

SUMMARY OF ACTIVITIES 2024

Geoscience BC Report 2025-01





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Geoscience BC (2025): Geoscience BC Summary of Activities 2024; Geoscience BC, Report 2025-01, 88 p.

Summary of Activities (Geoscience BC) Annual publication

ISSN 2817-8564 (Online)

Geoscience BC 1107–750 West Pender Street Vancouver, British Columbia V6C 2T8 Canada

Front cover photo and credit: J.D. Bhakta, M.Sc. student at the University of Manitoba, at the end of a field day in the Ice River project area, southeastern British Columbia. Photo by A.R. Chakhmouradian.



Foreword

Geoscience BC is pleased to once again present results from our scholarship recipients and ongoing projects in our annual *Summary of Activities* publication. Geoscience BC would like to thank all authors and reviewers of the *Summary of Activities* papers for their contributions and RnD Technical for its work in editing and assembling this volume.

Geoscience BC would like to thank all members (corporate, individual, associate and student), project sponsors and advocates for their ongoing support of Geoscience BC's independent public geoscience. Teck Resources Limited is thanked for their sponsorship of the 2024 Geoscience BC Scholarship program, as are the Association for Mineral Exploration British Columbia, Canadian Discovery Ltd. and the Kamloops Exploration Group for their support of this volume. Finally, Geoscience BC would like to express our appreciation for the leaders and volunteers in British Columbia's mineral exploration, mining, clean energy and geological carbon storage sectors who support our organization through their guidance and their use and recognition of the data and information that we collect and publish.

Christa Pellett Vice President, Minerals Geoscience BC www.geosciencebc.com





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Role of Alteration in the Alaskan-Type Turnagain Complex, North-Central British Columbia (Parts of NTS 104I/07, 20)

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Broda, K.R., Williams-Jones, A.E. and Vasyukova, O.V. (2025): Role of alteration in the Alaskan-type Turnagain Complex, north-central British Columbia (parts of NTS 104I/07, 20); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 1–6.

Introduction

Although numerous studies have investigated the magmatic processes producing sulphide mineralization in mafic–ultramafic igneous complexes, fewer studies have examined the influence of hydrothermal processes on this type of mineralization. Most of the latter studies have focused on the serpentinization of the olivine and pyroxene in these complexes and rarely have considered the effect of this alteration on the distribution of the metals. Notable exceptions are the studies by Keays and Jowitt (2013) and Kamenetsky et al. (2016) that documented the mobilization of nickel during the serpentinization of olivine in the McIvor Hill mafic–ultramafic complex of Tasmania, in Australia.

The Turnagain Complex, situated in northern British Columbia (BC; Figure 1), is an example of an Alaskan-type mafic–ultramafic complex containing a potentially economic concentration of magmatic nickel-sulphide mineralization. Serpentinization resulted in the replacement of some of the magmatic olivine by serpentine and the remobilization of the primary sulphides to form secondary sulphides. In addition, deserpentinization led to the formation of secondary olivine that replaced serpentine and locally created a near-complete pseudocumulate texture.

Owing to the presence of nickel in both the sulphide and silicate minerals, as well as to the complex interplay between magmatic and hydrothermal processes, the nickel-to-sulphur ratio within the deposit varies considerably (Ni:S = 0.5-14.2; Broda et al., 2024), posing challenges for metallurgical recovery. The Turnagain Complex, with the highest known nickel-sulphide endowment of any Alaskan-type complex (1574 Mt of ore grading 0.21 wt. % Ni; Giga Metals Corporation, 2023), is an ideal natural laboratory in which to

study the processes leading to the development of economic concentrations of nickel-sulphide.

Preliminary results of a petrographic study of representative samples from the Horsetrail-Northwest zone of the Alaskan-type Turnagain mafic–ultramafic complex undertaken to investigate the role of alteration on the nickel-tosulphur ratio are presented in this paper. These results show that the rocks of the Turnagain Complex experienced two stages of alteration: 1) early serpentinization related to fluids introduced along faults and 2) subsequent dehydration of the serpentine as a result of thrusting of the complex to higher crustal levels.

Geology of the Turnagain Complex

The Turnagain Complex is located approximately 70 km east of Dease Lake in northern BC and comprises four spatially and temporally distinct mafic-ultramafic intrusions (units 1-4; Figure 1a) that were emplaced in the late Paleozoic Road River metavolcanic-metasedimentary succession between 189 and 185 Ma (Scheel, 2007; Jackson-Brown, 2017; Nixon et al., 2020). These intrusions comprise, from north to south (Figure 1a, b): wehrlite, olivineclinopyroxenite and minor dunite in unit 1; dunite and wehrlite in unit 2; diorite in unit 3; and clinopyroxenite and hornblendite in unit 4. The complex is bounded to the north and east by a major thrust fault and juxtaposed against pyrite-bearing graphitic phyllite along its northern and eastern margins; it is also in intrusive contact with metavolcanic and metasedimentary rocks on its southern boundary (Scheel, 2007; Jackson-Brown, 2017; Nixon et al., 2020). Enclaves of pyrite-bearing graphitic phyllite and metavolcanic rock are found in all the intrusions except the diorite (Figure 1b). All of the intrusive rocks have undergone varying degrees of alteration, mainly serpentinization, particularly close to faults and shear zones that are present in the complex (Clark, 1980).

The Turnagain deposit is located along the southern margin of intrusive unit 2, in the Horsetrail-Northwest zone (Figure 1b). This zone consists predominantly of dunite,

¹The lead author is a 2024 Geoscience BC Scholarship recipient.

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Figure 1. Geology and location of the Alaskan-type Turnagain mafic–ultramafic complex: **a)** map showing the generalized geology, the boundaries of intrusive units (1–4) that represent separate intrusive units composed of the rock types identified in the legend of panel (b), the Horsetrail-Northwest resource zone (HTNW) study area outlined by the dashed rectangle in the southeastern part of the complex and the location of the complex (bottom right inset); **b)** detailed geology of the Turnagain Complex study area. All co-ordinates are in UTM Zone 9N, NAD83.

Methodology

wehrlite, olivine-clinopyroxenite and clinopyroxenite. A distinguishing feature of the Horsetrail-Northwest zone is the association of a large body of low-grade nickel-sulphide mineralization with abundant pyrite-bearing graphitic phyllite enclaves.

The primary silicate and sulphide textures in the Horsetrail-Northwest zone have been described in Broda et al. (2024). The fresh dunite, wehrlite and olivine-clinopyroxenite consist primarily of equigranular, 1-3 mm (locally up to 15 mm in diameter), euhedral to anhedral cumulus olivine, with varying proportions of interstitial clinopyroxene comprising mainly diopside (Clark, 1980; Scheel, 2007); and subhedral to euhedral cumulate spinel clusters, less than 1 mm in diameter. In the clinopyroxene-dominant rocks, the clinopyroxene occurs as 1-3 mm subhedral cumulate crystals.

The main sulphide minerals are disseminated to net-textured interstitial pyrrhotite ($\{(Ni,Fe)_{x-1}S\}$; x = 0–0.2) and pentlandite $\{(Ni,Fe)_9S_8\}$, the primary ore mineral. They are concentrated in the dunite and wehrlite. The pyrrhotite occurs as aggregates of 50 to 500 µm subhedral to euhedral crystals. Two types of pentlandite are present: subhedral to euhedral blocky crystals, ranging from <20 to 150 µm in diameter and occurring as aggregates (up to 1500 µm) enclosed within pyrrhotite or as isolated crystals; and fine lamellae within pyrrhotite. The blocky pentlandite is interpreted as a product of peritectic reaction between monosulphide solid solution (MSS) and sulphide liquid, whereas the fine lamellae likely exsolved from MSS (Broda et al., 2024). The presence of significant interstitial pyrrhotite and pentlandite in the cumulus olivine is consistent with their crystallization from an immiscible sulphide liquid (Scheel, 2007; Jackson-Brown, 2017; Nixon et al., 2020; Broda et al., 2024).

The rocks of the Horsetrail-Northwest zone have undergone varying degrees of serpentinization of the primary olivine and clinopyroxene. This alteration generally proceeded along microfractures, crystal cleavages and grain boundaries. In weakly altered rocks, serpentine formed fibrous, lath-like crystals perpendicular to the olivine grain boundaries. In more strongly altered rocks, the olivine has been completely replaced by serpentine. Veinlets of magnetite and fibrous serpentine are commonly concentrated along grain boundaries, mantling the olivine and preserving the original cumulate texture. Magnetite, produced during serpentinization, is disseminated as anhedral crystals intergrown with serpentine. The clinopyroxene was more resistant than the olivine, with alteration typically restricted to grain boundaries, fractures and cleavages, where it manifests itself as prismatic to tabular tremolite and minor serpentine. The sulphide mineral pyrrhotite was replaced extensively by magnetite and the pentlandite, much less so by magnetite, heazlewoodite and millerite (Broda et al., 2024).

A total of 30 drillcore samples, representative of both altered and unaltered rocks from the Horsetrail-Northwest zone, were selected for petrographic analysis. The selection was based on the degree of alteration and the textural relationships of the minerals in the sulphide ores. The petrographic analyses were performed at the Department of Earth and Planetary Sciences of McGill University using polished thin sections. An Olympus BX51-P polarizing microscope and a Hitachi High-Tech Canada, Inc. SU5000 field-emission scanning electron microscope, equipped with an X-Max^N 80 silicon drift detector from Oxford Instruments plc, were employed for this purpose.

Results

Alteration

Alteration in the Horsetrail-Northwest zone of the Turnagain Complex is evident by the variable replacement of cumulus olivine by serpentine, magnetite and minor brucite. In weakly altered samples, serpentine replaced olivine along crystal boundaries and microfractures (Figure 2a). However, in samples or parts of samples in which the alteration was intense (samples taken closer to faults), cumulus olivine crystals were completely replaced by serpentine (Figure 2b). Magnetite occurs as disseminated anhedral crystals that are intergrown with serpentine along relict grain boundaries and microfractures; minor brucite is intergrown with serpentine, but due to its fine-grained nature, is hard to discern.

The alteration of interstitial diopside to tremolite, and locally serpentine, is only evident where the alteration was intense. Replacement of the diopside by tabular tremolite grains (<100 μ m) is patchy and localized to grain boundaries, microfractures and cleavage planes.

Locally, secondary olivine replaced serpentine along relict grain boundaries, producing veins as well as rims on the serpentine pseudomorphs after primary olivine (Figure 2b). Additionally, rare disseminated anhedral grains of secondary olivine replaced mesh-to hourglass-textured serpentine. The grain size of the olivine pseudomorphs is erratic and varies according to alteration intensity. In rare cases, secondary olivine is juxtaposed with relict cumulus olivine.

The alteration of the sulphide minerals, pyrrhotite and pentlandite, occurred mainly along grain boundaries, fracture surfaces and cleavage planes. Where the alteration was strong, pyrrhotite was partially to almost completely replaced by magnetite (Figure 2c). In contrast, pentlandite was replaced only locally by magnetite and, to a lesser extent, by heazlewoodite, millerite and violarite.





Figure 2. Photomicrographs of primary and secondary silicate and sulphide minerals of samples from the Turnagain Complex illustrating their textural relationships: **a**) image in cross-polarized light of subhedral-euhedral cumulus olivine (OI), with traces of serpentine (Srp) along the fractures, completely encased in interstitial clinopyroxene (Cpx); **b**) image in cross-polarized light of secondary olivine developed along the rim and crosscutting a crystal of primary olivine that has been completely replaced by serpentine and is surrounded by pyrrhotite (Po) and pentlandite (Pn); **c**) image in reflected light of magnetite (Mag) locally replacing pyrrhotite, and of pyrrhotite and pentlandite replacing serpentine, which itself has replaced primary olivine, encased in interstitial pyrrhotite and serpentine; **d**) image in cross-polarized light of secondary olivine (the secondary olivine and pyrrhotite replacing serpentine, which itself has replaced pyrrhotite.

The secondary sulphides, pyrrhotite and pentlandite, are disseminated in serpentine and magnetite veinlets. Additionally, olivine and serpentine were replaced along their crystal boundaries by pyrrhotite and minor pentlandite, which take the form of irregularly shaped dendrites (Figure 2c, d).

Discussion

The silicate and sulphide textures observed in the Horsetrail-Northwest zone of the Turnagain Complex record two distinct alteration events. The first event was marked by the alteration of primary silicate and sulphide minerals by hydrothermal fluids. In this event, olivine was variably replaced by serpentine+magnetite±brucite (Brc), according to the following reaction assuming an olivine composition of Fo₉₃ (preliminary microprobe data):

$$\begin{array}{l} 10Mg_{1.86}Fe_{0.14}SiO_{4(Ol)} + 14.2H_2O_{(l)} \Leftrightarrow \\ 5Mg_3Si_2O_5(OH)_{4(Srp)} + 3.8Mg_{0.95}Fe_{0.05}(OH)_{2(Brc)} + \\ 0.4Fe_3O_{4(Mag)} + 0.2H_{2(g)} \qquad (\text{reaction 1}) \end{array}$$

and clinopyroxene (as diopside [Di]) by tremolite (Tr) plus or minus serpentine via the following reactions:

$$2MgCaSi_2O_{6(Di)} + 4H_2O_{(l)} + 3Mg^{2+}_{(aq)} + 4SiO_{2(aq)} \Leftrightarrow Ca_2Mg_5Si_8O_{22}(OH)_{2(Tr)} + 6H^+_{(aq)}$$
 (reaction 2)

and/or



$$Mg_3Si_2O_5(OH)_{4(Srp)} + 2MgCaSi_2O_{6(Di)} + 2SiO_{2(aq)} \Leftrightarrow Ca_2Mg_5Si_8O_{22}(OH)_{2(Tr)} + H_2O_{(l)}$$
 (reaction 3)

In addition, pyrrhotite was replaced by magnetite, and pentlandite by magnetite or, less commonly, by heazlewoodite, millerite and violarite. These findings are consistent with the textures and reactions previously described and reported in Broda et al. (2024). An increase in the intensity of the serpentinization with increasing proximity to faults suggests that this alteration was due to the interaction of the ultramafic rocks of the complex with fluids focused along these faults.

The second alteration event involved the dehydration of serpentine to form secondary olivine, according to the following reaction:

$$Mg_{3}Si_{2}O_{5}(OH)_{4(Srp)} + Mg_{0.95}Fe_{0.05}(OH)_{2(Brc)} \Leftrightarrow 2 Mg_{1.975}Fe_{0.025}SiO_{4(Ol)} + 3 H_{2}O_{(l)}$$
(reaction 4)

As magnetite produced during serpentinization was not involved in this reaction, the secondary olivine is predicted to exhibit a higher magnesium content than the primary magmatic olivine (Filippidis, 1982). Indeed, semiquantitative scanning electron microscope–energy dispersive spectrometry (SEM-EDS) analyses show this to have been the cause. Additionally, nickel was mobilized from olivine during serpentinization to form secondary sulphide minerals, implying that the secondary olivine should have a lower concentration of nickel than the primary olivine.

As the Turnagain Complex and its hostrock are part of a thrust sequence, it appears that this dehydration event was triggered by the decrease in pressure that accompanied the displacement of the complex to higher crustal levels. This decrease in pressure would have favoured the product (low volume) side of Reaction 4, leading to the replacement of the serpentine and brucite by olivine.

During one of these alteration events, pyrrhotite (with minor pentlandite) pseudomorphs may have formed via the reaction:

$$3Mg_{1.8}Fe_{0.2}SiO_{4(Ol)} + 3H_2O_{(l)} + 0.6H_2S_{(l)} + 0.6SiO_{2(aq)} \Leftrightarrow 1.8Mg_3Si_2O_5(OH)_{4(Srp)} + 0.6FeS_{(Po)}$$
(reaction 5)

as proposed in Broda et al. (2024), and may indicate that sulphur fugacity was high during serpentinization (Filippidis, 1982).

Conclusions and Future Work

The Alaskan-type Turnagain mafic–ultramafic complex experienced two distinct stages of alteration: 1) hydrothermal alteration causing the serpentinization of olivine, the alteration of diopside to tremolite and the replacement of sulphides by magnetite, followed by the remobilization of sulphide; and 2) the dehydration of serpentine to form secondary olivine. The first event is attributed to the introduction of hydrothermal fluids along faults within the complex and the second event, to the decrease in pressure that accompanied the thrusting of the complex to higher crustal levels.

The interplay between magmatic processes and the mobilization of nickel and sulphur during alteration events was likely responsible for the observed variability and unpredictability in nickel-to-sulphur ratios. A deeper understanding of these processes will be crucial in developing a comprehensive genetic model for mineralization in Alaskan-type complexes. The preliminary interpretations presented above, which are based on petrographic observations of textural relationships, will be tested by detailed analyses of the compositions of the primary and secondary silicate and sulphide minerals using a combination of electron probe microanalysis (EPMA) and laser-ablation inductively coupled plasma-mass spectrometry (ICP-MS) methods. Proven correct, they should show, among other things, that the secondary olivine is depleted in both iron and nickel. In addition, the analyses will help in assessing spatial and temporal variations in mineral compositions during alteration. Sulphur isotope compositions will be measured using laser-ablation, multiple-collector inductively coupled plasma-mass spectrometry (LA-MC-ICP-MS) to evaluate sulphur mobilization and physicochemical parameters such as oxygen fugacity in hydrothermal fluids. Furthermore, ultraviolet femtosecond laser-ablation, multiple-collector inductively coupled plasma-mass spectrometry (UV-fs-LA-MC-ICP-MS) will be employed to measure in situ iron and nickel isotope compositions in rock-forming minerals. This will aid in the evaluation of the sources of iron and nickel and their isotopic fractionation between minerals during the primary magmatic and secondary alteration processes. The resulting data will be used to constrain temperature and redox conditions during magmatism and serpentinization.

Acknowledgments

This project was funded by the Natural Sciences and Engineering Research Council of Canada (NSERC) and Giga Metals Corporation. The lead author extends sincere thanks to Geoscience BC for financial support through the 2023 and 2024 Geoscience BC Scholarship programs. The authors also gratefully acknowledge the technical and logistical support provided by Giga Metals Corporation. Finally, they thank K. Rempel for her insightful and constructive feedback during the review process.

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Camp Creek Copper-Molybdenum-(Gold) Porphyry-Alteration Mineralogy and Geochemistry: Exploring for Blind Copper Mineralization in Northern British Columbia (NTS 104K/10W)

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Introduction

Porphyry Cu-Mo-(Au) deposits are a significant source of critical metals for technologies required to facilitate a transition toward electricity-based energy systems. Porphyry Cu deposits are the primary source of Cu globally, but the world-class discoveries needed to meet increasing demand are not being made. As most near-surface porphyries have likely been discovered, developing tools for exploring 'blind' subsurface porphyry deposits is vital to meet increasing global resource demand (Sillitoe, 2010).

The Intermontane Belt corresponds to the Intermontane Superterrane, which includes the Quesnel and Stikine island-arc terranes and the Cache Creek ocean-floor terrane (Mihalasky et al., 2010). This belt features postaccretionary porphyry Cu deposits of both calcalkaline porphyry Cu±Mo±Au and alkaline porphyry Cu-Au subtypes. The Camp Creek prospect is a calcalkalic porphyry Cu-Mo prospect located in the northwestern portion of the Canadian Cordillera of British Columbia (BC; Figure 1) with other intermontane island-arc porphyry Cu deposits, including the Highland Valley Copper Mine, the Gibraltar Mine and the Schaft Creek joint venture (Mihalasky et al., 2010). The Camp Creek prospect is part of the Thorn property, which is located approximately 94 km east of Juneau, Alaska, 120 km northwest of Telegraph Creek, BC and 130 km southeast of Atlin, BC (Figure 1a). In 2019, significant zones of porphyry-style veining with Cu and Mo mineralization were intersected at Camp Creek at depths of more than 300 m, with further drilling programs in 2021–2024 focused on discovering extensions of porphyry-style mineralization (Lemiski et al., 2023).

In the early 2000s, a project was initiated by the Mineral Deposit Research Unit (MDRU) at The University of British Columbia to investigate the late Cretaceous igneous complexes in the Taku River area of the Stikine terrane in northwestern BC. This study focused on characterizing intrusions and establishing a geochronology framework for the larger Thorn district (Simmons, 2005; Simmons et al., 2005). More recently, MDRU completed a further geochemical and petrographic study following the discovery of Camp Creek, using data collected in 2021, to help distinguish various intrusive phases at the deposit (Bouzari and Barker, 2021).

¹*The lead author is a 2024 Geoscience BC Scholarship recipient.*

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Figure 1. Simplified geology of the Camp Creek porphyry, part of the Thorn property in northwestern British Columbia. The location of the Thorn property is shown on the inset map, outlined in red. All co-ordinates are in UTM Zone 8N (NAD83).

This paper summarizes previous work, discusses preliminary field observations from fieldwork carried out in 2023– 2024 and outlines next steps for the coming year. This project, part of the lead author's M.Sc. research, is undertaking the first major geological review of the Late Cretaceous calcalkalic Cu-Mo-(Au) Camp Creek porphyry prospect, focused on characterizing the alteration, mineralization and time scales of magmatism. Ongoing research is integrating petrography, micro X-ray fluorescence (micro-XRF) element mapping, shortwave infrared studies, wholerock geochemistry and U-Pb geochronology, and will contribute insights into potential geochemical vectors for discovering blind porphyry Cu-Au-(Mo) mineralization.

Regional Geology

The Thorn property is located in the northwest-trending Stikine terrane, the largest and westernmost terrane of the Intermontane Belt in the Canadian Cordillera. The Stikine terrane comprises Triassic island-arc volcanic rocks and related sedimentary sequences of the Stuhini Group, with accretion of the Stikine Arc to western North America being assumed to have taken place in the Middle Jurassic at 172 Ma (Mihalynuk, 1999). Following accretion, Late Cretaceous igneous rocks were emplaced into the Stikine terrane as part of a continental arc, forming a north-northwest-trending belt that extends for 300 km from Yukon into northern BC (Mihalynuk, 1999; Simmons, 2005).

Two distinct periods of Late Cretaceous magmatism have been recognized at the Thorn property, the older magmatic pulse being informally referred to as the Thorn Stock, with exposure of this unit shown in Figure 1 in the study area (Mihalynuk, 1999; Simmons, 2005). The Thorn is composed of dominantly tholeiitic diorite porphyry intrusions, which are the major hostrocks for hydrothermal mineralizing systems at the Thorn property (Awmack, 2002; Simmons, 2005). The Thorn Stock has been constrained to an age of 93.3 ± 2.4 Ma by U-Pb geochronology of zircon from multiple intrusions (Mihalynuk et al., 2003). The younger Windy Table suite was emplaced into the Thorn stock, and is characterized by subaerial, dominantly felsic volcanic rocks and associated calcalkaline, equigranular monzonite to granodiorite (Awmack, 2002). Windy Table suite volcanic rocks nonconformably overlie the Thorn stock (Figure 2). The tuff overlying this unconformity has a U-Pb SHRIMP-RG zircon age of 84.7 ±0.8 Ma, providing an age constraint for the onset of Windy Table volcanism (Simmons, 2005).





Figure 2. Camp Creek field area: a) Camp Creek looking northeast, highlighting quartz diorite (Thorn stock) outcrop; b) Camp Creek valley looking east; c) unconformity at the base of the Windy Table suite, previously sampled by Simmons (2005); d) Windy Table volcanic strata looking east.

Geology of the Camp Creek Property

Initial whole-rock geochemical analysis of core samples from Camp Creek by Bouzari and Barker (2021) revealed the occurrence of at least five porphyry units. These units are geochemically distinct from the Stuhini volcanosedimentary hostrocks, which are characterized by high Sc (>10 ppm). These porphyry units have been termed X, Y, Z, W and V, and are distinguished by different ratios of several trace elements, including Zr, Ti and Sc, as seen in Figure 3. Petrography by Bouzari and Barker (2021) provided an initial correlation between mineralogy and geochemistry, providing preliminary criteria for characterizing and distinguishing each of the porphyry units. Representative hostrock samples from drillcore were collected by the author during 2023–2024 fieldwork for further geochemical analysis to improve the previously established characteristics. Examples are shown in Figure 4 and field observations are summarized in the following section. The effective subdivision of the porphyry stocks is critical to refine the geological model of Camp Creek and to understand grade distribution.

Volcanosedimentary Units (Stuhini Group)

The Stuhini Group hostrock at Camp Creek contains siltstone and mafic volcanic rocks with pervasive biotite alteration. It is exposed at surface to the southwest of the study area and is typically first intersected at depths of 500– 700 m at Camp Creek. Bedding is occasionally preserved and the unit is crosscut by an intense quartz-vein stockwork when proximal to porphyry X.

Porphyry Z

Porphyry Z is a quartz-biotite-hornblende-plagioclase porphyry. Diagnostic features of porphyry Z include abundant medium-grained hornblende phenocrysts, medium- to coarse-grained biotite and 1–4 mm rounded quartz pheno-





Figure 3. Classification of hostrocks at Camp Creek based on Zr, Ti and Sc trace-element data from whole-rock geochemical analysis: a) Sc versus Ti; b) Sc versus Zr; c) Ti versus Zr (Bouzari and Barker, 2021; reproduced with permission).

crysts. Biotite phenocrysts are 1–4 mm, typically coarse and commonly replaced by fine-grained sericite and rutile. Hornblende is the next most abundant phase after plagioclase, typically tabular and commonly replaced by chlorite. Porphyry Z is distinguished by its high Ti (>0.2%) and Nb (>5 ppm; Bouzari and Barker, 2021), and is the least altered and mineralized of the porphyry phases, dominantly pyrite (1–5%) with trace chalcopyrite (<0.1%).

Porphyry Y

Porphyry Y is a hornblende-quartz-biotite-plagioclase porphyry and is mineralogically and texturally similar to porphyry Z but with significantly less hornblende phenocrysts. Porphyry Y typically has stronger alteration and contains higher amounts of sulphide (5–10%) and Cu grades (>0.1%) when compared to porphyry Z. Geochemically, porphyry Y has lower Ti (<0.2%) and lower Zr (<40 ppm) than porphyry Z (Bouzari and Barker, 2021). Porphyry W was originally distinguished by slightly elevated Zr contents compared to porphyry Y, but is mineralogically and texturally indistinguishable and has been classified as part of porphyry Y during core logging. There were no clear crosscutting relationships observed between porphyries Y and Z.

