Characteristics and Origin of Rare Earth Mineralization at the Alces Lake Property, Northern Saskatchewan

A Thesis Submitted to the College of Graduate and Postdoctoral Studies In Partial Fulfillment of the Requirements For the Degree of Master of Science In the Department of Geological Sciences University of Saskatchewan, Saskatoon

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ABSTRACT

Rare earth elements (REEs) are critical metals that are essential to the low-carbon economy, especially the rapidly-growing green energy production and storage industry. REEs have been produced from a great variety of deposit types related to magmatic, metamorphic, hydrothermal, chemical weathering, and sedimentary processes. Massive monazite mineralization at the Alces Lake REE property in the Beaverlodge Domain of the Rae Province of northern Saskatchewan has attracted recent exploration and has been suggested to represent a paleo-placer type deposit, which would imply significant resources; however, this interpretation for the origin of the REE mineralization is questionable.

The massive monazite mineralization at the Alces Lake property is situated on the eastern limb of a fault truncated synform fold. This synform fold is likely related to the nearby St. Louis Fault adjacent to an unconformity with Archean aged or Arrowsmith Orogeny-related gneisses and granites on the western limb and younger Paleoproterozoic gneisses and granites alongside the Murmac Bay Group metasedimentary rocks, metavolcanics, and quartzite on the eastern limb (Normand, 2014). The REE showings occur as monazite-biotite-mineralized pegmatites, massive biotite pegmatites, granitic pegmatites, allanite-rich veinlets, and large leucogranite bodies scattered throughout the property.

The REE mineralization at the Alces Lake property is primarily in the form of monazite, zircon, and allanite. The elevated REE contents largely correspond to an increase in radioactivity caused by the thorium contents in monazite and allanite. Therefore, radioactivity can be used as an exploration tool to locate REE mineralization. Exceptions to this do occur with some less radioactive pegmatites and leucogranites containing high total REE contents (>1.5 wt.%), whereas some highly radioactive biotite-rich pegmatites have <0.5 wt.% of total REE.

Field and petrographic examinations show that the Ivan-Dante Zone has a semicontinuous pegmatite of abundant monazite and zircon with the highest discovered radioactivity and REE mineralization at the Alces Lake property. Monazite-(Ce) is characterized by 5.43-10.54 wt.% ThO₂, incorporated largely by the substitution Th⁴⁺ + Ca²⁺ = 2REE³⁺, while zircon contains 1.03-1.48 wt.% HfO₂. U-Pb geochronological analyses yielded an age of 1926±3 Ma for zircon and 1926±15 Ma for monazite, which confirm their crystallization during the Taltson Orogeny.

Therefore, the Alces Lake pegmatite-hosted REE mineralization is not a placer, but most likely had a late magmatic-hydrothermal origin related to Taltson-aged leucogranites. The leucogranites most likely formed from anatexis as a result of granulite-facies regional metamorphism during the Taltson Orogeny and were evolved to produce pegmatites and associated hydrothermal veins containing REE mineralization. The fold hinges and faults at Alces Lake provided the conduits for focused emplacement of the REE-mineralized pegmatites and hydrothermal veins.

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Contents

PERMISSION TO USEII
ABSTRACT III
AcknowledgmentsIV
List of Figures and Maps:VII
List of Tables:IX
Chapter 1 INTRODUCTION
1.1 Purpose of Thesis
1.2 Layout of Thesis2
1.3 Objectives
1.4 Methodology
1.5 Location and Accessibility
1.6 Previous Geological Work4
1.7 Overview of the Alces Lake Deposit5
Chapter 2 CURRENT UNDERSTANDING OF REE DEPOSITS
2.1 Introduction
2.2 Historic Emphasis of Research
2.3 Types of REE Deposits
2.4 Magmatic REE Deposits7
2.4.1 Carbonatitic REE Mineral Deposits8
2.5 Hydrothermal REE Deposits
2.6 Ion-Adsorbing Clay REE Deposits13
2.7 Placer/Paleoplacer REE Deposits14
Chapter 3 REGIONAL GEOLOGY OF THE ALCES LAKE PROPERTY15
3.1 Introduction
3.2 Early Archean Rocks within the Rae Craton17
3.3 Arrowsmith Orogeny17
3.4 Murmac Bay Group19
3.5 Taltson Orogeny22
3.6 Waugh Lake Group23
3.7 Sedimentary Basins
3.7.1 Thluicho Lake Group24
3.7.2 Martin Group

3.7.3 Athabasca Group	27
Chapter 4 GEOLOGY AND STRUCTURES OF REE MINERALIZATION AT THE ALCES LAKE PROPERTY	29
4.1 Introduction	29
4.2 Lithologies of the WRCB Area	
4.2.1 Orthogneiss	
4.2.2 Paragneiss	
4.2.3 Massive Biotite	
4.2.4 Pegmatites	
4.2.5 Granites	
4.2.6 Mylonitic Granite	
4.2.7 Amphibolite	
4.2.8 Massive Monazite	40
4.3 Structures at the WRCB area	41
4.4 Metamorphism and Alteration	43
4.5 History of the Alces Lake Area	
4.5 Detailed Description of REE Occurrences	45
4.5.1 Wilson and Richard Zones	46
4.5.2 Charles – Bell Zone	48
4.5.3 Ivan – Dante Zone	50
4.6. Allanite in the WRCB Area	55
4.7 Discussion of REE Mineralization	57
Chapter 5 COMPOSITIONS AND TIMING OF REE MINERALIZATION	60
5.1 Whole-Rock Geochemical Compositions	60
5.2 Compositions of Monazite and Zircon	62
5.3 Monazite and Zircon U-Pb Geochronology	69
Chapter 6 CONCLUSIONS AND SUGGESTIONS FOR FUTURE RESEARCH	73
REFERENCES	75
APPENDIX A: DESCRIPTION OF ALCES LAKE MINERALIZED SAMPLES IN THIS STUDY	85
APPENDIX B: LA-ICP-MS U/PB GEOCHRONOLGY DATA FROM ZIRCON AND MONAZITE GRAINS	86

List of Figures and Maps:

Figure 1.1: Location of Alces Lake REE property relative to the geological domains of Nort	hern
Saskatchewan and to Uranium City	4
Figure 2.1: Pie chart displaying the production of REE as of 2020 weighted by country	7
Figure 2.2: Geological map of the Mountain Pass carbonatite, USA	10
Figure 2.3 Geological Map of the Bayan Obo REE deposit, Inner Mongolia, China	11
Figure 3.1: Geological map of Northwestern Saskatchewan delineating the study area	16
Figure 3.2: Simplified diagram illustrating the tectonic activity of the Rae Craton in Nort	hern
Saskatchewan and the North West Territories	18
Figure 3.3: Geological map showing the Murmac Bay Group at the Uranium City area	20
Figure 3.4: Geological map of the Taltson Magmatic Zone	22
Figure 3.5: Geological Map showing the distribution of the Martin Group sediments	26
Figure 4.1 Geological map of the Alces Lake area	29
Figure 4.2: Structural features of the Alces Lake area	30
Figure 4.3: Geological map overlain on top of an orthophoto of the WRCB area	32
Figure 4.4: Photographs of outcrops showing the main lithologies in the WRCB area	34
Figure 4.5: Photographs of representative drill samples from WRCB area	35
Figure 4.6: Photomicrographs of major lithologies collected at outcrops in the WRCB area	36
Figure 4.7: Photographs of representative pegmatites at the Alces Lake property	38
Figure 4.8: Photographs of the Ivan and Dante outcrops showing deformation	40
Figure 4.9: Photograph of drill cores displaying monazite mineralization in the WRCB area	41
Figure 4.10: Schematic age diagram of geological events affecting Alces Lake	44
Figure 4.11: Geological map of the Wilson and Richard Zones	46
Figure 4.12: Structural map of the Wilson and Richard Zones	47
Figure 4.13: Geological map of the Charles – Bell Zone	49
Figure 4.14: Structural map of the Charles – Bell Zone	50
Figure 4.15: Geological map of the Ivan – Dante Zone	51
Figure 4.16: Structural map of the Ivan – Dante Zone	52
Figure 4.17: Photographs of massive monazite pods surrounded by chloritized biotite hoste	ed in
paragneiss at the Ivan – Dante Zone	53

Figure 4.18: Photomicrographs of radioactive pegmatites from the Ivan Zone	54
Figure 4.19: Photomicrographs of allanite mineralized core sample	56
Figure 4.20: Satellite photo of radioactive occurrences in the Alces Lake property	58
Figure 5.1: Chondrite-normalized REE plots illustrating whole-rock REE compositions	of
representative mineralized samples from the Ivan Zone	62
Figure 5.2: BSE and CL images of monazite and zircon grains in Ivan Zone samples	63
Figure 5.3: Chondrite normalized REE plots showing the REE compositions monazite-(Ce)	in
samples 805-727 and 805-735 from the Ivan Zone	66
Figure 5.4: BSE and CL images showing internal structures of representative zircon grains from	om
the Ivan Zone	67
Figure 5.5: Concordia diagram showing U-Pb data of sample grains from the Ivan Zone	71

List of Tables:

Table 4.1: 2019 World Production of REEs by Country	1
Table 4.1: Magnetitic Susceptibilities of Lithologies at Alces Lake	33
Table 4.2. Deformation history in the Beaverlodge Domain near Alces Lake	42
Table 5.1. Whole-rock compositions (ppm) of REE and other selected elements in Ivan Zone	
samples	61
Table 5.2 Compositions (wt.%) of monazite in the Ivan Zone from EMPA analyses	64
Table 5.3 Compositions (wt.%) of zircon in the Ivan Zone from EMPA analyses	68

Chapter 1 INTRODUCTION

1.1 Purpose of Thesis

Rare Earth Elements (REEs) as a strategic resource have never been of more importance to the global economy than they are now. REEs are used in a wide variety of industries including electronics, automotive components, medicine, clean energy devices, and defense, although their most important end uses are for permanent magnets and rechargeable batteries. As of 2019, it is estimated that permanent magnets account for 38% of all REE demand and this is only forecasted to continue (Natural Resources Canada Rare Earth Element facts: https://www.nrcan.gc.ca/our-natural-resources/minerals-mining/minerals-metals-facts/rare-earth-elements-facts/20522). Beyond their relative rarity to be concentrated sufficiently for economic extraction, future access to existing resources is becoming increasingly critical, due to increasing political tension between North American and European nations with China who currently possesses the world's majority of REEs (Table 1; Kynicky et al., 2012; Balaram, 2019). Consequently, there has been a concerted effort to secure new sources of these critical elements independent of China's control. This urgent need has spurned new exploration across the globe for viable sources of REEs as well as evaluation of previously found REE occurrences for potential mining using newly developed extraction techniques.

	Tons	Percentage of Total	
China	168000	60.7	
USA	A 43000 15.5		
Burma	Burma 26000 9.4		
Australia	22000	8.0	
Thailand	8000	2.9	
Other Countries	9700	3.5	
Total:	276700	100	

Table 1.1 2021	World Production	of REEs b	y Country
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Table 1.1: Data from U.S. Geological Survey, Mineral Commodity Summaries, January 2022. (https://pubs.usgs.gov/periodicals/mcs2022/mcs2022-rare-earths.pdf)

Historically, the most important sources for REEs have been carbonatites, alkaline igneous systems, ion-adsorption clay deposits, and monazite-xenotime placer deposits (Balaram, 2019). Though Canada is currently not a major producer of REEs, it is estimated to possess large reserves of REO (Rare Earth Oxides) in excess of 15 million tonnes. Coupled with its political stability, developed infrastructure, and export friendly deposition, Canada is ideally placed to emerge as a global leader in REE production, assuming viable REE resources can be found. To date, the only producing REE mine in Canada is the Nechalacho deposit of Cheetah Resources Ltd, which commenced its extraction of bastnaesite concentrates in June 2021. The Nechalacho deposit is the upper part of the larger Thor Lake REE-Y-Be deposit in the Mackenzie Mining District of the

Northwest Territories, approximately 100 km southeast of the city of Yellowknife (Pinckston and Smith 1995; Sheard et al. 2012; Möller and Williams-Jones, 2017).

One such potential REE resource is the Alces Lake REE property in northern Saskatchewan. The Alces Lake property contains some of the highest reported grades of REE anywhere in the world and may represent a major source of REE; however, the circumstances leading to its mineralization is poorly understood. This thesis aims to provide details and context for the Alces Lake REE mineralization, contributing both to the better understanding of REE mineralization and to hopefully aid in the development of future REE mines in Canada.

1.2 Layout of Thesis

The basic layout of the thesis starts with a description of the objectives, methods, and history of Alces Lake in Chapter 1, followed by a brief review of the current understanding of REE deposits in Chapter 2 to provide a general framework for the Alces Lake REE property. Chapter 3 provides an overview of the regional geology of the Beaverlodge Domain and Chapter 4 focuses on the local geology at Alces Lake. Chapter 5 discusses the geochemistry and U-Pb monazite and zircon geochronology of REE mineralization from samples gathered in the main WRCB area. Proposed theories on the origins of the REE mineralization at Alces Lake and recommendations for further research will be presented in Chapter 6.

1.3 Objectives

The Archean and Paleoproterozoic rocks of northern Saskatchewan are a potential source for increasingly critical REE resources. Small scale occurrences of REEs have been identified over the decades of Uranium exploration surrounding the Athabasca basin (Normand, 2014). One such REE occurrence is the massive biotite and massive monazite hosted, high grade REE occurrence at Alces Lake, which has been tentatively interpreted by Normand (2014), to be paleoplacer in origin.

The primary objective of this thesis is to provide details on the current understanding of REE deposits and how they relate to the REE occurrence at Alces Lake. Of particular interest are the lithological types found at Alces Lake, the mineralogy, petrography, and geochemistry of the rock types, the extent of the REE mineralization, and the ages of the high-grade monazite and zircon mineralization. These ages will be compared to ages obtained from nearby Archie Lake which has also been interpreted as a paleoplacer (Normand, 2014).

The field relationships and petrogenesis of the Alces Lake deposit will be explored via field mapping, core logging, petrographic examination of various lithologies both at surface and at depth, and an examination of the high-grade REE bearing monazite and zircon mineralization through electron microprobe analysis (EMPA), and laser ablation inductively coupled plasma micro spectrometry (LA-ICP-MS) to provide constraints on their genesis. During the 2021 field season, geological field mapping was conducted across the Alces Lake property by myself and other geologists working on behalf of Appia Rare Earths and Uranium Corp. The purpose of the field mapping was to visit every radiometric anomaly that was discovered using aerial radiometric

survey while simultaneously identifying the major lithologies and collecting structural measurements of the major and minor folds and faults in the area. Any outcrops that displayed sufficiently high radioactivity, had visible monazite, or were otherwise of any interest were sampled and sent to the Saskatchewan Research Council (SRC) Geoanalytical Laboratories in Saskatoon, Saskatchewan for the determination of their REE contents and whole-rock geochemical analysis using the ICP-MS2 package.

1.4 Methodology

Geological mapping and oriented core logging of diamond drill holes intersecting the areas of mineralization provided information of the lithologies and how they are expressed at depth. Fieldwork provided "ground truth" for aerial radiometric and magnetic geophysics and was used to confirm the presence of radioactive pegmatites each containing variable amounts of REE. Structural measurements taken in the field provide strong evidence for their relationship to the main sequences of folding in the region. Monazite- and zircon-rich channel samples taken from the Ivan-Dante Zone, which has the highest grade REE showing yet discovered at Alces Lake, were made into polished thin sections for petrographic and EMPA analyses, along with detailed backscattered electron (BSE) and cathodoluminence (CL) imaging.

Of the polished thin sections that were investigated by EMPA, two were chosen for laser ablation inductive coupled plasma mass spectrometry (LA-ICP-MS) U-Pb dating representing the highest percentage of monazite and zircon mineralization in pegmatites. Due to the large size of the monazite and zircon grains, multiple spots were chosen per grain for EMPA and LA-ICP-MS. This allowed for evaluating any heterogeneity in composition and age within individual grains.

In addition to the field work, 39 diamond drill holes were completed at the "WRCB" (Wilson-Richard-Charles-Bell) area, which includes the Ivan-Dante Zone, on the Alces Lake property. These holes were completed to variable depths dependant on their intersection with a basal unit of quartzite which defined the drill hole depth limit. All drilling was done perpendicular to strike to the dominant fabric developed during regional metamorphism as lithology bedding was largely parallel to the fabric. Drilling at this orientation maximized the value of any potential intercepts with REE bearing layers as it was interpreted based on core obtained from previous drilling efforts to more likely represent true thickness. During logging of these diamond drill holes any sections that have high counts of radioactivity (>250 counts per second) as measured by a handheld scintillometer were selected for analyses at the SRC using the same packages for the surface samples.

1.5 Location and Accessibility

The Alces Lake property is located 35 kilometres east-northeast of Uranium City. Winter weather permitting, the area is accessible from late March to Mid-December via scheduled flights to the Uranium City airport followed by a helicopter flight to the site or via float plane directly to the Appia Corp. campsite that is located along the eastern shore of Alces Lake. Approximately 132 kilometres to the southeast is the town of Stony Rapids, which can also be used to access the

Alces Lake property via helicopter or float plane. Stony Rapids maintains an all-weather road that can reach the rest of the province if need be. The Alces Lake property itself claims a 35,420-acre area. The WRCB area, which is the primary focus of this study, is located approximately 2 km south from the shore of Alces Lake.



Figure 1.1 Location of Alces Lake REE property relative to the geological domains of northern Saskatchewan and to Uranium City (UC). Grey dashed lines represent major shear zones. Modified after Ashton (2010), Bethune et al. (2013), and Maxeiner et al. (2021).

1.6 Previous Geological Work

Initial discovery of the REE mineralization at the Alces Lake property was made by J.H Wilson in 1954, who was conducting airborne uranium exploration. With the use of explosives, he exposed a number of monazite-rich trenches alongside biotite-rich segregation bands in gneissic rocks. These trenches were approximately 1.2 to 1.8 meters wide and measured up to 91 meters long. Donald A. W. Blake, a geologist working for the Geological Survey of Canada, visited the area in 1955 to perform additional exploration of the nearby Oldman River area. He spent a year mapping and classifying the lithologies during his evaluation of Oldman River for further uranium exploration (Blake, 1956).

In 1966, Numac Oil and Gas would return to the trenches blasted by J.H. Wilson and perform assays of the monazite mineralized bands. As noted by Charles Normand in his 2014 report of the area, the assay returned 2.91% ThO₂ (2.56% Th), 0.04% Y_2O_3 (315 ppm Y), 0.046 and 0.03% U_3O_8 (390 to 254 ppm U), and <15 ppm Sc (Norman, 2014).

In 2010, Charles Normand of the Saskatchewan Geological Survey visited Alces Lake and performed field mapping of the area. In his 2011 report, he described three separate types of monazite mineralization, which he encountered over a 300-meter strike length (Normand, 2011). The first was minor monazite in strongly deformed granitoid pegmatites. The second was <30-

centimeter-thick monazite-(Ce) rich layers growing increasingly rich in biotite-allanite-pyrite towards the layer's contact with the gneissic country rock. The third type of monazite occurred in biotite rich layers up to 2 meters thickness exposed in two trenches - presumably the same trenches blasted by J.H. Wilson. Bulk rock chemical analysis was performed on two samples taken from the trenches yielded 13 wt. % and 29 wt. % TREE (Normand, 2011). Following Charles Normand's report, the property was staked by Scott Bell of La Ronge in 2013, who took some samples of the trench and subsequently partnered with Appia Rare Earths & Uranium for further exploration of the property. Mr. Bell made his trench samples from the Ivan Zone available for the initiation of this study in 2014 (Appendix A).

1.7 Overview of the Alces Lake Deposit

The Alces Lake REE property is a complex system of many hundreds of discontinuous, highly deformed, occasionally sheared, variably radioactive pegmatites hosted in an intensely metamorphosed rock of dominantly granitic composition. These pegmatites are almost universally oriented north-north east consistent with the axial plane of the large fold whose fold nose lies in close proximity to the main biotite-monazite mineralized outcrop described by J.H. Wilson and Charles Normand. Alongside these pegmatites, centimeter scale allanite-rich bands or veinlets appear erratically occasionally traceable for tens to sometimes a hundred meters. The REEs are hosted mainly in monazite and allanite with zircon containing HREE in monazite rich zones. Total Rare Earth Oxide (TREO) percentages are highly variable with some samples reporting values as high as 15 wt. % TREO before sharply falling to as little as 0.6 wt. % TREO, reflecting a highly heterogenous distribution of monazite and allanite.

Chapter 2 CURRENT UNDERSTANDING OF REE DEPOSITS

2.1 Introduction

Discovering relatively large occurrences of minerals hosting notable quantities of rare earth elements (REEs) is not uncommon in the Earth's crust (Balaram, 2019). Rare earth elements occur in a wide variety of settings and are associated with many geological processes. A REE deposit that hosts minerals that are able to be efficiently processed without causing lasting environmental damage is rare. Finding a REE deposit that is also sufficiently concentrated to make the endeavour worthwhile is considerably rarer.

Our collective understanding of the processes leading to the mineralization of REEs in increasingly diverse settings is growing, aided through organizations such as the newly founded Critical Minerals Mapping Initiative (CMMI) which includes Canada, the United States, and Australia (https://www.usgs.gov/centers/gggsc/science/critical-minerals-mapping-initiative-cmmi). Careful examination of the various REE occurrences that occur across Canada and how they relate to our current understanding of how REEs mineral deposits form is of great academic value and will be a useful tool for future exploration and study. It is important to broaden our understanding of rare earth deposits and to not consider them purely in a historic context. Exclusion of potentially viable resources due to their lack of obvious similarities with more established settings can lead to either a misclassification of a resource to make it fit one of these models, or to a potential dismissal of what could have been a producing resource.

2.2 Historic Emphasis of Research

With production of REE mineral dependant technologies only predicted to rise in the coming years, exploration and development of new REE resources to meet surging demand must be ongoing to satiate the increasingly hungry market (<u>https://www.nrcan.gc.ca/our-natural-resources/minerals-mining/minerals-metals-facts/rare-earth-elements-facts/20522</u>).

