**Paleoproterozoic (1.75 Ma) Granitoid Rocks and Uranium Mineralization in the
Baker Lake – Thelon Basin region, Nunavut**

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ABSTRACT

The Nuelin suite is comprised of the Nuelin granite, Pitz Formation rhyolite and related basaltic dykes. Despite the extensive range of emplacement, petrographic and geochemical analysis show that all Nuelin suite components are feldspar-phyric, strongly potassic, Fe-rich and highly fractionated, with enrichment in rare earth elements, except for strongly depleted Eu. Overall, the Nuelin suite shares strong affinity with anorogenic (A-type) igneous rocks, with an interpreted petrogenetic model of crustal melting due to ponding of basaltic melt in the lower crust within an extensional regime. Fractionation is likely responsible for enrichment of U, Th and REE. Significant U mineralization, located in the Kiggavik area, is associated with the Lone Gull granite, a geologically complex granite. The presence of secondary uranium mineralization in association with Paleoproterozoic Nuelin igneous rocks implicate the suite as prospective for exploration.

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TABLE OF CONTENTS

ABSTRACT	ii
ACKNOWLEDGEMENTS.....	iii
LIST OF TABLES.....	vii
LIST OF FIGURES	viii
Chapter 1 : Introduction.....	1
Chapter 2 : Geologic and Tectonic Setting.....	3
2.1. REGIONAL GEOLOGY.....	3
2.1.1. Western Churchill Province	3
2.1.2. Hearne Domain.....	3
2.1.3. Rae Domain.....	4
2.1.4. Trans Hudson Orogeny.....	5
2.1.5. Dubawnt Supergroup.....	5
2.1.6. Lower Dubawnt Supergroup	6
2.1.6.1. Christopher Island Formation and Dubawnt Minettes.....	7
2.1.6.2. Hudson Granitoid.....	8
2.1.6.3. Martell Syenite.....	9
2.1.7. Middle Dubawnt Supergroup	10
2.1.7.1. Nuelin Granite.....	11
2.1.7.2. Pitz Rhyolite.....	11
2.1.7.3. McRae Lake Dyke	12
2.1.8. Upper Dubawnt Supergroup.....	13
2.2. LOCAL GEOLOGY	14
2.2.1. Nuelin Lake	15
2.2.2. Mallery Lake	17
2.2.3. Kiggavik uranium deposit	19
2.2.3.1. Lone Gull granite	21
2.2.4. Amer plutonic complex	21
Chapter 3 : Methods.....	31
3.1. FIELD WORK	31

3.2. ANALYTICAL METHODS.....	31
3.2.1. Petrography	31
3.2.2. Whole Rock Geochemistry	32
3.2.3. Neodymium-Isotope Geochemistry.....	33
3.2.4. U-Pb SHRIMP Geochronology.....	34
Chapter 4 : Results.....	37
4.1. PETROGRAPHY	37
4.1.1. Nueltn Lake - Nueltn granite.....	37
4.1.1.1. Fine-grained aplite dyke.....	38
4.1.1.2. Pegmatitic granitic seam	39
4.1.2. Mallery Lake - subvolcanic Nueltn granite.....	40
4.1.2.1. Mallery basalt.....	42
4.1.2.2. Pitz basalt	43
4.1.3. Kiggavik U-deposit - Lone Gull granite.....	44
4.1.4. Amer plutonic complex - rapakivi granite	46
4.1.4.1. Amer plutonic complex - aplite dykes	47
4.2. GEOCHEMISTRY	58
4.2.1. Nueltn Lake - Nueltn granite.....	59
4.2.1.1. Fine-grained aplite dyke.....	60
4.2.1.2. Pegmatitic granitic seam	61
4.2.2. Mallery Lake - subvolcanic Nueltn granite.....	62
4.2.2.1. Mallery basalt.....	62
4.2.2.2. Pitz basalt	63
4.2.3. Kiggavik U-deposit - Lone Gull granite.....	64
4.2.4. Amer plutonic complex - rapakivi granite	65
4.2.4.1. Aplite dyke	66
4.3. NEODYMIUM-ISOTOPE CHEMISTRY	74
4.4. URANIUM-LEAD SHRIMP GEOCHRONOLOGY	79
4.4.1. Lone Gull granite - 08JP-52A	79
4.4.2. Lone Gull granite - 09PHA-J39	80
Chapter 5 : INTERPRETATION AND DISCUSSION	88
5.1. CHARACTERISTICS OF THE NUELTN SUITE.....	88
5.1.1. Nueltn Suite Geochemistry	88
5.1.2. Nueltn Suite REE Geochemistry	90
5.1.3. Nueltn Suite Isotope Geochemistry.....	92
5.2. NUELTN SUITE - ANOROGENIC (A-TYPE) SUITE	93

5.3. NUEL TIN (A-TYPE) SUITE AND URANIUM PROSPECTIVENESS	97
5.4. SOURCES OF UNCERTAINTY	102
: CONCLUSIONS	116
REFERENCES	121
Appendix A: Latitude and longitude co-ordinates for samples	130
Appendix B: Whole Rock Analysis.....	131
Appendix C: Bulk continental crust normalization values	137
Appendix D: REE chondrite normalization values	138

LIST OF TABLES

Table 4.1. List of Th to U ratios for units from this study.	69
Table 4.3. Sm and Nd isotopic analyses from Nueltin suite rocks.....	76
Table 4.4. U-Pb SHRIMP isotopic analyses from Lone Gull granite.....	82

LIST OF FIGURES

Figure 2.1. Map of the western Churchill Province highlighting Proterozoic igneous rocks. Areas of study are highlighted. Map modified from Peterson et al. (2002).	24
Figure 2.2. Nueltin Lake study area. Sample locations marked on inset, with anomalous U-Th samples labelled. The filled square indicates a "vein" occurrence containing Th and REE and the open square represents a polymetallic skarn deposit, both located by Charbonneau and Swettenham (1986).	25
Figure 2.3. Geology map of Mallery Lake area modified from LeCheminant et al. (1981). Outcrop sample locations of Pitz basalt are located north of Ruby Lake. Samples collected at the Mallery Lake peninsula are highlighted on the inset.	26
Figure 2.4. Close-up map of Mallery Lake peninsula: a) geological units and field stations with contacts inferred according to b) magnetic map.	27
Figure 2.5. Simplified geology map of the Kiggavik region, western NTS 66A and eastern 66B, from Jefferson et al. (GSC Open File, in prep.)	28
Figure 2.6. Cartoon cross-section of the Kiggavik deposit Main Zone, showing inferred relationship of uranium mineralization with the Lone Gull (Nueltin) granite.	29
Figure 2.7. Geology of Amer plutonic complex (after maps in (after maps in Jefferson et al., 2011b). Geological units and field stations (upper) and expression of geological units in total magnetic field lower after Tschirhart et al. (2011).	30
Figure 4.1. Field photos of Nueltin Lake granite outcrops: a) field photo of Nueltin granite showing weak igneous alignment (general direction oriented potassium feldspars); b) photo of Nueltin granite in shallow water showing numerous angular to sub-rounded xenoliths (up to 30 cm), aligned sub-parallel; c) apophysis of coarse Nueltin granite cutting sandstone; d) pegmatitic seam of primary U-, Th-, and REE-rich minerals within Nueltin granite.	48
Figure 4.2. Photos of slab sections and photomicrographs of Nueltin granite showing mineralogy: a) megacrysts (upper left is 5.5 cm and upper right is 4.5 cm) of potassium feldspar within Nueltin granite; b) polished slab of Nueltin granite showing typical mineralogy and coarse texture, with larger potassium feldspar crystals commonly oikocrystic and a resorbed plagioclase feldspar (at centre) with inclusions of quartz and biotite; c) anhedral green coloured K-hastingsite (amp), spatially associated with annitic biotite included by titanite and opaques (Fe-Ti-oxide), in plane polarized light (ppl); d) inclusion-rich coarse biotite, shows pleochroic halos around zircon and monazite, in cross polarized light (xpl); e) Carlsbad twinned plagioclase crystal showing fine polysynthetic twinning (lower twin) and diffuse chemical zoning (upper twin) with mymekitic texture; f) weakly zoned allanite (blue and yellow) adjacent to biotite and titanite, in xpl.	49

Figure 4.3. Photos of slab and thin section of aplite dyke (09PHA-J05) which was hosted by Nueltin granite, at Nueltin Lake: a) polished slab of aplite dyke showing heterogeneity and weak foliation defined by biotite with medium-grained greenish colour allanite; b) accompanying autoradiograph negative highlights radioelement bearing phases that coincide with allanite (larger) and thorite (smaller); c) rare earth-bearing allanite (al) is more coarse than that found in main phase of Nueltin granite with common U-bearing thorite (th). 50

Figure 4.4. Photos of slab sections from Nueltin granite hosted pegmatitic seam (09PHA-J06): a) polished slab of pegmatite seam showing coarse fluorite (fl) and allanite (greenish colour); b) autoradiograph film exposed over two weeks of the polished slab which highlights spatial association of radioelement bearing minerals, allanite (light gray) and urano-thorite (black); c) primary igneous allanite with visible twinning, adjacent to chloritized biotite (lower right) and microcline (upper left), with heavily metamictized, isotropic thorite (th); d) back scatter electron image of larger allanite from pegmatitic seam dominated by subhedral allantie (al) with contrasting core and rim, due to difference in rare earth element (REE) content and brighter inclusions consist of urano-thorite (U) and clots of REE-carbonates are in the brighter core.... 51

Figure 4.5. Photos of samples from Mallery Lake area: a) field photo of low relief, areally small outcrop of subvolcanic Nueltin granite; b) polished slab of subvolcanic granite, hematite stained, rounded potassium feldspar phenocryst (some show concentric biotite inclusions) and chloritic mafic clots; c) photomicrograph of granophyric texture surrounding potassium feldspar phenocrysts in subvolcanic Nueltin granite; d) photomicrograph of Fe-rich chlorite, with inclusions of titanite (centre); e) field photo of quartz stockwork zone hosted by fractured subvolcanic Nueltin granite (scale in cm); f) sample from stockwork with vuggy quartz and coarse, dark purple fluorite cutting silicified rhyolite. 52

Figure 4.6. Photos of samples from Mallery Lake mafics: a) field photo of mafic bedded volcanoclastic rocks overlain by b) hematized feldspar-phyric felsic materials (scale card is 7 cm); c) polished slab showing primary bedding of tuff; c) photomicrograph of Fe-rich chlorite, with inclusions of grainy titanite and opaques (commonly magnetite). 53

Figure 4.7. Photos of samples from Pitz basalt (10PHA-J06) north of Ruby Lake: a) field photo of feldspar-phyric basalt with maroon weathered groundmass (scale card is 7 cm); b) polished slab showing aligned chloritic clots (appear black) and epidote (ep) amygdules; c) close-up showing larger phenocryst (upper right) incorporating matrix material with arrows indicating elongated chloritic masses; d) photomicrograph of epidote-filled amygdule, with Fe-rich chlorite rim and pseudomorph (indicate by arrow); e) resorbed plagioclase, with adjacent chlorite (arrows). 54

Figure 4.8. Photos from sample of Lone Gull granite (08JP-52A) at 75.5 m depth: a) polished slab showing typical porphyritic texture of Lone Gull granite with large mafic inclusion (right) and smaller mafic blebs (left); b) photomicrograph of Lone Gull granite (08JP-52A), arrows highlight potassium feldspar and plagioclase with overgrowths. Quartz phenocrysts (lower right) are highly rounded (resorbed?) and biotite (centre) is altered completely to chlorite. A typical wedge of common pale titanite is observed (top left). 55

Figure 4.9. Photomicrographs of the Lone Gull granite from depth 17.3 m: a) potassium feldspar with rim showing numerous quartz inclusions and poikilitic plagioclase; b) microphenocryst of chlorite pseudomorph after amphibole. 56

Figure 4.10. Rapakivi granite from Amer plutonic complex: a) field photo of rapakivi granite (scale card 7 cm); b) example of tabular potassium feldspar with well defined plagioclase rim; c) photomicrograph of plagioclase mantle over ragged potassium feldspar crystal boundary; d) photomicrograph of strongly zoned plagioclase (lower left) with euhedral titanite and weakly altered biotite; e) brittle deformation of the rapakivi granite (station 10PHA-J25) followed by sericitization; f) contact between late-crystallizing aplite dyke and strongly hematized host rapakivi granite. 57

Figure 4.11. Selected Harker diagrams for all Nueltn granite and related rocks. Data from previous work from Peterson et al. (2002) and references therein..... 67

Figure 4.12. Trace element geochemistry diagrams for granitic rocks from Nueltn Lake: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974). 68

Figure 4.13. Trace element geochemistry diagrams for rocks from Mallery Lake: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974). 70

4.14. Trace element geochemistry NMORB-normalized after (Sun and McDonough, 1989) diagram for mafic rocks from Mallery Lake. The Mallery basalt has trace element patterns consistent with an enriched MORB normal mid-ocean ridge basalt (NMORB). This contrasts strongly with the Pitz basalt, which has slight negative anomalies in Nb and Ti, characteristic of subduction related basalts. 71

Figure 4.15. Trace element geochemistry diagrams for granitic rocks from Kiggavik: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974). 72

Figure 4.16. Trace element geochemistry diagrams for granitic rocks from Amer plutonic complex: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974). 73

Figure 4.18. Diagram showing $\epsilon\text{Nd}(t)$ versus Age (Ga) for Nueltn granites and related rocks, relative to depleted mantle (DM) and chondritic uniform reservoir (CHUR). Depleted mantle (DM) evolution curve is calculated from $\epsilon\text{Nd}(t) = 2.37$ to $-9.38(t)$ where (t) is the age (DePaolo, 1981). Symbols are the same as Figure 4.16, with the addition of the Mallery Lake basalt (10PHA-J07A2) indicated by the open circle. 78

Figure 4.19. Images of zircon in backscattered electron mode showing the general quality of zircon, analyzed where indicated: a) 08JP-52A, a large pyramidal, embayed Proterozoic zircon analyzed twice, the second analysis failed due to an inclusion in the target located in reflected light; b) 08JP-52A, an inherited Archean core showing very little subsequent rim growth; c) 09PHA-J39, a zircon showing recrystallization at the core revealed Nueltnin-age analysis from an area of fine oscillatory zoning closer to the rim; d) 09PHA-J39, a Hudson age was inferred from the result of analysis over the unzoned core; e) 09PHA-J39, xenocrystic Archean zircon with diffuse zoning and very fine oscillatory zoning along the rim; f) 09PHA-J39, (not analyzed) a conspicuously large embayed and fractured pyramidal zircon, morphologically similar to those from 08JP-52A. 86

