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AN ORDOVICIAN, $^{40}Ar/^{39}Ar$ STEP-HEATING AGE FOR FABRIC-FORMING HORNBLENDE IN AMPHIBOLITE, THE GREAT BEND COMPLEX, CENTRAL NEWFOUNDLAND (NTS 2D/5)

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ABSTRACT

The Great Bend Complex is a strongly dismembered ophiolite remnant in the eastern Exploits Subzone of the Dunnage Zone (Newfoundland Appalachians) and inferred to correlate with the peri-Gondwanan, Late Cambrian Pipestone Pond and Coy Pond (ophiolite) complexes. The Great Bend Complex structurally overlies turbiditic metasedimentary rocks of the Mount Cormack Subzone along part of its northwestern margin. It is intruded by gabbro of the Silurian Mount Peyton intrusive suite to the east and north, and thrust imbricated with Sandbian graptolitic black shale, and Silurian sandstone, siltstone and polymictic conglomerate inferred to be part of the Indian Islands Group.

The Great Bend Complex comprises fault-bounded blocks of variably deformed and altered ensimatic rocks including spinel-facies dunite and peridotite, gabbro, amphibolite, diabase and pillow lava spatially associated with black shale and common black-shale mélange. Although undated, metamorphosed basaltic rocks of the complex are comparable to those of the Pipestone Pond Complex and derived from a subduction-zone-modified depleted mantle. One sample of a small exposure of hornblende+plagioclase flaser-textured gabbro, yielded a previously unpublished $^{40}Ar/^{39}Ar$ step-heating, bulk hornblende-grain separate quasi-plateau-age of 471 ± 4 Ma (56.0% of the ^{39}Ar released). This overlaps, within error, the U–Pb zircon crystallization age of 474 +6/-3 Ma for the terrane-stitching Partridgeberry Hills Granite. Furthermore, it is slightly older than the age of a fossiliferous, ophiolite-detritus-bearing, late Dapingian to early Darriwilian limestone conglomerate unconformably overlying the Mount Cormack Subzone. The ca. *471 Ma cooling age is therefore a minimum estimate for the time of terminal eastward transport of the Great Bend Complex over Ganderia.*

INTRODUCTION

The Great Bend Complex (GBC) is exposed in the Exploits Subzone of the eastern Dunnage Zone (Figure 1), situated ~70 km south of the Trans-Canada Highway (Route 1) and lying trans-longitudinally athwart the Bay D'Espoir Highway (Route 360). The GBC and the spatially associated Penobscot ophiolite complexes of the eastern Dunnage Zone, including the Coy Pond Complex (510 ± 4 Ma; Colman-Sadd, 1980, 1985; Colman-Sadd *et al.*, 1992; Dickson, 1992, 1996; Sandeman *et al.*, 2012) and the Pipestone Pond Complex (493.9 +2.5/-1.9 Ma; Colman-Sadd and Swinden, 1984, 1985; Dunning and Krogh, 1985; Jenner and Swinden, 1993) are exposed around, and partially encircle, the Neoproterozoic to Cambrian turbiditic metasedimentary rocks of the Mount Cormack Subzone (Colman-Sadd and Swinden, 1985; Colman-Sadd *et al.*, 1992; Valverde-Vaquero *et al.*, 2006; Figures 1 and 2). The Mount Cormack Subzone is inferred to represent a tectonic window into the "structural basement" of the eastern Exploits Subzone, exposing the greenschist- to upper amphibolite-facies grade, metasedimentary and granitoid rocks of the Ganderian Mount Cormack Subzone (Colman-Sadd and Swinden, 1984, 1985; Williams *et al.*, 1988; Colman-Sadd *et al.*, 1992; Valverde-Vaquero *et al.*, 2006).

The ultramafic complexes were first recognized during a reconnaissance, river-based exploration (*e.g.*, Howley, 1918). This recognition led to a number of middle 20th century programs of exploration for chromite, nickel, copper and other base metals (*e.g.*, Snelgrove, 1934; Grady, 1953; Coleman, 1954; Kean, 1974; Zwicker and Strong, 1986), using the model that the Great Bend and Coy Pond complexes possibly represented layered ultramafic intrusions into the crust. However, with the advent of the plate tectonic theory, the ultramafic rocks of the region were re-interpreted to represent remnants of Ordovician oceanic crust preserved on the eastern margin of the Exploits Subzone

