

Geochemical characteristics of gold-related granitoids in southwestern New Brunswick, Canada

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Abstract

Two groups of granitoids associated with gold mineralization in the Appalachian orogen of southwestern New Brunswick are recognized: a Late Silurian to Early Devonian (423–396 Ma) granodioritic to monzogranitic series (GMS), and a Late Devonian (370–360 Ma) granitic series (GS). The GMS granitoids are relatively low in silica, calc-alkaline, metaluminous to weakly peraluminous, and show characteristics of normal (oxidized) to reduced I-type granites depending on the properties of country rocks. They may have been derived from partial melting of lower crustal rocks triggered by underplated basaltic magmas; and country rocks bearing reduced organic carbon and/or graphite may have played an important role in the reduction of normal I-type intrusions to reduced I-type, which is essential in the formation of intrusion-related gold systems. In contrast, the GS granites, although calc-alkaline and metaluminous to peraluminous, are relatively rich in silica, incompatible elements, and high field strength elements. They are fractionated I-type granites, and are probably related to the coeval Mount Douglas granite in the Saint George batholith through fractional crystallization. Their parental magmas may have been derived from partial melting of quartzofeldspathic sources at relatively low temperatures. Both GMS and GS intrusions are orogenic, although some of them display the affinity of those emplaced into a within-plate environment. The origin of intrusion-related gold systems in this region appears to be controlled by several factors, including magma sources, magmatic processes, redox conditions (country-rock nature), and local structural regimes.

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1. Introduction

Recently discovered gold deposits and occurrences (e.g., Clarence Stream, Lake George, Poplar Mountain, Jimmy Hill, McDougall Brook, and Tower Hill) associated with Late Silurian to Late Devonian granitoid intrusions in southwestern New Brunswick (McLeod and McCutcheon, 2000; Chi, 2002; Lentz et al., 2002; Yang et al., 2002, 2003), a well-known W–Sn–Mo–Sb–Bi mineralization province (McLeod, 1990; Whalen, 1993),

share some similarities with the recently recognized intrusion-related class of gold deposits. This type of gold deposit differs from porphyry copper–gold systems in fluid composition, redox state, mineralization style, alteration, elemental distribution, mineral assemblages, and tectonic setting. Intrusion-related gold deposits are characterized by carbonic fluids, reduced I-type intrusions, association with Bi, As and Te, narrow fracture-controlled alteration zones, reduced sulfide assemblages, and a continental setting well inboard of convergent plate boundaries (McCoy et al., 1997; Thompson et al., 1999; Thompson and Newberry, 2000; Lang and Baker, 2001; Baker, 2002; Groves et al., 2003; Blevin, 2004; Hart et al., 2004). The newly developed model put forth for intrusion-related gold systems has stimulated exploration efforts for gold in southwestern New Brunswick, and has highlighted the need for studies to determine

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the essential controlling factors for the formation of these gold deposits.

Previous research on granitoids in southwestern New Brunswick (Butt, 1976; Cherry, 1976; Fyffe et al., 1981; Lentz et al., 1988; Sinclair et al., 1988; McLeod, 1990; Taylor, 1992; Whalen, 1993; Whalen et al., 1994, 1996; McCutcheon et al., 1997) provided a sound basis for further evaluation of the petrogenesis of these granitoids and their relationships to gold mineralization, which is the focus of this study. Their petrogenesis is examined from petrochemistry, isotopic compositions, and magma temperatures, and their relationship to gold mineralization is discussed based on magma types and sources, magmatic and hydrothermal processes, redox conditions, and tectonic regime.

2. Geological setting and two groups of granitoids

The Canadian Appalachian orogen is divided into several Lower Paleozoic tectonic zones, which are overlain by Middle Paleozoic basins and then by an Upper Paleozoic successor basin (Williams et al., 1999). In New Brunswick, from northwest to southeast, Lower Paleozoic tectonic zones include the Dunnage, Gander, and Avalon zones (Fig. 1). The Dunnage and Gander zones collided during the closure of the Iapetus Ocean

in Late Ordovician to Late Silurian time (Williams et al., 1999). The Avalon zone was amalgamated with the Gander zone by accretion of the Ellsworth arc in the Early Ordovician, and the Caledonian terrane in the Late Silurian (Fyffe et al., 1999). Various basins, including (from north to south) the Late Ordovician to Early Devonian Matapedia and Fredericton basins, and the Late Silurian Mascarene basin, overlie and obscure the boundaries between the Dunnage, Gander, and Avalon zones. Subsequently, these zones are overlain in part by the Late Devonian to Early Permian Maritimes successor basin (Fig. 1).

Two major granitoid batholiths occur in southwestern New Brunswick: the Pokiok batholith (415–402 Ma; Whalen, 1993; McLeod et al., 1994), which intruded the Cambrian to Middle Ordovician strata of the Miramichi Terrane (Fyffe and Fricker, 1987) of the Gander Zone and Late Silurian strata of the Fredericton basin (Fig. 1), and the Saint George batholith (composed of five intrusive suites ranging in age from 423 to 360 Ma; Cherry, 1976; McLeod, 1990; Whalen, 1993; McLeod et al., 1994; Davis et al., 2004), which intruded the Neoproterozoic to Cambrian rocks of the peri-Gondwanan New River terrane, Ordovician strata of the St. Croix terrane, and Silurian sequences of the Mascarene basin (Fyffe and Fricker, 1987; Fyffe et al., 1999; Fig. 1). Although the batholiths are located

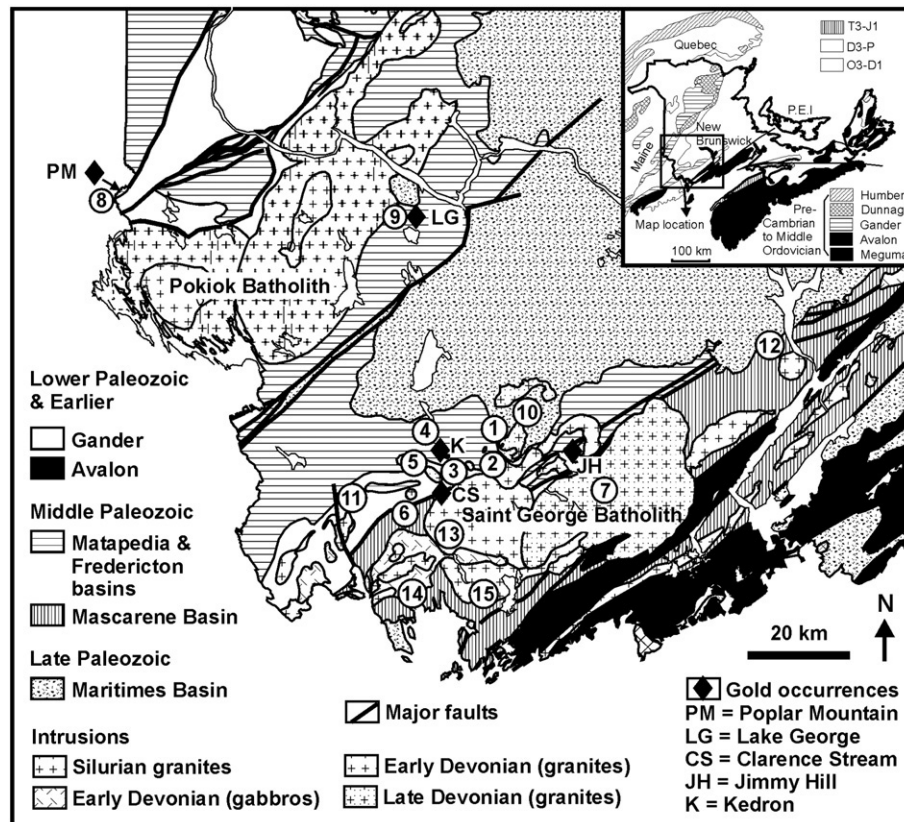


Fig. 1. Regional geological map of southwestern New Brunswick (modified from New Brunswick Department of Natural Resources and Energy, 2000; Chi, 2002), showing the location of the two granitoid series, namely, the Late Devonian granitic series (GS) granites (1 = Mount Pleasant granite suite, 2 = True Hill granite, 3 = Beech Hill granite, 4 = Kedron granite, 5 = Pleasant Ridge granite, 6 = Sorrel Ridge granite, and 7 = Mount Douglas granite; 1–6 occur on the northwest margin of the Saint George batholith, and 7 on its eastern side), and the Late Silurian to Early Devonian granodioritic to monzogranitic series (GMS) granitoids (8 = Poplar Mountain volcanic suite, 9 = Lake George granodiorite, 10 = McDougall Brook granitoid suite, 11 = Tower Hill granite, 12 = Evandale granodiorite, 13 = Magaguadavic granitoid suite, 14 = Bocabec granitoid suite, and 15 = Utopia granite; 13–15 constitute the western part of the Saint George batholith).

near the terrane boundaries, their emplacement occurred after the amalgamation of these zones.

A number of satellite intrusions occur along the northwest margin of the Saint George batholith: McDougall Brook, Mount Pleasant, True Hill, Beech Hill, Kedron, Pleasant Ridge, Sorrel Ridge, and Tower Hill, as well as the Evandale stock situated ~20 km northeast of the batholith (Fig. 1; Butt, 1976; Lentz et al., 1988; Lentz and MacAllister, 1990; McLeod, 1990; Whalen, 1993; McLeod et al., 1994; Yang et al., 2003). The Lake George granodiorite stock and the Poplar Mountain dacite dome are associated with the Pokiok batholith (Fig. 1; Seal et al., 1987; Whalen, 1993; Chi and Watters, 2002; Yang et al., 2002).

On the basis of field relationships, geochemical features, and ages, two major groups of granitoid intrusions are recognized: a Late Devonian (370–360 Ma) granitic series (GS), and a Late Silurian to Early Devonian (423–396 Ma) granodioritic to monzogranitic series (GMS). GS intrusions include the Mount Douglas granite of the Saint George batholith, and the Mount Pleasant, True Hill, Beech Hill, Kedron, Pleasant Ridge, and Sorrel Ridge granitic stocks (Butt, 1976; Lentz et al., 1988; Sinclair et al., 1988; McLeod, 1990; Taylor, 1992; Whalen et al., 1996; Yang et al., 2003). The granitic stocks intrude into Ordovician to Devonian sedimentary rocks (quartzite, slate, siltstone, and greywackes) along NE-trending fault zones northwest of the Saint George batholith (Fig. 1). Several generations of penetrative deformation and foliation (pre- to post-Acadian orogeny) have been defined in the region (Ruitenberg and McCutcheon, 1982; Castonguay et al., 2003), but deformation is rarely evident within the GS intrusions.

The GS rocks are characterized by high abundances of quartz, K-feldspar, and sodic plagioclase, with variable minor and accessory minerals (Butt, 1976; Lentz et al., 1988; Sinclair et al., 1988; McLeod, 1990; Whalen et al., 1996; Yang et al., 2003; Yang and Lentz, 2005). Although they are mostly monzogranites, some are syenogranites (sensu Le Maitre et al., 2002). Fe-rich biotite and Li-bearing muscovite are present, but amphibole is absent. Primary topaz occurs in some of the intrusions (e.g., the Mount Pleasant, Pleasant Ridge, and Kedron granites). They resemble haplogranites, and consist of texturally variable (i.e., fine- to coarse-grained, equigranular, seriate, and porphyritic) subsolvus granites because of the presence of discrete plagioclase and K-feldspar crystals. They exhibit highly evolved compositional characteristics (e.g., high silica contents) and petrologic features (e.g., porphyritic texture) indicative of high-level emplacement (Yang and Lentz, 2005). Aplite, pegmatite, miarolitic cavities, comb quartz layers and quartz veins, unidirectional solidification textures, and myrmekitic and micrographic textures commonly occur (Butt, 1976; Sinclair et al., 1988; Yang et al., 2003), suggesting water saturation during crystallization (Lentz et al., 1988; Sinclair et al., 1988; Candela, 1997; Yang and Lentz, 2005). GS rocks have relatively low values of magnetic susceptibility (e.g., Mount Pleasant granite $0\text{--}5 \times 10^{-5}$ SI, True Hill granite $0\text{--}15 \times 10^{-5}$ SI), similar to ilmenite-series granites of Ishihara (1981, 2004), consistent with low $f(\text{O}_2)$ conditions.

The GMS granitoids include the western part of the Saint George batholith (i.e., the Magaguadavic and Bocabec granitoid

suites, and the Utopia granite), the McDougall Brook granitoid suite, Tower Hill granite, and Evandale granodiorite, as well as the Lake George granodiorite and the Poplar Mountain volcanic suite related to the Pokiok batholith (Fig. 1; Butt, 1976; Cherry, 1976; McLeod, 1990; Whalen, 1993; McLeod et al., 1994; Chi and Watters, 2002). Most GMS rocks are granodioritic in terms of the IUGS classification (sensu Le Maitre et al., 2002), although a few are syenogranite, monzogranite, monzonite, quartz diorite, diorite, and gabbro (e.g., Bocabec granitoid suite). Most GMS intrusions are of Late Silurian to Early Devonian age, except the McDougall Brook granitoid suite which is Late Devonian (McCutcheon, 1990; McCutcheon et al., 1997; Yang et al., 2003). GMS rocks are less evolved than GS granites, and were emplaced at various levels (Yang and Lentz, 2005). Hybridization with the country rocks commonly occurs at their margins. Muscovite and garnet are locally present in the marginal phase of Tower Hill granite intrusion, whereas the Bocabec granitoid suite comprises various rocks from gabbro to monzogranite. Highly fractionated phases occur in the Magaguadavic granitoid suite and other intrusions (cf. Chappell and White, 1992). GMS rocks contain hornblende, variable amounts of biotite, and more abundant plagioclase than K-feldspar. They are typically subsolvus granites. Primary magnetite is present in some GMS rocks (e.g., the Magaguadavic and Bocabec granitoid suites), but absent in the others that are dominated by ilmenite (e.g., the Lake George granodiorite). Aplite and pegmatite dikes, as well as late sulfide-bearing quartz veins, are commonly associated with these intrusions.