Figure 4.20. Concordia diagrams of Proterozoic U-Pb SHRIMP data from Lone Gull granite at the Kiggavik Main Zone. Uncertainties for the analyses shown at the 2σ level. Concordia for sample 08JP-52A (a) has an age of 1822.3 ± 11 Ma. The calculated $^{207}\text{Pb}/^{206}\text{Pb}$ (207/206) age for the sample (b) has a weighted average of 1820 ± 12 Ma, which overlaps with the concordia age. Concordia diagram (c) for Paleoproterozoic aged U-Pb SHRIMP data from Lone Gull granite sample (09PHA-J39). Discrete populations can be resolved by Th/U ratios (d) giving weighted averages (e) and (f) that are 1806 ± 41 Ma (*ca.* Hudson-aged) and 1748 ± 9.4 Ma (Nueltnin-aged), respectively. Re-analysis of the sample, split from the original material, does show *ca.* Nueltnin 207/206 ages (g), but are strongly affected by common Pb contamination (> 0.8 counts per second). Removal of samples with high common Pb contamination leaves a concordia (h) with calculated age 1836.6 ± 14 Ma, which overlaps with the inset that shows the weighted average of 1836 ± 22 Ma. 87

Figure 5.1. Total alkalinity versus silica for Nueltnin suite rocks after Cox et al. (1979). Data from previous work from Peterson et al. (2002) and references therein..... 104

Figure 5.2. Alumina saturation versus silica content shows Nueltnin granite and related rocks progress from strongly metaluminous to strongly peraluminous as rocks become more felsic. 105

Figure 5.4. Continental crust-normalized (Taylor and McLennan, 1995) extended trace element diagram for a) Nueltnin granite and b) related rocks. 106

Figure 5.5. Chondrite-normalized (Nakamura, 1974) rare earth element diagram for a) Nueltnin granite and b) related rocks. 107

Figure 5.6. Increased silica generally correlates with a more pronounced Eu-anomaly, likely due to continued plagioclase fractionation as the melt evolves. 108

Figure 5.7. Diagram showing epsilon Nd vs. age (Ma) for rock suites throughout the western Churchill Province. Nueltnin suite rock from this study appear cross-hatched, overlap with Nueltnin rocks (Dudás et al., 1991; Peterson et al., 2002). Epsilon Nd values from other suites found in western Churchill Province suites are plotted for comparison (Byrne, 2005; Cousens et al., 2001; Peterson et al., 2002). Depleted Mantle evolution curve is calculated assuming $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.51315 and $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.215 for modern depleted mantle. 109

Figure 5.8. Geotectonic diagrams after Pearce et al. (1984) which indicates melt generation environment.	110
Figure 5.9. Geotectonic diagrams after Whalen et al. (1987), differentiating A-type granites (A denotes calculated average) from: a) and b) fractionated felsic granites (FG) and unfractionated I- and S-type granites (OTG); c) to f) I-type and S-type granites.	111
Figure 5.10. Geotectonic diagrams after Eby (1992) differentiating types of anorogenic granites.	112
Figure 5.11. Geotectonic diagram after Frost et al. (2001), outlines indicate established fields for A-type granites, including rapakivi granites.	113
Figure 5.12. Bivariate plots show the covariation of Th and U with increasing potassic character.	114
Figure 5.13. Bivariate plot shows U content (ppm) is typically highest when the alumina index (A/CNK) approaches 1	115
Figure 6.1. Adapted model after (Rämö and Haapala, 2005) for generation of rapakivi granite, with analogous units from this study at approximate levels of exposure.....	120

CHAPTER 1: INTRODUCTION

The late Paleoproterozoic Nueltin suite outcrops across a large portion of the western Churchill Province and marks the last period of extensive igneous activity in the Kivalliq Region of Nunavut. Understanding and characterizing this group of rocks has immediate implications for distinction from regional Hudson suite rocks, and can contribute to a broader understanding of the evolution of the regional geology during the Paleoproterozoic. Based on petrographic, geochemical and geochronological evidence the Nueltin granites are related to an anorogenic event which generated megacrystic and rapakivi-type igneous plutons, typical of those emplaced globally during the Paleoproterozoic (Larin, 2009). Uranium-lead isotope analyses of zircon from granite and rhyolite outcrops have constrained the timing of emplacement from 1765 to 1750 Ma (Rainbird and Davis, 2007; van Breemen et al., 2005). This is well after the Trans-Hudsonian orogeny and the protracted emplacement of the post-orogenic Hudson granitoid suite (1845 to 1795 Ma), and the contemporaneous extrusion of Christopher Island Formation. Occurrences of uranium, rare earth elements (REE) and precious metals (Au-Ag), found in association with Nueltin suite rocks implicate the suite as prospective for exploration and economic potential.

U exploration has been primarily conducted within the Baker Lake and Thelon basins and immediately surrounding basement rocks, as these basins are considered coeval to the Athabasca and Martin basins that contain world class U-deposits (Miller and LeCheminant, 1985). Within the western Churchill Province some showings appear

to be related spatially to Proterozoic granitoid and volcanic rocks. The most prominent association between Nuelin suite rocks and U mineralization is found at the Kiggavik uranium deposit. U mineralization is primarily pitchblende disseminated within local sedimentary rocks and partly within the complex Lone Gull granite (Weyer et al., 1987). Miller and LeCheminant (1985) record thermally altered miarolitic fluorite-bearing granites as potential source rocks for U-bearing polymetallic systems, citing vein U and Cu mineralization northwest of Forde Lake. In the Nuelin Lake area, west of Sandybeach Lake, several examples of vein U-, Th- and REE-mineralization (Charbonneau and Swettenham, 1986; this study) have been reported. The Mallery Lake Au-Ag deposits are within quartz-chalcedony epithermal stockworks, hosted dominantly in fluorite-bearing Pitz Formation rhyodacite (Turner et al., 2003). Locales with substantial glass-rich Pitz Formation volcanoclastic deposits are also prospective source areas for uranium upon its release from devitrifying volcanic glass, drawing analogues from the Duobblon occurrence in northern Sweden (Miller and LeCheminant, 1985).

The purpose of this thesis is to characterize the Nuelin granite, comagmatic Pitz Formation rhyolite, contemporaneous McRae Lake diabase and related mafic units. This is accomplished by means of geological context, petrographic, geochronologic and geochemical analysis. Using trace element and isotope chemistry, the geotectonic and petrogenetic nature of the Nuelin suite can be inferred. Furthermore, understanding the formation processes of the Nuelin suite could lead to a better understanding of granite hosted or spatially associated U mineralization in the Baker Lake - Thelon basin.

CHAPTER 2: GEOLOGIC AND TECTONIC SETTING

2.1. Regional Geology

2.1.1. Western Churchill Province

The western Churchill Province is a composite Archean craton welded to the adjacent Slave and Superior Archean cratons the Paleoproterozoic Taltson-Thelon and Trans-Hudson orogens (Hoffman, 1988). The western Churchill Province is divided into the Rae and Hearne domains, both of which comprise 3.0-2.6 Ga low-metamorphic grade greenstone belts and intrusive granitoid rocks juxtaposed with high-grade gneiss and clastic sedimentary sequences (Hoffman, 1988). The Archean Snowbird Tectonic Zone (STZ), a crustal scale fault reactivated during the Paleoproterozoic, juxtaposes the two domains.

2.1.2. Hearne Domain

The Hearne domain, located on the southeast side of the STZ, extends from northern Saskatchewan to Hudson Bay, an extent of approximately 700 km. Neoarchean tonalite to granodiorite gneiss and plutons predominate regionally. In the northwest and central portions of the Hearne domain, Neoarchean volcanosedimentary sequences and associated intrusive packages constitute rocks of the Ennadai-Rankin greenstone belt (Hanmer et al., 2004). The southern portion of the domain is dominated by metasedimentary rocks of the Wollaston Supergroup (Annesley et al., 2005). The

Neoarchean belts are unconformably overlain by areally restricted, earliest Paleoproterozoic coarse clastic rocks of the Montgomery Group and more widespread platformal quartzite, mudstone, carbonate strata, and minor tholeiitic volcanic rocks of early Paleoproterozoic Hurwitz Group (Sandeman et al., 2003).

2.1.3. Rae Domain

The Rae domain extends more than 2000 km from northern Saskatchewan to Baffin Island to the west and north of the STZ. The dominant rocks in the central Rae domain are Archean para- and ortho-gneiss which form basement to the supracrustal Woodburn Lake and Prince Albert groups. The latter supracrustal package comprises komatiitic volcanic, mature to relatively immature siliciclastic and rare mafic-felsic extrusive rocks (Skulski et al., 2003). To the southwest, the *ca.* 2.7-2.8 Woodburn Lake group is a sequence of early komatiitic flows, felsic-intermediate-mafic volcanic rocks and epiclastic strata (Zaleski et al., 2000). Late Archean igneous activity was recorded by the 2.6 Ga Committee Bay granitoids intrusions (Byrne, 2005). The youngest Neoarchean strata are widespread quartz eye rhyolite dated by LeCheminant and Roddick (1991) at *ca.* 2.6 Ga. The overlying quartzite, thought by Zaleski et al. (2000) to be Neoarchean, is now considered to be early Paleoproterozoic, a component of the Ketyet River Group correlative with the basal quartzite of the Hurwitz Group (Rainbird et al., 2010). The Neoarchean and Paleoproterozoic sequences were together affected by four phases of deformation (Pehrsson et al., 2010), the earliest involving nappe tectonics (Jefferson et al., 2011a; Jefferson et al., 2011b).

2.1.4. Trans Hudson Orogeny

Following terminal collision of the Hearne domain with the Superior Province, an extensive mountain belt was formed with roots preserved as the Trans Hudson Orogen (THO; Aspler and Chiarenzelli, 1998). Tectonic escape structures and distal extension resulted in the rapid uplift and subsequent denudation to exposure granulites, such as the Kramanituar Complex, to the north of Baker Lake and Rankin Inlet (Sanborn-Barrie et al., 2001). In the western Churchill Province, magmatism continued for some time, in two forms. Sill-like bodies of late syn- to post-orogenic Hudson granitoid (*ca.* 1.85 Ga – 1.8 Ga) and contemporaneous ultrapotassic minette to bostonite dykes intruded at mid-crustal levels, locally generating Martell syenite plutons. At the same time in the intracontinental trans-extensional Baker Lake Basin, the ultrapotassic magmas erupted to form the Christopher Island Formation.

2.1.5. Dubawnt Supergroup

The sedimentary and volcanic components of the Dubawnt Supergroup (Gall et al., 1992) constitute three unconformity-bounded groups (Baker Lake, Wharton and Barrenslund) that were accommodated by three successive trans-extensional basins, overlapping both the Hearne and Rae domains of the western Churchill Province (Jefferson, pers. comm, 2012). The total depositional timeframe of the Dubawnt Supergroup is bracketed by the age of the oldest volcanic rocks in the Baker Lake Group (1.84 Ga; Rainbird et al., 2006) and the age of the youngest volcanic rocks in the

Barrensland Group (1.54 Ga; Chamberlain et al., 2010). The Baker Lake Group fills a series of intracontinental trans-extensional rift basins straddling the Snowbird Tectonic Group, termed the greater Baker Lake Basin (Rainbird et al., 2003). The Wharton Group unconformably overlies the Baker Lake Group and overlaps onto Archean basement in what has been termed the middle Dubawnt Supergroup (Gall et al., 1992). The Wharton Group comprises conglomerate and aeolian quartz arenite of the Amarook Formation, rhyolite and basalt (this study) and epiclastic rocks of the Pitz Formation, and covers a much wider area than the Baker Lake Group, extending west as far as the Beverly Lake (Figure 2.1) and Akiliniq Hills area, with evidence of mixed mafic and felsic magmas (LeCheminant et al., 1984) and south transecting Dubawnt Lake (Peterson, 2006). The upper Dubawnt Supergroup rocks constitute the Barrensland Group, a succession of feldspathic sandstone, conglomerate and mudstone of the Thelon Formation, overlain by potassic mafic Kuungmi Formation lavas and the Lookout Point Formation dolostone. The Thelon Formation partly overlaps the Baker Lake Basin but is mostly preserved to the west within the Thelon Basin. The Thelon Basin is inferred to be thermal sag in origin (Rainbird et al., 2003) but shows significant evidence of strike slip and dip slip faulting (Jefferson et al., 2011a; Jefferson et al., 2011b). Sedimentation is constrained by the age of diagenetic cementation by U-bearing fluorapatite at 1667 Ma (Davis et al., 2011).

2.1.6. Lower Dubawnt Supergroup

The Baker Lake Group comprises three formations. At the base, the South Channel Formation lies unconformably on earlier crystalline basement rocks and consists

of alluvial/fluvial immature basement derived conglomerates (Donaldson, 1965; Macey, 1973). The Kazan Formation is also considered to be fluvial, with a major conglomeratic marker bed containing volcanic clasts inferred to be sourced from the Christopher Island Formation extrusive event (Blake, 1980). The Christopher Island Formation represents the thickest section (2500 m) of the Dubawnt Supergroup (Peterson and Rainbird, 1990) and is characterized by subaerial to subaqueous lamprophyre flows and epiclastic rocks (Gall et al., 1992). Finally, the Kunwak Formation consists of massive alluvial debris flows and conglomerate composed of Christopher Island Formation and Archean detritus (Gall et al., 1992).

2.1.6.1. Christopher Island Formation and Dubawnt Minettes

The Christopher Island Formation volcanic rocks are considered the extrusive equivalent of the Dubawnt minettes (Peterson et al., 2002). The dykes are described as ultrapotassic, transitional minette-lamproites. The Christopher Island Formation flows are a bimodal succession recording alternating felsic-mafic-felsic event (Peterson et al., 1994). The lower felsic sequence is sandine-phryic, with phlogopite and clinopyroxene, whereas the upper felsic sequence is relatively crystal-poor. The Christopher Island Formation extrusive event is dominated by the mafic sequence of strongly porphyritic minettes with phenocrysts of phlogopite and clinopyroxene and microphenocrysts of apatite, and less commonly olivine (Peterson et al., 1994). Potassium feldspar, rarely present as a phenocryst phase, is predominant in the groundmass (LeCheminant et al., 1987). Amphibole, where present, is commonly K-Ti rich and included within potassium

feldspar oikocrysts (Peterson et al., 1994). Geochemically, the minette flows range in SiO₂ from 42.9 - 59.5 wt % and are typically Ca-poor, Mg-rich and generally enriched in large ion lithophile elements (LILE) such as K, Rb, Ba and Th. They are commonly depleted in high field strength elements (HFSE) such as Ti, Nb, and Ta (Peterson et al., 1994). The minette magmas are interpreted to have been derived from an upper lithospheric mantle previously depleted by basaltic melt extraction and subsequently enriched in LILE (Cousens et al., 2001).