Porphyry X

Porphyry X is a hornblende-quartz-biotite-plagioclase porphyry. Diagnostic features include coarse biotite books, quartz-rich groundmass and strong K-feldspar alteration, dominantly within the groundmass. It is cut by a stockwork of A-type granular quartz veins and has the highest concentration of Cu mineralization among intrusive units (>0.2%). Geochemically, porphyry X is distinguished by its low Zr (<20 ppm) relative to porphyries Y and Z (Bouzari and Barker, 2021). No clear crosscutting relationships were observed between porphyry X and porphyries Y and Z.

Porphyry V

Porphyry V is a biotite-hornblende-plagioclase porphyry, typically occurring as smaller dikes that crosscut porphyries Z, Y and X. Diagnostic features include fresh to weakly altered biotite phenocrysts, 3–5% disseminated magnetite and rare quartz phenocrysts. It is weakly altered and contains trace pyrite but no Cu-bearing mineralization. Geochemically, porphyry V is characterized by its high Zr concentration (>75 ppm), which is significantly higher than the other porphyry phases observed at Camp Creek (Bouzari and Barker, 2021). Clear crosscutting relationships were observed between porphyry V and porphyries Z, Y and X.

Polymictic Diatreme Breccia (Oban breccia)

The Oban zone is a porphyry-related diatreme breccia hosting polymetallic (Ag-Au-Pb-Zn-Cu) mineralization within the breccia matrix. The Oban breccia varies significantly, and typically contains subangular and well-rounded fragments of fine- to medium-grained quartz-diorite porphyry with altered biotite-feldspar-quartz phenocrysts. Porphyry clasts host quartz veins with remnants of K-feldspar alteration that resemble A-type quartz veins, which suggests the presence of porphyry mineralization at depth. Weakly al-





Figure 4. Examples of hostrocks observed at Camp Creek, with corresponding scale card shown to the right of each sample: **a**) siltstone with pervasive and veinlet-controlled biotite alteration overprinted by chlorite and sericite, THN23-261 841.20 m; **b**) quartz-biotite-hornblende-plagioclase porphyry (porphyry Z) with pervasive and phenocryst-replacement sericite alteration, THN23-277 348.60 m; **c**) hornblende-quartz-biotite-plagioclase porphyry (porphyry Y) with pervasive and phenocryst-replacement sericite alteration, overprinted by chlorite, THN23-276 875.25 m; **d**) hornblende-quartz-biotite-plagioclase-porphyry (porphyry X) with pervasive K-feldspar alteration and quartz-vein stockwork, THN23-276 1235.80 m; **e**) biotite-hornblende-plagioclase porphyry (porphyry V), THN22-201 845.97 m. Mineral abbreviations: Bt, biotite; Chl, chlorite; Hbl, hornblende; Ilt, illite; Pl, plagioclase; Py, pyrite; Qz, quartz; Ser, sericite.

tered volcanic and sedimentary clasts are hosted within rock flour, and fractures are commonly filled with sulphides and sulphosalts. Generally, the contact of the Oban diatreme breccia with the hostrock is poorly constrained. Dominant alteration minerals include sericite and pyrophyllite with minor diaspore, suggesting shallower depths of a porphyry environment. The breccia is heterogeneously altered and locally contains fragments that have variable degrees of sericite-pyrophyllite-diaspore alteration (Lemiski et al., 2023). Minerals such as pyrophyllite and diaspore indicate an initially hot, acidic hydrothermal system on the order of 250 °C or greater (Hedenquist et al., 2000; Simmons, 2005).

Alteration

Observations from drilling in the Camp Creek porphyry indicate multiple distinct types of alteration that are evident in the classic porphyry-alteration model of Sillitoe (2010). At Camp Creek, pyrophyllite with rare diaspore occurs at the roots of the preserved advanced argillic zone (see Figure 5), with residual quartz and quartz-kaolinite occurring at higher levels. Zones of advanced argillic alteration occur in the upper part of porphyry Cu deposits that can form a large near-surface footprint of the hydrothermal system, providing a vectoring tool toward concealed porphyry Cu mineralization at depth (Bouzari et al., 2022). A challenge to establishing alteration zoning is the subtle mineralogical changes across porphyritic units that can be difficult to identify (Bouzari et al., 2022). MicroXRF is an effective tool to gauge modal abundances and review mineralogy by element mapping, particularly when phases are difficult to distinguish by eye. Based on previously analyzed shortwave infrared (SWIR) data, a variety of downhole alteration trends and observations is shown in Figure 5 (from Bouzari et al., 2023).

Sericite alteration increases in intensity with depth, being weaker in porphyry Z and typically increasing in porphyry Y before transitioning to a dominantly K-silicate alteration in porphyry X (Bouzari et al., 2023). Sericitic assemblages are typically classified by a quartz-sericite-pyrite assemblage, which transitions into zones of intense chlorite replacement of biotite and hornblende with depth. K-silicate alteration is classified by pervasive fine-grained K-feldspar alteration (Figure 6), common 1–3 mm K-feldspar haloes and occasional biotite alteration. Propylitic alteration is characterized by abundant epidote and chlorite occurring distally to Camp Creek.

Mineralization

Camp Creek is a blind porphyry target, and exploration drilling intersected porphyry-style mineralization in 2019. Drill programs from 2021 to 2024 aimed to push holes beyond 1000 m depth, targeting deeper Cu-Au-Ag-Mo mineralization (Lemiski et al., 2023). Copper mineralization is predominantly hosted by a coarse-grained feldspar-quartz-biotite porphyry, with intervals of Stuhini intermediate volcanic and sedimentary rocks. Porphyry-type vein density, following the model of Gustafson and Hunt (1975), increases gradually with depth. An example interval with higher grade mineralization occurred in hole THN22-201 from 335 m to 1302.71 m, with 967.71 m of 0.25% Cu, 0.09 g/t Au, 2.39 g/t Ag and 186 ppm Mo (Lemiski et al., 2023). Mineralization assemblages comprise mainly chalcopyrite, pyrite and molybdenite within porphyry-style





Figure 5. Downhole shortwave infrared data on drillcore samples collected in 2019–2022, plotted along northward-facing drillcore crosssections (also shown is the corresponding location of the cross-section shown in Figure 1 [modified from Bouzari et al., 2023]): a) lithology; b) chlorite to sericite ratio, showing the relative abundance of chlorite to sericite using the spectral depth of the minerals; c) illite to sericite ratio, with muscovite abundance increasing with depth relative to illite; d) sericite composition, dominated by paragonite illite near surface and transitioning to illite and then phengitic illite with depth; e) sericite crystallinity increasing with depth toward zones of K-silicate alteration; f) pyrophyllite occurring at shallow levels with rare diaspore.

quartz and quartz-anhydrite veins, as well as disseminated sulphides.

postmineral dikes, each of approximately 2 kg, were submitted for U-Pb geochronology.

Sampling and Analytical Work

Detailed observations of, and sample selection for, further geochemical analysis were completed over two summer field seasons in 2023 and 2024. Four drillholes were relogged to capture additional information on mineralogy, alteration, veining and crosscutting relationships (Table 1). Detailed observations in drillcore have been recorded following logging advice from Williams (2022), aiding in the consistent characterization of intrusive units. As of October 2024, 40 samples have undergone whole-rock geochemical analysis. From this sample set, 20 were selected for petrographic analysis. MicroXRF analysis was completed on 14 samples to assess zircon content prior to U-Pb geochronology and hostrock classification. Four samples of More than 400 rocks were sampled from drillcore for follow-up analysis. The majority of these samples have been collected at 50 m intervals to be sent for whole-rock geochemical analysis and thin-section preparation. These samples came from the least-altered sections with the least quartz veining and will be used to understand downhole trends and assist in distinguishing intrusive phases. A short fieldwork stint in August 2024 focused on investigating the unconformity between the Thorn stock and the overlying Windy Table volcanics, with three samples collected for U-Pb geochronology (locations in Figure 1). A complete suite of Shortwave Infrared (SWIR) data has been gathered for 14 prioritized drillholes at Camp Creek, and further work will aim to constrain previously identified zones of alteration.





Figure 6. Flatbed optical scans and microXRF element maps of hostrocks at Camp Creek: **a**) scan of porphyry X at 1013.30 m in drillhole THN23-261; **b**) microXRF element map of porphyry X at 1013.30 m in drillhole THN23-261, with a legend showing elements and their corresponding colours; **c**) scan of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element map of siltstone at 1340.23 m in drillhole THN23-261; **d**) microXRF element ma

Future Work

Ongoing studies will utilize a range of techniques to investigate the evolution of the Camp Creek porphyry prospect. A subsequent detailed petrographic study will provide insight into the textural, mineralogical and alteration variations of the different porphyry phases downhole. Additional samples have been selected for whole-rock geochemistry, and a complete geochemical analysis will be performed on the collected data. Further SWIR data will be processed to build an understanding of the alteration model. More microXRF scans will be collected from existing samples to gain insight into modal abundance of mineralogy and textural relationships. Further U-Pb dating will be completed for intrusive phases, to better constrain the timing and formation of the porphyry stocks at Camp Creek. Research outputs will focus on establishing the evolutionary model of the system, as well as developing an alteration model of the different porphyry phases.

This project will inform mineral-exploration development by adding vectoring tools to assist in more efficient exploration programs, enabling more effective exploration of subsurface cryptic and 'blind' mineralization that may constitute future high-grade Cu-Mo-(Au) discoveries. Feasible programs may help to boost investment in British Columbia's mineral-exploration industry and enable more rapid, cost-effective exploration across the province. Furthermore, understanding the geometry of these porphyries and the spatial association of intrusive phases can help predict where mineralizing phases are likely to be encountered at the deposit scale, increasing the likelihood of exploration success. Alteration footprints can be linked to multiple porphyry phases and help distinguish the paragenetic evolution of intrusive phases at Camp Creek.

Conclusions

Mineralization is hosted primarily as disseminated and vein-hosted chalcopyrite in porphyry X, as well as siltstone and mafic volcanic rocks of the Stuhini Group. Porphyry Z has the weakest sericite alteration, transitioning to more intense sericite alteration and overprinting chlorite alteration in porphyry Y, and weak sericite alteration in shallower zones

Table 1. Diamond-drill hole data for THN22-213, THN23-161, THN23-276 and THN23-285 (Lemiski et al.,2023). All co-ordinates are in UTM Zone 8N (NAD83).

Drillhole ID	Year Completed	Easting	Northing	Azimuth (°)	Dip (°)	Depth (m)
THN22-213	2022	627659.34	6491855.06	319.72	-85.21	1243
THN23-261	2023	628263.704	6491844.726	353	-81	1650
THN23-276	2023	628613.622	6492065.301	216	-82	1470
THN23-285	2023	627876	6491944	110	-84	1602



of porphyry X. These porphyry units overlap in time, based on U-Pb dating, and further analytical work will be needed to resolve uncertainties on timing. Additionally, crosscutting relationships do not provide information on the relative ages of porphyries Z, Y and X through evidence gathered to date. Porphyry V clearly postdates porphyries Z, Y and X, with observed postmineral dikes clearly postdating all other intrusive phases. SWIR data show multiple alteration zones, with mineralogy dominated by pyrophyllite at shallow levels, transitioning to illite and then muscovite at greater depths, and eventually to K-feldspar in porphyry X. Future characterization work will aim to constrain field observations by detailed petrography, additional SWIR analysis and geochemical analysis.

Acknowledgments

This study is being undertaken by the Mineral Deposit Research Unit at the University of British Columbia, in collaboration with Brixton Metals and supported by the MDRU BC Porphyry Project sponsors. Thank you to Geoscience BC, the Society of Economic Geologists and the International Association on the Genesis of Ore Deposits (IAGOD) for providing financial support for the author. Thanks also go to S. Barker and F. Bouzari for providing continued guidance during this research. The author thanks the team at Brixton Metals for supporting summer fieldwork, particularly D. Guestrin, C. James and C. Anstey, and R.G. Lee from BHP for providing input on the project scope. Thank you to M. Manor, M. Rodriguez and N. Moerhuis for providing continuous guidance in the field and lab. Thank you to K.B. Riedell for being a mentor during the project and providing a review as a co-author, and to M. Hohl for providing a peer review of the manuscript.

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Directly Dating Molybdenum-Copper Porphyry and Zinc-Silver-Lead-Copper-Gold Carbonate Replacement Deposit Mineralization in Northwestern British Columbia (NTS 104M/1)

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Bowie, S., Mottram, C., Kellett, D., Barrington, M. and Pike, C. (2025): Directly dating molybdenum-copper porphyry and zinc-silverlead-copper-gold carbonate replacement deposit mineralization in northwestern British Columbia (NTS 104M/1); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 15–26.

Introduction

Understanding tectonic controls on mineralization is central to locating and modelling ore deposits and reducing the socio-environmental footprint of exploration efforts. During collisional mountain-building events, faults and shear zones are known to act as planes of weakness, facilitating the migration of magma and associated mineralizing fluids up through the crust (e.g., Arndt, 2005; Sillitoe, 2010). Evidence of this fluid flow can be preserved in the form of mineralized quartz-carbonate veins that are localized or controlled by fault zones. Methods appropriate for dating magmatic and high-temperature ore-forming processes (such as U-Pb zircon and Re-Os molybdenite dating) cannot quantify important hydrothermal processes that may (re)deposit metals in many systems in the mid- to upper crust. However, recent developments in in-situ U-Pb carbonate geochronology now allow for these lower temperature fluid-flow and mineralization processes to be directly dated (e.g., Rasbury and Cole, 2009; Roberts et al., 2020; Mottram et al., 2024), providing important timing controls for petrogenetic models.

This project aims to directly date fluid flow, faulting and mineralization in a porphyry-epithermal-carbonate replacement deposit (CRD) to understand the timing, rates and duration of fluid flow and metal (re)distribution. It focuses on a new discovery in northwestern British Columbia (BC) associated with the Llewellyn fault, which structurally controls many epithermal and porphyry Cu-Au deposits in the Atlin Mining District north of the Golden Triangle (Figure 1a; e.g., Hart and Pelletier, 1989; Love et al., 1998; Mihalynuk et al., 2003; Ootes et al., 2017, 2018; Millonig et al., 2017). The Blue property, owned and operated by Core Assets Corp., hosts several mineralization styles: Mo-Cu porphyry, Zn-Cu-Ag skarn and Ag-Zn-Pb-Cu CRD in several target areas (e.g., Sulphide City, Jackie; Figure 1). Mineralized zones display evidence of multiple complex magmatic faulting and fluid-flow events that currently have no quantitative geochronological constraints. Therefore, the case-study system is ideal for applying U-Pb carbonate dating methodology to link the timing(s), rate and duration of mineralizing events to the tectonic and metallogenic framework of northwestern BC. Where possible, carbonate dates will be combined with higher temperature geochronometers (e.g., U-Th-Pb zircon, monazite, garnet, apatite and titanite, and/ or Re-Os molybdenite), and trace-element data, to track the evolution of mineralization from the magmatic to hydrothermal regime (~800 °C to ~200 °C).

Geological Setting

The western margin of North America has undergone >200 million years of accretionary orogenesis, processes that continue to be active today (e.g., Coney, 1980; Colpron et al., 2007; Monger and Gibson, 2019). This long-lived tectonic history has resulted in a patchwork of commonly fault-bounded crustal fragments that make up the terranes of the Canadian-Alaskan Cordillera (Coney, 1980; Nelson and Colpron, 2007). These terranes can be grouped into five distinct morphogeological belts (Figure 1a; e.g., Nelson and Colpron, 2007; Soucy La Roche et al., 2022). Of these, the Intermontane Belt extends from south-central BC to central Yukon and comprises allochthonous pericratonic terranes associated with subduction-related arcs (Yukon-

¹The lead author is a 2024 Geoscience BC Scholarship recipient.

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Figure 1. **a)** Morphogeological belts of the Canadian-Alaskan Cordillera (modified from Gabrielse et al., 1991). Atlin Mining District boundaries shown in purple, and area shown in Figure 1b is indicated by black box. Abbreviations: IS, Insular; IM, Intermontane; C, Coast; O, Omineca; F, Foreland. **b)** Simplified geology of southwestern Yukon and northwestern British Columbia, showing locations of mineral occurrences and past-producing mines. Cache Creek terrane units are undifferentiated and shown in grey. Sulphide City and Jackie sites within the Blue property are shown with green squares. Geology after Doherty and Hart (1988), Hart and Pelletier (1989), Hart and Radloff (1990), Mihalynuk (1999) and Ootes et al. (2018), and includes data from MapPlace (BC Geological Survey, 2006) and Yukon MINFILE (Yukon Geological Survey, 2020),. U-Pb zircon, K-Ar hornblende and Lu-Hf garnet sample locations from, Currie and Parrish (1997), Armstrong (unpublished data, BC Geological Survey, 2006) and Dyer (2020); see text for ages. Estimated ages of mineralization shown in black are from Hart and Pelletier (1989; Montana Mountain), Love et al. (1998), Mihalynuk et al. (2003; Middle Ridge), Millonig et al. (2017; Engineer and Mount Skukum) and Ootes et al. (2018; Bennett Plateau). Base map information from Yukon map data (https://map-data.service.yukon.ca/geoyukon/Administrative_Boundaries/Territorial_Provincial_Borders_1M).



Tanana, Stikinia, Quesnellia), and accretionary complexes made up of oceanic-floor sedimentary and volcanic rocks (Cache Creek, Slide Mountain; e.g., Colpron and Nelson, 2011; Monger and Gibson, 2019).

In northwestern BC, rocks that constitute Stikinia include the upper Triassic to middle Jurassic volcano-sedimentary Stuhini and Laberge Groups, as well as associated plutonic suites (Mihalynuk, 1999, Figure 1b). Within the Yukon-Tanana terrane, units include pervasively deformed greenschist to upper-amphibolite grade Proterozoic to early Paleozoic continental margin strata (Mihalynuk, 1999). Stratigraphic units include the Florence Range metamorphic suite of quartz-rich metapelitic rocks, marble, quartzite and amphibolite (Figure 1b; e.g., Jackson et al., 1991). This suite has been correlated with the pre-Late Devonian Snowcap assemblage in Yukon (Mihalynuk, 1999; Piercey and Colpron, 2009) and may have undergone several tectonometamorphic events between ca. 270 Ma and ca. 120 Ma (Currie, 1994; Dyer, 2020; Soucy La Roche et al., 2022). Also present in this region is the Boundary Range metamorphic suite, which consists of marble and limestone as well as chlorite-actinolite and quartzofeldspathic schists (Soucy La Roche et al., 2022). At present, the age and provenance of this unit remain unclear, although metamorphism likely postdates the emplacement of an Early Mississippian pluton (Currie and Parrish, 1997; Soucy La Roche et al., 2022) and Lu-Hf ages indicate garnet growth occurring from ca. 202 to 192 Ma (Dyer, 2020).

Llewellyn Fault System

The Llewellyn fault is a steeply dipping, northwest-striking brittle fault network that stretches ~180 km from northwestern BC to southwestern Yukon, where it may overprint the ductile Tally Ho shear zone (Ootes et al., 2017, 2018). In the Atlin Mining District, this fault may lie along a major terrane boundary between the Yukon-Tanana terrane to the west and the Stikine terrane to the east (Figure 1b; e.g., Hart and Radloff, 1990; Mihalynuk, 1999). Few previous studies have directly dated this fault; however, it is thought to have both ductile (ca. 120-75 Ma) and brittle deformation phases (ca. 56-50 Ma; Hart and Radloff, 1990; Ootes et al., 2017, 2018). Along strike, rocks are penetratively deformed, with south-southeast-trending foliations, lineations and folds that may indicate an overall sinistral sense of shear (Mihalynuk, 1999; Ootes et al., 2018). It is suggested that ductile deformation associated with the Llewellyn and Tally Ho deformational corridor postdates deposition of the Jurassic Laberge Group (Kellett and Zagorevski, 2022); and, while parallel deformational fabrics may be present in Triassic and older units, it remains difficult to definitively link them to a specific fault zone (Mihalynuk, 1999; Ootes et al., 2018). Along the Llewellyn fault zone, brittle deformation has been constrained to ca. 56-50 Ma north of the study site using U-Pb zircon ages from crosscutting plutons and ⁴⁰Ar-³⁹Ar vein adularia and vanadian illite ages from the Engineer and Mount Skukum mines (Love et al., 1998; Millonig et al., 2017; Ootes et al., 2018). This deformation overprints earlier ductile deformation fabrics that are likely analogous to the Tally-Ho shear zone (Figure 1b; Ootes et al., 2018).

The Llewellyn fault is spatially associated with more than 50 mineral occurrences in this region (Figure 1b), including epithermal, mesothermal and intrusion-related Au (Hart and Pelletier, 1989; Mihalynuk, 1999, Mihalynuk et al., 2003; Millonig et al., 2017; Ootes et al., 2017, 2018). At the Engineer and Mount Skukum mines (Figure 1b), it is thought that epithermal Au mineralization may be related to Eocene faulting following pluton emplacement (Love et al., 2017). However, more recent evidence also suggests that at least some intrusion-related Au mineralization at Middle Ridge and Bennett Plateau likely underwent ductile deformation and therefore predates the brittle Eocene phase of faulting in this region (Ootes et al., 2018).

Case-Study Ore Deposit

The Blue property (Core Assets Corp.), located ~50 km southwest of Atlin, BC and ~20-60 km east-southeast of Skagway, Alaska, is spatially proximal to the Llewellyn deformation corridor and comprises Jurassic, Cretaceous and Eocene intrusive units of the Boundary and Florence Range metamorphic suites (Figures 1b, 2; Currie and Parrish, 1993). Recent work focused on the Wann River shear zone near the base of the Willison glacier has shown that ductile deformation within the pre-Late Devonian Florence Range suite may be related to two regional metamorphic events, ca. 270-240 Ma and ca. 195-170 Ma, followed by localized contact metamorphism after ca. 120 Ma (Soucy La Roche et al., 2022). Similarly, Lu-Hf ages from the same area indicate that garnet growth occurred at ca. 190 Ma and ca. 185 Ma (Dyer, 2020). Proximal to the Llewellyn fault, what are interpreted to be zircon crystallization ages of ca. 103 to ca. 101 Ma were obtained from a large plutonic unit, and ages of ca. 352, ca. 335, ca. 191 and ca. 189 Ma were obtained from small granitic dikes near the Llewellyn glacier (Figure 2b; Currie and Parrish, 1997). Near the Willison glacier, previous studies obtained hornblende and biotite K-Ar cooling ages of ca. 144 Ma and ca. 97-96 Ma from amphibolite and granodiorite units, respectively (Armstrong and Werner, unpublished data, BC Geological Survey, 2006). At present there are few absolute geochronological constraints on brittle faulting, mineralized igneous units or fluid-flow processes in the region. This study aims to quantify the thermal-fluid-mineralization evolution of this case-study deposit system and test whether mineralization at the Blue property is temporally related to 1) major regional intrusive units; 2) deformation along the



Figure 2. Schematic deposit-scale maps of the Sulphide City (**a**) and Jackie (**b**) showings within the Blue property (Core Assets Corp.). Locations of samples with U-Pb carbonate dates included in this paper are shown with numbered orange stars: 1) 23SB-013B1; 2) SLM22-006-396.0; and 3) SLM22-002-95.53. Stereonet displays field data from both showings and includes Rose diagrams showing the trends of dike contacts, gneissic foliation and carbonate veins. Slickenlines are plotted by dominant shear sense (dextral, sinistral, normal or reverse). Poles to fault planes are contoured at 1 intervals. Maps modified, with permission, from a Core Assets Corp. assessment report (Rodway and Barrington, 2024) and plotted with EPSG:4326–WGS 84 datum. Abbreviations: Bt, biotite; Cb, carbonate; Grt, garnet; Pb, lead; Py, pyrite; Qz, quartz; U, uranium.



Llewellyn fault or other smaller fault splays; and/or 3) several overprinting mineralizing magmatic-hydrothermal events.

Field Observations and Relationships

Fieldwork carried out in July 2023 and July 2024 focused on the 'Sulphide City' and 'Jackie' showings (Figures 1b, 2) within the Blue property. Hostrocks comprise southeast-striking, steeply dipping and polyphase folded marble and garnet-biotite gneiss of the Florence Range metamorphic suite (Figures 1b, 2; e.g., Mihalynuk, 1999). Mineralization styles range from Mo-Cu porphyry and Fe-Zn-Cu skarn at Sulphide City to Zn-Ag-Pb-Cu-Au CRD, massive to semimassive sulphide skarn and distal epithermal veins at Jackie. Mineralized zones correlate with proximity to surficial exposures of ~5-200 m wide porphyritic tonalitic to granodioritic intrusions and ~15-50 cm wide mafic to intermediate dikes. Northeast-striking, steeply dipping brittle faults commonly crosscut the dominantly northwest-trending lithological contacts between gneiss and marble units, as well as younger dikes and plutonic units (Figure 2), and may also correlate with areas of increased mineralization and fluid alteration. Fault planes variably contain ~1-2 cm thick quartz, carbonate and, more rarely, zinc- and/or iron-oxide infill.

Sulphide City

Intrusive units, \sim 5–200 m wide, are observed at Sulphide City and include mineralized tonalitic to granodioritic rocks that can be divided into four phases based on field relationships. Exposed at the edges of the central intrusive body (phase 3; Figures 2a and 3), phases 1 and 2 may be defined as diatreme/magmatic-hydrothermal breccias and are challenging to differentiate. Due to pervasive sericite alteration, the clasts, ranging from <1 cm to 10 cm, are difficult to identify but may be composed primarily of felsic igneous material containing ~1–2 mm quartz eyes and biotite and feldspar porphyroclasts. In places, clasts of Zn-Fe-Cu massive sulphide associated with phase 1 occur within phase 2 (Figure 3b). The matrix contains quartz and sericite-altered feldspar <1 mm in size and, in places, disseminated pyrite and pyrrhotite mineralization ~2-5 mm in size.