Magnetic susceptibility values vary significantly among GMS intrusions. For example, Lake George granodiorite samples have low magnetic susceptibility values ($4 \times 10^{-5}\text{--}21 \times 10^{-5}$ SI), reflecting the low $f(\text{O}_2)$ character of the granodiorite intrusion (Seal et al., 1987; Yang and Lentz, 2005) that is comparable to ilmenite-series granites, whereas the Bocabec granitoid suite granite samples display very high magnetic susceptibility values ($864 \times 10^{-5}\text{--}4790 \times 10^{-5}$ SI), similar to magnetite-series granites of Ishihara (1981, 2004), revealing high $f(\text{O}_2)$ conditions.

3. Sampling and analytical methods

The freshest samples were collected either from drill cores, road cuts, quarries and abandoned trenches, or natural outcrops. Drill cores were available for granitoids from the Mount Pleasant, True Hill, Kedron, Pleasant Ridge, Sorrel Ridge, Tower Hill, and Lake George. The drill cores of the True Hill Granite, Kedron Granite, Pleasant Ridge Granite, Sorrel Ridge Granite, and Tower Hill Granite are stored in the New Brunswick Department of Natural Resources, Geological Survey Branch, Sussex, whereas those for the Mount Pleasant Granite and Lake George Granodiorite are stored at Mount Pleasant and Lake George mines, respectively. The other granitoid samples (e.g., Beech Hill, Mount Douglas, Poplar Mountain, McDougall Brook, Magaguadavic, Bocabec, and Utopia) were collected either from road cuts, quarries and trenches, or natural exposures using a diamond saw or a 10 lb sledgehammer. For surface samples, any weathered surfaces

Table 1
Representative geochemical analyses of granitoids in southwestern New Brunswick

Unit		MPGS	MPGS	TRHG	TRHG	BHG	BHG	KG	KG	PRG	PRG	SRG	SRG	MDG	MDG	PMVS	PMVS	PMVS	
sample		PRL95-1-1771	PRL95-1-1914	DTRH-01-121	TH80-6-120	DBH-01-109	BH135	BR84-4-129	BR84-4-147	C82-4-63	C82-5-75	C80-9-266	C80-10-353	198 ^a	219 ^a	PM-03	PM-04	GC01-140	
Rocks		GII	GIII	Bi-G	Bi-G	Bi-G	Aplite	G	G	Topaz G	Topaz G	Bi-G	Bi-G	Bi-G	Bi-G	Dacite	Dacite	DV	
<i>wt. %</i>																			
SiO ₂	ICP-OES	74.7	75.4	75.1	75.5	76.8	78.6	75.9	74.9	72.6	76.1	77.2	76.5	73.3	76.0	65.7	68.3	59.8	
TiO ₂	ICP-OES	0.03	0.02	0.10	0.04	0.14	0.02	0.02	0.02	0.02	0.02	0.09	0.10	0.31	0.16	0.47	0.55	0.56	
Al ₂ O ₃	ICP-OES	12.60	13.28	15.62	18.83	12.29	12.68	13.20	14.23	15.31	13.35	11.78	12.51	13.20	12.30	13.47	15.48	16.00	
Fe ₂ O ₃	ICP-OES	1.45	1.55	1.79	0.85	1.28	1.11	1.23	1.34	1.20	1.42	1.10	1.41	1.68	1.24	3.98	4.47	8.95	
MnO	ICP-OES	0.04	0.10	0.05	0.01	0.03	0.02	0.10	0.09	0.08	0.07	0.02	0.02	0.03	0.04	0.07	0.04	0.09	
MgO	ICP-OES	0.08	0.22	0.33	0.22	0.30	0.19	0.51	0.33	<LD	0.31	0.51	0.52	0.38	0.08	0.97	0.22	1.77	
CaO	ICP-OES	0.79	1.08	0.40	1.72	0.09	0.09	0.02	0.02	0.34	0.01	0.09	0.07	1.24	0.64	2.68	0.29	0.63	
Na ₂ O	ICP-OES	3.46	3.17	3.35	0.04	3.50	4.42	4.06	4.69	4.50	4.54	2.90	3.64	3.59	3.67	0.01	3.86	5.21	
K ₂ O	ICP-OES	4.66	4.26	4.61	0.77	4.97	1.89	3.74	5.27	4.80	4.24	5.13	5.05	5.11	4.89	3.98	2.34	1.69	
P ₂ O ₅	ICP-OES	<0.025	<0.026	0.01	0.01	<0.025	<0.026	<0.026	<0.026	<LD	<0.026	<0.025	<0.026	0.07	0.03	0.17	0.20	0.18	
LOI		0.91	1.57			0.62	0.85	0.82	0.49	0.38	0.34	0.73	0.46	0.42	0.73	6.10	3.00	2.92	
total		98.7	100.7	101.4	98.0	100.0	99.8	99.6	101.4	99.2	100.3	99.5	100.2	99.4	99.8	97.6	98.8	97.7	
ACNK		1.04	1.13	1.39	4.67	1.09	1.34	1.23	1.06	1.16	1.11	1.13	1.08	0.96	0.98	1.46	1.65	1.39	
<i>ppm (unless otherwise stated)</i>																			
S	XRF	98	85	149	2391	<LD	702	161	43	<LD	44	255	64					6207	
Cl	XRF	241	65	48	168	140	46	319	153	<4	54	87	227	<25	<25				
F														835	2100				
Rb	ICP-MS	681.5	813.7	326.4	85.5	382.7	157.1	865.8	1301.6	1410.1	1201.6	445.5	390.3	255	425	49	59	39.2	
Cs	ICP-OES	7.8	10.0	5	3	5.1	5.8	14.5	25.5	49.3	46.3	7.8	6.6	7.7	14.2	3.0	3.0	1.6	
Ba	ICP-OES	16.2	58.5	96	<10	105.4	60.7	17.6	19.5	17.3	6.0	96.5	36.0	511	63	770	440	326	
Sr	XRF	16.5	30.9	19.5	5.9	35.5	39.5	17.6	14.0	9.80	7.40	34.3	14.8	102	30			87	
Ga	XRF	33	36	20	13	20	21	34	36	33	34	18	18	16	17	17	15		
Li														60	140				
Ta	ICP-MS	22.2	24.1	7.4	15.1	6.1	3.5	23.7	21.9	20.6	15.3	8.1	4.9	1.9	4.6	<0.5	<0.5	0.59	
Nb	ICP-MS	106.5	84.4	67.7	93.6	51.6	24.8	65.1	60.3	57.6	60.0	45.9	38.3	18	35	17.0	13.0	8.4	
Hf	ICP-MS	10.2	9.4	6	3	6.5	2.4	6.9	6.8	5.9	5.4	4.6	4.2	5.8	5.6	4	6	3.0	
Zr	ICP-MS	139.4	122.1	151.3	118.4	174.6	50.8	95.4	82.2	75.1	74.7	101.1	109.9	155	123	136	161	114.6	
Y	ICP-MS	178.6	139.8	83.4	130.8	85.9	16.7	97.6	82.1	108.3	93.4	83.0	68.8	36	83	1	31	21.2	
Th	ICP-MS	49.9	40.5	51.2	59.3	50.5	7.3	18.0	19.0	27.5	24.4	41.6	43.9	28.0	46.3	8.8	10.0	7.8	
U	ICP-MS	31.5	28.7	16.7	29.6	21.1	7.1	29.4	6.0	24.0	18.7	31.9	25.0	6.9	15.0	2.6	2.6	1.3	

La	ICP-MS	41.2	41.4	47.2	24	44.0	8.2	26.0	24.6	28.9	29.6	26.9	28.4	48	56	29.0	35.0	12.2
Ce	ICP-MS	111.7	111.2	93	63	100.9	21.2	75.5	67.8	84.5	82.4	66.8	66.9	87	110	49.0	61.0	31.7
Pr	ICP-MS	13.2	13.4			11.6	2.3	9.1	8.4	10.5	10.0	7.7	7.7	-	-			4.1
Nd	ICP-MS	44.5	44.8	32	17	40.2	8.1	31.3	28.1	35.3	32.9	28.3	27.6	30.5	43.0	20.00	24.00	18.6
Sm	ICP-MS	11.9	11.5	6.1	5.2	8.8	2.3	9.3	8.3	11.5	10.3	7.3	6.5	5.09	7.56	3.50	6.50	4.6
Eu	ICP-MS	0.02	0.01	0.4	0.5	0.25	0.19	0.03	0.02	0.01	0.02	0.14	0.15	0.85	0.28	1.30	1.40	0.9
Gd	ICP-MS	12.1	10.9			8.3	2.3	8.5	8.1	10.9	9.5	7.8	6.5	-	-			4.1
Tb	ICP-MS	2.4	2.2	1.3	1.8	1.5	0.4	1.9	1.8	2.6	2.2	1.4	1.2	0.90	1.70	<0.5	0.70	0.6
Dy	ICP-MS	19.0	17.6			11.3	2.7	15.4	14.4	21.2	18.2	11.3	8.7	-	-			4.1
Ho	ICP-MS	4.1	3.8			2.4	.5	3.4	3.0	4.5	3.8	2.4	1.9	-	-			0.8
Er	ICP-MS	14.2	13.4			8.2	1.5	12.0	11.0	15.8	13.7	8.1	6.2	-	-			2.4
Tm	ICP-MS	2.6	2.5			1.4	.2	2.3	2.1	3.2	2.7	1.4	1.0	-	-			0.3
Yb	ICP-MS	18.8	18.9	9.3	20	9.5	1.5	18.0	15.9	24.6	20.2	9.8	7.0	4.26	8.80	2.10	3.20	2.2
Lu	ICP-MS	2.9	2.9	1.47	3.18	1.5	.2	2.8	2.5	3.7	3.1	1.5	1.1	0.67	1.39	0.41	0.49	0.3
Cr	INAA	<10.1	<10.3	7	<5	<9.8	<10.5	<10.4	<10.3	<10	<10.5	<10	<10.2	2	<1	25	24	<10
Ni	XRF	8	8	5	5	<2	<2	10	13	14	8	7	7	2	1			<25.9
Co	INAA	<10.1	<10.3	3	2	<9.8	<10.5	<10.4	<10.3	<10	<10.5	<10	<10.2	3.0	1.3	6	3	30
Sc	ICP-OES	2.76	2.45	3.4	4.8	2.76	3.13	3.92	4.92	11.2	10.1	3.03	2.24	3.3	4.3	16	19	28
V	ICP-OES	<10.1	<10.3	<3	<3	<9.8	<10.5	<10.4	<10.3	<10	<10.5	<10	<10.2	19	6	66	79	102
Cu	ICP-OES	12.2	24.8	24	<1	7.10	29.3	12.3	19.5	41.3	33.9	73.5	21.1	1	2			5.8
Pb	ICP-OES	21.8	23.5	41	16	23.1	25.5	23.3	25.2	32.4	24.3	22.4	19.0	12	24	2350	16	25.1
Zn	ICP-OES	134.2	88.4	44	6	28.7	66.7	64.6	64.0	22.3	15.2	19.7	35.9	18	17	118	<50	102
Bi	ICP-OES	2.8	0.7			0.5	26.9	72.1	1.5	2.9	1.6	2.6	0.5	0.5	<0.1			0.2
Cd	ICP-OES	<20.2	<20.7			<19.7	<21	<20.7	<20.7	<20.1	<20.9	<20	<20.4	<0.5	<0.5			<20.8
Sn	ICP-OES	27.6	18.8	<0.02	<0.01	8.0	18.0	20.7	16.5	9.4	5.8	5.5	4.0	1.0	1.5			1.8
W	INAA	5	11	<1	19	<1	<1	8	4	23	19	<1	3	3.3	3.3	12.00	2.00	<4
Mo	INAA	8	9	6	16	7	6	33	<1	14	<1	<1	7	<0.5	3.5	<1	<1	<5
Ag	INAA	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5					<5
As	INAA	2.7	3.5	5.7	10.1	3	24.5	9.2	11.8	5.4	4.1	4.1	3.4	1.3	6.5	5200.0	12.0	21
Se	INAA	<3	<3	<3	<3	<3	<3	<3	<3	<3	<3	<3	<3	<0.4	<0.4			<5
Sb	INAA	0.3	0.6	<0.1	0.8	0.4	0.2	0.5	0.5	1	1.1	0.5	0.1	0.5	0.8	25.0	10.0	12.3
Au (ppb)	INAA	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	5	<2	1.0	1.5	1500	6	23

Table 1 (continued)

