish Highlands in the Avalon terrane, metasedimentary granulites from the structural basement of Meguma, sedimentary rocks from eastern Newfoundland (Avalonia), and metavolcanic schists from Creignish Hills (Avalonia) of southwestern Cape Breton [67, 69, 71, 72]. Sedimentary rocks of southwest New Brunswick and coastal Maine have a few detrital zircons with ages of ~650 Ma (12 zircons) but four of the six units are south of the Turtle Head Fault [73]. The Turtle Head Fault was considered to be the boundary between Gander and Avalon terrane although this interpretation seems to have fallen out of favour [6, 57]. Presently, the zircon inheritance suggests that the rocks of the Blair River inlier are more likely correlative to the Avalon terrane.

6.2. Postcollisional Origin of the Fox Back Ridge Intrusion? The metagabbro is chemically similar to within-plate alkali basalt with LREE-enriched chondrite-normalized patterns but dissimilar to gabbroic rocks of the Red River anorthosite suite (Figures 6(d), 7(b), and 8(a)–8(c)). It appears that the rocks were either unaffected or minimally affected by crustal contamination as they do not have negative primitive mantle-normalized Nb-Ta anomalies and their Nb/U (≥ 24) and Th/NbPM (≥ 1.1) ratios do not trend toward average crustal values (Figure 8(d)). Furthermore, it does not appear that fractional crystallization of plagioclase or Fe-Ti oxide minerals played a role in their formation as the Eu/Eu∗ values are ∼1 and they do not have negative anomalies of Sr, Ba, or Ti (Figure 7(a)). This suggests, if early fractional crystallization occurred and produced the metagabbro, that only mafic silicate minerals (olivine, pyroxene) fractionated. Therefore, it is likely that the chondrite-normalized rare earth element patterns are primarily the result of partial melting. The high Tb/YbN (1.8–1.9) and Sm/YbPM (4.2–4.5) ratios of the metagabbro indicate that garnet was likely a residual phase in the mantle source suggesting that the minimum depth of melting was equivalent to the garnet-spinel transition (70 ± 20 km) zone [74]. Farther to the west, two diorite rocks from the Fox Back Ridge intrusion reported by Miller [22] are interpreted to be within-plate basalt but also indicate that the intrusion likely differentiated from mafic to intermediate.

The chemistry of the metagabbro is typical of mafic rocks from within-plate (Figure 8(b)) settings but the regional tectonic setting just prior (>430 Ma) to emplacement was likely one of compression (i.e., Salinic orogeny), and thus the gabros should be similar to volcanic-arc basalt or more likely demonstrate subduction-related geochemical signatures [43, 62]. The tectonomagmatic dichotomy is difficult to reconcile. However, compositionally similar rocks to the metagabbro are known to occur in postcollisional settings. The Late Ediacaran (~380 Ma) Kekem (Cameroun) and Guéra Massif (Chad) gabroic rocks have low SiO2 (45.7 to 51.0 wt%) and high TiO2 and P2O5 (TiO2 = 1.6 to 3.6 wt%; P2O5 = 0.8 to 2.0 wt%) contents (Figures 8(b) and 8(c)). They were emplaced 15-25 million years after the terminal collision (600–590 Ma) between the Saharan Metacraton and the Congo-São Francisco Craton. Both the Kekem and Guéra gabroids are considered to be derived from a subduction-modified lithospheric mantle source from the spinel-garnet transition zone [75, 76].