Phase 3 is composed of light grey, massive granodiorite. It has a porphyritic texture, with 2–3 mm plagioclase and 1-2 mm biotite phenocrysts (~5%) in a fine crystalline matrix composed dominantly of quartz (~50%), plagioclase (~30-40%) and K-feldspar (~10-20%; Figure 3a). It also hosts ~0.1–0.5 cm crystals of porphyry-style pyrite, pyrrhotite and molybdenite mineralization, both in generally westdipping ~0.2–0.5 cm wide stringers and disseminated throughout the matrix. Chalcopyrite mineralization is less common at the surface but may occur with pyrrhotite as ~0.1–0.5 cm crystals disseminated within the matrix and increasing in abundance with depth. This unit is not pervasively brecciated, but in places contains rare clasts of massive sulphide and garnet skarn mineralization. At the surface, phase 3 generally displays a lower degree of sericite alteration than phases 1 and 2, and plagioclase phenocrysts are less pervasively altered.

Phase 4 is a northwest-striking, steeply dipping, unmineralized diorite dike approximately ~5-6 m wide. It has a porphyritic texture containing ~1-2 mm biotite phenocrysts (~10-15%) and less quartz than phases 1 to 3 (~30–40%). It appears largely unaltered by sericite or skarn alteration, and phenocrysts of plagioclase and biotite appear euhedral and intact on fresh surfaces. Therefore, this unit likely postdates igneous mineralizing activity. It has a ~1 m wide chilled margin defined by a fining of crystal size from ~5-10 mm to <2 mm that visibly crosscuts phases 2 and 3 and contains rounded xenoliths similar in composition to phase 3.

Carbonate minerals are present at Sulphide City in an array of settings, including as $\sim 0.5-3$ cm wide deformed veins within the gneissic and marble country rocks; fault breccia infill and slickenfibres; crystals intergrown with pyrite, pyrrhotite, chalcopyrite, sphalerite, epidote and garnet; and as barren and undeformed veins crosscutting mineralized units (Figure 3c, d). Skarn mineralization at Sulphide City is observed at the margins of phases 1 to 3, occurring mainly at northwest-trending, steeply dipping contacts between marble and garnet-biotite gneiss hostrock units (Figure 3e), at contacts between hostrock units and predeformational felsic sills and dikes, and within north- to west-northwest-trending hinge zones in doubly plunging marble folds. Both exoskarn and endoskarn mineralization occurs across the showing. Exoskarn is observed as massive replacement of marble and as smaller ~10-50 cm pods along fracture surfaces (Figure 3d). Prograde assemblages include $\sim 0.1-1$ cm wollastonite needles, $\sim 0.2-1$ cm red to light brown zoned garnet and dark green augite in a very fine (<0.5 mm) crystalline sericite-quartz-carbonate matrix (Figure 3d, e); ~0.2-1 cm, light green- to brown-zoned vesuvianite may also be present in places. These areas are often spatially associated with retrograde assemblages of quartz, carbonate, epidote, chlorite and massive sulphide mineralization (Py+Ccp+Po+Sp±Gn; Figure 3e). Although also observed in phases 2 and 3, endoskarn alteration is especially well developed in phase 1 as massive garnet, augite, epidote and chlorite, along with rusty orange-weathering massive sulphide (Py+Po+Gn+Sp) mineralization. With increasing distance (e.g., >50 m) from intrusive phases 1-3, both prograde and retrograde skarn assemblages reduce in intensity.

Jackie

The Jackie showing comprises mainly southeast-striking marble and garnet-biotite gneiss country rocks with gener-





Figure 3. Examples of intrusive units and mineralization styles at Sulphide City: a) porphyry-style mineralization in granodiorite (phase 3) containing disseminated and vein-hosted pyrite, molybdenite and chalcopyrite; b) massive sulphide skarn clasts within the highly sericitealtered matrix of phase 2; c) pyrite mineralization within a carbonate vein crosscutting deformed and altered garnet-biotite gneiss hostrock; d) carbonate-garnet-augite-magnetite skarn mineralization partially replacing marble hostrock; e) rusty-weathering, massive sulphide mineralization and wollastonite-carbonate-garnet skarn alteration at the margin of marble hostrock. Inset shows detail of texture within the massive sulphide mineralization zone. Abbreviations: Aug, augite; Bt, biotite; Cb, carbonate; Grt, garnet; Gn, galena; Mag, magnetite; PI, plagioclase; Po, pyrrhotite; Py, pyrite; Sp, sphalerite; Wo, wollastonite.

ally east-southeast-trending complex folds crosscut by several intrusive units (Figure 2b). Crosscutting intrusions include 1) ~2–3 m wide, northeast- to east-striking hornblende gabbro dikes that, in places, host pyrite mineralization in ~1 mm wide veins; 2) <1 m wide, typically northeast-striking, often aphanitic, light green–weathering intermediate 'alaskite' dikes (Figure 4a); 3) ~0.5–1 m wide north-northwest- to northwest-trending mineralized pebble dikes that crosscut host rocks and are composed of ~1–15 cm, rounded to subrounded clasts of marble and garnetbiotite gneiss, mineralized tonalite to granodiorite, and massive sulphide skarn mineralization (Figure 4b). Pebble dikes are largely clast supported and have matrices consisting of ~85–95% hydrothermal quartz and carbonate, and ~5–15% rock flour and sulphide minerals. Pyrrhotite-

sphalerite-galena-chalcopyrite CRD mineralization commonly occurs in centimetre- to metre-scale amorphous pods and massive replacement zones, ~1-2 cm wide veins and at northwest-striking contacts between marble and the 'alaskite' dikes (Figure 4a, c).

Steeply dipping, northeast-trending brittle faults crosscut all lithologies present at Jackie, including mineralized units and typically show evidence for centimetre- to metre-scale offsets (Figure 4d). Carbonate, wollastonite and/or ironoxide slickenfibres are commonly observed on fault surfaces and suggest a mixture of normal, reverse and strikeslip motion, with both sinistral and dextral displacement, across the showing (Figure 2b). Although the 'alaskite' dikes are often crosscut and offset by faults, faulting is also observed parallel to and at northwest-striking contacts be-





Figure 4. Mineralization at the Jackie showing: **a**) pyrite and chalcopyrite mineralization at the contact between an 'alaskite' dike and marble host rock; **b**) pebble dike containing marble, granodiorite and garnet-biotite gneiss clasts; inset shows clast containing sulphide mineralization; **c**) carbonate replacement-style sulphide mineralization in marble; massive sulphide texture detail shown in inset; **d**) ~50–75 cm offset on a conjugate fault set crosscutting mineralized bands within marble host rock; **e**) hydrothermal Fe-rich carbonate with cockade textures infilling a ~1–2 m wide fault zone and containing disseminated pyrite mineralization. Abbreviations: Bt, biotite; Cb, carbonate; Ccp, chalcopyrite; Gn, galena; Grt, garnet; Po, pyrrhotite; Py, pyrite; Sp, sphalerite.

tween the dikes and host rocks. At the NW margin of the showing, a north-northeast-striking fault contains a $\sim 2-3$ m wide zone of hydrothermal breccia comprising $\sim 1-5$ cm clasts surrounded by cockade texture Fe-rich carbonate and disseminated pyrite (Figure 4e).

Methodology

In situ U-Pb carbonate dating via laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) was carried out at the University of Portsmouth to determine the temporal evolution of fluid flow within the two study areas described above (methodology follows that described in Parrish et al. (2018) and Mottram et al. (2024)).

Table [•]	1 Summar	v of sample	locations :	for U-Ph	dating and	preliminary	age information
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Showing	Field sample	Drillhole no.	Sample depth (m)	Drillhole dip	Latitude	Longitude	Mineral analyzed	n	Date (Ma) ¹	MSWD	Average U (ppm) ²
Sulphide City	23SB-013B1	n/a	Surface	n/a	59.165569	-134.355588	Carbonate	57	56.2 ±2.5	2.0	1.4
Sulphide City	n/a	SLM22-006	396.0	50°	59.165556	-134.357505	Carbonate	45	54.0 ±6.5	1.5	0.2
Jackie	n/a	SLM22-002	95.53	90°	59.153966	-134.323381	Carbonate	33	51.0 ±4.8	1.3	0.3

 1 Quoted uncertainties are 2σ and include propagated uncertainty

² Semiquantitative estimate

Abbreviation: MSWD, mean square weighted deviation

Latitude and longitude co-ordinates use EPSG:4326-WGS 84 datum



Samples were first characterized via optical and reflected light, scanning electron microscopy (SEM) and electron dispersive spectroscopy (EDS). Carbonate minerals were targeted from a range of pre-, syn- and post-mineralization vein and fault settings (Figures 3, 4, 5). Quoted uncertainties in Table 1 and Figure 5 are 2σ and include 2% additional propagated uncertainty to account for the dispersion and long-term reproducibility of the secondary reference materials: Duffbrown limestone (64 ±2 Ma; Hill et al., 2016) and Mudtank zircon (732 ±5 Ma; Black and Gulson, 1978; Jackson et al., 2004). Where applicable, in situ U-Pb skarn garnet dating and any targeted accessory minerals will follow a similar methodology (e.g., Seman et al., 2017).

Preliminary Geochronology Results

Preliminary U-Pb carbonate dates from samples collected at the Sulphide City and Jackie showings are presented in Table 1 and Figure 5. At Sulphide City, an ~1 cm wide blocky carbonate vein crosscutting mineralized granodiorite (phase 3) yielded an age of 56.2 \pm 2.5 Ma (n = 57; MSWD = 1.3). Similarly, an ~0.5 cm wide quartz-carbonate vein crosscutting large centimetre-scale pyrite cubes at Sulphide City yielded an age of 54.0 ± 6.5 Ma (n = 45; MSWD = 1.5). At Jackie, ~ 0.2 cm wide pull-apart veins infilled with fine crystalline carbonate crosscut the foliation in garnet-biotite schist and yielded an age of $51.0 \pm 4.8 \text{ Ma} (n = 33; \text{MSWD} = 2.0)$. Approximately 30 additional carbonate samples, representing a wide range of textures within the Sulphide City and Jackie showings, are currently being analyzed, as well as skarn garnets and igneous zircon in intrusive units.

First-Order Interpretations

Combined with field and drillcore observations, the preliminary dates obtained here provide the first temporal constraints on the timing of magmatic-hydrothermal fluid activity at the Sulphide City and Jackie showings. Carbonate ages date the timing of faulting, extensional crack-seal vein formation and fluid flow (e.g., Roberts et al., 2020; Mottram et al., 2020, 2024).

The preliminary ages obtained here broadly correlate with ages for epithermal gold mineralization at the Mount Skukum mine at ca. 54 Ma (adularia ⁴⁰Ar-³⁹Ar age) and Engineer mine at ca. 50 Ma (vanadian illite ⁴⁰Ar-³⁹Ar age), and

Figure 5. Preliminary U-Pb dates from carbonate samples (see Table 1 for details and Figure 2 for map locations): **a)** U-Pb date from a coarse crystalline carbonate (CB) vein crosscutting mineralized granodiorite at Sulphide City (sample SLM22-002-95.53); **b)** U-Pb date from quartz (Qz)-carbonate vein crosscutting ~3 cm pyrite (Py) crystals at Sulphide City (sample SLM22-006-396.0); **c)** U-Pb date from small pull-apart veins crosscutting marble host rock at Jackie (sample 23SB-013B1). All quoted uncertainties are 2ó and include propagated uncertainty. Black scale bars in all insets are 1 cm.







Figure 6: Compilation of current geochronological constraints on the Llewellyn fault–Tally Ho shear zone, deformation and metamorphism, and mineralization, northwestern British Columbia and southwestern Yukon. Dates included are from this study and [1] Bultman (1979), [2] Hart and Pelletier (1989), [3] Currie (1994), [4] Currie and Parrish (1997), [5] Love et al. (1998), [6] Mihalynuk et al. (2003), [7] Tizzard et al. (2009), [8] Ootes et al. (2018), [9] Millonig et al. (2017), [10] Dyer (2020), [11] Soucy La Roche et al. (2022) and [12] Armstrong and Werner (unpub. data, BC Geological Survey, 2006). See Figure 1b for spatial distribution of age constraints. Abbreviations: BP, Bennett Plateau; E, Engineer mine; MM, Montana Mountain; MR, Middle Ridge; MS, Mount Skukum mine.

mesothermal gold mineralization at Montana Mountain occurring after ca. 66 Ma, based on volcano-stratigraphic relationships (Figures 1, 6; Hart and Pelletier, 1989; Love et al., 1998; Millonig et al., 2017). Combined with existing regional ages, results indicate the widespread and protracted relationship between brittle deformation occurring along the Llewellyn fault, the intrusion of several overprinting phases of porphyritic granodioritic-tonalitic units, and a wide variety of mineralization styles.

Future Work

The next phase of this project will focus on combining several additional geochronometers, including zircon, garnet, titanite and apatite, spanning from igneous to hydrothermal temperatures, to thoroughly document the temporal evolution of the complex and interrelated igneous and mineralizing processes at the Blue property. U-Pb zircon dating of the major igneous phases, including the mineralized tonalite and granodiorite, will provide absolute constraints for the timing of igneous activity. The next steps for this project will therefore include analysis of additional carbonate and garnet samples, and the improvement of precision, where possible, for existing ages by carrying out more analyses to locate relatively high-U domains. Ongoing SEM and EDS imaging will be used to better characterize microstructures and minor mineral phases present in key samples. This imaging will also aid in identifying the textural context of additional accessory minerals that may be dated, including titanite, apatite and/or molybdenite. In addition, trace element analysis of dated carbonate and garnet crystals will aid in linking these minerals to mineralization processes, and provide context for differentiating fluid flow events. Comparing the age of intrusive units, skarn garnet growth, and carbonate vein and breccia crystallization will be critical for temporally constraining the rate and duration of both porphyry and skarn mineralization, as well as faulting events at the study sites.

Acknowledgments

The authors thank M. Shtereva, D. Šilerová, R. Martin, and L. Nania for their assistance in the field in 2023 and 2024, and G. Long, J. Dunlop and G. Chapman for laboratory support at the University of Portsmouth. S. Porter, L. Piccott, N. Rodway, S. Coutts, K. Smallwood and the rest of the Core Assets team are thanked for their help, both in the core library and at field sites. Thanks also to J. Luck and J. Cleveland of Interior Helicopters Ltd. for their expert helicopter support, and M. Mihalynuk for helpful discussion in the field. The authors gratefully acknowledge funding for this project provided by the University of Portsmouth and Core Assets Corp. (to S.B. and C.M.), the GEM-GeoNorth program of the Geological Survey of Canada (to D.K.), and NSERC PGS-D, Geoscience BC and SEG PhD scholarship programs (to S.B.). J. Darling and J. Cutts are thanked for their constructive reviews of this manuscript.

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Platinum-Group Element and Gold Deportment in the Kwanika Copper-Gold±Molybdenum Porphyry System, North-Central British Columbia (Parts of NTS 093N/06, 11)

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Karatas Ahmadli, C., Harraden, C., Barker, S.L.L., Lang, J.R. and Manor, M.J. (2025): Platinum-group element and gold deportment in the Kwanika copper-gold±molybdenum porphyry system, north-central British Columbia (parts of NTS 093N/06, 11); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 27–34.

Introduction

The demand for critical minerals essential for clean energy technologies like wind turbines, solar panels, electric vehicles and battery storage is increasing due to the climate emergency, and the global aim for decarbonization by 2050 (Dallas et al., 2021). In Canada, platinum-group elements (PGEs), listed as critical minerals, play a crucial role in supporting clean-energy technologies, advancing a net-zero future, and in manufacturing aircraft turbines, industrial catalysts and electronics (Bossi, 2012; Natural Resources Canada, 2022). Primary economic PGE mineralization is presently extracted from stratiform chromitite reefs located within ultramafic-mafic intrusions formed between the late Archean and early Proterozoic (Maier, 2005). However, PGEs, primarily palladium, are also recovered as byproducts from porphyry copper deposits (John and Taylor, 2016), such as at the Bingham Canyon mine, in the United States (Krahulec, 2018), and Kal'makyr mine, in Uzbekistan (Pas va et al., 2010).

In this paper, the deportment of PGEs and gold in the Kwanika copper-gold±molybdenum porphyry deposit in British Columbia (BC) was investigated, with a particular focus on the potential of palladium to add value as a by-product through metallurgical processes. The Kwanika porphyry deposit, located in the central Canadian Cordillera, is a Late Triassic to Early Jurassic calcalkaline system comprising the Central Zone (copper-gold-silver) and the South Zone (copper-gold-silver-molybdenum). Palladium

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mineralization was first identified by NorthWest Copper Corp. in 2022 but the formation mechanisms and deportment characteristics of palladium and gold remain poorly understood. The aims of this study are to determine the primary host mineral(s) for palladium and gold, identify associated mineralogy affecting metallurgical deportment, and investigate geological factors controlling palladium and gold precipitation in the deposit.

Background Information

Platinum-Group Elements (PGEs)

Platinum-group elements, which are referred to as platinum-group metals in engineering and metallurgy, are a group of six metallic elements: platinum, palladium, rhodium, ruthenium, iridium and osmium (Gunn, 2014). The PGEs rarely occur in nature as native metals but typically form alloys with each other or other metals like iron, tin, copper, lead, mercury and silver. The PGEs can also bind with metalloids (such as antimony, arsenic, bismuth and tellurium) and nonmetals (such as selenium and sulphur). They also exist in solid solutions within base-metal sulphides, where PGEs substitute for other metals within the crystal lattice without altering the mineral structure (Zientek et al., 2017). Minerals containing one or more PGE are called platinum-group minerals, with over one hundred recognized types (Cabri, 2002; Gunn, 2014). Platinum-group minerals in rocks are typically very fine grained, varying in size from submicron to hundreds of microns in diameter (Zientek et al., 2017).

PGEs in Porphyry Deposits

Porphyry copper deposits around the Pacific rim, the Mediterranean and Carpathian regions of Europe, and the Al-

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pine-Himalayan mountain system may contain PGEs as byproducts, with some economic advantage for the primary producer (Eliopoulos et al., 2014; John and Taylor, 2016). Elevated concentrations of combined PGEs (>5 ppm palladium+platinum) have been identified in high-grade bornite-chalcopyrite ores and/or flotation concentrates from several alkaline porphyry deposits in various regions worldwide (Economou-Eliopoulos, 2005). Platinumgroup element deportment is crucial for the recovery of PGEs as a byproduct, as it facilitates the design of the most appropriate flowsheet (Sahu et al., 2021). In BC, Thompson et al. (2002) reported elevated palladium and platinum contents at five alkaline porphyry deposits formed in island-arc tectonic settings: Galore Creek, Lorraine, Mount Milligan, Mount Polley and Ajax East. The heavy mineral concentrates they studied comprise a mixture of sulphides (bornite, chalcopyrite and pyrite) and oxides (magnetite and hematite). The PGE mineralization, especially palladium and platinum, in copper-gold porphyry deposits in the BC Cordillera is linked to postsubduction, alkali-enriched magmatism in the upper crust, with hydrothermal processes enriching palladium, gold, copper and tellurium from porphyry igneous rocks into hydrothermal deposits (Holwell et al., 2019).

Geological Setting

The Kwanika porphyry deposit is hosted in the Hogem batholith at the western margin of the Quesnel terrane. The Quesnel terrane consists of Late Triassic to Early Jurassic volcanic, sedimentary and plutonic rocks (Nicola and Takla groups) and hosts numerous porphyry copper±gold±molybdenum±silver deposits (Logan and Mihalynuk, 2014). The Hogem batholith is a composite plutonic body that contains calcalkalic and alkalic suites, as well as Alaskantype ultramafic-mafic intrusions, and hosts several coppergold porphyry deposits (Osatenko et al., 2020). In the project area, Late Triassic volcanic and sedimentary rocks of the Takla Group are cut by several suites of Late Triassic, Early Jurassic and Middle Jurassic plutons of the Hogem suite (Garnett, 1978; Woodsworth et al., 1991). Mineralization within the Kwanika property is associated with an Early Jurassic quartz monzonite phase of the Hogem batholith (Osatenko et al., 2020). The Central Zone (copper-gold-silver) hosts mineralization within diorite and monzodiorite porphyritic intrusions, whereas the South Zone (copper-gold-silver-molybdenum) exhibits structurally controlled mineralization in equigranular quartz monzonite to quartz monzodiorite and breccia bodies (Ausenco Engineering Canada Inc., 2023). A more detailed geological setting is provided in Osatenko et al. (2020); Figure 1 illustrates the local geology and the spatial relationships between these zones.

The Central Zone (Figure 1) comprises Takla Group andesites and Early Jurassic Hogem batholith intrusions, both cut by quartz monzonite porphyry hosting the highest grade copper-gold mineralization. Hydrothermal alteration includes an inner potassic core, an outer potassic shell that transitions into a peripheral propylitic zone and a patchy sericite alteration covering other alteration types (Osatenko et al., 2020). Palladium concentrations from fire assay analyses done on core samples 2 m long range from 0.5 to 1655 ppb (average: 17.65 ppb), with fire-assay subsample concentrations of up to 5280 ppb. Gold concentrations range from 0.5 to 12 700 ppb (average: 281 ppb), reaching up to 3180 ppb in fire-assay subsamples. Figure 2 shows a representative sample from the Central Zone used in the analysis.

In the South Zone (Figure 1), mineralization is hosted mainly in an equigranular quartz monzonite intrusion of the Hogem batholith, showing potassic, minor propylitic and sericite alteration. The mineralization in the South Zone is younger than that in the Central Zone (Osatenko et al., 2020). Based on fire assay analyses done on core samples 2 m long, palladium concentrations range from 0.5 to 73 ppb, with an average of 4.60 ppb. Gold concentrations range from 0.5 to 856 ppb, with an average of 32 ppb. Figure 3 shows a representative sample from the South Zone used in the analysis.

Ultramafic rocks, mainly hornblende pyroxenite (Figure 4), have been drilled in various locations across the Kwanika property, including north-northwest of the Central Zone (Figure 1), west of the West fault, and east of the South Zone (NorthWest Copper Corp., 2022; Lang, 2023). These ultramafic rocks of unknown age show weak alteration with chlorite, epidote and hematite after magnetite but lack copper-gold or molybdenum mineralization (Lang, 2023). Within the ultramafic rocks, palladium concentrations range from 0.5 to 410 ppb (average of 37.47 ppb), whereas gold concentrations determined from fire assay analyses done on core samples 2 m long range from 0.5 to 108 ppb (average of 4.5 ppb).

Methodology

In this study, palladium and gold deportment in three distinct geological zones were examined using mineralogical characterization (optical petrography, micro X-ray fluorescence [μ -XRF], scanning electron microscope [SEM], electron probe microanalysis [EPMA]), geochemistry and heavy mineral separation methods. Heavy mineral separation was conducted with bench-top tests performed at the Coal and Mineral Processing Laboratory of The University of British Columbia to determine mineralogy, grain size and deportment of palladium- and gold-bearing minerals, producing copper, pyrite and magnetic concentrates. All concentrates and head assay samples were analyzed using aqua-regia partial digestion followed by inductively coupled plasma–mass spectrometry (ICP-MS), fire assay fol-




Figure 1. Geology of the Kwanika porphyry deposit area (location shown by red dot on inset map; modified from Osatenko et al., 2020). Due to overburden, the geology is projected to the surface based on drillcore data. Green circles represent drillholes selected for this study, blue circles indicate drillholes with palladium concentrations determined from fire assay analyses done on core samples 2 m long and white circles show drillholes where no analysis was conducted for palladium. Figure from Porphyry Deposits on the Northwestern Cordillera of North America: A 25 Year Update, page 489. Reprinted and revised with permission of the Canadian Institute of Mining, Metallurgy and Petroleum. Base map was created using ArcGIS[®] software by Esri. ArcGIS[®] is the intellectual property of Esri and is used herein under license. Copyright © Esri. All rights reserved. For more information about Esri[®] software, please visit https://esri.ca/. All co-ordinates are in UTM Zone 10N, NAD83.





Figure 2. Representative half-core sample of monzonite cut by quartz veins from the interval with the highest palladium concentration (1655 ppb), based on fire assay analyses done on core samples 2 m long, from the Central Zone of the Kwanika porphyry deposit. The red outline indicates the sampling location of a thin section collected for scanning electron microscope analysis. Abbreviations: Cal, calcite; MON, monzonite; Qz, quartz.



Figure 3. Representative half-core sample of quartz monzonite cut by molybdenitequartz vein from the interval with the highest palladium concentration (73 ppb), based on fire assay analyses done on core samples 2 m long, in the South Zone of the Kwanika porphyry deposit. The red outline indicates the sampling location of a thin section collected for scanning electron microscope analysis. Abbreviations: Mol, molybdenite; QMON, quartz monzonite.



Figure 4. Representative half-core sample of hornblende pyroxenite ultramafic rock, with magnetite and hematite veins from the interval with the highest palladium concentration (410 ppb), based on fire assay analyses done on core samples 2 m long, from the Kwanika porphyry deposit. The red outline indicates the sampling location of a thin section collected for scanning electron microscope analysis. Abbreviations: Hem, hematite; HORN. PYX, hornblende pyroxenite; Mag, magnetite.

lowed by inductively coupled plasma–optical emission spectrometry (ICP-OES), and four-acid near-total digestion followed by ICP-OES+ICP-MS. Additional reporting on heavy mineral separation methods is ongoing.

