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MINERAL CHEMISTRY OF THE GREAT BEND COMPLEX RECORDS SUBDUCTION INITIATION WITHIN THE PENOBSCOT BACK-ARC BASIN

G. Santos, I.C.Y. Yip and H.A.I. Sandeman[1]
Regional Geology Section
[1]Mineral Deposits Section

ABSTRACT

The Great Bend Complex (GBC) is one of three frequently correlated ophiolitic complexes at the base of the Exploits Subzone, a major tectonic division in Newfoundland. This contribution presents new mineral chemistry data from five representative samples of dunite from the GBC, central Newfoundland, collected during the 2024 field season. The samples are medium-grained, strongly serpentinized, and consist of olivine, disseminated euhedral to subhedral Cr-spinel grains and minor interstitial clinopyroxene.

The composition of olivine, clinopyroxene and Cr-spinel are all consistent with the formation of dunite via *a replacive reaction at low pressures between a depleted peridotite and an exotic percolating boninite melt. The record of boninitic magmatism, typically preserved in the GBC is inconsistent with the prevailing tectonic models where the GBC is assumed to represent a back-arc basin ophiolite. The data discussed here suggest at least one previously unrecognized episode of basin inversion and subduction initiation within the Penobscot back-arc basin between its formation* ca. *494 Ma and ophiolite obduction* ca. *476 Ma. An episode of ridge inversion during induced subduction in the Penobscot back-arc basin would be consistent with the absence of a mature intra-oceanic arc and the short timeline between basin formation, inversion, boninite magmatism, and ophiolite obduction.*

INTRODUCTION

Ophiolites, exposed remnants of non-subducted oceanic crust, are key exposures for the petrology of the oceanic crust and the nature of plate tectonics. Mineral chemistry of peridotites is critical for interpreting the magmatic processes that generated them and for constraining their original tectonic setting. The Great Bend Complex (GBC) is one of several ophiolitic complexes and fragments thought to define the eastern margin of the Dunnage Zone in central Newfoundland (Figure 1; Blackwood, 1982; Colman-Sadd *et al.*, 1992). It is also the most poorly exposed and least studied compared with the better understood Coy Pond and Pipestone Pond complexes (Colman-Sadd and Swinden, 1984; Dunning and Krogh, 1985; Jenner and Swinden, 1993).

This contribution addresses the knowledge gap by presenting electron microprobe data of olivine, clinopyroxene and Cr-spinel from the ophiolites of the GBC. The data indicates the GBC preserves the mineral chemical signature of subduction initiation in a forearc setting, and not back-arc spreading providing further constraints on the Late Cambrian tectonic history of the Penobscot arc/back-arc basin.

LOCAL GEOLOGY

The GBC forms the eastern edge of the Exploits Subzone of the Dunnage Zone, one of the major tectonostratigraphic subdivisions of the northern Appalachians of central Newfoundland (Figure 2) (*e.g.*, Williams, 1979; Williams *et al.*, 1988). The Exploits Subzone comprises arc and back-arc terranes and associated marine sedimentary and volcanic rocks formed on the leading Ganderian margin of the Iapetus Ocean. The oldest rocks of the Dunnage Zone consist of three ophiolitic complexes, from east to west, Great Bend, Coy Pond and Pipestone Pond (*e.g.*, Colman-Sadd and Swinden, 1984). These three complexes are often stratigraphically correlated in the literature (*e.g.*, Colman-Sadd and Swinden, 1984; Jenner and Swinden, 1993; Sandeman *et al.*, 2012, van Staal and Barr, 2012).

During its drift through the Iapetus Ocean, Ganderia was the continental locus of at least two arc–back-arc systems. The earlier of the two being the 515–485 Ma Penobscot arc (Zagorevski *et al.*, 2010). Rifting at *ca.* 494 Ma led to the opening of the Penobscot back-arc basin and rifting of the now remnant Penobscot arc from the Ganderian passive margin (Colman-Sadd *et al.*, 1992; van