2.1.6.2. Hudson Granitoid

The Christopher Island Formation was broadly contemporaneous with emplacement of the Hudson granitoid (Peterson et al., 2002). The ultrapotassic Dubawnt minette is locally mingled with the calc-alkaline Hudson granitoid intrusions locally producing syenitic plutons (Martell syenite; Peterson et al., 2010). The characteristics of the Hudson granitoid laccoliths and sills are consistent with minimum granitic melts of lower to middle crustal rocks (Peterson et al., 2002). Extrusive and hypabyssal facies equivalents of this suite have not been identified. The plutons lack chilled margins, are rich in xenocrysts such as inherited zircon, and inherited fabrics (Peterson et al., 2002). In many places the Hudson granite is coarsely equigranular, although potassium feldspar-phyric variants are locally documented. Accessory phases are typically biotite (< 5%) and magnetite, with less common hornblende, fluorite, pyrite and muscovite. Compositions cluster around a point in Qz-Plag-Or space consistent with water-saturated minimum melts at *ca.* 0.5 GPa or 15-20 km depth (Berman et al., 2000). The average Hudson

granite has low Y and high light rare earth (LREE) to heavy rare earth (HREE) ratios, with moderate to weak europium anomalies and markedly negative Ta-Nb-Ti anomalies (Peterson et al., 2002). Hudson initial Nd-isotopic compositions are nearly indistinguishable from wall rocks (van Breemen et al., 2005). Hudson intrusions usually have a prominent positive magnetic signature which is isotropic, but corrugated on a scale of 0.5 km, giving a rough knobby appearance to expanses of laccoliths (Jefferson, pers. comm., 2012). The genesis of these late syn- to post-orogenic emplacements is proposed to be the coincidence of radiogenic heating, thicker crust and thermal relaxation after the THO event (Peterson et al., 2002).

2.1.6.3. Martell Syenite

The ultrapotassic, mafic to felsic Martell syenite dykes (bostonite) and plutons broadly overlap in emplacement area with Dubawnt minette dykes and Hudson plutons. Irregular contacts are observed where the syenite is in contact with Hudson granitoid plutons and can display textures inferred to have developed by mixing (e.g., Sandeman et al., 2000) between Hudson granitic melts and contemporaneous Dubawnt minettes (LeCheminant et al., 1987; Peterson et al., 2010). Textures range from reddish microsyenite (e.g., Peterson et al., 2010) to pink, coarse-grained quartz syenite (e.g., Tella et al., 1985). It has been observed previously that a pseudo-hypabyssal texture is developed locally where Hudson granite has mingled with Dubawnt minette to produce an apparent porphyritic syenite in which partly digested potassium feldspar xenocrysts float in a syenitic matrix along with mafic clots (Jefferson, pers. comm., 2011).

Mineralogy generally consists of (perthitic) potassium feldspar and varying modal abundances of biotite, (sodic) amphibole, clinopyroxene, with rare olivine, and accessory minerals sphene, apatite, zircon, titanian garnet and fluorite (LeCheminant et al., 1987).

2.1.7. Middle Dubawnt Supergroup

The Wharton Group of the middle Dubawnt Supergroup is characterized by basal polymict conglomerate and aeolian quartz arenite of the Amarook Formation, succeeded by dominantly felsic subaerial volcanic rocks of the Pitz Formation (Gall et al., 1992). Pitz volcanic edifices are locally intruded by comagmatic Nueltin granite (Peterson et al., 2002). Field relationships suggest reworked Pitz rhyolite epiclastic rocks were also intruded by McRae Lake diabase dykes in the vicinity of Tebesjuak Lake, which are also traced into and comingle with Nueltin granite (e.g., Booth, 1983). In contrast to the THO scale Hudson granitoid plutons, Nueltin granite and Pitz rhyolite (*ca.* 1.75 Ga) are restricted to mainly the southern Kivalliq Region of Nunavut. Plutons of Nueltin granite extend SSW through a broad corridor from northwest of Baker Lake southward into northern Manitoba and Saskatchewan (Figure 2.1). Within the main Nueltin corridor, Hudson granitoids plutons are less common, indicating that the corridor may have down-faulted during or after Nueltin emplacement (Peterson et al., 2002). Moreover, the volcanic Pitz Formation rhyolite, at the north end of the corridor is mostly absent at the south end. where the granite plutons (i.e., in the type area at Nueltin Lake) are exceptionally coarse-grained to megacrystic. Therefore, the Nueltin corridor forms a tilted crustal block with shallower crustal levels exposed to the north (Peterson and

Pehrsson, 2010) inferred that tilting syn- or post-emplacement exposed successively shallower hypabyssal and volcanic facies toward the north, indicating that down-faulting in the corridor was significantly greater at its northern end.

2.1.7.1. Nueltin Granite

Anorogenic Nueltin granite plutons are typically elongate bodies, intruded between 1765 and 1750 Ma (van Breemen et al., 2005) into metamorphosed Archean to early Paleoproterozoic crustal and supracrustal rocks. Nueltin plutons are typically stocks, with sharp, well chilled contacts. In places they incorporate angular xenoliths from the local country rocks. Toward the northern end of the corridor, sub-volcanic stocks grade into hypabyssal granite, rhyolite breccia, and rhyolite flows (i.e. Pitz Formation). Localized mingling textures between Nueltin granite and dioritic rocks (i.e. McRae Lake diabase) are observed within the 700 km² circular, rapakivi-textured Pamiutug pluton (Booth, 1983; LeCheminant et al., 1979). Nueltin granite is most clearly distinguished from Hudson granite by the multi centimetre-scale zoned alkali feldspar phenocrysts. Plagioclase phenocrysts are subordinate, whereas the typically interstitial quartz also forms as centimetre sized spheroidal phenocrysts.

2.1.7.2. Pitz Rhyolite

The volcanic facies are typically feldspar-quartz phyric rhyolite (Blake, 1980; LeCheminant et al., 1987), although less silicic rhyodacite to dacite have been noted

(Biczok, 1996; Turner, 2002), and some flows can be fluorite and/or topaz bearing (LeCheminant et al., 1979). Less commonly the rhyolite outcrops as aphyric glassy lava flows or domes, and welded crystal-lithic tuff (LeCheminant et al., 1979; LeCheminant et al., 1987). Rhyolitic rocks commonly are dark red to maroon in colour due to finely disseminated hematite (Blake, 1980; Booth, 1983; LeCheminant et al., 1979; Rainbird and Davis, 2007). Dating of the Pitz rhyolite, by U-Pb zircon analysis, gave ages of 1757.6 ± 3.3 and 1753.7 ± 1.6 Ma (Rainbird et al., 2003). Some Pitz rhyolite samples contain microgabbro clots and fragments (LeCheminant et al., 1984; Peterson et al., 2002). These observations are consistent with the presence of a coeval basaltic magma, such as the McRae Lake diabase (LeCheminant et al., 1987), which is exposed in a series of dykes between Pamiutuq Lake and Mallery Lake.

2.1.7.3. McRae Lake Dyke

The McRae northeast-trending Lake diabase dyke was first identified by LeCheminant et al. (1979) at the south end of Tebesjuak Lake, where it intrudes Archean and Pitz Formation rocks. The dyke primarily comprises labradorite and clinopyroxene, with accessory biotite, amphibole, and magnetite-ilmenite (LeCheminant et al., 1979). The McRae Lake diabase is traced as several linear to en-echelon dykes beneath the northeast Thelon Basin and farther north, via distinct linear aeromagnetic highs that correspond to mapped outcrops at locations such as south of Aberdeen and Tebesjuak lakes, and further south at the Pamiutuq pluton. Baddeleyite from the McRae Lake dyke had a U-Pb date of 1750 ± 1 Ma (LeCheminant, unpublished data). The dyke is mingled

with Nueltin granite at the Pamiutuq pluton (Booth, 1983). The dyke also has negative Nd-isotopic values (Peterson et al., 2002) that overlap with the granite.

2.1.8. Upper Dubawnt Supergroup

The Barrenslund Group, at the top of the Dubawnt Supergroup, is preserved mainly within the Thelon basin and at the northern end of the Baker Lake sub-basin. It is thought to have had a broader original distribution (Rainbird and Davis, 2007), although Davis et al. (2011) noted that syn-depositional faulting influenced deposition and diagenesis. The Thelon Formation initiated unconformably overlying basement regolith comprising (> 2000 m) multiple upward-fining sequences of locally derived conglomerate, feldspathic sandstone and mudstone (Davis et al., 2011) unconformably overlying basement regolith and containing recycled zircon from the Baker Lake and Wharton groups (Rainbird and Davis, 2007). The Thelon Formation is relatively flat-lying and dominantly fluvial with some marine and aeolian facies (Rainbird and Davis, 2007). In the centre of the main Thelon Basin, two formations conformably overlie the Thelon Formation. The Kunngmi Formation, a thin unit (10 m) of amygdaloidal, sub-aerial basaltic flows dated at 1540 Ma by U-Pb in baddeleyite (Chamberlain et al., 2010) is topped by stromatolitic dolostone of the Lookout Point Formation (40 m; Gall et al., 1992).

2.2. Local Geology

This study examined mainly Nueltin granite rocks at four locations: Nueltin Lake, Mallery Lake, the Kiggavik uranium deposit and the Amer plutonic complex. The Nueltin Lake area shows plutonic facies of the Nueltin granite and represents the type location for Nueltin granite, therefore it represents an important study location for characterizing the granite. The Mallery Lake was previously investigated for significant Au-Ag mineralization within quartz stockworks. Local studies suggest these stockworks formed in association with a shallow level intrusive Nueltin granite. Areas south of Mallery Lake were explored for outcrops of basaltic dykes inferred coeval to Nueltin granite. The Kiggavik area, south of Schultz Lake, is the location of one of the most significant uranium deposits in the Thelon Basin area. Uranium mineralization is hosted in metasedimentary rocks and within fractures of the Lone Gull granite. The Lone Gull granite has been inferred Nueltin granite based on textural and mineralogical features. spatially related to the Lone Gull granite, interpreted as a Nueltin hypabyssal pluton. Finally, the rapakivi granite comprising the northern half of the Amer plutonic complex, represents the most northerly outcrop of Nueltin granite.

2.2.1. Nueltin Lake

.....The Nueltin Lake area, which straddles the Nunavut-Manitoba border, lies within the Southern Hearne

Figure 2.2). The area was extensively intruded by plutons of the Hudson granitoid suite (van Breemen et al., 2005) and Nueltin granite suite (Calder, 1970; Wanless and Eade, 1974). Further investigation by Gandhi (pers. comm., 2010) revealed uraniferous Hudson granodiorite in the area, described as massive to weakly foliated and tourmaline bearing. Additionally, significant abundances of coarse cordierite and sillimanite indicate the granodiorite is aluminum-rich.

Geochronology of inherited zircon, hosted by Hudson granite samples, and Sm-Nd model ages from both Hudson and Nueltin suites, indicate that crust in this part of the Southern Hearne is mainly of *ca.* 2.7 Ga vintage, with smaller *ca.* 3.0 Ga old domains (Peterson et al., 2002). Although contacts are obscured by cover, Nueltin granite is inferred to have intruded in contact with older Hudson granitoids bodies and locally, Nueltin granite apophyses are observed to intrude local arenite (Aspler et al., 2001) sandstone. Angular xenoliths (up to 30 cm) of similar sandstone, and visibly deformed metamorphic country rock were identifiable in the Nueltin plutons and are commonly aligned subparallel to weak magmatic alignment of feldspars. Nueltin plutons typically show little deformation, having been intruded well after the cessation of regional tectonism related to the THO, although Wanless and Eade (1974) note minor faulting and shearing along contacts due to emplacement processes.

Localized uraniferous mineralization in the Nueltin Lake area was previously described by Charbonneau and Swettenham (1986) and consists of uraninite and U-bearing silicates within a tourmaline-bearing granite-quartz monzonite and locally derived float of sulphide bearing calc-silicate assemblages; the latter is inferred to be pre-

Hurwitz. Also, two anomalous occurrences of U and Th were identified by Scott et al. (2012) hosted within a pegmatitic seam and an aplitic dyke located within Nuelin granite, which represent primary magmatic concentrations of U, Th, and REE mineralization.

2.2.2. Mallery Lake

The oldest rocks in the Mallery Lake study area are Neoarchean supracrustal rocks and granite of the inferred 2.6 Ga Snow Island intrusive suite (Figure 2.3). The supracrustal rocks are predominantly migmatitic schist and quartzofeldspathic paragneiss (Biczok, 1996). The Snow Island intrusive suite rocks here constitute a mafic to felsic complex: basal peridotite, biotite hornblende diorite, crosscutting megacrystic granite, and a late biotite leucogranite dyke (Peterson and Born, 1994). Archean units are exposed mainly along the western and northern edges of Mallery Lake. Additionally, Archean rocks are exposed between a plug of Nuelin granite and Christopher Island Formation rocks along the south shore of Mallery Lake. The rocks comprise strongly hematitized and gossanous gneiss, with lesser amphibolite and tonalite.

Two pulses of Paleoproterozoic igneous activity are recorded in the area. The first pulse is represented by rocks of the Christopher Island Formation, which are mapped (Figure 2.3) extending in a northeast trending belt covering a large area from south to east of Mallery Lake, unconformably overlying Archean assemblages (Biczok, 1996; LeCheminant et al., 1979). The Christopher Island Formation here is composed of minette lavas, pyroclastic rocks and volcanoclastic strata, variably overlain by arkose to

conglomerate of the Kunwak Formation (LeCheminant et al., 1981). The Kunwak Formation is commonly exposed conformably above the Christopher Island Formation, with Pitz Formation unconformably overlying (LeCheminant et al., 1981). The Pitz Formation dominates the map area from south to west of Mallery Lake (Figure 2.3). In this area it comprises flows, volcanoclastic and epiclastic strata, which unconformably overlie both the crystalline basement and Christopher Island Formation (Rainbird et al., 2003). The volcanic rocks range from basalt (this study) to rhyolite, and generally appear red to mauve coloured, with abundant feldspar and quartz phenocrysts. More rarely, flow-banded and sparsely porphyritic tuff containing fluorite and topaz is documented by LeCheminant et al. (1979) at Topaz Lake.

The plug of Nueltin granite at Mallery Lake occupies a 6 km wide circular peninsula along the south shore surrounded by a magnetic high (Figure 2.4). Nueltin granite is more abundant to the southwest of the map area, e.g. the Pamiutuq pluton. Numerous Au- and Ag-bearing quartz stockwork zones intrude the Pitz Formation and Nueltin granite around the Mallery Lake (Biczok, 1996; Turner et al., 2001). McRae Lake diabase dykes are located to the southwest and north of the map area and are roughly paralleled, 3.5 km to the southeast, by a porphyritic leucogabbro dyke. The leucogabbro is inferred to be contemporaneous with the diabase, and can also be traced by a magnetic anomaly signature to an outcrop of amygdaloidal basalt northeast of Ruby Lake (Jefferson, pers. comm., 2012). The leucogabbro is described as labrodorite-phyric with subordinate clinopyroxene rimmed by amphibole, magnetite-ilmenite and apatite (LeCheminant et al., 1979).