For petrographic and SEM analysis, grain mounts (<106 µm) were prepared using the copper concentrate (792 ppb palladium and 13 000 ppb gold) from the Central Zone, copper concentrate (1520 ppb palladium and 396 ppb gold) from the South Zone, and high-intensity magnetic concentrate (360 ppb palladium, 52 ppb platinum and 9 ppb gold) and tailings of the ultramafic rocks (357 ppb palladium, 85 ppb platinum and 15 ppb gold) due to their elevated palladium and platinum concentrations. Thin sections were taken from core samples with elevated palladium and gold concentrations based on fire assay analyses done on core samples 2 m long. Thin sections from the Central and South zones were prepared from half-core samples exhibiting sulphide mineralization and quartz veins, whereas in the ultramafic rocks, the presence of magnetite or hematite was also a factor considered for sample collection. The SEM analysis was performed on four grain mounts and four thin sections, the locations of three of which are shown on Figures 2-4, where possible palladium and gold minerals had been observed during microscopic examinations. The SEM imaging was conducted in the Electron Microbeam and X-ray Diffraction Facility of The University of British Columbia using a SU3900 scanning electron microscope fitted with Advanced Mineral Identification and Characterization System software, both developed by Hitachi High-Tech Canada, Inc. A brightphase search method was used to identify platinum-group minerals and gold-bearing minerals, with thin sections and grain mounts analyzed at different resolutions (0.72, 0.63 and 0.58 micron/pixel). Furthermore, the major-element composition of a complex platinum-palladium grain identified through SEM analysis in the grain mount of tailings from the ultramafic rocks was determined using a JEOL JXA-iHP200F field emission electron probe microanalyzer with Bruker 30 mm² dual silicon drift detector-energy dispersive spectrometry. Table 1 lists plati-



num-group minerals and gold-silver minerals identified through SEM analysis, corresponding to the details of the platinum-group minerals displayed in Figures 5–7.

Results

In the Central Zone, minerals containing palladium are found both enclosed in quartz, chalcopyrite and pyrite, and at the boundaries between these minerals. The size of palladium minerals in the Central Zone ranges from 0.8 to $341.1 \ \mu m^2$. Minerals containing gold and silver were primarily within or at the boundaries of chalcopyrite, pyrite and bornite, and less commonly found within quartz. The gold-silver minerals in the Central Zone range in size from 0.4 to $70.0 \ \mu\text{m}^2$. Gold-silver minerals with over 75% gold are classified as native gold (Alguacil, 2006), whereas some selenium-bearing types are identified as fischesserite (Bindi and Cipriani, 2004). Gold and silver tellurides were also detected but did not match known mineral compositions. Backscattered electron (BSE) images of these minerals are shown in Figure 5a–f.

In the South zone, a grain of merenskyite 0.7 μ m² in size was identified at the quartz-chalcopyrite boundary (Figure 6a), and an unknown mineral containing silver and palladium of 1.2 μ m² was found enclosed in biotite (Figure 6b). The elemental composition of the silver-palladium mineral does not match any known silver-bearing palla-

Table 1. Platinum-group minerals and gold-silver minerals identified in the Kwanika porphyry deposit using scanning electron microscopy. Sperrylite and isomertieite in the ultramafic rocks share an unclear boundary; the area (32.6 μm²) represents their total combined area.

Area	Hole number	Sample depth (m down hole)	MDRU Lab ID	Area (µm²)	Host mineral	Host type	Mineral name	Figure label
Central Zone	K-21-206	469.6	BPKW0031BT	341.1	Ccp-Qz On boundary Merenskyite (PdTe		Merenskyite (PdTe ₂)	5a
Central Zone	K-21-211	36.7	BPKW0033BT	232	Qz Enclosed Kotulskite		Kotulskite (PdTe)	5b
Central Zone	K-21-206	469.6	BPKW0031BT	4.7	Ccp-Qz	On boundary Isomertieite (Pd ₁₁ Sb ₂ A		5c
Central Zone	K-21-211	36.7	BPKW0033BT	7.6	Сср	On boundary	Fischesserite (Ag ₃ AuSe ₂)	5d
Central Zone	K-21-211	36.7	BPKW0033BT	6	Сср	On boundary	Merenskyite (PdTe ₂)	5d
Central Zone	K-22-242	476	BPKW0281M05B	5	Py	On boundary	Native gold (>75%)	5e
Central Zone	K-21-206	469.6	BPKW0031BT	26.04	Qz	Enclosed	Native gold (>75%)	5f
Central Zone	K-21-206	469.6	BPKW0031BT	30	Qz	Enclosed	Hessite (Ag ₂ Te)	5f
South Zone	K-22-250	200.2	BPKW0197T	1.2	Bt	Enclosed	-	6a
South Zone	K-22-250	199.8	BPKW0289M05B	0.7	Ccp-Qz	On boundary	Merenskyite (PdTe ₂)	6b
Ultramafic rocks	K-22-247	477.6	BPKW0184BT	17.1	Срх	Enclosed	Temagamite (Pd ₃ HgTe ₃)	7a
Ultramafic rocks	K-22-247	477.6	BPKW0184BT	46.6	Срх	Enclosed	Kotulskite (PdTe)	7a
Ultramafic rocks	K-22-247	477.6	BPKW0184BT	32.6	Срх	On boundary	Sperrylite (PtAs ₂)	7b
Ultramafic rocks	K-22-247	477.6	BPKW0184BT	52.0	Срх	On boundary	Isomertieite (Pd ₁₁ Sb ₂ As ₂)	7b
Ultramafic rocks	K-22-247	476	BPKW0295M06B	33.5	Срх	On boundary	Pd-Pt mineral	7c

Abbreviations: Bt, biotite; Ccp, chalcopyrite; Cpx, clinopyroxene; MDRU, Mineral Deposit Research Unit; Py, pyrite; Qz, quartz.



Figure 5. Scanning electron microscope (SEM) backscattered electron (BSE) images of palladium minerals and gold-silver minerals from the Central Zone of the Kwanika porphyry deposit: **a**) merenskyite (PdTe₂) and enargite (Cu_3AsS_4) at the boundary between chalcopyrite and quartz; **b**) kotulskite (PdTe) enclosed in quartz; **c**) isomertieite ($Pd_{11}Sb_2As_2$) at the boundary between chalcopyrite and quartz; **d**) fischesserite (Ag_3AuSe_2) and merenskyite on a chalcopyrite boundary; **e**) native gold (>75% gold) on a pyrite boundary; **f**) native gold adjacent to hessite (Ag_2Te) enclosed in quartz. Abbreviations: Ccp, chalcopyrite; Eng, enargite; Py, pyrite; Qz, quartz.



Discussion

dium minerals from the literature. Backscattered electron images of these minerals are shown in Figure 6.

In the ultramafic rocks, palladium minerals, ranging in size from 0.4 to 46.6 μ m², were primarily enclosed in clinopyroxene. Minerals containing platinum were also identified, with two grains containing both palladium and platinum minerals (32.6 μ m² and 33.5 μ m², respectively) at the boundary with clinopyroxene (Figure 7b, c). Electron probe microanalysis (EPMA) was conducted on one of these grains to further analyze its composition (Figure 7d). Backscattered electron images of these minerals are shown in Figure 7a–c.

The distribution of precious minerals across different geological zones provides valuable insights into the mineralization processes within the project area. A comparison of palladium- and platinum-bearing minerals identified in this study with those discussed in the literature (Cabri, 2002) reveals five palladium-bearing minerals and one platinumbearing mineral. Table 1 summarizes the occurrence of platinum-group minerals across geological zones. Two distinct styles of PGE mineralization are observed in the Kwanika deposit. The first is palladium dominant, occurring in monzonite and quartz monzonite hostrocks in the Central and South zones. Elevated tellurium levels, along



Figure 6. Scanning electron microscope (SEM) backscattered electron (BSE) images of merenskyite and a silver-palladium mineral from the South Zone of the Kwanika porphyry deposit: **a)** merenskyite (PdTe₂) at the boundary between chalcopyrite and quartz; **b)** silver-palladium mineral enclosed in biotite. Abbreviations: Ag, silver; Bt, biotite; Ccp, chalcopyrite; Pd, palladium; Qz, quartz.



Figure 7. Scanning electron microscope (SEM) backscattered electron (BSE) images showing palladium and platinum minerals within ultramafic rocks (primarily associated with clinopyroxene) from the Kwanika porphyry deposit, including temagamite (Pd_3HgTe_3) and kotulskite (PdTe) enclosed in clinopyroxene (**a**), isomertieite ($Pd_{11}Sb_2As_2$) and sperrylite ($PtAs_2$) on the boundary with clinopyroxene (**b**), a palladium-platinum mineral on the boundary with clinopyroxene (**b**), a palladium-platinum mineral on the boundary with clinopyroxene (**c**) and an electron probe microanalysis (EPMA) image of the palladium-platinum mineral from image (**c**) showing isomertieite ($Pd_{11}Sb_2As_2$), sperrylite ($PtAs_2$) and palladoarsenide (Pd_2As) in a complex grain (**d**). Abbreviations: Cpx, clinopyroxene; Pd, palladium; Pt, platinum.



with the occurrence of platinum-group minerals at sulphide-silicate boundaries or enclosed within hydrothermal quartz and sulphides, suggest a process similar to the semimetal collector mechanism proposed for the Skouries porphyry deposit (McFall et al., 2018); this mechanism likely facilitated the transport and concentration of PGEs in these zones. Furthermore, a significantly higher abundance of gold-silver minerals in the Central Zone and the association of platinum-group minerals with gold-silver minerals in the Central and South zones suggest that PGEs, gold and silver were transported by the same telluride melt. The second PGE style is palladium and platinum dominant, confined to hornblende pyroxenite, with little to no sulphide mineralization. Absence of gold-silver minerals, low tellurium content and platinum-group minerals (kotulskite, isomertieite, sperrylite, temagamite and palladoarsenide) enclosed in clinopyroxene indicate a magmatic origin.

Conclusion

The Kwanika porphyry deposit exhibits an unusual enrichment in platinum-group elements, characterized by three distinct geological zones. Six types of platinum-group minerals that average 22 μ m² in size are identified in the deposit. Two distinct styles of platinum-group element mineralization are identified; palladium dominant in the Central and South zones, and palladium and platinum dominant in the ultramafic rocks. This study constitutes part of a broader investigation into platinum-group element and gold deportment in the Kwanika deposit. Ongoing research will focus on the geological factors controlling palladium and gold precipitation, and the interplay between these two styles of platinum-group element mineralization. Future studies will also address the potential for lattice-hosted platinum-group elements.

Acknowledgments

The authors would like to thank NorthWest Copper Corp. and the Mineral Deposit Research Unit of The University of British Columbia for their financial support of this project. The lead author is also grateful to Geoscience BC for its financial support through the Geoscience BC Scholarship program. Special thanks to M.A. Rodriguez Mustafa for reviewing the manuscript.

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Carbonatite-Associated Rare-Earth-Element Mineralization in the Eastern Part of the Ice River Carbonatite Complex, Southeastern British Columbia (NTS 082N/01)

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Bhakta, J.D., Chakhmouradian, A.R. and Brown, J.A. (2025): Carbonatite-associated rare-earth-element mineralization in the eastern part of the Ice River Carbonatite Complex, southeastern British Columbia (NTS 082N/01); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 35–46.

Introduction

Carbonatites are mantle-derived, silica-poor igneous rocks with >50 vol. % primary carbonate minerals (Yaxley et al., 2021). Despite being silica-poor, the modal ratio of carbonate to silicate minerals is highly variable in carbonatites (Christy et al., 2021) and is dependent on how their parental magma is generated and evolves before emplacement. Carbonatites are of significant interest to exploration companies due to their high concentrations of rare-earth elements (REEs). As carbonatite magma evolves, REEs and other incompatible elements concentrate within residual melts (Anenburg et al., 2021). Anorogenic, rift-related settings facilitate the emplacement of small-volume magmas produced by decompressional melting of the underlying mantle. This allows the emplacement of alkali-rich, silicaundersaturated magmas (which can crystallize and form rocks such as carbonatite, nephelinite and phonolite) within large-scale, intraplate extensional structures (Chakhmouradian and Zaitsev, 2012). Examples of such anorogenic alkaline rock-carbonatite igneous provinces include the Kola Peninsula of Russia and the East African Rift system. Carbonatites and alkaline rocks in orogenic settings may also host concentrated REEs (Song et al., 2016). In orogenic settings, carbonatite emplacement occurs either during postorogenic collapse or before a transition from a convergent to divergent tectonic setting (Yaxley et al., 2022). The cooling and crystallization of carbonatite magmas are accompanied by the release of alkali-rich fluids, which metasomatically alter (fenitize) surrounding country rocks (Le Bas, 2008). Fluids derived from carbonatite magmas contain elevated levels of alkalis and volatiles, which can transport incompatible elements like REEs (Elliott et al., 2018). Therefore, metasomatically altered rocks (fenites) can in principle provide information on the magmas from which the fluids were sourced and any associated mineral deposits.

To form economic concentrations of REEs in igneous rocks, some sort of evolutionary process (liquid immiscibility, fractional crystallization or the release of fluids from magma) is required (Chakhmouradian and Zaitsev, 2012). Furthermore, carbonatites are often affected by hydrothermal fluids, which facilitate the development of late-stage REE mineralization. The interaction of primary magmatic REE-bearing minerals (e.g., burbankite and carbocernaite) with carbonate-, chloride-, fluoride- or phosphate-rich hydrothermal fluids enables their dissolution and precipitation of secondary REE phases (Williams-Jones et al., 2012; Anenburg et al., 2021). The presence of fenitized rocks and REE-bearing minerals, such as bastnäsite, parisite, synchysite, Ba-REE fluorocarbonate and monazite, indicate REE enrichment due to hydrothermal interaction (Wall et al., 2008; Chakhmouradian and Wall, 2012; Hoshino et al., 2016). As with hydrothermal dissolution-reprecipitation, weathering can also be responsible for the breakdown of primary REE phases allowing the transport of REEs into later formed REE phases (Anenburg et al., 2021). Because carbonatites are susceptible to subsolidus re-equilibration (Chakhmouradian et al., 2016), understanding how the mineralogy of a given carbonatite has been affected by postmagmatic processes is critical for determining its mineralization potential.

Ice River Complex

The lead author's graduate research is focused on a carbonatite-associated REE exploration prospect in the eastern aspect of the Ice River Complex (IRC). The prospect is located 42 km southeast of Golden, in the British Columbian portion of the Vermillion Range, where it is sandwiched between, and bordered by, Yoho and Kootenay National parks. Since 2003, Eagle Plains Resources has been conducting exploration on the property. TerraLogic Exploration Inc. (on behalf of Eagle Plains Resources) ran six exploration projects (in 2006–2010 and 2012) to ad-

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vance their understanding of the bedrock geology of the IRC within its contact aureole (Brown and Holowath, 2022). The IRC is a late Devonian to early Carboniferous alkaline intrusion, which has a J-like outline in plan view and measures approximately 18 km in length and 29 km² in total exposure. A geological map and description of the complex are provided in Figure 1. Weakly metasomatized Cambrian and Ordovician limestone, dolomite and shale country rocks (Chancellor and Ottertail formations and McKay Group) host the intrusion. The oldest parts of the intrusion (hereafter referred to as the layered series) comprise a rhythmically stratified, feldspar-poor, ultramafic to mafic series ranging in modal composition from jacupirangite to urtite. The layered series is crosscut by feldspathoid syenites (hereafter referred to as the syenite series) and carbonatite, which are in turn crosscut by lamprophyre dikes (Locock, 1994; Peterson and Currie, 1994; Mumford, 2009; Brown and Holowath, 2022). Emplacement of the IRC is believed to have resulted in the formation of a contact metasomatic skarn at the margins of the complex (Currie, 1975).

During the Laramide orogeny, the IRC and surrounding hostrocks were thrust approximately 200 km to the east, along a west-dipping décollement surface (Gabrielse, 1991). The IRC behaved as a relatively competent mass during transport, whereas the surrounding hostrocks were strongly sheared (Locock, 1994). Marginal units of the IRC show strong deformation (e.g., gneissosity and boudinage; Mumford, 2009). Both the IRC and hostrocks were affected by regional metamorphism of prehnite-pumpellyite facies with pressures of 2–9 kbar and temperatures of 200–400 °C (Currie, 1975; Locock, 1994). Gabrielse (1991) suggested that regional metamorphism was coeval with the eastward thrusting of the IRC. There is a large diversity of carbonatite types in the IRC, differing in petrography and structural characteristics. The following description is a summary of previous work on these rocks. Carbonatites occur as dikes and plugs, intruding alkaline igneous, silicate igneous and metasedimentary hostrocks (Currie, 1975; Peterson and Currie, 1994). The carbonatites vary in colour from white to almost black. Currie (1975) identified three carbonatite bodies within the IRC. The largest one is a lenticular plug (400 by 1500 m) of white- to buff-weathering calcite carbonatite (unit 10, Figure 1) containing minor aegirineaugite, fluorapatite and phlogopite and accessory pyrochlore, pyrite, strontianite and ilmenite. This largest carbonatite body is visible east of Aquila Mountain on Figure 1. Calcite-carbonatite dikes (thin, elongate bodies of unit 10, Figure 1) commonly intrude the layered series (units 5–9, Figure 1) and the syenite series (units 11–15, Figure 1; Pell, 1987). Associated with the large calcitecarbonatite body are black-weathering dikes with major calcite, siderite and berthierine; also present are minor biotite, aegirine, edingtonite, perovskite and ilmenite, as well

as trace quantities of sphalerite and pyrite (Locock, 1994). A younger red-weathering carbonatite is mineralogically similar to the black-weathering carbonatite, but contains up to 10 vol. % noncarbonate minerals, including serpentine, pyrochlore and xenotime (Pell, 1987). In addition, carbonatites with ropy textures, possibly indicating ductile deformation, were reported by Currie (1975). More silicarich, massive, dark green carbonatites have also been documented, but no detailed information on their mineralogy is currently available.

In the eastern part of the IRC (the Moose Creek valley [MCV], outlined in purple in Figure 1), the carbonatites have complex, poorly understood structural and temporal relationships to the syenites (Mumford, 2009); carbonatite and syenite bodies locally crosscut each other. This ambiguous structural relationship may be due to multiple phases of carbonatite emplacement (Pell, 1987). It has also been proposed that syenite-associated carbonatites are related to the red-weathering carbonatite described above (Peterson, 1983). Syenite-associated carbonatites are distinct due to low abundances of silicate minerals in these rocks, and the presence of ankerite, barytocalcite and strontianite in addition to calcite (Peterson, 1983; Pell, 1987). Associated silicate minerals in syenite-associated carbonatites include phlogopite, alkali feldspars (dominantly albite with minor microcline) and zeolites (natrolite, analcime and minor edingtonite; Pell, 1987). Whole-rock geochemical analyses of the carbonatites show variable Nb and REE contents (Pell, 1987). The syenites are relatively more enriched in Nb, but relatively less enriched in REE in comparison with the carbonatites (Brown, 2013).

The aims of this study are to 1) describe the mineralogy and paragenetic characteristics of the different carbonatites in the MCV, with an emphasis on REE-bearing phases and 2) identify distinct mineralization styles, which correlate to geochemical variations and anomalies in the hostrocks. Using new analytical data, an attempt will be made to constrain the source of REE mineralization in the eastern part of the IRC and identify the relationships between mineralization styles and carbonatite interaction with their hostrocks.

Methodology

Identification of paragenetic sequences is fundamental in understanding the REE-mineralization styles in the IRC and has implications for the petrogenesis of the complex. For this study, there is a particular emphasis on the identification of REE-bearing minerals and the processes that have led to their formation. An in-depth understanding of their compositional and paragenetic characteristics is essential to determining the economic potential of the IRC. For this purpose, 57 rock samples were collected during the 2022 field program. Sample sites were selected based on field



Figure 1. Geology of the Ice River Complex (IRC) showing the major units in the complex. The study area (the extent of Eagle Plains Resources' [EPL] tenure) is shown as a black outline along the eastern margin of the complex. The area of the Moose Creek valley (MCV), as considered for this study, is outlined in purple. The original field mapping was undertaken by Currie (1975), later digitized by TerraLogic Exploration Inc. Geological data from Cui et al. (2017) has been added. BC MINFILE entries: Moose Creek, MINFILE 082N 027; Shining Beauty, MINFILE 082N 025; Waterloo, MINFILE 082N 028; Yippie, MINFILE 082N 097; Zinc Creek, MINFILE 082N 026 (Brown and Holowath, 2022; BC Geological Survey, 2024). All co-ordinates are in UTM Zone 11N, NAD 83. Legend on the following page.

Age	Subgroup	Map unit	Key	Unit description (modified from Currie, 1975)				
Quaternary	Unconsolidated sedimer		17	Glacial sediments, fluvial sediments, talus rock flour and related materials				
Ice River Alkaline Complex units								
Devonian or Carboniferous		Lamprophyre (minette and sannaite)	16	Melanocratic rocks occasionally ocellar, with phenocrysts of pyroxene and kaersutite, sometimes in a feldspar-bearing matrix				
		Altered zeolite-rich syenite	15	Buff-weathering rocks rich in secondary natrolite with minor carbonate and aegirine				
		Sodalite-nepheline syenite		Massive to faintly gneissic leucocratic green to blue rocks, including veins and dikes				
Devonian or	Syenite series	Leucocratic nepheline syenite	13	Pale greenish, distinctly gneissic rock commonly bearing kaersutite and sodalite				
Carboniferous		Meso- to melanocratic syenite	12	Migmatite and agmatite gneiss				
		Silica-saturated contact breccia	11	Pale, fine-grained hybrid rocks, commonly saturated, with metasediment blocks, includes minor amounts of units 12, 13				
Devonian or Carboniferous	Carbonatite	Buff-weathering carbonatite	10	White to buff, foliated carbonate and massive, altered, melanocratic silicocarbonatite				
	Layered series	Urtite	9	Coarse-grained to pegmatitic pale green rocks				
		Ijolite	8	Medium- to coarse-grained bluish rocks with cubic nepheline set in euhedral pyroxene lathes				
Devonian or		Titaniferous andradite ijolite	7	Medium- to coarse-grained brownish-black rocks with cubic nepheline, vitreous brown garnet				
Carboniferous		Mela-ijolite	6	Coarse-grained blue-black rocks with prominent biotite phenocrysts and clumps				
		Jacupirangite		Black to greenish pyroxenite with small cubes of nepheline, and minor biotite				
		Ice River Alkaline Compl	ex ho	strock units				
Devonian or Carboniferous		Contact skarn	4	Banded skarn and calcsilicate rocks, contact- altered sedimentary rocks formed during emplacement of the complex				
Cambrian to Ordovician		McKay Group	3	Limy shales with interbanded thin limestones generally grey-green with local brownish stripings and dolomite layers				
Cambridge		Ottertail Formation	2	Massive grey-blue limestone, shaly horizons toward the base and top				
Camorian		Chancellor Formation	1	Reddish slaty shale, commonly thinly fissile, minor limy horizons and dolomite				

Figure 1 (continued). Legend for geology of the Ice River Complex, showing the major units in the complex.



observations and the results of a scintillometer-assisted survey (Brown and Holowath, 2022), which targeted anomalies outlined in Brown (2013). From this sample suite, a total of 35 polished thin sections were prepared from selected samples chosen on the basis of whole-rock geochemical data (major- and trace-element concentrations), using such geochemical indicators as total REE, Ba, Zr, Nb, Th, SiO₂ and Al₂O₃ contents. The incompatible trace elements were chosen for their association with mantle-derived magmatism and generally low levels in metasedimentary hostrocks, whereas the SiO₂ and Al₂O₃ values are useful for discriminating between carbonatites, alkaline silicate rocks and their metasomatically modified variants.

The polished thin sections were examined using a Leica Microsystems Leica DM750 P polarizing microscope to identify major constituent minerals and textural characteristics. The sections were then examined using a variety of analytical techniques, including cathodoluminescence (CL) and backscattered-electron (BSE) imaging, energy dispersive X-ray spectrometry (EDX) and wavelength-dispersive X-ray fluorescence (WDXRF), and Raman vibrational microspectroscopy (Raman). By using a combination of imaging and quantitative techniques, a comprehensive understanding of textural (BSE and CL) and compositional (EDX, WDXRF, Raman) characteristics of the selected sample suite was achieved.

Cold-cathode CL imaging is used for luminescent minerals (such as calcite, apatite and fluorite; Mitchell et al., 2017) to identify trace-element-induced compositional variations and zonation, which would be undetectable in BSE images, including variations arising from REE substitutions in these minerals. In tandem with CL imaging and EDX, BSE imaging was used to elucidate interrelations among different minerals, and in particular, microtextural characteristics beyond the resolution of an optical microscope (such as exsolution lamellae and symplectic intergrowths). Optical, CL and BSE imaging were also used to identify and map representative areas of interest for further quantitative analysis.

Quantitative microbeam techniques enable major- and trace-element compositional variations to be determined in situ and with a spatial resolution of 10–30 m. A nondestructive microbeam technique capable of detecting major and minor elements in mineral grains at concentrations >200 ppm is WDXRF. For accurate petrographic characterization of all rock types in this study, WDXRF was used to determine the major-element composition of all constituent minerals (feldspars, clinopyroxenes, calcite, etc.) and REE phases. Raman was used to identify polymorphs and compositionally similar phases that cannot be reliably discriminated using a combination of EDX and CL imaging. To enable accurate beam positioning, matching BSE and reflected-light microscopic images and maps of target areas were used.

Moose Creek Valley

Structure

Igneous rock exposures in the MCV are abundant at higher elevations in the western flank of the valley, where they are hosted within the Ottertail Formation and McKay Group. Igneous rock exposures in the MCV generally dip gently westward. Throughout the MCV, three phases of deformation are recognized. Folding F₁ is characterized by smallscale, tight folds around the margins of the syenite series. This deformation event was probably caused by syenite emplacement. Folding F2 represents a structurally dominant compressive event caused by the Laramide orogeny, and is characterized by open, north- to northwest-trending axial traces. In the MCV, the main exposure of F_2 is a sigmoidal-shaped, open- to isoclinal-plunging anticline, which trends north-northwest to west-northwest. Postorogenic dextral shear rotated the jacupirangite-urtite layered series, carbonatite plug and surrounding hostrock $\sim 30^{\circ}$ and is also responsible for the sigmoidal shape of the F₂ anticline and local tertiary folding (F₃; Currie, 1975).

Late-Stage Igneous Features of the MCV

Numerous syenite, lamprophyre and carbonatite dikes and sills are exposed within the MCV. Three distinct groups of intrusions were identified: 1) a nepheline syenite-nephelinolite group, 2) a syenite-monzodiorite group and 3) an alkali feldspar granite dike. The syenite dikes are typically grey, follow a north-northwest trend, and dip between 35° and 65° (Figure 2a; Mumford, 2009). Their texture is fine grained to pegmatitic and commonly inequigranular; individual bodies range from 0.20 m to several metres in width and may be traced in outcrop for ~200 m. Syenite bodies, which intrude the Ottertail Formation, generally conform to bedding in the hostrock and, therefore, can be classified as sills. Syenite intrusions are strongly deformed, which is observed as boudinage at the outcrop scale and as foliated biotite aggregates at the thin-section scale (Mumford, 2009).