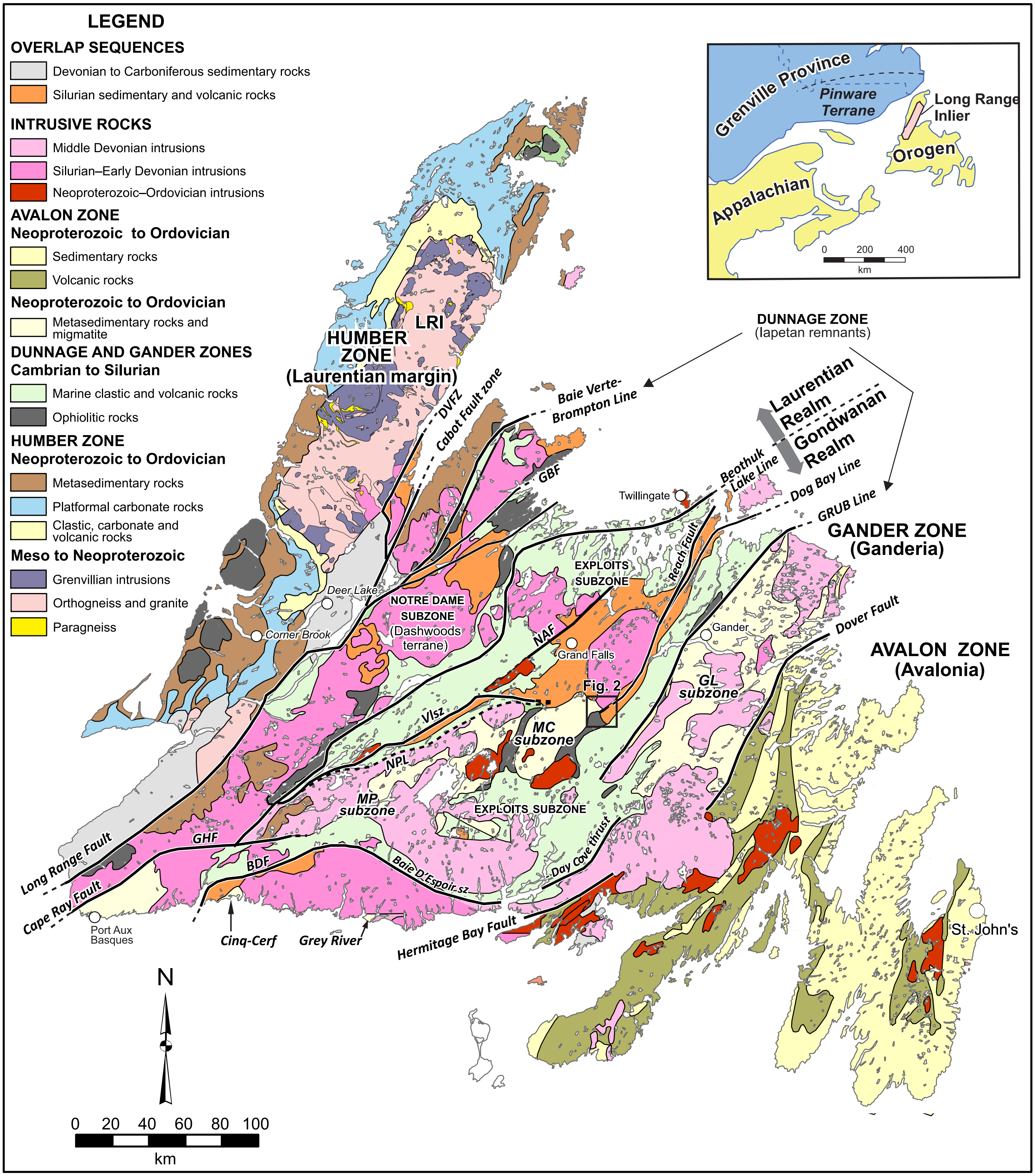


Figure 1. *Simplified regional geology of south-central Newfoundland (modified after Hinchey* et al.*, 2025) and information from the GSNL geoscience atlas (Colman-Sadd* et al.*, 1990). Abbreviations: BDF–Baie d'Est Fault, DVFZ–Doucers Valley Fault system, GBF–Green Bay Fault, GHF–Gunflap Hills Fault, GL–Gander Lake subzone, LRI–Long Range Inlier, MC–Mount Cormack subzone, MP–Meelpaeg, NAF–Northern Arm Fault, NPL–Noel Paul's Line, Vlsz–Valentine Lake shear zone.*

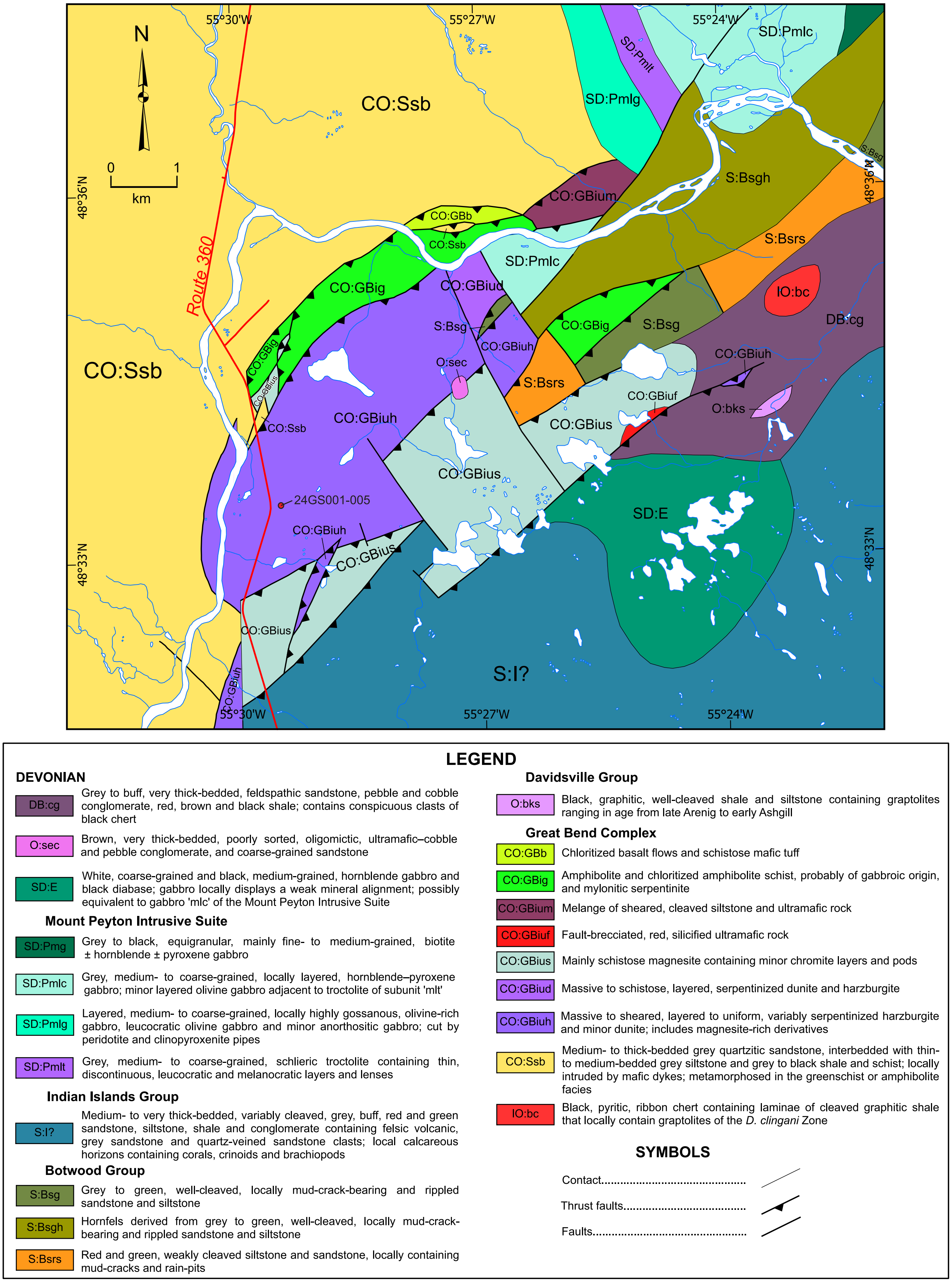


Figure 2. *Local geology of the Great Bend Complex, central Newfoundland. Modified after Colman-Sadd* et al. *(1990).*