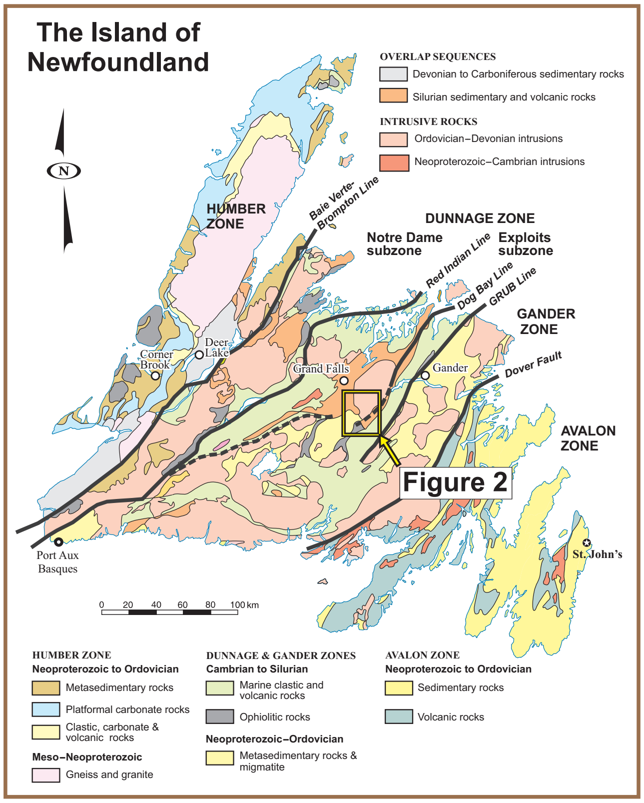

Figure 1. *Simplified geological map of Newfoundland showing the location of the study area with respect to major geological terranes and tectonic boundaries (after Colman-Sadd* et al.*, 1990).*

during amalgamation of the differing terranes of the Appalachian Orogen (Church and Stevens, 1971; Dewey and Bird, 1971). Such a hypothesis garnered significant geoscience interest during the following decades (Dunning and Krogh, 1985; Dec and Colman-Sadd, 1990; Colman-Sadd *et al.*, 1992; Jenner and Swinden, 1993), resulting in a renewed focus on these rocks, particularly for mother-lode-style, orogenic gold systems (*e.g.*, Vaskovic, 1987; Butler, 1988; Mercer, 1988a, b; Graham, 1989, 1990; Bradley and Graham, 1989) spatially associated with ophiolitic rocks (*cf.* listwanite; Halls and Zhao, 1995).

This short contribution reports a previously unpublished $^{40}Ar/^{39}Ar$ cooling age for K-poor amphibole from a strongly foliated, flaser-textured metagabbro of the GBC. The strongly deformed and metamorphosed flaser-textured gabbro and amphibolite (*see* Dickson, 1992) was inferred to occur near the structural base of the complex. It forms the basal dynamothermal aureole of the ophiolite (*e.g.*, Williams and Smyth, 1973; Woodcock and Robertson, 1977), accompanying its tectonic emplacement and transport onto the Gander margin in the Ordovician (Colman-Sadd and Swinden, 1984; Colman-Sadd *et al.*, 1992). Along with new trace-element lithogeochemical data for the specimen, the compositional and thermochronological data verify the intraoceanic back-arc character of the gabbro and its minimum age of emplacement onto the leading edge of the Iapetan Ganderian substrate. This age also provides a maximum age constraint on the emplacement of epigenetic, orogenic gold mineralization in the rocks of the region. All ages herein refer to the International Commission on Stratigraphy International Chronostratigraphic Chart (v2018/8).

REGIONAL SETTING

The GBC lies within the eastern Exploits Subzone of the Dunnage Zone (Figure 1) in the Newfoundland Appalachians. The eastern Exploits Subzone consists of accreted arc and back-arc volcanic and marine sedimentary rocks that formed near the Ganderian-leading margin in the Iapetus Ocean during the Cambrian and Ordovician (van Staal *et al.*, 1996; van Staal and Barr, 2012). These sequences of rocks were subsequently tectonically emplaced eastward (present-day coordinates) onto Ganderia's Laurentian-facing margin during the Middle Ordovician Penobscot orogeny (475–465 Ma; Neuman, 1967; Colman-Sadd *et al.*, 1992; Zagorevski *et al.*, 2010). In the study area, the Ganderian basement of greenschist- to amphibolite-facies metaturbidites of the Mount Cormack Subzone, occur as a structural window through the Exploits Subzone (Colman-Sadd and Swinden, 1984; Colman-Sadd, 1985; Colman-Sadd *et al.*, 1992; Dickson, 1992, 1996; Valverde-Vaquero *et al.*, 2006). The Mount Cormack Subzone is a broadly oval, 25–35 km antiformal dome partially encircled by the Great Bend, Coy Pond and Pipestone Pond (ophiolite) complexes (Colman-Sadd and Swinden, 1984; Williams *et al.*, 1988; Colman-Sadd *et al.*, 1992). In the central and western parts of the Mount Cormack Subzone, the turbiditic rocks locally attain upper amphibolite-metamorphic facies and are intruded by the peraluminous Through Hill granite (Colman-Sadd and Swinden, 1984; Colman-Sadd *et al.*, 1992; Valverde-Vaquero *et al.*, 2006). The Through Hill granite formed through crustal anatexis and crystallized during upper amphibolite- to granulite-facies metamorphism at *ca.* 468–458 Ma (Colman-Sadd *et al.*, 1992; Valverde-Vaquero *et al.*, 2006), in response to tectonic loading on the Gander margin immediately post-ophiolite obduction. Significantly, the metamorphic isograds in the Mount Cormack Subzone show a concentric decrease in metamorphic grade outward toward the Cambro-Ordovician ophiolitic rocks (Colman-Sadd and Swinden, 1984; Colman-Sadd *et al.*, 1992). The Mount Cormack Subzone has therefore been up-domed during subsequent lower metamorphic grade, broadly greenschist tectonothermal activity (Early Devonian?).