Unit	LG	LG	MBGS	MBGS	THG	THG	EG	EG	MGS	MGS	BGS	BGS	BGS	BGS	UG	UG
sample	LG81-14-1291	LG83-2-1995	289 ^a	29B2	C80-3-45	C80-5-72	SEV-01-125	SEV-01-127	DM-01-130	DM-01-89	SDB-01-84	DB-01-104	DB-01-93	202 ^a	SU-01-129	180 ^a
Rocks	GD	GD	MBG1	MBG2	Mus-Bi-G	Mus-Bi-G	GD	GD	CG-GD	FG-G	fg-Gabbro	fg-diorite	mg-GD	G	VCG-G	VCG-G
<i>wt. %</i>																
SiO ₂	66.3	68.3	67.8	68.3	70.3	68.7	66.1	64.6	67.5	77.7	48.1	59.9	67.6	72.0	73.3	74.1
TiO ₂	0.57	0.52	0.51	0.37	0.12	0.18	0.60	0.63	0.41	0.09	1.79	1.52	0.33	0.36	0.23	0.18
Al ₂ O ₃	13.91	14.72	14.6	14.2	16.68	16.24	16.30	15.82	15.90	14.95	14.07	14.35	17.66	13.95	14.77	12.90
Fe ₂ O ₃	3.24	2.94	3.44	3.26	1.27	1.44	3.62	3.59	2.43	0.75	12.04	8.57	2.68	2.32	2.31	1.62
MnO	0.06	0.06	0.19	0.10	0.05	0.03	0.06	0.06	0.06	0.01	0.17	0.16	0.04	0.05	0.03	0.03
MgO	1.17	1.00	0.85	0.58	0.32	0.38	2.59	2.77	1.45	0.07	7.40	2.90	0.46	0.56	0.41	0.13
CaO	3.10	2.72	0.72	1.37	0.94	1.10	3.41	3.27	2.36	0.62	9.13	5.00	2.43	1.78	0.21	0.75
Na ₂ O	3.19	3.53	2.50	2.61	3.94	3.46	4.01	3.73	3.88	4.47	2.93	3.63	5.78	4.09	3.39	4.11
K ₂ O	2.76	3.66	6.84	6.23	4.52	5.20	3.59	3.78	4.27	4.27	0.89	1.90	3.08	4.19	5.24	5.00
P ₂ O ₅	0.14	0.12	0.13	0.09	0.16	0.14	0.17	0.18	0.15	<LD	0.28	0.44	0.04	0.08	0.04	0.04
LOI	3.90	1.30	1.54	2.23										0.47		0.73
total	98.4	98.8	99.1	99.3	98.3	96.9	100.4	98.5	98.4	102.9	96.8	98.4	100.1	99.9	99.9	99.6
ACNK	1.00	1.00	1.14	1.05	1.28	1.22	0.98	0.98	1.04	1.14	0.63	0.84	1.02	0.96	1.27	0.95
<i>ppm (unless otherwise stated)</i>																
S	6121	1471			100	48	136	112	45	43	875	386	41		163	
Cl					<4	<4	254	252	68	53	416	646	399	100	110	<25
F														255		1500
Rb	112	158	323	207	265.1	236.6	140.8	147.4	191.0	232.5	23.6	72.7	49.7	149	202.8	240
Cs	7	10	10	10	22	12	4	3	4	5	2	4	<1	4.6	3	5.9
Ba	360	360	1950	1200	157	333	653	701	549	<10	119	313	1051	475	119	65
Sr	223	267	160	80	60.3	101.2	387.9	382.0	265.0	33.8	257.1	205.7	185.4	90	37.7	23
Ga	16	16			23	23	21	21	22	19	20	20	27	17	23	24
Li														18		57
Ta	1.6	1.6	2.5	<0.5	3.1	<0.5	<0.5	1.5	<0.5	<0.5	<0.5	<0.5	2.8	1.2	2.2	3.3
Nb	13.3	14.9	5	5	18.1	12.6	12.4	13.1	15.8	13.1	7.6	16.8	13.9	11	35.2	36
Hf	4	5	12	13	2	2	5	5	5	3	4	8	18	5.7	5	7.4
Zr	166.5	179.4	520	480	48.2	89.3	197.4	186.1	182.1	60.7	159.1	310.2	770.8	149	215.0	158
Y	13	22	50	50	10.7	10.6	12.7	13.0	15.6	4.4	32.4	59.9	42.4	43	68.8	88
Th	13.2	14.7	14.8	15.9	6.4	9.8	18.6	16.9	15.6	24.2	2.3	10.3	9.7	20.5	19.1	22.8
U	5.8	6.2	4.6	4	19.8	9	3.2	4.1	3.4	5.7	1.6	2.9	3	5.4	4	6.3

La	31.4	30	62.1	106	10.8	21.7	36.8	41.7	28.8	23.3	13.1	35.6	28.9	30	66.3	62
Ce	52	49	142	178	21	41	64	69	55	22	30	77	58	57	134	124
Pr														6.7		-
Nd	15	15	44	75	8	17	26	19	21	12	<5	37	25	24.9	49	55.5
Sm	4.1	4.3	9.5	13.5	1.4	3.9	4.1	4.3	4.9	0.5	5	9.2	6.5	5.35	12.5	11.40
Eu	1	1	2.6	2.9	<0.2	<0.2	1.1	1.1	0.9	<0.2	1.9	2.7	5.2	0.81	0.6	0.32
Gd														5.61		-
Tb	<0.5	0.8	0.9	<0.5	<0.5	<0.5	<0.5	<0.5	0.8	<0.5	1.4	<0.5	0.8	1.00	1.6	2.40
Dy														6.53		-
Ho														1.48		-
Er														4.39		-
Tm														0.68		-
Yb	1.9	2.4	4.2	5.2	0.9	0.8	1.4	1.3	4	0.7	3.7	7.1	4.5	4.90	6.9	9.41
Lu	0.29	0.36	0.65	0.8	0.15	0.13	0.19	0.22	0.6	0.12	0.56	1	0.85	0.73	1.09	1.41
Cr	21	11	<5	<5	<5	<5	30	33	5	<5	181	18	<5	5	<5	6
Ni	11	7	5	6	<2	<2	13	13	5	<2	28	5	<2	3	<2	2
Co	9	7	4	5	<1	<1	11	9	6	1	46	18	4	4.6	2	4.1
Sc	7.9	7	9.4	11.5	3.2	2.8	7.1	7.5	6	2	46.4	23.1	6.7	6.5	13.6	14.2
V	59	50	0	0	8	<3	54	55	37	6	319	134	14	31	16	6
Cu	140	21			<1	<1	13	17	7	<1	71	18	<1	3	<1	2
Pb	23	27			27	34	14	14	22	34	6	17	11	19	15	14
Zn	39	48	352	<50	21	25	28	26	24	<2	49	60	<2	28	26	35
Bi	0.8	16.3												-		<0.1
Cd	<0.3	<0.3												<0.5		<0.5
Sn	1	<1	<0.01	<0.03	<0.01	<0.01	<0.01	<0.01	<0.01	<0.02	<0.02	<0.02	<0.02	3.5	<0.01	6.0
W	<1	<1	<1	<1	<1	<1	<1	<1	<1	<1	<1	<1	<1	1.3	<1	2.0
Mo	6	6	7	3	<1	<1	<1	<1	<1	<1	<1	<1	<1	2.5	5	<0.5
Ag	0.4	0.4	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5		>5	
As	365	3.5	22.3	8.3	2.5	2.8	2.9	3.5	1.9	21.2	4.9	3.9	4	1.0	11.9	2.5
Se	0.8	0.5	<3	<3	<3	<3	<3	<3	<3	<3	<3	<3	<3	1.8	4	<0.4
Sb	6.5	1.2	1.5	1.9	0.5	0.2	0.4	0.3	0.4	1	<0.1	0.3	0.8	0.5	0.6	0.5
Au (ppb)	2	77	<2	6	<2	<2	<2	3	5	<2	<2	<2	3	<0.3	5	2.5

Notes: <=below detection limit (LD); blank cells denote no analysis. Unit abbreviations: BGS = Bocabee granitoid suite; BHG = Beech Hill granite; EG = Evandale granodiorite; KG = Kedron granite; LG = Lake George granodiorite; MDG = Mount Douglas granite; MGB = McDougall Brook granitoid suite; MGS = Magaguadavic granitoid suite; MPGS = Mount Pleasant granite suite; PMVS = Poplar Mountain volcanic suite; PRG = Pleasant Ridge granite; SRG = Sorrel Ridge granite; THG = Tower Hill granite; TRHG = True Hill granite; UG = Utopia granite. Lithological abbreviations: Bi-G = biotite granite; CG-GD = coarse-grained granodiorite; DV = dacitic volcanoclastics; FG-diorite = fine-grained diorite; FG-G = fine-grained granite; GB = gabbro; GD = granodiorite; GRII and GIII = monzogranites from MPG; Mus-Bi-G = muscovite-biotite granite; Topaz-G = topaz-bearing granite; VCG-G = very coarse-grained granite. a -Data from Whalen (1993).

Table 2
Two groups of granitoids in southwestern New Brunswick

	Granitic series (GS)	Granodioritic to monzogranitic series (GMS)
SiO ₂	>72%	<70%
K ₂ O/Na ₂ O	>1, or <1	<1, or >1
*σ	1.5–3.4	1.3–3.4
τ	23–774	6–133
A/CNK	0.95–1.34	0.60–1.23
A/NK	>1.0	>1.0
Type	evolved I-type	I-type or reduced I-type

Notes:

$\sigma = (\text{Na}_2\text{O} + \text{K}_2\text{O})^2 / (\text{SiO}_2 - 43)$, unit in wt.% (Rittmann, 1973).

$\tau = (\text{Al}_2\text{O}_3 - \text{Na}_2\text{O}) / \text{TiO}_2$, unit in wt.% (Gottini, 1968).

A/CNK = $\text{Al}_2\text{O}_3 / (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$, unit in moles; A/NK = $\text{Al}_2\text{O}_3 / (\text{Na}_2\text{O} + \text{K}_2\text{O})$, unit in moles (Maniar and Piccoli, 1989).

See text for details.

and (or) sawn surfaces that might have metal contamination were carefully removed before crushing for determination of chemical compositions. Any visible veinlets in drill core samples were also removed prior to crushing. Approximately 1 kg of material for each sample was reduced to 0.5–2.0 cm chips in a steel jaw crusher. The chip samples were then pulverized to powder below 200 mesh with a soft-iron shatter box. Before pulverizing, the shatter box was cleaned with quartz sand and contaminated with the sample to be pulverized.

166 samples, including 80 taken from drill cores, were selected for determination for major elements and selected trace elements, including rare earth elements (REE). The methods used in this study include inductively coupled plasma-optical emission spectrometry (ICP-OES), inductively coupled plasma-mass spectrometry (ICP-MS), instrumental neutron activation analysis (INAA), and X-Ray fluorescence spectrometry (XRF).

Elements systematically analyzed with ICP-OES include SiO₂, TiO₂, Al₂O₃, Fe₂O₃(total), MnO, CaO, MgO, Na₂O, K₂O, P₂O₅, S, As, Ba, Cd, Co, Cr, Cu, Ni, Sc, Sr, V, and Zn). ICP-MS was used to determine Rb, Y, Zr, Nb, Cs, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, Pb, Th, U, Bi, and Sn at the Geological Survey of Canada–Quebec. Elements such as Au, Ag, As, Ba, Br, Ca, Co, Cr, Cs, Fe, Hf, Hg, Ir, Mo, Na, Ni, Rb, Sb, Sc, Se, Sn, Sr, Ta, Th, U, W, Zn, La, Ce, Nd, Sm, Eu, Tb, Yb, and Lu, were analyzed by INAA at Activation Labs, Ancaster, Ontario. XRF was used to measure SiO₂, TiO₂, Al₂O₃, Fe₂O₃(total), MnO, CaO, MgO, Na₂O, K₂O, P₂O₅, Rb, Ba, Sr, Ga, Nb, Zr, Y, Th, U, Ce, Cr, Ni, Sc, V, Cu, Pb, Zn, Cl, S, and As on pressed pellets (Longerich, 1995) at the Memorial University of Newfoundland, St. John's. Certified standard samples SY2, and JG1 (Govindaraju, 1994), and an internal standard RHY-1 (Lentz, 1995) were used to monitor the quality of data produced by these laboratories. Generally, the analytical results are comparable for these laboratories in terms of analytical precision. Analytical precision of major elements is better than 3% relative standard deviation (RSD), except for SiO₂ (5% RSD); and for most trace elements is better than 10% RSD. However, the analyzed values beyond 5% of the recommended values in the standards were not used. For Example, Rb value analysed by XRF in the Memorial University of Newfoundland for a standard was different by

more than 5% from the accepted value, then Rb values for the samples analysed by this laboratory were not reported. The analytical methods for each element reported in this study are indicated in Table 1.

4. Geochemical characteristics

Table 1 presents the representative analysis for each rock type from fifteen intrusions in southwestern New Brunswick (Fig. 1). Detailed analytical data are deposited in a data bank as a supplemental electronic file (e-Table 1), which are also available from the authors upon request. The principal petrochemical characteristics of the two groups of granitoids are summarized in Table 2 and Fig. 2, and are discussed below.

4.1. Major elements

Granitic series (GS) rocks are characterized by high silica (>72.2 wt.%), K₂O>Na₂O (most samples are high-K; cf. Le Maitre et al., 2002) except the Pleasant Ridge granite that has variable K₂O/Na₂O ratios, low CaO (<1.88 wt.%), low P₂O₅ (<0.13 wt.%), low TiO₂ (<0.22 wt.%), and low MgO/(MgO+Fe₂O₃(total)) ratios (Tables 1). Although differences in major elements between various intrusions are relatively constant, some typical variations are observed within individual intrusions, such as P₂O₅ and CaO decreasing with increasing SiO₂. According to the classification of Frost et al. (2001), some GS rocks are ferroan granites, whereas others are magnesian granites within a narrow silica range. Intra-intrusion FeO_(total)/(FeO_(total)+MgO) ratios typically increase with increasing SiO₂.

Compared to the GS suite, GMS series rocks have lower silica contents (<70 wt.%), except for the evolved phases within some intrusions (e.g., the Magaguadavic and Bocabec granitoid suites), variable K₂O/Na₂O ratios (<1 or >1), and higher CaO

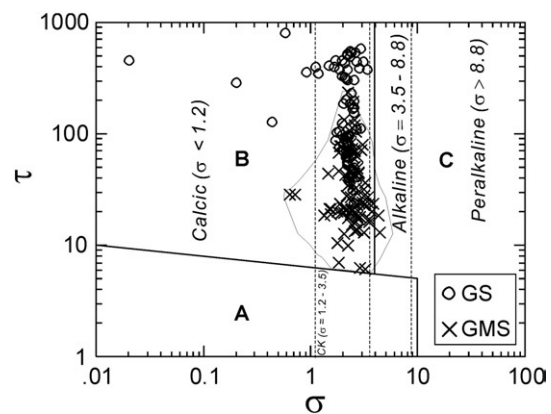


Fig. 2. σ - τ diagram for southwestern New Brunswick granitoids (modified after Rittmann, 1973). Formulations for parameters τ (Gottini, 1968) and σ are given in Table 2. Field A is for lavas formed in anorogenic regions, field B represents lavas from orogenic belts and island arcs, and field C is that of alkaline derivatives of both A and B; i.e., trachyte, phonolite, and tephrite, among which sodic types are generally linked to A, and potassic types to B. Fields of calcic series ($\sigma < 1.2$), calc-alkaline (CK; $\sigma = 1.2-3.5$), alkaline (3.5–8.8), and peralkaline (>8.8) igneous rocks are from Yang et al. (2003) and Yang (2007). Shaded area indicates data from GMS rocks. Data sources: this study, Taylor (1992), and Whalen (1993).