2.2.3. Kiggavik uranium deposit

The Kiggavik deposit is located in the Schultz Lake area, within the Rae domain of the western Churchill Province, at the eastern edge of the Thelon Basin (Figure 2.5). Locally, the gently dipping Woodburn Lake Group (*ca.* 2.7-2.8 Ga) Neoarchean meta-sedimentary and meta-volcanic rocks unconformably overlie Archean basement para- and ortho-gneisses (Zaleski et al., 2000). Highly resistive quartzite, of inferred early Paleoproterozoic age, outcrops directly north and west of the Kiggavik U deposit.

Numerous dykes, broadly contemporaneous with the Christopher Island Formation, are common in outcrop and abundantly expressed by high susceptibility aeromagnetic lineaments throughout the local area. Present to the west is a sparsely outcropping, composite pluton, the Schultz Lake intrusive complex, comprised of Hudson granite and Martell syenite. Based on outcrop and geomagnetic signature (Jefferson et al., 2011b), the complex is a southwest-northeast elongate (60 km), body some 15-20 km across. The outcrops observed in this study are generally consistent with initial descriptions by Miller and LeCheminant (1985) of the Hudson outcrops, as non-foliated, medium grained, sulphide- and fluorite-bearing subporphyritic to equigranular granite, locally cut by phogopite-phyric lamprophyre and syenitic (bostonite) dykes. This pluton is cut by the informally named Thelon fault (East Aberdeen fault), and offset by a *ca.* 25 km right lateral (Jefferson et al., 2011a) and *ca.* 550 m dip slip offsets (Rainbird et al., 2010).

Smaller bodies of Hudson granite occur between Kiggavik and the Schultz Lake intrusive complex, in an area informally named Granite Grid (Scott et al., in prep). These

smaller, equigranular granitic bodies are inferred to be part of the Hudson suite and include variably pegmatitic to aplitic dykelets lacking chilled margins. Some outcrops expose abundant pyrite, especially where cut by dykelets. Less than 15 km to the southeast of the Kiggavik deposit, the Nutaaq pluton of REE-rich Martell syenite cut Neoproterozoic metasediment rocks. At surface, immediately above the Kiggavik Main Zone deposit is a composite outcrop of medium-grained Hudson monzogranite, reddish Martell (micro-)syenite and a Hudson-aged porphyritic dyke that cuts local metasediments (Peterson et al., 2010).

Approximately 10 km to the north, all local units are abruptly truncated by the East Aberdeen and Thelon Faults, juxtaposing older units on the south side against downdropped upper DSG Thelon Formation on the north side. Fluorapatite cement initially described by Miller et al. (1989) has now been found from the basal unconformity to the upper sequence of the Thelon Formation, and dates mesodiagenesis throughout the Thelon Formation at *ca.* 1667 Ma (Davis et al., 2011). The fluorapatite cement is proposed to be sourced in part from phosphorous- and fluorine-rich units of the lower and middle Dubawnt Supergroup that underlie or had shed detritus into the Thelon Formation (Davis et al., 2011; Gall and Donaldson, 2006). The Pitz Formation rhyolite is mapped along the south and southwestern margins of the Thelon Basin in the Aberdeen Lake area (LeCheminant et al., 1984). Pitz Formation rhyolite is also inferred by petrographic analysis of drill core cutting the Andrew Lake deposit just 15 km southwest of Kiggavik (Peterson, pers. comm., 2010).

2.2.3.1. Lone Gull granite

Early work by Fuchs and Hilger (1989) define the Lone Gull granite as a small stock intruding local metasedimentary rocks. A detailed description of the Lone Gull granite was initially produced by Fuchs et al. (1986), indicating the assemblage as subporphyritic, sulphide-bearing (predominantly pyrite and molybdenite), fluorite granite. Weyer et al. (1987) subsequently outlined the modal mineralogy as containing abundant potassium feldspar and lesser amounts of biotite/chlorite and muscovite/sericite, with accessory titanite, apatite and zircon. Whole rock geochemistry of the Lone Gull granite indicated an overall geochemical signature were consistent with S-type granite (Weyer et al., 1987). Alteration assemblages were also explored by Fuchs et al. (1986), reporting that in the area of highly altered (and mineralized) Lone Gull granite had varying depletions in most major elements. Attempts to date the Lone Gull granite via the K-Ar system yielded an age range of 1648 to 1563 Ma, later interpreted as a reflection of age resetting (Fuchs et al., 1986).

2.2.4. Amer plutonic complex

The Amer plutonic complex straddles and is cut by the Amer Mylonite Zone (AMZ), some 50 km west of Amer Lake. The geology surrounding the Amer plutonic complex is dominated by Neoarchean granitic and supracrustal Woodburn Lake Group basement rocks, unconformably overlain by early Paleoproterozoic, dominantly sedimentary rocks of the Amer Group (Rainbird et al., 2010; Tella, 1994; Tella and

Heywood, 1978), subsequently incorporated into the Amer fold and thrust belt (e.g., Patterson, 1986). Archean rocks, mostly exposed directly to the northeast, include orthopyroxene-bearing felsic granulites, and foliated or layered tonalitic gneiss. Variably deformed (up to mylonitic) rocks of dioritic to granitic composition form the northwest and southeast margins of the belt. The Amer Belt trends ENE-WSW and shows increasing metamorphic grade toward the northeast (Patterson, 1986).

The Amer plutonic complex exposes both Hudson-aged (1.83 Ga), potassium feldspar-phyric, hornblende syenogranite and Nuelin-aged (1.75 Ga), fluorite-bearing rapakivi granite (Tella, 1994; Tella, unpublished). The plutons truncate early dextral ductile fabrics established along the AMZ, but were later affected by reactivation of the AMZ, expressed as cataclasis (Tella, 1994). The AMZ is a crustal scale, strike-slip fault that extends discontinuously for more than 130 km to the northeast across the Rae subdomain, with metamorphic grade increasing, toward the northeast into the lower greenschist facies (Tella et al., unpublished). The AMZ can be traced southwesterly beneath the Thelon Basin into the McDonald Fault system of the East Arm of Great Slave Lake (Jefferson et al., 2011a; Jefferson et al., 2011b).

The Hudson and Nuelin granite components are expressed as lobate plutons, corresponding to a strong aeromagnetic high. The aeromagnetic high associated with the younger and larger, Nuelin rapakivi granite pluton to the north, has relatively sharp margins. The aeromagnetic high associated with the older sub-porphyritic Hudson granite to the south has gradational margins. Smaller satellite plutons of Hudson granite through diorite and syenite are present southeast of the main plutonic complex. A tiny plug of

relatively fine-grained Nueltin granite is also present on the southeast side of the AMZ
(sample collected by C.W. Jefferson examined by the writer).

Figure 2.1. Map of the western Churchill Province highlighting Proterozoic igneous rocks. Areas of study are highlighted. Map modified from Peterson et al. (2002).

Figure 2.2. Nueltin Lake study area. Sample locations marked on inset, with anomalous U-Th samples labelled. The filled square indicates a "vein" occurrence containing Th and REE and the open square represents a polymetallic skarn deposit, both located by Charbonneau and Swettenham (1986).

Figure 2.3. Geology map of Mallery Lake area modified from LeCheminant et al. (1981). Outcrop sample locations of Pitz basalt are located north of Ruby Lake. Samples collected at the Mallery Lake peninsula are highlighted on the inset.

Figure 2.4. Close-up map of Mallery Lake peninsula: a) geological units and field stations with contacts inferred according to b) magnetic map.

Figure 2.5. Simplified geology map of the Kiggavik region, western NTS 66A and eastern 66B, from Jefferson et al. (GSC Open File, in prep.)

Figure 2.6. Cartoon cross-section of the Kiggavik deposit Main Zone, showing inferred relationship of uranium mineralization with the Lone Gull (Nueltin) granite.

Figure 2.7. Geology of Amer plutonic complex (after maps in (after maps in Jefferson et al., 2011b). Geological units and field stations (upper) and expression of geological units in total magnetic field lower after Tschirhart et al. (2011).

CHAPTER 3: METHODS

3.1. Field work

Hand-sized samples were collected from mapped outcrops at Nueltin Lake, Mallery Lake and Amer plutonic complex and from drill core of the Lone Gull granite at the Kiggavik U deposit. The use of a calibrated RS-230 portable multi-spectral spectrometer (manufactured by Radiation Solutions Inc.), in survey mode, was necessary to measure background radiation levels of Nueltin granite as well as locate any anomalous occurrences uranium and/or thorium. Radiation values are reported in counts per second (cps). Hand samples intended for further analysis were hammered from outcrop and further trimmed of weathered surfaces by rock hammer. When sampling extremely coarse-grained (heterogeneous) outcrops, enough material was collected to approximate a representative sample. Splits of each sample are archived at the Geological Survey of Canada (GSC), Ottawa.

3.2. Analytical methods

3.2.1. Petrography

Samples selected for analysis were initially cut into slabs and then into billets using wet saws. Polished thin sections were prepared by Vancouver Petrographics Ltd. and studied in detail in via transmitted and reflected light microscopy, scanning electron

microscope (SEM), utilizing backscatter electron (BSE) imaging and electron dispersion spectrometry (EDS), and electron microprobe (EMP) at the GSC and Carleton University, both in Ottawa. Autoradiograph films were used as a preliminary tool to locate radioelement bearing minerals in polished slabs. Exposure times were taken over a 3 month period, unless otherwise indicated, and developed at the GSC.

3.2.2. Whole Rock Geochemistry

Samples selected for geochemical analysis were cut into slabs using wet (water) saws. Weathered surfaces were further removed using a trim saw. The samples were crushed using a Braun Chipmunk and pulverized in a Rocklabs Ltd. shatterbox using a carbon steel mill head. Contamination from the preparation procedure was only considered significant in the case of fourth period transition metals (V, Cr, Mn, Co, Ni), which are typically found in very low quantities in granites (a full list can be obtained from Rocklabs Ltd. webpage). Major and trace elements were analyzed at Acme Analytical Laboratories (Vancouver) Ltd., using the 4X-4B whole rock lithogeochemical package with major elements determined by x-ray fluorescence (XRF) and trace elements determined by inductively coupled mass spectrometry (ICP-MS) after a lithium metaborate/tetraborate fusion and nitric acid digestion or Aqua Regia for metals.

Accuracy and precision of geochemical results for major elements was determined by comparing analytical results for the CANMET Diorite Gneiss reference standard SY-4(D) and a blind standard (99CR-7) of known composition. All results were within 95% confidence except for MnO on the SY-4(D) standard. An internal standard

(STD SO-18) of known composition was run in quadruplicate to assess precision and accuracy of trace elements, which were precise to within 95% confidence for all but Ba, Co, Nb, Rb, Sr, Zr, Y and Ce and accurate to within a 95% confidence interval except for Nb, Rb, Sr, Th, U, Ce, Sm. Furthermore, incomplete dissolution of zircon is known to affect the reporting of Zr when analyzed via ICP-MS. A comparison of Zr values determined by XRF and ICP-MS was conducted for six of the granite samples. It was found that the average discrepancy was 22 ppm (maximum discrepancy of 46 ppm) between values reported by ICP-MS and XRF, with latter method reporting higher ppm for five samples. The maximum represents a significant underreporting for Zr contents, even where typical values for granite are over 200 ppm. Additionally, the content of fluorine was run in quadruplicate against an internal standard (STSD-1), which showed good accuracy, but had poor precision within a 95% confidence interval. Geochemical data were plotted in GCDKit 3.0 (Janoušek et al., 2006).

3.2.3. Neodymium-Isotope Geochemistry

Isotopic results were for $^{143}\text{Nd}/^{144}\text{Nd}$ were determined at Carleton University using thermal ionization mass spectrometry (TIMS) via methods outlined by Cousens (1996). Whole-rock powders from ten of the samples were spiked (Table 4.2) with a mixed ^{148}Nd - ^{149}Sm spike prior to dissolution. Due to high Nd (ppm) concentration, two samples from the same pluton at Nueltin Lake were analyzed unspiked; the $^{147}\text{Sm}/^{144}\text{Nd}$

ratio was instead calculated using the Sm and Nd concentrations determined by ICP-MS.[†] A split of sample 09PHA-J06 was also run unspiked and had a difference of 0.000010, well within the uncertainty (0.000019), when compared to the spiked sample. The uncertainties in Sm and Nd concentrations are $\pm 1\%$, but $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are reproducible to better than 0.5%. A total of 26 runs of the La Jolla standard average $^{143}\text{Nd}/^{144}\text{Nd} = 0.511852 \pm 11$ (October 2011 - March 2012). Epsilon Nd (ϵNd^T) values (DePaolo, 1981) were calculated relative to a modern Chondrite Uniform Reservoir (CHUR) value of 0.512638 and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$, using an age of 1750 Ma. The precision of the ϵNd^T values are ± 0.5 epsilon units, based on duplicate analyses of geochemical standards and other rock samples. Depleted mantle model ages (T_{DM}) were calculated assuming a $^{147}\text{Sm}/^{144}\text{Nd}$ of 0.2140 and $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.513115 for modern depleted mantle.

3.2.4. U-Pb SHRIMP Geochronology

Large sample splits, taken from the hand samples crushed for geochemistry, were ground on a Rock Labs Inc. shatterbox using a tungsten carbide mill head at the GSC. To remove fine particles (i.e. possible contamination), the chips were sieved and decanted, then a mineral concentrate was obtained by heavy liquid techniques. The separates were subsequently sorted by magnetic susceptibility via FranzTM isodynamic separator. It was determined that due to the presence of complex, inherited cores and general poor zircon

[†] $^{147}\text{Sm}/^{144}\text{Nd} = (\text{Sm}/\text{Nd})/1.6606$

quality that use of the sensitive high-resolution ion microprobe (SHRIMP) II would yield the most meaningful results. The drill core samples presented a problem due to a volumetrically small rock sample size and thereby reducing the total amount of recovered zircon to select for analysis. Zircon were mounted on 2.5 cm epoxy disks along with fragments of the standard z6266 (559 Ma) and secondary standard z1242 (2679.6 Ma). Polished mount surfaces were coated with 10 nm of high purity Au. Zircon features (morphology, zoning, inclusions, and alteration) were characterized by reflected and transmitted light microscopy, backscattered electron (BSE) and cathodoluminescence (CL). The latter two techniques were conducted at the GSC, utilizing a Cambridge Instruments SEM. Selected areas of the zircon were sputtered using a mass-filtered O⁻ primary beam operating in Kohler mode. Corrections for common Pb were estimated from monitored ²⁰⁴Pb, and error are reported with 1 σ analytical errors. More detailed descriptions on the analytical procedure using a SHRIMP II at the GSC are after Stern (1997) with data processing using SQUID2 excel plug-in (Ludwig, 2000). Mean standard weighted deviates were calculated at the 2 σ level using the program Isoplot/Ex (Ludwig, 2000).