Another abundant igneous rock type in the MCV is lamprophyre. It occurs as dark grey to dark green dikes ranging from 0.15 to 5 m in width. The dikes crosscut both the IRC and surrounding metasedimentary sequence, follow a northwest trend and have similar dips to the syenite intrusions (Figure 2b). Two principal varieties of lamprophyre have been identified: aphanitic to fine-grained, amphibolerich sannaite and conspicuously porphyritic, biotite-rich minette. The lamprophyres are locally deformed, as indicated by calcite-filled tension gashes and boudinage. Mumford (2009) noted isotopically indistinguishable syenite-lamprophyre dike assemblages occurring throughout





Figure 2. Photographs of late-stage igneous features in the Moose Creek valley, southeastern British Columbia: **a**) ridge between Buttress Peak and Manganese Mountain (looking south) showing syenite dikes emanating from the syenite series into surrounding Ottertail Formation limestone (unit 2, Figure 1); **b**) minette dike (outlined with yellow dashed line) on the ridgeline between Sentry Peak and Zinc Mountain (looking south) intruding meso- to melanocratic syenite (unit 12, Figure 1); **c**) carbonatite dike crosscutting carbonate metasedimentary rocks south of the ridge between Manganese Mountain and Buttress Peak (note F₃ folding in the hostrock).

the MCV, and presumably they were derived from the same source as the lamprophyres. Both lamprophyre and syenite intrusions exhibit a similar orientation with respect to the regional tectonic fabric, indicating their postemplacement rotation into the main foliation plane (S_3 ; Figure 2c).

Carbonatite Field Relationships

The least abundant intrusive rock in the MCV (both in volume and number of mapped bodies) is carbonatite. This rock is typically found as sills or dikes measuring from 0.10 to 2 m in width. However, numerous thinner crosscutting veins and veinlets comprising similar mineral assemblages are also observed in the syenites and metasedimentary hostrocks (Figure 3a-c). For this study, large carbonatite bodies and pervasively metasomatized rocks were preferentially sampled because they can be more readily recognized in the field based on their distinctive structural characteristics (such as crosscutting relationships or recessed units). Also, a scintillometer-assisted survey in the MCV showed the radiometric response from the carbonatites was low but greater than the hostrock background (Brown and Holowath, 2022). Areas of elevated radiometric response were targeted for sampling. The sampled intrusions are moderately to steeply dipping (24-90) bodies that generally follow a south-southwest trend and range from 2 to >50 m in exposed length.

Three varieties of carbonatite were identified on the basis of field, geochemical and petrographic characteristics: pale greyish dolomite carbonatite; white to light grey, saccharoidal, calcite carbonatite; and cream to pink, barytocalcite carbonatite. Oxidized carbonatite varieties are orangebrown to dark brown owing to the replacement of accessory pyrite by Fe-oxyhydroxides. Potentially related to the barytocalcite carbonatite is a fracture-hosted hydrothermal carbonate-rich assemblage observed in both calcite carbonatites and syenites. The texture of carbonatites ranges from fine to coarse grained and from subequigranular to brecciated. All samples show some textural evidence of postemplacement ductile and brittle deformation.

Mineralogy of the Eastern Ice River Complex Carbonatites

Dolomite Carbonatite

Dolomite carbonatite is interpreted to be the earliest phase of carbonatite magmatism in the eastern IRC, where it is confined to the area surrounding the Yippie mineral showing (MINFILE 082N 097, BC Geological Survey, 2024; Figure 1). Of particular note in this area is an anastomosing carbonatite-syenite sill (Figure 3a) associated with the greatest REE enrichment recorded on the property (Brown and Holowath, 2022). This carbonatite is composed predominantly of mosaic-textured, fine- to medium-grained dolomite (Figure 4a). Its most notable petrographic feature





Figure 3. Photographs of carbonatite dikes in the Moose Creek valley (MCV), southeastern British Columbia: a) carbonatite–syenite sill at the Yippie showing (MINFILE 082N 097, BC Geological Survey, 2024), associated with the greatest rare-earth-element enrichment within the study area; b) brown, oxidized, calcite carbonate dike, viewed looking northward toward Sentry Peak; c) white, calcite carbonatite breccia crosscutting the McKay Group, southeast of Mount Mollison.



Figure 4. Images of dolomite and calcite carbonatites, Ice River Complex, southeastern British Columbia: **a**) backscattered-electron (BSE) image of dolomite (Dol) carbonatite intruded by calcite (Cal) carbonatite; **b**) BSE image of relict accessory monazite (Mnz) surrounded by a poikilitic reaction rim comprising fluorapatite (Fap), ancylite (Anc) and strontianite (Str); **c**) cathodoluminescence image of b) and surrounding area showing yellow-luminescing calcite crosscutting dolomite and blue luminescing fluorapatite rims on nonluminescent monazite. Abbreviations: Bbn, burbankite; BCal, barytocalcite; Ca, calcium; Sr, strontium.



is accessory equant monazite, which is partially replaced by secondary REE-Sr-Ba phases (Figure 4b, c).

Calcite Carbonatite

The most common variety of carbonatite found throughout the MCV is calcite carbonatite. The best-studied example is a dike composed predominantly of mosaic-textured, fineto coarse-grained calcite, minor platy phlogopite and fibrous magnesioarfvedsonite near fenitized contacts. Toward the rim, there is a decrease in Mg content in calcite grains, which indicates their crystallization from an initially Mg-rich source (Viladkar, 2000; Chakhmouradian et al., 2016). Phlogopite crystals are also zoned and characterized by Mg enrichment in their cores relative to more Ferich rims (Figure 5a). Accessory phases include monazite and burbankite, which occur as ovoid inclusions encapsulated in calcite, pyrrhotite or sphalerite (Figure 5b). Monazite and burbankite appear to have reacted with postmagmatic fluids (evidenced by the precipitation of hydrous minerals), producing reaction rims or pseudomorphs composed of ancylite, apatite, strontianite, barite and calcite (Figures 4b, c, 5c). Silicate-rich varieties of calcite carbonatite containing elevated levels of SiO₂, K₂O, FeO and MgO contents are typically weathered to an orange-brown colour owing to oxidation of ferromagnesian silicates (in particular, replacement of biotite by chamosite and, to a lesser extent, clinochlore). Throughout the MCV, silicaterich calcite carbonatite is more abundant than silicate-poor calcite carbonatite and contains brecciated fragments of the country rocks (Figure 5d).

Barytocalcite Carbonatite

A vein of cream to pink barytocalcite carbonatite with a glimmerite contact margin was observed at the intersection between jacupirangite and melasyenite units. Its association with glimmerite indicates contact-metasomatic changes similar to those reported for other carbonatites in British Columbia (Chakhmouradian et al., 2015; Rukhlov et al., 2018). Barytocalcite carbonatite is a newly recognized rock type for the IRC, making it one of just a few localities worldwide where barytocalcite has a rock-forming status (Zaitsev et al., 1998; Reguir, 2001). Whole-rock geochemical analysis of the sample collected from this carbonatite shows anomalously high concentrations of Ba, Sr and REE (~12.5, 1.9 and 1.4 wt. %, respectively). Fine- to

Figure 5. Images of calcite carbonatite assemblages, Ice River Complex, southeastern British Columbia: a) backscattered-electron (BSE) image of zoned phlogopite (PhI) in calcite (Cal) carbonatite; b) BSE image of ovoid monazite (Mnz) and titanium-rich magnetite (Ti Mag) inclusions in pyrrhotite (Po); c) BSE image of a pseudomorph after burbankite composed of ancylite (Anc), strontianite (Str), apatite (Ap), barite (Brt) and partially resorbed dolomite (Dol); d) crossed-polarized light image of comb-textured calcite carbonatite containing xenoliths of foliated McKay Group metasedimentary rock. Abbreviation: Gn, galena.







Figure 6. Images of barytocalcite carbonatite vein, Ice River Complex, southeastern British Columbia: a) cathodoluminescence image of barytocalcite (BCal) showing trace-element zoning, which is undetectable optically; note red-luminescing calcite (Cal) near the glimmerite-barytocalcite vein contact; b) corresponding plane-polarized light image of the barytocalcite vein and glimmerite composed of phlogopite (PhI) and ilmenite (IIm).

medium-grained barytocalcite makes up the bulk of this vein and is also developed interstitially within the glimmerite (Figure 6a, b). The remainder of the vein is composed of accessory strontianite and ancylite, which are found as very fine-grained overgrowths on incorporated barite and ilmenite grains. In CL images, barytocalcite luminesces dull to bright yellow; smaller patches of red-luminescing calcite are found within the glimmerite and at its contact with barytocalcite (Figure 6a).

Hydrothermal Carbonate Assemblages

Mineralogically complex hydrothermal carbonate assemblages are hosted in thin veinlets crosscutting syenites (Figure 7a) and carbonatites (Figure 7b, c). These assemblages share many similarities with the barytocalcite carbonatite, in particular, an abundance of Ba and Sr carbonate phases. The principal difference between the assemblages and the barytocalcite carbonatite is the abundance of chlorite (dominantly chamosite) in the former. The instability of biotite, paucity of hematite and abundance of chamosite, il-



Figure 7. Images of veinlet-hosted hydrothermal carbonate assemblages, Ice River Complex, southeastern British Columbia: a) backscattered-electron (BSE) image of fracture infill of barytocalcite (BCal), strontianite (Str), ancylite (Anc) and thorite (Thr) in K-feldspar (Kfs)-rich syenite with minor biotite (Bt); in syenites, smaller-scale fractures are composed of thorite 'rims' surrounding fracture edges, strontianite and ancylite 'blockages' in fluid pathways, and barytocalcite forming around the strontianite-ancylite blockages; b) BSE image of fracture infill of alstonite (Asn), ancylite, strontianite and thorite in carbonatite; in comparison with a), there is more ancylite and less thorite in this assemblage; the alstonite present in this assemblage is a polymorph of barytocalcite; the fracture infill in carbonatites is distinctly less ordered, compared to a); c) cathodoluminescence image of zoned apatite (Ap) occurring close to the hydrothermal assemblage shown in b); wavelength-dispersive X-ray fluorescence analysis of the apatite indicates light rareearth-element enrichment in blue-luminescing zones. Abbreviation: Cal, calcite.



menite and locally siderite in this hydrothermal assemblage imply acidic conditions conducive to K^+ and Fe^{2+} mobility (Mücke, 2005; Chakhmouradian et al., 2015). The fractureinfilling material is very fine grained and typically comprises alstonite, strontianite, ancylite, biotite, chamosite and ilmenite; thorite, pyrrhotite, sphalerite, galena, monazite, kukharenkoite, synchysite, euxenite and allanite are much less abundant. Minerals such as biotite, chamosite and ilmenite are particularly abundant in thicker veinlets associated with pervasive alteration and oxidation.

Fenite

Metasomatic mineral assemblages that can be confidently interpreted as fenite were observed along fractures and in xenoliths hosted by carbonatites and syenites. All known examples of fenitization in the MCV samples are sodic. These metasomatic rocks are composed predominantly of albite and aegirine (Figure 8a), with magnesioarfvedsonite (Na amphibole) observed only in the endocontact zone of calcite carbonatite. At the contact between syenite and fenitized jacupirangite, primary K-feldspar is rimmed by hyalophane and Ba-rich phlogopite (Figure 8b), indicating Ba mobility. In addition to Na- and Ba-silicate phases, fenitization produced several oxide minerals (Figure 8c). Oikocrysts of thorutite (ThTi₂O₆) are intergrown with Nbbearing anatase (tetragonal TiO₂), whereas pyrochlore (Na-Ca-Nb oxide) occurs as inclusions in Nb-bearing brookite (orthorhombic TiO₂). Titanite crystals in the precursor rock were replaced by ilmenite and calcite. Minor quantities of very fine-grained calcite, ancylite and barytocalcite also occur in the fenitized rocks.

Conclusions and Future Work

Dolomite, calcite and barytocalcite carbonatites have been identified throughout the Moose Creek valley (MCV). The most abundant carbonatite type, calcite carbonatite, is composed of calcite and minor proportions of dolomite and ferromagnesian silicate minerals. Predominantly dolomitic or barytocalcitic rocks are less common throughout the MCV; however, minor quantities of dolomite and barytocalcite are observed in calcite carbonatites. In MCV dolomite carbonatites, monazite and burbankite are primary, early-crystallizing rare-earth-element (REE)-bearing minerals. Both minerals show evidence of a reaction with postmagmatic fluids, which produced a texturally complex secondary mineral assemblage of Sr-Ba carbonates, barite and apatite. In this assemblage, the principal REE host is ancylite. Monazite is an early-crystallizing REE phase in both dolomite and calcite carbonatites, as indicated by its euhedral equant morphology, occurrence as inclusions in other minerals, and the presence of late-stage reaction assemblages developed at the expense of monazite. All examined calcite carbonatites contain ancylite and apatite; these minerals are most abundant in samples where the replace-



Figure 8. Images of fenite assemblages, Ice River Complex, southeastern British Columbia: **a)** plane-polarized light image of aegirine (Aeg)- and albite-rich fenite overprinting syenite; **b)** back-scattered-electron image of the contact between syenite (at left) and fenitized jacupirangite; note a reaction rim of hyalophane (Hya) and phlogopite (PhI) between the two rocks; **c)** BSE image of thorutite (Thrt) oikocrysts intergrown with anatase (Ant) and rimmed by brookite (Brk) containing pyrochlore (PcI) inclusions; the identity of anatase and brookite was confirmed by Raman microspectroscopy. Abbreviations: Ab, albite; Anc, ancylite; Ap, apatite; Cal, calcite; ChI, chlorite; IIm, ilmenite; Kfs, K-feldspar; Ntr, natrolite; Ttn, titanite.



ment of monazite or burbankite is observed. The nature or source of a fluid responsible for this hydrothermal overprint is at present unknown. It can be hypothesized that it was carbonatitic, given the absence of other voluminous intrusives enriched in Sr, Ba and REE in the study area. However, the relative contribution of nonmagmatic sources to this fluid remains to be determined. A study of C-O stable isotope variations in the MCV rocks is presently underway to address this question.

Silica-rich varieties of calcite carbonatite are widespread throughout the MCV; it is not clear if these rocks represent products of silica contamination or crystallized from a silica-rich melt. These rocks are pervasively altered, which resulted in biotite replacement by chlorite and deposition of diverse secondary REE-bearing minerals (monazite, kukharenkoite, synchysite, euxenite and allanite) in association with chlorite. The late-stage REE-bearing minerals are less abundant than monazite and ancylite. An investigation of compositional variability of these minerals, with emphasis on petrogenetically sensitive trace elements (Mn, REEs, Th and U) is currently underway to analyze the distribution of REE among different constituent minerals in the carbonatites and constrain the processes of their formation.

This study recognized barytocalcite carbonatite as a new rock type for the Ice River Complex (IRC). Barytocalcite is accompanied by strontianite and ancylite (as a major REE carrier), which is mineralogically akin to fracture-hosted hydrothermal assemblages observed throughout the MCV. It is therefore possible that this type of carbonatitic magmatism had a larger footprint prior to the exhumation and erosion of the IRC during the Laramide orogeny. The provenance of barytocalcite is presently being investigated using C-O stable isotope, Sr radiogenic isotope and traceelement geochemical analyses.

Acknowledgments

The authors thank Eagle Plains Resources Ltd., TerraLogic Exploration Inc., the Natural Sciences and Engineering Research Council of Canada, University of Manitoba and Geoscience BC (through the Geoscience BC Scholarship program) for their financial and logistical support. X.M. Yang (Manitoba Geological Survey) is thanked for his constructive review of this paper.

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Preliminary Insights into Carbonatite Genesis, and Future Directions Using Uranium-Lead Zircon Ages and Hafnium Isotopic Signatures of Alkaline Rocks from the Canadian Cordillera, Southeastern British Columbia (Parts of NTS 082L, M, 083D)

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Abdale, L., Nelson, J.L., Millonig, L.J. and Groat, L.A. (2025): Preliminary insights into carbonatite genesis, and future directions using uranium-lead zircon ages and hafnium isotopic signatures of alkaline rocks from the Canadian Cordillera, southeastern British Columbia (parts of NTS 082L, M, 083D); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 47–56.

Introduction

Carbonatites are uncommon igneous rocks (Bell et al., 1999) that contain over 30% primary carbonate minerals (Mitchell, 2005) and are of considerable economic interest due to their enrichment in rare-earth elements (REEs) and niobium, making carbonatites the major source of these elements in the world (Rankin, 2005). In addition to their economic importance, carbonatites are critical for understanding mantle processes and tectonic history. Their frequent occurrence in Precambrian cratonic areas (~88%) and association with extensional tectonics or continental rifting (Bell and Tilton, 2001; Veevers, 2007) make them valuable markers for reconstructing past continental configurations. Their distinctive geochemical signatures, characterized by a high hafnium content, enable radiogenic isotope studies that can provide insights into the composition and evolution of the subcontinental mantle.

Despite significant advances in understanding carbonatites, several key questions remain unresolved, particularly concerning their mantle source regions and genesis. Geochemical studies have linked carbonatite formation to the chemical heterogeneity of the mantle, with different isotopic compositions, such as depleted mid-ocean–ridge basalt (MORB) mantle (DMM), high- μ (HIMU) and enriched mantle (EM) types, associated with carbonatites and oceanic-island basalts (OIBs). Whereas OIBs provide insights into the sub-oceanic mantle, carbonatites, predomi-

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nantly found in continental settings, offer a window into the subcontinental mantle. However, there is ongoing debate regarding whether carbonatites originate from distinct mantle components or from heterogeneous lithospheric mantle sources. For example, carbonatites from the East African Rift have been attributed to mantle-plume metasomatism (Bell and Tilton, 2001), whereas alternative interpretations suggest a localized lithospheric source without plume involvement (Woolley and Bailey, 2012). The lack of consensus on these issues highlights the need for further isotopic and geochemical research to clarify the processes driving carbonatite genesis and mantle heterogeneity. The formation of carbonatites and their potential to become oregrade deposits are controlled by factors such as the source region of the parent magma, the depth and degree of melting, and the subsequent evolution of the magma during its ascent and emplacement (Simandl and Paradis, 2018).

Preliminary results and discussion aimed at addressing the gaps in understanding the processes leading to carbonatite formation are presented in this paper, focusing specifically on an area of the Canadian Cordillera in southeastern British Columbia (BC). Late Devonian-Early Mississippian alkaline magmatism in this region offers a unique opportunity to investigate mantle processes at a point of significant tectonic transition from passive to active along the western Laurentian margin. The primary objective of future research will be to investigate the timing and mantle source characteristics of carbonatite and syenite magmatism, with a focus on testing the hypothesis that this activity reflects the involvement of a metasomatized mantle source. By integrating new zircon U-Th-Pb geochronological data with Hf isotopic signatures, future research will aim to clarify the role of tectonic processes and mantle heterogeneity in

¹The lead author is a 2024 Geoscience BC Scholarship recipient.

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driving alkaline magmatism during this period of tectonic transition to a subduction zone setting. An introduction to these ongoing investigations is presented in this preliminary paper that serves as a foundation for forthcoming studies, which will contribute to a broader understanding of the role of metasomatized mantle sources in shaping carbonatite genesis.

Regional Geology

The Canadian Cordillera can be subdivided into five morphogeological belts from west to east: the Insular, Coast, Intermontane, Omineca and Foreland belts (Figure 1a; Monger et al., 1982; Monger and Price, 2002). The Omineca belt represents the westernmost section of the miogeocline of the North American craton, where allochthonous terranes are structurally juxtaposed with parautochthonous (Kootenay terrane) and autochthonous North American margin. The study area is located within the southern Omineca belt (Figure 1a, b) and comprises two tectonic terranes: the Selkirk allochthon and the Monashee Complex. The Selkirk allochthon features deformed and metamorphosed rocks from accreted terranes, metasedimentary strata and mafic sills thrust over the Monashee Complex, which includes basement rocks and a cover sequence of metasedimentary and metavolcanic units (Read and Brown, 1981).

Carbonatites and associated alkaline rocks in the Canadian Cordillera are restricted to the Omineca and Foreland belts (Figure 1a), where they intruded or erupted in contact with Neoproterozoic and early to middle Paleozoic miogeoclinal strata during periods of rifting or extensional tectonics (Okulitch, 1984; Parrish and Scammell, 1988; Pell, 1994). There have been three distinct episodes of alkaline magmatism within the Cordillera, spanning approximately 460 m.y. (Millonig et al., 2012). The youngest and most extensive episode occurred around 360–340 Ma (Figure 1b), associated with extensional tectonics related to a back-arc regime during the early stages of subduction along the western margin of Laurentia (Pell, 1994; Nelson and Colpron, 2007).

Local Geology

Blue River Area (BRA)

The Blue River area (BRA) is underlain by the Selkirk allochthon, situated structurally above the Paleoproterozoic basement represented by the Malton and Monashee complexes (Figure 1b; Digel et al., 1998). It is bounded by the North Thompson normal fault to the west and the Rocky Mountain Trench to the east (Figure 1b; Struik, 1993). The area features brittle normal faults and folds, collectively known as the Selkirk fan, characterized by a southwestward to northeastward change in vergence (Ewing, 1981; Gibson et al., 2008). The Trident Mountain syenite occurs on the eastern flank of the fan, where the deepest crustal levels are exposed (Figure 1b).

The principal lithological unit in the BRA is the Neoproterozoic Horsethief Creek Group, a clastic turbidite sequence that forms the basal succession of the upper Proterozoic and Paleozoic Cordilleran miogeocline (Figure 1b; Wheeler, 1965; Brown, 1978; Perkins, 1983). It is overlain by the upper Paleozoic-lower Cambrian Hamill Group quartzites, the lower Cambrian Badshot Formation marbles and the deep-water facies of the lower Paleozoic Lardeau Group, which includes carbonates and metavolcanics (Figure 1b; Colpron et al., 2002). Approximately 18 carbonatites and accompanying alkaline rocks in the BRA occur as sill-like bodies and lenses that intruded these units during the late Cambrian and Late Devonian-Late Mississippian (Figure 1b; Millonig et al., 2012). The Trident Mountain syenite (TMS) is found within a lower pelite unit in the eastern part of the BRA.

Frenchman Cap Dome (FCD)

The Monashee Complex, which includes the Frenchman Cap dome (FCD), represents the deepest exposed structural level within the Shuswap metamorphic core complex (Figure 1b; Monger et al., 1982; Okulitch, 1984). The FCD is bounded to the west by the Monashee décollement, a ductile reverse shear zone, and to the east by the Columbia River fault (Figure 1b). The Mount Grace carbonatites are located on an overturned limb of the Mount Grace syncline, a west-verging fold nappe. The Three Valley gap (TVG) alkaline rocks occur in the southwestern part of the FCD, within a series of tightly spaced normal faults related to the Victor Lake fault.

The Monashee core gneiss comprises Proterozoic orthogneiss and subordinate paragneiss and is unconformably overlain by the Monashee cover gneiss, which includes a Mesoproterozoic to Late Devonian metasedimentary sequence (Figure 1b; Wheeler, 1965; Reesor and Moore, 1971). Within the FCD, carbonatites and alkaline rocks ap-

Figure 1. Morphogeological belts of the Canadian Cordillera and generalized geology of the study area: a) distribution of carbonatite and alkaline complexes (black dots) in western North America (after Pell, 1994; McMillan and McLemore, 2004; Lund, 2008; Woolley and Kjarsgaard, 2008; Millonig et al., 2012) and simplified map showing the boundaries (blue dashed lines) of the five morphogeological belts (modified from Monger and Price, 2002), the pressure-temperature diagram of amphibolite-facies metamorphism (purple polygon) and the study area (red box); b) sample localities (red dots) plotted on a tectonic assemblage map of the study area, and other carbonatite and syenite localities plotted (small white dots) with igneous ages (after Parrish and Scammell, 1988; Pell, 1994; Parrish, 1995; Colpron et al., 2002; Millonig et al., 2012). Abbreviations: BCF, Birch Creek fault; Cr., creek; CRF, Columbia River fault; Gl., glacier; L., lake; LtCh, Little Chicago; MD, Monashee décollement; Mt., mount; Mtn., mountain; NEF, Northeast fault; NTF, North Thompson fault; RMT, Rocky Mountain Trench; Serp., Serpentine.



Sample location

Thrust fault

Normal fault Fault, unclassified

River, lake

0

Other carbonatites (unless denoted)

10 km

NFF

Hamill Group

. Ar

5.3 Ma

118°W

52° N



51° N



pear as large, sill-like bodies that intruded or erupted into the basal few hundred metres of the Monashee cover gneiss (Höy, 1987; Scammell and Brown, 1990). The Mount Grace carbonatites (MGCs) are located within the upper cover gneiss and the TVG rocks are located within the lower cover gneiss of the Monashee Complex.

Zircon Isotopic Sampling Program

Previous Geochronological Studies

This study focuses on the TMS, MGC and TVG carbonatite and syenite, each of which has been the subject of previous geochronological analyses. Table 1 lists the locations of samples from this study and related carbonatites and syenites along with any known igneous or metamorphic age. Millonig et al. (2012) conducted U-Pb zircon analysis on the TMS, this syenite yielding an inferred intrusion age of ca. 360 Ma, along with a metamorphic age of ca. 57 Ma, based on the observed linear discordia arrays. Millonig et al. (2012) also investigated zircon from the MGC at the Blais Creek location, concluding that the eruption age was ca. 359.3 ± 2 Ma. Parrish (1995) analyzed samples from the MGC at Blais Creek and determined that the growth of metamorphic zircon occurred at ca. 55 Ma, indicating significant post-eruption metamorphism. Uranium-lead analysis of zircon from an intrusive carbonatite gneiss near the Three Valley gap locality by Parrish (1995) suggested an intrusion age of 359 ±46 Ma for the TVG carbonatite and syenite. Notably, the TVG syenite has not been previously dated, making this study crucial for establishing its geochronological context.