(<4.39 wt.%, P₂O₅ (<0.24 wt.%) and MgO/(MgO+Fe₂O_{3(total)}) ratios (Tables 1). Intra-intrusion and inter-intrusion variations in major elements are very large. Generally, MgO, CaO, TiO₂, P₂O₅, and Fe₂O_{3(total)} decrease with increasing SiO₂, whereas alkalis, A/CNK, and FeO_(total)/(FeO_(total)+MgO) ratios increase with increasing SiO₂ (Table 1). The GMS granites are mostly magnesian, although highly differentiated phases are classified as ferroan granites based on the classification of Frost et al. (2001).

Both GMS and GS rocks are metaluminous to weakly peraluminous, although most GS samples are peraluminous (Fig. 3). GS rocks have a higher Gottini index (τ ; Gottini, 1968) than GMS rocks, reflecting their more evolved nature (Fig. 2; Table 2). The differences in aluminum saturation status (A/CNK) of these two series granitoids are attributed to either their distinct sources, differentiation (magmatic evolution), or related hydrothermal systems (Clarke, 1992). High A/CNK values of some GS rocks result from extreme differentiation and related hydrothermal reaction (e.g., greisenization). Strong supracrustal assimilation could also lead to significant aluminum oversaturation (e.g., the marginal phase of the Tower Hill granite).

4.2. Large Ion Lithophile Elements (LILE)

Taken collectively, the abundances of LILE (e.g., Rb, Cs, Sr, Ba) vary systematically with differentiation (i.e., increasing SiO₂) in both GS and GMS rocks as a whole (Table 1). Rb and Cs generally increase, whereas Sr and Ba broadly decline with increasing SiO₂. These trends are observed on both intra-intrusion and inter-intrusion plots, which are generally consistent with fractional crystallization of plagioclase (cf. Ayuso and Arth, 1992; Clarke, 1992; Christiansen and Keith, 1996; Yang et al., 2003). Intriguingly, Eu/Eu* ratios are positively correlated with Sr and Ba but are negatively correlated with Rb (Fig. 4), suggesting that these elements are mainly controlled by fractionation of plagioclase and K-feldspar. The scattered data distribution (Fig. 4) suggests a role of supracrustal assimilation in genesis of these granitoids. GS rocks contain higher Rb (≤ 1588 ppm) and Cs (≤ 50 ppm),

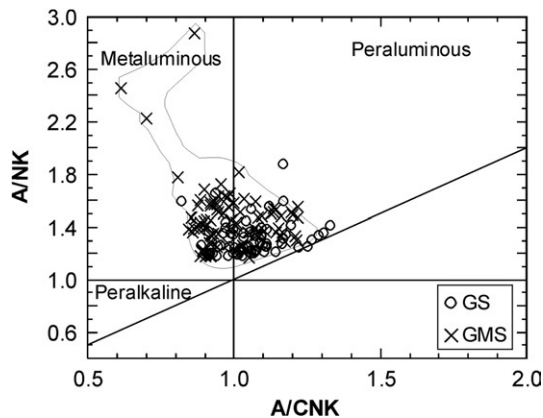


Fig. 3. Shand index plot for southwestern New Brunswick granitoids. Fields are from Maniar and Piccoli (1989); A/CNK=Al₂O₃/(CaO+Na₂O+K₂O), A/NK=Al₂O₃/(Na₂O+K₂O), in moles. Shaded area encloses GMS rocks.

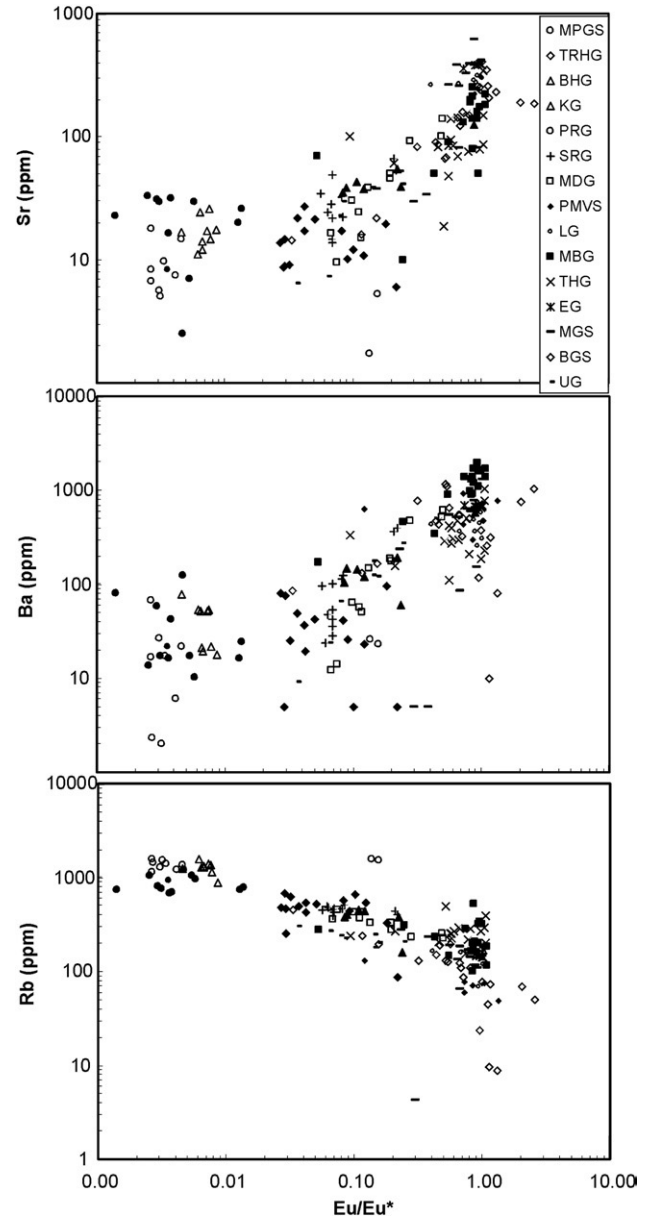


Fig. 4. Plot of Eu/Eu* vs. Sr, Ba, and Rb for southwestern New Brunswick granitoids. The unit abbreviations are listed in Table 1.

but lower Sr and Ba, compared with GMS rocks. Intra-intrusion variations in LILE are larger for GMS than for GS rocks (Table 1).

Based on criteria for fractionated granites from the Lachlan Fold Belt of SE Australia (e.g., Rb>270 ppm; Chappell and White, 1992), GS rocks are fractionated granites (Table 1), whereas GMS rocks are mostly unfractionated, except for some evolved phases (e.g., the Magaguadavic granitoid suite).

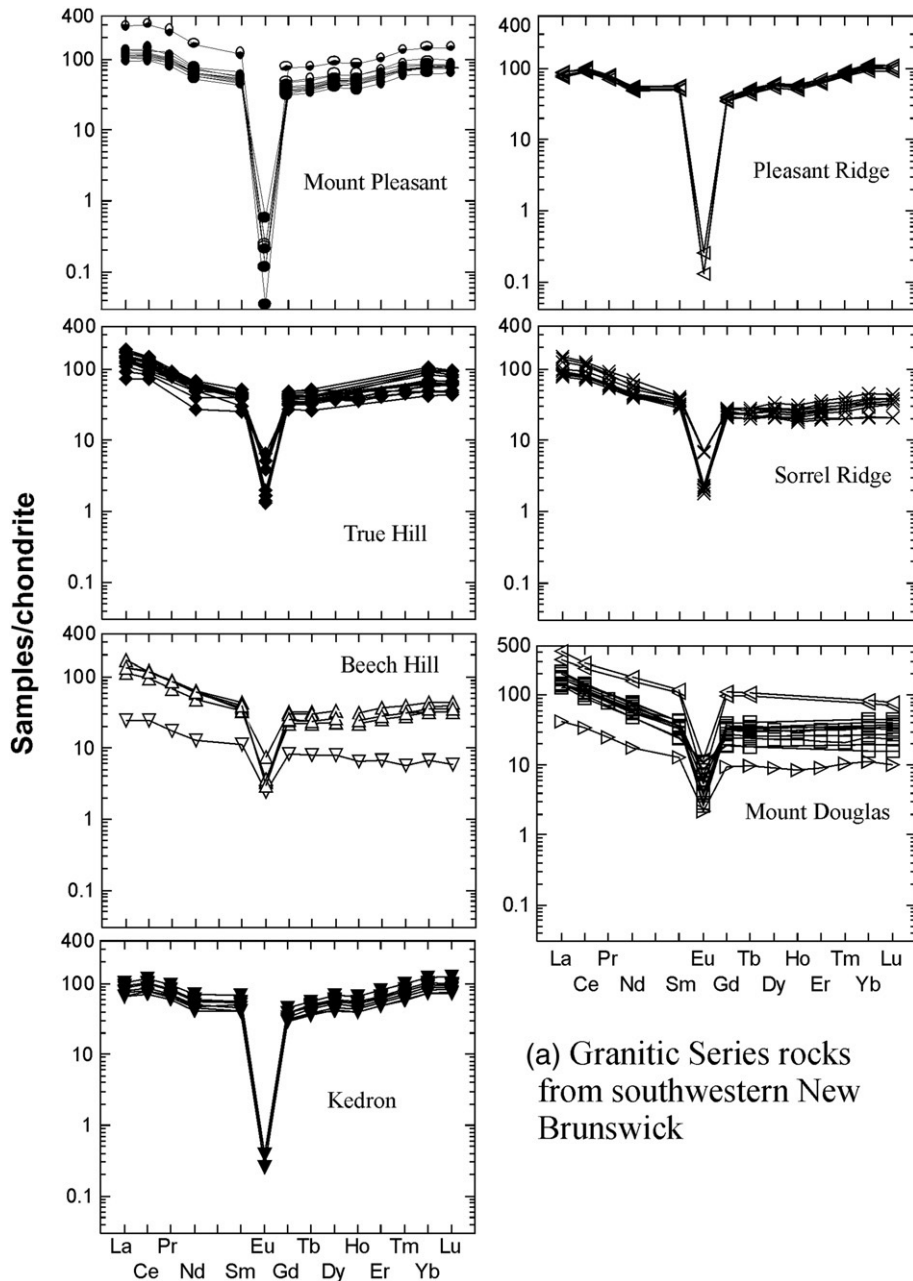
4.3. Rare Earth Elements (REE)

Chondrite-normalized REE patterns of GS are distinctive from those of GMS rocks, with pronounced negative Eu anomalies and lower (La/Yb)_N ratios (<7.3; Fig. 5). The REE patterns of different GS intrusions are similar (Fig. 5a), although

there are differences in Eu/Eu^* . The REE patterns for the granitic rocks from Mount Pleasant, Kedron, and Pleasant Ridge are indistinguishable. These granites contain primary topaz and Li-bearing mica (Sinclair et al., 1988; Taylor, 1992; Yang et al., 2003). Granitic rocks from True Hill, Beech Hill, Sorrel Ridge, and Mount Douglas are very similar, although the True Hill granite containing Fe-rich biotite shows higher HREE. Notably, the REE patterns of GS rocks also display small positive Ce anomalies (Fig. 5a). The $(La/Yb)_N$ ratios and Eu/Eu^* values decrease with fractionation. From Mount Douglas granite → Beech Hill granite, Sorrel Ridge granite, True Hill granite → Mount Pleasant granite → Pleasant Ridge granite and Kedron granite, the $(Ce/Yb)_N$ ratios decrease with increasing

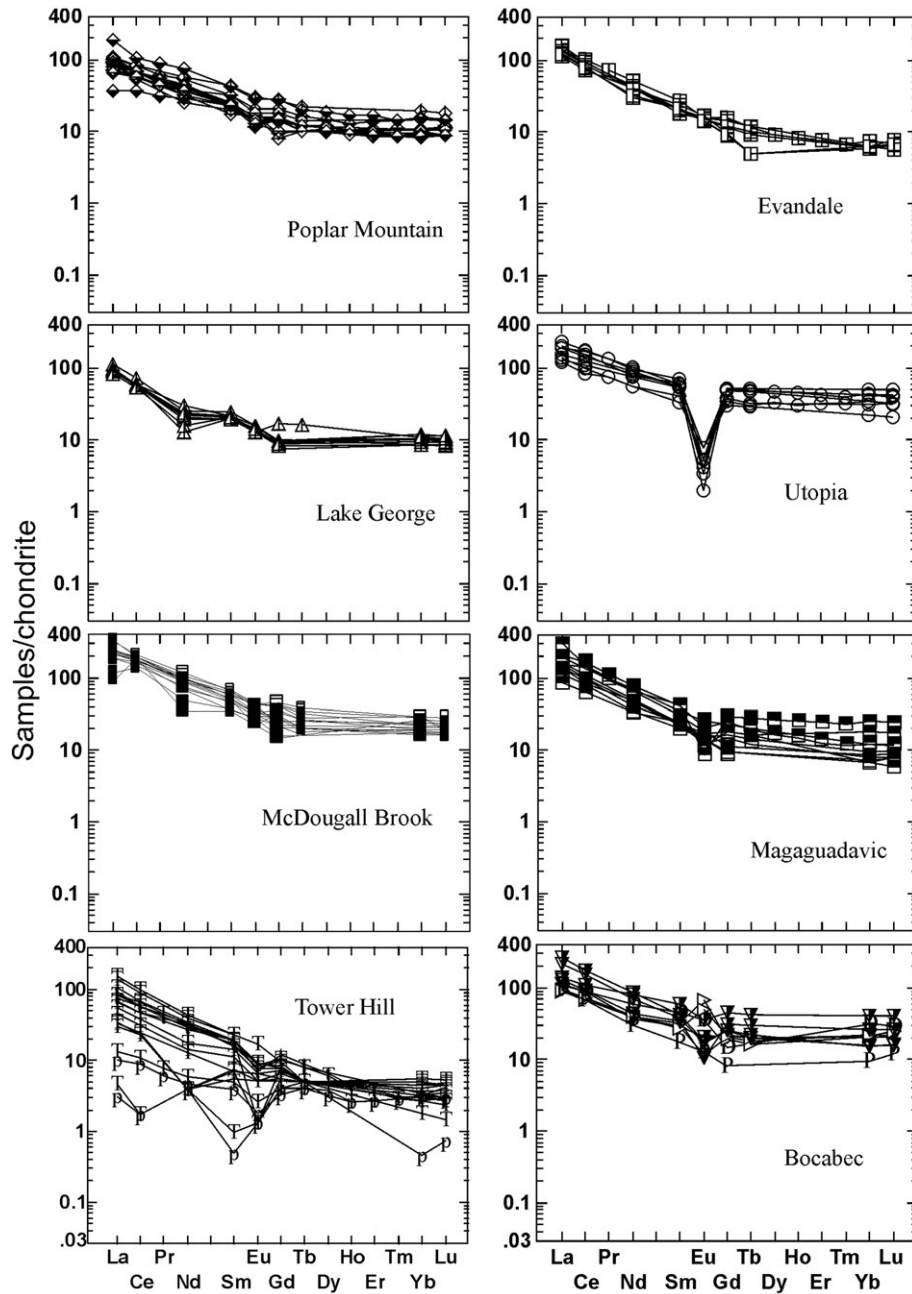
$(Yb)_N$ (Fig. 6), suggesting that fractional crystallization is a dominant process governing evolution of GS rocks. This supports the suggestion that the Pomeroy Suite granites (McLeod, 1990) are cogenetic with Mount Douglas granite (Lentz and McAllister, 1990; McLeod, 1990) via fractional crystallization.