Zircon analyses were conducted over the course of multiple sessions. For the first session analytical spot diameter sizes were 10 μ m (K70) at 1.5 nA. Standard z1242 reported a concordia age of 2681 ± 9.4 Ma and the standard z6266 reported a 207/206 age of 514 ± 48 Ma. The second session was run at 2 nA with spot diameter sizes of 15 μ m (K100). The standards z1242 reported a concordia age of 2676.3 ± 5.8 Ma and standard z6266 reported a 207/206 age of during this session was 563 ± 27 Ma. The

general poor quality of the grains necessitated further evaluation of results for validity, especially in cases where strong negative discordance was recorded. The main factor generally attributed to strong reverse (negative) discordance is resultant from the presence of labile Pb or "parentless" radiogenic Pb, yielding an unusually low $^{204}\text{Pb}/^{206}\text{Pb}$ ratio (Mattinson et al., 1996; Wiedenbeck, 1995). The labile Pb may be contained in amorphous microdomains caused by diffusion and/or alpha-recoil implantation (Wiedenbeck, 1995). Calibration effects can result when these domains are analyzed. Stronger reverse discordance can result from the removal of metamict, high uranium zones by corrosive hydrothermal fluids. This should be considered in areas of known hydrothermal fluid movement, e.g. Kiggavik. If disturbance of the zircon is significant, it may present itself as an area of alteration of the zircon (observed in BSE or CL) along fractures or inclusions within the target spot (observed by transmitted light microscopy). The results are reported in Table 4.3, although the focus is on Proterozoic age dating. Archean ages are reported for discussion.

CHAPTER 4: RESULTS

4.1. Petrography

4.1.1. Nueltin Lake - Nueltin granite

The Nueltin granite, in the type location Nueltin Lake area, appear as outcrops of distinctly potassium feldspar megacrystic and rarer rapakivi-textured monzogranite (Figure 4.1). In outcrop potassium feldspar is predominantly perthitic white to pink microcline-albite, with interstitial coarse plagioclase, smoky quartz and coarse, black biotite. Typical size of potassium feldspar are 2-3 cm (in length), but up to 5 cm is not uncommon and megacrysts can reach over 7 cm (Figure 4.2). Potassium feldspar phenocrysts have coarse perthitic texture and are often poikilitic, concentrically enclosing euhedral biotite and quartz. Plagioclase, up to 3 cm in size, occurs as oligoclase to andesine and can have diffuse chemical zoning (Figure 4.2). Commonly plagioclase has undergone sericitization. More rarely resorbed phenocrysts are observed with inclusions of biotite and quartz (Figure 4.2). The biotite is Fe-rich (annite), often spatially associated with zircon and apatite, which also commonly occur as inclusions (Figure 4.2). Zircon typically occur as large (100-200 μm), prismatic, euhedral crystals, with faint (in BSE) discontinuous zoning. Dark purple fluorite also occurs with biotite and can reach sizes of nearly 0.5 cm. Hornblende (potassium-hastingsite) is common, though not widespread, as fine- and rarely medium-grained, anhedral crystals, typically partially replaced by biotite and quartz (Figure 4.2). Hornblende tends to be more common in centimetre-sized areas

of the sample where the coarse-grained nature has been disrupted yielding to a finer-grained assemblage and accompanied by minor magnetite. Titanite is very common as small euhedral wedges with few larger, subhedral crystals, reaching 1.6 mm. Allanite typically occurs as fine-grained (0.4 mm), subhedral crystals, displaying very fine zoning in cross-polarized light, with an often associated Ca-REE alteration product (Figure 4.2). Prismatic, euhedral zircon crystals were generally elongate (100-200 μm) and, in backscatter electron (BSE) imaging, showed faint discontinuous oscillatory zoning. The granite had a background radiation count of 200-300 cps. A REE-rich pegmatitic seam and a U-bearing aplite dyke of similar mineralogy to the host granite were located within a pluton.

4.1.1.1. Fine-grained aplite dyke

The sample from this location is a heterogeneous, weakly foliated, fine- to medium-grained, aplite dyke (Figure 4.3). The dyke is primarily quartz and potassium feldspar, with up to 30% biotite. Fine-grained crystals of purple fluorite were common, along with thorite and allanite as significant accessory phases. The thorite was typically weathered ruddy-brown, occurring as fine-grained (0.1-0.3 mm), anhedral crystals. The allanite had a sub-metallic luster and occurred as greenish, anhedral pools occurring randomly throughout the sample. The weak foliation is defined by the biotite and is discontinuous throughout the hand sample but appears to be most prevalent at the transition between the fine and medium-grained feldspar and quartz. Contact relationships were obscured by cover but from the inferred centre of the dyke grain size

appeared to grade outward from medium- to fine-grained. An *in situ* reading of up to 4000 cps was measured by the gamma-ray spectrometer for the dyke. In thin section, vermicular intergrowths of plagioclase and quartz was frequently observed as myrmekite. The composition of feldspars (as determined by EMP) in the aplite are identical to those of the host pluton, although some plagioclase is moderately sericitized. The primary source of radioactivity identified by autoradiograph (Figure 4.3) was predominantly attributed to 0.2-0.3 mm thorite crystals. Thorite ranged from heavily metamict and altered, to cores containing blebs of nearly pure thorium with marginally altered rims. Larger exposed areas on the autoradiograph corresponded to allanite crystals. Several allanite crystals, up to 3 mm in size, are visible in a polished slab. Adjacent minerals are noticeably damaged due to metamictization induced by the allanite, causing radial and concentric fracturing in the neighbouring grains. In thin section the allanite is dominantly yellow in plane polarized light and can be isotropic in crossed-polarized light, consistent with intense metamictization (Figure 4.3).

4.1.1.2. Pegmatitic granitic seam

This sample was taken from a 10 cm wide pegmatitic dyke hosted by Nueltin granite (Figure 4.1). The seam outcropped at surface for approximately 2 feet with a strike of 280°. Contact relationships with host granite were difficult to discern in outcrop due to cover. The outcrop sample is dominated by coarse potassium feldspar, quartz, biotite, allanite and dark purple fluorite. The seam gave an approximate *in situ* reading of 9000 cps. An autoradiograph (2 week exposure) of the polished slab for this sample

indicated radioelements are dominantly within the allanite crystals (Figure 4.4). These crystals, which often show well developed simple twinning, average 1 cm in length with a sub-metallic luster, greenish colouration and have a preferred orientation perpendicular to the inferred contacts (comb layering). Energy-dispersive X-ray spectroscopy (EDS) analysis determined that the primary host minerals of U and Th are subhedral to anhedral, internally heterogeneous urano-thorite crystals, which also contained minor quantities of Ce, Nd and Sm. EMP confirmed the sample contained LREE predominantly within subhedral allanite crystals. BSE images and EDS showed many allanite crystals have rims that are depleted in REE relative to the cores (Figure 4.4). The urano-thorite and allanite show a very strong spatial correlation, as observed by SEM, the former often observed as inclusions in the latter. Parallel to sub-parallel fracturing can be traced across the sample, unaffected by mineral boundaries, and are also observed radially from urano-thorite and could therefore be related to metamictization. Typically, fractures adjacent to allanite are in-filled by a secondary REE carbonate mineral, observed less commonly in the host Nueltn granite. Allanite was also observed in contact with an intergrowth symplectite of rutile and titanite which could indicate rapid cooling.

4.1.2. Mallery Lake - subvolcanic Nueltn granite

On the south shore of Mallery Lake locally topographic high, circular peninsula juts into the lake. Nueltn granite is exposed in small (1-5 m) mounds, across much of the peninsula (Figure 4.5). The presence of open miarolite cavities and granophyric texture would suggest low-pressure crystallization within subvolcanic facies. Outcrops and

felsenmeer of the unit are strongly porphyritic and coloured deep red to pink, occasionally exhibiting miarolitic cavities (Figure 4.5). Within the vicinity of several quartz stockwork zones, the granite can be fractured and infilled by drusy silica and fluorite (Figure 4.5). The fine-grained groundmass is comprised of about 80-90 vol. % of potassium feldspar and quartz, with the dominant phenocryst phase occurring as ovoid potassium feldspar, with lesser quartz and rare plagioclase (Figure 4.5). Potassium feldspar phenocryst size (0.5-1 cm) and abundance (10-40 vol. %) is variable between sample locations. Potassium feldspar phenocrysts are variably turbid, hematite-stained and altered to epidote and sericite. These are partially or wholly surrounded by rims, including embayments, of outwardly coarsening granophyric intergrowths (Figure 4.5). Quartz phenocrysts are fluid inclusion-rich and can also have granophyric textured rims. Plagioclase phenocrysts are discernible but often heavily sericitized. Mafic phases have been replaced by epidote, sericite, and blue chlorite. A few very small (< 0.2 mm) primary biotite crystals remain. Minor accessory titanite, anhedral to subhedral, and zircon, subhedral to euhedral, up to 0.2 mm typically occur spatially with secondary chlorite, epidote and opaques (Figure 4.5). Some granite outcrops in the southwest (i.e. 10PHA-J03) contain chloritized mafic blebs (up to 1 cm sized) within the groundmass and as inclusions in phenocrysts. These blebs can also contain minor pyrite and are reminiscent in appearance of mafic blebs in Pitz Formation rhyolite identified by Peterson et al. (2002). In the vicinity of silicic stockworks, the host granite has undergone brecciation and quartz veining with coarse drusy quartz and dark purple fluorite (single crystals > 1 cm) lined cavities (Figure 4.5). Miarolite cavities were also lined with coarse

drusy quartz and fluorite. Veins of adularia and fluorite are locally present at one of the southwest sample sites.

4.1.2.1. Mallery basalt

A mafic volcanic sequence of fine-grained rocks outcrop on the west side of the peninsula, at the shoreline, and a possibly related outcrop on the southeast flank, away from the shore, correlate with a strongly magnetic ring structure (Figure 2.4). Texturally the unit shows gently dipping, primary layering, indicative of volcanoclastic fallout (Figure 4.6). The layers vary in thickness on a scale of 1 to 10 mm, with contrasting thicker greenish (epidote-bearing) layers to thin white coloured laminae. The unit on the west flank, appears columnar jointed, and contains large, discontinuous lenses of intermediate, maroon weathered material, rich in 1 cm sized feldspar phenocrysts (Figure 4.6). Petrographic analysis of the mafic unit at this site shows strong recrystallization and alteration in the form of secondary epidote, Fe-rich chlorite, magnetite and calcite veins, obscuring much of the primary mineralogy (Figure 4.6). The unit appears rich in plagioclase, with accessory titanite. Relict textures may include pseudomorphs after clinopyroxene. Secondary magnetite is also very common, and likely accounts for the magnetic high. This unit could therefore represent a unit subsequently intruded by Nueltin granite. The sample from the southeast flank shows similar rocks, but more strongly altered and is cut by a granitic vein.

4.1.2.2. Pitz basalt

A helicopter traverse, 25 km south of Mallery Lake, was conducted to a series of outcrops northeast of Ruby Lake (Figure 2.3) and along trend with the porphyritic leucogabbro dyke identified by LeCheminant et al. (1979), which outcrops approximately 10 km to the southwest. The first outcrop was comprised of vesicular and amygdaloidal basalt, coloured mauve to red due to weathering. This unit shares textural similarities with a unit located south of Ruby Lake (LeCheminant et al., 1979). The exposure sampled was from a low relief, white to pink feldspar-phyc unit (Figure 4.7). Modal percentages of phenocrysts are, 15 - 20 % potassium and plagioclase feldspar (1 - 5 mm), 10 % Fe-rich chlorite (average 1 mm), 5 % bright green coloured epidote and lesser (< 2%) microcrystalline quartz; the latter three are inferred to be amygdules (Figure 4.7). Larger phenocrysts are often intergrowths of potassium feldspar and plagioclase, although solitary crystals of both are more common. Phenocrysts are anhedral, with resorbed boundaries and distinctive, but typically fine feldspar rims (Figure 4.7). The largest feldspar phenocrysts appear in loose glomerophyric clusters of resorbed crystals with incorporated matrix material between some of the crystals (Figure 4.7). Chlorite occurs as brilliant blue masses in cross-polarized light (xpl), with visible inward coarsening and abundant opaque minerals or microcrystalline quartz along the edge of the larger clusters. Chlorite also can wholly or partially rim epidote masses and, less commonly occurring fluorite amygdules. Highly birefringent epidote (Fe-rich?) occurs as typically radiating coarse crystals (Figure 4.7). Many of the larger amygdules appear aligned and/or drawn out, up to 3 cm, which could indicate a flow direction. Commonly

observed aggregate clusters of microphenocrystic apatite, opaques (mostly magnetite), epidote and chlorite are reminiscent of interstitial material described by LeCheminant et al. (1979) in the leucogabbro dyke to the south. The groundmass is comprised of microlitic plagioclase and potassium feldspar, with lesser stubby microphenocrysts of apatite, minor hematite staining and very fine disseminated magnetite.

A related second outcrop, 50 m to the north, overlooks the previous outcrop and correlates with a melanocratic basalt mapped by LeCheminant et al. (1979). The unit contains abundant amygdules and plagioclase phenocrysts, commonly in glomerophyric clusters, in a microlitic plagioclase groundmass. Plagioclase microlites are typically aligned and can show kinking. Observed in the previous outcrop, the unit contained similar but more abundant epidote filled amygdules, frequently with blue chlorite jacketing.

4.1.3. Kiggavik U-deposit - Lone Gull granite

Two drill core samples of Lone Gull granite were studied. The samples are from outside, but texturally continuous with, the granitoid that hosts U mineralization at the Kiggavik main zone.

Sample 08JP-52A was from a hole that intersects the Lone Gull granite at a depth of 75.5 m. This sample is dominantly a leucocratic porphyritic granitoid, possibly hypabyssal (Figure 4.8). A series of opaque, chlorite and fluorite lined fractures crosscut the entire sample. This sample displays seriate potassium feldspar (up to 1.2 cm) and less abundant plagioclase phenocrysts, both show fine potassium feldspar overgrowths

(Figure 4.8). Plagioclase phenocrysts tend to show similar potassium feldspar overgrowths, even those completely pseudomorphed by sericite, have evidence of relict rims. Phenocrysts of quartz (up to 7 mm) are rounded, embayed, and milky white due to the abundance of fluid inclusions. Biotite phenocrysts are completely altered to chlorite, but still display partially altered or secondary titanite inclusions and pleochroic haloes around zircon. Fluorite occurs as disseminated crystals and also lining quartz filled fractures. Apatite and titanite are common. Very minor quantities of allanite and thorite were found by SEM. The groundmass is fine-grained, predominantly quartz, potassium feldspar and sericite, with minor plagioclase. This specific piece of core was of interest since it hosts a 5 cm mafic inclusion. The ferromagnesian minerals within the inclusion have been completely chloritized, but relict textures suggest biotite with wedge-shaped inclusions of epidote-allanite. Opaque minerals such as pyrite are also common within the fragment. In thin section, zircon were not apparent in the mafic phase, but slivers of titanite and anhedral apatite are frequently observed, in both the mafic the granitic phases. The contact between the inclusion and granite host lacks any discernible chill margin and is generally lobate. The mafic inclusion appears to have incorporated a 0.4 cm quartz inclusion identical to those in the host granite, as well as smaller millimetre-scale (partially digested?) felsic clots are present. Smaller inclusions of mafic material are found throughout the piece of core (Figure 4.8). The textural evidence could indicate commingling while both units were still warm. Likely, as the Lone Gull granite magma ascended, it mechanically incorporated still warm, and hence likely coeval, mafic material.