Methods

Zircon grains were separated using conventional density and magnetic methods. The entire separate was annealed in a muffle furnace at 900 °C for 60 hours to repair radiation damage, enhance cathodoluminescence (CL) emission and improve laser-ablation inductively coupled plasma–mass spectrometry (LA-ICP-MS) performance. After annealing, individual grains were hand-picked, mounted, polished and imaged via CL using a scanning electron microscope (SEM), and spots for LA-ICP-MS analysis were selected.

Uranium, thorium and lead isotope analyses were performed using LA-ICP-MS at the Goethe-Universität Frankfurt, following the method presented in Gerdes and Zeh (2006, 2009), with slight modifications. A Thermo-Scientific Element II High Resolution ICP-MS was coupled with a 193 nm excimer laser system for ablation. Spot sizes ranged from 17 to 80 μ m, depending on uranium content. Data were acquired in time-resolved mode, with a 20 second background measurement followed by 21 seconds of sample ablation. Raw data were corrected offline for background signal, common lead, elemental fractionation and instrumental mass bias using an in-house spreadsheet program. The method was verified using reference zircon samples.

Hafnium isotopes were measured in zircon by monitoring masses ¹⁷²Yb, ¹⁷³Yb, ¹⁷⁵Lu and ¹⁷⁶Hf, with isobaric interference corrections applied. Bias corrections used an exponential law (based on ¹⁷⁹Hf/¹⁷⁷Hf = 0.7325) and the GJ-1 zircon served as the reference standard. The average ¹⁷⁶Hf/¹⁷⁷Hf value obtained for GJ-1 was 0.282008 ±16 (2 σ), consistent with published values.

Sample Descriptions

Blue River Area

The Trident Mountain syenite is a concordant lenticular body within the lower pelite unit of the Horsethief Creek Group rocks and outcrops of the Selkirk terrane on the slopes of Trident Mountain and adjacent ridges (Perkins, 1983; Pell, 1994). The syenite is white to gray, medium grained and moderately well foliated parallel to the margins of the intrusive body, with compositional layering (Perkins, 1983; Pell, 1994). Sample TMS (Table 1) is a nephelinesyenite gneiss composed of K-feldspar (30–40 vol. %), nepheline (34–45 vol. %), plagioclase (10 vol. %), sodalite (2–5 vol. %) and biotite (~2 vol. %), with accessory zircon and pyrochlore.

Frenchman Cap Dome

The Mount Grace carbonatites are pyroclastic deposits and occur as thin (~0.5-4 m), laterally discontinuous stratabound mappable lenses on a single stratigraphic horizon within the Monashee cover gneiss (Höy and McMillan, 1979; Höy, 1987). They can be traced and projected for at least 60 km along strike on the inverted southwestern limb of the Mount Grace syncline (Figure 1; Höy, 1987; Pell, 1994). The Mount Grace carbonatites (at Blais Creek, Mount Grace, Perry River and Bourne Glacier; Figure 1b) were extensively described in Abdale et al. (2024) and comprise a range of lithofacies, from tuffs, lapilli tuffs and tuff breccias to country-rock breccia. Sample MGC (Table 1) is a calciocarbonatite from the Mount Grace location: it is a brown- to black-weathered, massive to poorly bedded and mildly foliated clast-supported lapilli tuff that is poorly sorted and weakly normally graded. The grains appear to be ~75-85% juvenile carbonatitic in composition, with ~15-25% albite-rich xenoliths.

The Three Valley gap carbonatite and syenite occur as thin, discontinuous bedding-parallel lenses within the Monashee cover gneiss. Carbonatite lenses are generally 20– 60 cm in width and have mafic fenites 10–30 cm thick developed between them and adjacent rocks. Everywhere they were observed, the fenites are in direct contact with, and gradational to, syenites. Commonly the carbonatite occurs as lenses within the fenite. Sample TVG-C (Table 1) is

Area	Name/ rock type	Sample	Latitude (N), Iongitude (W)	U/Th-Pb age (Ma ±2ஏ)	Phase (method)	Interpretation	Source	Sample description and other information
Blue River area	Trident Mountain syenite	TMS	51°53'38.48", 118°05'43.25"					Intrusive; foliated
		TR-042-1	51°53'38.48", 118°05'43.25"	359.2 ±1.4	Zircon (weighted average)	Igneous age	Millonig et al. (2012)	
		TR-042-1		57.2 ±1.2	Zircon (lower intercept)	Metamorphic age	Millonig et al. (2012)	
		Unnamed	51°54"00", 118°09'00"	378 ±7	Zircon (upper intercept)	Igneous age	Pell (1994)	
		Unnamed		138 ±9	Zircon (lower intercept)	Metamorphic age	Pell (1994)	
		Unnamed		60	Pyrochlore (weighted average)	Metamorphic age	Pell (1994)	
Frenchman Cap dome	Mount Grace carbonatite	MGC	51°31'20.56", 118°48'40.63"					Extrusive; pyroclastic
		BL-081-3	51°35'30.45", 118°48'23.45"	359.3 ±2	Zircon (weighted average)	Igneous age	Millonig et al. (2012)	
		BL-081-3		50.9 ±0.8	Pyrochlore (weighted average)	Metamorphic age	Millonig et al. (2013)	
		RS-3	51°35'08.32", 118°48'23.18"	55 ±5	Zircon	Metamorphic age	Parrish (1995)	
		Unnamed	51°31'20.56", 118°48'40.63"	60	Pyrochlore	Metamorphic age	Pell (1994)	
	Three Valley gap carbonatite	TVG-C	50°55'37.36", 118°23'31.47"					
		PCA-303	50°55'49.20", 118°23'30.16"	359 +46/-43	Zircon (upper intercept)	Igneous age	Parrish (1995)	
				86 +12/-14	Zircon (lower intercept)	Metamorphic age	. ,	
		Unnamed	50°55'34", 118°23'29"	70–100	Zircon	Metamorphic age	Pell (1994)	
	Three Valley gap syenite	TVG-S	50°55'37.36", 118°23'31.47"					

 Table 1. Rock type, sample name, coordinates and geochronological data for the Blue River and Frenchman Cap dome areas.



a calciocarbonatite and consists of calcite (60–70 vol. %), biotite (10–20 vol. %), amphibole (hornblende, 5– 20 vol. %), apatite (5–10 vol. %) and augite (1–5 vol. %), as well as trace amounts of titanite, ilmenite, pyrochlore, pyrrhotite, monazite, epidote (allanite), kyanite, REE-carbonates and zircon. Sample TVG-S (Table 1) is a nepheline syenite and consists of K-feldspar (50–60 vol. %), plagioclase (40–50 vol. %), augite (5–15 vol. %) and titanite (1– 2 vol. %).

Preliminary Data Interpretation

Interpretation of the initial data has revealed key insights into the magmatic history of Late Devonian–Early Mississippian carbonatites and syenites in the southeastern Canadian Cordillera. Results from preliminary zircon U-Pb dating suggest that these rocks may record a significant magmatic pulse ca. 360 Ma. This aligns with a regional pulse of alkaline magmatism that affected the FCD and BRA ca. 360–340 Ma (Pell, 1994; Millonig et al., 2012; Millonig and Groat, 2013). While several carbonatite and syenite complexes in the BRA have been found to date from the Late Devonian–Early Mississippian (Figure 1b), the MGC and TVG carbonatite and syenite are the only known occurrences of this age in the Monashee Complex (Figure 1b). This period corresponds to a significant tectonic transition along the western Laurentian continental margin, which occurred ca. 390 Ma (Monger and Price, 2002). The tectonomagmatic setting is displayed schematically in Figure 2 (see Abdale et al., 2024, Figure 9). During this time, the margin shifted from an intraplate continental margin to an interplate convergent margin, followed by extensional tectonics along rejuvenated crustal faults and back-arc basin formation due to slab rollback further west during the Late Devonian to early Carboniferous (Figure 2; Roback et al., 1994; Smith et al., 1995; Colpron et al., 2007; Nelson and Colpron, 2007; Lund, 2008; Lund et al., 2010). In this tectonic setting, carbonatites likely intruded near the continental margin, where a thin continental lithosphere was present (Figure 2; Millonig et al., 2012; Abdale et al., 2024). Protoliths to the BRA carbonatite-syenite hostrocks (the Horsethief Creek Group) represent relatively deepwater, likely back-arc basin strata (Brown et al., 1986; Journeay, 1986; Scammell and Brown, 1990), whereas the FCD carbonatite-syenite hostrocks (the Monashee cover gneiss) represent shallow-water, continental-margin sediments (Figure 2; Wheeler, 1965; Reesor and Moore, 1971;



Figure 2. Schematic diagram of ancestral North American margin in Devonian–Mississippian times. Top panel shows a hypothetical depositional setting with extensional arrows and the location (small red box) of the bottom panel that shows the depositional mechanism of Mount Grace carbonatite lapilli tuff (LT1) and tuff breccia (TB1 and TB2) lithofacies as well as the calcareous mud, sandstone, evaporite and limestone in a transgressive marine sequence (modified from Abdale et al., 2024, Figure 9).





Figure 3. Hafnium evolution diagrams for carbonatites worldwide and late Archean syenitic complexes from the Canadian Shield (adapted from Rukhlov et al., 2015, Figure 5b). Depleted mid-ocean-ridge basalt (MORB) mantle (DMM), enriched mantle 1 and 2 (EM1, EM2), 'Focus zone' (FOZO) and high-²³⁸U/²⁰⁴Pb or (HIMU) mantle components are shown on the y-axis of this Hf(t) versus time (Ga) diagram; Hf(t) = ([¹⁷⁶Hf/¹⁷⁷Hf_(sample))¹⁷⁶Hf/¹⁷⁷Hf_(CHUR)]-1) 10⁴, where ¹⁷⁶Hf/¹⁷⁷Hf_(sample) is the initial ratio in the sample and ¹⁷⁶Hf/¹⁷⁷Hf_(CHUR) is the ratio in the chondritic uniform reservoir (CHUR) at that time (Rukhlov et al., 2015). Also shown are data from the oldest silicate rocks and detrital zircons from South Africa, Western Australia and West Greenland (Rukhlov et al., 2015). Error bars are 2 uncertainties that include propagated errors associated with age, measured ¹⁶⁷Lu/¹⁷⁷Hf and ¹⁷⁶Hf/¹⁷⁷Hf ratios, ¹⁷⁶Lu decay constant, and CHUR parameters. Copyright Province of British Columbia. All rights reserved. Used with permission.

Journeay, 1986; Scammell and Brown, 1990; Crowley, 1997; Höy, 2001). The abundance of Late Devonian–Early Mississippian alkaline rocks in the BRA reflects the thin lithosphere and concentration of reactivated deep-seated crustal faults within the back-arc basin, whereas further east on the continent, the thick crust and fewer faults may have limited such ultra-low viscosity intrusions (Figure 2).

Preliminary U-Pb and Hf isotopic data indicate correlations with mantle source characteristics, suggesting HIMU-EM2-like components that are consistent with metasomatized mantle sources. This is similar to the findings of Ruhklov et al. (2015), who presented Sr, Pb, Nd and Hf isotopic data for carbonatite occurrences mainly from the northern hemisphere, including a few samples from the Canadian Cordillera, along with published global data (Figure 3). Figure 3 presents carbonatite-evolution data, including HIMU (high- μ , where $\mu = {}^{238}U/{}^{204}Pb[t = 0]$); a radiogenic Pb source found in MORBs, OIBs, kimberlites and carbonatites associated with recycled oceanic and continental crust subducted into the mantle (Hofmann and White, 1982; Zindler and Hart, 1986; Hofmann, 1997; Stracke et al., 2005); and EM2, a radiogenic Hf source formed from metasomatic enrichment of ancient oceanic

lithosphere followed by long-term, deep-mantle storage (Zindler et al., 1979 and Roden et al., 1984; Workman et al., 2004). These signatures are consistent with a subduction zone, back-arc, thin-lithosphere setting. Results from this study establish the presence of a metasomatized mantle below the western continental margin in the Late Devonian– Early Mississippian (Figure 3). These initial observations will be explored further in future studies to better define the role of mantle heterogeneity in these magmatic systems.

Summary and Next Steps

Future research will focus on providing a comprehensive analysis of these findings, with implications for understanding mantle metasomatism and tectonic-magmatic interactions along continental margins. The successful application of zircon U-Pb geochronology and Hf isotopic analysis in this study demonstrates a robust methodology for understanding the magmatic history and mantle processes beneath western North America. These initial data hint at connections between regional tectonic shifts and magmatic processes; however, further studies are essential to fully elucidate how mantle dynamics and crustal structure influenced magmatism in this back-arc setting. Mov-



ing forward, additional data will be collected from approximately ten samples of carbonatites and syenites sourced from the FCD and BRA, with ages ranging from 800 to 360 Ma. Zircon Hf isotopic analysis of these samples will assist in further elucidating the Hf composition of the subcontinental mantle, thus enhancing understanding of its evolution and the tectonic context of the region. This expanded dataset will contribute to a more comprehensive interpretation of subcontinental mantle dynamics and the geochemical signatures associated with alkaline magmatism in this area. The results presented in this paper establish a preliminary framework, laying the groundwork for future detailed analyses on the interaction between mantle heterogeneity and tectonics in the Cordillera.

Acknowledgments

The authors are grateful for the support from Geoscience BC (in the form of a Geoscience BC Scholarship to the lead author) and the Natural Sciences and Engineering Research Council of Canada (in the form of a Discovery Grant to the fourth author) that helped fund part of this research. The lead author thanks E. Ye for assistance in the field. The authors also thank M. Parker for reading and reviewing various versions of this document.

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Tourmaline as a Recorder of Geochemical Evolution and an Exploration Tool in the Petalite-Subtype Prof Pegmatite, Revelstoke, Southeastern British Columbia (NTS 082M/01)

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Breasley, C.M., Groat, L.A., Martins, T. and Linnen, R.L. (2025): Tourmaline as a recorder of geochemical evolution and an exploration tool in the petalite-subtype Prof pegmatite, Revelstoke, southeastern British Columbia (NTS 082M/01); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 57–66.

Introduction

The Prof pegmatite is a petalite-subtype pegmatite located in southeastern British Columbia (BC). This pegmatite was recently discovered to contain petalite, marking it as the first known pegmatite in BC to contain significant lithium mineralization (Breasley et al., 2024). The pegmatite contains a suite of minerals indicative of a highly evolved geochemical pegmatitic melt including petalite, elbaite, lepidolite and Ta-Nb oxides. The multiple textural and geochemical varieties of tournaline within the Prof pegmatite record a complex geochemical history of crystallization and metasomatism. This paper presents the preliminary results of a study into the tournalines of the Prof pegmatite, and forms part of the lead author's Ph.D. research into lithium mineralization in petalite-subtype pegmatites across Canada.

Regional Geology

The Prof pegmatite is located on Boulder mountain², west of Revelstoke, BC. It is situated in the Monashee Complex, a division of the larger Shuswap Complex, which contains two major culminations: the Thor-Odin dome in the south and the Frenchman Cap dome in the north (Figure 1). Boulder mountain is located between these two structural culminations within the Monashee Complex cover sequence (Norlander et al., 2002). The cover sequence contains a variety of schists, quartzite and calcsilicate and quartzofeldspathic gneisses, which overlie the higher grade ortho- and

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paragneisses and migmatites of the Monashee Complex basement (Hinchey et al., 2006). These basement rocks underwent high-grade metamorphism and deformation and have experienced multiple periods of anatexis (Hinchey et al., 2006). Between 60 and 50 Ma, the Monashee Complex experienced rapid exhumation and decompression, leading to the dome expressions seen today (Spalla et al., 2011). Anatexis was a major process during this time period and is recorded in the form of migmatites, leucosomes and pegmatites (Hinchey, 2005). Pegmatites of the region have been dated, for example, the aplite-pegmatites in the Blanket Mountain and Grizzly flats³ area (25 km south of Boulder mountain) crystallized between 52.2 ± 0.5 Ma and 50.2 ±0.5 Ma (U-Pb zircon dating; Johnston et al., 2000). Additionally, the youngest S-type granite of the region, the Ladybird suite (Hinchey and Carr, 2006), has a zircon U-Pb age of 62.1 ± 0.3 Ma for the granites and as young as 55.5 ±0.3 Ma for associated pegmatites (Carr, 1992). Eoceneage normal brittle faults cut through some of the pegmatites in the region (Kruse and Williams, 2005; Hinchey et al., 2006).

Prof Pegmatite

The Prof pegmatite lies within an extensive field of pegmatites, which has not been explored in detail. The pegmatites on Boulder mountain were briefly described by Lane (2017) and the Prof pegmatite was described in detail by Breasley et al. (2024). The Prof pegmatite is a bilobate intrusion 70 m long and 5 m wide that strikes 60° and was divided into four mineralized zones by Breasley et al. (2024): 1) border zone, 2) intermediate zone including the graphic and overgrowth subzones, 3) central zone and 4) quartz zone. The border zone forms a millimetre-scale band surrounding the pegmatite and contains quartz, K-feldspar, muscovite and dravite-schorl (Mg- and Fe-bearing tourmaline) with minor magnetite and biotite. Within the interme-

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²Latitude 51.005292, longitude -118.394507

³Latitude 50.820186, longitude -118.349559

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Figure 1. a) Regional geology of the Revelstoke area (modified after Wheeler and McFeely, 1991). The red square shows the location of Boulder mountain (unofficial name) and the Prof pegmatite. b) The overview map shows the location of the Monashee Complex in relation to the morphogeological belts of British Columbia.



Preliminary Results

diate zone, the graphic subzone contains graphic intergrowths of K-feldspar and quartz and hosts multiple aplite bands, which contain quartz, perthite, albite and schorldravite. The overgrowth subzone lacks graphic intergrowths and contains more abundant tourmaline. Multiple minerals form rims around older phases in this subzone. The central zone hosts an evolved pegmatite core, which contains the main lithium mineralization of petalite, elbaite (Li-bearing tourmaline) and lepidolite. A subzone found at the contact between the intermediate zone and the central zone is termed the distal tourmaline subzone, due to the abundance of different coloured tourmalines. The quartz zone is 1-3 m thick and is almost monomineralic with minor beryl, schorl and mica. Three phases of metasomatism altered the Prof pegmatite including an albitization event, a later more Na-Li-F-enriched metasomatic event and a sericitization event (Breasley et al., 2024).

Methods

During fieldwork in 2021, 2022 and 2023, a representative sample set of tourmalines was collected from the Prof pegmatite. Samples of tourmalines with variable colours and textures and from different zone locations form the basis of this study. Mineral compositions were determined using a JEOL Ltd. JXA-iHP200F field emission electron probe microanalyzer in the Department of Earth, Ocean and Atmospheric Sciences at The University of British Columbia (Vancouver, BC). Compositional data were acquired for tourmalines and X-ray intensities were processed together within the Probe for EPMA software (Probe Software, Inc.). The instrument was operated at an accelerating voltage of 15 kilovolts (kV) with a beam current of 15 nanoamperes (nA) and a beam diameter of 5 µm. Tourmaline compositions were quantified and corrected using the following standards: albite (CMTaylor) for Si Ka, Na Ka; corundum (CMTaylor) for Al $K\alpha$; spessartine (CMTaylor) for Mn $K\alpha$; ScPO₄ (National Museum of Natural History [NMNH] 168495) for Sc Ka; RbTiPO₅ (Astimex Standards Ltd.) for Rb $L\alpha$; pollucite (SPI Supplies [SPI]) for Cs $L\alpha$; chromium oxide (SPI) for Cr Ka; diopside (SPI) for Ca Ka, Mg $K\alpha$; fluorite (SPI) for F $K\alpha$; hematite (SPI) for Fe $K\alpha$; rutile (SPI) for Ti $K\alpha$; orthoclase (SPI) for K $K\alpha$; tugtupite (SPI) for Cl Ka; and Cu (JEOL (Germany) GMBH) for Cu $K\alpha$. Structural formulae were calculated on the basis of 31 anions, assuming stoichiometric amounts of H₂O as (OH), that is, OH + F = 4 atoms per formula unit (apfu), B_2O_3 (B = 3 apfu) and Li₂O (as Li⁺; MacDonald et al., 1993; Burns et al., 1994). The amount of Li assigned to the Y mineralogical site corresponds to the ideal sum of the cations occupying the T + Z + Y mineralogical sites (15 apfu) minus the sum of the cations actually occupying these sites (Li = 15 -[T+Z+Y] or Li = 15 - [Si+A]+Mg+Fe+Mn+Zn+Ti+Sc + Cr]); the calculation was iterated to self-consistency.

Border Zone

Tourmalines of the border zone typically contain concentric zoning and occur throughout the border zone as individual black crystals, either disseminated or in a comb structure oriented perpendicular to the contact. Detailed compositions of these tourmalines are yet to be analyzed, but they were shown to be dravite-schorls from Raman spectroscopy.

Intermediate Zone

Tourmalines are highly variable in the intermediate zone and occur in comb structures (Figure 2a), in aplite bands (Figure 2b) and as individual millimetre-scale crystals disseminated throughout the zone. The tourmalines typically have a strong concentric zonation but locally contain no zonation (Figure 3a, b). In the aplitic portions of the intermediate zone, the tourmaline cores contain high contents of Mg and Fe and are dravites. These tournalines are typically black in representative samples and appear brown to deep blue in thin section. These dravites are often surrounded by a dull brown-green rim in representative samples. This corresponds to schorl compositions with Fe > Mg or Li+Al at the Y mineralogical site. Locally, the dravites and schorls are coated by a final F- and Li-rich tourmaline rim with a fluor-elbaite composition (Figure 3a), which is often colourless in thin section and cannot be distinguished by eye in representative samples.

Central Zone

Multiple colours of tourmaline exist in the central zone of the Prof pegmatite including green tourmalines in the more distal regions of the zone, pale pink tourmalines and rare blue tourmalines. These tourmalines are all elbaite to fluorelbaite in composition. A variety of these multicoloured tourmalines are located in the distal tourmaline subzone, which is close to the contact with the intermediate zone. Tourmalines of the core of the central zone form elongate radial sprays or distinct isolated crystals (Figure 2c). These pink tourmalines can either show oscillatory zonation or can contain abundant quartz (Figure 3c) and, rarely, both of these textures can be seen within individual crystals (Figure 3d). The fluor-elbaite commonly host elevated Mn concentrations (up to 0.32 apfu).



Table 1. Selected compositions of tourmalines from the Prof pegmatite, southeastern British Columbia. Oxides are in wt. %, ions are in atoms per formula unit (apfu), B = 3 apfu and Li = 15 – (T+Z+Y). Abbreviations: n.d., no data; r, vacancy; T, Z and Y, mineralogical sites.

Thin section ID	PFS1A	PES1A	PFI 2A	P34A
Zone	Intermediate	Intermediate	Central, core	Quartz
(subzone)	(graphic)	(graphic)		
Position within mineral	Middle	Middle	Core	Middle
Point	49	50	51	78
Mineral	Dravite	Schorl	Fluor-elbaite	Fluor-schorl
Oxide/ion				
SiO ₂	34.42	34.36	37.43	34.08
TiO ₂	0.46	0.71	0.02	0.01
Al ₂ O ₃	32.73	33.26	41.23	33.77
Sc ₂ O ₃	0.01	0.03	0.01	n.d.
FeO	7.87	9.8	0.02	13.14
MgO	6.19	4.38	n.d.	0.2
CaO	0.85	0.5	0.36	0.08
MnO	0.05	0.1	1.44	1.13
Na ₂ O	2.09	2.04	2.05	2.18
K ₂ O	0.08	0.07	0.02	0.04
Rb ₂ O	0.07	0.05	0.02	0.06
F	0.54	0.41	1.32	1.01
H ₂ O*	3.36	3.41	3.25	3.08
$B_2O_3^*$	10.48	10.45	11.23	10.32
Li ₂ O*	0.31	0.31	2.4	0.72
Total	99.52	99.91	100.82	99.82
T: Si	5.7	5.72	5.79	5.74
AI	0.29	0.29	0.21	0.26
В	3	3	3	3
Z: AI	6	6	6	6
Mg	n.d.	n.d.	n.d.	n.d.
Y: Al	0.1	0.24	1.31	0.45
Ti	0.06	0.09	n.d.	n.d.
Mg	1.53	1.09	n.d.	0.05
Mn	0.01	0.01	0.19	0.16
Fe ²⁺	1.09	1.36	n.d.	1.85
Li*	0.3	0.21	1.5	0.49
Total Y	3	3	3	3
X: Ca	0.15	0.09	0.06	0.02
Na	0.67	0.66	0.62	0.71
к	0.02	0.02	n.d.	0.01
Rb	0.01	n.d.	n.d.	n.d.
r	0.16	0.23	0.32	0.26
ОН	3.71	3.78	3.35	3.46
F	0.28	0.22	0.65	0.54

*calculated value

Quartz Zone

Large blue-black tourmalines are present at the edges of the quartz zone and have a fluor-schorl composition.

Discussion

The Prof pegmatite contains multiple minerals that record the fractionation and geochemical evolution of melt, as evidenced by their incompatible element contents. The geochemical evolution of the micas and Nb-Ta oxides of the pegmatite are described in detail in Breasley et al. (2024). Tourmalines are a common mineral that can record geochemical evolution trends in pegmatites. They are a useful mineral to analyze in pegmatites as they have the potential to form throughout all stages of crystallization, if there is sufficient B in the system, capturing the geochemical environment throughout the paragenesis of the pegmatite, from emplacement to the final stages of metasomatism. In general, tourmalines show a trend of Fe- and Mg-bearing schorl to dravite to Li-bearing elbaites with melt evolution. This has been noted by multiple authors in the literature (Jolliff et al., 1986; Selway et al., 1999; Roda-Robles et al.,





Figure 2. Outcrop and representative samples from the Prof pegmatite, southeastern British Columbia: **a)** tourmalines in a comb structure mark the contact zone between the subzones of the intermediate zone; graphic subzone to the left and overgrowth subzone to right; pencil is 14 cm long; **b)** aplite band from the graphic subzone, composed of tourmaline, feldspars and quartz; **c)** evolved core of the central zone containing radial pink elbaites with purple lepidolite roses and white feldspar; notebook is 12 cm high.