The Late Ediacaran tectonomagmatic setting of Central Africa is broadly similar to that of the Blair River inlier before and during emplacement of the metagabbro. The collision between the Avalon terrane and Laurentia was likely preceded by westward subduction toward Laurentian and the formation of Late Silurian arc and later back-arc (post-collisional) magmatic rocks throughout Newfoundland, New Brunswick, and Nova Scotia [42, 43]. Collision and crustal thickening occurred before 428 Ma which could account for subsequent mantle melting within the garnet stability field to produce the metagabbro. Keppie et al. [16] suggests that
| Pluton No. | Location | Terrane | Age (Ma)             | Mineral          | Reference                  |
|-----------|----------|---------|----------------------|------------------|-----------------------------|
| Red River sheet | Cape Breton Island | Blair River inlier | 425.1 ± 2.2 | Zircon | This study |
| Meta-gabbro | Blair River inlier | 423.8 ± 2.5 | Zircon | This study |
| Red River anorthosite | Blair River inlier | 421 ± 3 | Zircon | Keppie et al. [16] |
| Fox Back ridge | Blair River inlier | 423 ± 3 | Titanite | Miller et al. [17] |
| Fox Back ridge | Blair River inlier | 417 ± 6 | Hornblende | Miller et al. [17] |
| Red River syenite | Blair River inlier | 425 ± 2 | Titanite | Miller et al. [17] |
| Lowland Brook syenite | Blair River inlier | 428 ± 7 | Muscovite | Miller et al. [17] |
| Lowland Brook syenite | Blair River inlier | 424 ± 3 | Titanite | Miller et al. [17] |
| Otter Brook gneiss | Blair River inlier | 423 ± 6 | Titanite | Miller et al. [17] |
| Gneissic anorthosite | Blair River inlier | 427 ± 2 | Titanite | Miller et al. [17] |
| Otter Brook gneiss | Blair River inlier | 421 ± 6 | Phlogopite | Miller et al. [17] |
| Red River anorthosite | Blair River inlier | 424 ± 4/3 | Titanite | Miller et al. [17] |
| Sammys Barren | Blair River inlier | 435 ± 7/3 | Zircon | Miller et al. [17] |
| MacLean Brook | Aspy | 442.4 ± 2 | Zircon | Slaman et al. [35] |
| Lavis Brook | Aspy | 440 ± 1.6 | Zircon | Slaman et al. [35] |
| Gillanders Mountain | Aspy | 436.7 ± 2.6, 428.6 ± 1.9 | Zircon | Barr et al. [33], Lin et al. [19] |
| Gills Brook | Aspy | 436.4 ± 1.5 | Zircon | Barr et al. [33] |
| Lake Ainslie | Aspy | 428.5 ± 1.7 | Zircon | Barr et al. [33] |
| Grand Falaise | Aspy | 427.9 ± 2.1 | Zircon | Slaman et al. [35] |
| Leonard MacLeod Brook | Aspy | 420.9 ± 2.8 | Zircon | Barr et al. [33] |
| Pass Island | Newfoundland | Avalonia | 423 ± 4 | Zircon | Kellet et al. (2014) |
| Welsford | New Brunswick | Ganderia | 422 ± 1 | Zircon | Bevier [49] |
| Utopia | Ganderia | 428.3 ± 1, 425.5 ± 2.1, 420.4 ± 2.4, 420 ± 3.5Ma | Zircon, monazite | Barr et al. [48], Mohammadi et al. [46] |
| Bocabec gabbro | Ganderia | 422.1 ± 1.3 Ma | Zircon | Clarke et al. [44] |
| Jake Lee Mountain | Ganderia | 418 ± 2 | Zircon | Mohammadi et al. [46] |
| Mooshorn | Maine | Coastal Maine | 421.1 ± 0.8 | Zircon | McLaughlin et al. [53] |
| Penobscot | Coastal Maine | 419 ± 2 | Zircon | Stewart et al. [56] |
| Spruce Head | Coastal Maine | 421 ± 1 | Zircon | Tucker et al. [57] |
| Cadillac Mountain | Coastal Maine | 424 ± 2, 419 ± 2 | Zircon | Seaman et al. [55] |
Table 2: Continued.