The second sample (09PHA-J39) is from drill core intersecting the Lone Gull granite at a much shallower depth 17.3 m. The groundmass of this sample is more orange in colour due to intense hematization. Similar to the previous sample the mineral assemblage includes abundant phenocrysts of abundant potassium feldspar and milky white quartz, with lesser but more highly altered plagioclase in a quartzofeldspathic groundmass. All potassium feldspar phenocrysts appear to have rims of fine quartz inclusions (up to 0.2 mm width) and can have poikilitic inclusions of sericitized plagioclase (Figure 4.9). The sample contains up to 5% (modal) chlorite, which forms 1 cm clots, and along fractures throughout the sample. Distinguishable chlorite pseudomorphs (up to 1.5 mm) are after amphibole (Figure 4.9) and biotite, the latter can contain inclusions of titanite and zircon. Feldspar within the groundmass is dominantly potassium feldspar, with much less altered plagioclase. Few titanite wedges and zircon are present in the groundmass. Quartz phenocrysts exhibit mineral strain and feldspars intense sericitic alteration when adjacent to fractures.

4.1.4. Amer plutonic complex - rapakivi granite

The granite (herein referred to as the Amer plutonic complex) outcropping north of the AMZ in the Amer plutonic complex area exhibits a strong rapakivi texture (Figure 4.10). Ovoid (reaching over 4 cm) and tabular perthitic potassium feldspars, averaging 2 cm, range in colour from pink to white and are variably mantled by plagioclase rims up to 0.25 cm thickness (Figure 4.10). Interstitial plagioclase and quartz, typically not exceeding 1cm in size, are euhedral to subhedral, with myrmekitic texture frequently

observed in thin section. Plagioclase shows variation in extinction angles consistent with fine chemical zoning in thin section, with larger crystals commonly mildly sericitized (Figure 4.10). Biotite is the most common mafic mineral, and spatially associated with accessory titanite, allanite, fluorite, and magnetite. Titanite is very abundant and can occur as wedges up to 1 mm in length. Allanite, typically associated with titanite, is often zoned and occurs as much smaller crystals (100-60µm). Allanite is LREE enriched and has undergone some degree of alteration, leaving a Ca-REE alteration product.

Radioelement-bearing mineral phases are mostly absent, but rare thorite and urano-thorite blebs were found within allanite. Euhedral zircon were generally 100-200 µm in size and showed fine oscillatory zoning, occurring frequently with smaller, occasionally zoned apatite. Atypical of Nuelin granite, this unit contains abundant magnetite, which accounts for the large magnetic anomaly to which this pluton corresponds.

4.1.4.1. Amer plutonic complex - aplite dykes

Fine-grained, pink coloured, aplitic dykes, which range from several centimetres up to a metre in width, were observed at most outcrops to cut the rapakivi granite (Figure 4.10). Some aplite dykes were then cut by (0.5 cm) quartz veins. Contacts with the rapakivi granite are sharp, where the host granite is hematized, with sericitized plagioclase and biotite undergoing replacement by chlorite and Fe-oxides.

Figure 4.1. Field photos of Nueltin Lake granite outcrops: a) field photo of Nueltin granite showing weak igneous alignment (general direction oriented potassium feldspars); b) photo of Nueltin granite in shallow water showing numerous angular to sub-rounded xenoliths (up to 30 cm), aligned sub-parallel; c) apophysis of coarse Nueltin granite cutting sandstone; d) pegmatitic seam of primary U-, Th-, and REE-rich minerals within Nueltin granite.

Figure 4.2. Photos of slab sections and photomicrographs of Nueltin granite showing mineralogy: a) megacrysts (upper left is 5.5 cm and upper right is 4.5 cm) of potassium feldspar within Nueltin granite; b) polished slab of Nueltin granite showing typical mineralogy and coarse texture, with larger potassium feldspar crystals commonly oikocrystic and a resorbed plagioclase feldspar (at centre) with inclusions of quartz and biotite; c) anhedral green coloured K-hastingsite (amp), spatially associated with annitic biotite included by titanite and opaques (Fe-Ti-oxide), in plane polarized light (ppl); d) inclusion-rich coarse biotite, shows pleochroic halos around zircon and monazite, in cross polarized light (xpl); e) Carlsbad twinned plagioclase crystal showing fine polysynthetic twinning (lower twin) and diffuse chemical zoning (upper twin) with mymekitic texture; f) weakly zoned allanite (blue and yellow) adjacent to biotite and titanite, in xpl.

Figure 4.3. Photos of slab and thin section of aplite dyke (09PHA-J05) which was hosted by Nueltin granite, at Nueltin Lake: a) polished slab of aplite dyke showing heterogeneity and weak foliation defined by biotite with medium-grained greenish colour allanite; b) accompanying autoradiograph negative highlights radioelement bearing phases that coincide with allanite (larger) and thorite (smaller); c) rare earth-bearing allanite (al) is more coarse than that found in main phase of Nueltin granite with common U-bearing thorite (th).

Figure 4.4. Photos of slab sections from Nuelin granite hosted pegmatitic seam (09PHA-J06): a) polished slab of pegmatite seam showing coarse fluorite (fl) and allanite (greenish colour); b) autoradiograph film exposed over two weeks of the polished slab which highlights spatial association of radioelement bearing minerals, allanite (light gray) and urano-thorite (black); c) primary igneous allanite with visible twinning, adjacent to chloritized biotite (lower right) and microcline (upper left), with heavily metamictized, isotropic thorite (th); d) back scatter electron image of larger allanite from pegmatitic seam dominated by subhedral allanite (al) with contrasting core and rim, due to difference in rare earth element (REE) content and brighter inclusions consist of urano-thorite (U) and clots of REE-carbonates are in the brighter core.

Figure 4.5. Photos of samples from Mallery Lake area: a) field photo of low relief, areally small outcrop of subvolcanic Nueltin granite; b) polished slab of subvolcanic granite, hematite stained, rounded potassium feldspar phenocryst (some show concentric biotite inclusions) and chloritic mafic clots; c) photomicrograph of granophyric texture surrounding potassium feldspar phenocrysts in subvolcanic Nueltin granite; d) photomicrograph of Fe-rich chlorite, with inclusions of titanite (centre); e) field photo of quartz stockwork zone hosted by fractured subvolcanic Nueltin granite (scale in cm); f) sample from stockwork with vuggy quartz and coarse, dark purple fluorite cutting silicified rhyolite.

Figure 4.6. Photos of samples from Mallery Lake mafics: a) field photo of mafic bedded volcaniclastic rocks overlain by b) hematized feldspar-phyric felsic materials (scale card is 7 cm); c) polished slab showing primary bedding of tuff; d) photomicrograph of Fe-rich chlorite, with inclusions of grainy titanite and opaques (commonly magnetite).

Figure 4.7. Photos of samples from Pitz basalt (10PHA-J06) north of Ruby Lake: a) field photo of feldspar-phyric basalt with maroon weathered groundmass (scale card is 7 cm); b) polished slab showing aligned chloritic clots (appear black) and epidote (ep) amygdules; c) close-up showing larger phenocryst (upper right) incorporating matrix material with arrows indicating elongated chloritic masses; d) photomicrograph of epidote-filled amygdule, with Fe-rich chlorite rim and pseudomorph (indicate by arrow); e) resorbed plagioclase, with adjacent chlorite (arrows).

Figure 4.8. Photos from sample of Lone Gull granite (08JP-52A) at 75.5 m depth: a) polished slab showing typical porphyritic texture of Lone Gull granite with large mafic inclusion (right) and smaller mafic blebs (left); b) photomicrograph of Lone Gull granite (08JP-52A), arrows highlight potassium feldspar and plagioclase with overgrowths. Quartz phenocrysts (lower right) are highly rounded (resorbed?) and biotite (centre) is altered completely to chlorite. A typical wedge of common pale titanite is observed (top left).

Figure 4.9. Photomicrographs of the Lone Gull granite from depth 17.3 m: a) potassium feldspar with rim showing numerous quartz inclusions and poikilitic plagioclase; b) microphenocryst of chlorite pseudomorph after amphibole.

Figure 4.10. Rapakivi granite from Amer plutonic complex: a) field photo of rapakivi granite (scale card 7 cm); b) example of tabular potassium feldspar with well defined plagioclase rim; c) photomicrograph of plagioclase mantle over ragged potassium feldspar crystal boundary; d) photomicrograph of strongly zoned plagioclase (lower left) with euhedral titanite and weakly altered biotite; e) brittle deformation of the rapakivi granite (station 10PHA-J25) followed by sericitization; f) contact between late-crystallizing aplite dyke and strongly hematized host rapakivi granite.

4.2. Geochemistry

Major and trace elements can plot on discrimination diagrams and normalized trace element diagrams to identify relative geochemical patterns and affinities. Diagrams used include: Harker diagrams, total alkalis versus silica after Irvine and Baragar (1971), A/CNK versus SiO_2 , bulk continental-normalized trace element discrimination diagrams after (Taylor and McLennan, 1995) and rare earth element chondrite-normalized discrimination diagrams after (Nakamura, 1974). The usefulness of Harker diagrams is somewhat diminished by the high SiO_2 nature effectively diluting other major elements as the rock become fractionated towards higher silica contents. A caveat is also necessary when using total alkalis versus silica as Na and K are mobile under mild degrees of alteration. When profiling granitic rocks, continental crust normalization should provide the most useful information as granite is often derived from partial melting of crustal material. Trace element diagrams that are normalized to mid-ocean ridge basalt (MORB) is more apt for differentiating mafic rock tectonic affinities. The LILE elements are mobile, so the usefulness is diminished where alteration of the rock is apparent. Comparisons are more robust using relatively immobile HFSE. Additionally, REE are typically immobile under most conditions of alteration, and are therefore good indicators of the original sample. Mineral fractionation can often be inferred based on significant depletion (negative anomalies) in elements that have known affinities for specific minerals. Conversely, accumulation of minerals can also be resolved where significant positive anomalies occur. Geotectonic discrimination diagrams can also be produced by understanding typical element behaviour. Although, the use of these diagrams can be

more contentious as in nature there are always exceptions and unanticipated during the generation of a melt. Their weaknesses are addressed in the discussion section.

4.2.1. Nueltin Lake - Nueltin granite

Nueltin granite geochemistry is relatively homogeneous within plutons, considering the coarse-grained nature of these granites. Within each pluton, the weight percent of the major elements have a standard deviation of 0.7 or less, with the only exception of SiO_2 varying by more than 1 wt. %. Geochemistry begins to diverge when comparing between plutons. Overall, SiO_2 content is relatively low, averaging 66.7 wt. % for the north pluton and 70.4 wt. % for the south pluton. Harker diagrams all show strong, negative curvilinear trends for major elements as silica content increases, except for K_2O which shows a positive trend (Figure 4.11). The northernmost plutons is weakly metaluminous and the southern, more silicic, is peraluminous. Both are potassic with an average $\text{K}_2\text{O}/\text{Na}_2\text{O}$ value of 2.09.

When plotted on a bulk continental crust-normalized (Taylor and McLennan, 1995) trace element diagram, patterns are generally similar within plutons, but again contrast between the two plutons (Figure 4.12). Geochemistry is consistent with slight differences in modal mineralogy, i.e. elevated Th (and U), and total LREE in the southern pluton which has more abundant allanite relative to the northern pluton, which has more abundant amphibole and shows higher total HREE. Both plutons show strong negative anomalies for Sr and Ti relative to neighbouring elements, though stronger for the southernmost pluton. Overall, increases in average Th (ppm) are correlated with slight

increases in average U (ppm) content (). Nueltin granites are conspicuous for hosting high REE (especially LREE) enrichments. Chondrite-normalized (Nakamura, 1974) REE patterns have a moderate to steep negative trend, average $(La/Yb)_N = 21$ (Figure 4.12). The more southern pluton has the more marked LREE enrichment for all samples collected with average $(La - Nd = 830 \text{ ppm})$. The northern pluton has an average lower abundance of LREE $(La - Nd = 240 \text{ ppm})$, but higher HREE $(Er - Lu = 13 \text{ ppm})$, yet hosts highly localized, extremely REE enriched (up to 8 wt. %) seams and dykes. Eu-anomalies $(Eu/Eu^*)^\dagger$ are strongly negative for both plutons, 0.50 and 0.22, for north and south, respectively. Fluorine concentrations are consistent between plutons and average nearly 3000 ppm. The variations highlight the effect that differences in fractionation can affect the Nueltin granites, even at the local scale.

4.2.1.1. Fine-grained aplite dyke

The aplite dyke is characterized by a higher silica content (70 wt. %) and slightly lower values of CaO and Fe_2O_3 (total), relative to the host granite (northern pluton), and is weakly metaluminous. Overall, the sample has trace element abundances similar to the host granite (Figure 4.12), with a few exceptions. This sample has notably higher Zr (1062 ppm) and Hf (37 ppm). The anomalously high Th (770 ppm) and U (57 ppm) values (ratio of 13.5), are reflective of the high modal abundance of thorite. REE trends correlate with the host Nueltin granite, total LREE (366 ppm) and HREE (10 ppm), but

$^\dagger Eu/Eu^* = Eu_N / [Sm_N \times Gd_N]^{0.5}$

have a slightly higher $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratio of 16.3 (Figure 4.12). The negative Eu-anomaly (0.5) is similar to that of the host Nuelin pluton.

4.2.1.2. Pegmatitic granitic seam

In comparison to the host Nuelin granite (northern pluton), the seam has a much lower silica content (< 50 wt. %), high CaO content and a very low Na_2O content ($\text{CaO}/\text{Na}_2\text{O} = 10$) and higher Fe_2O_3 (total) content. The sample is strongly metaluminous. The trace element chemical signature of this sample strongly parallels that of Nuelin granite in general but relative to the host granite, element enrichments and depletions can be more pronounced (Figure 4.12). The most pronounced depletions, relative to the host granite are in phosphorus (80 ppm), and Zr (79 ppm) and Hf (3 ppm). Pronounced enrichments, relative to the host, are Th (8838 ppm) and U (610 ppm), giving a Th/U ratio of 14.4, compared to average Nuelin granite value of 11.9. This sample is strongly enriched in REE, as exemplified by Figure 4.12, with a concentration of approximately ΣREE 84500 ppm ($\Sigma \text{LREE} = 83460$ ppm; $\Sigma \text{HREE} = 168$ ppm) and $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratio of 208. The seam exhibits a strong negative europium anomaly value of approximately 0.1, more pronounced than that found in the host granite. Also, F concentration approaches 1% of the analyzed sample.