Staal 1994; Zagorevski *et al.*, 2010; Sandeman *et al.*, 2012). Zagorevski *et al.* (2010) and van Staal and Barr (2012) suggested the Great Bend, Coy Pond and Pipestone complexes formed in this stage of back-arc basin spreading. Closure of the Penobscot back-arc basin is constrained by the age of the "stitching", *ca.* 474 Ma Partridgeberry Hills Pluton (Colman-Sadd *et al.*, 1992) and may have been caused by the shallowing of the subducting Iapetus slab (van Staal *et al.*, 2009). During the Ordovician, the collapsed Penobscot arc became the infrastructure of the second, 478–474 Ma Victoria arc (Tucker *et al.*, 1994; Valverde-Vaquero *et al.*, 2000; Zagorevski *et al.*, 2010). A significant slab-roll-back episode along the northwestern leading edge of the Victoria arc led to arc-rifting and the formation of the Tetagouche–Exploits back-arc basin at *ca.* 475–465 Ma (van Staal *et al.*, 2003; Valverde-Vaquero *et al.*, 2006).

The GBC has been described as comprising several fault bounded rocks, mostly composed of peridotite and altered ultramafic cumulates, with minor gabbro, basalt and trondhjemite (Colman-Sadd and Swinden, 1984; Dickson, 1992). The larger blocks commonly preserve ophiolitic associations, with metaperidotite, gabbro, diabase dykes, basalt (with rare pillow structures preserved), and trondhjemite (Sandeman and Dickson, 2019). The limited preserved stratigraphy was interpreted by Colman-Sadd and Swinden (1984) as a correlative to the better-preserved Coy Pond and Pipestone Pond complexes. Direct geochronological constraints for the GBC are limited. Dunning and Krogh (1985) reported a U–Pb zircon age of 494 ± 3 Ma for the Pipestone Pond Complex and a minimum $^{207}Pb/^{206}Pb$ zircon age of 489 Ma for a trondhjemite from the Coy Pond Complex. Sandeman and Dickson (2019) reported a hornblende quasi-plateau $^{40}Ar/^{39}Ar$ age of 471 ± 4 Ma for a gabbro in the GBC. This age overlaps with the 474 +6/-3 Ma Partridgeberry Hills Pluton (Colman-Sadd *et al.*, 1992), which intrudes several of the surrounding units, including the Coy Pond Complex, and is therefore interpreted as a stitching pluton. Sandeman and Dickson (2019) interpreted this 474 Ma as the maximum age constraint for the thrusting of the GBC over the passive margin of Ganderia.

SAMPLES AND LITHOLOGY

Metaperidotite from the GBC is well exposed in a quarry alongside the Bay d'Espoir Highway (611230.79 mE, 5379306.48 mN: WGS1984, Zone 21N), where five representative samples were collected (24GS001-005) in the summer of 2024. The quarry exposes a ~320 x 250 m body of dunite, exhibiting a dark orange to light brown weathered surface (Plate 1A). Talc and serpentine veins, up to 5-cm thick and 0.5–2-m long, are ubiquitous and show no discernible preferential orientation in the quarry (Plate 1B). The dunite is medium grained, containing olivine (>95 Vol %), Cr-spinel (<2 Vol %), and minor clinopyroxene. Olivine grains are extensively replaced by serpentine, resulting in widespread mesh textures in the collected samples (Plate 1C). Cr-spinel is fine to medium grained (<2 mm), and largely euhedral (Plate 1D). Cr-spinel is typically disseminated but locally forms schlieren 0.5–1-cm thick and up to 40-cm long. Minor clinopyroxene is preserved as inclusions in Cr-spinel grains in samples 25GS003 and 25GS005. Denser concentrations of Cr-spinel define a weak foliation through the outcrop.

MINERAL CHEMISTRY

ANALYTICAL CONDITIONS

Mineral analytical data for olivine, clinopyroxene and Cr-spinel in five thin sections were obtained using a JEOL JXA-8230 SuperProbe electron probe microanalyzer at Memorial University. Analyses were carried out in wavelength dispersive mode under an accelerating voltage of 15 kV with a beam current of 20 nA. JEOL software was used to quantify raw X-ray intensities using standard ZAF techniques. Olivine was analyzed for Si, Al, Ti, Fe, Mn, Mg, Ca, Co and Cr. Clinopyroxene was analyzed for Si, Al, Ti, Fe, Mn, Mg, Ca, Na, Zn, Ni, Co and Cr. Spinel was analyzed for Fe, Mg, Cr, Al, Ti, V, Mn, Zn, Ni and Co. The complete dataset consists of 49 olivine analyses, 6 clinopyroxene analyses, and 114 Cr-spinel analyses. The whole dataset is available in the supplementary table S1.

Olivine

Olivine grains in the samples are mostly serpentinized. None of the analyzed olivine grains from the GBC plot within the mantle array of Takahashi (1986), with grains invariably skewed towards lower NiO contents (Figure 3). Olivine shows uniformly high Fo values (Mg# 90.2 to 92.3) with little variation. The variation in NiO content is larger, with samples 24GS001, 24GS003 and 24GS005 clustering at 0.28–0.34 wt. %, and having a wider range between 0.11–0.24 wt. % for samples 24GS002 and 24GS004; all at nearly constant Mg#. MnO and CaO concentrations are low, up to 0.24 wt. % and 0.20 wt. %, respectively. The Mg# and NiO contents of olivine plot within the fields of abyssal peridotite (samples 24GS001 and 24GS003; Niu, 2024) and forearc peridotite (sample 24GS005; Pagé *et al.*, 2009), with the data from samples 24GS002 and 24GS004 closely following the mantle-melt interaction vector towards lower NiO values (Hirai *et al.*, 2024) from the forearc peridotite field (Figure 3).