2015) and common elements that reflect the geochemical change of pegmatitic systems include Fe, Mn, Li, Al and F. Due to the abundance, textural varieties and geochemical variation of tourmalines at the Prof pegmatite, tourmaline is a highly valuable mineral to study to reveal insights into the geochemical change throughout the paragenetic crystallization sequence of the pegmatite.

The dravite and schorl within the Prof pegmatite are interpreted to be primary magmatic in nature due to their association with blocky minerals and the pristine nature of the crystals. The elbaites show complex textural relationships and exhibit oscillatory zonation and can be found with abundant quartz inclusions. The elbaites show evidence of primary magmatic origins due to their oscillatory zoning, which is not usually found in replacement minerals (corresponding to the blue/green crystals). In addition to this, the elbaites, which contain abundant quartz inclusions and appear more internally mottled (pink in colour), are found associated with crosscutting veins containing albite and lepidolite. This could indicate that these elbaites are secondary metasomatic in origin. Both of these textures can be observed within a single crystal (Figure 3d), with a relic crystal partially altered to tourmaline and quartz.

Three geochemical plots in Figure 4 show the tourmaline composition from different zones within the Prof pegmatite. The tourmalines in general show an increase in F, Mn and Na and a decrease in Fe as the pegmatite crystallized. These trends indicate that the process of incompatible element accumulation in the residual melt was significant (Selway et al., 1999). Locally, tourmalines within the intermediate zone, which have been influenced by the previously mentioned late-stage Li- and F-rich metasomatism, contain an entire progression from more primitive Fe- and Mg-bearing dravites and schorls to more evolved elbaite rims. Figure 4a, b and c all show that the intermediate zone crystallized first, recorded as comparatively low values of F, Mn, Na and elevated values of Fe in tourmalines. The initial low values and vertical spread in Mn and F in the intermediate zone can potentially be explained by garnets preferentially incorporating Mn during contemporaneous crystallization (Tindle and Breaks, 2000). A similar trend is seen in Na values, which can be attributed to primary albite crystallization in the intermediate zone. The elevated Fe values in the intermediate zone are unsurprising, as this is a more compatible element and is preferentially taken into the tourmaline crystal lattice before more incompatible elements such as Mn. At the conclusion of the primary crystal-





Figure 3. Backscattered-electron images of textural varieties of tourmaline from the Prof pegmatite, southeastern British Columbia: **a**) concentric-zoned dravite-schorl core with fluor-elbaite rim from the intermediate zone (thin section PFS1A); **b**) unzoned schorl from the intermediate zone (thin section PFS1C); **c**) mottled fluor-elbaite with abundant quartz inclusions from the central zone (thin section PFL1A); **d**) two generations of fluor-elbaite with preserved primary oscillatory zonation (green dashed line) and secondary included recrystallization (blue dashed line; thin section P28B), from the central zone.

lization of the intermediate zone, garnet and primary albites stopped crystallizing, allowing Mn and Na values to increase in the residual melt. As the central zone crystallized, the F, Mn and Na content of tourmalines increased, and this can also be correlated with later stage metasomatism of the pegmatite recorded in fluor-elbaite rims. The quartz zone contains fluor-schorls, which are thought to be a product of country rock contamination of the pegmatite coupled with later stage metasomatic recrystallization as observed in other pegmatites (Selway et al., 2000a). The most evolved tourmalines in the Prof pegmatite are fluor-elbaites, a trend which is commonly seen in evolved pegmatites (Selway et al., 2000a; Henry and Dutrow, 2011).

The texture of the comb-structured tourmalines in the Prof pegmatite is commonly seen in pegmatites such as the Tanco pegmatite, Manitoba (Selway et al., 2000b) and in Minas Gerais, Brazil (Webber and Simmons, 2007). This texture is also associated with rapid crystallization of an undercooled melt (Baker and Freda, 1999).

Conclusions

This ongoing study highlights the importance of understanding the mineralogy and paragenesis of pegmatites when exploring pegmatite fields in southeastern British Columbia. Preliminary conclusions from this study are summarized below:

The Prof pegmatite hosts a textural and geochemical suite of tourmalines that recorded magmatic and metasomatic crystallization processes throughout the entire paragenetic sequence of the deposit.

Fluorine in tourmalines is a good elemental tracer of geochemical evolution in the pegmatite as concentrations increase throughout the crystallization sequence.





Figure 4. Plots of tourmaline geochemistry from the Prof pegmatite with colour indicating the pegmatite zone/subzone and the shape reflecting the electron probe microanalysis point position within the mineral: **a**) F versus MnO, **b**) F versus Na₂O, **c**) FeO versus MnO. Abbreviations: distal tur, distal tourmaline subzone; graphic, graphic subzone.



The Prof pegmatite lies within an extensive unmapped field of pegmatites on Boulder mountain. Geochemical analysis of tourmalines within other pegmatites of the field can be used as an exploration tool via plotting data against established evolution trends from the Prof pegmatite. The presence of F- and Li-enriched rims on tourmalines in seemingly barren pegmatite surficial expressions could reflect Li mineralization at depth. This Li enrichment has the potential to be expressed as spodumene, the higher pressure lithium aluminosilicate of petalite.

Future Work

Future work aims to further investigate the tourmaline geochemistry of the Prof pegmatite in depth and provide an overview of the tourmaline textures and geochemical evolution throughout crystallization. Future analyses will include using laser-ablation, inductively coupled plasmamass spectrometry (LA-ICP-MS) to determine accurate lithium concentrations within the tourmalines. Additional data will be collected from the tourmalines of the border zone to make the dataset more complete. Once all geochemical data has been collected, links between the Prof pegmatite and the neighbouring pegmatites on Mount Begbie, described and analyzed by Dixon et al. (2014), will be drawn. These pegmatite's field, and comparing the tourmalines will provide geochemical evidence for a link between them.

This work will provide vital geochemical information, which could be used as an exploration tool for the region. Tourmalines from multiple pegmatites within the Prof pegmatite's field could be analyzed and plotted against the known tourmalines from the Prof pegmatite to show how comparatively geochemically evolved the system is. The more geochemically evolved a pegmatitic is, the more likely it is to host increased levels of incompatible elements such as Li, Cs or Ta. Therefore, understanding the tourmaline geochemistry can help vector for more economic mineralization in the region.

Acknowledgments

The authors are grateful for support from Geoscience BC (in the form of a Geoscience BC Scholarship to C. Breasley) and the Natural Sciences and Engineering Research Council of Canada (in the form of a Discovery Grant to L. Groat). This research was supported by a Geological Society of America Foundation Lipman Research Award and a SEG Canada Foundation Student Research Grant to C. Breasley. The authors thank T. Naber and T. Muilu for invaluable field assistance, and E. Johnstone of Glacier Helicopters Ltd. for helicopter support. The authors also thank A. von der Handt for electron probe microanalysis assistance at The University of British Columbia. Thanks are also ex-

tended to T. Cawood from the Geological Survey of Canada for peer reviewing this document and C. Pellet and B. Clift from Geoscience BC for their help in improving this document.

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Experimental Research on Mineralogical Characteristics and Mechanical Properties of Montney Siltstone from Northeastern British Columbia (Part of NTS 092A) Using Indentation Tests and Scanning Electron Microscope–Energy Dispersive Spectrometry Analysis

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Yu, H. and Zheng, W. (2025): Experimental research on mineralogical characteristics and mechanical properties of Montney siltstone from northeastern British Columbia (part of NTS 092A) using indentation tests and scanning electron microscope–energy dispersive spectrometry analysis; *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 67–74.

Introduction

Natural gas production from unconventional formations has surged over the past two decades, driven by advancements in horizontal drilling and multistage hydraulic fracturing (Sharma et al., 2019; Cheng et al., 2022). In northeastern British Columbia (BC), the Montney gas play has become the largest gas producer in the province, with twothirds of drilled wells targeting the siltstone-rich Montney Formation (Rivard et al., 2014; Vishkai et al., 2017). The effectiveness of hydraulic fracturing in creating fracture networks is closely linked to the mechanical properties of the target formations, which are significantly influenced by the mineralogical composition of the rock (Rivard et al., 2014; Cheng et al., 2022). Additionally, localized mechanical properties of the rock affect proppant embedment within fractures, thereby impacting fracture conductivity (Zheng et al., 2019). Thus, understanding the relationship between the mechanical parameters of the rock and its mineralogy at the microscale is critical.

The micro-indentation test was initially developed to determine the hardness and elastic modulus of thin films and small solid samples by measuring indentation depth as a function of increasing force applied by an indenter (Poon et al., 2008). Compared to traditional compression tests, the indenter applies force to a small area, creating high, localized stress that enables the measurement of mechanical properties at the microscale (Kasyap et al., 2021; Song et al., 2022). Meanwhile, scanning electron microscopy (SEM) provides high-resolution micromorphologies of sample surfaces by using a focused high-energy electron beam to excite various physical signals through its interaction with the sample (Fandrich et al., 2007). Additionally, the energy dispersive spectrometry (EDS) system, attached to the SEM, allows for chemical analysis of the scanned area (Meyer et al., 2013;

Five disk-shaped rock samples obtained from cores drilled at depths ranging from 2068.29 to 2337.39 m within the Lower Triassic Montney play in northeastern BC were used in this study. Nine instrumented indentation tests were performed on each sample to generate force-displacement curves, from which the local Young's modulus and hardness at each indentation site were determined. Following the indentation tests, SEM and EDS analyses were conducted to create digital mineral maps, enabling identification of the mineral composition and localized microstructure at the indentation sites. Finally, the influence of the mineralogical composition of the rock on its mechanical properties is discussed.

Materials and Methods

Materials

The five samples were prepared from the drillcore from well TOURMALINE HZ TOWN C-031-H/094-B-09 (well authorization 28232, unique well identifier 200C031H094B0900; BC Energy Regulator, 2024a), located in the Lower Triassic Montney play in northeastern BC, as shown in Figure 1. The measured depths of the cores ranged from 2068.29 to 2337.39 m. For comparison purposes, the cores were selected from two formations: samples S1 (2068.29 m) and S2 (2068.55 m) were taken from the Doig Formation, while samples S3 (2265.85 m), S4 (2336.04 m) and S5 (2336.37 m) were extracted from the Montney Formation.

The samples were obtained by drilling horizontally into cores (Figure 2a) and then trimmed into disk shapes with a nominal diameter of 25.4 mm. The lateral surfaces of the samples were coated with an epoxy resin (Figure 2b). To minimize the effects of surface roughness on the indentation tests, the flat surfaces were polished using a Buehler

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Figure 1. Location of the Montney unconventional gas play (modified from BC Energy Regulator, 2024b). The Montney play area and other oil and gas resource basins, including the Liard and Horn River basins, and Cordova Embayment are shown; the yellow star indicates the location of the sampled well.



Figure 2. Instrumented indentation test performed on samples from the Montney play: a) drillcore from well sampled at approximately 2068 m (red arrow indicates the location and orientation of the sample plug extracted); b) sample plug S1 (diameter of 25.4 mm); c) nine test sites, labelled L1–L9, for the Brinell hardness testing indenter arranged in a three-by-three grid (red polygon in panel b shows test site location); d) Nanovea-M1[™] instrumented indentation machine.



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Figure 3. Typical force-displacement curve for instrumented indentation test results from test site L5 of drilled sample S1 from the Montney play. The force-displacement curve is divided into two segments: the loading and unloading stages. The range of 90 to 50% of the maximum force (P_{max}) for the unloading-stage curve is selected and fitted versus indentation depth (*h*) using a power-law function. The initial slope of this fitted curve represents the contact stiffness, as illustrated by the triangle in the figure. Abbreviations: f, full; max, maximum; N, newton; m, micrometre.

AutoMet[®] 250 Pro grinder-polisher; surface roughness was also scanned using a profiler, yielding values less than $0.4 \mu m$.

Experimental Method

Instrumented Indentation Test

The Nanovea-M1[™] instrumented indentation machine (Figure 2d) was used to perform the instrumented indentation test, employing a Brinell hardness testing indenter—a spherical tungsten carbide ball with a 1 mm diameter. Each sample had nine test locations, labelled L1 to L9, arranged in a three-by-three grid, as shown in Figure 2c. During the test, the indentation force increased at the loading stage to a maximum of 35 newtons (N), then decreased to 0 N at the unloading stage. A constant rate of 17.5 N/min was applied for both the loading and unloading stages. The typical force-displacement curve is presented in Figure 3.

Notably, the final indentation depth did not fully recover to 0 after unloading. This remaining depth is known as the plastic depth, whereas the difference between the maximum and final depths is referred to as the elastic depth. The unloading curve can thus be used to calculate local mechanical properties, such as Young's modulus and hardness. This method assumes that the recoverable depth represents only elastic deformation during unloading (Oliver and Pharr, 2004) and the unloading curve of indentation force (P) versus indentation depth (h) is used for this calculation.

The *P*-*h* curve is a fitting power-law equation:

$$P \qquad (h \quad h_f)^m$$

where α , *m* and h_f are fitting constants.

S is the contact stiffness, which is determined by the initial slope (90-50%) of the *P*-*h* curve:

$$S m (h h_f)^{m-1}$$

The contacted depth (h_c) between the indenter and sample can be calculated as:

$$h_c \quad h_{\max} \quad P_{\max} / S$$

where h_{max} is maximum depth, P_{max} is maximum force, and

is the constant depending on the geometry of the indenter; for the Brinell hardness testing indenter, it is 0.75. Based on the h_c , the contact projected area (A_c) between the Brinell indenter and sample can be calculated as:

$$A_c \qquad (r^2 \quad [r^2 \quad h_c]^2)$$

where *r* is the radius of the spherical indenter (0.5 mm). When A_c is determined, the indentation hardness (*H*) can be obtained as:

$$H P_{\text{max}} / A_c$$

The efficient Young's modulus (E_{eff}) presents the apparent Young's modulus between sample and indenter, which is calculated as:

$$E_{eff} = \sqrt{s}/2 \sqrt{A_c}$$

where β is the effective coefficient of all physical processes, which is 1.05 in this study.

Moreover, Young's modulus (E) of the sample is calculated as:

$$E (1 v^2) / (\frac{1}{E_{eff}} \frac{1 v_i^2}{E_i})$$

where v and E are the Poisson's ratio and Young's modulus of the sample, respectively, and v_i and E_i are the Poisson's ratio and Young's modulus of the indenter. In this study, v_i equals 0.31 and E_i equals 600 GPa are used for the tungsten carbide Brinell indenter, and v equals 0.30 is used for the rock sample.

SEM and EDS Analysis

Scanning electron microscopy (SEM) and energy dispersive spectrometry (EDS) analyses were performed on indents from test sites L1 to L9 following the completion of the indentation tests. The electron images of SEM and elemental distributions from EDS were obtained using a Tescan Mira3 XMU scanning electron microscope equip-



ped with a field emission gun and a X-Max Energy Dispersive Spectrometer detector made by Oxford Instruments. Typical SEM and EDS images are presented in Figure 4. The high-resolution SEM images revealed anisotropic micromorphological features of the localized rock (Figure 4a), whereas the EDS images, shown in different colours, displayed the distributions of chemical elements corresponding to various mineral types (Figure 4b–i).

The online library of the BC Energy Regulator (previously BC Oil and Gas Commission) houses a digital archive of well data and reports available to the public. The X-ray diffraction (XRD) data for rock samples from the same study well and at similar depths were retrieved from this archive to inform this research (Mineralogy, Inc., 2014). Consequently, the mineral composition of the samples at each depth was determined. Using the chemical formulas of known minerals, the chemical elements identified in the layered EDS images were used to determine local mineral types at the same locations. For example, quartz (SiO_2) contains only Si and O; pyrite (FeS_2) contains only Fe and S; muscovite $\{KAl_2(Si_3AlO_{10})(OH)_2\}$ was determined by the presence of Si, Al and K; and albite $(NaAlSi_3O_8)$ was determined by the presence of Si, Al and Na.

Results

Young's Modulus and Hardness

Based on the results of the unloading P-h curves from the instrumented indentation tests, Young's modulus and hardness were determined. To ensure precision, the results from some test sites were discarded due to fluctuations in the instrumented indentation tests and the failure of EDS analysis to detect certain elements. Table 1 presents the average, standard deviation, and coefficient of variation for Young's modulus and hardness for each sample. The results show



Figure 4. Typical scanning electron microscopy (SEM) image of a) test site L1 of sample S1 from the Montney play and energy dispersive spectrometry (EDS) elemental maps of b) silicon (Si); c) calcium (Ca); d) magnesium (Mg); e) sulphur (S); f) iron (Fe); g) sodium (Na); h) aluminium (AI); i) potassium (K).



that samples from similar depths exhibit comparable values for Young's modulus and hardness. Additionally, mechanical properties vary between samples from different depths within the Montney Formation. Test locations at the shallower depth (S3 at approximately 2265 m) show a lower average Young's modulus compared to those at greater depths (S4 and S5 at approximately 2336 m). The average values of Young's modulus of the Doig Formation samples (S1 and S2) are higher than that of the Montney Formation at around 2265 m (S3), but lower than those at around 2336 m (S4 and S5). For hardness, S2 from the Doig Formation exhibits the highest average value (0.94 GPa), whereas S5 from the Montney Formation has the lowest (0.69 GPa).

Figure 5 presents the scatter plot of Young's modulus and hardness for all samples at various depths. Due to the similarity in mechanical properties among samples from similar depths, the scatter plot illustrates three distinct groups corresponding to the three depth levels. The red dots represent samples from the Doig Formation at approximately 2068 m (S1 and S2), whereas the blue triangles at 2265 m (S3) and the black squares at 2336 m (S4 and S5) correspond to sam-

ples from the Montney Formation. The dashed lines in Figure 5 represent the linear fit between Young's modulus and hardness for the three groups. Results indicate significant correlation between Young's modulus and hardness for samples from similar depths within the same formation, with R² values of 0.98, 0.97 and 0.98. Additionally, the coefficient (y) of the fitting equation reflects the ratio between Young's modulus and hardness. For the Montney Formation, the sample from 2265 m (S3) exhibits a lower linear coefficient (23.14) compared to samples from 2336 m (S4 and S5) with a higher linear coefficient (71.07). In contrast, the linear coefficient (40.61) of Doig Formation samples (S1 and S2) is larger than that of S3 at 2265 m but smaller than that of S4 and S5 at 2336 m.

Mineral Composition and Microstructure

Digital mineral maps were generated from EDS images of chemical elements using in-house MATLAB[®] code. A typical mineral map for samples from different depths is shown in Figure 6a. According to the map, samples from various depths exhibit differences in mineral grain sizes and com-

Table 1. Average (Avg), standard deviation (std) and coefficient of variation (Cv) of Young's modulus and hardness for the five samples (S1–S5) from the Montney play.

Specimen	Dataset	Depth	Young's modul	us (GPa)	Hardness (GPa)		
number	count	(m)	Avg ±std	Cv	Avg ±std	Cv	
S1	5	2068.29	32.62 ±15.35	0.44	0.81 ±0.14	0.17	
S2	9	2068.55	36.80 ±2.23	0.06	0.94 ±0.06	0.07	
S3	8	2265.85	20.14 ±3.38	0.17	0.87 ±0.10	0.12	
S4	7	2336.04	63.34 ±3.66	0.06	0.83 ±0.03	0.04	
S5	6	2336.37	42.75 ±4.41	0.1	0.69 ±0.13	0.19	



Figure 5. Young's modulus versus hardness for samples from different depths in the Montney play: circles, dashed line and values in red represent indentation results and fitting result of samples S1 and S2 from the Doig Formation at approximately 2068 m; triangles, dashed line and values in blue represent indentation results and fitting result of sample S3 from the Montney Formation at approximately 2265 m; squares, dashed line and values in black represent indentation results and fitting result of samples S4 and S5 from the Montney Formation at approximately 2336 m. Abbreviation: GPa, gigapascal.



positions. The mineral grain sizes of samples from approximately 2068 m in the Doig Formation range from 20 to 50 μ m, with the primary load-bearing framework composed of coarse minerals, including calcite, dolomite, albite and quartz. Samples from around 2265 m in the Montney Formation have grain sizes between 10 and 50 μ m. The carbonate mineral (calcite and dolomite) content in samples from 2265 m is of smaller grain size than that of samples from 2068 m. Samples from 2336 m in the Montney Formation have grain sizes of less than 20 μ m where clay minerals form the rock matrix, with harder minerals embedded within the clay.

The average area percentages (A%) of pores and minerals for each sample, including all indentation locations, were measured from the digital mineral maps and the results are presented in Figure 6b. Quartz had the highest average area percentage across all samples, whereas albite, pyrite and chlorite had the lowest values. Small amounts of titanium dioxide (TiO₂) were also identified in samples S3, S4 and S5. The average area percentage of carbonate minerals was relatively high in samples S1 and S2, whereas clay minerals (muscovite and chlorite) had a higher average area percentage in samples S4 and S5 than carbonate minerals. Sample S3 exhibited a relatively balanced distribution of carbonate and clay minerals. Table 2 presents the average, standard deviation and coefficient of variation for the area percentage of pores and each mineral across all test sites. The results show a very low coefficient of variation of only 0.25 for quartz, whereas it is very large (1.2) for titanium dioxide; however, the average area percentage is very small and titanium dioxide is found only in samples S3, S4 and S5.

Discussion

Impacts of Mineral Types on Mechanical Properties

The Pearson correlation coefficient was used to assess the impact of mineral content on Young's modulus and hardness. The p-values for the Pearson correlation range from 0 to 1, with a p-value less than 0.05 indicating a statistically significant correlation between the two variables. The correlation coefficient (R) ranges from -1 to 1, where a negative R indicates a negative correlation and a positive R indicates a positive correlation. Note that results from test sites due to fluctuations in the instrumented indentation test were discarded, as were those from EDS analyses in which elements had been missed. Figure 7 presents scatter plots of mineral average area percentage (A%) versus hardness and Young's modulus, along with the corresponding correlation coefficients. In these plots, the horizontal axis represents the average area percentage of each mineral, whereas the vertical axis indicates hardness and Young's modulus.

Based on the p-values and R coefficient, calcite (P = 0.048, R = 0.337) shows a positive correlation with hardness. Meanwhile, a positive relationship is observed between dolomite and hardness (P = 0.057, R = 0.325), although its p-value is slightly above 0.05. This finding aligns with the conclusions presented earlier in this paper, where S2 exhib-

Table 2. Average (Avg), standard deviation (Std) and coefficient of variation (Cv) for the area percentage of pores and each mineral across all test sites in the Montney play.

	Area percentage (%)									
	Porosity	Albite	Calcite	Dolomite	Muscovite	Orthoclase	Pyrite	Chlorite	Titanium dioxide	Quartz
Avg	2.75	6.91	13.75	15.39	14.25	8.67	1.68	2.47	0.12	34.01
Std	2.43	3.98	6.71	11.47	7.64	4.02	0.87	2.39	0.14	8.47
Cv	0.89	0.58	0.49	0.75	0.54	0.46	0.52	0.97	1.20	0.25



Figure 6. Mineral map and average area percentage of minerals for samples from the Montney play: **a)** typical digital mineral map for test site L1 of sample S1, with colours representing different minerals and the scale bar indicating mineral types; **b)** average area percentage (A%) of pores and minerals for each sample, with close-up (red box) showing the locations of small amounts of titanium dioxide.



	Porosity	Albite	Calcite	Dolomite	Muscovite	Orthoclase	Pyrite	Chlorite	TiO ₂	Quartz
Н	i Maria Sanara	37.1		· 帮 按		14.15 M				
Ρ	0.127	0.607	0.048	0.057	0.002	0.025	0.010	0.029	0.085	0.849
R	0.263	0.090	0.336	0.325	-0.508	-0.379	-0.428	-0.369	-0.295	-0.033
Ε	ka sana Alabaran Alabar	2000 - 100 100 100		14 - 14 3 - 454 4 - 1		ing a syl of again the		144 - 18 14 - 1917 - 1		1.481.4 2.982.1 2.10
Ρ	0.878	0.032	0.087	0.093	0.130	0.846	0.000	0.007	0.007	0.038
R	-0.027	-0.364	-0.293	-0.289	0.261	0.034	0.628	0.447	0.450	0.351

Figure 7. Scatter plots of indentation depth versus mineral area percentage (A%) and porosity for samples from the Montney play, with Pearson correlation coefficient (R) and p-value (P): red scatter plots show the relationships of hardness (H) and Young's modulus (E) with A%. For plots in the same row, the vertical axis is consistently scaled, whereas the horizontal axis varies according to the A% of the minerals. These plots visually demonstrate the relationships between different minerals and Young's modulus and hardness. The p-value indicates whether a significant linear relationship exists, whereas the R coefficient reflects the strength of this relationship. Abbreviation: TiO₂, titanium dioxide.

its the greatest hardness with the higher average area percentage of calcite and dolomite (Figure 6b), whereas orthoclase (P = 0.025, R = -0.379), pyrite (P = 0.010, R = -0.428), muscovite (P = 0.002, R = -0.508) and chlorite (P = 0.029, R = -0.369) exhibit negative correlations with hardness. Similarly, S5 has the highest average area percentage of muscovite and orthoclase (Figure 6b) and exhibits the lowest hardness. In addition, pyrite (P = 0.000, R = 0.628), chlorite (P = 0.007, R = 0.447) and titanium dioxide (TiO₂; P = 0.007, R = 0.450) exhibit positive correlations with Young's modulus, whereas albite (P = 0.032, R = -0.364) is negatively correlated with Young's modulus. This further corroborates the earlier result shown in Figure 5, where sample S3, with the highest albite content, exhibits the lowest Young's modulus.

These findings indicate a correlation between the mechanical properties of the rock and its mineral composition. An increase in carbonate minerals (calcite and dolomite) enhances rock hardness. Conversely, an increase in clay mineral content (muscovite and chlorite) decreases hardness but increases Young's modulus, whereas higher albite content reduces Young's modulus. However, in this study, the lack of observed correlations between porosity and other minerals, such as quartz, may be due to the limited variation in quartz content, as evidenced by a coefficient of variation of only 0.25 for quartz (Table 2). Therefore, it cannot be concluded that these factors have no impact on the mechanical properties of the rock.