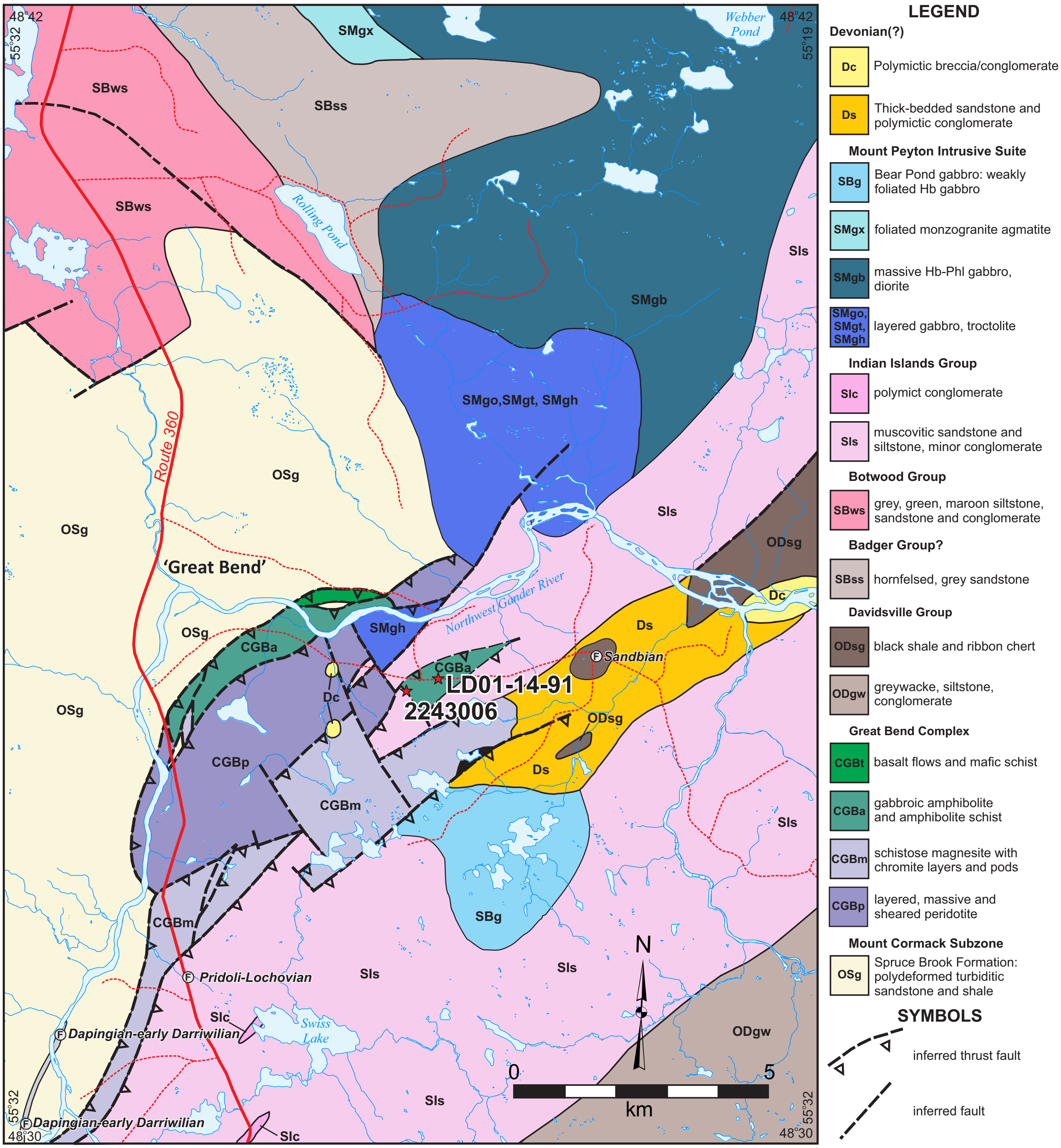

Figure 2. *Simplified geological map of the Great Bend area (adapted from Colman-Sadd* et al.*, 1990, 1992; Dickson, 1992, 1996; Sandeman, unpublished data, 2018) showing the location of the* $^{40}Ar/^{39}Ar$ *flaser-textured gabbro sample (LD01-14-91) and the second, amphibolite geochemistry sample (2243006).*

The rocks of the study area form a structurally imbricated series of lithologically diverse Ordovician back-arc, sedimentary–volcanic assemblages that include the Davidsville Group (Blackwood, 1982; Blackwood and Green, 1982, 1983; Dickson, 1996), the Baie D'Espoir Group (Colman-Sadd, 1985; Colman-Sadd *et al.*, 1992), the mafic–ultramafic and spatially associated sedimentary rocks of the Coy Pond and Great Bend complexes and, the Partridgeberry Hills Granite (Colman-Sadd, 1985; Colman-Sadd *et al.*, 1992; Dickson, 1992, 1996). The study area is also characterized by a range of green–red–maroon, sandstone–siltstone and pebble-to-cobble conglomerate of inferred Silurian, and possibly Devonian, ages (Dickson, 1992, 1996). South of the Northwest Gander River and immediately west of Route 360, an exposure of fine-grained, buff-grey sandstone yielded Late Silurian (Pridoli) to earliest Devonian (Lochovian) crinoids and brachiopods from a 20-cm-thick fossiliferous limestone debris flow (Donovan *et al.*, 1997). All of these units, with the possible exception of the youngest, (inferred Devonian (?) serpentine-clast-bearing conglomerate; Figure 2), are collectively structurally imbricated, with vergence to the north-northwest (Dickson, 1992, 1996).