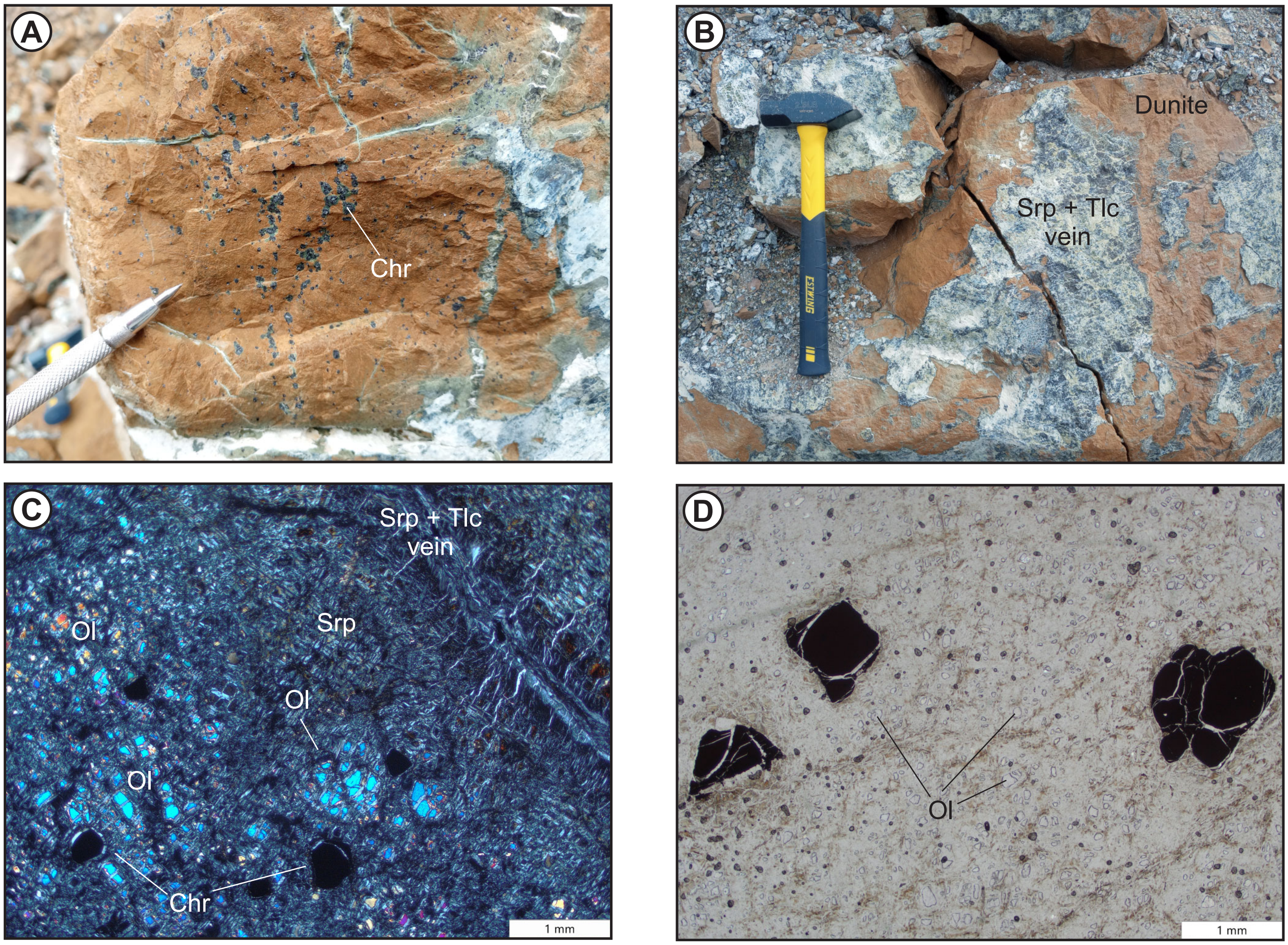


Plate 1. *Great Bend Complex Dunite. A) General aspect of the dunite, showing denser concentration of Cr-spinel grains and serpentine + talc veins; B) Major serpentine + talc veins cutting the dunite; C) Cross-polarized image of sample 24GS003, showing relic olivine grains displaying well-developed mesh textures in a serpentine ground mass. Chromite grains and a serpentine + talc vein are also visible; D) Plane-polarized image of coarse Cr-spinel grains from sample 24GS002, showing euhedral to subhedral habits. Note the small relic olivine grains widespread in the field of view. Abbreviations: Chr–Cr-spinel, Ol–olivine, Srp–serpentine, Tlc–talc.*

Clinopyroxene

Clinopyroxene grains exhibit uniformly high Mg# (94.8–95.4), with low TiO_2 (0.01 to 0.04 wt. %) (Figure 4A) and low Al_2O_3 (0.72 to 1.18 wt. %; Figure 4B). Measured Cr_2O_3 values have much wider variation, ranging between 0.62 to 1.12 wt. % for sample 24GS003 and 1.17 to 1.16 wt. % in sample 24GS004 (Figure 4C), reflecting Cr# (Cr/[Cr + Al]) ranges of 27.3 to 44.1 and 52.3 and 55.7, respectively. Clinopyroxene compositions fall within the forearc peridotite field of Morishita *et al.* (2011) and form an array with varying Cr_2O_3 at low Al_2O_3 concentrations anchored in the secondary clinopyroxene field in Figure 4C.

Cr-Spinel

Cr-spinel compositions are divided into three groups. Group I consists of Cr-spinel rims and one core from sample 24GS003. This group has Mg#'s between 16 and 20.2, and the highest measured Cr# (87.7 to 92.7) values in the dataset (Figure 5A). TiO_2 for these Cr-spinel grains largely overlap with the other two groups, ranging from 0.14 to 0.20 wt. % (Figure 5B). These grains plot in the boninite field of Barnes and Roeder (2001; Figure 5B).