Economic REE deposits have historically been associated with either alkaline/peralkaline igneous systems, carbonatites, ion-adsorption clays, and placer deposits, with the latter three being the most important. Prior to the 1960s, monazite-bearing placer deposits supplied the world's limited demand for REEs. With the 1964 discovery of the Mountain Pass carbonatite deposit, the United States would become the world's leader in REE production. This would continue until 1985 whereupon China would rapidly surpass the United States as the world's dominant producer and would maintain a de facto monopoly on REE production until 2010 (Van Gosen et al., 2017), when China drastically cut down its REE exports for political and environmental considerations.

REEs occur with an average concentration in bulk continental crust of 125 ppm (Rudnick and Gao, 2004). The collective name of rare earth elements is therefore a misnomer. They are "Rare" only in that a few of their members, such as thulium and lutetium, are actually rare. These elements however are 200 times more enriched in the Earth's crust than gold (Haxel, et al., 2002).



Instead, the rarity of REE is owed to the rarity of the REE hosting minerals that are suitable for mining and extraction (Chakhmouradian and Wall, 2012).

Figure 2.1 Pie chart displaying the production of REE as of 2021 weighted by country. Data from U.S. Geological Survey, Mineral Commodity Summaries, January 2022. (https://pubs.usgs.gov/periodicals/mcs2022/mcs2022-rare-earths.pdf)

Carbonatites remain the world's most important source of REEs with China's massive carbonatite deposits – most notably the world's largest REE deposit Bayan Obo, followed by Australia's Mount Weld deposit and the Mountain Pass deposit in USA (Fig. 2.1).

2.3 Types of REE Deposits

With such abundance, it is to be expected that the geological settings in which REE deposits can form are more varied than many other deposits currently known. The versatility, with which REEs can become concentrated - either magmatically, hydrothermally, or through sedimentary processes - is rivaled only by the seemingly endless explanations put forward for these same occurrences in the literature. Across the world there are also many notable REE deposits associated with regional and global scale geological events and processes whose origin is not yet decisively known.

2.4 Magmatic REE Deposits

The definition of magmatic REE deposits is that crystallization of REE-bearing primary minerals must be coeval with the magmatic melts/liquids in which they crystallize (Mariano and Mariano, 2012). As a result, there are a few notable magmatic REE occurrences.

Alkaline magmas have been demonstrated to have an increased solubility of high field strength elements (HFSE) and have much higher concentrations of REEs, Y, Zr and Nb (Sheard,

2010). As such, alkaline magma is an important source for many REE/HREE deposits, though rarely is it sufficiently concentrated itself to be a viable REE deposit.

The cumulate REE-rich zircon layers that occur at Thor Lake in the Nechalacho deposit are an example of a magmatic REE deposit. The Nechalacho deposit occurs in the Blachford Lake Intrusive Complex which intrudes the Slave Craton in the North West Territories. Zircon and eudialyte occur in separate cumulate layers that are interpreted to represent the injection and cooling of separate phases of an alkaline and peralkaline-rich magma, respectively (Sheard et al. 2012). The cumulate grains of zircon are interpreted by Sheard et al. (2012), to have formed before the pegmatitic zircon, crystallizing during the convection, overturns and injection of fresh magma pulses into the intrusive body. This resulted in the layered, alkaline, Thor Lake syenite with monomineralic crystallization of zircon. These zircon-dominated layers formed cumulates via gravitational settling out of the melt. Coupled with the zircon layers, eudialyte-dominated layers formed from the more peralkaline magma pulses that formed separate cumulate layers. HREE are more mobilized in alkaline melt than LREE (which tend to be mobilized in hydrothermal settings) with cumulate zircon becoming concentrated in HREE while the pegmatitic, lower layers are concentrated in the LREEs. However, viable metallurgical process is currently lacking for the extraction of HREE from zircon- and eudialyte-dominant ores (Zhang et al. 2016).

2.4.1 Carbonatitic REE Mineral Deposits

REE deposits associated with carbonatites comprise the majority of the world's supply of REE partly because of the size of the Mesoproterozoic Bayan Obo REE-Nb-Fe deposit (though its exact classification as a carbonatite is disputed), and the smaller deposits found in the Himalayan Mianning-Dechang Orogenic Belt (Yuan et al., 1992; Liu et al., 2020). There are several important examples associated with carbonatitic settings in North America, including the Niobec (Tremblay et al., 2015), Montviel (Nadeau et al., 2015), Wicheeda (Trofanenko et al., 2016), Bear Lodge (Moore et al., 2015), and Mountain Pass deposits (Castor, 2008). All of these examples include REE mineralized carbonates, yet they differ widely in terms of their geological setting and origin. Carbonatitic REE deposits can originate from magmatic settings, hydrothermal systems, supergene enrichment, or some combination of these three processes (Nadeau et al., 2015).

Carbonatitic magmas, as well as alkaline/peralkaline igneous bodies, are usually enriched in REE and act as a potential source for REE + Nb, Th, and U. As an important source of REE, these igneous rock types are associated with a wide variety of REE deposits across the world (Kynicky et al., 2012; Lai et al., 2012; Sheard et al., 2012; Gysi and William-Jones, 2013; Nadeau et al., 2015). Mineralization associated with carbonatitic magma is a complicated process as both the emplaced magmas can serve as a source for mineralization (either in the immediate host rock or in the dykes themselves) or carbothermal fluids (that being fluids enriched with respect to CO₂) expelled from the intrusive body and its resulting interactions with the host rock (Lai et al., 2012; Fan et al., 2014). Because the distinction between carbothermal and magmatic REE mineralization processes in carbonatites is either difficult or intimately associated, the two can

easily be confused and carbonatites will frequently contain at least some carbothermal/hydrothermal components.

Indeed, according to the strict definition of magmatic REE deposits posited by Mariano and Mariano (2012), the only known example of a purely magmatic carbonatitic REE deposit is the Mountain Pass carbonatite located near the Nevada and California border within the Southwest Mojave Desert (Fig. 2.3)(Mariano and Mariano, 2012; Nadeau et al., 2015).

The Mountain Pass LREE deposit is the second largest LREE deposit in the world and is a carbonatitic intrusion composed of multiple enriched dykes and plutons that belong to the Sulphide Queen carbonatite. As ultrapotassic magmas are commonly enriched with respect to REEs, the origin of the Mountain Pass deposit is tied to the origin of the magma itself, and there are several proposed explanations for the generation of these unusual ultrapotassic magmas (Castor, 2008). It has been proposed that ultrapotassic rocks, as seen at Mountain Pass, are derived from a mantle sourced alkaline parent melt assimilating thickened crust before emplacement (Castor, 2008). The exact means by which the Mountain Pass magma was generated is beyond the scope of this paper; however, it is enough to note the importance of ultrapotassic magmas to the generation of alkaline magma suites that may lead to the genesis of REE deposits.

LREE at Mountain Pass are primarily hosted in bastnaesite, with lesser amounts of parisite, synchysite, monazite, and allanite (Castor, 2008; Mariano and Mariano, 2012). Proven reserves are 16.7 million tonnes at 7.98 wt% REO contained within the Sulphide Queen carbonatite, which is a ~75m thick tabular sill dipping at 45° (Mariano and Mariano, 2012). The carbonatitic intrusion is composed of a series of 1.38 Ga carbonatitic dykes that lie in a narrow 130-kilometer-long belt hosted in granulite facies gneiss and schist. This belt is characterized by the occurrence of potassic and ultrapotassic rocks that are relatively unmetamorphosed and unusually rich in LREE and Th (Castor, 2008). The source for the magma that crystallized the Mountain Pass deposit is considered to represent a metasomatized mantle derived magma emplaced at ~2-3 kilometers depth (Castor, 2008). The Mountain Pass deposit is relatively well understood whereas the massive Bayan Obo deposit typifies the complexity of a carbonatitic REE deposit.



Figure 2.2 Geological map of the Mountain Pass carbonatite, USA. Modified from Castor (2008).and Olson et al. (1954)

Discovered in Inner Mongolia in 1927, the Bayan Obo deposit is, relative to other carbonatite deposits, average in its REE concentration though massive in size. With reserves of 48 million tonnes at ~6 wt% TREO grade, the Bayan Obo REE deposit is by far the largest in the world (Kynicky et al., 2012; Fan et al., 2014; Nadeau et al., 2015). The Bayan Obo deposit is dominated by LREE in bastnaesite and monazite along with lesser Ca-REE and Ba-REE flourocarbonates and is temporally and spatially related to carbonatitic-syenite complexes (Kynicky et al., 2012). The exact age of mineralization at Bayan Obo is complicated due to the multiple stages of mineralization. Some have suggested an age as young as ~500 Ma for its first stage of mineralization (Kynicky et al., 2012), while others have suggested mineralization as old as ~1400 Ma (Fan et al., 2014). Also, as the REE minerals are situated mostly in dolomite, which is layered, conformably bound by upper and lower contacts, and has sedimentary features, yet has an igneous geochemical profile the origin of the REE mineralization at Bayan Obo deposit is still debated (Lai et al., 2012).



Figure 2.3: Geological Map of the Bayan Obo REE deposit, Inner Mongolia, China. Modified from Fan et al. (2014).

The currently-favoured carbonatitic origin of the Bayan Obo deposit is extremely complex and is comprised of 11 stages including multiple intrusions, metamorphic events, and fluid infiltration (Kynicky et al., 2012). REE mineralization at Bayan Obo occurs in four types of dolomites, as recognized in the field. These are sedimentary carbonates, fine grained marble, coarse grained metasedimentary marble that is geochemically comparable to carbonatites and carbonatitic dykes (Kynicky et al., 2012; Lai et al., 2012). This can be further complicated by the presence of "transitional carbonates" which, based on their geochemical profile, cannot be easily classified as igneous or sedimentary in origin (Lai et al., 2012). The Bayan Obo group itself is comprised mostly of sandstones and slates alongside the REE hosting dolomite unit that are believed to have been deposited during the Mesoproterozoic Bayan Obo marginal rift. Sedimentation comprising the host rocks at Bayan Obo is bound by an upper age limit of approximately 1900 Ma (Fan et al., 2014).

Although the interpretation of the Bayan Obo deposit is contentious, it is clear that Bayan Obo formed, at least in large part, due to the interaction of carbonatitic magma with dolomite (Kynicky et al., 2012; Lai et al., 2012; Fan et al., 2014;). One proposal suggests that the REE mineralized dolomites at Bayan Obo formed as a result of hydrothermal metasomatism of marine dolomitic sediments. The source for the REE-rich hydrothermal fluids was from the intrusion of carbonatitic dykes. There are currently about 100 such dykes that intruded throughout the mineralized dolomite rock unit (Fan et al., 2014). These intrusions are interpreted to have developed during the Mesoproterozoic rifting of the North China Craton. This is supported by the geochemistry of the most intensely REE mineralized dolomites being similar to igneous rocks and the high REE contents of the dykes themselves (Lai et al., 2012). As is typical when discussing REE deposits, the distinction between magmatic and hydrothermal processes are blurred as the carbonatitic dykes themselves are enriched in REE. It is reasonable to suggest that magmatism directly resulted in the mineralization of the LREE. However, a magmatic origin does not account for the extent of the REE mineralization and the characteristic fenitization (consisting of sodic amphiboles, aegirine, albite, and phlogopite) surrounding the dykes which are indicative of hydrothermal processes (Lai et al., 2012).

2.5 Hydrothermal REE Deposits

Hydrothermal REE deposits are the most diverse in terms of their geological setting as the processes by which REE can become concentrated and deposited by fluid-flow systems are both varied and complex. Many proposed REE models include a hydrothermal component, either for transportation or further concentration of REE. Most often hydrothermal processes are considered in magmatic or pegmatitic deposit types, where it is difficult to distinguish the limits or extent to which mineralization was influenced or controlled by secondary fluids, particularly if primary igneous textures have been overprinted by hydrothermal processes as is commonly the case in pegmatitic occurrences.

Fluids are enriched in REE in peralkaline and alkaline igneous complexes. Fluids that are derived from these systems, or that interact with these systems following deposition, are a commonly considered source for hydrothermal REE deposits (Kynicky et al., 2012; Sheard et al., 2012). REE contained within igneous bodies can be mobilized under many different conditions and can be transported by fluids via structural features to form distal deposits with no obvious association to the source rock (Pandur et al., 2016). These fluids can also be enriched via fluid/host rock alteration reactions, such as by the formation of acid haloes surrounding intrusive bodies and pegmatites that increase rock porosity and enable mobilized fluids to form in distal regions surrounding the host rock (Gysi and William-Jones, 2013). The means by which REE can be mobilized in hydrothermal fluids depends on the specific temperature, pressure, pH and salinity conditions and the aqueous complexes available. Due to the presence of volatiles inherent in magmatic systems, associated REE deposits likely contain a hydrothermal component in the ore mineralization excepting those granites undergoing weathering such as is the case of ion-adsorbing clays (which formed in ambient surface environments).

Based on the reaction pathway observations that Gysi and William-Jones (2013) made on the Strange Lake pegmatite-hosted REE deposit, hydrothermal fluid interactions can be roughly divided into LREE and HREE interactions. In general, the mobility of LREE is controlled by the stability of fluorocarbonates, and the mobility of HREE is controlled by the zirconosilicate group and secondary gadolinite group minerals (Gysi and William-Jones, 2013). The solubility of each REE-bearing mineral in these mineral groups varies according to its environment and subtle changes in temperature and fluid chemistry control mineralization. Simulations performed by Gysi and William-Jones (2013) indicated that the mobilization of REE and Zr is favoured at low pH. However, REE mobility is favoured at temperatures in excess of 400°C, whereas Zr mobility is favoured by lower temperatures at ~250°C. Mobility is enhanced by the presence of $(OH)_2F_3^{2-1}$ which forms a complex with Zr and REE. The resulting REE fluoride complexes are less mobile at high temperatures, except in fluids that also contain HCl. In this case, mobility is enhanced with temperature due to the competing speciation of the HCl complex. Hydrofluoric complexes also likely play a role as they, too, reduce liquidus and solidus temperatures and enhance mobility of REE bearing fluids by decreasing viscosity and enhancing H_2O solubility in melts. Accompanying an increase in the mobility of REEs and other HFSE, these volatile phases also act to increase the fractionation of the fluid, further concentrating these elements (McKeough et al., 2013). While this model was used to explain the occurrence of REE enriched zones of pegmatites at Strange Lake, this same model can be applied to other predominantly hydrothermal systems that occurred under similar conditions.

2.6 Ion-Adsorbing Clay REE Deposits

Ion absorbing clay-type REE deposits account for the majority of the worlds HREE production (Li et al., 2019; Feng et al., 2022). The largest and most well-known of this type of REE deposit are the HREE-enriched residual clay deposits in southern China, namely the Huanglongpu and Huayangchuan deposits belonging to the Qinling belt (Kynicky et al., 2012). The REE-rich clays are formed by the in-situ alteration and disintegration of REE-bearing granitic host rock from intense tropical weathering followed by physical adsorption on negatively charged surfaces on aluminosilicate clays (Moldoveanu and Papangelakis, 2016). As outlined by Kynicky et al., (2012), development of this type of deposit requires minimal erosion of the clays following deposition, as well as a source of REE-enriched minerals susceptible to chemical weathering. These minerals include REE-bearing bastnaesite, monazite, apatite, xenotime and allanite (Moldoveanu and Papangelakis, 2016). Secondary REE- rich minerals that form from the weathering process (such as fluorocarbonate and phosphates) will be deposited with the clay in addition to REE that are adsorbed on to the surface of the clays via cation exchange at the clay surface. These adsorbed REE constitute the majority of the REE resource in the deposit and are steadily enriched by adsorbtion-desorption processes that concentrate HREE due to preferential adsorption of cations with larger atomic radii (Li et al., 2019).

The weathering conditions that form these deposits produce REE in three forms, which are colloid-phase REE (REE are deposited as insoluble oxides or hydroxides as part of a colloidal

polymeric organometallic compound), exchangeable-phase REE (REE occur as soluble free cations on the clay surface, and which account for the majority of REE in ion adsorption clay deposits), and mineral phase-REE (REE in minerals from the host rock that have survived the weathering process, usually occurring in distal regions to the main deposit) (Kynicky et al., 2012; Moldoveanu and Papangelakis, 2016; Li et al., 2019; Feng et al., 2022).

Although lower in grade to the other REE deposit models, the low cost of processing and low concentration of radioactive elements, such as Th and U, make ion adsorbing clay type REE deposits very attractive as a potential source of HREE (Moldoveanu and Papangelakis, 2016). The appropriate climatic conditions are required because these types of deposits need intense weathering, related to acidity, microbial activity, temperature and humidity (Kynicky et al., 2012). These types of deposits also likely require rocks that are Mesozoic to Cenozoic in age because the weathering processes that enable these deposits is ultimately self-destructive and they are unlikely to be preserved in pre-Mesozoic age rocks (Kynicky et al., 2012). Subtropical zones similar to those in southern China such as can be found in Madagascar, Brazil, and across South East Asia where it is hypothesized that other HREE deposits may exist given the presence of a suitable REE-bearing host rock have become areas of great interest (Moldoveanu and Papangelakis, 2016). Currently exploration projects are ongoing at Serra Verde, Brazil and the TRE project on Nosy Be Island, Madagascar (Natural Resources Canada, 2016; Moldoveanu and Papangelakis, 2016).

2.7 Placer/Paleoplacer REE Deposits

REE deposits that occur in placer/paleoplacer settings are arguably the most widespread in the world. Many of the REE-bearing minerals, such as xenotime, monazite, and zircon are mechanically durable and can be readily accumulated in sediments. REE placer deposits are particularly attractive because the cost of extraction and processing the REE-bearing minerals is much lower than in the hardrock REE deposit types (Bern et al., 2016; Zhang et al., 2016). Because of these two factors, many placer deposits that have been mined for other minerals have been examined as potential sources of REE (Bergen and Fayek, 2012).

The processes by which mechanically durable minerals undergo concentration by detrital accumulations in sedimentary settings are relatively well understood. Placer deposits form from the weathering and recycling of less durable ore-bearing or nonmineralized rocks and subsequent transportation and concentration by sedimentary processes involving wind or water (Foley et al., 2015; Simandl, 2014). In theory, placer REE deposits have the potential to be both high tonnage and high concentration based on the relative abundance of REE, yet the majority of the most enriched REE deposits occur in small, localized alluvial settings (Foley et al., 2015).

Chapter 3 REGIONAL GEOLOGY OF THE ALCES LAKE PROPERTY

3.1 Introduction

The geological evolution of the Rae Craton is both poorly understood and complicated. Remote outcrops in the difficult to navigate northern Canadian wilderness contribute to this collective lack of understanding in the literature. However, the geological significance of the massive Rae Craton is equally daunting. The Rae Craton represents a vast period of time in the geological evolution of the Canadian Shield and contains within itself incredible diversity in terms of the observed rock types and the inferred environments of its constituents. The Rae Craton in northwest Saskatchewan records at least two Meso-Paleoproterozoic orogenic events, numerous sedimentary basins, both felsic and mafic intrusions, volcanics, and many regional metamorphic and structural events. Providing a thorough summary of all that has occurred in the Rae Craton is beyond the scope of this thesis. As the Rae Craton is so large, this summary will confine itself to the region stretching from the northern Alberta Border of the Taltson Domain to the Black Lake shear zone along the Tantato and Dodge Domains in northwest central Saskatchewan (Fig. 1.1). In addition, the study area itself is confined to northwest Saskatchewan and so the extensive history and lithologies comprising the Rae Craton will be narrowed so as to focus on that area.





Figure 3.1: Geological map of northwestern Saskatchewan delineating the study area. At the margin of the Rae Province, Rae and Hearne rocks are characterized by the severity of metamorphism and the timing of these metamorphic events. Rae Craton rocks were distinctly metamorphosed to granulite facies from two separate events: a 2.55 Ga event and a 1.91-1.90 Ga event, due to the Taltson-Thelon Orogeny (Ashton et al., 2012); this map was created from a composite of maps from Jefferson et al. (2007), Ashton et al., (2009a;b), and Maxeiner et al. (2021).

3.2 Early Archean Rocks within the Rae Craton

The Rae Craton is a geologically diverse assemblage of many different rock types in northwestern Saskatchewan. The Rae Craton is predominantly comprised of Meso - Neoarchean, amphibolite to granulite facies, felsic orthogneiss with northeast striking greenstone belts. These belts extend from the southern Athabasca Basin to Baffin Island, with felsic to mafic plutons (ca 2.68 – 2.32 Ga) being common (Thiessen et al., 2018).

One of the most notable aspects of the Rae Craton in northwestern Saskatchewan (Fig. 3.1) is that it is host to a variety of rare Paleoarchean to Mesoarchean rock types. These occurrences are geologically significant and represent one of the few ways that we gain insight into the formational processes that occurred during the early stages of the Archean. The rarity of these rocks and their value provide insight into the geological processes during the early Earth. Studies of the radiogenic isotope signatures of the Rae Craton have been conducted to locate potential areas for these elusive rocks. These rocks are found within the tectonically complex Arrowsmith Orogeny zone, and the younger 2.0-1.9 Ga Taltson-Thelon Magmatic Zone in the western Beaverlodge Belt within the Churchhill Province (Hartlaub et al., 2005; Ashton et al., 2012).

Prior to the development of the Arrowsmith Orogeny, there is evidence of older suturing events with some rocks in the northern Rae Craton, such as in the vicinity of Baffin Island which have been dated to 3.6 Ga or older (Ashton et al., 2012; Thiessen et al., 2018). Based on a U-Pb dating study of detrital zircon in granitic, rocks near Uranium city, ages >3.7Ga, were reported with intrusive bodies of similar age located beneath the Murmac Bay Group, notably the 3072 ± 41 Ma Lodge Bay, 3014 ± 10 Ma Elliot Bay and Cornwall Bay Granites (Hartlaub et al., 2005). These intrusive bodies are primarily exposed near Beaverlodge Lake, in the Crackingstone Peninsula, and along the north shore of Lake Athabasca (Ashton et al., 2009a).

Therefore, the Rae Craton is a composite of various accretion events, orogenies, and sutures between rocks that have subsequently undergone multiple, extensive deformational and metamorphic processes across a broad unit of geologic time. One of the best examples of these suturing events is the Black Bay Shear Zone. The Black Bay Shear Zone is the surface expression of a suture between a Paleoarchean to Mesoarchean crustal block and Neoarchean crust that corresponds to the Rae and Hearne Cratons, respectively (Hartlaub et al., 2005; Ashton et al., 2012). These geological events will be explored below.