REE patterns of GMS rocks are typically enriched in LREE with high $(La/Yb)_N$ ratios (3.5–35.8) and small or no Eu anomalies (Fig. 5b). The mafic end member even exhibits a pronounced positive Eu anomaly (e.g., gabbro in Bocabec granitoid suite). However, the Utopia granite has a similar REE pattern to those of the Mount Douglas granite, which are LREE enriched with negative Eu anomalies. In this intrusion, both $(La/$



(a) Granitic Series rocks from southwestern New Brunswick

Fig. 5. Chondrite-normalized rare earth element patterns for granitoids from southwestern New Brunswick: (a) granitic series rocks; (b) granodiorite to monzogranite series rocks. Normalizing values from Sun (1982); abbreviations: D = diorite; P = pegmatite; T = muscovite granite.



(b) Granodioritic to Monzogranitic Series rocks from southwestern New Brunswick

Fig. 5 (continued).

$Yb)_N$ and Eu/Eu^* values appear to decrease with increasing silica. Similarly, total REE abundances in evolved phases within individual intrusions (e.g., the Magaguadavic and Bocabec granitoid suites) also decrease with fractionation. In general, these trends are consistent with magmatic fractional crystallization (cf. Miller and Mittlefehldt, 1982; Christiansen and Keith, 1996).

It is notable that the REE patterns of the Tower Hill granite cross over to some extent, both in heavy and light REEs (Fig. 5b), suggesting involvement of additional magmatic processes (e.g., crustal contamination) rather than simple differ-

entiation. The La versus (La/Yb) plot shows a large variation that cannot be accounted for by simple fractional crystallization (Fig. 7a). Major elements (e.g., MgO, CaO, TiO_2 , P_2O_5) are scattered, but broadly decline with increasing SiO_2 (Fig. 7b–e); alkalis remain relatively constant (Fig. 7f).

4.4. High Field Strength Elements (HFSE)

The abundances of HFSE (i.e., Nb, Ta, Th, U) in the granitoids generally increase with fractionation, consistent with the incompatible nature of these elements; their abundances

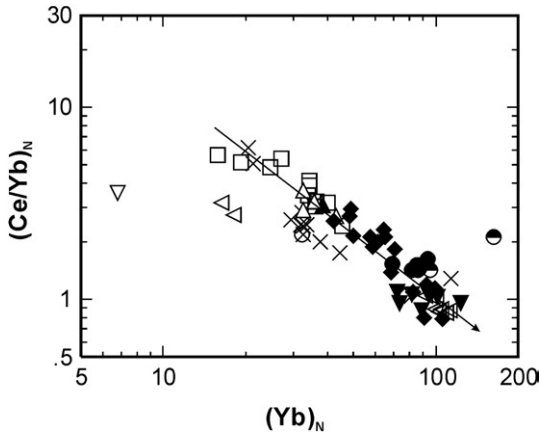


Fig. 6. Plot of chondrite-normalized $(Yb)_N$ vs. $(Ce/Yb)_N$ for southwestern New Brunswick granitic series rocks. Chondrite values from Sun (1982). Symbols as in Fig. 5a.

are higher in GS rocks than in GMS rocks (Tables 1 and 2; e-Table 1).

Nb/Ta and Zr/Hf ratios in GS are significantly lower than GMS rocks (Fig. 8). In GS rocks, Nb/Ta and Zr/Hf ratios appear to decrease systematically from Mount Douglas granite, Beech Hill granite, Sorrel Ridge granite, True Hill granite, Mount Pleasant granite, Pleasant Ridge granite, to Kedron granite, consistent with the trends of $(La/Yb)_N$ and Eu/Eu^* mentioned above. This suggests that the formation of GS granites may have been controlled by fractional crystallization. It is noted that Nb/Ta ratios in GS rocks are lower than mid-ocean-ridge basalts and ocean island basalts (average $Nb/Ta = 17.5 \pm 2$; Green, 1995) that are similar to chondrites (Sun and McDonough, 1989), and also lower than average continental crust ($Nb/Ta \approx 11$; Green, 1995). Zr/Hf ratios in GS rocks are lower than in common igneous rocks, chondrites, and crust (average = 33 to 40; Jochum et al., 1986).

Inter-intrusion variation in Nb/Ta ratios of GMS rocks is large, and intra-intrusion variation is also evident (i.e., Nb/Ta

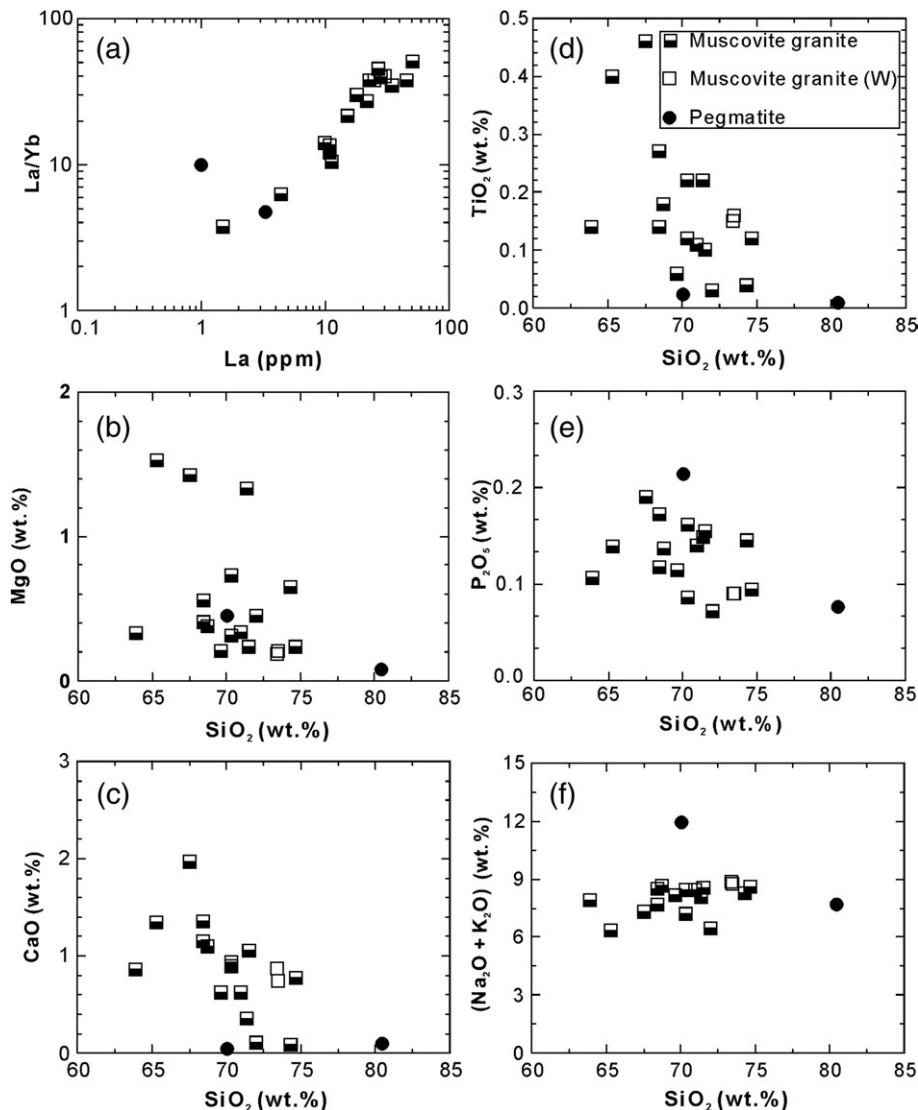


Fig. 7. Representative Harker diagrams for the Tower Hill granite: (a) La vs. (La/Yb) ; (b–f) variations of selected elements versus SiO_2 .

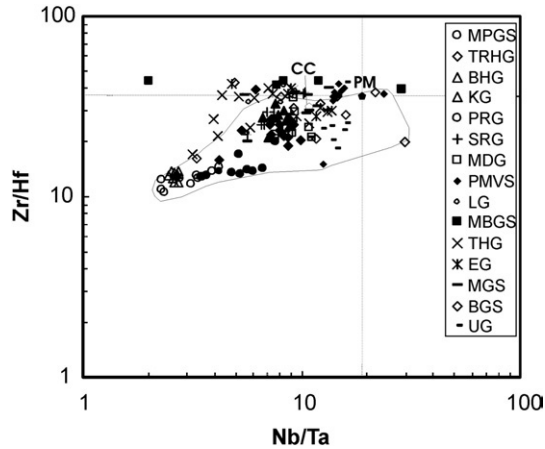


Fig. 8. Plot of Nb/Ta vs. Zr/Hf ratios for GS and GMS rocks. Shaded area represents GS rocks. PM—primitive mantle (data from McDonough and Sun, 1995), CC—continental crust (data from Taylor and McLennan, 1985). The unit abbreviations are listed in Table 1.

ratios decrease with differentiation). GMS Zr/Hf ratios are fairly constant, mostly falling into the range of common igneous rocks, chondrites, and crust (Jochum et al., 1986). General intra-intrusion consistency in Zr/Hf ratios suggests similarity in magma sources, although the cause of intra-intrusion variations requires further study. Assimilation and fractional crystallization may explain most of these variations.

The Bocabec granitoid suite is used to illustrate geochemical variations on Harker diagrams (Fig. 9). Firstly, it is noteworthy that there is a compositional gap present in SiO₂ between 52 and 58 wt.%; and secondly, Zr, Nb, TiO₂, and P₂O₅ show similar trends that decrease with differentiation above 60 wt.% SiO₂ (Fig. 9a–d), suggesting that minerals such as zircon (Zr), columbite or Fe–Ti oxides (Nb, TiO₂), and apatite (P₂O₅) do not crystallize until SiO₂ exceeds this level. In contrast, MgO and CaO display a gradual decrease (Fig. 9e,f), whereas alkalis and A/CNK correlated positively with SiO₂ (Fig. 9g,h). Correlations for other elements are poorer (Fig. 9i–n).

4.5. Transition elements

Abundances of transition elements (i.e., Cr, Ni, Co, Sc, Zn, and V) in GS rocks are generally lower than in GMS granitoids (Tables 1 and 2). These elements decrease with increasing silica within a magmatic suite (e.g., the Magaguadavic and Bocabec granitoid suites), consistent with a normal fractional crystallization trend. Intra-intrusion transition element variations are not pronounced for GS rocks, and concentrations of these elements remains consistently low, reflecting their highly evolved nature. However, large variations in transition elements occur in GMS rocks, both within and between intrusions.

Figs. 10 and 11a suggests that most GS rocks are fractionated granites, as Zn broadly decreases with increasing SiO₂, consistent with trends for felsic igneous rocks defined by Lentz (1998; references therein). Only samples from Mount Pleasant granite, two from Kedron granite, and six samples from McDougall Brook granitoid suite fall into the A-type

field; high Zn contents in these rocks may be attributed to fluid involvement (Yang et al., 2003) because Zn partitions preferentially into a fluid phase.

4.6. Volatile elements

Compared to GMS, F contents are high but Cl and S contents are relatively low in GS rocks (Table 1). Intra-intrusion S variation is significant, i.e., S is high in gabbro relative to evolved granite. However, Kedron granite samples contain high but extremely variable S contents (average = 558 ± 1172 ppm S), suggesting some S may be hydrothermally derived because S is relatively insoluble in evolved granite magmas (Poulson et al., 1991; Yang et al., 2006).

4.7. Gold contents and correlations with other trace elements

GS rocks have low Au contents (<0.2 to 7 ppb, except one sample from True Hill granite containing 21 ppb Au; Table 1; Fig. 12). Gold contents of GMS rocks are higher than GS rocks, with contents ranging from <0.2 ppb up to 77 ppb (Fig. 12). It is remarkable that U and Au do not correlate in the southwestern New Brunswick granitoids (Fig. 12), suggesting that fractional crystallization did not lead to Au enrichment in residual magmas, because U is a highly incompatible element in granitic rocks (Yang et al., 2003). However, intra-intrusion Au variation suggests that it is a compatible element during magmatic evolution, which is mainly controlled by the state of sulfur saturation in granitic melts (Yang et al., 2006).

5. Isotopic compositions

The sources of granitic series (GS) rocks can be constrained with stable and radiogenic isotope data. A few whole-rock δ¹⁸O data for GS rocks have been published previously (Taylor, 1992; Whalen, 1993; Whalen et al., 1996): Mount Pleasant granite δ¹⁸O = 8.2 to 8.6‰, Tower Hill granite = 8.0 to 8.2‰, Beech Hill granite = 8.5‰, Kedron granite = 6.8 to 7.2‰, Pleasant Ridge granite = 8.1 to 8.6‰, Sorrel Ridge granite = 8.0 to 9.4‰, and Mount Douglas granite = 7.2 to 7.7‰.