Group II Cr-spinel are present in 24GS001, 24GS002, 24GS004 and 24GS005, and encompass most of the measurements, with high Mg# values (37.4 to 49.24) and inter-

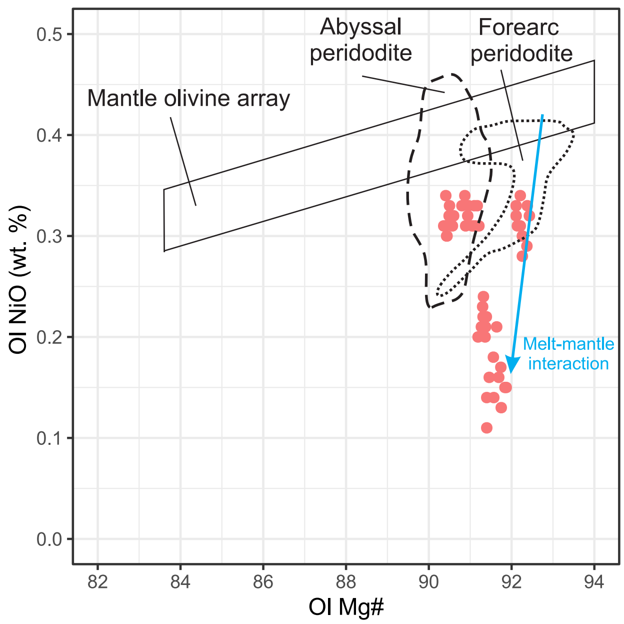


Figure 3. *NiO (wt. %)* vs. *Mg# for olivine in the Great Bend Complex. Mantle olivine array is referenced to Takahashi (1986). Approximately half of the olivine measurements plot within the abyssal peridotite field of Niu (2004) and the forearc peridotite field of Page* et al. *(2009). The other half of the samples have low NiO contents, forming an array parallel to the mantle-melt interaction vector of Hirai* et al. *(2024).*

mediate Cr# values (74.1 to 78.4; Figure 5A), and a wide range in TiO_2 concentrations (0.05 to 0.21 wt. %; Figure 5B). Group II analyses overlap the island-arc tholeiite (IAT; Allan, 1994) and forearc peridotite (Ishii *et al.*, 1992) fields in Figure 5A, and plot at the intersection between the forearc (Kamenetsky *et al.*, 2001), suprasubduction zone (SSZ) (Kamenetsky *et al.*, 2001), and boninite fields (Barnes and Roeder, 2001) in Figure 5B.

Group III comprises Cr-spinel cores in sample 24GS003, with Mg# (Figure 5A) and TiO_2 contents (40.1 to 43.5 and 0.10 to 0.16 wt. %; Figure 5B) overlapping that of Group II and the lowest Cr# (67.5 to 69.8) values in the GBC. It also plots in the intersection of the IAT (Allan, 1994) and forearc peridotite (Ishii *et al.*, 1992) fields in Figure 6A and the intersection of forearc (Kamenetsky *et al.*, 2001) and SSZ (Kamenetsky *et al.*, 2001) fields in Figure 5B.

DISCUSSION

ORIGIN OF THE DUNITE

Several competing models have been proposed to explain the origins of dunite. It may represent the refractory residue after high amounts of partial melting of mantle peridotite (*e.g.*, Nicolas and Prinzhofer, 1983), the products of olivine accumulation from mafic or ultramafic melts (*e.g.*, Moores and Vine, 1971), the products of reactive replacement of pyroxene-bearing peridotite by percolating melts (Batanova *et al.*, 1998; Suhr, 1992), or a combination of replacive and cumulative models (*e.g.*, Merseburger *et al.*, 2025). Mineral chemistry is a fundamental tool for discriminating between these hypotheses.

Olivine in the GBC dunite has lower NiO at high Mg# than would be expected in mantle olivine. In Figure 4, approximately half of the analyses plot within the abyssal and forearc peridotite fields, with the other half displaced below the forearc peridotite field (toward low NiO values) in a trend parallel to the melt-mantle interaction vector of Hirai *et al.* (2024). This melt-mantle interaction vector is based on the assimilation and fractional crystallization (AFC) model of Kelemen (1990), where the reaction at low pressure between an olivine-saturated and pyroxene unsaturated melt precipitates olivine and assimilate pyroxenes from the host peridotite. As olivine crystallizes, it preferentially removes Mg from the melt, which would normally decrease the melt's Mg# and thus lower the Fo of subsequent olivine. However, the dissolution of mantle pyroxenes buffers the Mg and Fe content of the melt, keeping the Fo of crystallizing olivine stable and high (primitive-like) even as the reaction progresses (*e.g.*, Suhr *et al.*, 2003). Ni is highly compatible in olivine and has comparatively lower concentrations in pyroxene. As pyroxenes are assimilated by the percolating melt, they do not replenish NiO in the melt sufficiently to offset the loss from olivine crystallization. Consequently, as the AFC process progresses along the reaction front, the melt becomes increasingly NiO-depleted, leading to the crystallization of olivine with progressively lower NiO contents, while maintaining high Fo (Kelemen, 1990; Kelemen *et al.*, 1992; Suhr *et al.*, 2003). The low NiO content in GBC dunite is incompatible with residual mantle olivine, which would have retained elevated NiO values.

The chemistry of Cr-spinel is also consistent with the origin of the GBC dunite as a reaction product. Group I Cr-spinel plots within the boninite field in Figure 5B, and close to boninite melts in the Cr# *vs.* TiO_2 diagram (Pearce *et al.*, 2000; Figure 5C), with Group II Cr-spinel largely overlapping the Izu-Bonin-Mariana boninite field, and Group III showing slightly lower Cr#'s. Overall, Cr-Spinel from the GBC define a trend between a refractory peridotite (after ~20% melt extraction from a fertile NMORB mantle) and boninite. The morphology of Group I Cr-spinel, which mainly consists of rims, further suggests the modification of pre-existing Cr-spinel by the subsequent percolation of a boninite melt.