3.3 Arrowsmith Orogeny

The ca. 2.5 - 2.3 Ga Arrowsmith Orogeny is a magmatic belt located in the Beaverlodge and Zemlak Domains. The rocks linked to the Arrowsmith Orogeny extend across most of the Rae Craton all the way to Baffin Island with an overall strike length of approximately 2000km (Fig, 3.2). However, despite its size the Arrowsmith Orogeny is poorly understood as it has been overprinted by many younger thermotectonic episodes over the history of the Rae Craton (Hartlaub et al., 2007). The Arrowsmith Orogeny includes two protracted, dominant magmatic

phases: a 2.5-2.45 Ga pulse and a 2.3 Ga pulse (Pehrsson et al., 2010; Ashton et al., 2012; Berman et al., 2013).



Figure 3.2: Simplified diagram illustrating the tectonic activity of the Rae Craton in northern Saskatchewan and the North West Territories ca 2.56 – 2.26 Ga. Also shown is the location of Alces Lake, SK. Modified from Berman et al. (2013).

The processes that led to the formation of the Arrowsmith Orogeny are currently being debated, resulting in many proposed formational processes in the literature. Hartlaub et al. (2007) proposed that the orogenesis was related to the collision of unidentified terranes along the western margin of the Churchill Craton occurring in relation to subduction and syn to post collisional magmatism ca. 2.33-2.29 Ga of granites. Schultz et al. (2007) interpreted the Arrowsmith Orogeny as having occurred in an extensional setting to explain the Tibetan plateau-like characteristics of the Queen Maud Block in northern Laurentia. A study by Berman et al. (2013), interpreted the Arrowsmith Orogeny to be a Cordilleran-like accretionary continental arc, which is possibly occurring on a global scale. This interpretation is correlated by monazite dating of foliation in various magmatic rocks occurring throughout the orogeny, zircon dating of granites, and the geochemical similarities between granitic plutons. An example would be those granites observed in the Beaverlodge and Zemlak Domains, notably the 2.32-2.28 Ga Buffalo Head terrane in Alberta and the North West Territories, to the interior granites produced by the Cordilleran arc

(Hartlaub et al., 2007; Berman et al., 2013). However, granitic rocks of similar geochemical signature are not exclusive to continental-arc and collisional arc settings, and can form in a variety of tectonic settings (Chacko et al., 2011). In northwestern Saskatchewan, this collisional event continued until sometime before the 2.2-2.1 Ga mafic magmatism, at which point rifting occurred until the 2.0-1.9 Ga Taltson Orogeny.

Metamorphic events associated with the Arrowsmith Orogeny are dated to be younger in the Beaverlodge and Zemlak Domains (2.47-2.28 Ga) relative to the northern regions (2.54-2.35 Ga). This implies that the timing of the collision was not uniform across the entirety of the Arrowsmith Orogeny, which is reasonable given its extent (Berman et al., 2013). However, the timing of the regional metamorphism was coeval with the timing of the periods of major magmatism, and the majority of the magmatic rocks in the Beaverlodge Domain have been dated to ca. 2.3 Ga (Hartlaub et al., 2007; Berman et al., 2013). Rocks associated with the Arrowsmith Orogeny are a mixture of mafic and felsic intrusive rocks, and a few of the rocks, particularly the North Shore Plutons, have a Nd isotopic signature indicative of magma mixing of juvenile and older crust, consistent with an arc setting (Hartlaub et al., 2007).

Following the Arrowsmith Orogeny, a number of sedimentary successions are found throughout the Rae Craton that appear to be the products related to this event. These post orogenic successions are made more complicated in their study as they have the added context of environment/depositional setting, regional correlation and less well constrained ages in addition to metamorphic and deformational processes (Ashton et al., 2012). The Murmac and Waugh Lake groups are two such sedimentary successions, with the Murmac being correlated to the Arrowsmith Orogeny and the Waugh Lake being correlated to the Taltson Orogeny (Pana, 2010; Ashton et al., 2012).

3.4 Murmac Bay Group

The Murmac Bay Group is a post-Arrowsmith supracrustal sedimentary succession that is a component of the Beaverlodge Domain, best exposed in lower amphibolite rocks near Uranium City (Fig. 3.3) (Hartlaub et al., 2004; Ashton et al., 2012). Although the Murmac Bay Group is most well known near Uranium City, it occurs in numerous outcrops throughout the study area. However, it is principally found in the western Beaverlodge Domain (Card et al., 2007).

The lithologies of the Murmac Bay Group itself consist of a quartzite-basalt-pelite succession with basal quartzite, psammite intercalated with dolostone, patchy iron formations, mafic flows, and overlying psammopelite-pelite coupled with ca 2.3 Ga and 1.9 Ga felsic intrusions known as the North Shore plutons (Hartlaub et al. 2004, 2007; Ashton et al. 2009b; Ashton and Card, 2010; Ashton et al., 2012). The Murmac Bay Group unconformably overlies a predominantly granitic basement ca. 3.0 Ga in age which is itself intruded by multiple granites, mafic dykes, and komatiites (Hartlaub et al., 2004; Ashton et al., 2009a). The Murmac Bay Group also overlies the 2321±3 Ma Gunnar Granite (Hartlaub et al., 2007).



Figure 3.3: Geological map showing the Murmac Bay Group at the Uranium City area. Note that the Alces Lake property is located 34km to the east-northeast of Uranium City outside of the map boundary. Modified from Hartlaub et al. (2004).

Determining the age of the upper and lower Murmac Bay Group has been problematic due to the number of deformational and metamorphic events that have occurred in the area, obscuring primary geological relationships and resetting ages. The Murmac Bay Group itself likely represents a significant unconformity that can be up to 320 Ma (Shiels et al., 2017). This unconformity separates the Murmac Bay Group into the upper and lower respectively and likely represents a change in the sedimentary source between the two groups (Shiels et al., 2017). The lower group contains zircons of early Paleoproterozoic and Archean age while the upper group only contains zircons aged <2.33 Ga (Shiels et al., 2017).

There have been a number of interpretations as to the upper and lower Murmac Bay Group's age. Recent re-examination of U/Pb ages from zircon obtained from the Donaldson Lake granite which crosscuts the Murmac Group indicates that the upper Murmac Bay Group has a maximum depositional age of 1999 +/- 13 and a minimum depositional age of 1.93 Ga (Shiels et al., 2017). Based on the work by Ashton et al., (2013) and Bethune et al., (2013), the maximum age limit of the lower Murmac Bay Group is 2323 +/- 2.3 Ma from samples of the basal conglomerate zircons, while samples from zircons of a pelite provided a minimum depositional age of 2171 +/- 31 Ma.

The depositional setting of the Murmac Bay Group has been interpreted on the upward lithological and stratigraphic progression coupled with co-relatable detrital zircon U-Pb ages (Ashton et al., 2012; 2013; Shiels et al., 2017). The basal conglomerate, defining the earliest

known member of the Murmac Bay Group, is believed to have been deposited in a high energy and/or high relief environment due to the size and poor sorting of its clasts. The source of these clasts was the erosion of exposed hypabyssal phases of the 2.33 Ga granites, described by Hartlaub et al. (2004), and/or extrusive felsic flows (Ashton et al., 2012).

Quartzite in the Murmac Bay Group likely formed from sandstones in a distal fluvial or shallow marine environment due to the rare presence of planer and trough crossbedding, ripple marks and scour channels as well as the presence of calcareous-dolostone-quartzite sequences observed near the top of the quartzite, which is consistent with these environments (Hartlaub et al., 2004; Ashton et al., 2012). The exact source for the quartzite is not clear. Detrital zircon characteristics of the Arrowsmith associated felsic intrusions or zircons found in pre-Arrowsmith Paleo-Mesoarchean basement, which have been observed in the basal conglomerate, are not observed in the quartzite. If these sediments formed in an extensional depositional setting, the quartzite may represent recycling of older quartzite sequences (Ashton et al., 2012).

Following the shallow marine quartzite in the sedimentary sequence, psammatic gneiss occurs locally interbedded with quartzite, (observed at Macintosh Bay) at the contact and intruded by gabbroic dykes. These dykes are interpreted to be feeding dykes that supplied mafic flows stratigraphically above the psammite (Ashton et al., 2012). Detrital zircons from the psammite indicate that the source for these sediments was from an unknown ca. 3.0 -3.3 Ga basement, presumably older than the Murmac Bay Group Basement, and certainly older than the collisional granites produced by the Arrowsmith Orogeny (Hartlaub et al., 2004; Ashton et al., 2012). Detrital zircons that correspond to the ca. 2.5 - 2.35 Ga period, which contains the most Arrowsmith produced granites, are absent in the psammite. This was interpreted by Ashton et al. (2012), to imply that these granites were too distal to contribute to psammite deposition and/or that the rocks had not yet been uncovered to contribute to sedimentation.

The extrusive igneous rocks overlying the quartzite exhibit both cooling cracks and pillow structures in different areas, characteristics that are indicative of both sub-aerial and sub-aqueous eruption indicative of either varying sea levels or differences in relief (Ashton et al., 2012). Based on the sedimentary textures exhibited in the Murmac Bay Group, the depositional environment was likely that of a developing basin with early fluvial and shoreline sediments progressing to lower energy pelitic, fine grained sediments. This is consistent with the proposed origin of the Arrowsmith Orogeny being a compressional arc, gradually moving from near shore/higher energy environments, implied by the basal conglomerates and quartzites, to low energy marine environments, implied by the pelite. The Murmac Bay Group, along with the Tantato and Beaverlodge Domains, underwent a regional, high grade metamorphic and deformational event at 1.90 Ga, producing mylonite (Ashton and Hunter 2003; Card et al., 2007). This metamorphic event distinguishes the basement upon which the sedimentary basins - the Thluicho Lake and Martin Group(s) - are deposited.



Figure 3.4: Geological map of the Taltson Magmatic Zone and Thelon Magmatic Zone alongside the various tectonic elements and cratons of western Laurentia. GLSZ is Great Lake Shear Zone and STZ is Snowbird Tectonic Zone. Modified from Chacko et al. (2011) and Card et al. (2014).

3.5 Taltson Orogeny

As with the Arrowsmith Orogeny, the Taltson Magmatic Zone (Fig. 3.4) covers a wide extent of the Rae Craton, and its formational processes have been the subject of considerable debate. Originally considered part of the Thelon Orogeny, the Taltson Magmatic Zone has been interpreted as being the result of the 2.0 – 1.9 Ga subduction of the Slave Craton and Buffalo Head terrane beneath the Rae Craton (Card et al., 2007; Card et al., 2014). The rock types in the Taltson Magmatic Zone are a collection of older, dioritic to quartz dioritic I–type granites and younger S-type granitic rocks that are highly variable in their quartz contents and Sm – Nd isotopic signatures (Theriault 1990; Card et al., 2014).

Early explanations proposed that the Taltson and Thelon Orogenies were representative of a Himalayan type compressional tectonic regime. In this regime I-type granites were generated in a continental margin arc setting following the closure of an ocean basin and the collision of the previously separated continental crusts that generated peraluminous S-type granites. This was based on the chemical classification of the granitic components in the Taltson and Thelon Orogenies being similar to Himalayan analogs (Gibb and Thomas 1977; Hoffman 1987, 1988; Theriault 1990; Chacko et al., 2011).

Chacko et al. (2011) noted that the mantle component observed in continental margin collision produced granites was absent in plutons of the Taltson Magmatic Zone. They proposed an intra plate setting for the formation of the Taltson Orogeny. This was based on the evolved Sm – Nd isotope signature of the older 1.986 – 1.959 Ga I –type granites and the increased silica content observed in the Taltson plutons in Alberta (Chacko et al., 2011; Card et al., 2014). The negative Sm-Nd isotope signature and the silica composition of the I-type granites are more characteristic of partial melting of continental crust rather than a subduction generated melt (Thériault, 1990; Chacko et al., 2011). The partial melting to generate these granites would rely on continent-continent collisions that produced significant crustal thickening due to thrusting. This intracratonic thrusting was likened to the Tian Shan Belt in central Asia, which has considerable (50-55 km) crustal thickening. As noted by Chacko et al. (2011), the Tian Shan Belt is also associated with foreland basins that are the proposed sedimentary environment of the Thuicho, Martin and Athabasca Basin Groups.

However, it has been argued that the geochemical trends observed in the Taltson Magmatic Zone I-type granites can still be observed in subduction generated granitic plutons, and can be explained by crustal contamination of the melt as it migrates through the crust (Thériault 1992; McDonough et al., 2000; Card et al., 2014). Based on a study by Card et al. (2014) on the overall geochemical composition of the I-type granites coupled with an extensive aeromagnetic and geochronological study of the progressively more evolved I-types granites they concluded that the intermediate composition of the granites was unlikely to be the product of an intra-plate collision. The oldest plutons in the Taltson Magmatic Zone are characterized as being more primitive and compatible with arc generated granites but became more evolved over the course of the collision due to steadily increasing crustal contamination of the magma. This crustal contamination imparted the negative Nd profile and contributed quartz to the melt. Based on the extensive work that has been performed on the Taltson Orogeny, the interpretation that the Taltson Magmatic Zone is a product of continental margin plate collision seems likely. In the study area, rocks belonging to the Taltson Orogeny cover a very limited extent but the thermotectonic overprint and deformation affects most of the Rae Craton. The resulting sedimentary basins define most of the Zemlak and Beaverlodge Domains.

3.6 Waugh Lake Group

The Waugh Lake Group, known primarily from Alberta but also in small patches at the Alberta border in northwest Saskatchewan, is interpreted as a product of the Taltson Orogeny. The Waugh Lake Group has been dated to ca. 2.01 - 1.97 Ga and likely formed in a back arc basin (McDonough et al., 1997; Pana, 2010). The Waugh Lake Group consists of large, mylonitized shear zones and igneous sediments that were produced from the erosion of Taltson granitoids and have experienced at least greenschist facies metamorphism. These rocks are a major component of

shear zones that occur along the Taltson Magmatic Zone and are best seen in northwest Saskatchewan near the Tazin River shear zones (Pana, 2010).

3.7 Sedimentary Basins

Within northern Saskatchewan, there are numerous post-Taltson sedimentary rock groups that are known within the Zemlak and Beaverlodge Domains. Despite the extensive metamorphism and deformation that occurred in these regions, the sedimentary rocks that were preserved primarily in fault-controlled basins are remarkably well preserved (Bethune et al., 2010). The three best known sedimentary basins in northwest Saskatchewan are the Thluicho Lake Group, Martin Group, and Athabasca Group.

Besides these groups, metasedimentary rocks within the Beaverlodge and Zemlak Domains are commonly reported, but the exact relation they have to these groups is still unclear. While the exact area and nature of these sediments are still being investigated, they are largely considered to have been deposited in environments that formed due to the Paleoproterozoic crustal reworking of Archean terranes initialized by the Taltson Orogeny (Morelli et al., 2009). Based on numerous studies examining these sediments, it has been proposed that, prior to erosion, these sedimentary basins covered a large area across the mountain belt. These sediments characteristically lie unconformably over a crystalline basement rock composed of a variety of Archean gneiss, mylonite, and Paleo to Mesoarchean plutons that have all been extensively metamorphosed to, at the least, greenschist facies (Bethune et al., 2010).

3.7.1 Thluicho Lake Group

The Thluicho Lake Group (Fig. 3.1) is composed of coarse grained, siliciclastic rocks and is the oldest of three post-Taltson, Paleo to Mesoproterozoic sedimentary successions preserved within fault bound basins in northwest Saskatchewan (Hunter, 2007). These sediments lie on an angular unconformity over an upper amphibolite crystalline basement rock characterized by having been affected by at least two regional deformational events and by ca. 1.93 Ga late-Taltson, greenschist to amphibolite facies metamorphism (Ashton and Hunter 2004; Bethune et al., 2010).

The initial deposition of the Thluicho Lake Group sediments is believed to have coincided with the onset of the D2 event that uplifted the Taltson associated granitic rocks (Card et al., 2007; Bethune et al., 2010). After the D2 event enabled deposition, the Thluicho Lake Group sediments underwent two more deformational events, D3 and D4, before the Martin Group deposition began. The maximum age of deposition is defined by a Pb²⁰⁷/Pb²⁰⁶ isotope study of zircons within the base of the group at 1922 +/- 6 Ma (Bethune et al., 2010). This occurs between the 1.93 Ga thermotectonic front, correlated to the metamorphism of the Thluicho basement rocks and the development of the Black Bay Fault between 1.91 and 1.90 Ga. The minimum age is defined by the 1.82 Ga Uranium City dyke swarm which intrudes through the Thluicho Lake Group sediments (Ashton and Hunter 2004; Hunter, 2007; Bethune et al., 2010).

The Uranium City dyke swarm defines the maximum age of the Martin Group and represents the feeder dykes to the basalt flows found in the Martin Group (Ashton and Hunter, 2004). The Thluicho Lake Group primarily occurs within two fault bound, structural basins, called the Thluicho and Gulo Lakes basin and the Waterloo Wellington Basin. Besides these basins, sediments belonging to the Thluicho Lake Group have also been observed in small, isolated patches along the north shore of Lake Athabasca, near the Burntwood Group, and near the Tazin River fault (Ashton and Hunter, 2004; Hunter, 2007; Bethune et al., 2010).

The Thluicho Lake Group sediments occur as approximately 1.5 km thick sequences but the original thickness of the Thluicho Lake Group may have been as thick as 5 km. This is due to the observation of Thluicho Lake Group sediments being recycled into the neighbouring Martin Group and the interpreted environments of these systems (Hunter 2007). Despite the metamorphism and deformation that occurs within the Thluicho Lake Group, sedimentary features, such as ball and pillow structures and bedding, are preserved and have been used to determine the depositional environment of this group (Hunter, 2007; Bethune et al., 2010). Based on these structures, the Thluicho Lake Group is interpreted to have been deposited within a retrograde, intermontane alluvial basin from the erosion of the mountain belt produced from the Taltson Orogeny. These sediments grade from a basal alluvial fan conglomeratic environment, to high energy braided streams producing arkose, to a river delta system producing argillite that was deposited into a lake (Hunter, 2007). Interestingly, the Thluicho Lake Group sediments do not contain the 1.90 Ga zircon associated with late-stage plutons of the Taltson Orogeny. This implies either that the plutons containing these zircons were not exposed at this time or that the deposition of the Thluicho Lake Group occurred in the foreland of the Taltson arc, separated by a continental divide (Bethune et al., 2010). While the exact position of the Thluicho Lake Group in relation to the Taltson Orogen is unclear, the proposed depositional environments seem consistent with a mountainous setting.

Despite occurring in a relatively small area in comparison to the extent of the Taltson Orogeny, the Thluicho Lake Group likely covered a vast area but is now only persevered within a few structural basins. This is likely due to extensive erosion. However, besides the Thluicho Lake Group, there are other isolated sediments within the Zemlak and Beaverlodge Domains that may be co-relatable with the same sedimentary processes. One such group is the Burntwood Group, which occurs on the north shore of Lake Athabasca and Slate Island. The ca. <1.92 Ga Burntwood Group has been interpreted as being coeval with the Thluicho Lake Group deposition (Ashton and Hunter, 2004). Based on correlations between the Thluicho Lake and Burntwood Groups of northwest Saskatchewan and the Nonacho Group that occurs in the North West Territories, it is likely that these sediments represent the remnants of erosion from the uplifted mountains produced from the continental collision of the Taltson-Thelon Orogeny (Ashton and Hunter, 2004).



Figure 3.5: Geological Map showing the distribution of the Martin Group sediments. The red arrow indicates the direction to Alces Lake which is located approximately 20 km beyond the eastern edge of the map. Modified from Maxeiner et al. (2021).

3.7.2 Martin Group

The Martin Group is a continental redbed, composed of a variety of siliciclastic rocks interbedded by mafic volcanics that were deposited in a fault controlled intracratonic basin. (Morelli et al., 2009; Ashton and Card, 2010; Bethune et al., 2010). The contact between the Martin Group sediments and the underlying crystalline basement varies from a sharp angular unconformity to a poorly defined, graded regolith (Morelli et al., 2009). Deposition of the Martin Group sediments is believed to have been initialized by the 1.84 Ga terrane accretion of the Fort Simpson-Nahanni Terrane in Alberta and the 1.83 Ga Trans-Hudson Orogeny in eastern Saskatchewan and Manitoba (Ashton and Card 2010). This also roughly coincides with the Martin Group basaltic flows that are interpreted as having been fed by the 1.82 Ga Uranium City dyke swarm (Ashton and Hunter, 2004). Therefore, the Uranium City dyke swarm, which defines the minimum age of the underlying Thluicho Lake Group, also defines the maximum age of the Martin

Group (Bethune et al., 2010). While not very well defined, they are currently the most accepted ages for the onset of deposition.

The Martin Group igneous rocks, which include mafic dykes and sills, occur over an ~3000km² area and occurs in outcrops surrounding Uranium City (Fig. 3.5) (D'Souza 2012). Geochemically, the Martin Group igneous rocks are consistent with mafic volcanics that formed in an extensional setting and are heavily enriched in LREE but depleted in HFSE (Morelli et al., 2009). These igneous rocks are classified as tholeiitic basalts to basaltic andesites and may be related to the geochemically similar Uranium City mafic dykes. Morelli et al. (2009), proposed that the Uranium City mafic dykes and the Martin Group igneous rocks were related but represented slightly differently evolved magmas based on minor differences in composition. It was observed that the Uranium City mafic dykes are slightly more tholeiitic, incompatible element deficient and compositionally uniform in comparison to the Martin Group igneous rocks.

It has also been suggested that due to geochemical similarities, the Martin Group igneous rocks are related to the ultrapotassic plutonic rocks in the Baker Formation of the Dubawnt Supergroup in Nunavut and may be regionally related to the massive Christopher Island Formation (Ashton et al., 2009a; D'Souza 2012). The Martin Group igneous rocks (1833±3 Ma, U-Pb zircon age), Baker Lake igneous rocks (1812 Ma, ⁴⁰Ar/³⁹Ar phlogopite age), Uranium City mafic dykes (1818 Ma, U-Pb zircon age), and the Christopher Island Formation (1830-1810 Ma) all have a strong temporal correlation in addition to their compositional similarities (Rainbird et al., 2006; Morelli et al., 2009).