The GBC comprises a number of fault-bounded blocks (Figure 2) that contain many of the classical ophiolite sequence rocks including peridotite, harzburgite and dunite (extensively hydrated and carbonatized); variably foliated, metabasic, gabbroic and basaltic rocks; diabase dykes; rare pillow basalts and trondhjemite. These large blocks (m- to km-scale) are commonly separated by intervals of tectonized black shale, olistostromal deposits and black shale mélange. Black and dark-grey shale is common in the Great Bend area and, in proximity to the igneous ophiolitic rocks, typically forms the preferentially flattened and sheared matrix to tectonic mélange containing diverse local rock fragments (*e.g.*, peridotite, sandstone, felsic tuff; Sandeman *et al.*, 2012). The age of the black shale is not always unambiguous. For example, a well-constrained graptolite assemblage from the Coy Pond Complex (Williams *et al.*, 1992) indicates that some of the shale is late Floian (*ca.* 475-470 Ma), whereas other exposures containing graptolitic fauna indicates that much of the shale is likely Sandbian (*e.g.*, Williams, 1991; S.H. Williams, personal communication 1992, *in* Dickson, 1992; Williams and Tallman, 1995). The majority of the exposed contacts between shale and other units of the region appear to be tectonized (Dickson, 1992, 1996) and represent zones of intense deformation. Geological contacts throughout the region are, therefore, highly strained and structurally complicated owing to polyphase deformation, resulting in uncertain primary rock relationships. All evidence indicates that the rocks of the region, ranging in age from Late Cambrian (Furongian) to earliest Devonian (Lockovian) are intercalated and likely thrust imbricated.

SAMPLE DESCRIPTION AND ^{40}Ar/^{39}Ar GEOCHRONOLOGY

A conventional ^{40}Ar/^{39}Ar step-heating age was determined for hornblende grains (Sample LD01-14-91) from a foliated, hornblende–plagioclase-bearing flaser-textured metagabbro cropping out near the inferred structural base of the GBC (Figure 2; *see* Plate 1; Dickson, 1992, 1996). The hornblende is typically ≤400 μm long, facilitating isolation of relatively pure mineral separates. The ^{40}Ar/^{39}Ar age was determined at the University of Maine ^{40}Ar/^{39}Ar Thermochronology Laboratory using the methods outlined by O'Neill and Lux (1989). The ^{40}Ar/^{39}Ar data are given in Table 1 and the age spectra are shown in Figure 3. At the time of the analysis, errors in the isotopic ratios were not routinely reported and therefore all are assumed to be 5 relative %.

Plate 1. *Photograph of the flaser-textured metagabbro in outcrop. Camera lens is 5 cm in diameter.*

The gas steps used in the calculation of the plateau ages are indicated by bold type in Table 1 and by shaded boxes in Figure 3. The data was processed using the ^{40}Ar/^{39}Ar age spectrum module of ISOPLOT v. 3.60 (Ludwig, 2008). The ^{40}Ar/^{39}Ar plateau ages are typically defined by at least 3 contiguous gas release steps (consisting of >60% of released ^{39}Ar), with ^{40}Ar/^{39}Ar ages overlapping within 2σ error (McDougall and Harrison, 1988 and references therein; Snee *et al.*, 1988; Singer and Pringle, 1996). A plateau must also be defined by ^{40}Ar/^{39}Ar steps with reasonably low excess scatter (MSWD <2.2).

These criteria were not fully satisfied by the gas-release spectrum for the sample under investigation, as the analysis yielded a "quasi-plateau" containing only 56% of the total ^{39}Ar released (Table 1; Figure 3). However, the internal consistency, relatively high Ca/K ratios and small volumes of contained atmospheric argon for the last 4 concurrent, high-temperature gas steps indicate that this quasi-plateau likely

Table 1. The $^{39}Ar/^{40}Ar$ thermochronological data for a hornblende grain separate from sample LD01-14-91. Bold steps are used in the quasi-plateau age