Initial (⁸⁷Sr/⁸⁶Sr)_i ratios of GS rocks are available only for the Mount Pleasant (0.7126; Kooiman et al., 1986) and Beech Hill granites (0.7124 to 0.7186; Butt, 1976; Fyffe et al., 1981); these data suggest derivation from crustal sources. Whalen et al. (1996) also presented ε_{Nd}(T) data for the Mount Pleasant (−0.2), Beech Hill (+0.1), Pleasant Ridge (+0.7), Kedron (+0.3), and Sorrel Ridge (+0.6), and they interpreted these values to reflect the presence of continental basement with different compositions beneath the boundary zone between the Gander and Avalon zones.

Available whole-rock δ¹⁸O and ε_{Nd}(T) data for GMS rocks (Whalen, 1993; Whalen et al., 1994, 1996; Yang et al., 2004) include the Lake George granodiorite (δ¹⁸O = 9.4–10.7‰, ε_{Nd}(T) = −2.4), McDougall Brook granitoid suite (7.1‰, −0.2), and Tower Hill granite (9.2–10.5‰, −0.4), suggesting that they are derived from igneous protoliths although supracrustal contamination may have been involved in petrogenesis.

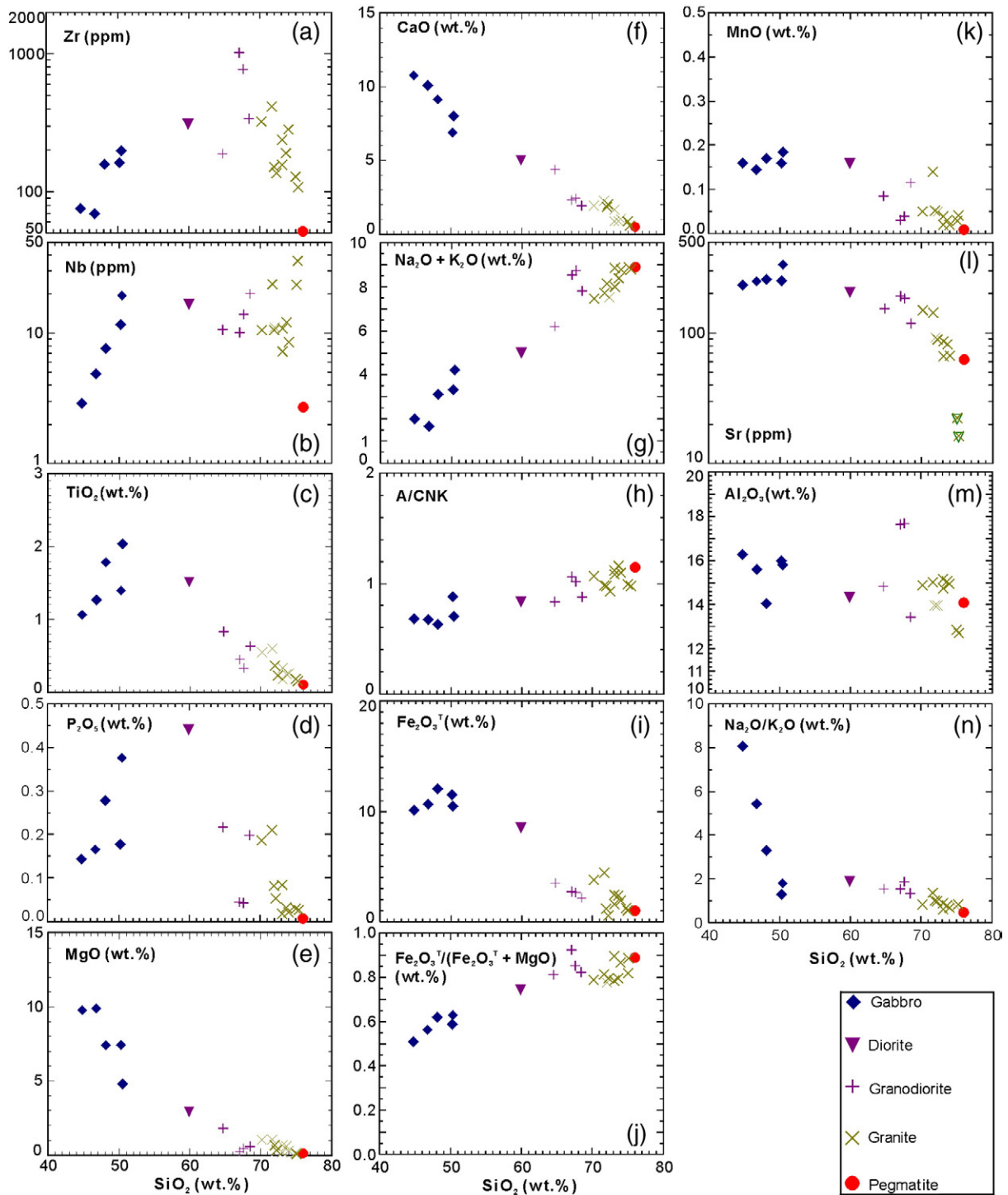


Fig. 9. Harker diagrams showing variations of selected elements and ratios with SiO_2 for the Bocabec granitoid suite. Data sources: this study, and Whalen (1993).

The Tower Hill granite initial ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratio (0.7027; Butt, 1976) requires a depleted mantle and (or) juvenile igneous source. It is noted that the Tower Hill granite displaying variations in Zr/Hf and Nb/Ta ratios (Fig. 13) contains garnet and muscovite in marginal phases, but the garnet is typically almandine-spessartine series without detectable Cr_2O_3 and is absent in the inner phases of the intrusion. These features suggest that the Tower Hill granite was contaminated by supracrustal

material such as the Cookson Group slate and muscovite-garnet schist that contain low radiogenic Sr and are characterized by high Zr/Hf and Nb/Ta ratios and high A/CNK values.

6. Magma temperatures

Magma temperatures can be estimated by using zircon, apatite, and monazite solubility models, which provide three independent

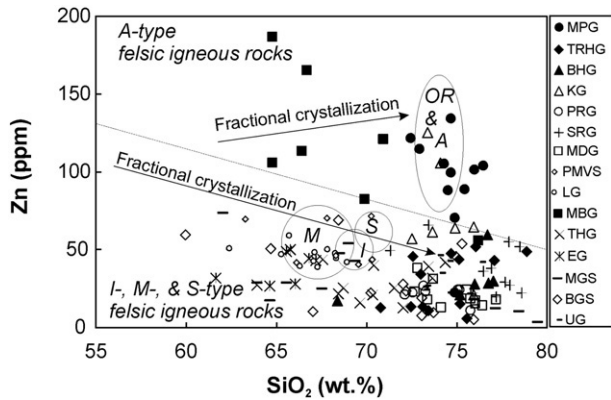


Fig. 10. SiO_2 vs. Zn discrimination diagram for various granitoid rocks (modified from Lentz, 1998). Abbreviations: M-type = depleted mantle source; I-type = igneous source (mantle-derived or crustal); S-type = sedimentary crustal source; A-type = anorogenic type (mantle or crustal); OR = oceanic ridge. The unit abbreviations are listed in Table 1.

temperatures for a granitoid sample. These techniques are applied to southwestern New Brunswick granitoids.

6.1. Zircon-saturation temperatures

Using the bulk rock Zr composition to represent the melt composition (Tables 1) in the model of Watson and Harrison (1983), zircon-saturation temperatures (T_{Zircon}) were calculated for southwestern New Brunswick granitic rocks (Table 3). T_{Zircon} of Mount Douglas granite (745 to 812 °C, average = 770 °C) are the highest in GS rocks (Table 3), but lower than those of GMS rocks (ranging from 745 to 986 °C, average 831 °C) with exceptions being the evolved Tower Hill granite and pegmatites that display low T_{Zircon} value (Table 3).

6.2. Apatite-saturation temperatures

Assuming the bulk rock composition (P_2O_5 and SiO_2) to represent the melt composition (Tables 1) and the apatite solubility model of Harrison and Watson (1984), apatite-saturation temperatures (T_{Apatite}) were calculated (Table 3). The calculated T_{Apatite} values appear higher than the T_{Zircon} values (Table 3). This may be attributed to the presence of some other P-bearing phases (e.g., monazite) that host a proportion of P. Hence, the calculated T_{Apatite} values may represent maximum apatite-saturation temperatures. Notably, T_{Apatite} values decrease from Mount Douglas granite (average 863 °C) → Sorrel Ridge granite (809 °C), Beech Hill granite (806 °C) and aplite (807 °C) → Kedron granite (773 °C), Mount Pleasant granite (767 °C) → Pleasant Ridge granite (739 °C) and True Hill granite (728 °C), consistent with magma temperature variation during fractional crystallization. Again, GS rock T_{Apatite} values are lower than GMS rocks (Table 3), consistent with the results of T_{Zircon} .

6.3. Monazite-saturation temperatures

Using the bulk rock REE composition ($\sum \text{REE}_i = \text{La} + \text{Ce} + \text{Pr} + \text{Nd} + \text{Sm} + \text{Gd}$) to represent the melt composition (Tables 1) and

the equation of Montel (1993), monazite-saturation temperatures (T_{Monazite}) were calculated for southwestern New Brunswick granitoids (Table 3). Considering that both GS and GMS granitoids are subsolvus granites and display textural evidence of water saturation during crystallization, reasonable water contents (5 wt.% H_2O) were assumed for each intrusion during calculation procedures. The T_{Monazite} values of some samples are higher than T_{Zircon} and T_{Apatite} values (Table 3), suggesting the presence of other REE-bearing phases.

It is noteworthy that GS rock T_{Monazite} values (Table 3) are higher than those (730 ± 50 °C) estimated on the basis of Q–Ab–Or– H_2O phase diagrams (Butt, 1976; Cherry, 1976; Lentz et al., 1988; Lentz and McAllister, 1990; McLeod, 1990). Because temperatures estimated from phase diagrams usually represent eutectic temperatures (close to solidus) for multiple component silicate melts (Rollinson, 1993), they may reflect either final crystallization temperatures for the magmatic system or initial partial melting conditions of the system. Monazite-saturation temperatures may represent conditions under which monazite remains in equilibrium with melts (Montel, 1993). Two interpretations are possible for monazite-saturation temperatures: (1) temperatures are those at which melts equilibrated with monazite in their source regions; (2) monazite is a liquidus phase, providing liquidus temperatures. Petrographic evidence shows that monazite commonly crystallizes as an early phase in granitic rocks, and therefore T_{Monazite} most likely reflects liquidus temperatures.

7. Geochemical constraints on tectonic regimes

Fig. 2 shows that almost all granitoid samples and previously published data plot in field B (lavas from orogenic belts and island arcs), except for a few in field C (e.g., one dacitic volcanoclastic sample from Poplar Mountain volcanic suite) because of hydrothermal alteration. This suggests that both GS and GMS granitoids are orogenic, rather than anorogenic. It is noteworthy that the biotite in both GS and GMS rocks (Yang and Lentz, 2005) are mostly located in calc-alkaline fields based on the discrimination diagram of Abdel-Rahman (1994), consistent with the bulk-rock major element data (Fig. 2).

On granitoid tectonomagmatic discrimination diagrams (Fig. 14; Pearce et al., 1984), GS rocks mainly plot in the within-plate granite field, whereas GMS rocks principally fall into the volcanic-arc granite field. The spread of data from the syn-collision granite field into the within-plate granite field for some intrusions probably reflects fractionation and supracrustal contamination (Fig. 14).

Gorton and Schandl (2000) employed HFSE to discriminate felsic to intermediate volcanic rocks from oceanic arcs, active continental margins, and within-plate volcanic zones (Fig. 15). On their Yb versus Th/Ta diagram, GS rocks plot mainly in the within-plate volcanic zone field, although some (e.g., Sorrel Ridge and Beech Hill granites) fall into the field of active continental margins (Fig. 15a). Collectively, from the Mount Douglas and Sorrel Ridge granites, through the True Hill, to the Kedron and Mount Pleasant granites, a differentiation trend is well defined, consistent with other trace element variations

described above (Figs. 4, 10, and 13). In contrast, GMS rocks mostly plot in the fields of oceanic arcs and active continental margins, and some evolved phases plot in the within-plate volcanic zone field (Fig. 15b).