Besides the geochemically similar igneous intrusions, there is also considerable similarity both in the timing and depositional setting between the sediments of the Martin Group and the Dubawnt Supergroup. Located ~850km to the northwest of the study area, the Dubawnt Supergroup is a fault associated silicicalstic redbed that likely formed in the same, intracratonic, fault controlled basin setting as the Martin Group (Ashton et al., 2009b). The two also share remnants of an older redbed succession, implying they both recycled detritus from the same source. Unfortunately, linking the two stratigraphically is challenging because they both formed as a result of fault induced processes that obscure stratigraphic relationships. Nevertheless, it is highly likely that the two are related based on their sedimentary characteristics, temporal relationship and mafic volcanic similarity (Rainbird et al., 2006; Ashton et al., 2009b; Morelli et al., 2009; D'Souza, 2012).

3.7.3 Athabasca Group

While the Athabasca Basin represents a significant geological event in Saskatchewan - hosting world class uranium deposits - these sediments will not be examined in detail in this thesis. The Athabasca Group sediments are primarily located on the south shore of Lake Athabasca, although they also occur in small localities along the north shore of Lake Athabasca as well (Cumming and Krstic, 1992; Ramaekers, 2004; Yeo et al., 2007). In contrast to the rock groups found in the Beaverlodge and Tantato Domains, the Athabasca Group sediments are characteristically
unmetamorphosed (Card et al., 2007). Deposition of the Athabasca Group occurred between 1730 Ma and 1600 Ma within a series of four unconformity bound basins which are nearly entirely composed of sandstone successions (Ramaekers, 2004, Jefferson et. al. 2007). These four unconformity bound sandstones occur in five fluvial deposition systems that are described in detail by Yeo et al. (2007), with Jefferson et al. (2007) providing an excellent summary of the processes enabling mineralization.

Chapter 4 GEOLOGY AND STRUCTURES OF REE MINERALIZATION AT THE ALCES LAKE PROPERTY

4.1 Introduction

The Alces Lake property (Fig. 4.1) of Appia Rare Earths and Uranium Corporation (Appia hereafter), which is located approximately 34 km east of Uranium City, is situated in the Murmac Bay Group in the Beaverlodge Domain, in close proximity to the St. Louis Fault (Fig. 4.2). The Alces Lake REE occurrence discovered by J.H. Wilson was staked in 2012 by Mr. Scott Bell (prospector of La Ronge, SK) and was subsequently acquired by Appia. The original trench that was blasted by J.H. Wilson contains highly radioactive (>60,000 CPS) massive monazite described by Blake (1956). Channel sampling by Appia confirmed the occurrences of massive monazite and discovered elevated REE in surrounding biotite pegmatites. This historic occurrence was initially referred to as the Ivan Zone. Subsequently, additional radiometric surveys and stripping by Appia exposed another REE occurrence approximately 25 meters away containing pods of red massive monazite, which appear to follow the same general trend as the monazite pods exposed in the Ivan Zone (though much smaller and discontinuous). These two occurrences were hereafter referred to collectively as the Ivan – Dante Zone (Fig. 4.3).



Figure 4.1 Geological map of the Alces Lake area (modified after Blake 1956). The total claim area for the Alces Lake property (marked by the red line) is 57,048 hectares.



Figure 4.2: Structural features of the Alces Lake area (modified after Blake 1956).

Additional stripping and radiometric surveys by Appia discovered an outcrop of highly radioactive (30,000 – 40,000 CPS) massive biotite located approximately 150 meters to the southwest of the Ivan-Dante Zone. Geochemical analyses confirmed that this outcrop of massive biotite contains elevated TREO and the occurrence was termed the Wilson Zone (Fig. 4.3). Another occurrence of similar massive biotite with high radioactivity (>20,000 CPS) is located approximately 50 meters to the northwest of the Wilson Zone, and is hereafter referred to as the Richard Zone. Because the Wilson Zone and Richard Zone are in close proximity and appear to share the same REE bearing massive biotite, they are frequently referred together as the Wilson - Richard zone. Directly north of the Richard Zone, two more radiometric anomalies were discovered and contain radioactive (>20,000 CPS) biotite pegmatites in highly deformed gneisses. Interspersed within these pegmatites are several pods of massive monazite. These radiometric anomalies have been called the Charles-Bell Zone (Fig. 4.3). These six occurrences of REE mineralization (Ivan – Dante, Wilson – Richard, and Charles – Bell) make up the area termed the "WRCB" by Appia (Fig. 4.2). Other REE occurrences have been discovered across the Alces Lake property. However, massive monazite appears to be restricted to the WRCB area with the highest REE contents contained in the original trench blasted by J.H. Wilson in the Ivan Zone.

Diamond drilling in the WRCB area has been ongoing at Alces Lake since 2018. However, only drill data collected during the 2021 season will be examined for this thesis. All drill holes were orientated with an azimuth of 73° and dips ranging from 45° to 90°. The azimuth orientation was chosen based on surface fabric and magnetic imaging data yielding a perpendicular to lithological contacts that would best represent a true thickness for any encountered lithologies. Because of the presence of magnetite, orientation was maintained by gyroscopic drill survey. The maximum hole depth (up to 210 m) was usually determined by an intersection to a unit of barren quartzite, which appears to be continuous in the WRCB area.

During 2021, over a dozen additional radioactive occurrences (reporting at least 10,000 CPS) were found across the Alces Lake property that contained elevated REE (1-5% TREO). Several occurrences (such as the Biotite Lake occurrence) will be mentioned and used to provide an example of a typical lithology, but they will not be explored in detail. As of 2022, exploration of the Alces Lake property is ongoing with many new REE bearing pegmatites being found regularly.

The terrain of the Alces Lake property varies from a lightly wooded forest of pine, birch, and black spruce, to densely wooded new growth of alders and black spruce. Old growth sections of the forest are relatively uncommon with the majority of the trees being less than twenty-five years old. Large burned areas indicating frequent forest fires are common and rock exposure in the area is excellent. The majority of the property alternates between forested inclines separating marshes, ponds, and small lakes with rocky cliffs. The cliffs vary in size, with the largest being over 50 meters in height. Evidence of abundant wildlife is common across the property.



Figure 4.3: Geological map overlain on top of an orthophoto showing the Wilson (bottom), Richard (middle), Charles - Bell (top centre) Zones, and the Ivan -Dante (top right) Zone. Also displayed in green dots are the drill collars (21-WRC-###) for the diamond drill holes at WRCB that were completed in 2021. Since its initial acquiring of the property, Appia has done extensive stripping of the topsoil and bush at WRCB to expose the outcrops. All data collected by Howard Zip (2021) and other geologists working for Appia.

4.2 Lithologies of the WRCB Area

A combination of granulite-facies regional metamorphism, migmatitization, and intense deformation makes differentiating the protoliths of various gneisses at Alces Lake and its neighboring areas likely impossible because all primary features have been destroyed. One of the key tools used for identifying lithologies at the Alces Lake property is magnetitic susceptibility, which has been observed to be quite consistent for the major rock types (Table 4.1). In this study, eight distinct lithologies are recognized on surface at the WRCB area (Fig. 4.3). Additional lithologies have been found at depth from drilling in the WRCB area; however, only those lithologies which contained significant amounts of REE will be discussed.

Major Lithologies (excluding pegmatites)	Magnetitic Susceptibility (K) 10 ⁻³ SI		
Orthogneiss	0.15 - 0.25		
Paragneiss	0.35 – 0.4		
Mylonitic granite	1->100		
Amphibolite	0.3 - 1		
Hematitized zone	1 – 25		
Quartzite	0.1 - 0.5		

Table 4.1 Magnetitic Susceptibilities of Lithologies at Alces Lake

Data collected using a Terraplus KT-10C magnetic susceptibility meter. With the exception of the hematitized zone, these values represent typical rock samples that have not underwent hematite or magnetite alteration. Pegmatite values vary considerably depending on the quartz content with quartz rich pegmatites being similar to quartzite.

4.2.1 Orthogneiss

The orthogneisses are probably Archean Arrowsmith age and are highly deformed. They are brown to pink to red on weathered surface (Fig. 4.4A), but light grey to dark grey in colour on fresh surface (Fig. 4.5A). Orthogneiss comprises the majority of the rocks on the western and southern side of WRCB, although isolated pods of orthogneiss can be found interbedded with younger (probably Paleoproterozoic Taltson) paragneiss throughout the Alces Lake property. The contact between the orthogneiss and the paragneiss, which runs directly through the Wilson Zone and the Richard Zone, is tentatively interpreted to represent an unconformity.

The orthogneiss varies from course to medium grained, and is characterized by magnetic susceptibility in the range of 0.15-0.25 (Table 4.1). In un-migmatized outcrops, the orthogneiss is granitic to tonalitic in composition often grading between the two even within a single outcrop. The orthogneiss typically contains, 0-30% quartz, 5-20% plagioclase and 0-10% K-feldspar 5 mod.% pyroxenes, 5 mod. % amphiboles, 5% biotite (Fig. 4.6A). Garnet porphyroblasts (up to 3 cm in diameter) also occur locally and are commonly altered partly to sillimanite (±cordierite), pyroxenes, amphiboles, chlorite, and biotite. Locally, an outcrop of orthogneiss can become highly garnetiferous transitioning into garnetite.

The great variability in the compositional characteristics of the orthogneiss is due to the high extent of migmatization. The formation of granitic leucosome and pyroxene- and hornblende-rich melanosome is widespread with outcrops changing repeatedly into felsic and mafic rich bands. The contacts between these bands are very sharp. The majority of the

leucosome bands are a millimeter to centimeter wide, although meter-scale migmatites can be found across the property as well.



Figure 4.4: Photographs of outcrops showing the main lithologies in the WRCB area. A) granite, B) orthogneiss, C) amphibolite in contact with a pod of massive monazite, D) massive biotite, E) pegmatite grading into quartz pegmatite, and F) biotite pegmatite in contact with a pod of massive monazite, which is in sharp contact with paragneiss.



Figure 4.5: Photographs of drill core samples illustrating representative major lithologies in the WRCB area: A) orthogneiss taken from drill hole 21-BIO-004 at depth 24.75 meters, B) paragneiss taken from drill hole 21-WRC-013 at depth 66.20 meters, C) massive biotite from drill hole 21-BIO-004 at depth 24.50 meters, D) granitic pegmatite from drill hole 21-DIA-001 at 10.07 meters, E) pegmatite at 21-DIA-001 at 13.25 meters, F) massive monazite from drill hole 21-WRC-015 at 22.26 meters, G) biotite pegmatite at 18.75 meters, and H) massive biotite containing monazite-(Ce) and allanite-(Ce) from 21-WRC-015 at 21.75 meters.



Figure 4.6: Photomicrographs (plain polarized light or PPL: A,F,H; cross-polarized light or XPL: B, C, D, E, G) of major lithologies collected at outcrops in the WRCB area: A) orthogneiss of the Wilson Zone showing quartz (Qtz), Kfeldspar (Afs), and biotite (Bio) as the major minerals, B) paragneiss containing quartz (Qtz), garnet (Grt), and sillimanite (Sill) altering garnet. C) massive biotite containing monazite grains surrounded by biotite, and D) pegmatite containing quartz and feldspars including microcline (Mc). E) granite containing quartz and feldspar with intergrowths of quartz and Kfeldspar in the groundmass. Biotite lathes occur throughout. F) mylonitic granite containing magnetite (mag), quartz, and biotite. Clusters of very fine grained (<10 micron) titanite (Ttn) and quartz occur in the groundmass. G) Amphibolite showing hornblende (Hbl), clinopyroxene (cpx), and biotite in a feldspar and sercite groundmass. H) massive monazite showing monazite and clinopyroxene in a quartz and Kfeldspar groundmass.

4.2.2 Paragneiss

The paragneiss primarily occurs on the north eastern side of the unconformity. It is brown to dark green on weathered surface (Fig. 4.4F), but light grey to dark grey on fresh surface (Fig 4.5B). Paragneiss comprises the majority of the host rocks exposed at the WRCB area. They are compositionally and texturally similar to the orthogneisses (oftentimes being identical) and do not have any sedimentary features preserved (Fig. 4.6B) Their classification as paragneiss follows that proposed by Normand (2014) and is currently based on its magnetitic susceptibility value of $^{0.35}$ (Table 4.1). In the field, paragneiss typically contains more garnet than orthogneiss and can usually be found to grade into garnetite.

4.2.3 Massive Biotite

The massive biotite rock is classified as "massive" due to its composition being dominated by biotite (>70-90% by volume) as opposed to its textural characteristics. Alternatively, the massive biotite could be classified as a biotite-schist. Massive biotite is a muddy brown on weathered surface (Fig. 4.4D), but black on fresh surface (Fig. 4.5H). Massive biotite is a fine to medium grained with disseminated monazite (with the largest grains being <1mm though typically they are too small to see in hand specimen) (Fig. 4.6C). Occurring alongside the monazite grains are clinopyroxene, orthopyroxene, and titanite which are of similar size and crystal habits. Massive biotite has a wide range of radioactivity with some samples, such as those depicted in figures 4.4D and 4.5H, recording values as high as 5000+ CPS. Uncommon xenocrysts (up to 1 mm in diameter) of K-feldspar and secondary quartz occur as well. Some massive biotite, such as that at the Ivan-Dante Zone, may represent a mylonitic shear zone.

Outside of the WRCB area, massive biotite at the Biotite Lake zone, the Oldman River zone, and the Danny Zone appears to be identical to its counterparts at the Richard-Wilson Zone and often contains visible monazite as well. However, these occurrences of massive biotite are not accompanied by massive monazite pods.

4.2.4 Pegmatites

Pegmatites are a catch-all term used to refer to the irregular, multi-generational, very coarse grained (>1 centimeter to as large as 20 centimeters in diameter) intrusive rocks that occur throughout the Alces Lake property. On weathered surface, pegmatites can vary in colour from greyish white to dark brown and in fresh surface they are usually grey to brown to dark grey depending on the amounts of biotite or pyroxenes (Fig. 4.5E). They are granitic to monzogranitic in composition (dominantly 10 – 50 mod.% quartz, 10 – 50% feldspar, 10 – 50% plagioclase, 10 – 50% K-feldspar, and 0 – 30% biotite) and may contain trace amounts of monazite, pyroxenes, titanite, or allanite (Fig. 4.6D). Chloritization is common alongside epidote, hematite, K-feldspar, and/or pyrite occurring in micro-fractures and micro-veins.

Most pegmatites are discontinuous over a 2 to 3 meters, although in the vicinity of folds and/or faults they occur in much greater abundance and frequently occur along strike of these structures. In these instances, pegmatites typically crosscut each other and form combined multigenerational pegmatite bodies that can be 10 or more meters in length and width. All pegmatites in the Alces Lake property are sheared, occurring alongside the limbs of minor or major folds (Fig. 4.7A) and becoming concentrated in the fold noses (Fig 4.7B). The grains in pegmatites typically display pure shear, but dextral shear has also been observed in the field.



Figure 4.7: Photographs of representative pegmatites at the Alces Lake property. A) large deformed feldspar grains in a sheared pegmatite exhibit pure shear along the limb of a minor fold. B) a minor fold with a small 20-centimeter pegmatite occurring at the fold nose. The pegmatite breaks through the fold, and crystalizes into a small fan shape. The axal plunge of this fold and this pegmatite is identical to the axial plane of the folds in the WRCB area and suggests that this is an F4 fold.

Pegmatites typically grade into one of two additional sub types: 1). Quartz pegmatites are white on weathered surface (Fig. 4.4E) and fresh surface (Fig. 4.5D). Quartz pegmatites contain $>70 - 100 \mod .\%$ quartz as centimeter-sized crystals and occasionally have large (mm in diameter) crystals of plagioclase and K-feldspar and may contain porphyroblasts of hornblende or biotite. Quartz pegmatites are rarely radioactive. 2). Biotite pegmatites are red to dark brown in outcrop (Fig. 4.4F) and are predominantly black to reddish brown depending on K-Feldspar alteration (Fig. 4.5G). Biotite pegmatites contain 20 – 30 mod.% biotite, but occasionally as high as >50%. Some biotite pegmatites having >70% biotite may be classified as massive biotite and

indeed the determination between these two has been muddled in drill core logging. At surface, biotite pegmatites can be seen grading into massive monazite such is as the case at the Charles – Bell Zone. Similar to massive biotite, biotite pegmatites have a highly variable radioactivity with some examples (such as the outcrop depicted in Fig 4.4F) having values as high as 5000 CPS.

4.2.5 Granites

Granites are grey on weathered surface (Fig. 4.4A) and vary in composition from granite to tonalite and monzogranite. Texturally, they can be relatively undeformed but they may grade into a well foliated gneissic rock making them difficult to separate from the orthogneiss. They can also coarsen and grade into a pegmatite or be crosscut by pegmatites of identical composition. Across the Alces Lake property, there are multiple occurrences of exposed granites, which have low (<500 CPS) to high (~1000 – 1500 CPS) radioactivity.

The granites in the WRCB area are alkaline with (>35 mod. % of feldspars) with microcline exhibiting microcline twinning and occurring as large 500 micron – 3mm sized crystals alongside equally large quartz crystals which appear brownish red in thin section (Fig. 4.6E). Thin biotite lathes (5 mod. %, 10 – 100 microns long) and euhedral titanite (<1 mod. %, 10 – 100 microns in diameter) crystals occur in a quartz rich groundmass alongside lathes of chlorite (5 mod. %, 10 – 100 microns).

4.2.6 Mylonitic Granite

In and around the vicinity of the WRCB area, Normand (2011; 2014) described an amphibolite unit in contact with the gneisses and quartzite. This amphibolite covers an area approximately 200m² along the northeastern limb of the fold. However, I did not find any large amphibolite outcrop in the WRCB area. Instead, a magnetite-rich gneissic granite (Fig. 4.6F) occurs in the WRCB area leading towards Alces Lake. This is the dominant exposed lithology in the area and, given its size and location, it was likely the "amphibolite" described by Normand (2011, 2014). Blake (1956) described a quartz-feldspar mylonite in the vicinity of the radiometric anomalies at Alces Lake. On the basis of my field observation and petrographic examination, the original description of Blake (1956) was accurate.

4.2.7 Amphibolite

The amphibolite is a dark green, fine-grained rock notable for being found in close proximity to and in direct contact with massive monazite (Fig. 4.4C, 4.8C). It occurs as highly deformed bands, commonly 5 cm to 50 cm wide, but increasing to near 5-meter-wide bands that appear in several outcrops within the WRCB area and across other zones in the Alces Lake property. They commonly display the local deformation, wrapping around fold noses and becoming pinched in areas of high strain. Amphibolite does not exhibit brecciation into any other rock types, instead becoming highly deformed. Highly deformed amphibolite in the proximity of major folds seems to accompany intense mineralization of REE bearing pegmatites and massive monazite. Many of the smaller pods in the Ivan Zone are completely enveloped by amphibolite (Fig. 4.8C).



Figure 4.8 : Photographs of the outcrops in the Ivan - Dante Zone showing: *A*, boudinaged quartz filled tension gashes in paragneiss in close proximity to amphibolite and massive monazite mineralization; *B*, a highly deformed amphibolite (green rock) in direct contact with massive monazite (bottom reddish-grey rock); *C*, small red pods of massive monazite (red rock outlined in white spray paint) contained in amphibolite that varies greatly in width from over 5 meters (closest to camera) to less than a meter (furthest from camera); and *D*, amphibolite tapering off and disappearing (at the yellow book) into quartz pegmatite (potentially quartzite).

Amphibolite consists of dominantly subhedral to anhedral hornblende (50 mod. %), orthopyroxene (10%), and clinopyroxene (10 mod. %). Biotite laths (5-10 mod. %, 100 – 1000 microns long) and chlorite lathes (5-10 mod. %, 100 - 1000 microns long) occur randomly with no obvious preferred orientation (Fig. 4.6G).

4.2.8 Massive Monazite

Similar to the massive biotite rock unit, the massive monazite lithology was termed "massive" because of the high percentage of monazite it contains (~20-80% mod.% monazite). Texturally, the massive monazite could be classified as a monazite breccia. Massive monazite is red on weathered surface (Fig. 4.4F) and red in fresh surface (Fig. 4.5 F). Massive monazite occurs as highly irregular pods ranging in size from 5 cm in diameter to as large as 5 meters long by 3 meters wide, which are primarily concentrated in the fold nose at the Ivan – Dante Zone and the biotite pegmatite clusters at the Charles – Bell Zone. Massive monazite is spatially related to the amphibolite. Compositionally, the massive monazite contains ~20-80 mod.% monazite with variable amounts of zircon, pyroxenes, and garnet in a biotite rich matrix (Fig 4.5 H).



Figure 4.9: Photograph of diamond drill cores displaying the types of monazite mineralization in the WRCB area with features highlighted: A) drill hole WRC-015 in the Richard Zone showing three intersects of massive monazite. B) monazite mineralization (5000 – 8000 CPS) but with primarily amber-coloured grains at the depth of 21.75m. C) biotite pegmatite with a radioactivity of 4000 – 5000 CPS at the depth of 19 m. D) Close-up view of massive monazite with a radioactivity of 10,000 – 14,500 CPS at the depth of 22.25 m. Abbreviations Bio, Mnz, Kfs, Qtz, Pl, and Py are biotite, monazite, K-feldspar, quartz, plagioclase and pyrite.

The massive monazite pods appear to be discontinuous at depth. Drill sections targeting massive monazite (Fig 4.9A) can have highly different intercepts, suggesting that the monazite pods have highly irregular dimensions. In core sample, massive monazite occurs as disseminated coarse to medium grained crystals (<1mm – 2mm) in a randomly oriented matrix of coarse (1mm) to very fine-grained biotite (Fig 4.9D). The thickness of the massive monazite varies based on contact depth and drill orientation, but are typically 20 cm – 50 cm thick units at contact. Massive monazite grades into K-feldspar rich biotite pegmatites (Fig. 4.9C) which may or may not contain late stage, disseminated pyrite alteration and massive biotite type mineralization (Fig. 4.9B). These biotite pegmatites are dominated by randomly orientated, fine to coarse grained biotite (>80%) with monazite occurring as disseminated, medium to coarse grains.