Gas Step	T (°C)	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	%Rad	Ca/K	Cumm. % ^{39}Ar Released	Age (Ma)	Error (Ma)
1	850	44.98	1.302	0.0577	49.8	62.3	2.66	6.9%	462.0	5.5
2	930	34.91	2.441	0.0230	71.1	81.0	5.00	9.8%	466.1	5.0
3	1000	31.63	8.188	0.0158	94.5	87.3	16.67	13.0%	457.9	4.9
4	1060	31.45	9.155	0.0154	49.6	87.8	18.87	6.2%	458.1	5.7
5	1110	30.16	10.857	0.0110	61.7	92.1	22.22	8.2%	461.1	4.2
6	**1170**	**29.68**	**11.278**	**0.0072**	**112.8**	**95.9**	**23.26**	**15.5%**	**471.1**	**4.2**
7	**1220**	**29.66**	**11.296**	**0.0070**	**113.4**	**96.0**	**23.26**	**15.6%**	**471.4**	**4.2**
8	**1260**	**30.41**	**12.128**	**0.0104**	**120.8**	**93.0**	**25.00**	**16.6%**	**468.9**	**4.2**
9	**1400**	**32.01**	**12.703**	**0.0155**	**59.0**	**88.8**	**26.32**	**8.2%**	**471.2**	**4.4**

Note: Quasi-plateau steps indicated by bold; Rad – radiogenic; Cumm. – cumulative; Total-gas age: 466.3 ± 4.6 Ma; Quasi-plateau age: 470.6 ± 3.9 Ma; Isotope correlation age: 467 ± 5 Ma ($^{40}Ar/^{36}Ar$ = 312 ± 13)

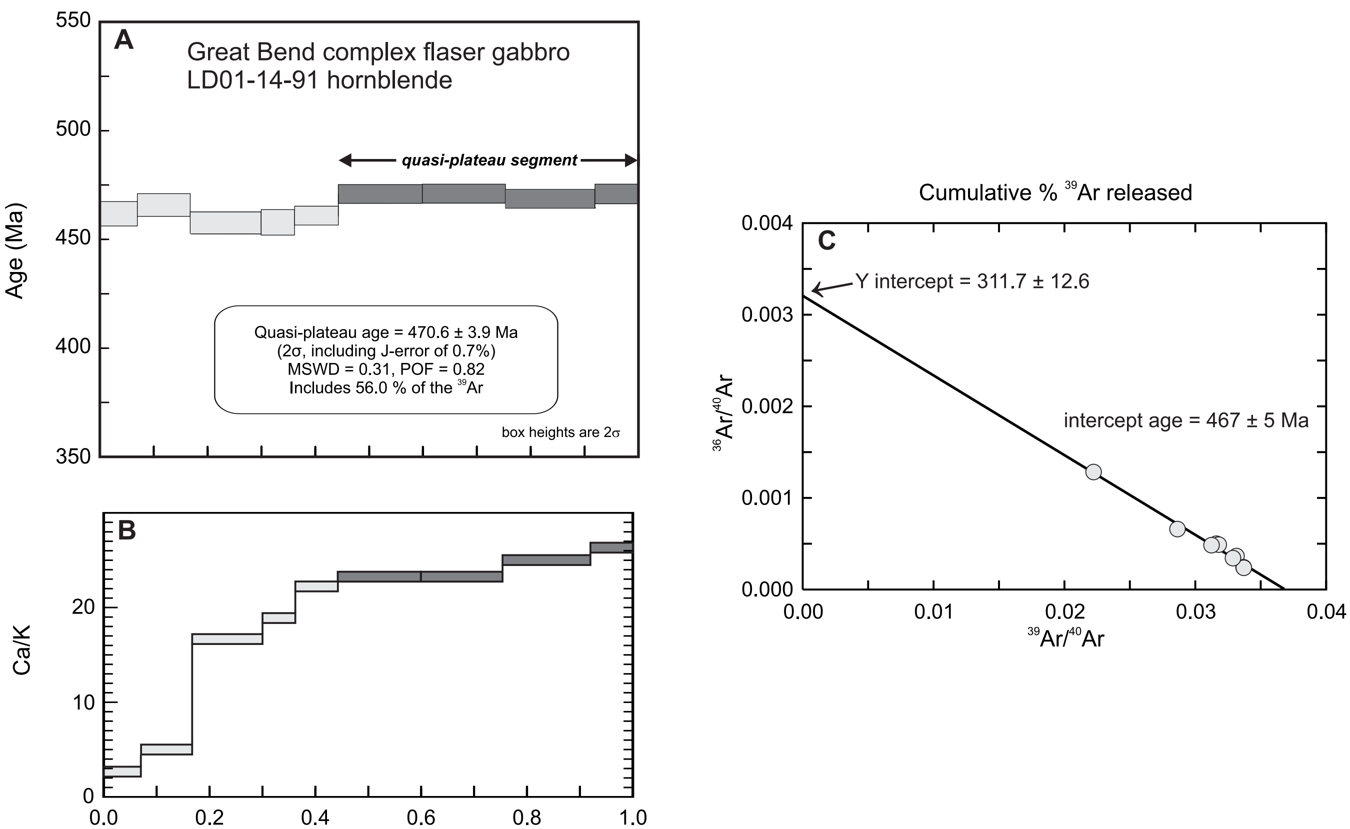

Figure 3. *A)* $^{40}Ar/^{39}Ar$ *age spectrum for hornblende from specimen LD01-14-91; B) Ca/K* vs. *cumulative fraction of* ^{39}Ar *released showing the relatively consistent and high Ca/K for the quasi-plateau steps; C)* $^{36}Ar/^{40}Ar$ vs. $^{39}Ar/^{40}Ar$ *inverse isochron plot for the sample.*

represents the cooling age of the hornblende grain separate. Age uncertainties are quoted at the 2σ uncertainty level.