8. Discussion

8.1. Petrogenesis of GS rocks

Multiple lines of geochemical evidence suggest that the Devonian GS granites are highly fractionated I-types and could be linked via fractional crystallization to the Mount Douglas granite of the Saint George Batholith (Figs. 2, 3, 5). This group

of granites display well-developed fractionation trends (Figs. 5, 13, 14); their $\delta^{18}\text{O}$ values fall into the range of normal granitoids (6 to 10‰; see Taylor, 1978), consistent with I-type granites (Chappell and White, 1992). It is noted that P_2O_5 , Zn, and A/CNK values decrease with differentiation, in keeping with their calc-alkaline, metaluminous to weakly peraluminous nature. Furthermore, the calculated zircon-, monazite-, and apatite-saturation temperatures of GS rocks with low Nb/Ta and Zr/Hf ratios are lower than GMS rocks (Table 3). This requires that GS rocks are sourced from crustal quartzofeldspathic rocks (e.g., granitoid rocks, or equivalent igneous rocks; cf. Creaser et al., 1991) and experienced extensive fractionation, and they cannot be derived from residual sources that experienced the

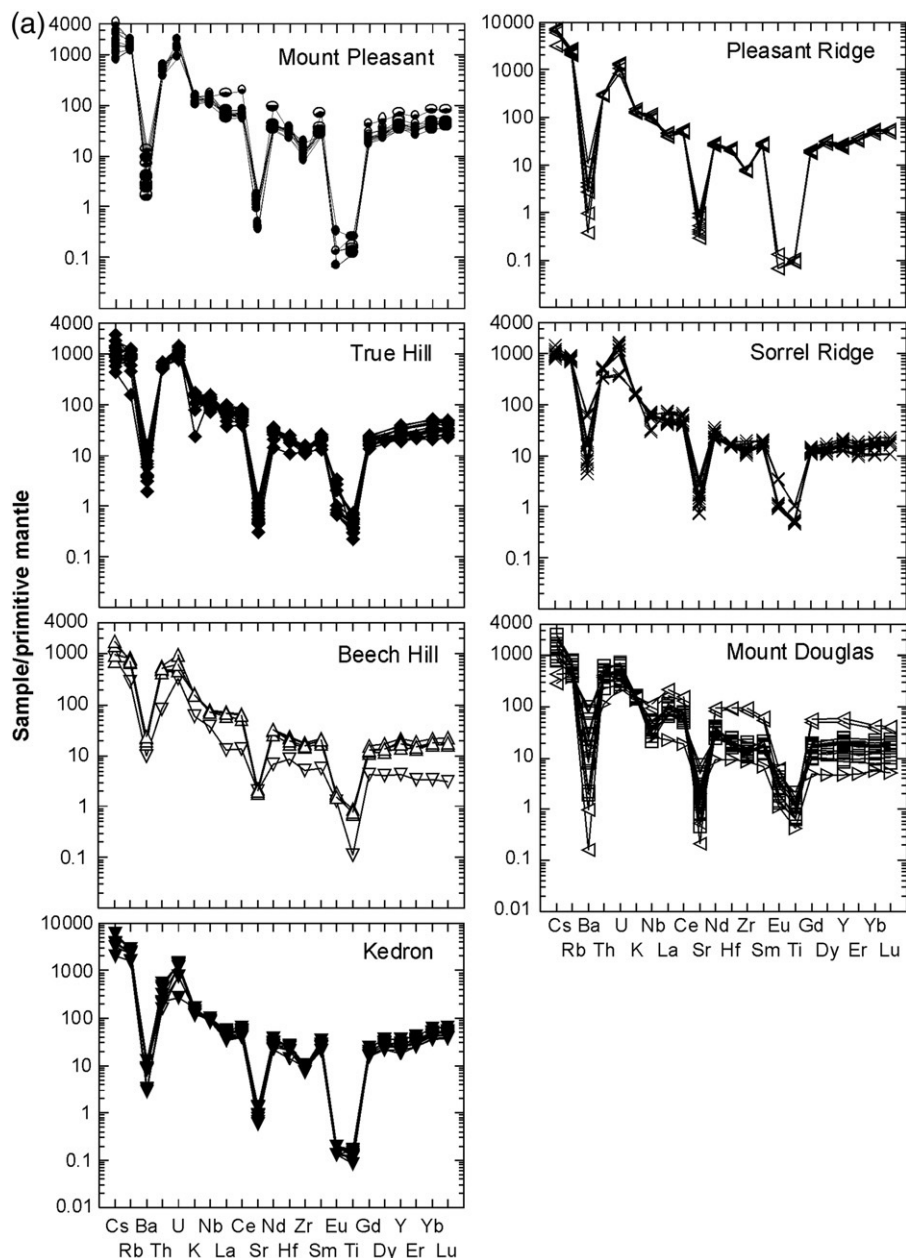


Fig. 11. Primitive mantle-normalized spider diagrams for southwestern New Brunswick granitoids: (a) granitic series (GS) rocks; (b) granodiorite to monzogranite series (GMS) rocks. Primitive mantle values from Sun and McDonough (1989).

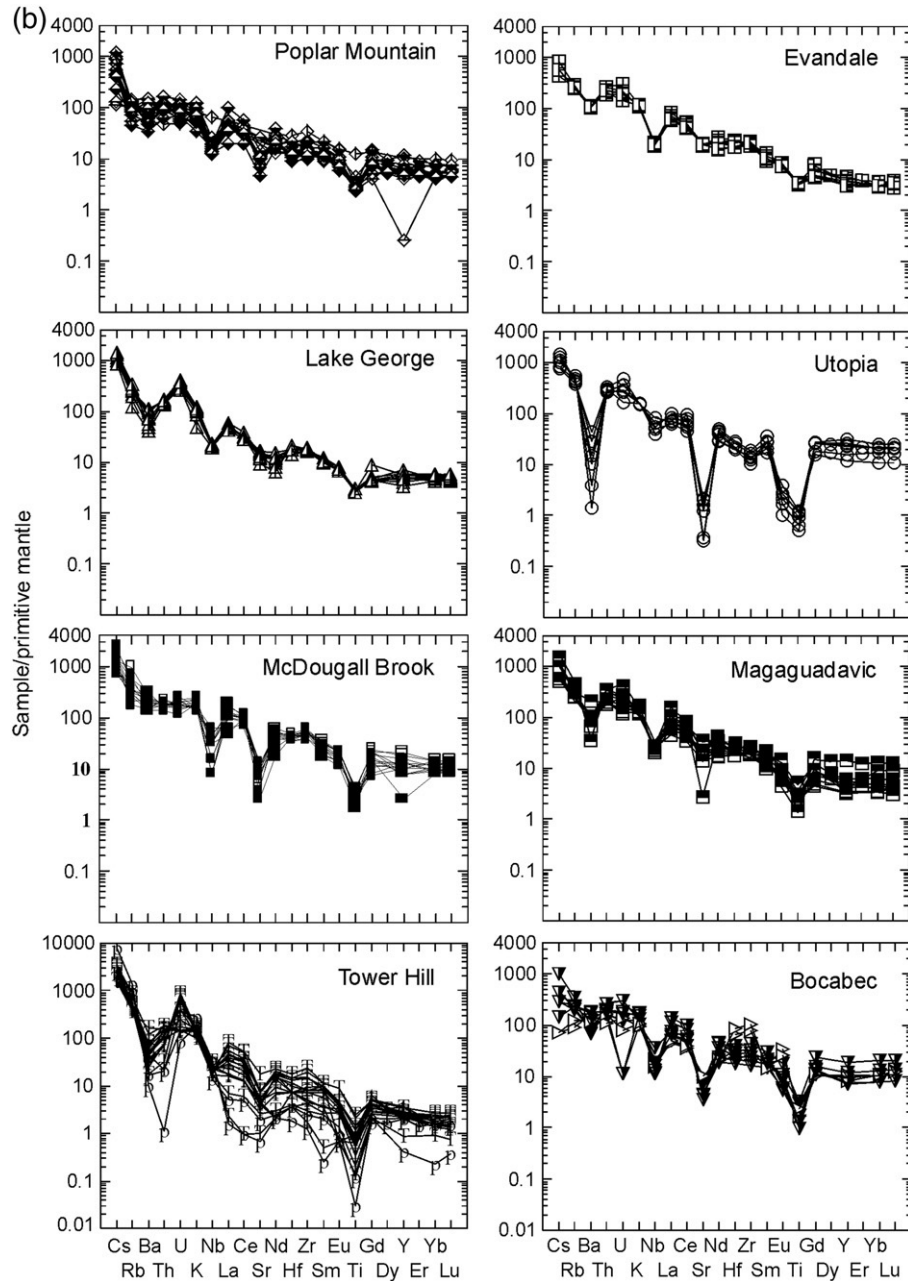


Fig. 11 (continued).

prior extraction of granite magmas, a commonly used model for A-type granites based on the residual source model (cf. Collins et al., 1982; White and Chappell, 1983; Clements et al., 1986; Whalen et al., 1987; Eby, 1990, 1992; King et al., 1997; Lentz, 1998).

Published radiogenic isotope data for the GS granites, although limited, are worthy of further evaluation. Their high initial ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios (0.7126–0.7186) and $\epsilon_{\text{Nd}}(\text{T})$ values (–0.2~+0.7) (Butt, 1976; Fyffe et al., 1981; Kooiman et al., 1986; Whalen et al., 1996) appear to be controversial in terms of potential magma sources. The Sr and O isotopic data require that GS granites be sourced from old crustal materials (igneous), whereas Nd isotopic data need a juvenile crustal source or a

slightly depleted mantle or a crustal source similar to GS rocks themselves. All geochemical features (i.e., high silica and incompatible elements, low Sr, Ba and Ti, flat REE patterns with apparent negative Eu anomalies; Table 1, Figs. 4 and 11) support crustal sources (igneous) for the highly evolved GS granites. Therefore, the GS granites are interpreted to derive from old crustal sources with igneous quartzofeldspathic composition and relatively flat REE patterns.

8.2. Petrogenesis of GMS rocks

The GMS rocks display calc-alkaline, and metaluminous to slightly peraluminous features (Figs. 2 and 3), whose variations

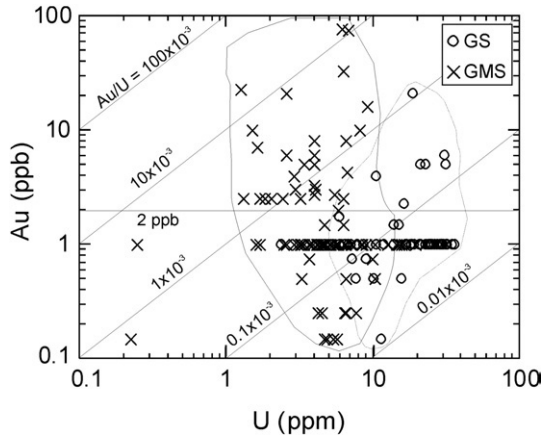


Fig. 12. U vs. Au diagram for southwestern Brunswick granitoid rocks. Au data below 2 ppb are plotted as 1 ppb; those samples with Au below 1 ppb are cited from Whalen (1993). Shaded area denotes GMS rocks.

in major and trace elements as well as trace element ratios are compatible with those of I-type granites elsewhere (Chappell and White, 1974, 1992; White and Chappell, 1983; Ayuso and Arth, 1992; Whalen, 1993; King et al., 1997). The presence of hornblende in some of the GMS intrusions (e.g., Lake George granodiorite, Magaguadavic granitoid suite) and the dominance of plagioclase in feldspars reflect their geochemical characteristics (Yang and Lentz, 2005). Bulk $\delta^{18}\text{O}$ values and $\epsilon_{\text{Nd}}(\text{T})$ data (Whalen, 1993) suggest that supracrustal contamination played a role in their origin. A typical example is the Tower Hill granite which represents a contaminated I-type granite (cf. Ague and Brimhall, 1988).

In some GMS intrusions (e.g., Bocabec and Magaguadavic granitoid suites), minor gabbroic phases are present, which may be important with respect to their origin (McLeod, 1990). Field relationships indicate that gabbros are older than granite intrusions in the Bocabec granitoid suite. The gabbros may be derived from depleted mantle ($\epsilon_{\text{Nd}}(\text{T})$ values +1.1~+5.3; Whalen, 1993; Thorne and Lentz, 2001). Hybridization at the contact zone between gabbro and granite leads to the formation of hybrid diorite to granodiorite (Fyffe et al., 1981). There is a compositional gap between gabbros and granites from the Bocabec granitoid suite (Fig. 9), and compositional trends for these two groups of rocks differ, suggesting distinct non-cogenetic magmas. Gabbro data show increases in ($\text{Na}_2\text{O} + \text{K}_2\text{O}$), TiO_2 , P_2O_5 , Zr, and Nb, and decreases in CaO, MgO, and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ with fractionation. In comparison, granitoid data are more scattered, except for TiO_2 and P_2O_5 that decrease with increasing silica. Scattered Nb and Zr distributions in granitoids suggest a role of supracrustal assimilation. Fig. 11b exhibits a pronounced negative Nb anomaly, but lesser Ti, Ba, and Sr anomalies, features suggesting arc-like source rocks. Furthermore, with fractionation, the negative Nb anomaly becomes smaller, but the negative Ti, Ba, and Sr anomalies become more pronounced.

McLeod (1990) emphasized that underplated basaltic magmas induced partial melting of thickened crust as terranes were juxtaposed. His model could explain GMS genesis because these rocks were emplaced after Gander and Dunnage

zone assembly, and also that of the Gander and Avalon zones (i.e., magmatism took place in a within-plate environment). However, as stated in the previous section, GMS intrusions are interpreted to be the products of magmatism during the waning stage of the Acadian Orogeny, rather than being post-orogenic. Our model differs from that of McLeod (1990) in that younger GS intrusions do not represent products of granulitic source remelting at higher temperatures. Magma temperatures of early Magaguadavic granitoid suite phases are higher than those of younger Mount Douglas granite in the Saint George batholith (Table 3), ruling out the possibility of Mount Douglas granite being derived by partial melting of residual source.

In summary, GMS rocks represent I-type granites derived from lower crustal partial melting, which was triggered by underplated basaltic magmas. Inter-intrusion geochemical similarities generally reflect common magmatic processes during petrogenesis. Intra-intrusion differentiation produced evolved phases. Supracrustal assimilation coupled with fractional crystallization can mostly explain the petrogenesis of GMS intrusions.

8.3. Implications for intrusion-related gold system

Source controls are essential for intrusion-related gold systems (McCoy et al., 1997; Thompson et al., 1999; Thompson

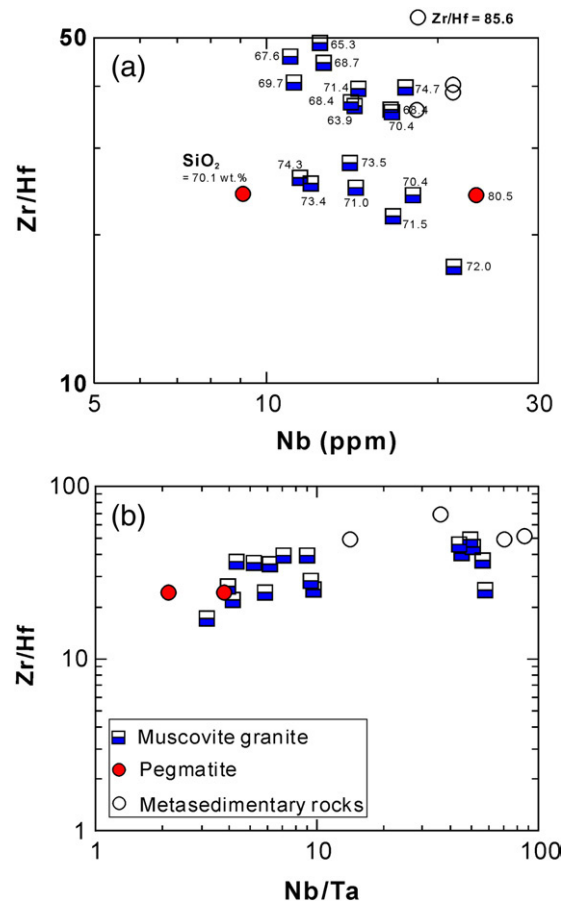


Fig. 13. Supracrustal contamination signatures recorded in the Tower Hill granite intrusion: (a) Nb vs. Zr/Hf; (b) Nb/Ta vs. Zr/Hf. Data from this study and Whalen (1993).