| Pluton              | No. | Location         | Terrane | Age (Ma)     | Mineral | Reference             |
|---------------------|-----|------------------|---------|--------------|---------|-----------------------|
| Sedwick             | 22  | Coastal Maine    |         | 419.5 ± 1    | Zircon  | Stewart et al. [56]   |
| Cranberry Island    | 23  | Coastal Maine    |         | 424 ± 1      | Zircon  | Seaman et al. [55]    |
| Brenton pluton      | 24  | Nova Scotia      | Meguma  | 439 + 4/−3   | Zircon  | Keppie and Krogh [50] |
| White Rock formation| 25  | Meguma           |         | 438 ± 3/−2   | Zircon  | MacDonald et al. [52] |
| Mavillette gabbro   | 26  | Meguma           |         | 426 ± 2      | Baddeleyite | Warsame [58] |
the fault and shear zone pattern of NW Cape Breton is indicative of a positive flower structure and that the Blair River inlier was rapidly exhumed during the Early Devonian (415-410 Ma) probably due to sinistral reverse motion along the Cape Ray suture that is related to sinistral transpression along the Dover fault in Newfoundland. Furthermore, the metagabbro is contemporaneous with the emplacement of the 'A-type' granites of the Saint George Batholith of southern New Brunswick. The Saint George Batholith (Welsford, Utopia, and Jake Lee Mountain plutons and Bocabec gabbro) intruded rocks of the southern Gander terrane and is very close to the boundary with Avalonia [46]. In other words, it is possible that the gabbro was emplaced during a period of postcollisional crustal relaxation after the Avalon-Gander collision but prior to Early Devonian sinistral fault motion in Newfoundland.
6.3. Petrogenetic Model of the Anorthositic Rocks. Anorthosite is an uncommon plutonic rock that is almost entirely (≥ 90%) comprised of plagioclase feldspar [77, 78]. There are at least six different varieties of anorthosite identified by Ashwal [77, 79] that include: Archean anorthosite, Proterozoic massif anorthosite, layered intrusion anorthosite, oceanic anorthosite, anorthosite inclusions, and extraterrestrial anorthosite. The most abundant type of terrestrial anorthosite is the Proterozoic massif-type anorthosite [80, 81]. Proterozoic massif-type anorthosite is associated with large (up to ~18 000 km²) composite batholiths that range in age from ~2.6 Ga to ~0.5 Ga. The batholiths also include silicic rocks (mangerite, charnockite, and granitoids) that are coeval but not petrogenetically related to the anorthosite [81, 82]. The plagioclase compositions are commonly between An 70 and An 90 although some anorthosite masses may have a restricted range (An 45–65) whereas others have a wide range [81]. The formation of massif-type anorthosite is not entirely understood but is ultimately related to efficient segregation and accumulation of plagioclase from a mafic parental magma [77, 80–83].

In order to test the possible petrological relationship between the metagabbro and anorthositic rocks, we apply fractional crystallization modeling using MELTS [84, 85]. For the model, we use a parental magma composition equal to the metagabbro. The precise magmatic conditions of the system at the time of emplacement are unknown; however, given the likely collisional/post-collisional tectonic setting [62], we based the modeling conditions (H₂O content and fO₂) on such settings (Table 3). We use a relatively moderate water content (~2.0 wt%) for the parental magma and a relative oxidation state equal to the Ni-NiO buffer (ΔFMQ +0.7). The moderate water content is selected for two reasons: (1) the rock has high P₂O₅ content that suggests it was derived from a volatile-rich system and (2) the rock composition indicates it already experienced fractionation of mafic silicate minerals (olivine, pyroxene) that likely increased the bulk water content of the residual magma. The choice of the redox conditions is based on the fact that mafic magmas associated with collisional settings tend to be +1 to +4 log units (ΔFMQ +1 to +4) above that of mid-ocean ridge basalt although our choice of the NNO (ΔFMQ +0.7) buffer is not anomalously high [86, 87]. The pressure (0.2 GPa) of the model is equal to a depth associated with the lower upper crust or upper middle crust.