The aliquot yielded a relatively simple argon release spectrum (Figure 3A). The ages of the first 5 gas fractions form a rough, humped plateau at *ca.* 461 Ma, but then rise abruptly to an older, flat plateau segment for the remainder of the higher temperature gas fractions. The quasi-plateau segment (steps 6 through 9), representing 56.0% of the ^{39}Ar released (MSWD = 0.31: POF = 0.82), exhibits high, but generally internally consistent Ca/K ratios (Figure 3B), is composed of a high proportion of radiogenic argon (93.4%) and yields an age of 471 ± 4 Ma. This quasi-plateau overlaps, within error, the total gas integrated age of 466 ± 5 Ma. Moreover, a $^{36}Ar/^{40}Ar$ *vs.* $^{39}Ar/^{40}Ar$ inverse isochron plot of all gas steps (Figure 3C) yields a simple linear regression age of 467 ± 5 Ma, which overlaps, within error, the quasi-plateau age. The resultant calculated atmospheric argon ratio is 311.7 ± 12.6.

GEOCHEMISTRY

Major- and selected trace-element data for sample LD01-14-91 (Lab #2243003) are available on the Government of Newfoundland and Labrador Natural Resources Geoscience Atlas (http://geoatlas.gov.nl.ca/Default.htm) in the Volcanic Rock Geochemical Database tab. Although useful, this database does not include low-detection limit, inductively-coupled plasma mass spectrometry (ICP-MS) analytical data for many of the critical incompatible trace elements (*e.g.*, Th, Nb, Hf, rare-earth elements (REE)). In order to address this lack of critical trace element data, aliquots of the archival rock powders for two mafic schists of the Great Bend Complex, the dated sample LD01-14-91 and amphibolite sample 2243006, were analyzed by ICP-MS at the Geological Survey Howley Building Geochemical Laboratory. The methods for ICP-MS analysis are outlined in Finch *et al.* (2018) and the complete lithogeochemical data for the two samples are presented in Table 2.

Samples LD01-14-91 and 2243006 are similar in composition to some of the basalt and diabase of the Pipestone Pond Complex (Jenner and Swinden, 1993) and Coy Pond Complex (Sandeman *et al.*, 2012). The two Great Bend samples are subalkaline, tholeiitic basalts (Figure 4A, B) with major, compatible trace-, and incompatible trace-element abundances comparable to intra-oceanic mid-ocean-ridge basalt (N-MORB: Figure 4). Diagnostic incompatible trace-element ratios such as Th/Yb, Nb/Yb and TiO_2/Yb are also characteristic of N-MORB (Figure 4C, D), but the two Great Bend samples have elevated Nb, relative to the majority of, but not all, Pipestone Pond and Coy Pond samples. Rocks from all three complexes have relatively low TiO_2/Yb, suggesting derivation from a shallow, spinel- or plagioclase-facies, asthenospheric mantle (Figure 4D). Samples LD01-14-91 and 2243006 have REE abundances and extended multi-element patterns comparable to N-MORB and to the lavas of the Pipestone Pond Complex (Figure 5), however, some diabase dykes of the Pipestone Pond and Coy Pond complexes have prominent Nb troughs indicative of their formation in a primitive, intra-oceanic arc setting.

DISCUSSION

Regional 1:50 000 scale mapping, U–Pb geochronology, lithogeochemistry, and paleontological investigations (Colman-Sadd, 1985; Dunning and Krogh, 1985; Dec and Colman-Sadd, 1990; Colman-Sadd *et al.*, 1992; Dickson, 1992, 1996; Jenner and Swinden, 1993; Sandeman *et al.*, 2012) have established that the Pipestone Pond and Coy Pond complexes, and correlative GBC, are Late Cambrian to earliest Ordovician, ocean-floor fragments (back arc?) tectonically emplaced eastward (present-day coordinates) over Ganderian substrate in the Early to Middle Ordovician. Three important constraints on this central Newfoundland Early Ordovician orogenesis (Penobscot orogeny) are particularly salient. First, an ophiolite–detritus-bearing, limestone conglomerate containing late Dapingian to early Darriwilian (469–466 Ma) brachiopods and trilobites is inferred to unconformably overlie the Mount Cormack Subzone (Dec and Colman-Sadd, 1990; Colman-Sadd *et al.*, 1992). Second, the 474 +6/-3 Ma medium-grained, biotite ± muscovite Partridgeberry Hills Granite cuts the Coy Pond Complex, the Mount Cormack Subzone and perhaps the Ordovician North Steady Pond Formation, and is therefore considered a terrane-stitching pluton (Colman-Sadd *et al.*, 1992). Third, the maximum attained metamorphic grade in all of the rocks around the GBC is upper amphibolite- to lower granulite-facies in the core of the Mount Cormack Subzone. These high-grade rocks formed immediately after east-directed thrusting as a result of crustal thickening and loading, and metamorphism was accompanied by the onset of dehydration melting of the Mount Cormack metasedimentary rocks and emplacement of the 464 +4/-3 Ma peraluminous Through Hill granite (Colman-Sadd *et al.*, 1992; Valverde-Vaquero *et al.*, 2006). At *ca.* 464 Ma, the high-grade rocks were at a much deeper structural level in the crust than the Great Bend Complex, but they are now exposed near the core of a younger structural dome. It is inferred, therefore, that the rocks of the Great Bend Complex and the flaser-textured gabbro have not been thermally disturbed above greenschist facies, since their obduction. The *ca.* 471 Ma horn-