Table 3
Calculated results of zircon-, apatite-, and monazite-saturation temperatures (°C) for southwestern New Brunswick granitoids

Intrusions	Rocks ^a	N	T _{Zircon}	T _{Apatite}	T _{Monazite}
<i>GS rocks</i>					
Mount Pleasant Granite Suite (MPGS)					
GRII	GRII	5	747–767 (av. 758)	755–791 (av. 767)	777–869 (av. 813)
GRIII	GRIII	7	725–789 (751)	711–782 (766)	759–807 (782)
Average	GRII+GRIII	12	754	767	795
True Hill Granite (TRHG)	Bi-G	17	716–768 (750)	673–796 (728)	774–937 (816)
Beech Hill Granite (BH)					
Bi-G	Bi-G	6	745–777 (767)	783–908 (806)	730–782 (766)
Aplite	Aplite	1	672	807	700
Kedron Granite (KG)	Topaz-G	10	680–728 (714)	753–851 (773)	740–886 (785)
Pleasant Ridge Granite (PRG)	Topaz-G	10	702–711 (706)	623–817 (739)	727–776 (760)
Sorrel Ridge Granite (SRG)	Bi-G	12	730–785 (752)	791–856 (809)	720–790 (755)
Mount Douglas Granite (MDG)	Bi-G	11	745–812 (770)	817–910 (863)	745–776 (761)
<i>GMS rocks</i>					
Poplar Mountain Volcanic Suite (PMVS)					
Dacite	Dacite	10	745–808 (776)	875–965 (937)	717–853 (747)
Lake George Granodiorite (LG)					
GD	GD	11	801–826 (810)	848–920 (899)	706–742 (720)
McDougall Brook Granitoid Suite (MBGS)					
MBG1	MBG1	11	769–900 (864)	833–963 (901)	765–866 (812)
MBG2	MBG2	7	868–933 (901)	875–958 (925)	766–842 (806)
Tower Hill Granitoid (THG)					
Mus-Bi-G	Mus-Bi-G	17	640–816 (712)	845–949 (928)	629–752 (732)
Pegmatite	Pegmatite	2	521–634 (578)	974–987 (981)	599
Evandale Granodiorite (EG)					
GD	GD	8	824–854 (833)	887–927 (910)	717– (729)
Magaguadavic Granitoid Suite (MGS)					
CG-GD	CG-GD	10	807–867 (821)	912–964 (929)	707–784 (751)
FG-G	FG-G	5	668–784 (715)	758–885 (798)	629–767 (706)
Aplite	Aplite	1	746	801	688
Bocabec Granitoid suite (BGS)					
FG-diorite	FG-diorite	1	924	967	696
GD	GD	4	853–986 (932)	798–963 (844)	690–745 (715)
Bi-G	Bi-G	10	746–887 (803)	830–1000 (878)	712–822 (763)
Pegmatite	Pegmatite	1	683	754	748
Utopia Granite (UG)					
VCG-G	VCG-G	8	757–797 (777)	821–901 (860)	742–840 (774)

^a Rock type abbreviations are the same as Table 1.

and Newberry, 2000; Lang and Baker, 2001; Baker, 2002; Groves et al., 2003). All gold-related granitoids are I-type, i.e., derived from igneous sources. From this respect, GS and GMS intrusions in this region are prospective for generating gold mineralization, as they are either I-type or highly fractionated I-type granites.

From gold contents in the granitoids it appears that less differentiated phases in a granitoid suite have a greater potential for generating gold deposits (Yang et al., 2003). However, some fractionated phases, such as aplite-pegmatite dikes from the Clarence Stream Gold deposit, contain significant gold (Thorne et al., 2002). LA-ICPMS analysis indicates that gold in pyrite from an aplite dike offshoot from the Beech Hill granite intrusion contains as high as 21 ppm Au (Yang et al., 2006). McCoy et al. (1997) also observed that Au is enriched in aplite-pegmatite systems from Alaska. High Au in pegmatite and aplite dikes may be ascribed to its high solubility in exsolving magmatic brines (Frank et al., 2002) involved in dike formation.

Magmatic processes play a crucial role in the origin of intrusion-related gold systems. This study shows that GS rocks may be linked by fractional crystallization, whereas GMS rocks are formed through partial melting of lower crustal rocks coupled with AFC processes. Higher GMS magma temperatures (Table 3)

may have enhanced supracrustal assimilation and fractional crystallization processes. Thus, mineralizing components (e.g., Au, Mo, and S) present in assimilated rocks particularly black shale (or slate) of Lower Cambrian to Lower Ordovician age (McLeod, 1990; Fyffe and Pickerill, 1993) may play a role in mineralization related to GMS intrusions (McLeod, 1990; Lentz, 1998). Besides geochemical constraints, a preliminary study of whole-rock and sulfide S isotopes confirmed that, although much of the sulfur was magmatically derived, a part of it was likely scavenged from country rocks (Yang and Lentz, 2004). The association of gold mineralization with CO₂-enriched fluids that are generally exsolved at an early stage of magma evolution, such as observed in Lake George, Poplar Mountain, and Clarence Stream (Chi, 2002; Yang et al., 2004), suggests that early-stage, less-fractionated magmas are favorable for gold mineralization. However, as some greisenized granites contain significant gold (e.g., 1 g/t at Kedron, 0.1 g/t at True Hill), GS rocks could also be prospective for gold mineralization. This may reflect the effect of hydrothermal systems on gold enrichment, which is related to the redox conditions and the state of sulfur saturation in magmas (Yang et al., 2006).

Previous studies of intrusion-related gold systems empirically recognized that reduced conditions are favorable for gold

mineralization (McCoy et al., 1997; Thompson et al., 1999; Thompson and Newberry, 2000; Lang and Baker, 2001; Baker, 2002; Groves et al., 2003; Hart et al., 2004), although weakly oxidized plutons are also associated with gold as shown by Blevin (2004) and Mustard (2004). However, minimal research has evaluated why low $f(O_2)$ conditions are favorable for Au mineralization. Possibly reduced I-type intrusions and peripheral wallrocks enhance reduction of auriferous fluids and promote gold precipitation, for gold mostly exists as 1+ valence state in ore-fluids, regardless of the complexing mode (Gammons and Williams-Jones, 1997; Frank et al., 2002). This leads to decomposition of gold complexes and hence gold deposition, with ore-fluids likely being derived from progressive cooling of magma at depth. Remobilization of gold (and copper) in early magmatic sulfide minerals by these fluids (cf Keith et al., 1997) could be an important mechanism of mineralization if concentrated in shear zones or intrusion-related fracture systems.

GS rocks have relatively low values of magnetic susceptibility, consistent with low redox conditions. This promotes sulfide saturation in granitic magmas or facilitates sulfide (bearing gold) precipitates from ore-fluids (e.g., Yang et al., 2006). However, there are significant variations in magnetic

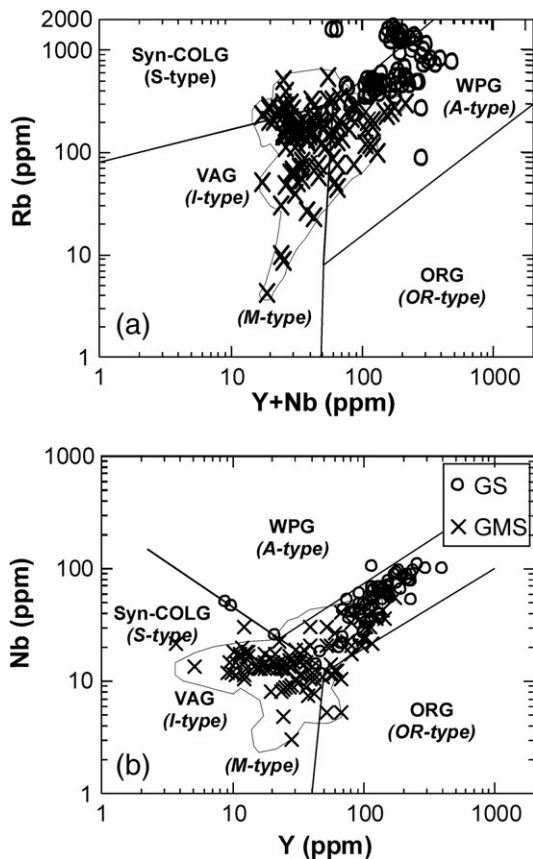


Fig. 14. Tectonomagmatic discrimination diagrams for southwestern New Brunswick granitoids (after Pearce et al., 1984; modified by Christiansen and Keith, 1996). (a) $(Y+Nb)$ vs. Rb ; (b) Y vs. Nb . ORG: ocean-ridge granitoids; Syn-COLG: syncollision granitoids; VAG: volcanic-arc granitoids; WPG: within-plate granitoids. Shaded area encloses GMS rocks.

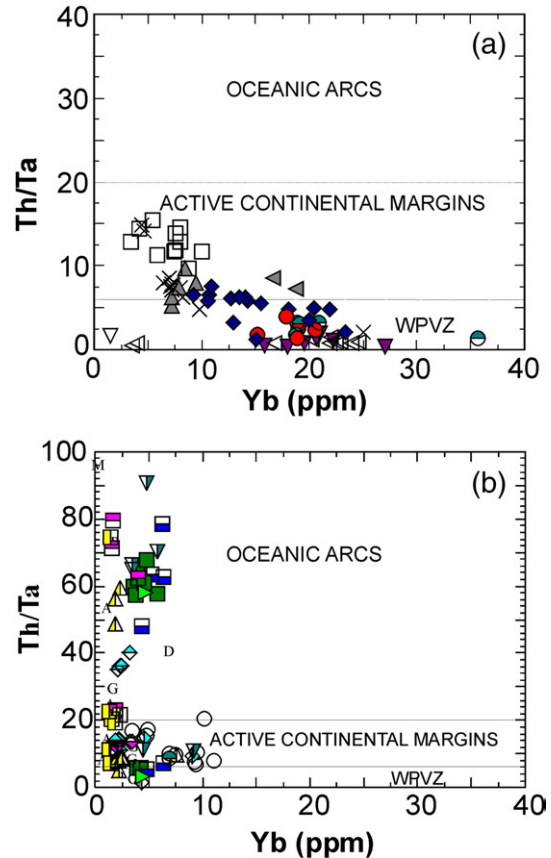


Fig. 15. Yb vs. Th/Ta discrimination diagram for southwestern New Brunswick granitoids: (a) GS rocks; (b) GMS rocks. The boundaries for felsic and intermediate rocks from oceanic arcs, active continental margins, and within-plate volcanic zones (WPVZ) are from Gorton and Schandl (2000). Symbols as in Fig. 5.

susceptibility values among GMS intrusions. For example, Lake George granodiorite samples displays reduced features as shown by low magnetic susceptibility and the presence of CH_4 -bearing fluid inclusions trapped in quartz phenocrysts (Yang et al., 2004), whereas the Bocabec granitoid suite granite samples exhibit oxidized features by very high magnetic susceptibility. The reduced nature of the Lake George granodiorite may have been caused by interaction with the Silurian Kingsclear Group that contains reduced carbon and graphitic materials, whereas the Bocabec granitoid suite retained its primary high redox condition because it intruded a gabbro body that contains appreciable magnetite. This difference in redox conditions may explain why there is gold mineralization associated with the Lake George granodiorite, while the Bocabec intrusion is barren.

9. Conclusions

In southwestern New Brunswick, two broad groups of granitoids are distinguished based on geochemical characteristics: the Late Silurian to Early Devonian Granodiorite to Monzogranite series (GMS), and the Late Devonian granite Series (GS). Emplacement of both GMS and GS intrusions took

place in an overall within-plate setting, subsequent to terrane accretion associated with the Acadian Orogeny.

GMS rocks are calc-alkaline, metaluminous to weakly peraluminous, characterized by relatively low silica (<70 wt.%), high CaO (<4.39 wt.%), Sr, Eu, variable K₂O/Na₂O ratios (0.77–14.25), low incompatible elements, but moderate to high (La/Yb)_N (3.5–35.8) and Zr/Hf (20.1–48.9) ratios, and a pronounced negative Nb anomaly with small Eu, Ti, Sr, and Ba anomalies. These rocks are I-type granitoids, and may have been derived from partial melting of lower crustal sources with arc-type chemical signatures. Melting likely occurred due to crustal thickening and underplating of basaltic magmas. Intra-intrusion fractionation coupled with supracrustal assimilation is common. The geochemical characteristics of GMS rocks are generally similar to those of magmatic systems associated with gold deposits elsewhere. In contrast, GS rocks, although mostly calc-alkaline and metaluminous to strongly peraluminous, contain high silica (>72 wt.%), alkalis, incompatible elements, and HFSE, but low CaO (<<1.88 wt.%), TiO₂ (<0.22 wt.%), (La/Yb)_N (<7.3) and Zr/Hf (12.8–43.0) ratios, and do not have (or very small) a Nb anomaly, but have more pronounced negative Eu, Ti, Sr, and Ba anomalies. They may have been derived from partial melting of crustal quartzofeldspathic sources under relatively low temperatures.

Gold abundances in GS and GMS rocks are relatively low. Magma processes, such as AFC processes, may have played a role in related gold mineralization. Supracrustal country rocks (e.g., black-shale) may have acted as a *f*(O₂) buffer, resulting in some GMS intrusions with reduced I-type features. This study suggests that a reduced I-type intrusion together with surrounding country rocks may promote gold precipitation from auriferous ore-fluids derived from progressive cooling of voluminous magmas at depth.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.lithos.2008.01.002.

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