4.2.2. Mallery Lake - subvolcanic Nueltin granite

Samples of subvolcanic Nueltin granite from the Mallery Lake showed pervasive hematitic alteration. Two samples that showed the least alteration and are away from the quartz stockwork zones, were selected for analysis. The difference between samples for SiO_2 content is 2.6 wt. % but otherwise major element compositions vary only slightly. The outcrops are mildly metaluminous (Figure 4.13), potassic and strongly alkaline, average $\text{Na}_2\text{O} + \text{K}_2\text{O}$ is 8.93%.

Plotted bulk continental crust-normalized (Taylor and McLennan, 1995) trace elements show nearly identical patterns, with the exception of Sr and Ba. This confirms a genetic relationship and indicates little post-emplacement alteration between samples (Figure 4.13). Indicative of a higher degree of fractionation, relative depletions in Ba, Sr and Ti are significantly stronger than those observed in the granites. Chondrite-normalized (Nakamura, 1974) REE patterns showed a moderately negative trend for REE, the average $(\text{La}/\text{Yb})_{\text{N}}$ is 21.3, with a strong negative Eu-anomaly (0.19). The fluorine content of 2500 ppm is similar to the granites at Nueltin Lake.

4.2.2.1. Mallery basalt

The tuff located at the shoreline outcrop on the western flank of the Mallery Lake peninsula represents analysis of the most mafic rock to date associated with this suite. The sample is silica-saturated, SiO_2 content is low (46 wt. %). Concentrations of CaO (11 wt. %), MgO (7 wt. %) and total iron (11 wt. %) are consistent with plagioclase and

ferromagnesian mineral content. Relative to primitive mantle-derived basalt, MgO and Ni contents are slightly low, which likely indicates a loss of olivine and/or clinopyroxene, possibly through fractionation. The unit is also strongly metaluminous.

Plotting the Mallery basalt on an NMORB-normalized trace element diagram the basalt has an enriched MORB (EMORB) signature. When plotted on a bulk continental crust-normalized (Taylor and McLennan, 1995) trace element diagram the sample shows enrichments in Cs and Rb, consistent with apparent alteration and epidotization. At ten times greater than the chondrite-normalization (Nakamura, 1974), the REE abundances trend nearly flat $(La/Yb)_N = 1.3$, with a negligible Eu-anomaly of 0.96 (Figure 4.13).

4.2.2.2. Pitz basalt

The outcrop of Pitz basalt sampled at Ruby Lake has a relatively low SiO₂ content (55 wt. %). This basalt has the highest Al₂O₃ content for all rocks sampled. The CaO content is 2 wt. %, relatively low for the abundant plagioclase content and may be lost with alteration. The sample is metaluminous and alkali-rich ($Na_2O + K_2O = 10.41$).

Plotting the sample on a bulk continental crust-normalized (Taylor and McLennan, 1995) trace element diagram show trace element patterns similar to the Mallery Lake Nueltin granite, but LILE are relatively less abundant with the exception of an increase in Ba and a relatively smaller Sr-anomaly (Figure 4.13). Plotting the Pitz basalt on a NMORB-normalized trace element diagram, depletions in Nb and Ti, give the basalt a signature with affinities to derivation from a subduction zone. In the multi-element discrimination diagram by Nakamura (1974) REE patterns (Figure 4.13) are

negative and moderately steep $(\text{La/Yb})_N = 20.7$, very similar to the granite samples at Mallery Lake, but differ in that the negative Eu-anomaly was markedly weaker (0.74).

4.2.3. Kiggavik U-deposit - Lone Gull granite

The geochemistry of the two Lone Gull granite samples are very similar. Minor differences in major element abundances are consistent with a slightly higher degree of fractionation and/or a greater degree of alteration in the shallow sample (09PHA-J39). The slightly deeper sample (08JP-52A) has a lower SiO_2 content (70 wt. %), and marginally higher abundances of major elements, except for Al_2O_3 and K_2O . Both samples are peraluminous and strongly potassic.

Bulk continental crust-normalized trace elements between the samples are sub-parallel (Figure 4.15). Normalized plots show the two samples are similarly enriched in LILE, have strong negative anomalies in Ta and Nb, and are depleted with respect to Ti and P. Indicative of the closer proximity to the bulk of uranium mineralization, U content for 08JP-52A is somewhat elevated (12 ppm), resulting in a lower Th/U ratio (3.8). Chondrite-normalized REE concentrations show strongly fractionated La/Yb ratios for the Lone Gull granite (Figure 4.15). The less silicic, deeper sample (08JP-52A) shows stronger fractionation ($(\text{La}_N/\text{Yb}_N = 83)$), compared to the shallower 09PHA-J39 ($(\text{La}_N/\text{Yb}_N = 67)$). Eu-anomalies are similarly very weak for both samples (0.91 - 0.86).

4.2.4. Amer plutonic complex - rapakivi granite

The samples of rapakivi granite collected at the north pluton at Amer plutonic complex show a markedly homogeneous composition for the pluton. Major elements have a standard deviation of less than 1.0 wt %, except for SiO_2 content at 1.5 wt % ($n = 6$). Typically geochemical variation occurs in samples collected at the margin of the pluton relative to those collected at the interior, e.g. samples at the margin had a lower SiO_2 , slightly higher CaO, MgO and total iron ($\text{Fe}_2\text{O}_3 + \text{FeO}$). Harker diagrams show decreasing trends for major elements with increasing SiO_2 content, excluding K_2O , which increases. The granite is mildly metaluminous, has a potassic character with $\text{K}_2\text{O}/\text{Na}_2\text{O}$ values ranging from 1.74 - 2.28 (average 1.94) and is strongly alkaline ($\text{K}_2\text{O} + \text{Na}_2\text{O} = 8.36 - 8.92\%$).

Interplutonic trace elements appear to be somewhat variable, but form sub-parallel patterns when plotted normalized to average continental crust (Taylor and McLennan, 1995) in a trace element discrimination diagram (Figure 4.16). Similar magnitude depletions in elements Sr, Ta, Nb, P and Ti are apparent and could indicate mineral fractionation during melt evolution. A strong positive Th anomaly is significant, as Th values average 55.3 ppm and are similarly reflected by elevated U concentrations (average 9.9 ppm) with a Th/U ratios between (5.3 - 7.9). Rare earth elements show a mild enrichment, more apparent for LREE. Chondrite-normalized (Nakamura, 1974) REE patterns (Figure 4.16) have a moderate negative trend $(\text{La}/\text{Yb})_N$ averaging 21.8, with a distinct negative Eu-anomaly (0.48 - 0.66).

Fluorine was also elevated averaging approximately 1600 ppm. All samples from this pluton had detectable traces of gold (0.7 - 3.3 ppb).

4.2.4.1. Aplite dykes

The aplite dykes represent strongly fractionated, silica-rich intrusive rocks. Relative to the host rapakivi granite, the aplite dykes have similar continental crust-normalized (Taylor and McLennan, 1995) patterns for LILE, while HFSE and REE are markedly different (Figure 4.16). Patterns for two of the dykes (10PHA-J26A and 10PHA-J33B) appear nearly parallel, while the third (10PHA-J24A) contrasts, especially for Sr, HFSE and REE. The two similar dykes have a higher chondrite-normalized (Nakamura, 1974) content of HREE relative to LREE (Figure 4.16). Increasing values toward the heaviest REE (average $Gd_N/Yb_N = 0.9$) and a negative (0.4 - 0.5) Eu-anomaly, giving a 'gull-wing' shaped REE pattern. The contrasting dyke (10PHA-J24A) has higher normalized LREE than HREE. Similar to the previous two dykes, values increase towards the heaviest REE and a smaller (0.6) Eu-anomaly, give the dyke a more 'bowl-shaped' shaped REE pattern.

Figure 4.11. Selected Harker diagrams for all Nueltin granite and related rocks. Data from previous work from Peterson et al. (2002) and references therein.

Figure 4.12. Trace element geochemistry diagrams for granitic rocks from Nueltin Lake: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974).

Table 4.1. List of Th to U ratios for units from this study.

Sample	Intrusion	Th	U	Th/U
10PHA-J09	Mallery Lake granite	25.8	3.8	6.8
10PHA-J08	Mallery Lake granite	26.4	4.4	6.0
10PHA-J06A*	Pitz basalt	14.0	2.7	5.3
10PHA-J07A2	Mallery basalt	0.2	0.1	2.0
09PHA-J03*	Nueltin Lake granite	20.9	5.2	4.0
09PHA-J04	Nueltin Lake granite	15.5	5.3	2.9
09PHA-J08	Nueltin Lake granite	23.4	5.7	4.1
09PHA-J10*	Nueltin Lake granite	41.3	4.6	9.0
09PHA-J11	Nueltin Lake granite	126.2	6.5	19.4
09PHA-J13	Nueltin Lake granite	109.5	9.2	11.9
09PHA-J14	Nueltin Lake granite	116.7	6.4	18.2
09PHA-J18*	Nueltin Lake granite	189.7	6.1	31.3
09PHA-J21	Nueltin Lake granite	123.6	4.7	26.3
09PHA-J22	Nueltin Lake granite	100.6	6.3	16.0
09PHA-J23*	Nueltin Lake granite	99.6	11.8	8.5
09PHA-J39	Lone Gull granite	49.9	7.0	7.1
08JP-52A	Lone Gull granite	46.5	12.1	3.8
10PHA-J24B	Amer plutonic complex	26.2	4.6	5.7
10PHA-J25A	Amer plutonic complex	33.7	6.4	5.3
10PHA-J28	Amer plutonic complex	53.4	10.0	5.3
10PHA-J29	Amer plutonic complex	72.1	15.2	4.7
10PHA-J33A	Amer plutonic complex	92.9	16.2	5.7
82TX-S188	Amer plutonic complex	53.6	6.8	7.9
10PHA-J24A*	Amer plutonic complex aplite	56.6	11.1	5.1
10PHA-J26A	Amer plutonic complex aplite	21.6	11.8	1.8
10PHA-J33B	Amer plutonic complex aplite	22.3	11.2	2.0
09PHA-J05	Nueltin dyke	787.7	56.8	13.9
09PHA-J06	Nueltin dyke	8461.4	589.3	14.4

Figure 4.13. Trace element geochemistry diagrams for rocks from Mallery Lake: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974).

4.14. Trace element geochemistry NMORB-normalized after (Sun and McDonough, 1989) diagram for mafic rocks from Mallery Lake. The Mallery basalt has trace element patterns consistent with an enriched MORB normal mid-ocean ridge basalt (NMORB). This contrasts strongly with the Pitz basalt, which has slight negative anomalies in Nb and Ti, characteristic of subduction related basalts.

Figure 4.15. Trace element geochemistry diagrams for granitic rocks from Kiggavik: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974).

Figure 4.16. Trace element geochemistry diagrams for granitic rocks from Amer plutonic complex: a) bulk continental crust-normalized after Taylor and McLennan (1995); b) chondrite-normalized rare earth element diagram after Nakamura (1974).

4.3. Neodymium-isotope Chemistry

Results for isotopic analysis of Paleoproterozoic rocks are presented in Table 4.2. Data for all samples can be compared by calculating initial $^{143}\text{Nd}/^{144}\text{Nd}$ at the age of crystallization (Figure 4.17). U-Pb zircon age dating of the Nueltin granite and Pitz Formation rhyolite has well constrained emplacement from 1750 to 1765 Ma, the inferred peak of activity is shortly after Nueltin magmatic event initiated (van Breemen et al., 2005). At 1750 Ma, the $\epsilon_{\text{Nd},1750 \text{ Ma}}$ values for the granitoids range from -4.8 to -9.7, except for the Mallery basalt sample at +2.4. Within the granitoid samples, the value for the Lone Gull granite is the lowest at -9.7. $\epsilon_{\text{Nd},1750 \text{ Ma}}$ values determined for the Amer rapakivi pluton are also quite low at -9.2 and -9.4. The highest values are from the southernmost Nueltin Lake pluton are -4.8 and -5.2, which differ considerably from the adjacent Nueltin Lake granite, northern pluton at -7.1 and -7.3. The latter pluton hosts the U and Th-bearing aplite dyke (09PHA-J05) with $\epsilon_{\text{Nd},1750 \text{ Ma}} = -6.6$ and a REE-rich pegmatitic seam (09PHA-J06) with $\epsilon_{\text{Nd},1750 \text{ Ma}} = -7.6$. The Pitz basalt, is crustally contaminated as it has a $\epsilon_{\text{Nd},1750 \text{ Ma}} = -6.7$, virtually indistinguishable from the felsic Nueltin units. A simple mixing calculation (Faure, 1986) between assumed endmembers, Mallery basalt ($\epsilon_{\text{Nd}} = 2.4$ and $\text{Nd} = 7.1 \text{ ppm}$) and an average ($n = 11$) of Hudson granitoid (as a proxy for crust, $\epsilon_{\text{Nd}} = -8.8$ and $\text{Nd} = 50 \text{ ppm}$) show that approximately a 25 to 30 % assimilation of Mallery and Hudson granitoid could form a rock with an ϵ_{Nd} (-6.7) that approximates Pitz basalt.

Depleted mantle model ages (T_{DM}) were calculated for the basaltic samples, 10PHA-J06 and 10PHA-J07A2, using the formulation of DePaolo (1981). The Mallery

basalt also has an older, although slightly contaminated T_{DM} age at 2.6 Ga and the Pitz basalt at 2.5 Ga. The T_{DM} values indicate Archean source materials provided much of the Nd in the samples. The T_{DM} ages represent minimum ages of the source rocks, since fractionation of Sm and Nd during melting tends to produce model ages that are younger than the crystallization age of the magma.

Table 4.2. Sm and Nd isotopic analyses from Nueltin suite rocks.