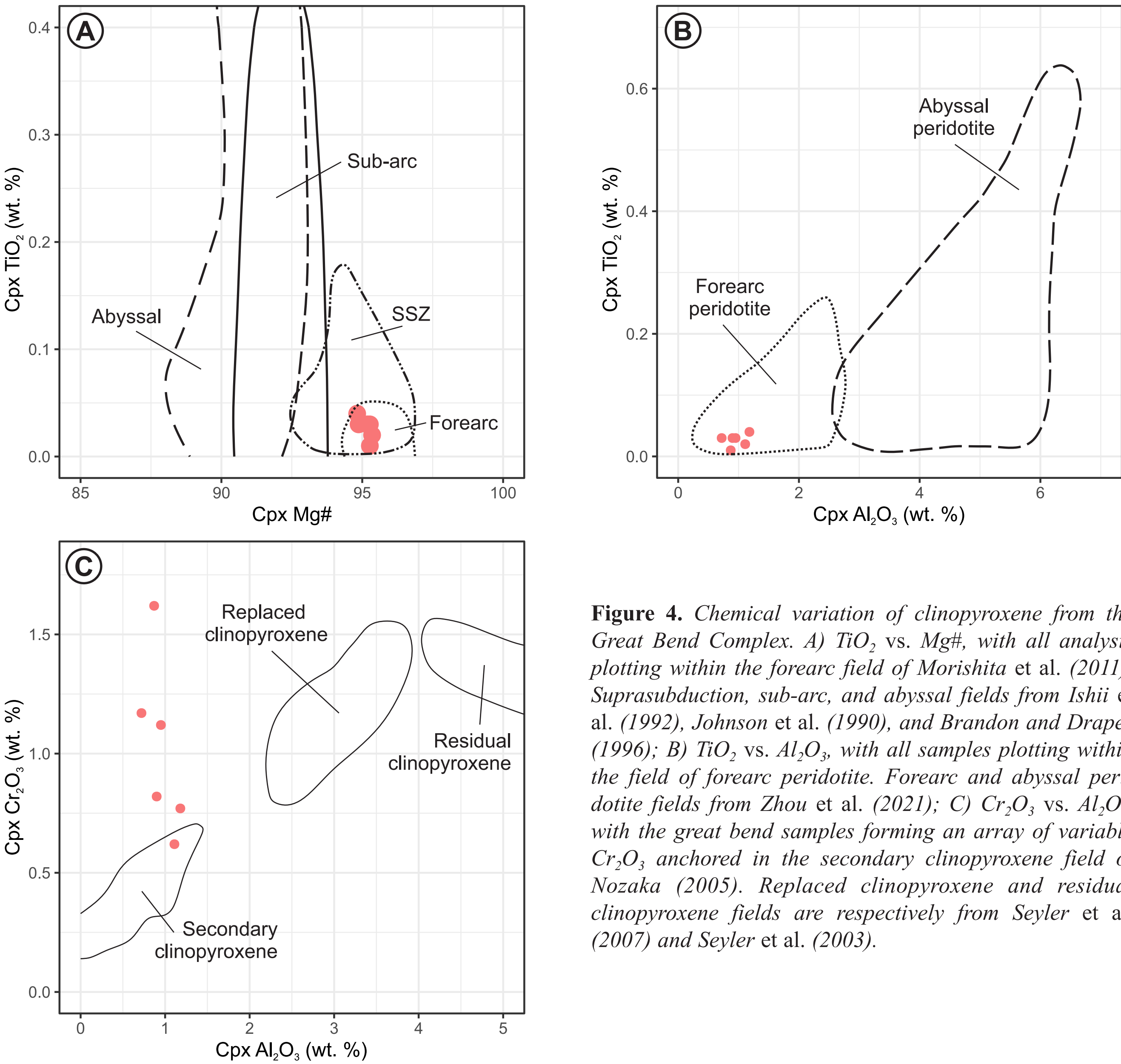
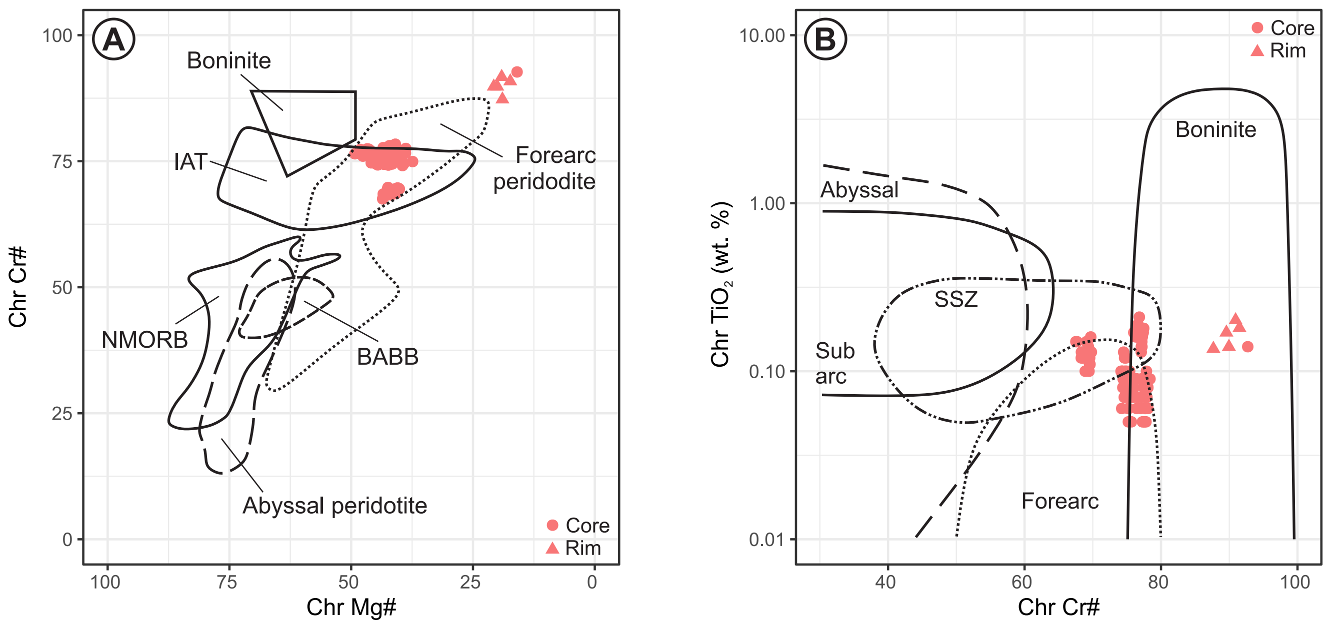