Table 2. Petrochemical data for amphibolitic gabbro LD01-14-91 and amphibolite 2243006

Sample	LD01-14-91	2243006	
Lab #	2243003	2243006	
rock type	amphibolitic gabbro	amphibolite	
UTMEAST	616050	615400	
UTMNORTH	5381950	5381600	
UTMZONE	21	21	
DATUM	NAD27	NAD27	Analytical Method
SiO_2	48.75	49.30	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
TiO_2	1.22	1.10	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
Al_2O_3	14.17	14.00	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
Fe_2O_3	4.60	1.98	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
FeO	7.81	8.97	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
MgO	7.20	8.16	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
MnO	0.21	0.20	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
CaO	10.01	9.10	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
Na_2O	2.99	3.60	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
K_2O	0.51	0.32	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
P_2O_5	0.10	0.07	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
LOI	2.40	2.40	GRAV
H2O	2.62	2.78	IGA
CO2	0.18	0.24	IGA
S	0.03	0.04	IGA
Total	100.40	99.86	
Ag	0.1	0.1	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
As	2	2	INAA
Au	1	1	INAA
Ba	55	124	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
Be	1	1	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Bi	0.7	b.d.	ICP-MS
Br	1	1	INAA
Cd	0.1	0.1	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Co	52	48	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Cr	123	131	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
Cs	0.2	0.6	ICP-MS

Sample	LD01-14-91	2243006	
Lab #	2243003	2243006	
rock type	amphibolitic gabbro	amphibolite	
UTMEAST	616050	615400	
UTMNORTH	5381950	5381600	
UTMZONE	21	21	
DATUM	NAD27	NAD27	Analytical Method
Cu	80	38	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
F	169	241	ISE
Ga	22	21	ICP-MS
Ge	1.6	1.3	ICP-MS
Hf	1.7	1.7	ICP-MS
Li	14.6	14.3	ICP-MS
Nb	3.4	3.2	ICP-MS
Ni	66	69	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Pb	1	1	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Rb	12	8	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Sb	0.3	0.2	INAA
Sc	51	48	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Se	1	1	INAA
Sn	2	2	ICP-MS
Sr	69	61	ICP-MS
Ta	0.1	b.d.	ICP-MS
Th	0.3	0.2	ICP-MS
U	b.d.	b.d.	ICP-MS
V	343	327	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
W	1.8	1.5	ICP-MS
Y	22.8	20.7	ICP-MS
Zn	104	104	ICP-OES ($HF/HCl/H_3BO_3$ digestion)
Zr	64	59	ICP-OES ($LiBO_2$ fusion, $HF/HCl/H_3BO_3$ digestion)
La	2.52	2.57	ICP-MS
Ce	6.63	6.81	ICP-MS
Pr	1.20	1.15	ICP-MS
Sm	2.48	2.16	ICP-MS
Nd	6.45	6.17	ICP-MS
Eu	0.89	0.81	ICP-MS
Gd	3.82	3.19	ICP-MS
Tb	0.63	0.58	ICP-MS
Dy	4.26	3.96	ICP-MS
Ho	0.97	0.85	ICP-MS
Er	2.75	2.58	ICP-MS
Tm	0.45	0.38	ICP-MS
Yb	2.86	2.37	ICP-MS
Lu	0.48	0.41	ICP-MS

Key: b.d. = below detection; ICP-MS = inductively coupled plasma mass spectrometry (after Finch *et al.*, 2018); ICP-OES = inductively coupled plasma optical emission spectrometry (after Finch, 1998); INAA = instrumental neutron activation analysis (Bequerel Laboratories, now Maxxam); GRAV = gravimetric analysis; ISE = ion specific electrode; IGA = infrared gas analyser