4.3 Structures at the WRCB area

The Beaverlodge Domain rocks at the Alces Lake area have recorded a long and complicated history of folding and faulting since their deposition. At least four phases of folding have been recognized in the region and termed F_1 to F_4 (Bethune et al., 2013). The earliest folds (F_1) are identified by Hartlaub (2004) to have been cryptic folding, which occurred during the

compression of the Murmac Bay Group. The F₂ folds are tight to isoclinal folds trending east south east, which transposed bedding in the early Murmac Bay Group that formed the earliest surviving foliation designated $S_{0/1/2}$ (Hartlaub, 2004; Bethune et al., 2013). This event (D_{1/2}) is rarely preserved at the Alces Lake property but was found in Macintosh Bay granite (Hartlaub, 2004). Following D₁/D₂ deformation, NE trending folds (D₃/F₃) that predominantly plunge northwest are recorded in the Murmac Bay Group at Uranium City. The F₄ folding is believed to have formed in a dextral, transpressive regime in response to dextral-oblique slip along the Black Bay Fault. This series of folding is considered responsible for the recrystallization and resetting of fabric about the north-east trending F₄ fold axes (Ashton et al., 2010).

Event	Macro/Mesoscopic Features in Beaverlodge Domain Rocks	Fabric Elements at Alces Lake		
D1	Early folds of bedding and the earliest observed foliation.	S_0/S_1 foliation. Not visible at Alces		
		Lake.		
D2	Tight to isoclinal folds trending east – southeast.	Transposition of foliation. Likely not		
		visible at Alces Lake.		
D3	Large scale north – northwest trending folds.	Likely not visible in the WRCB, but		
		visible around Oldman River.		
D4	Northeast trending, predominantly southwest plunging	Most prominent folding in the WRCB		
	folds of earlier ductile transposition foliation (S ₂).	area. Responsible for the north		
		trending, western plunging F₃ folds.		
D5	North trending open folds.	Realignment of mica axial planar to		
		folding. No mica samples were		
		examined for a specific stress		
		alignment.		

Table 4.2. Deformation history in the Beaverlodge Domain and Alces Lake area

Data from Hartlaub (2004) and Bethune et al., (2013).

At Alces Lake, the most prominent regional feature is the large synformal folds (though interpreted by Blake to be synclinal), which overlap south of Alces Lake towards Forget Lake. These folds, which represent the youngest folding event (F_4), plunge between 40 degrees to becoming near horizontal in some locations and trend north east (Blake, 1956). Folded rock units include all lithologies across the property except possibly some granitic intrusive bodies (though there is a distinct lack of measurable features) and lamprophyre dykes. In the WRCB area, there are two distinct phases of folding. The F_3 folding plunges west 20° (D4) with its axial plane dipping south 30°. The F_4 folds (D5) are sub-vertical, with the axial plane striking north-south and plunging south 30°.

At WRCB, the F₄ folding seems to have the most impact on the structural features with shear zones striking identically to the axial plane of the synform. Interestingly where foliation intersects the axial plane of the fold along these shear zones, massive monazite occurs at WRCB. The WRCB itself is just to the south west of the fold nose of this large synformal fold. Pegmatites frequently

occur in fold noses (either parasitic or minor to the major fold) (Fig. 4.6B). It is possible that the magmatic/hydrothermal fluids derived from the pegmatites exploited a structural point of weakness or that the low-pressure environment of the fold noses was favourable for fluid migration (Cosgrove, 2015).

At the macro scale this is best observed at the Ivan – Dante Zone, which is the nose of a minor drag fold with the lower competency of the massive monazite and potentially the amphibolite being trapped between paragneiss. Similar structurally-re-activated shear zones in the Beaverlodge area have been suggested to likely play a part in the many uranium deposits at Uranium City (Dieng et al., 2015). Brittle shearing in the rocks surrounding the St. Louis Fault, which strikes through Alces Lake, is likely due to reactivation of the Zenith Lake shear zone (Hartlaub, 1999).

Given the complexity of the region, determining the timing of events without further geochronological analyses is difficult. Massive monazite mineralization appears to be very late with the only features cross-cutting monazite pods and pegmatites being brittle faults and quartz veins/quartz tension gashes. The faults, which trend east-northeast, are similar to the trend of the St. Louis Fault.

Other rocks that may be younger are the east-northeast trending and vertically dipping lamprophyre dykes. These lamprophyre dykes appear grey – brownish red in colour, are resistive, mildly radioactive (300 - 400 CPS), and contain large crystals of biotite (>1mm) and K – feldspar (>1cm). Lamprophyre dykes do not appear at WRCB, but are common in the southern part of the Alces Lake property near Oldman River (Blake, 1956; Hartlaub, 1999). These dykes are likely the same as the lamprophyre dykes occurring in other parts of the Beaverlodge Domain described by Ashton et al. (2006), and are interpreted to be 1.8 - 1.85 Ga (Ashton et al., 2009. These late-stage east-west structures are believed to be the product of the compressional regime caused by the terminal collision of the Trans Hudson Orogeny to the east and plate interactions to the west (Ashton et al., 2009a; Ashton et al., 2010).

4.4 Metamorphism and Alteration

Granulite facies metamorphism at the Alces Lake property is consistent with the regional metamorphism of the Beaverlodge Domain (Blake, 1956). The granulite facies metamorphism was accompanied by extensive migmatization, yielding common leucosome and melanosome bands. Occasional "rafts" of relatively undeformed gneisses occur throughout the property. Partial melting is interpreted to be the result of anatexis that occurred during the ~1.93 Ga Taltson – Thelon Orogeny, which reached granulite facies and generated a variety of granitoid gneisses throughout the Beaverlodge Domain (Ashton et al., 2010).

Granulite facies metamorphism at the Alces Lake property are characterized by the common occurrences of the orthopyroxene-bearing assemblages in amphibolites and the garnet-sillimanite-cordierite-K-feldspar assemblages in the paragneiss. Epidote veins and chlorite

alteration of biotite are present in nearly all lithologies (excluding quartzite), suggesting that retrograde metamorphism occurred after the peak metamorphism.

Potassic alteration, silicification, hematization (usually paired with extensive chloritization and epidote alteration), pyritization, and other styles of alteration involving dolomitic carbonates, calcium carbonate, rhodochrosite, and fluorite have been found to occur regularly in drill cores, typically as patchy occurrences or veins. These late styles of alteration typically post-date regional metamorphism. All rock types across the Alces Lake property are subjected to late alteration, although there is no obvious association between REE mineralization and late alteration.



4.5 History of the Alces Lake Area

Figure 4.10: Schematic age diagram summarizing the relative ages of the Alces Lake REE occurrence. This summary includes inferred ages for the amphibolite and pegmatitic rock units on the context of field observations and the timing of regional deformational events occurring in the western Beaverlodge Domain. This diagram also includes LA-ICP-MS ages of monazite and zircon grain samples obtained from the Ivan Zone (see Chapter 5). Modified from Pandur et al., 2020.

Based on previous studies examining the western Beaverlodge Domain, the geological history of the Alces Lake area can be inferred. The orthogneiss and variably deformed granitic rocks interpreted to be Paleoproterozoic or Archean in age found throughout the Alces Lake property may be related to the 2.29 – 2.32 Ga Arrowsmith post-collisional granitic suite which occurs regionally in the western Beaverlodge Domain (Hartlaub et al., 2007; Ashton et al., 2009a). Similarly, metasedimentary gneissic rocks described by Bethune et al., 2013, and Ashton et al., 2013 that likely belong to the lower Murmac Bay Group could be the lithologies interpreted by Blake (1956) and Normand (2011; 2014) to be paragneiss. They may also belong to the upper Murmac Bay group which would make them significantly younger (Shiels, et al., 2017). Based on the deformational timings, the amphibolite rock unit in the WRCB may be related to either the D3 or D4 events. Larger, regional amphibolite found in the southern region of the property surrounding the Oldman River are likely related to the D3 event as they are primarily affected by the large scale north – northwest trending folds.

The REE bearing pegmatites are strongly affected by the D4 deformation and assuming that the zircon and monazite ages obtained in this study represent crystallization, they could indicate that the D4 deformation extends to an older age than previously thought. D5 folding and realignment of grains was not directly observed at Alces Lake, but is likely present. 1.91 - 1.90 Ga movement on the Black Bay Fault (Bergeron, 2001), may be responsible for similar movement on the St. Louis Fault. Identically trending east-northeast minor faults could have a similar age. Related 1.91 - 1.90 Ga peak regional metamorphism of the Beaverlodge Domain caused by the Taltson Orogeny is reflected in the extensively migmatized and granulite facies gneisses occurring throughout the Alces Lake area (Ashton et al., 2012). Finally, the Trans-Hudson orogeny associated lamprophyre dykes that occur across the property are likely the same as those described by Ashton et al., (2009) and yield an age of 1850 – 1800 Ma.

4.5 Detailed Description of REE Occurrences

REE showings occur throughout the Alces Lake property and are hosted primarily in biotite-allanite-monazite mineralized pods typically <5 centimeter thick and in pegmatites primarily focused around the limbs of the major synformal fold, in minor associated folds, and along the amphibolite contact. REE mineralized pegmatites also appear in shear zones recognizable due to the presence of graphite, blue quartz, and boudinaged feldspars and quartz (Fig. 4.8A). Pegmatites containing elevated REEs are also distributed across the property hosted in gneissic and granitic rocks that appear to be unrelated to any structural feature and are likely formed from anatexis. However, massive monazite mineralization occurs only within the Ivan – Dante Zone. As of 2021, more than ten REE bearing occurrences with several hundred radioactive pegmatites have been found at the Alces Lake property, with many more undocumented. However, only the WRCB area has been drilled extensively.

4.5.1 Wilson and Richard Zones



Figure 4.11: Geological map of the Wilson and Richard Zones. Each series of points represents a channel sample that was cut beginning at 1 meter of gneissic country rock, and extending across all lithologies until the channel intersected the radioactive, REE mineralized rock before being extended an additional 1 meter into the gneissic country rock. All total rare-earth oxide (TREO) analyses were conducted at the SRC Geoanalytical Laboratories in Saskatoon, Saskatchewan (see more details in Chapter 5). All data collected by Howard Zip (2021) and other geologists working for Appia.

The Richard and Wilson Zones occur at the north central part of the Alces Lake property (Fig. 4.11). Prior to their discovery, these zones were buried by trees, boulders, and overburden that were cleared between 2016 – 2021 to expose a mixture of highly deformed gneisses, relatively undeformed granites, granitoid boulders, pegmatites, and a radioactive massive biotite outcrop. The massive biotite outcrop strikes approximately 60 meters in total length at surface and is continuous at shallow depths across the Richard and Wilson Zones.

The massive biotite occurrence is on strike with the fold axis of the major synformal fold and is adjacent to the lithological contact between orthogneiss and paragneisses. They are roughly parallel to the dominant foliation (Fig. 4.12). The dips of the pegmatites are south 30° and the mineralized massive biotite are presumably similar, although this cannot be determined from surface measurements.





The massive biotite is likely structurally controlled by the fold, but no direct evidence other than its occurrence being axially planar to the major fold could be found. At surface, the massive biotite is in sharp contact with monzogranitic gneisses and granites, but has a gradational contact with biotite pegmatites that themselves become increasingly felsic towards granitic pegmatites. Pegmatites near the margins are frequently brecciated with angular, 1 - 2 cm blocks of pegmatites interbedded with gneisses. Mineralized biotite pegmatites and massive biotite are much more enriched in biotite than adjacent monzogranitic gneisses.

There is no consistent alteration affecting the mineralized biotite pegmatites at surface. Potassic alteration, pyritization, and hematitization are sporadic throughout the Wilson and Richard Zones. All styles of alteration appear to be late with pyrite or hematite mostly filling veinlets as well as patchy, solitary crystals being most common. Moderate chloritization of biotite is also common. At depth, extensive hematite – epidote – K-feldspar alteration has been found. In addition, contacts with hematite rich layers can be traced at depth with magnetic survey.

The proportion of minerals inside the biotite pegmatites reaches nearly 100 percent biotite in some sections from channel sampling. Visible monazite crystals (<1mm in diameter) occur irregularly in the biotite groundmass and occasionally form clusters, the accumulations of which do not always coincide with measured radioactivity. Xenoliths of gneisses are absent in massive biotite but occur in the biotite pegmatites where they grade into granitic pegmatites.

4.5.2 Charles – Bell Zone

The Charles – Bell Zone is located closest to the major fold hinge (Fig. 4.2; Blake, 1956) and has been significantly folded, sheared, and faulted (Figs. 4.13 and 4.14). The majority of the REE mineralizations at the Charles – Bell Zone are found in biotite pegmatites hosted in paragneiss, although highly irregular pods of massive monazite also occur. The shape of the massive monazite pods at the Charles – Bell Zone is difficult to separate from the surrounding biotite pegmatites and the pods may be smaller than they appear in the map. The biotite pegmatites are identical to the biotite pegmatites at the other zones. Interestingly, the massive monazite and massive biotite rocks at the Charles – Bell Zone have much lower REEs than their counterparts at the Ivan – Dante Zone and the Wilson - Richard Zones (Fig. 4.13).

The Charles – Bell outcrop is located approximately 200 meters to the south-southwest from the main fold hinge. Consequently, it is more difficult to obtain accurate measurements of structural features at the Charles – Bell Zone than at the other zones, with the majority of the features present exhibiting pure shear. However, accurate measurements were obtained and they are consistent with the overall stress regime operating in the WRCB area. The mineralization of the radioactive pegmatites appears to be controlled by the large shear zone cutting across the outcrop (Fig 4.14).



Figure 4.13: Geological map of the Charles – Bell Zone. All TREO analyses were conducted at the SRC Geoanalytical Laboratories in Saskatoon, Saskatchewan. All data collected by Howard Zip (2021) and other geologists working for Appia.

The massive monazite at the Charles – Bell Zone is generally fine grained and often contains abundant, mm to cm scale feldspars which become increasingly common as the massive monazite grades into diffuse contact with the biotite pegmatites. The massive biotite outcrop appears identical to its counterparts at the Wilson and Richard Zones, except that the former is more granitic and occurs perhaps in a diffuse contact with granitic pegmatites. The massive monazite rock unit in the Charles – Bell zone appears to be a very thin lens, with its depth at times being exceeded by the channels cut across its surface. The massive monazite is extremely discontinuous.



Figure 4.14: Structural map of the Charles – Bell Zone. Folds measured in the area all plunge south with an average plunge of ~56°. Brittle faults striking west – east crosscut the mineralization and offset pegmatites and massive monazite mineralization. Massive monazite mineralization also occurs along shear zones roughly parallel to the fold axis of minor folds. A single 10 meter long shear zone is roughly parallel to the main foliation which is consistent with other shear zones across the property. All data collected by Howard Zip (2021) and other geologists working for Appia.

4.5.3 Ivan – Dante Zone

The Ivan – Dante Zone is characterized by the occurrences of massive monazite pods hosting the highest grade REE mineralization (up to 22.71 wt.% TREO from channel sample) at Alces Lake. The Ivan Zone occurs at the top of a small rocky cliff alongside a hill that has been stripped of vegetation and soil. Approximately 15 meters from the main mineralization, a helipad has been constructed against the back of a small cliff. This same cliff wraps around east and continues north, gradually becoming less severe before leveling into a gently slope where the host paragneiss and amphibolite disappear.



Figure 4.15: Geological map of the Ivan – Dante Zone. All TREO analyses were conducted at the SRC Geoanalytical Laboratories in Saskatoon, Saskatchewan. All data collected by Howard Zip (2021) and other geologists working for Appia.

The host lithology at the Ivan – Dante Zone is compositionally indistinguishable from gneisses elsewhere in the WRCB area, but is more intensely deformed. Needle-like grains of hornblende and fine-grained to medium grained biotite are common and large porphyroblasts of garnet (up to several centimeters in diameter) occur locally in high abundance (80 vol.%) and

become garnetite. Extensive shearing and minor folds result in a complicated fabric that is overall consistent with the eastern limb of the major synformal fold (Fig. 4.16). Measurements obtained from several smaller folds' axes are consistent with the plunge of the major fold (Fig. 4.16). The minor fold along the southern contact of the shear zone in contact with the massive monazite mineralization is the nose of a drag fold and it is likely that the monazite mineralization itself occurs in a set of drag fold noses, suggesting a structural control on REE mineralization (Blake, 1956; Normand, 2014).



Figure 4.16: Structural map of the Ivan – Dante Zone showing multiple folds with an average plunge of \sim 40°, similar to the folds measured at the Charles – Bell Zone. All data collected by Howard Zip (2021) and other geologists working for Appia.

Massive monazite pods (up to ~3 meters wide and 4 meters long) are characterized by a distinct red color at surface (Fig. 4.17). Both the size and the percentage of monazite grains increase towards the centre of these pods from ~20-40 vol.% at the rims to ~80% at the core. The

cores of these pods are also the most radioactive reaching approximately 60,000 CPS in the largest pod, which is the highest radioactivity measured at the Alces Lake property. Compositionally these pods are dominated by monazite alongside (in order of abundance) feldspars, pyroxenes, hornblende, biotite (altering to chlorite), zircon, and occasionally garnet. Monazite grains are commonly several millimeters in diameters (Fig. 4.18A,C), while zircon is particularly abundant in the monazite-rich cores (Fig. 4.18B,D).



Figure 4.17: Photographs of massive monazite pods surrounded by chloritized biotite hosted in paragneiss at the Ivan – Dante Zone. A) the pink rim of the red massive monazite pod is likely a product of surface weathering, because it is absent in massive monazite pods intersected at depth (Fig. 4.5). B) an isolated pod located approximately 3 meters from the main mineralization.

Beside a slight increase in the amount of chlorite, there is no distinct alteration in the massive monazite pods or their host lithologies (Fig. 4.17). Contacts between the red pods of massive monazite and host lithologies are very sharp and easily distinguished by the difference in colour, but is especially pronounced in the green amphibolite. The largest pod of massive monazite is crosscut by a mylonitic shear zone of very fine to medium grained biotite (that has been extensively chloritized), which is 20 - 50 centimetres in width and is perpendicular to the

main foliation. The massive monazite mineralization continues along a steep angle into a crevasse (striking north ~10-20°). Biotite crystals in the shear zone are semi-randomly oriented. This shear zone is not radioactive and does not contain any visible monazite.

Approximately 25 meters from the main mineralization at the Ivan Zone is an outcrop of ~2 m in diameter, consisting of abundant hematite and magnetite that is highly magnetic and easily distinguishable for its red-black colour from the brown host paragneiss. While localized occurrences of hematite/magnetite are common across the Alces Lake property, they occur within the gneisses and are likely the products of hydrothermal alteration. However, their exact relationships with the other rock types are unclear, although intensely magnetic, iron rich, K-feldspar rich, and epidote rich lithologies have also been found at depth. Currently these magnetic rocks at depth are tentatively interpreted to have formed from late hydrothermal alteration.



Figure 4.18: Photomicrographs of radioactive pegmatites from the Ivan Zone. A) Biotite pegmatite showing large K-feldspar (Kfs) and plagioclase (PI) (~300 microns in diameter) with quartz. B) and D) abundant grains of zircon in massive monazite pods and locally as clusters. C) Massive monazite in the Ivan Zone. Monazite grains are relatively homogenous, but contain inclusions of K-feldspar and quartz.

4.6. Allanite in the WRCB Area

Petrographic examination of a radioactive biotite pegmatite obtained from WRC-015 (Fig. 4.9D) first confirmed the presence of allanite (Fig. 4.19). Allanite occurs both as small, 10 - 20 micron grains and as 200 - 1000 micron larger grains similar in size to the monazite grains. Allanite comprises $10 - 20 \mod .\%$ of the sample's total mineral percentage.

Subsequently, allanite has been found to occur in various lithologies, especially amphibolite, across the Alces Lake property. Indeed, allanite may be present in other radioactive occurrences where monazite was thought to be the main source of radioactivity.



Figure 4.19: Photomicrographs (plain polarized light or PPL: C, B, D, E, G; cross-polarized light or XPL: A, F) of core sample of a mineralized biotite pegmatite from 21-WRC-015 at 21.75m depth A) Extensively sericitized feldspars, pyrite, and quartz with 500 – 1000 micron clinopyroxene and monazite grains. B) Garnet grains are partly altered to sericite along fractures. C) heavily fractured clinopyroxene grain. 300 micron allanite grain is visible causing radiation damage to the feldspars. D) Heavily fractured and altered titanite crystal. E) displays the PPL view of a solitary allanite grain approximately 10-15 microns in diameter in a quartz and feldspar dominated groundmass. F) and G) close up radiation damage in quartz next to the allanite grain.

4.7 Discussion of REE Mineralization

The REE mineralization at the WRCB area is hosted primarily in gneisses that have been tentatively interpreted to be migmatized paragneiss of Paleoproterozoic age based on the description of the gneissic rock occurring in the Beaverlodge Domain (Bethune et al., 2013). Previous attempts have been made to further differentiate the gneisses by Blake (1956), and later Normand (2011; 2014) into psammitic gneiss, psammopelitic - pelitic gneiss, and garnetiferous diatexite, however, with the exception of garnetite, which occurs irregularly at surface and grades into garnet bearing gneiss, it is difficult to justify the usage of these terms considering how extensively deformed and migmatized they are. Even separating the gneiss into either orthogneiss or paragneiss is questionable given that there are no primary igneous or sedimentary features. Usage of the term paragneiss to describe the rock is largely based on the initial interpretation of the rock type by Blake (1956), who described it as "coarse grained, grey to pink weathering, garnetiferous, biotite gneiss of granodioritic composition". This description is accurate, but is lacking a sedimentary association. The classification specifically of the gneiss as being sedimentary in origin is likely owed to previous work done by the Saskatchewan Geological Survey on the outcrops surrounding Gibbs, Gale, Stephens, Hutton, MacRae, and Dodds Lakes, and the shore of Lake Athabasca. where paragneiss was previously identified (Blake, 1956).

Protolith aside, there is a distinguishable difference between two gneissic types on the basis of their magnetitic susceptibility (Table 4.1). Orthogneiss is less magnetically susceptible and compositionally is more dioritic in composition when viewing a fresh surface. Paragneiss is in general more granodioritic in composition. There is considerable compositional variation due to the high degree of migmatiziation, however, and a gneissic rock type can, at certain outcrops, shift between being melanosome and leucosome dominant across centimeter scale layers. There is a poorly defined contact between the two gneisses along the top of an approximately 20-meter slope in the Wilson outcrop. The entirety of REE mineralization in the WRCB area lies on the eastward limb of an overturned, major synformal fold.