Figure 4. *Chemical variation of clinopyroxene from the Great Bend Complex. A) TiO_2* vs. *Mg#, with all analysis plotting within the forearc field of Morishita* et al. *(2011). Suprasubduction, sub-arc, and abyssal fields from Ishii* et al. *(1992), Johnson* et al. *(1990), and Brandon and Draper (1996); B) TiO_2* vs. *Al_2O_3, with all samples plotting within the field of forearc peridotite. Forearc and abyssal peridotite fields from Zhou* et al. *(2021); C) Cr_2O_3* vs. *Al_2O_3, with the great bend samples forming an array of variable Cr_2O_3 anchored in the secondary clinopyroxene field of Nozaka (2005). Replaced clinopyroxene and residual clinopyroxene fields are respectively from Seyler* et al. *(2007) and Seyler* et al. *(2003).*

Boninite melts are undersaturated in orthopyroxene, leading to orthopyroxene dissolution in the host peridotite along flow channels and in the crystallization of clinopyroxene. Despite the extensive serpentinization of the GBC dunite, the observed clinopyroxene grains form small inclusions in Cr-spinel. This morphology is consistent with the formation of clinopyroxene as reaction products along limited percolation channels or reaction fronts (*e.g.*, Le Roux *et al.*, 2014; Le Roux and Liang, 2019). The combination of low Al_2O_3 and high Cr_2O_3 in dunite clinopyroxene (Figure 4C) is consistent with formation through mineral replacement (Nozaka, 2005; Seyler *et al.*, 2007). Trace-element analysis would be needed to confirm the reactive origin of clinopyroxene.

The co-variation of co-existing Cr-spinel and olivine can be used to infer the degree of melt extraction and the pressure at which the host peridotite was formed (Pearce *et al.*, 2000). Data from the GBC forms an array within the olivine–spinel mantle array (Figure 6), specifically in the compositional field of oceanic suprasubduction zone peridotite. This provides further corroboration of the refractory nature of the peridotite. Furthermore, the array defined by Cr-spinel Cr#, and olivine Mg# indicates a shallow (~5 kbar



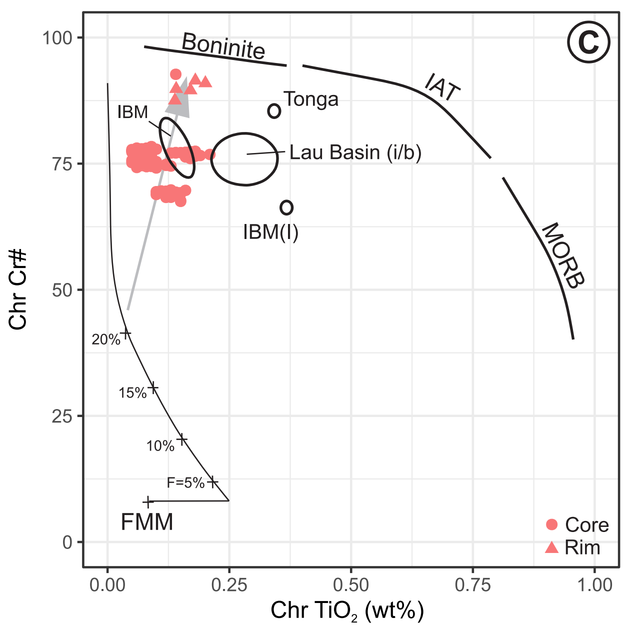


Figure 5. *Chemical variation of Cr-spinel from the Great Bend Complex. A) Ch#* vs. *Mg# where groups II and III plot at the intersection of IAT (Allan, 1994) and forearc peridotite (Ishii* et al.*, 1992) fields. Group I Cr-spinel plots adjacent to the forearc-peridotite field. NMORB, abyssal and boninite fields are from Dick and Bullen (1984), and the back-arc field is after Allan (1994); B) TiO_2* vs. *Cr#, where group I Cr-spinel plots within the boninite field of Barnes and Roeder (2001), and groups II and III plot at the intersection of the suprasubduction zone (Ishii* et al.*, 1992), forearc (Morishita* et al.*, 2011), and boninite fields. Abyssal and sub-arc fields are from Kamenetsky* et al. *(2001); C) Cr#* vs. *TiO_2 diagram from Pearce* et al. *(2000), where the Great Bend Complex samples form array between a heavily depleted mantle peridotite and boninite melts. IBM stands for Izu-Bonin-Mariana, "b" stands for boininite and "i" to island-arc tholeiite.*

i.e., ~15 km) mantle provenance, with perhaps a minor influence from fractional crystallization processes.

TECTONIC IMPLICATIONS

The GBC has not been the target of detailed mapping or geochemical studies and is typically discussed collectively with other correlated ophiolites in central Newfoundland, including the Coy Pond and the Pipestone complexes (*e.g.*, Colman-Sadd and Swinden, 1984). As such, the tectonic setting of the stratigraphically correlated ophiolitic complexes can be used to elucidate the genesis of the GBC. While discussing lithogeochemistry and Sm–Nd isotope systematics of basalt, gabbro, peridotite and plagiogranite of the Pipestone Pond Complex, Jenner and Swinden (1993) described a large depletion in incompatible elements and an isotopic signature suggesting that the analyzed rocks formed from a mantle source that was not contaminated by a subducting slab. They proposed two possible tectonic settings for the formation of the Pipestone Pond Complex. Their first hypothesis was formation in a subduction initiation or intraplate setting, and the second was formation at an arc-rift or back-arc basin. Based on regional correlations, such as the presence of older volcanic arc rocks in the Victoria Lake Supergroup, Jenner and Swinden (1993) favoured an arc-rift or back-arc basin setting. Zagorevski *et al.* (2010) proposed