The results of the fractional modeling are summarized in Table DR3 and shows that the system is dominated by feldspar (55.6%), pyroxene (10.9%), and spinel (11.9%) fractionation with ~10% residual liquid remaining. The residual liquid will not match the bulk composition of the anorthositic rocks given that they are ‘cumulate’ rocks, but we selected the mineral compositions generated by the model that are within the compositional range of the minerals of the Red River anorthosites presented by Miller [22] in order to ‘reconstruct’ the rocks by mass balance. Tables 4 and 5 summarize the results of mass balance and show that the modeled mineral compositions derived from the metagabbro can yield compositions that are very similar to the anorthosite dykes and main body anorthosite. Miller [22] observed that the anorthositic rocks are 90% to 97% plagioclase with a compositional range from An 13-50 to An 30-56 in the least altered samples. The amount of plagioclase required in both mass balance calculations are 94.2% and 90.2%, respectively. The compositions of the feldspars used in the calculations are An₄₉ for calculation 1 (Table 4) and An₃₉ for calculation 2 (Table 5). The lower An content reflects the higher SiO₂ and Na₂O contents and lower Al₂O₃ content of the ‘sodic’ anorthosite but also the lowest measured An component (An₃₀) observed by Miller [22]. The mafic minerals observed in the anorthosite include biotite, chloride, epidote, Fe-Ti oxide minerals with recrystallized orthopyroxene, and altered clinopyroxene [22]. In addition to 94.2% plagioclase (An₄₉) for calculation 1, 1.7% titanomagnetite, 4.0% orthopyroxene (Mg# 70), and 0.1% apatite can reproduce the main body anorthosite composition. In contrast, calculation 2 requires 0.4% titanomagnetite, 0.17% apatite, and 9.0% biotite to reproduce the bulk composition of the anorthositic dykes.

The modeled mineral compositions we used in our calculation did not necessarily crystallize at the same time. For example, the modeled orthopyroxene composition used in the first mass balance calculation (Table 4) crystallized at a temperature of 950°C, whereas the modeled plagioclase compositions crystallized at 920°C and the Ti-magnetite at 820°C. Conceptually, this suggests that some minerals (e.g., orthopyroxene) within the anorthosite were entrapped within a plagioclase mush whereas others (e.g., Ti-magnetite) may have crystallized from an interstitial liquid within the plagioclase mush.

| Sample | BR-2-1 | Model |
|--------|--------|-------|
| SiO₂ (wt%) | 46.61 | 47.36 |
| TiO₂ | 1.95 | 1.98 |
| Al₂O₃ | 17.19 | 17.47 |
| Fe₂O₃ | 11.56 | 11.75 |
| MnO | 0.16 | 0.16 |
| MgO | 4.85 | 4.93 |
| CaO | 7.68 | 7.80 |
| Na₂O | 3.04 | 3.09 |
| K₂O | 1.82 | 1.85 |
| P₂O₅ | 1.58 | 1.61 |
| H₂O | 2.0 | 2.0 |
| Pressure | | 0.2 GPa |
| fO₂ | | ΔFMQ 0.7 |

FMQ: fayalite-magnetite-quartz buffer.
processes that led to the concentration of the plagioclase, it was likely due to a combination of density stratification and residual melt segregation (c.f., [77]). Regarding the likely tectonic setting of this system, as mentioned previously, the Avalon terrane collided with the margin of Laurentia during the Late Silurian [43, 61, 88]. The emplacement of older (>428 Ma) silicic plutons in Cape Breton Island and the southern margin of the Gander terrane prior to collision between the accreted margin of Laurentia and Avalonia are similar to I-type (cordilleran) granites, whereas younger (≤ 428 Ma) plutons are

Table 4: Mass balance calculation of the Red River anorthosite.