Granitoid rocks outcrop in contact with the gneisses, massive biotite, and pegmatites at the Wilson Zone. Compositionally, granites across the Alces Lake property vary considerably in their composition. However, the granites in the Wilson Zone are primarily dioritic in composition with some tonalitic outcrops appearing as well. They are not foliated and lack the gneissosity of the surrounding orthogneiss and paragneiss making identification relatively straightforward.

The REE mineralization at WRCB occurs primarily in monazite-(Ce) (Chadirji-Martinez et al., 2022) and is found in four distinct types (Normand, 2014). The first two types are hosted by pegmatites and biotite pegmatites (Normand, 2014). Cataclasis is frequently observed in folded pegmatites or pegmatites occurring near to shear zones such as those in the Ivan Zone, although not all of the pegmatites at WRCB are brecciated. Contacts between the pegmatites and country rocks are sharp with no obvious alteration. These pegmatites can be found across the property (Fig 4.20).

The third type of REE mineralization occurs in the massive biotite with coarse grained biotite (0.1mm – 1.0mm long plates) and monazite (0.5mm – 3mm in diameter) (Fig. 4.5C). The fourth and the most interesting type of REE mineralization occurs as massive monazite pods (Fig. 4.4F and 4.7). The occurrences of allanites observed in this study, especially those as dissemination in the amphibolite and massive biotite closely associated with massive monazite pods, may represent an additional (fifth) type of REE mineralization at the Alces Lake property. However, allanites are silicate minerals containing dominantly light REEs such as Ce, La, and Nd (Pan and Fleet, 1990, 1991; Armbruster et al., 2006; Pandur et al., 2016), requiring different hydrometallurgical processes for the extraction of REEs (Zhang et al., 2016).



Figure 4.20: Radioactive occurrences in the Alces Lake property. Many hundreds of mildly radioactive (<1000CPS) pegmatites and well over a hundred moderately radioactive (1000 – 2000 CPS) pegmatites were not included during the data gathering phase of the field season due to time constraints. At most locations, individual dots represent a group of pegmatites, rather than a single pegmatite outcrop. Field data was obtained by Howard Zip (2021) and other geologists working for Appia. Satellite photo obtained from Google Hybrid.

Due to confidentiality concerns, Appia has requested that all data from the subsurface modelling of the WRCB area as well as the other occurrences at the Alces Lake property are not disclosed in this thesis. Nevertheless, a general description can be provided on the basis of my core logging and field mapping. The structural trends that are observed at surface are consistent at depth and the pegmatites appear parallel to the axial plane of the main fold. However, the pegmatites do not appear continuous at depth. Also, the 2021 drilling program revealed that REE mineralization is not continuous at depth. In particular, the massive monazite pods observed on surface are intercepted in only selected holes and at short intervals (mostly < 1 m, but occasionally 5 - 20 centimeters), suggesting that they do occur at depth but are highly irregular in distribution.

It has been suggested that the massive monazite mineralization at Alces Lake might have formed from a modified restitic magma (Normand, 2014). However, concentration by a residue *in situ* by anatexis requires an extremely high degree of partial melting (95%) and 100% melt extraction efficiency assuming a protolith contained >1 vol.% monazite (Normand, 2014). This is an unrealistically efficient system, although the partial melt generation and melt extraction efficiency can be lowered if the protolith contained a higher percentage of monazite. Modern placer deposits, such as those occurring along the coasts of Australia, India and Madagascar (Grimes, 1992; Anitha, 2020) have very high concentrations of heavy minerals (rutile, ilmenite, monazite, xenotime, garnet, zircon) that would greatly lower the amount of partial melting required to achieve the high volume % of monazite (and to a lesser degree zircon) in the massive monazite rocks that are located at Ivan – Dante. The compositional characteristics and geochronological data of monazite and zircon (see Chapter 5 below) would provide important information for the genesis of REE mineralization at Alces Lake (Normand, 2014).

Chapter 5 COMPOSITIONS AND TIMING OF REE MINERALIZATION

5.1 Whole-Rock Geochemical Compositions

Chapter 4 noted that a large suite of samples from the Alces Lake property has been analyzed for whole-rock geochemistry, with emphasis on REEs and other selected trace elements such as Th and U. In partnership with the Saskatchewan Research Council (SRC), Appia conducted extensive sampling of both surface outcrops and diamond drilling core materials. Sampling primarily targeted radioactive outcrops and pegmatites, but did include all representative lithologies across the property. All whole-rock geochemical analyses performed by SRC used the REE Ore Grade and ICP-MS2 (basement rock) packages.

The REE Ore Grade package is designed for the analysis of refractory REE ores. An aliquot of sample is fused with lithium borate flux in the Claisse TheOx[®] fusion instrument. The bead is then dissolved in dilute HNO₃ for analysis by ICP-OES. Quality control measures include the preparation and analysis of three reference materials, a replicate, and a blank. The elements analyzed are: Sc, Ga, Y, Nb, REEs (La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, and Lu), Hf, Th, and U with a detection limit of 0.002% each.

The ICP-MS2 package consists of three separate analyses. The first is an ICP-MS analysis on the partial digestion which is performed on an aliquot of the sample pulp. This aliquot is digested in a mixture of concentrated nitric and hydrochloric acids (HNO₃ and HCl) inside a test tube in a hot water bath before being diluted using deionized water. The second is an ICP-OES analysis for major and minor elements on the total digestion. Total digestions are performed on an aliquot of the sample pulp. The aliquot is digested in a Teflon tube in a hot block digestion system using a mixture of concentrated HF, HNO₃ and HClO₄ acids with the residue then being dissolved in dilute HNO₃. The third is an ICP-MS analysis for trace elements on the total digestion, which includes Al₂O₃, CaO, Fe₂O₃, K₂O, MgO, MnO, Na₂O, P₂O₅, TiO₂, Li, S, V, Cr, Sr, Zr, Ba, La, Ce. The ICP-MS analysis also includes Rb, Cs, and the Pb isotopes (²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, and ²⁰⁸Pb), in the partial and total digestion.

Although the total rare-earth oxide (TREO) contents at the WRCB area have been given in maps in Chapter 4, permission from Appia allows only a report and description of the REE compositions of representative samples from the Ivan Zone in this thesis (Table 5.1; Fig. 5.1).

Sample	24054	24065	24068	24069	24070
Sc					
(ppm)	13	37	52	57	55
Nb	14	43	49	24	36
La	3960	9600	6860	15800	12200
Се	8040	18700	13600	30100	23700
Pr	890	2050	1580	3320	2560
Nd	3330	8020	5800	12800	10200
Sm	441	1140	813	1790	1450
Eu	10.4	27.8	20	43.6	35.6
Gd	217	642	446	1030	824
Tb	33	86	61	141	109
Dy	44.1	120	95.8	201	163
Но	9	29	21	46	33
Er	20.4	52.3	43.3	84.2	69.6
Yb	2.1	5	8.6	9	9
Hf	1	86	29	103	49
Th	2350	2410	2020	4900	3740
U	41	509	164	497	815
La _N /Yb _N	1274	1297	539	1186	916
Eu/Eu*	0.10	0.10	0.10	0.10	0.10

Table 5.1. Whole-rock compositions (ppm) of REE and other selected elements in channelsamples from the Ivan Zone

Whole-rock geochemical analyses of representative samples obtained from channel samples confirm that the Ivan Zone is characterized by exceedingly high REE contents, especially the light REEs or LREEs (Table 5.1). The LREE enrichment relative to heavy REEs or HREEs is particularly evident in chondrite-normalized REE plots (Fig. 5.1), with the La_N/Yb_N values up to 1297 (Table 5.1). Another notable feature of the chondrite-normalized REE patterns of the Ivan Zone samples is the pronounced negative Eu anomalies, with calculated Eu_N/Eu_N* values of 0.1 (Table 5.1), where Eu_N* = $(Sm_NxGd_N)^{0.5}$. However, the apparent negative Dy anomalies in the chondrite-normalized REE plots (Fig. 5.1) are likely an analytical artifact.



Figure 5.1: Chondrite-normalized REE plots illustrating whole-rock REE compositions of representative mineralized samples from the Ivan Zone chosen for LA-ICP-MS analysis. Chondritic values from Tylor and McLennan (1985). Tm and Lu were not analyzed for these samples.

The mineralized samples of the Ivan Zone are also characterized by elevated radioactive elements (up to 4900 Th and 815 U ppm; Table 5.1), consistent with the original discovery of this occurrence from a radioactivity survey. Other notable constituents in the Ivan Zone are Sc (up to 57 ppm), Nb (49 ppm), and Hf (103 ppm; Table 5.1).

5.2 Compositions of Monazite and Zircon

Monazite and zircon grains in four representative samples (024636, 805-727, 805-744, 805-735; Tables 5.2 and 5.3) obtained from channels cut through massive monazite in the Ivan Zone with the highest mod. % of monazite and zircon were chosen for quantitative wavelength dispersive spectrometry (WDS) analysis after BSE and CL imaging and qualitative energy dispersive spectrometry (EDS) analyses. Chadirji-Martinez et al. (2022) reported EMPA data of monazite from a massive monazite sample from the Wilson Zone, which are included in Table 5.2 for comparison.

All BSE and CL imaging as well as qualitative EDS and quantitative WDS analyses, including those reported by Chadirji-Martinez et al. (2022), were made on the JEOL JXA-8600 Superprobe equipped with one EDS spectrometer and three WDS spectrometers in the Department of Geological Sciences at the University of Saskatchewan. Operating conditions included: accelerating voltage of 15 kV, beam current of 50 nA, beam diameter of 3 μ m, counting times for Si, Ca, REE, P, and F of 40 s each, and 120 s each for U, Th, Y, and Pb. The following standards were used: quartz (Si K α), Y-Al garnet (Y L α), crocoite (Pb M β), diopside (Ca K α), synthetic REE phosphates (La L α , Ce L α , Pr L β , Nd L α , Sm L β , Gd L β , and Dy L α), U metal (U M β), Th metal (Th M α), fluorapatite (P K α), and fluorite (F K α).



Figure 5.2: CL (A, C, and F) and BSE (B, D and E) images of monazite and zircon grains in massive monazite samples from the Ivan Zone. A and B are typical monazite grains from the Ivan Zone. C and D are small zircon grains as interstitial infillings around large (1mm long, 500-600 microns wide), subhedral monazite. E and F with red circles indicate spots selected for quantitative EMPA analyses. Abbreviations Mnz, Zrn, Kfs, Qtz, PI, and Thr are monazite, zircon, K-feldspar, quartz, plagioclase, and thorite.
Combined BSE and CL imaging showed that all monazite grains in the samples of the Ivan Zone are generally homogeneous, without any obvious compositional zonation (Fig. 5.2). However, mineral inclusions of quartz, plagioclase, and microcline (50-100 micron in diameter) are occasionally observed in monazite grains of typically 2-3 mm in diameter. Other notable features of the monazite grains in the Ivan Zone are the local presence of very small grains of Th and U silicates such as thorite (ThSiO₄) along fractures or margins, which are indicative of preferential leaching of these radionuclides during late hydrothermal alteration (Pan et al., 1993; Ning et al., 2019).

Spot	Туре	ThO₂	UO2	Y ₂ O ₃	La ₂ O ₃	Ce ₂ O ₃	Pr ₂ O ₃	Nd ₂ O ₃	Sm ₂ O ₃	Gd ₂ O ₃	Dy ₂ O ₃	CaO	P2O5	Total:
					Iva	n Zone	Sample	805-73	5					=
1	iso	7.97	0.40	0.64	14.3	29.2	3.09	10.4	1.36	0.62	0.04	1.52	30.5	100.1
2	iso	8.75	0.40	0.19	14.5	29.2	3.10	10.7	1.46	0.70	0.00	1.36	29.8	100.1
3	iso	7.49	0.39	0.34	13.8	29.9	3.15	11.0	1.52	0.65	0.12	1.48	28.5	98.4
4	core	8.39	0.42	0.52	13.7	30.6	3.12	11.0	1.63	0.57	0.00	1.77	28.5	100.2
5	rim	8.39	0.42	0.33	14.8	30.5	3.22	10.4	1.43	0.76	0.01	1.50	27.3	99.1
6	core	7.88	0.42	0.28	14.2	30.6	2.97	10.6	1.70	0.69	0.00	1.57	29.0	99.9
7	rim	7.77	0.40	0.45	13.8	30.8	3.50	10.9	1.11	0.75	0.00	1.60	29.4	100.5
8	core	8.69	0.50	0.42	13.6	30.1	3.55	10.8	1.35	0.75	0.07	1.91	27.7	99.4
9	rim	8.23	0.54	0.77	14.0	29.7	3.22	10.8	1.08	1.04	0.13	1.84	28.3	99.7
10	core	9.56	0.54	0.46	13.4	28.6	3.08	10.9	1.41	0.79	0.00	2.12	29.1	100.0
11	rim	7.86	0.41	0.46	14.7	30.8	3.33	10.9	1.16	0.69	0.00	1.26	28.9	100.4
12	core	7.85	0.45	0.24	14.3	30.3	3.30	11.0	1.05	0.81	0.00	1.83	29.5	100.6
13	rim	8.45	0.52	0.24	13.7	30.0	3.31	11.1	1.38	0.89	0.00	1.75	29.7	101.0
Ivan Zone Sample 024636														_
14	core	8.32	0.45	0.33	14.0	29.8	3.36	10.7	1.22	0.80	0.00	1.76	29.4	100.2
15	rim	7.53	0.38	0.34	15.2	31.6	3.17	10.8	0.77	0.62	0.06	1.25	29.0	100.7
16	core	8.51	0.46	0.40	13.6	30.4	3.46	11.1	1.30	0.62	0.01	1.94	28.8	100.7
17	rim	8.42	0.40	0.13	14.2	30.6	3.10	10.9	1.23	0.81	0.02	1.19	28.0	98.8
18	core	7.50	0.40	0.20	14.8	31.4	3.43	10.4	1.37	0.70	0.00	1.60	29.2	101.0
19	rim	7.76	0.39	0.32	14.5	31.1	3.29	11.3	1.19	0.53	0.05	1.40	27.8	99.6
20	core	7.22	0.46	0.27	14.7	30.8	3.26	10.7	1.53	0.78	0.00	1.78	28.8	100.4
21	rim	8.01	0.38	0.42	14.4	31.2	3.26	10.9	1.26	0.73	0.00	1.47	28.8	100.9
22	core	7.19	0.38	0.36	14.8	31.6	3.21	10.7	1.46	0.67	0.11	1.72	29.2	101.3
23	rim	7.77	0.42	0.36	14.1	30.5	3.40	11.5	1.39	0.77	0.14	1.54	28.8	100.7
						Thin Se	ction 80)5-733						_
24	core	7.77	0.45	0.28	15.1	30.5	3.07	9.87	0.77	0.68	0.04	1.78	29.0	99.4
25	rim	6.25	0.39	0.09	14.7	30.4	3.18	10.8	0.85	0.79	0.12	1.82	30.0	99.4
26	core	7.73	0.42	0.05	14.6	29.7	3.40	10.7	1.06	0.35	0.18	1.98	29.6	99.7
27	rim	7.89	0.45	0.40	14.6	30.4	2.92	10.2	1.11	0.72	0.00	2.11	29.4	100.2

Table 5.2 Compositions (wt.%) of selected monazite grains from EMPA analyses

28	core	7.57	0.40	0.38	14.9	30.7	3.26	10.9	1.21	0.74	0.05	1.42	29.0	100.6	
29	rim	8.05	0.40	0.28	15.0	30.6	2.93	10.5	0.99	0.61	0.00	1.65	29.2	100.2	
30	core	7.96	0.40	0.32	15.3	30.7	3.17	10.6	1.12	0.38	0.00	1.36	28.6	99.9	
31	rim	8.05	0.43	0.33	15.0	30.7	3.39	10.6	1.13	0.64	0.00	1.58	28.8	100.7	
32	core	8.26	0.36	0.36	15.2	30.3	3.20	10.2	0.91	0.58	0.00	1.53	28.4	99.2	
33	rim	6.96	0.29	0.26	13.0	26.3	2.81	9.15	1.09	0.58	0.27	1.92	26.5	89.1	
	Ivan Zone Sample 805-727														
34	34 iso 7.75 0.46 0.38 14.6 29.2 3.29 10.4 0.99 0.63 0.00 2.04 28.5														
35	iso	5.43	0.94	0.13	14.3	30.2	3.37	11.2	1.67	0.86	0.11	1.49	29.6	99.3	
36	iso	7.71	0.99	0.06	13.7	29.0	3.28	10.9	0.85	0.17	0.00	1.92	29.1	97.6	
37	iso	8.05	0.46	1.20	13.9	29.1	3.05	11.0	1.65	1.33	0.34	1.66	28.6	100.3	
38	iso	6.87	0.42	2.29	13.2	28.8	3.18	10.8	1.69	1.83	0.37	1.52	30.3	101.3	
	Wilson Zone Sample (Chadirji-Martinez et al. 2022)														
1	-	9.49	0.47	0.24	14.9	30.7	3.39	9.28	1.96	0.36	0.00	0.93	28.2	100.0	
2	-	9.76	0.54	0.34	14.3	30.1	2.97	8.76	2.21	0.57	0.00	1.34	29.0	99.8	
3	-	8.65	0.54	0.45	14.4	30.2	2.86	8.48	1.69	0.39	0.00	1.30	30.3	99.3	
4	-	9.30	0.39	0.42	14.2	29.9	2.86	8.59	1.86	0.15	0.00	1.30	30.2	99.1	
5	-	10.5	0.47	0.25	14.3	29.5	3.42	8.47	1.55	0.06	0.00	1.00	29.8	99.3	
6	-	9.51	0.46	0.35	14.3	29.7	3.17	7.89	1.97	0.32	0.00	1.44	30.4	99.6	
7	-	9.66	0.49	0.43	14.2	27.5	3.11	7.26	2.18	0.20	0.00	1.30	31.1	97.4	
8	-	8.17	0.48	0.39	15.0	29.0	2.70	7.03	2.13	0.00	0.00	1.18	32.1	98.2	
						SPI	Monazi	te							
-	-	18.1	34.8	3.40	18.1	34.8	3.40	9.68	0.20	0.56	0.00	0.56	28.99	99.8	

Abbreviation "iso" refers to a single analyzed spot in one individual monazite grain, whereas "core" and "rim" refer to the analytical spots at the center and the margin of a monazite grain, respectively. Note that 8 analyses of monazite from the Wilson Zone (Chadirji-Martinez et al., 2022) are included for comparison. Also included for comparison are the data of the SPI monazite that was used as a standard.

Quantitative WDS analyses confirm BSE and CL imaging that all monazite grains in the Ivan Zone and the Wilson Zone are fairly compositionally homogenous within individual grains and among different grains in a given sample. Monazite in both the Ivan and Wilson Zones are invariably of the Ce dominant species (i.e. monazite-(Ce), with LREE-enriched chondrite-normalized REE patterns (Figs. 5.3 and 5.4). The most notable impurity constituent of monazite is ThO₂, which varies from 5.43 to 10.54 wt% (Table 5.2). The high ThO₂ contents of monazite are invariably accompanied by elevated CaO contents, indicative of a coupled substitution Th⁴⁺ + Ca²⁺ = 2REE³⁺ for the incorporation of Th in monazite (Uher et al., 2015). Other constituents such as U and Y are generally low in monazite of the Ivan and Wilson Zones (Table 5.2). Two notable differences between monazite grains from the Ivan Zone and the Wilson Zone are that the former have higher Th contents but lower REE (notably Nd, Gd, and Dy) than the latter (Table 5.2).



Figure 5.3: Chondrite normalized REE plots showing the REE compositions of monazite-(Ce) in samples 805-727 and 805-735 from the Ivan Zone (EMPA data in Table 5.2). Chondritic normalizing values are from Taylor and McLennan (1985).

Zircon grains in the massive monazite pods of the Ivan Zone are locally abundant (up to 10 mod.%) but are generally smaller (50-200 microns in diameter) (Figs. 5.2 and 5.5) than the dominant monazite grains. The zircon crystals are usually euhedral to subhedral prisms and are often locally clustered together, as interstitial fillings to monazite (Figs. 5.2 and 5.5). BSE and CL images show that the zircon crystals often show complex oscillatory or patchy zonation and, in contrast to the neighboring monazite grains, are generally free of mineral inclusions (Fig. 5.5).



Figure 5.4: BSE (A and B) and CL (C, D, E, and F) images showing internal structures of representative zircon grains from the Ivan Zone. Red dots represent sample spots that were used for analysis (see Table 5.3).

Quantitative WDS analyses showed that all zircon grains from the Ivan Zone, despite their complex zonation patterns revealed by BSE and CL imaging (Fig. 5.4), are compositionally homogeneous. The zircon grains are invariably close to the ideal stoichiometry ZrSiO₄ with most impurity constituents, such as ThO₂, UO₂, Y₂O₃, PbO and P₂O₅, being below 0.2 wt.% (Table 5.3). However, a few elevated Y₂O₃ contents (up to 0.5 wt.% Table 5.3) are obtained in zircon grains; and are consistent with the commonly HREE-enriched composition of this mineral (Yurimoto et al., 1990; Pan and Breaks, 1997; Breiter, 2016; van Gosen et al., 2017). Indeed, the unusually high abundance of zircon is likely responsible for the elevated HREE concentrations in the massive monazite pods of the Ivan Zone, although all REE (including HREE) contents of zircon from the Ivan Zone are below the detection limits of EMPA analyses. Note that several elevated CaO contents from the rims of zircon grains (Table 5.3) are obviously related to analytical problems (including exceedingly low totals and expected interference from adjacent plagioclase grains).