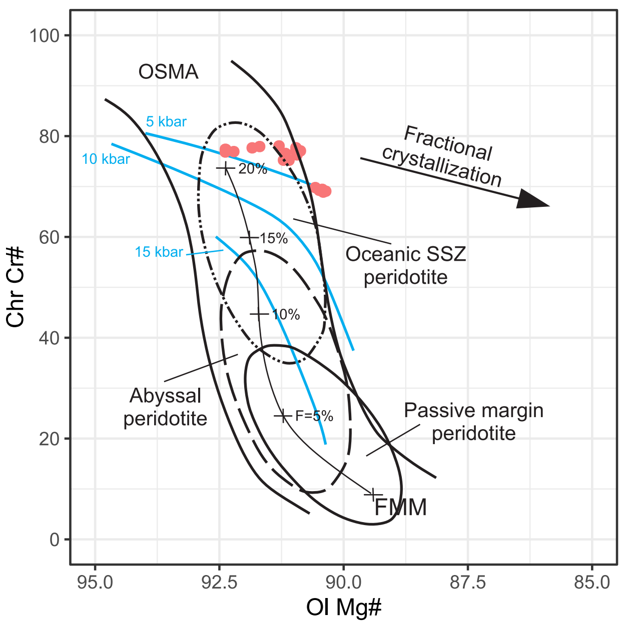


Figure 6. *Cr#* vs. *Mg# for coexisting Cr-spinel and olivine (respectively) after Pearce* et al. *(2000) in the dunite from the Great Bend Complex. Abyssal peridotite field after Dick and Bullen (1984), SSZ and passive margin fields after Pearce* et al. *(2000). OSMA (olivine–spinel mantle array) after Arai (1994). The diagram includes the melting curve of Pearce* et al. *(2000) and the 5 and 10 kbar curves from Sobolev and Batanova (1995) and the 15 kbar curve of Jaques and Green (1980).*

the Pipestone Pond Complex and Coy Pond Complex formed when incipient rifting of the Penobscot Arc resulted in the *ca.* 494 Ma opening of a back-arc basin in the Penobscot retro-arc. The remains of that back-arc basin are preserved in the *ca.* 494 Ma Coy Pond, Pipestone Pond and Great Bend complexes. This interpretation was reiterated by van Staal and Barr (2012).

The mineral chemistry data discussed here, however, strongly favours the hypothesis that the dunite in the GBC is replacive, formed by the reaction of a percolating bonitite melt through a refractory peridotite. Although rare instances of boninitic magmatism in back-arc basins (*i.e.*, Sinton *et al.*, 2003; Keller *et al.*, 2008) and arc rifting (Ribeiro *et al.*, 2013, 2015) have been documented, boninitic magmatism is strongly associated with and frequently used as a fingerprint for a subduction initiation setting (*i.e.*, Stern and Bloomer, 1992; Sklyarove *et al.*, 2016; Pearce and Reagan, 2019). There is no preserved intra-oceanic arc at this time within the Penobscot back-arc basin, indicating that if arc-magmatism took place, it was likely short lived, generating volumetrically small amounts of magma. Inversion of a back-arc basin can be caused by several mechanisms (*e.g.*, *see* Stern and Gerya, 2018), including ridge inversion (Maunder *et al.*, 2020), polarity reversal or subduction transference (Munteanu *et al.*, 2011), transform zone collapse (Sdrolias and Muller, 2006; Stern and Gerya, 2018), change in overriding plate velocity or trench advance (Sdrolias and Muller, 2006; Stern and Gerya, 2018; Balaz *et al.*, 2022), and collision-induced stresses (Balazs *et al.*, 2022). Obduction of the GBC and correlated ophiolites in central Newfoundland, and therefore closure of the Penobscot back-arc basin likely took place *ca.* 476 Ma, constrained by the age of the stitching Partridgeberry Hills Pluton (Colman-Sadd *et al.*, 1992) and a *ca.* 471 Ma $^{40}Ar/^{39}Ar$ age of hornblende in an amphibolite from the GBC (Sandeman and Dickson, 2019), leaving a window of ~20 My for basin formation, inversion, boninite magmatism and ophiolite obduction.

The absence of a preserved mature arc and the short timeline available for boninitic magmatism between back-arc basin formation and obduction favours ridge inversion during induced subduction initiation as the most likely mechanism (*e.g.*, Ishizuka *et al.*, 2011; Leng *et al.*, 2012). The likely cause of ridge inversion in the Penobscot back-arc basin is at present underdetermined. There are several examples of short Wilson cycles where basin formation, inversion, boninite magmatism, and basin closure/ophiolite obduction takes place within the 20 My time frame, including the Oman Ophiolite (Ishikawa *et al.*, 2002), Izu-Bonin system (no full basin closure, but re-start of back-arc opening: *i.e.*, Ishizuka *et al.*, 2011, Shervais *et al.*, 2019), and several intra-oceanic arc systems in the Central Asia Orogenic Belt (Furnes and Safanova, 2019) and western Pacific (Whattam, 2023a, b).

SUMMARY AND CONCLUSIONS

Five representative samples of dunite from the GBC were analyzed after the 2024 field season. The samples were medium-grained, strongly serpentinized, and consisting of olivine, disseminated euhedral to subhedral Cr-spinel grains and minor interstitial clinopyroxene. A weak foliation defined by higher-than-average concentrations of Cr-spinel is widespread. The composition of olivine (high Mg#, high Fo and low NiO), clinopyroxene (high Mg#, high Cr_2O_3, low TiO_2 and low Al_2O_3), and Cr-spinel (high Cr#, low Mg# and low TiO_2) are consistent with the formation of dunite *via* a reaction at low pressures (~5 kbar) between a host depleted peridotite (~20% melt extracted from a fertile NMORB mantle) and a percolating boninite melt.

The tectonic setting of the Great Bend, Coy Pond and Pipestone Pond complexes is generally assumed to be a back-arc basin formed during the *ca.* 494 Ma rifting of the Penobscot arc (Jenner and Swinden, 1993; Zagorevski *et al.*, 2010; van Staal and Barr, 2012). The record of boninitic magmatism preserved in the GBC is inconsistent with this hypothesis, requiring at least one previously unrecognized episode of basin inversion and subduction initiation within the Penobscot back-arc basin between its formation *ca.* 494 Ma and ophiolite obduction *ca.* 476 Ma. An ephemeral episode of subduction initiation causing ridge inversion would account for the boninitic magmatism and the absence of a developed intra-oceanic arc.

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