| Sample | SiO₂ | TiO₂ | Al₂O₃ | Fe₂O₃ | MnO | MgO | CaO | Na₂O | K₂O | P₂O₅ | Total  |
|--------|------|------|-------|-------|-----|-----|-----|------|-----|------|--------|
| BR-1-1 | 54.58| 0.20 | 26.60 | 2.33  | 0.03| 1.65| 8.89| 5.14 | 0.54| 0.03 | 100.00 |
| BR-1-2 | 54.49| 0.14 | 26.74 | 2.19  | 0.03| 0.94| 9.89| 4.82 | 0.69| 0.05 | 100.00 |
| Mass balance* | 54.65| 0.15 | 26.26 | 2.38  | 1.00| 9.66| 5.15 | 0.56 | 0.04| 99.99|

Mineral symbols: An: anorthite component; Pl: plagioclase; Ti-Mag: titanomagnetite; Ap: apatite; Opx: orthopyroxene.

Table 5: Mass balance calculation of the anorthositic sheet.

| Sample | SiO₂ | TiO₂ | Al₂O₃ | Fe₂O₃ | MnO | MgO | CaO | Na₂O | K₂O | P₂O₅ | Total  |
|--------|------|------|-------|-------|-----|-----|-----|------|-----|------|--------|
| BR-3-1 | 58.05| 0.14 | 23.28 | 2.93  | 0.06| 1.27| 5.62| 6.40 | 2.19| 0.07 | 100.00 |
| BR-3-2 | 57.59| 0.14 | 25.01 | 1.89  | 0.03| 0.58| 4.74| 6.40 | 3.30| 0.33 | 100.00 |
| Mass balance* | 58.08| 0.16 | 23.56 | 1.87  | 1.35| 5.60| 6.61| 2.10 | 0.07| 99.40|

Mineral symbols: An: anorthite component; Pl: plagioclase; Ti-Mag: titanomagnetite; Ap: apatite; Bt: biotite. Apatite composition taken from Deer et al. [100]. Biotite data from Shellnutt and MacRae [101].
similar to A-type (postcollision) granites and within-plate gabbros [33, 35, 46, 55, 89–91]. Given the emplacement age and the timing of A-type granite plutonism in southern New Brunswick (Saint George Batholith), it is likely that the anorthositic rocks of the Blair River inlier were emplaced at the same postcollisional setting, which may be an important environment for the creation of some Proterozoic and Phanerozoic massif-type anorthosites [81, 92, 93].

7. Conclusions
The anorthositic and mafic rocks collected from the southern Blair River inlier of Cape Breton are contemporaneous and were emplaced during a period of Late Silurian crustal relaxation. The anorthositic rocks of this study are different whereas one is compositionally similar to labradorite, the other is more similar to andesine although it has an appreciable amount of K~2O. The cumulate nature of the rocks is confirmed by their petrography, low concentrations of incompatible trace elements, and their high positive chondrite-normalized Eu anomalies (Eu/Eu* $>$ 2.5). The Fox Back Ridge metagabbro is compositionally similar to alkali basalt and was likely generated by partial melting of garnet-bearing peridotite. Fractional crystallization modeling coupled with mass balance calculations indicates that the anorthosite was emplaced during a period of Late Silurian crustal relaxation. The anorthositic rocks of this study are diagenetically different from a maﬁc parental magma similar in composition to the metagabbro of this study. It is likely that the emplacement of the Red River anorthosite occurred at a postcollision setting within the Avalon terrane and that the Laurentian boundary is located farther to the west.

Data Availability
All data relevant to the publication of manuscript L1111 was uploaded during the original submission. This includes data tables to be included within the manuscript and supplementary files.

Conflicts of Interest
The authors declare that they have no conflicts of interest.

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Supplementary Materials
Supplementary Figures DR1 and DR2: cathodoluminescence images of zircons. Table DR1: results of zircon geochronology for CBGR-1. Table DR2: results of zircon geochronology for CBCH-1. Table DR3: results of MELTS modeling. (Supplementary Materials)

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