Spot	Туре	SiO ₂	ZrO ₂	HfO ₂	ThO₂	UO₂	Y ₂ O ₃	P ₂ O ₅	CaO	PbO	Total	Zr/Hf
					Sa	mple 8	05-735					
1	core	33.5	65.0	1.32	0.00	0.17	0.10	0.00	0.47	0.00	100.54	43
2	mid	32.7	65.6	1.33	0.02	0.04	0.00	0.00	0.00	0.00	99.74	43
3	rim	32.7	65.7	1.27	0.01	0.11	0.00	0.00	0.02	0.00	99.75	45
4	core	33.6	65.7	1.33	0.00	0.11	0.00	0.00	0.00	0.00	100.72	43
5	mid	33.6	65.5	1.30	0.00	0.15	0.00	0.00	0.03	0.00	100.56	44
6	rim	32.3	63.5	1.19	0.07	0.11	0.00	0.00	0.26	0.00	97.40	46
7	core	32.6	64.6	1.26	0.02	0.13	0.04	0.00	0.02	0.00	98.67	45
8	mid	33.9	65.6	1.31	0.00	0.08	0.01	0.00	0.04	0.00	100.91	44
9	rim	33.7	65.9	1.23	0.00	0.09	0.00	0.00	0.00	0.00	100.95	47
10	core	33.5	66.0	1.38	0.00	0.02	0.00	0.00	0.02	0.00	100.90	42
11	mid	33.1	65.7	1.24	0.00	0.13	0.07	0.00	0.01	0.00	100.25	46
12	rim	33.6	65.7	1.39	0.03	0.09	0.00	0.00	0.00	0.00	100.82	41
13	core	32.1	65.9	1.03	0.03	0.03	0.11	0.00	0.00	0.00	99.15	56
14	mid	32.4	65.6	1.27	0.01	0.07	0.00	0.00	0.00	0.00	99.29	45
15	rim	33.5	65.9	1.39	0.00	0.06	0.00	0.00	0.06	0.00	100.93	41
					Sa	mple 0	24636					
16	core	33.6	64.8	1.32	0.06	0.19	0.17	0.00	0.03	0.00	100.16	43
17	rim	30.7	63.2	1.16	0.04	0.15	0.00	0.00	0.59	0.00	95.91	48
18	core	33.2	65.9	1.32	0.00	0.04	0.00	0.00	0.00	0.00	100.52	43
19	mid	32.1	64.1	1.21	0.08	0.12	0.05	0.00	0.33	0.00	98.06	46
20	rim	32.7	65.8	1.30	0.00	0.08	0.00	0.00	0.02	0.00	99.89	44
21	core	33.3	66.0	1.35	0.00	0.07	0.00	0.00	0.00	0.00	100.72	43
22	rim	32.9	66.0	1.23	0.02	0.15	0.09	0.00	0.00	0.00	100.33	47
23	core	33.1	66.2	1.35	0.05	0.11	0.03	0.00	0.00	0.00	100.86	43
24	rim	27.5	61.0	1.48	0.14	0.04	0.46	0.00	0.58	0.00	91.20	35

Table 5.3 Compositions (wt.%)) of zircon in the Ivan	Zone from EMPA analyses
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25	core	32.9	65.4	1.11	0.00	0.13	0.00	0.00	0.00	0.00	99.57	51			
26	rim	32.0	65.7	1.38	0.00	0.08	0.00	0.00	0.00	0.00	99.09	41			
27	core	33.2	65.5	1.31	0.00	0.12	0.00	0.00	0.02	0.00	100.12	44			
28	rim	24.0	54.3	1.41	0.01	0.01	0.11	0.00	1.49	0.00	81.30	34			
	Sample 805-733														
29	29 core 33.5 65.9 1.46 0.00 0.14 0.00 0.02 0.00 100.92 40														
30	rim	32.7	66.3	1.21	0.02	0.11	0.00	0.00	0.00	0.00	100.27	48			
31	core	33.7	65.7	1.25	0.04	0.09	0.00	0.00	0.05	0.00	100.88	46			
32	rim	32.4	65.5	1.42	0.00	0.06	0.00	0.00	0.00	0.00	99.38	40			
33	core	32.4	64.1	1.20	0.02	0.14	0.07	0.00	0.60	0.00	98.50	47			
34	rim	32.6	66.3	1.42	0.00	0.13	0.03	0.00	0.00	0.00	100.52	41			
35	core	32.6	65.6	1.34	0.01	0.08	0.00	0.00	0.04	0.00	99.61	43			
36	rim	33.7	65.8	1.21	0.00	0.07	0.00	0.00	0.00	0.00	100.76	47			
37	core	33.4	65.9	1.27	0.02	0.12	0.00	0.00	0.03	0.00	100.69	45			
38	rim	32.3	64.8	1.38	0.04	0.10	0.09	0.00	0.18	0.00	98.92	41			
					Sa	mple 8	05-727								
39	iso	32.3	64.2	1.33	0.04	0.23	0.05	0.00	0.29	0.00	98.39	42			
40	iso	29.9	59.0	1.21	0.01	0.14	0.51	0.00	0.33	0.00	91.10	42			
41	iso	32.9	65.1	1.30	0.03	0.19	0.00	0.00	0.07	0.00	99.53	44			
42	iso	28.0	60.3	1.29	0.00	0.21	0.51	0.00	0.53	0.00	90.81	41			
Abbrevi	ation "iso	" refers to	a single	analyzed	l spot in d	one indivi	dual mon	azite grain	. whereas	s "core" a	nd "rim" ref	er to the			

analytical spots at the center and the margin of a monazite grain, respectively.

Interestingly, the HfO₂ contents of all zircon analyses from the Ivan Zone fall within a narrow range from 1.03 to 1.48 wt.%, yielding the Zr/Hf values from 39 to 56 (Table 5.3). The Zr/Hf value in zircon in granites and other igneous rocks is often a sensitive indicator for the degree of fractional crystallization. For example, the chondritic Zr/Hf values are ~35-40 (Claiborne et al., 2006), while zircon in highly evolved granites and granitic pegmatites is commonly characterized by low Zr/Hf values (1-35) (Zozulya et. al., 2022). Therefore, the high Zr/Hf values in zircon from the Ivan Zone suggest that the REE mineralized pegmatites were not the product of a highly evolved system.

5.3 Monazite and Zircon U-Pb Geochronology

One of the key pieces of evidence necessary to determine the origin of the REE mineralization at the Alces Lake property is the precise age(s) of the dominant ore mineral monazite-(Ce) (Meldrum et al., 1998; Lenz et al., 2019; Schulz, 2021). In fact, had the massive monazite pods in the Ivan-Dante Zone formed from a placer deposit or a restitic melt of a preexisting placer, then the "detrital" monazite grains would have been expected to record pre-Taltson and variable ages, reflecting their accumulation in a sedimentary system from potentially diverse source rocks of different ages. On the other hand, our field-based hypothesis that the monazite mineralization in the Ivan-Dante Zone and other parts of the WRCB area was related to magmatic-hydrothermal processes as part of the high-grade regional metamorphism and associated anatexis would have more uniform ages of the Taltson Orogeny (Aston et al., 2009a, b; Bethune et al., 2010).

To test these proposed models for the REE mineralization at the WRCB area, monazite and zircon grains in the two mineralized samples from the Ivan Zone, following detailed BSE and CL imaging as well as qualitative EDS and quantitative WDS analyses above, have been selected for in-situ U-Pb isotope geochronological analyses using LA-ICP-MS. The inclusion of zircon grains for in-situ U-Pb geochronological analyses is based on the following two reasons. First of all, zircon grains in these two samples from the Ivan Zone are highly abundant and, therefore, allow a large number of analyses with statistical significance. Second, zircon is known to have a higher blocking temperature than monazite (Hancher and Miller, 1993; Connelly, 2001; Ning et al., 2019) and have a different annealing behavior from monazite (Hancher and Miller, 1993; Meldrum et al., 1998). The second reason is particularly relevant to REE mineralization at the Alces Lake property, which is situated in host rocks that have been subjected to granulite-facies regional metamorphism as well as extensive deformation and migmatization. Therefore, comparison of U-Pb isotopic ages from both zircon and monazite grains (Parrish, 1990) would be an excellent way of evaluating our field-based hypothesis versus the previously proposed placer origins (Normand, 2014).

All U-Pb geochronological analyses of monazite and zircon were made on a Nu–Plasma MC-1259 ICP-MS (Nu Instruments, UK) coupled to a frequency quintupled (= 213 nm) Nd:YAG laser 1260 ablation system (New Wave Research, USA) at the Radiogenic Isotope Facility (RIF), University of Alberta. The analytical methods used were modified from those of Simonetti et al. (2005). The collector configuration utilized a collector block design that includes three ion counters and twelve Faraday buckets. This allows for the simultaneous acquisition of ion signals ranging from mass U²³⁸ to Tl²⁰³, an important factor in obtaining highly precise and accurate ages (Simonetti et al., 2005). Isotopes Pb²⁰⁷, Pb²⁰⁶, Pb²⁰⁴ alongside Hg²⁰⁴ were measured on the ion counting channels with a 30 second blank between ablation to determ, ine the Hg²⁰⁴ contribution followed by another 30 second blank. All ablation analyses were Ti mass corrected to adjust for instrument bias using the ²⁰⁵Ti/²⁰³Ti ratio. The primary zircon reference used was LH9415 (Ashton et al., 1999) and the secondary reference was OG1 (Stern et al., 2009). The monazite reference from Western Australia was developed in house at the RIF (Simonetti et al., 2005). The zircon standards were rerun every 12 sample points and the monazite standards were rerun every 10 sample points. Analysis of zircon data occurred over the course of one entire day with all samples analyzed. All monazite samples were analyzed in one day and remaining time on the instrument was used to add more sample points at random to randomly selected monazite grains in order to obtain maximum value out of the time available with the instrument. All uncertainty values for all isotopic data are reported to a factor of 2 (2 σ), where standard deviation was σ = \pm (2/f)1/2and the Mean Squared Weighted Deviation (MSWD) had an expected value of 1 for a true isochron (Wendt and Carl, 1990).

Due to the large sizes of all monazite grains and some zircon grains in the two samples from the Ivan Zone (Figs. 5.2 and 5.4), it was possible to perform LA-ICP-MS analyses of multiple spots in individual grains. In this study, 20 monazite grains and 20 zircon grains (10 grains for each mineral per sample) were selected for LA-ICP-MS analyses, yielding a total of 41 and 118 spot analyses for zircon and monazite, respectively. All U-Pb geochronogical data are summarized in **Appendix B** and illustrated in concordia diagrams (Figs. 5.5).



Figure 5.5: Concordia diagram of zircon and age distribution diagram of monazite showing U-Pb data from samples taken from the Ivan Zone.

Despite the considerable variation in the sizes of zircon grains and their complex internal structures as revealed by BSE and CL imaging (Fig. 5.4), all 41 spot analyses obtained from 20 zircon grains from the two samples of the Ivan Zone are approximately concordant, yielding a uniform U-Pb age of 1926±3 Ma, with a MSWD = 0.016 (Fig. 5.5). This uniform zircon U-Pb age of 1926±3 Ma supports our field-based hypothesis that REE mineralization at the Ivan Zone (and by extension similar occurrences in the WRCB area at the Alces Lake property) formed by magmatic-hydrothermal processes related to the granulite-facies metamorphism and anatexis during the Taltson Orogeny. This result also casts doubt on the previously proposed placer or restitic placer models (Normand, 2014).

Of the total 118 spot analyses obtained from 20 monazite grains, two spots with anomalous ablation patterns and seven spots of unexpectedly large errors are excluded. Figure 5.5 shows the data of the remaining 111 spot analyses are not concordant but form a well-defined linear array, yielding an upper intercept age of 1926 ± 15 Ma with a MSWD value of 0.036.

This monazite age agrees with that from associated zircon grains within analytical uncertainty, further supporting the hypothesis that REE mineralization at the Ivan Zone formed by magmatic-hydrothermal processes related to the granulite-facies metamorphism and anatexis during the Taltson Orogeny.

Chapter 6 CONCLUSIONS AND SUGGESTIONS FOR FUTURE RESEARCH

Field mapping revealed that the REE mineralization in the WRCB and Ivan-Dante Zones at Alces Lake is hosted in monazite and biotite rich pegmatites that appeared to exploit structural conduits such as faults and shear zones before being deposited in the low-pressure environment of fold noses. Bounding these monazite and biotite rich pegmatites are highly deformed amphibolites, which may be acting as a trap for the mineralizing melt/fluid. To the south west of the WRCB area there are large leucogranite bodies, which are mildly radioactive (200 – 2000 CPS) and contain trace amounts of REEs. These leucogranites may have been scavenged for REEs during their formation related to granulite-facies metamorphism and anatexis and are tentatively interpreted to represent the magmatic sources for the formation of the REE-mineralized pegmatites. This interpretation is supported by available geochemical data (Htaniuk, 2021). However, proof of this interpretation requires detailed U-Pb geochronological analyses of the leucogranite intrusions at the Alces Lake property. Also, more detailed geochemical studies, especially Sm-Nd, Rb-Sr, Lu-Hf and Pb-Pb isotope analyses as geochemical tracers, of the leucrogranites and associated pegmatites at Alces Lake would provide additional constraints on their linkages to the high-grade regional metamorphism and their evolution to the formation of **REE** mineralization.

Petrographic and compositional characteristics of monazite and zircon at Alces Lake also support their crystallization in a pegmatitic environment. Moreover, the uniform U-Pb ages of 1926±3 Ma from zircon and 1926±16 Ma monazite at the Ivan Zone demonstrate that the formation of these pegmatites occurred during the late-stage tectonic processes of the Taltson Orogeny. Given the extensive and extreme metamorphic regimes that have occurred in the area, as evidenced by the gneissic rocks that have been intensely deformed and migmatized, anatexis seems a likely process for the generation of the leucogranites, which might have evolved to form REE-enriched melts or fluids to produce monazite mineralization.

In neither field observations nor petrographic examinations was there any association between monazite and Fe-oxides, which is typical of hydrothermal processes (Mahdy, 2021). However, in drill core at the Alces Lake property there are zones of monazite crystals that appear to have been heavily altered by, or have crystallized from, hydrothermal processes. These zones may not be directly related to the zircon and monazite rich pegmatites in the WRCB area, yet the presence of allanite in the WRCB requires further investigation. Also, the extent that late hydrothermal activities affected the monazite-rich pegmatites is not clear, although samples collected at surface outcrops from the Ivan Zone indicate that the monazite and zircon grains are likely magmatic-hydrothermal in origin.

The monazite rich pegmatites in the WRCB and Ivan-Dante Zones are only a few of the many REE occurrences that have been discovered across the Alces Lake property. The relationships between the formation of these intensely mineralized monazite bearing pegmatites and the regional structures are evident from field observations but require further study. In particular, the occurrences of monazite-rich pegmatites are clearly structurally controlled, with faults and fold limbs acting as their pathways for emplacement. However, the deformation that

was most beneficial to the emplacement of the REE mineralized pegmatites would be of crucial importance for any future prospecting ventures on the property. Beside the regional synformal folding, the association between the pegmatites and shear zones is especially interesting because folds intersected by shear zones are a common feature of monazite or biotite rich pegmatites.

The highly deformed amphibolites that appear to act as a trap for the monazite rich pegmatites are also of interest for additional research, especially considering the occurrences of disseminated REE mineralization in them and their possible use as a vector for future exploration in this region. Though only observed at the WRCB mineralization, the amphibolite has been a reliable indicator for the presence of increased REE contents – both at surface and at depth. An investigation of the REE bearing gabbroic dykes that occur at depth in the WRCB area would also be beneficial, although they do not have high REE contents as the pegmatites. These dykes of low REE grades are much larger and are more continuous at depth, and may become large enough at depth to be of economic interest.

One possible model for the formation of the unique massive monazite mineralization at Alces Lake was via the generation of an immiscible REE-phosphate melt. This model is analogous to the separation of immiscible carbonatitic melts from the parental silicate melts, which is commonly invoked for the formation of carbonatite hosted REE deposits (Anenburg et al., 2021). Also, the formation of immiscible melts has long been proposed as an essential component in pegmatites and related deposit types (Thomas and Davidson, 2016). Moreover, the evolution of the immiscible carbonatitic melts has been suggested to be responsible for the enrichment in LREE and the crystallization of REE minerals at lower temperatures in carbonatite hosted REE deposits (Thompson et al., 2007; Anenburg et al., 2021). Interestingly, the REE mineralization hosted by diopside-allanite veins and apatite breccia veins at Hoidas Lake, Saskatchewan, has also been proposed to form from the immiscibility enabled REE enrichment and mineralization (Pandur, 2015; Pandur et al., 2016). The presence of volatiles in the melt-fluid could readily allow for the generation of a low-temperature mineralization fluid as is evidenced at Hoidas Lake (Pandur, 2015; Pandur et al., 2016). However, zircon grains at Alces Lake had primitive Zr/Hf ratios suggesting that the mineralization melt-fluid was not extensively evolved. It is possible that the REE-phosphate melt-fluid at Alces Lake was proximal to the parental silicate melt and underwent only limited fractionation. Further geochemical studies of the neighboring leucogranites, the suspected parental silicate melts, could help to better understand the origin of the massive monazite pegmatites at Alces Lake.

Finally, it has been frequently noted by Appia that the Alces Lake REE occurrences are hosted primarily in monazite. This may be true for surface samples at WRCB, but there needs to be more investigation to confirm this assertion. Over the course of the 2021 field season, many biotite-rich radioactive pegmatites and veinlets were located and analyzed for their TREO contents. They often contain high REE contents, but many of these occurrences contain allanite-(Ce) as a major host of REEs. Allanite-(Ce) requires very different hydrometallurgical processes for REE extraction (Zhang et al., 2016). Therefore, further research to better document the distribution and compositions of allanite-(Ce) at Alces Lake is warranted in its possible further development as a REE resource.

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Sample Number	Map Location Number	Sample Type	UTM Co-ordinate NAD83 13	Radioactivity(cps)
024054	1	Boulder	329119E – 6618386N	9,538cps
024056	2	Boulder	329098E – 6618409N	13,624cps
024065	3 South of	Outcrop	329254E – 6618296N	53,500cps
	Trench 2			
024066	4	Outcrop	329274E – 6618126N	30,900cps
024067	5	Outcrop	329241E – 6618152N	46,600cps
024088	6 REE Trench 1	Boulder	329232E – 6618126N	51,700cps
	(North)			
024069	7	Outcrop	329204E – 6618212N	52,000cps
024070	8 REE Trench 2	Boulder	329239E – 6618109N	49,840cps
	(South)			
155247	9	Outcrop	329074E – 6618335N	5,500cps
155248	10 Beside Heli	Outcrop	329146E – 6618357N	8,753cps
	Pad			
446823	11	Outcrop	329192E – 6618230N	14,671cps
446824	12	Outcrop	329217E – 6618197N	23,100cps
446825	13	Outcrop	329245E – 6618192N	10,295cps
446826	14	Boulder	329267E – 6618247N	43,600cps
446827	15	Outcrop	329126E – 6618302N	22,300cps
Helicopter Pad		Outcrop	329160E – 6618370N	Below detection