Conclusion

Instrumented indentation tests and scanning electron microscope-energy dispersive spectrometry analyses were conducted on five samples from the Montney unconventional gas play in British Columbia. The indentation tests measured local Young's modulus and hardness, whereas mineral compositions were identified through scanning electron microscope–energy dispersive spectrometry analysis. The key findings of the study are summarized as follows:

- Images of chemical elements obtained through scanning electron microscope-energy dispersive spectrometry analysis facilitate their identification in different regions, enabling the construction of a localized mineral map. According to the mineral map, the grain size of minerals decreases with depth in the sampled well, with the Doig Formation characterized by the presence of more carbonate minerals than the Montney Formation, which contains more clay minerals.
- 2) Samples from similar depths exhibit comparable mechanical properties, with Young's modulus and hardness showing a linear correlation. For the ratio of Young's modulus to hardness, samples from the Montney Formation at 2336 m show the highest values, that from the Montney Formation at 2265 m the lowest, whereas the samples from the Doig Formation at 2068 m exhibit intermediate values.
- 3) A correlation is observed between the mechanical properties of the rock and mineral composition: an increase in carbonate minerals (calcite and dolomite) enhances hardness, whereas higher clay mineral content (muscovite and chlorite) reduces hardness but increases Young's modulus. Additionally, increased albite content is associated with a lower Young's modulus.



The findings highlight the influence of mineral composition on the localized mechanical properties of rock formations, revealing the effects of carbonate and clay minerals on rock behaviour. Future work will include a broader set of experimental samples to investigate the impact of minerals with limited variability in this study, with the aim of developing a predictive model from which rock properties based on mineral composition can be derived.

Acknowledgments

The authors gratefully acknowledge the financial support from Geoscience BC through the Geoscience BC Scholarship program, which was essential to the progress of this study. Additional support was provided by the Natural Sciences and Engineering Research Council of Canada (NSERC) and the University of Northern British Columbia (UNBC). The authors also thank C. Kang for offering valuable feedback during the peer review process.

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Thermal History of the Liard Basin in Northern British Columbia and Western Northwest Territories (NTS 095B, 095C, 094J, 094K and 094N)

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Pinto, T.F. and Enkelmann, E. (2025): Thermal history of the Liard Basin in northern British Columbia and western Northwest Territories (NTS 095B, 095C, 094J, 094K and 094N); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 75–82.

Introduction

The Liard Basin in British Columbia (BC) and the Northwest Territories (NWT; Figure 1) is a prolific energy-resource area that hosts hydrocarbon reservoirs and anomalously high geothermal gradients (40-55 °C/km; Grasby et al., 2012). Due to the economic potential of geothermal resources, areas such as the Liard Basin are of increasing interest for exploration. This is an economic opportunity for diesel-dependent communities such as Fort Nelson, where the Tu Deh-Kah First Nation geothermal-energy project can upscale their economy by providing electricity and heat for greenhouse food production and district heating. However, unlike other well-known geothermal-resource areas in Canada (e.g., Mount Meager; Grasby et al., 2012), many questions remain about the geological processes that allowed the high geothermal gradients in the Liard Basin to develop. The aim of this project is to investigate the thermal history of the basin.

The Liard Basin is part of the Western Canada Sedimentary Basin (WCSB; Figure 1). Its history and stratigraphy are closely linked to orogenies (e.g., Klondike, Cordilleran; e.g., Mossop and Shetsen, 1994; Beranek and Mortensen, 2011), resulting in a wide range of depositional environments (e.g., deltaic, shelf, deep marine). Basement structures (e.g., Bovie Structure, Liard transfer zone) also influenced sedimentation and erosional patterns within the basin (e.g., Leckie et al., 1991; Cecile et al., 1997). The multiple phases of subsidence and erosion that are expected to have affected the basin are potentially linked to the development of the geothermal anomaly. The speed at which hot rocks at depth are brought to surface by uplift and erosion could result in a lack of time for isotherms to equilibrate. This disequilibrium, in turn, would manifest itself as high geothermal gradients in the shallow subsurface. The hypothesis being proposed is that the anomalously high geothermal gradients in parts of the Liard Basin are a consequence of the basin's

erosional history. To test this hypothesis, low-temperature thermochronology is being carried out on outcrop and borehole samples from the Liard Basin.

The dataset comprises samples of different depositional age, depth and position in relation to the Cordilleran Fold-and-Thrust Belt. These characteristics enabled an examination of different time slices of the basin's history, which in turn can be used to build a more complete picture of the basin's history through time. The main questions of this project are 1) how is the heating and cooling of the Liard Basin related to major tectonic and erosional events that took place in western Canada? and 2) is there a spatial correlation between the thermal history of the Liard Basin and the geothermal gradients observed today? To answer these questions, 1) thermochronological analysis will be performed on 16 outcrop and 4 borehole samples from across the Liard Basin to constrain their thermal history; and 2) models will be developed that explain the different thermal histories within the region. Ultimately, a basin-wide model will be created that incorporates the different thermal histories, structures, stratigraphy and geothermal-gradient variations. Understanding the thermal evolution of the basin over time and in different regions is critical to understanding when and why the basin underwent heating events that led to the development of the hydrocarbon and geothermal, or potential geothermal, resources.

Geological Background

The WCSB is a wedge-shaped basin that records the long-lived history of the western margin of the North American continent (Figure 2; e.g., Mossop and Shetsen, 1994). Its stratigraphy overlies the Precambrian Canadian Shield and is defined by a Cambrian to Middle Jurassic passive-margin sequence (Figure 2) that evolved after the Neoproterozoic–Cambrian rifting. This sequence is overlain by an Upper Jurassic to Cenozoic foreland-basin sequence (Figure 2) that developed as the result of terrane accretion and collisional tectonics during the Klondike and Cordilleran orogenies (e.g., Beranek and Mortensen, 2011; Monger and Gibson, 2019).

The Liard Basin is part of the WCSB and is bounded to the east by the north-northeast-trending Bovie Structure (Fig-

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Figure 1. Western Canada Sedimentary Basin, with the study area marked by the blue box. Basins and sub-basins after Mossop et al. (2004). Schematic geothermal-gradient contours after Grasby et al. (2012). Dashed grey line on the inset map indicates the location of the schematic cross-section shown in Figure 2. Base map was created using ArcGIS® software by Esri. ArcGIS® is the intellectual property of Esri and is used herein under license. Copyright © Esri. All rights reserved. For more information about Esri® software, please visit ">https://esri.ca/>.

ure 2), which is a fault that extends to the basement, and to the west by the Cordilleran Fold-and-Thrust Belt (Figures 1 and 2). The Bovie Structure was reactivated several times during the Phanerozoic (Leckie et al., 1991; MacLean and Morrow, 2004), which resulted in the deposition of anomalously thick upper Paleozoic and middle Cretaceous strata (Leslie-Panek and McMechan, 2021). For instance, the Carboniferous (Mississippian–Pennsylvanian) Mattson Formation is more than a kilometre thick in the Liard Basin, whereas it is only a few metres thick on the Great Slave Plain, east of the Bovie Structure (Figures 1 and 2; Leckie et al., 1991; Wright et al., 1994).

The Liard Basin is also known for its hydrocarbon resources. Examples of hydrocarbon source rocks are the Devonian Road River and Triassic Toad-Grayling formations, and examples of reservoir rocks are the Devonian Nahanni and Cretaceous Chinkeh formations (Rocheleau and Fiess, 2014). They underwent at least two phases of heating within the oil-maturity window, one in the Devonian and the other between the Triassic and Cretaceous (e.g., Morrow et al., 1993; Jiang et al., 2021). This information sheds light on the complexity of the thermal history of the basin, where multiple phases of heating occurred. The timing of such events is not fully constrained, nor is the cause of such heating.

Geothermal Systems

Heat in the upper crust is generated by decay of radioactive elements (e.g., U, Th, K) and also originated from primordial heat related to Earth's formation (e.g., Grasby et al., 2012). Geothermal systems are generally characterized by enhanced heat flows accompanied by thermal blanketing to trap heat. In Canada, the heat flows in the Canadian Shield and Canadian Cordillera are approximately 30 MW/m² and 60 MW/m², respectively (Jessop, 1990). However, several locations in western Canada have much higher values, such as the Garibaldi Volcanic Belt region (Mount Meager,



Figure 2. Schematic cross-section across the Cordilleran Fold-and-Thrust Belt, Western Canada Sedimentary Basin and exposed Canadian Shield (dashed grey line in the Figure 1 inset). Stratigraphic units after Mossop et al. (1994) and Rocheleau and Fiess (2014).







southwestern BC; >200 MW/m²) and the Liard Basin and Great Slave Plain (70 MW/m²; Figure 1; Grasby et al., 2012). In the Liard Basin, thermal blanketing is key to the development of geothermal anomalies within lower conductivity rocks (i.e., coal, shale) that allow thermal trapping. The origin of the geothermal anomalies is not well understood. The subject has been studied extensively, considering parameters such as basin-wide fluid flow, basement heat production and the presence of younger magmatic rocks (e.g., Majorowicz and Jessop, 1981; Majorowicz, 1996). The most recent hypothesis is that basement heat production is higher than previously thought (Majorowicz, 2018). However, no evidence for this has been found.

Methods

Low-temperature thermochronology is used to constrain the thermal history of upper crustal rocks. These dating methods are temperature-sensitive and can therefore record processes that thermally affect rocks. Examples of heating-related processes are burial and hydrothermal activity, and cooling can be caused by rock exhumation and denudation. In this study, zircon (U-Th)/He (ZHe) analysis constrains heating and cooling between 140 and 200 °C (Reiners et al., 2004; Guenthner et al., 2013).

The ZHe is based on alpha decay of the parent nuclides uranium-238 (238 U), uranium-235 (235 U) and thorium-232 (232 Th) to their stable daughter isotope helium-4 (4 He, alpha particle; Farley et al., 2002). The dates determined by ZHe analysis will be a function of the thermal history of the rock. In the case of sedimentary rocks with complex thermal histories (postdepositional heating due to burial and cooling due to erosion), the dates are interpreted based on the extent to which the rock was exposed to temperatures in the Partial Retention Zone (PRZ; ZHe, 140-200 °C; Reiners et al., 2004; Guenthner et al., 2013). In sedimentary rocks, each grain already carries a cooling age that records its source-rock cooling history. If burial after deposition reaches temperatures >PRZ, the He is lost by diffusion (reset; Figure 3, red line), resulting in dates younger than deposition that record cooling after burial. If maximum burial is limited to temperatures <PRZ, most He is kept, resulting in dates older than deposition (not reset), and each grain records their source region (Figure 3, blue line). If temperatures were within the PRZ, single-grain dates will vary from younger to older than deposition age (partially reset; Figure 3, green line). Thermal history modelling should be performed to define possible time-temperature paths that statistically fit the data and further explore the rock's possible cooling/heating histories (Ketcham, 2005).

In order to obtain the zircon fraction and perform the ZHe dating, the study samples were first subjected to mineral-separation procedures. These include, in order, jaw crusher, disk mill, water table, magnetic separator and heavy-liquid separation (heteropolytungstate [LST] and methylene iodide [MI]; e.g., McKay et al., 2021). After mineral separation, the samples containing zircons were identified and dated. Individual zircons were picked under a stereomicroscope, aiming for euhedral, transparent and inclusion-free grains. However, this was not always possible due to the age and history of the rocks, which directly influence mineral yield and grain quality. After picking, width and length of the grains were measured for later use in alpha-ejection date correction (Ft correction). The Ft correction was necessary to account for He that was ejected from the grain during al-





Figure 4. Geology of the Liard Basin, showing locations of zircon (U-Th)/He dated samples. Samples are colour-coded according to their stratigraphic age. Coloured boxes mark samples with similar date patterns, the arrows indicating trends toward samples with younger dates. Geological units and structures from digital databases (Alberta Geological Survey, 2013; Lipovsky and Bond, 2014; Cui et al., 2017; Okulitch and Irwin, 2017).



pha decay. The grains were packed in Nb tubes and analyzed to quantify their daughter He content (using a He-extraction line) and parent U and Th content (using solution ICP-MS). Due to the expected complexity of the thermal history in the area, five grains were dated per sample. All analyses were conducted at the University of Calgary Geoand Thermochronology Laboratory.

Results

The 16 outcrop and 4 borehole samples taken across the Liard Basin (Figure 4) are from units that vary with respect to the depositional age. The majority are Carboniferous, with fewer Cretaceous, Triassic and Devonian ages. The results are presented for British Columbia and the Northwest Territories (Figure 4).

The depositional ages of the samples in BC are Triassic (Lundigton Formation) and Cretaceous (Dunvegan, Sikanni, Scatter and Garbutt formations). The Triassic samples present ZHe dates that range from Jurassic to Cretaceous. Due to their ZHe dates being younger than deposition, postdepositional burial temperatures reached >140 °C. These samples are the only ones in the entire study area that show a very similar range of dates (Figure 4). Conversely, all of the Cretaceous samples present ZHe dates that are Cretaceous and older. This suggests that postdepositional burial was limited to temperatures of 140-200 °C. The Cretaceous strata in the eastern part of BC yield ZHe dates ranging from middle Carboniferous to early Cretaceous. In the western part of the area, Triassic samples yield ZHe dates ranging from early to late Cretaceous. These dates portray a trend that becomes younger and less dispersed toward the west (Figure 4). This preliminary assessment of the ZHe dates in the area could indicate that the Triassic strata experienced higher postdepositional temperatures than the Cretaceous strata.

The depositional ages of samples in the Northwest Territories are primarily Carboniferous (Mattson and Prophet formations), and fewer samples are Devonian (basal clastics) and Cretaceous (unspecified). The Carboniferous outcrop samples yielded ZHe dates that generally range from Cambrian to Cretaceous. However, the range of ZHe dates varies from sample to sample, depending on their geographic location. For example, the easternmost sample has a Cambrian-Triassic ZHe date range, suggesting they are partially reset, while the westernmost sample has a Triassic-Cretaceous ZHe range, indicating that it has been reset. Overall, the ZHe dates and date range decrease toward the west (Figure 4). Based on this, it can be stated that the postdepositional burial temperatures were higher toward the west but are still limited to maximum temperatures of 140-200 °C. The westernmost sample likely experienced >200 °C. The Carboniferous borehole samples are located at depths of 1.2 km and 1.6 km. Their ZHe dates are older

than those of their outcrop counterparts, ranging from Neoproterozoic to Carboniferous, which suggests partial resetting. The Devonian borehole sample (from a depth of 2.4 km) is also partially reset but with a Cambrian–Cretaceous ZHe range. Finally, the Cretaceous borehole sample from 0.5 km presents a Neoproterozoic–Ordovician ZHe range and is therefore interpreted as not reset, having experienced postdepositional burial temperatures of <140 °C. This ZHe dataset from across the Liard Basin thus enables a better understanding of the maximum burial experienced across the basin. The upcoming thermal history modelling will link their histories with the distribution of the geothermal gradients in the basin (Figure 1).

Summary

This paper summarizes current progress in constraining the thermal history of the Liard Basin and its possible link to the development of the high geothermal gradients in the area. The southern Liard Basin presents Triassic strata that experienced postdepositional maximum temperatures >200 °C, while Cretaceous strata experienced 140-200 °C. In the northern Liard Basin, Carboniferous strata all experienced temperatures between 140 and 200 °C. However, the range of ZHe dates is gradually less dispersed and younger toward the west, suggesting that rocks in the west experienced higher temperatures than those in the east. The link between the maximum burial temperatures and the current geothermal gradient cannot be stated at the moment, given the need to quantify erosional/exhumation rates of the analyzed samples. Thermal-history modelling is currently underway to quantify the timing and rates of cooling of these rocks. Only through modelling it will be possible to constrain whether erosion was fast enough to result in isotherms not having time to equilibrate, hence generating the observed high geothermal gradients.

Understanding how geothermal anomalies develop is critical in guiding future exploration efforts. Geothermal energy is one of the most promising energy alternatives due to its potential for long-term use and low-carbon footprint. For diesel-dependent remote communities in northern BC and NWT especially, the use of geothermal heat would be beneficial in providing energy autonomy and reliability. Geothermal heat could be used by communities not only for district heating, but also for greenhouse food production and wood-pellet drying, creating many economic opportunities. Heating alone currently accounts for 80% of the energy demand (Majorowicz and Grasby, 2021), so making use of the geothermal resources would have a significant impact on the health, environment and economic and social well-being of all northern communities.



Acknowledgments

This research was funded by the Northwest Territories Geological Survey from 2020 to 2024, and by the University of Calgary. Additional support was received through Geoscience BC and NSERC grants RGPIN-2018-03932 and NRS-2018-517959 to E. Enkelmann. The authors would like to thank V. Terlaky for contributions to the project, and L. Currie and M. McMechan for providing samples for the project. They would also like to thank the Office of the Regulator of Oil and Gas Operations (OROGO), and the Core and Sample Repository of the Geological Survey of Canada for supporting the borehole sampling. The authors thank R. Hughes and B. Härtel for their constructive review that improved this report.

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Northeast BC Granite Wash Geological Carbon Capture and Storage Atlas (NTS 094A, H, I, O, P, parts of 093I, O, P, 094B, G, J, K, N)

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Sweet, N.L. (2025): Northeast BC Granite Wash Geological Carbon Capture and Storage Atlas (NTS 094A, H, I, O, P, parts of 093I, O, P, 094B, G, J, K, N); *in* Geoscience BC Summary of Activities 2024, Geoscience BC, Report 2025-01, p. 83–88.

Introduction

In 2023, Geoscience BC released the Northeast BC Geological Carbon Capture and Storage Atlas (Canadian Discovery Ltd., 2023), which included maps in both PDF and shapefile formats, as well as a comprehensive database of oil and gas pool and aquifer data. This study identified and assessed promising geological carbon dioxide (CO₂) sequestration sites in northeastern British Columbia (BC) within the Western Canada Sedimentary Basin, quantifying their CO₂ storage potential and evaluating their ability to support carbon capture and sequestration (CCS) and lowcarbon energy projects. As well, the atlas provided valuable information for policymakers, Indigenous groups, communities and industry stakeholders; recommended future CCS evaluation steps; and offered a methodology for assessing CO₂ sequestration potential in other BC basins, particularly those near major CO₂ emission sources.

The deep Granite Wash Formation (Granite Wash) was not included in the original atlas because of insufficient publicly available data. To address this data gap, Geoscience BC developed a new CCS project to assess the Granite Wash in northeastern BC (Figures 1, 2), and the technical work is being undertaken by Canadian Discovery Ltd. The primary goals of the project are to 1) regionally map the Granite Wash, 2) identify areas of northeastern BC where the Granite Wash has sufficient porosity, permeability and net reservoir thickness to be considered a potential CCS target and 3) calculate the CO_2 storage potential for two Granite Wash areas of interest (AOIs). This paper summarizes the project's key activities, expected outcomes and outputs.

Project results are expected to be made public in March 2026.

Geology of the Granite Wash

The Granite Wash is a diachronous lithostratigraphic unit composed primarily of siliciclastic rocks that unconformably rests atop a Precambrian basement comprising granites and metasedimentary rocks. The Granite Wash sands were deposited as part of an alluvial to braided plain in a deltaic and shallow marine environment surrounding an exposed Peace River arch land mass (Trotter and Hein, 1988). Regionally, Granite Wash siliciclastic units thin across the arch, thicken on the flanks and thin toward the distal edges. The structure of the underlying Precambrian basement influenced the deposition of the Granite Wash (O'Connell et al., 1990); the siliciclastic sediments typically blanketed the paleotopography of the Precambrian draping over highs and filling in lows (Dec et al., 1996). The Granite Wash can therefore be very thin or absent where the Precambrian was highly emergent at the time of deposition; these areas of nondeposition of the Granite Wash are referred to as 'bald highs'.

The siliciclastic units of the Granite Wash are composed predominantly of siltstones and fine-, medium- and coarsegrained feldspathic and quartz-rich sandstones. The most porous and permeable zones with the best potential to store CO_2 occur in the sandstones. The Granite Wash siliciclastic units are radioactive due to the high potassium content of the feldspar clastic component and are characterized by high readings on gamma-ray well logs (Figure 3).

In the study area, seals (caprocks), which acts as traps, occur both locally and regionally and at several levels in the subsurface. The seals overlying the potential storage reservoir need to provide containment and be geomechanically stable. Of particular note in the area is the Cretaceous Shaftesbury Formation, a thick regional shale unit that extends throughout the study area and is situated below the deepest groundwater level. This unit can act as a containment layer, preventing CO_2 migration to the surface.

Project Methodology

Sedimentological and stratigraphic data will be obtained from geophysical well logs, geological well files and core descriptions. Data from Alberta (four townships to the east of the BC border) is included in the evaluation, as there are wells that penetrated the Granite Wash and the well files can provide additional information. Gamma-ray, sonic, neutron density, bulk density, microlog and resistivity wireline well logs will all be used to determine gross isopach, average porosity and net reservoir. Although limited, core

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Figure 1. Project study area (red outline), northeastern British Columbia. The CO_2 emitters are shown as filled coloured circles, with the size and colour being proportional to annual CO_2 emission volumes. The two areas of interest in this study are outlined in blue. Granite Wash Formation (Granite Wash) well data from geoLOGIC systems Itd. (2024).







Figure 3. Petrophysical well log highlighting the stratigraphic tops and the typical log responses of the Granite Wash Formation (net reservoir highlighted in red; log from geoLOGIC systems ltd. [2024]). Abbreviations: bgs, below ground surface; DEN, density; DT, sonic delta-t; GAPI, gamma API units; GRS, gamma ray; ILD, deep induction; ILM, shallow induction; SFL, spherically focused log.





and drillstem test data will also be used to further characterize the reservoir.

Through the generous donation of industry partners, over 580 km^2 of 3-D seismic and 3540 km of 2-D seismic mapping is being used to map stratigraphy and structure, focusing on the Granite Wash and Precambrian basement. Areas of existing natural faulting need to be understood and assessed as to the risk of reactivation along their faults and potential for the fault to breach the seals. This project will assess and potentially categorize the various orientations and depths of seismically identified faults.

The seismic and geological data will be combined to create a suite of maps that will be used to identify areas with the greatest reservoir potential and to inform calculations of CO_2 storage potential. Based on the geological and geophysical data accessed for the project, two areas of interest (AOIs) in the study area were identified based on the greater availability of data, which allows for more detailed mapping (Figure 1). Figure 4 is an enlarged map of the southern AOI showing the position of the Peace River arch and the location of the schematic cross-section shown in Figure 5. The schematic section in Figure 5 highlights the generalized depositional geometry of the Granite Wash. Initial isopach mapping of the southern AOI indicates up to 200 m of Granite Wash.

Detailed porosity trends will be established for the AOIs. When calculating the net reservoir for the AOIs, a cutoff of 5% porosity will be used when sufficient porosity well logs are available, and when unavailable, net reservoir will be estimated using spontaneous potential, caliper and microlog responses. Initial mapping of the southern AOI showed up to 110 m of net reservoir.

Granite Wash saline aquifer data will be vetted for CO₂ storage suitability criteria, including cutoffs for porosity and permeability, as well as sufficient depth, temperature, pressure and trapping for effective storage, and aquifers with potential for CO₂ storage will be identified. Those aquifers with the potential for supercritical CO₂ storage will need to have initial reservoir pressures greater than 7500 kilopascals and temperatures greater than 31.1° C (the critical point for CO₂). At supercritical conditions, CO₂ has high density like a liquid, but low viscosity like a gasideal for injection and storage of greater volumes as compared to storage in the gaseous phase. The CO₂ density calculations require absolute pressure and temperature at reservoir conditions from equations of state (Span and Wagner, 1996). The CO₂ densities for this study will be calculated using a web computation tool (Wischnewski, 2007).

Theoretical and effective CO_2 storage potential for the Granite Wash aquifer(s) will be provided in the report. Theoretical storage is the mass of CO_2 that can be stored in the

aquifer and is based on the mapped pore volume of the reservoir and CO_2 density. Effective storage is the capacity for CO_2 storage after accounting for various reservoir conditions and fluid properties. For regional aquifers, given the large aerial extent and the fact that injected CO_2 must displace fluid in place, the calculations for effective storage capacity are provided on a ninetieth percentile (P90; high), fiftieth percentile (P50; mean) and tenth percentile (P10; low) basis, as a function of the percentage of total theoretical storage potential. This data will be gridded and summarized on a megatonne per township (or equivalent) map.

Conclusion

Mapping of the Granite Wash Formation (Granite Wash) will augment the existing maps and database generated for the *Northeast BC Geological Carbon Capture and Storage Atlas* and contribute to Geoscience BC's objective of providing atlas-style CO₂ storage assessment reports for the geological basins across British Columbia (BC).

The deliverables of the Northeast BC Granite Wash Geological Carbon Capture and Storage Atlas project include

- 1) catalogue of all relevant geoscience reports and data;
- structural map and gross thickness map for the Granite Wash, as well as a structural map for the underlying Precambrian basement;
- detailed maps and data tables for the Granite Wash in the areas of interest, including average porosity, net reservoir and CO₂ storage potential;
- preliminary assessment of carbon storage reservoir technical risks;
- 5) identification of data gaps and recommendations for future research phases; and
- 6) public report and atlas.

This project will ultimately provide further insights into the geological framework and carbon storage potential of the Granite Wash in northeastern BC.

Acknowledgments

The project is a partnership between Geoscience BC, the BC Ministry of Energy, Mines and Low Carbon Innovation and the BC Energy Regulator, with data contributions from Canlin Energy Corporation, Logan Energy Corp. and other industry partners. The City of Fort St. John and the Canadian Hydrogen Association is also supporting the research. Geoscience BC gratefully acknowledges the financial support of the Province of British Columbia through the BC Ministry of Energy, Mines and Low Carbon Innovation. The author would also like to thank W.S. Donaldson at Canadian Discovery Ltd. for taking the time to review this paper.





Figure 4. Southern area of interest with boundaries of the Peace River arch and the location of the schematic crosssection in Figure 5. Granite Wash Formation (Granite Wash) well data from geoLOGIC systems ltd. (2024).





Figure 5. Schematic cross-section of the Precambrian to the Devonian Wabamun Formation in the southern area of interest near the Peace River arch. See Figure 4 for the location of the cross-

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