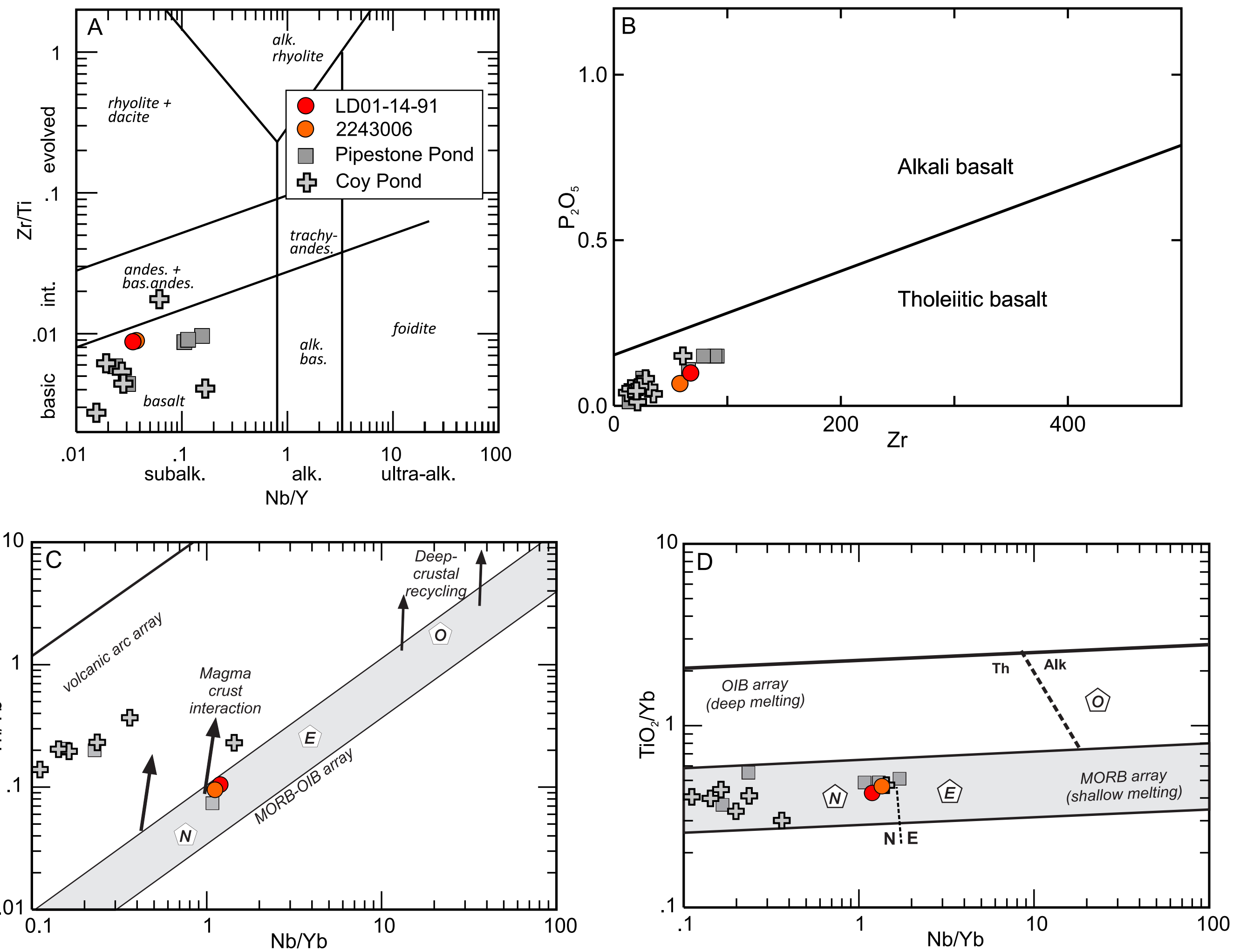

Figure 4. *Lithogeochemical diagrams comparing sample LD01-14-91 and 2243006 with historical data for the Pipestone Pond complex (Jenner and Swinden, 1993) and the Coy Pond Complex (Sandeman* et al*., 2012). A) Immobile element rock classification diagram after Pearce (1996); B)* P_2O_5 vs. *Zr demonstrating the subalkaline, tholeiitic composition of the samples (Winchester and Floyd, 1976); C) Th/Yb* vs. *Nb/Yb discrimination plot (Pearce, 2008). D)* TiO_2/Yb vs. *Nb/Yb discrimination plot (Pearce, 2008).*

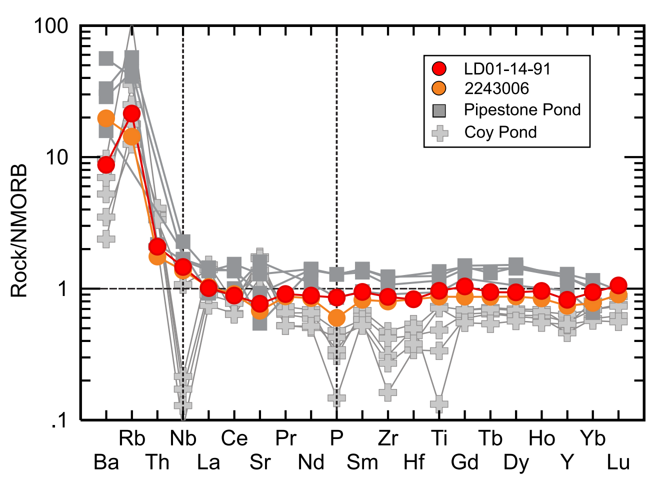

Figure 5. *NMORB normalized extended incompatible trace-element diagram comparing samples LD01-14-91 and 2243006 with historical data for lavas of the Pipestone Pond complex (Jenner and Swinden, 1993) and the Coy Pond Complex (Sandeman* et al*., 2012).*

blende cooling age reported here is in very good agreement with the other time constraints on Penobscott orogenesis and is, therefore, interpreted to be a minimum estimate for the time of terminal eastward transport of the Great Bend Complex over Ganderia. This cooling age is also a likely maximum age constraint on epigenetic orogenic gold mineralization in the greater area.

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