APPENDIX A: DESCRIPTION OF ALCES LAKE MINERALIZED SAMPLES IN THIS STUDY

APPENDIX B: LA-ICP-MS U/PB GEOCHRONOLGY DATA FROM ZIRCON AND MONAZITE GRAINS

										²⁰⁷ Pb*/ ²⁰⁶ Pb*	2 s	²⁰⁷ Pb*/ ²³⁵ U	2 s	²⁰⁶ Pb*/ ²³⁸ U	2 s	%
Identifier	²⁰⁶ Pb (cps) ²	²⁰⁴ Pb (cps)	²⁰⁷ Pb/ ²⁰⁶ Pt	28	²⁰⁷ Pb/ ²³⁵ U	28	²⁰⁶ Pb/ ²³⁸ U	28	Rho	age (Ma)	error (Ma)	age (Ma)	error (Ma)	age (Ma)	error (Ma)	discord.
Standard											. ,	,	. ,	,		
OG1-1	2492273	93	0.29988	0.00159	29.52014	0.85171	0.71395	0.02025	0.983	3470	8	3471	28	3473	76	-0.1
OG1-2	742949	23	0.29785	0.00159	28.76146	0.56004	0.70034	0.01311	0.962	3459	8	3445	19	3422	50	1.4
OG1-1	1871657	127	0.29902	0.00157	29.61844	1.30882	0.71840	0.03152	0.993	3465	8	3474	43	3490	117	-0.9
OG1-2	1782655	39	0.29802	0.00157	28.65441	1.16721	0.69733	0.02817	0.992	3460	8	3442	39	3411	106	1.8
0G1-1 0G1-2	222365	15	0.29859	0.00135	28.98150	1.02541	0.70396	0.02470	0.992	3465	7	3453	34 46	3456	93	0.3
0G1-2 0G1-1	337497	37	0.29948	0.00311	28.41182	1.47355	0.68806	0.03496	0.980	3468	16	3433	50	3375	132	3.4
OG1-2	311402	30	0.29889	0.00309	29.60027	1.33102	0.71826	0.03143	0.973	3464	16	3474	43	3490	117	-0.9
OG1-3	175594	8	0.29840	0.00310	29.05877	1.16517	0.70628	0.02736	0.966	3462	16	3456	39	3444	103	0.7
Samples																
727Z1-1	939636	102	0.11814	0.00053	5.75043	0.20598	0.35301	0.01255	0.992	1928	8	1939	31	1949	60	-1.2
727Z1-2	912166	54	0.11829	0.00051	5.73897	0.25488	0.35186	0.01555	0.995	1931	8	1937	38	1943	74	-0.8
727Z2-2	802805	29	0.11827	0.00056	5.60047	0.24346	0.34343	0.01484	0.994	1930	8	1916	37	1903	71	1.6
72773 2	1000709	41	0.11855	0.00052	5.91514	0.23009	0.30100	0.01399	0.994	1935	0	1903	33	1991	65	-3.4
72774-1	877286	17	0.11818	0.00055	5 58075	0.22512	0.34248	0.01568	0.995	1933	8	1942	39	1899	75	-0.9
727Z4-2	865612	16	0.11827	0.00051	5.62494	0.22461	0.34495	0.01369	0.994	1930	8	1920	34	1910	65	1.2
727Z5-1	693066	19	0.11858	0.00054	5.77772	0.28378	0.35338	0.01728	0.996	1935	8	1943	42	1951	82	-0.9
727Z5-2	579181	11	0.11829	0.00050	5.64104	0.28409	0.34588	0.01736	0.996	1931	8	1922	43	1915	83	0.9
727Z6-1	491213	6	0.11853	0.00059	5.78182	0.28609	0.35377	0.01742	0.995	1934	9	1944	42	1953	82	-1.1
727Z6-2	454292	8	0.11850	0.00054	5.91569	0.31389	0.36208	0.01914	0.996	1934	8	1964	45	1992	90	-3.5
727Z7-1	1080122	16	0.11797	0.00051	5.62482	0.21745	0.34580	0.01328	0.994	1926	8	1920	33	1915	63	0.7
727Z8-1	986177	10	0.11759	0.00051	5.56588	0.26021	0.34329	0.01598	0.996	1920	8	1911	39	1902	76	1.0
727Z8-2	1348528	17	0.11783	0.00048	5.60730	0.23288	0.34514	0.01427	0.995	1924	7	1917	35	1911	68	0.7
72728-3	1044408	6	0.11742	0.00051	5.00409	0.21203	0.34619	0.01597	0.993	1917	8	1917	32	1916	62 76	0.1
72770-2	608101	5	0.118/6	0.00055	5.68664	0.23007	0.33115	0.01330	0.995	1922	0 8	1001	40	1044	64	4.0
727Z10-1	573377	17	0.11876	0.00055	5.64537	0.21409	0.34476	0.01298	0.993	1938	8	1923	32	1910	62	1.7
727Z10-2	768786	5	0.11863	0.00053	5.53977	0.19218	0.33869	0.01165	0.992	1936	8	1907	29	1880	56	3.3
727M1-1	292735	70	0 11755	0 00113	5 54255	0 19177	0.34196	0 01137	0 961	1919	17	1907	29	1896	54	14
727M1-2	345810	78	0.11792	0.00115	5.43236	0.19746	0.33413	0.01170	0.963	1925	17	1890	31	1858	56	4.0
727M1-3	355832	64	0.11784	0.00112	5.45594	0.19211	0.33581	0.01138	0.963	1924	17	1894	30	1866	55	3.4
727M1-4	325552	45	0.11761	0.00111	5.39197	0.19993	0.33250	0.01192	0.967	1920	17	1884	31	1850	57	4.2
727M1-5	321466	72	0.11789	0.00112	5.40423	0.21037	0.33248	0.01255	0.970	1924	17	1886	33	1850	60	4.4
727M2-1	285552	65	0.11783	0.00114	5.58575	0.20039	0.34382	0.01188	0.963	1924	17	1914	30	1905	57	1.1
727M2-2	238150	50	0.11775	0.00115	5.60305	0.23876	0.34511	0.01431	0.973	1922	17	1917	36	1911	68	0.7
727M2-3	198152	60	0.11775	0.00112	5.42931	0.25621	0.33441	0.01546	0.979	1922	17	1889	40	1860	74	3.7
7271V12-4	277721	63 79	0.11855	0.00114	5.45844	0.23501	0.33394	0.01401	0.975	1934	17	1894	30	1857	67	4.6
727M2-6	275162	70 68	0.11824	0.00113	5.36157	0.23240	0.32888	0.01301	0.971	1934	17	1934	32	1833	58	5.8
727M2-1	401460	87	0.11766	0.00113	5.50516	0.23199	0.33935	0.01393	0.974	1921	17	1901	36	1884	67	2.2
727M3-2	430645	94	0.11778	0.00111	5.43323	0.19615	0.33458	0.01166	0.965	1923	17	1890	30	1861	56	3.7
727M3-3	358131	49	0.11789	0.00114	5.33866	0.21547	0.32843	0.01287	0.971	1925	17	1875	34	1831	62	5.6
727M3-4	364941	67	0.11846	0.00114	5.55649	0.26263	0.34020	0.01574	0.979	1933	17	1909	40	1888	75	2.7
727M3-5	385079	82	0.11783	0.00116	5.48140	0.22093	0.33738	0.01319	0.970	1924	17	1898	34	1874	63	3.0
727M4-1	463782	102	0.11703	0.00114	5.55675	0.25181	0.34436	0.01524	0.977	1911	17	1909	38	1908	73	0.2
727M4-2	592073	105	0.11750	0.00112	5.51446	0.22468	0.34038	0.01348	0.972	1919	17	1903	34	1888	65	1.8
727M4-3	455241	112	0.11725	0.00112	5.72366	0.24438	0.35404	0.01473	0.975	1915	17	1935	36	1954	70	-2.4
7271014-4	451796	90	0.11697	0.00113	5.49248	0.20608	0.34057	0.01235	0.966	1910	17	1899	32	1889	59	1.3
727M5-1	432220	103	0.11741	0.00113	5 3//06	0.20144	0.33441	0.01202	0.900	1917	17	1876	3/	18/1	50	3.5
727M5-2	451125	95	0.11720	0.00113	5 50983	0.18964	0.34102	0.01127	0.960	1914	17	1902	29	1892	54	1.3
727M5-3	515341	95	0.11770	0.00114	5.45408	0.18746	0.33608	0.01108	0.959	1922	17	1893	29	1868	53	3.2
727M5-4	340765	91	0.11762	0.00113	5.48767	0.19923	0.33838	0.01184	0.964	1920	17	1899	31	1879	57	2.5
727M5-5	358094	93	0.11726	0.00112	5.30037	0.22053	0.32784	0.01328	0.973	1915	17	1869	35	1828	64	5.2
727M6-1	413668	92	0.11731	0.00111	5.47789	0.23052	0.33867	0.01389	0.974	1916	17	1897	36	1880	67	2.1
727M6-2	415844	102	0.11722	0.00113	5.65156	0.25324	0.34968	0.01530	0.977	1914	17	1924	38	1933	73	-1.1
727M6-3	387155	73	0.11743	0.00112	5.38982	0.19355	0.33288	0.01153	0.964	1918	17	1883	30	1852	56	3.9
727M6-4	441432	77	0.11726	0.00113	5.37788	0.20384	0.33262	0.01219	0.967	1915	17	1881	32	1851	59	3.8
727M6-5	439833	72	0.11725	0.00112	5.25634	0.18658	0.32515	0.01111	0.963	1915	17	1862	30	1815	54	6.0
/2/M/-1	254/28	82 71	0.11779	0.00116	5.21002	0.23099	0.32079	0.01386	0.975	1923	18	1854	3/	1/94	67 FF	1.1
1211111-2	214002	11	0.11//0	0.00110	J.JJ00/	0.13140	0.00020	0.01134	0.902	1322	17	10/0	30	1029	55	4.9

727M7-3	279178	85	0.11766	0.00115	5.47386	0.20165	0.33741	0.01199	0.964	1921	17	1897	31	1874	58	2.8
727M7-4	233011	72	0.11765	0.00114	5.46489	0.23467	0.33690	0.01409	0.974	1921	17	1895	36	1872	68	2.9
727M7-5	239857	77	0 11809	0.00117	5 40477	0 21737	0 33193	0.01294	0.969	1928	18	1886	34	1848	62	4.8
727M8_1	242643	81	0.11835	0.00117	5 2/60/	0.21008	0.32154	0.01204	0.000	1020	18	1860	35	1707	63	4.0 8.0
727100-1	100026	70	0.11000	0.00117	5.24034	0.21000	0.32134	0.01304	0.072	1007	17	1070	31	1026	63 E7	5.0 E 4
727100-2	190930	07	0.11005	0.00115	5.30313	0.19921	0.32545	0.01101	0.905	1927	10	1079	31	1030	51	J.4
7271018-3	120404	87	0.11773	0.00125	5.36071	0.19207	0.33184	0.01129	0.954	1922	19	1003	30	1847	54	4.5
/2/M8-4	138633	69	0.11832	0.00116	5.36087	0.21670	0.32859	0.01289	0.970	1931	17	1879	34	1832	62	5.9
727M8-5	131255	70	0.11860	0.00118	5.40313	0.22270	0.33041	0.01322	0.971	1935	18	1885	35	1840	64	5.6
727M9-1	253206	70	0.11774	0.00116	5.43572	0.18000	0.33484	0.01059	0.955	1922	18	1891	28	1862	51	3.6
727M9-2	166968	64	0.11807	0.00116	5.31681	0.23369	0.32660	0.01399	0.975	1927	17	1872	37	1822	68	6.3
727M9-3	138782	76	0.11722	0.00125	5.17509	0.19583	0.32020	0.01162	0.959	1914	19	1849	32	1791	57	7.4
727M9-4	145455	88	0.11830	0.00118	5.30997	0.19507	0.32555	0.01151	0.962	1931	18	1870	31	1817	56	6.8
727M9-5	118962	68	0.11798	0.00122	5.30417	0.24382	0.32607	0.01460	0.974	1926	18	1870	39	1819	71	6.3
727M9-6	124198	77	0.11806	0.00122	5.39934	0.20307	0.33169	0.01200	0.962	1927	18	1885	32	1847	58	4.8
727M10-1	746950	111	0.11771	0.00112	5.37840	0.17819	0.33138	0.01051	0.957	1922	17	1881	28	1845	51	4.6
727M10-2	584363	106	0.11762	0.00112	5.43312	0.23642	0.33501	0.01422	0.976	1920	17	1890	37	1863	68	3.5
727M10-3	697547	116	0.11762	0.00110	5.51792	0.18683	0.34025	0.01107	0.961	1920	17	1903	29	1888	53	2.0
727M10-4	595224	132	0 11737	0.00113	5 19833	0 18234	0.32121	0.01083	0.961	1917	17	1852	29	1796	53	7.2
727M10-5	644061	98	0 11737	0.00112	5 26887	0.20690	0.32557	0.01240	0.970	1917	17	1864	33	1817	60	6.0
7271010-5	044001	30	0.117.57	0.00112	5.20007	0.20030	0.52557	0.01240	0.370	1317	17	1004	55	1017	00	0.0
115Z1-1	1217527	77	0.11786	0.00122	5.92266	0.32964	0.36446	0.01993	0.983	1924	18	1965	47	2003	93	-4.8
115Z1-2	1172925	46	0.11723	0.00122	5.59532	0.24747	0.34618	0.01488	0.972	1914	18	1915	37	1916	71	-0.1
115Z2-1	920039	84	0.11748	0.00121	5.76245	0.30531	0.35576	0.01849	0.981	1918	18	1941	45	1962	87	-2.7
115Z3-1	1212874	48	0.11780	0.00123	5.78402	0.24789	0.35611	0.01480	0.970	1923	19	1944	36	1964	70	-2.4
115Z3-2	849986	32	0.11751	0.00122	5.70710	0.28799	0.35223	0.01740	0.979	1919	18	1932	43	1945	82	-1.6
115Z4-1	1447668	68	0.11778	0.00125	5.64878	0.28475	0.34784	0.01714	0.978	1923	19	1924	43	1924	81	-0.1
115Z4-2	1304648	61	0.11741	0.00122	5.38572	0.25262	0.33268	0.01522	0.975	1917	18	1883	39	1851	73	4.0
115Z5-1	486891	34	0.11835	0.00124	5.74637	0.25584	0.35214	0.01524	0.972	1931	19	1938	38	1945	72	-0.8
115Z5-2	384188	37	0.11849	0.00125	5.75063	0.41828	0.35199	0.02533	0.990	1934	19	1939	61	1944	120	-0.6
115Z6-1	513437	38	0.11790	0.00125	5.66651	0.31703	0.34858	0.01915	0.982	1925	19	1926	47	1928	91	-0.2
115Z7-1	673528	22	0.11841	0.00123	5,59408	0.23266	0.34265	0.01380	0.968	1932	18	1915	35	1899	66	2.0
11577-2	629066	9	0 11851	0.00124	5 65028	0 26397	0 34578	0.01574	0 974	1934	19	1924	40	1914	75	12
11578-1	798806	26	0.11810	0.00124	5 60/87	0.25478	0.34304	0.01521	0.073	1029	10	1024	38	1006	73	1.4
11520-1	750000	20	0.11019	0.00124	5.00407	0.20470	0.34354	0.01321	0.973	1929	10	1004	30	1000	04	2.6
11528-2	152223	29	0.11799	0.00123	5.51991	0.29207	0.33930	0.01760	0.980	1920	19	1904	44	1003	84 66	2.0
11529-1	303830	32	0.11842	0.00127	5.92145	0.23920	0.36265	0.01412	0.964	1933	19	1964	34	1995	00	-3.7
11529-2	223454	22	0.11892	0.00130	5.81051	0.23784	0.35438	0.01398	0.964	1940	19	1948	35	1955	66	-0.9
115Z10-1	1179497	28	0.11869	0.00128	5.58582	0.30967	0.34133	0.01856	0.981	1937	19	1914	47	1893	89	2.6
115Z10-2	734917	25	0.11804	0.00123	5.59484	0.27796	0.34376	0.01670	0.978	1927	18	1915	42	1905	80	1.3
115M1-1	847833	70	0.11725	0.00132	5.53896	0.18949	0.34263	0.01106	0.944	1915	20	1907	29	1899	53	0.9
115M1-2	782057	72	0.11725	0.00133	5.59260	0.18197	0.34595	0.01055	0.938	1915	20	1915	28	1915	50	0.0
115M1-3	697122	66	0.11736	0.00133	5.67329	0.32906	0.35060	0.01995	0.981	1916	20	1927	49	1937	95	-1.3
115M1-4	495401	157	0.11841	0.00161	5.62396	0.22655	0.34447	0.01306	0.941	1932	24	1920	34	1908	62	1.4
115M2-1	506175	61	0.11770	0.00135	5.53528	0.23935	0.34109	0.01422	0.964	1922	20	1906	37	1892	68	1.8
115M2-2	542811	49	0.11848	0.00136	5.51837	0.24080	0.33780	0.01422	0.965	1933	20	1903	37	1876	68	3.4
115M3-1	475784	77	0.11742	0.00133	5,76305	0.27470	0.35596	0.01648	0.971	1917	20	1941	40	1963	78	-2.8
115M3-2	421297	56	0 11744	0.00135	5 56591	0 21298	0 34373	0.01255	0 954	1918	20	1011	32	1905	60	0.8
115M3_3	27/108	38	0.11736	0.00134	5 01118	0.21200	0.36530	0.01200	0.007	1016	20	1063	46	2007	01	-5.5
1151010-5	200246	60	0.11730	0.00134	5.01110	0.01007	0.26117	0.01001	0.070	1021	20	1060	40	1000	06	-0.0
110100-4	290240	75	0.11034	0.00139	5.69524	0.33914	0.30117	0.02033	0.979	1931	21	1900	49	1900	90	-3.4
1101010-0	444490	10	0.11770	0.00130	5.00423	0.27005	0.34391	0.01056	0.972	1923	21	1914	42	1905	79	1.0
1151015-0	430325	109	0.11809	0.00142	5.62035	0.21010	0.34519	0.01222	0.947	1927	21	1919	32	1912	56	1.0
115M4-1	506814	58	0.11801	0.00133	5.66053	0.23378	0.34788	0.01382	0.962	1926	20	1925	35	1924	66	0.1
115M4-2	317334	72	0.11748	0.00137	5.63098	0.19496	0.34764	0.01134	0.942	1918	21	1921	29	1923	54	-0.3
115M4-3	329094	64	0.11823	0.00140	5.80980	0.21357	0.35639	0.01241	0.947	1930	21	1948	31	1965	59	-2.1
115M4-4	347709	72	0.11822	0.00139	5.81010	0.22558	0.35643	0.01319	0.953	1930	21	1948	33	1965	62	-2.1
115M4-5	264711	60	0.11863	0.00136	5.80198	0.30076	0.35472	0.01793	0.975	1936	20	1947	44	1957	85	-1.3
115M4-6	286254	43	0.11837	0.00139	5.72114	0.26184	0.35053	0.01550	0.966	1932	21	1935	39	1937	74	-0.3
115M4-7	357758	27	0.11835	0.00136	5.75095	0.27789	0.35241	0.01654	0.971	1932	20	1939	41	1946	78	-0.9
115M4-8	339618	37	0.11718	0.00139	5.59355	0.22407	0.34619	0.01325	0.955	1914	21	1915	34	1916	63	-0.2
115M5-1	447075	26	0.11864	0.00137	5.35324	0.19697	0.32724	0.01143	0.949	1936	21	1877	31	1825	55	6.6
115M5-2	396200	29	0.11882	0.00137	5.60605	0.24623	0.34220	0.01451	0.965	1938	20	1917	37	1897	69	2.5
115M5-3	261146	26	0.11726	0.00134	5.35793	0.19669	0.33139	0.01156	0.950	1915	20	1878	31	1845	56	4.2
115M5-4	276912	9	0.11760	0.00134	5.56277	0.20104	0.34308	0.01177	0.949	1920	20	1910	31	1901	56	1.1
115M5-5	250502	24	0 11858	0.00152	5 48831	0 25860	0.33568	0.01523	0.962	1935	23	1899	40	1866	73	4 1
115M5-6	327177	11	0 11835	0.00135	5 51846	0 25901	0.33818	0.01540	0.970	1931	20	1903	40	1878	74	3.2
115ME 7	222166	20	0.11000	0.00100	5 54744	0.22500	0.34400	0.01204	0.060	1025	20	1000	26	1902	67	0.Z
	223 100	20	0.11/90	0.00137	5.547 14	0.23570	0.54109	0.01394	0.902	1920	21	1900	30	1092	0/	2.0

115M6-1	317200	19	0.11904	0.00136	5.63691	0.18764	0.34342	0.01074	0.939	1942	20	1922	28	1903	51	2.3
115M6-2	276268	14	0.11884	0.00137	5.71849	0.20108	0.34899	0.01160	0.945	1939	20	1934	30	1930	55	0.5
115M6-3	191183	9	0.11765	0.00135	5.55206	0.28699	0.34227	0.01725	0.975	1921	20	1909	44	1898	82	1.4
115M6-4	267712	19	0.11911	0.00138	5.45765	0.20159	0.33233	0.01166	0.950	1943	21	1894	31	1850	56	5.5
115M6-5	319423	13	0.11938	0.00138	5.70518	0.20115	0.34660	0.01155	0.945	1947	20	1932	30	1918	55	1.7
115M6-6	269064	4	0.11870	0.00140	5.50882	0.23648	0.33659	0.01389	0.962	1937	21	1902	36	1870	67	4.0
115M6-7	272803	2	0.11857	0.00136	5.54980	0.18464	0.33947	0.01061	0.939	1935	20	1908	28	1884	51	3.0
115M6-8	305624	4	0.11814	0.00137	5.58467	0.18095	0.34283	0.01037	0.934	1928	21	1914	28	1900	50	1.7
115M6-9	387632	18	0.11876	0.00136	5.54582	0.20776	0.33870	0.01208	0.952	1938	20	1908	32	1880	58	3.4
115M6-10	288541	9	0.11858	0.00137	5.69460	0.24151	0.34829	0.01421	0.962	1935	21	1931	36	1926	68	0.5
115M6-11	237442	6	0.11859	0.00135	5.57070	0.18604	0.34068	0.01069	0.940	1935	20	1912	28	1890	51	2.7
115M6-12	195421	1	0.11819	0.00140	5.59572	0.20415	0.34338	0.01185	0.946	1929	21	1915	31	1903	57	1.6
115M7-1	31585	3	0.11941	0.00183	5.36638	0.19443	0.32594	0.01070	0.906	1947	27	1880	31	1819	52	7.6
115M7-2	75059	0	0.11845	0.00140	5.37027	0.17519	0.32881	0.01000	0.932	1933	21	1880	28	1833	48	6.0
115M7-3	123940	2	0.11875	0.00139	5.46710	0.22653	0.33390	0.01328	0.960	1938	21	1895	35	1857	64	4.8
115M7-5	64799	1	0.11828	0.00141	5.43391	0.18306	0.33318	0.01050	0.935	1930	21	1890	28	1854	51	4.6
115M7-6	93580	1	0.11844	0.00150	5.47680	0.30862	0.33537	0.01842	0.974	1933	22	1897	47	1864	88	4.1
115M8-1	349737	2	0.11801	0.00134	5.41604	0.20914	0.33286	0.01229	0.956	1926	20	1887	33	1852	59	4.4
115M8-2	152870	1	0.11745	0.00134	5.67650	0.21829	0.35052	0.01287	0.955	1918	20	1928	33	1937	61	-1.2
115M8-3	211486	4	0.11775	0.00138	5.44229	0.21073	0.33522	0.01237	0.953	1922	21	1892	33	1864	59	3.5
115M8-4	232489	1	0.11827	0.00137	5.45080	0.21652	0.33426	0.01270	0.956	1930	21	1893	34	1859	61	4.2
115M8-5	185513	0	0.11846	0.00136	5.41247	0.19380	0.33139	0.01124	0.947	1933	20	1887	30	1845	54	5.2
115M8-6	329626	0	0.11851	0.00136	5.27303	0.23818	0.32271	0.01410	0.967	1934	20	1865	38	1803	68	7.8
115M8-7	211280	0	0.11847	0.00137	5.42457	0.20204	0.33209	0.01176	0.951	1933	21	1889	31	1848	57	5.0
115M9-1	506883	2	0.11812	0.00136	5.66918	0.23292	0.34811	0.01373	0.960	1928	20	1927	35	1926	65	0.1
115M9-2	385531	0	0.11793	0.00134	5.44416	0.22524	0.33482	0.01332	0.961	1925	20	1892	35	1862	64	3.8
115M9-4	427056	8	0.11805	0.00136	5.66513	0.30094	0.34806	0.01805	0.976	1927	20	1926	45	1925	86	0.1
115M9-5	388017	3	0.11840	0.00136	5.57795	0.21912	0.34169	0.01283	0.956	1932	20	1913	33	1895	61	2.2
115M9-6	309381	0	0.11796	0.00134	5.56630	0.27194	0.34225	0.01626	0.973	1926	20	1911	41	1897	78	1.7
115M9-7	269928	9	0.11958	0.00154	5.56424	0.23424	0.33748	0.01353	0.952	1950	23	1911	36	1875	65	4.5
115M10-1	716023	3	0.11766	0.00134	5.52631	0.21021	0.34063	0.01236	0.954	1921	20	1905	32	1890	59	1.9
115M10-2	642504	0	0.11754	0.00134	5.57055	0.21481	0.34372	0.01266	0.955	1919	20	1912	33	1905	60	0.9
115M10-3	483023	0	0.11791	0.00136	5.62666	0.21365	0.34609	0.01252	0.953	1925	20	1920	32	1916	60	0.5
115M10-4	439574	1	0.11717	0.00133	5.51022	0.22484	0.34109	0.01337	0.961	1913	20	1902	34	1892	64	1.3
115M10-5	448553	0	0.11771	0.00135	5.50703	0.19602	0.33932	0.01143	0.947	1922	20	1902	30	1883	55	2.3
115M10-6	559258	1	0.11766	0.00133	5.47440	0.23735	0.33744	0.01412	0.965	1921	20	1897	37	1874	68	2.8
115M10-7	332111	0	0.11797	0.00135	5.42229	0.24440	0.33337	0.01453	0.967	1926	20	1888	38	1855	70	4.2

*Not common Pb corrected