Sample Name	Locale	Unit	Nd	Sm
09PHA-J03	Nueltin Lake	Nueltin granite	51.24	11.92
09PHA-J05	Nueltin Lake	Nueltin granite	62.69	10.08
09PHA-J06	Nueltin Lake	Nueltin granite	15913.42	1839.14
09PHA-J10	Nueltin Lake	Nueltin granite	80.48	15.48
09PHA-J18	Nueltin Lake	Nueltin granite	173.80	21.96
09PHA-J23	Nueltin Lake	Nueltin granite	185.30	24.23
10PHA-J06	Ruby Lake	Pitz basalt	85.78	14.51
10PHA-J09	Mallery Lake	Nueltin granite	81.95	13.26
10PHA-J24B	Amer plutonic complex	Nueltin granite	65.95	10.36
10PHA-J33A	Amer plutonic complex	Nueltin granite	60.45	7.90
08JP-52A	Kiggavik	Nueltin granite	47.09	6.59
10PHA-J07A2	Mallery Lake	Mallery basalt	7.07	2.29

Table 3.1 (cont'd)

Sample Name	$^{147}\text{Sm}/^{144}\text{Nd}$	1σ	$^{143}\text{Nd}/^{144}\text{Nd}$	1σ	$\text{Nd}(\text{T})_{1750\text{Ma}}$	TNd,DM
09PHA-J03	0.1406	0.00141	0.511619	0.00001	-7.3	3085
09PHA-J05	0.0972	0.00097	0.511155	0.00001	-6.6	2545
09PHA-J06	0.0699	0.00070	0.510793	0.00001	-7.6	2444
09PHA-J10	0.1163	0.00116	0.511351	0.00001	-7.1	2737
09PHA-J18	0.0761	0.00076	0.511005	0.00001	-4.8	2279
09PHA-J23	0.0787	0.00079	0.511102	0.00001	-5.2	2309
10PHA-J06	0.1023	0.00102	0.511209	0.00001	-6.7	2587
10PHA-J09	0.0978	0.00098	0.511105	0.00001	-7.7	2622
10PHA-J24B	0.0950	0.00095	0.510988	0.00001	-9.4	2708
10PHA-J33A	0.0790	0.00079	0.510815	0.00001	-9.2	2583
08JP-52A	0.0846	0.00085	0.510851	0.00001	-9.7	2652
10PHA-J07A2	0.1957	0.00196	0.512747	0.00001	+2.4	3039

Figure 4.17. Diagram showing $\epsilon_{Nd}(t)$ versus Age (Ga) for Nueltin granites and related rocks, relative to depleted mantle (DM) and chondritic uniform reservoir (CHUR). Depleted mantle (DM) evolution curve is calculated from $\epsilon_{Nd}(t) = 2.37 - 9.38(t)$ where (t) is the age (DePaolo, 1981). Symbols are the same as Figure 4.16, with the addition of the Mallery Lake basalt (10PHA-J07A2) indicated by the open circle.

4.4. Uranium-lead SHRIMP geochronology

4.4.1. Lone Gull granite - 08JP-52A

Due to the small volume of sample after heavy liquid separation, all zircon present were picked from the recovered material. It was observed that the zircon were of moderate to poor quality, as they showed a high degree of fracturing and alteration in BSE. Sizes range from 50-150 μm and crystals typically have a 1:2 aspect ratio. Two conspicuously large, equant zircon have morphologies that are also present in the shallower sample of Lone Gull granite (Figure 4.18).

Analysis of eight zircon, from areas where oscillatory zoning was present to absent, resulted in a concordia age of 1822.3 ± 11 Ma probability of concordance (POC) = 0.30. The weighted average of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages is 1820 ± 12 Ma, mean standard weighted deviates (MSWD) = 1.8, which is consistent with the concordia age and therefore inferred to represent the age of crystallization (Figure 4.19).

Paleoproterozoic aged analyses show a limited Th/U ratio (0.28 - 0.63), with the exception of two zircon. These zircon, 33.1 and 34.1, had conspicuously high Th/U ratios of 2.24 and 3.39 respectively, which was due to high Th contents (1822 ppm and 2229 ppm). Both analyses were over unzoned cores (in BSE), and were proximal to minor inclusions visible in transmitted light. The ^{204}Pb concentration for these analyses is higher, resulting in a slight negative discordance and could indicate isotopic disturbance, possibly associated with the inclusions. Alternatively, in the absence of apparent recrystallization features, these two zircon may represent a discretely different population

of possibly inherited Paleoproterozoic zircon. Core versus rim age differences on zircon that showed overgrowths could not be established, as rims present on inherited zircon were too fine to analyze. Also, fluorite within fractures may indicate that fluids carrying F interacted with the sample which may have attacked zircon, thus contributing to the poor quality.

The presence of Archean inheritance is inferred from two discordant analyses that had low U (ppm) and Th (ppm) values, but gave $^{207}\text{Pb}/^{206}\text{Pb}$ ages that would suggest they are Neoarchean.

4.4.2. Lone Gull granite - 09PHA-J39

For this sample zircon picked from the least magnetic fraction was mounted for analysis. Similar to the previous Lone Gull sample, zircon morphology is highly varied, though aspect ratios rarely exceeded 1:2. Sizes are generally below 180 μm length. The largest is a morphologically conspicuous, highly resorbed, nearly equant zircon at 200 μm . Zircon quality appeared poor, due to fractures, inclusions and a high degree of alteration, with some apparent recrystallization textures (Figure 4.18).

Analysis revealed two discrete populations of zircon based on Th/U ratios (Figure 4.19). A weighted $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1806 ± 41 Ma (MSWD = 5.6) was determined for the population of zircon that had high Th/U ratios (1.63-2.16). These analyses took place over mostly unzoned or partially zoned (in BSE) areas within the zircon, which lack any obvious recrystallization textures. The high MSWD may indicate more than one population present within this set, a younger (*ca.* 1790 Ma, $n = 2$) and an older (*ca.* 1840

Ma, $n = 3$). The population of four zircon with lower Th/U ratios (0.51-0.72) were analyzed over areas of zoning to partial zoning, and revealed a weighted $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1748 ± 9.4 Ma (MSWD = 0.53). Three of the younger zircon showed negative discordance (Figure 4.19), but did not show any textures that would suggest recrystallization or isotopic disturbance. It was noted that analytical targets that had hit areas of disturbed textures within the zircon recorded high ^{204}Pb and typically showed strong negative discordance, which could cause matrix effects. Inherited Archean zircon were present, and appeared as diffusely zoned cores, with fine rims. Two analyses of the cores yielded ages *ca.* 2575 and 2650 Ma.

Residual material from the original hand sample was processed to test the reproducibility of the bimodal population result. These zircon were similarly of poor quality and six (of 12) had high ^{204}Pb (> 0.8 counts per second; Figure 4.19). A bimodal population of ages (*ca.* 1750 and 1830) was again determined, but did not correlated to discrete Th/U ratios as before. All younger age analyses (i.e., Nueltin-aged) showed high ^{204}Pb and consequently strong discordance (commonly negative). The remaining six analyses had a concordia age of 1836.6 ± 14 Ma (POC = 0.84), which overlaps with the calculated weighted $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1836 ± 22 Ma (MSWD = 4.0). As with the previous session, the high MSWD could indicate more than one population within this set, a younger group of zircon (*ca.* 1807; $n = 3$) and an older group (*ca.* 1847; $n = 3$), but otherwise cannot not be distinguished based on Th/U ratios or morphological differences.

Table 4.3. U-Pb SHRIMP isotopic analyses from Lone Gull granite.

IP#	Session	Spot Name	U	Th	Th	Pb*	²⁰⁴ Pb	± ²⁰⁴ Pb
			(ppm)	(ppm)	U	(ppm)	²⁰⁶ Pb	²⁰⁶ Pb
08JP-52A, Lone Gull granite								
565	1	10068-21.1	506	307	0.63	141	0.00018	22.244772
		10068-22.2	723	263	0.38	202	0.00005	32.648447
		10068-27.1	678	186	0.28	189	0.00005	34.30687
		10068-29.1	841	408	0.50	241	0.00001	316.41323
		10068-33.1	840	1822	2.24	242	0.00013	26.004712
		10068-34.1	701	2229	3.29	202	0.00019	17.74683
		10068-37.1	494	245	0.51	137	0.00011	27.893519
		10068-37.2	277	90	0.34	77	0.00000	33.90830
Inherited Zircon								
	10068-25.1	103	70	0.70	44	0.00015	39.532644	
	10068-40.1	258	125	0.50	86	0.00024	30.327758	
09PHA-J39, Lone Gull granite								
548	2	10064-7.1	642	1087	1.75	175	0.00008	33.800568
		10064-9.1	763	1207	1.63	215	0.00010	25.211125
		10064-15.1	736	512	0.72	199	0.00020	15.950948
		10064-29.1	434	692	1.65	121	0.00021	26.943547
		10064-35.1	248	484	2.02	68	0.00049	30.310709
		10064-42.1	700	348	0.51	200	0.00013	26.279069
		10064-50.1	514	1073	2.16	145	0.00020	26.864396
		10064-54.1	953	580	0.63	252	0.00011	22.282566
	10064-79.1	795	440	0.57	228	0.00012	19.092986	
Inherited Zircon								
	10064-39.1	224	82	0.38	96	0.00026	26.767542	
	10064-70.1	538	340	0.65	216	0.00029	15.13715	
607	rerun 3	10064-3.1	321	522	1.68	91	0.00013	25.693715
		10064-4.1	513	194	0.39	148	0.00010	29.939571
		10064-8.1	418	575	1.42	117	0.00014	22.01176
		10064-12.1	314	740	2.43	89	0.00002	30.535092
		10064-13.1	172	224	1.34	48	0.00039	27.820708
		10064-14.1	609	102	0.17	173	0.00001	26.669406

Notes (see Stern, 1997):

Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number.

Multiple analyses in an individual spot are labelled as x-y.z.z

Uncertainties reported at 1 σ (absolute) and are calculated by numerical propagation of all known sources of error

$f(206)^{204}$ refers to mole fraction of total ^{206}Pb that is due to common Pb, calculated using the ^{204}Pb -method

common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)

* refers to radiogenic Pb (corrected for common Pb)

Discordance relative to origin = $100 * ((207/206 \text{ age} - 206/238 \text{ age}) / (207\text{Pb}/206\text{Pb age}))$

Error in $^{206}\text{Pb}/^{238}\text{U}$ calibration 1%

Th/U calibration: $F = 0.03900 * \text{UO} + 0.85600$

Cont'd.

IP#	Session	Spot Name	$f(206)^{204}$	$\frac{^{208}\text{Pb}}{^{206}\text{Pb}}$	$\pm \frac{^{208}\text{Pb}}{^{206}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\pm \frac{^{207}\text{Pb}}{^{235}\text{U}}$
08JP-52A, Lone Gull granite							
565	1	10068-21.1	0.00058	0.19034	2.07316	4.94448	1.37145
		10068-22.2	0.00058	0.10975	2.33844	4.94467	1.39277
		10068-27.1	0.00058	0.08405	2.53955	4.95491	1.41883
		10068-29.1	0.00058	0.14163	1.96312	5.14675	1.21062
		10068-33.1	0.00058	0.66160	0.92865	5.14754	1.28192
		10068-34.1	0.00058	0.96897	0.82783	5.18512	1.66949
		10068-37.1	0.00058	0.15443	2.35329	4.92312	1.35669
		10068-37.2	0.00058	0.10708	3.80255	5.05172	1.56016
	Inherited Zircon						
		10068-25.1	0.00058	0.19506	3.96071	12.80899	2.17267
		10068-40.1	0.00058	0.10877	3.48193	8.76960	1.55128
09PHA- J39, Lone Gull granite							
548	2	10064-7.1	0.00058	0.49775	1.32892	4.80720	1.48896
		10064-9.1	0.00058	0.49823	1.22335	4.91377	1.41552
		10064-15.1	0.00058	0.22756	1.51200	4.62791	1.43474
		10064-29.1	0.00058	0.49517	1.72169	5.02724	1.63509
		10064-35.1	0.00058	0.59100	2.18674	4.93697	2.49660
		10064-42.1	0.00058	0.15824	2.31915	4.92759	1.46636
		10064-50.1	0.00058	0.66276	1.41407	5.15587	1.60033
		10064-54.1	0.00058	0.18324	1.81529	4.53830	1.39250
		10064-79.1	0.00058	0.17288	1.59208	4.93505	1.39968
	Inherited Zircon						
		10064-39.1	0.00058	0.10106	4.35793	12.41563	1.65536
		10064-70.1	0.00058	0.19712	1.88566	11.08185	1.44919
607	rerun 3	10064-3.1	0.00058	0.48025	2.47521	4.98792	1.29701
		10064-4.1	0.00058	0.12100	2.51786	5.22616	1.20946
		10064-8.1	0.00058	0.41718	1.51774	4.98035	1.23550
		10064-12.1	0.00058	0.72957	1.33055	5.10462	1.28826
		10064-13.1	0.00058	0.40903	3.35349	4.92660	1.93348
		10064-14.1	0.00058	0.05286	3.43001	5.13834	1.09991

Notes (see Stern, 1997):

Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number.

Multiple analyses in an individual spot are labelled as x-y.z.z

Uncertainties reported at 1 σ (absolute) and are calculated by numerical propagation of all known sources of error $f(206)^{204}$ refers to mole fraction of total ^{206}Pb that is due to common Pb, calculated using the ^{204}Pb -method

common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)

* refers to radiogenic Pb (corrected for common Pb)

Discordance relative to origin = $100 * ((^{207}/^{206} \text{ age} - ^{206}/^{238} \text{ age}) / (^{207}\text{Pb}/^{206}\text{Pb} \text{ age}))$ Error in $^{206}\text{Pb}/^{238}\text{U}$ calibration 1%Th/U calibration: $F = 0.03900 * \text{UO} + 0.85600$

Cont'd.

IP#	Session	Spot Name	^{206}Pb ^{238}U	$\pm^{206}\text{Pb}$ ^{238}U	Corr Coeff	^{207}Pb ^{206}Pb	$\pm^{207}\text{Pb}$ ^{206}Pb
08JP-52A, Lone Gull granite	1	10068-21.1	0.32373	1.17088	0.85	0.11077	0.71
		10068-22.2	0.32602	1.30630	0.94	0.11000	0.48
		10068-27.1	0.32431	1.21771	0.86	0.11081	0.73
		10068-29.1	0.33315	1.11820	0.92	0.11204	0.46
		10068-33.1	0.33562	1.13628	0.89	0.11124	0.59
		10068-34.1	0.33628	1.55193	0.93	0.11183	0.62
		10068-37.1	0.32219	1.19418	0.88	0.11082	0.64
		10068-37.2	0.32499	1.37973	0.88	0.11274	0.73
	Inherited Zircon						
		10068-25.1	0.50321	1.62089	0.75	0.18461	1.45
		10068-40.1	0.38898	1.29315	0.83	0.16351	0.86
09PHA- J39, Lone Gull granite	2	10064-7.1	0.31796	1.31227	0.88	0.10965	0.70
		10064-9.1	0.32741	1.31066	0.93	0.10885	0.53
		10064-15.1	0.31506	1.32855	0.93	0.10653	0.54
		10064-29.1	0.32502	1.36604	0.84	0.11218	0.90
		10064-35.1	0.31996	1.47710	0.59	0.11191	2.01
		10064-42.1	0.33196	1.32125	0.90	0.10766	0.64
		10064-50.1	0.32928	1.34923	0.84	0.11356	0.86
		10064-54.1	0.30753	1.29643	0.93	0.10703	0.51
		10064-79.1	0.33466	1.32523	0.95	0.10695	0.45
	Inherited Zircon						
		10064-39.1	0.49924	1.47504	0.89	0.18037	0.75
		10064-70.1	0.46796	1.36808	0.94	0.17175	0.48
	rerun						
607	3	10064-3.1	0.32917	1.09698	0.85	0.10990	0.69
		10064-4.1	0.33489	1.06251	0.88	0.11318	0.58
		10064-8.1	0.32662	1.07568	0.87	0.11059	0.61
		10064-12.1	0.32916	1.16319	0.90	0.11247	0.55
		10064-13.1	0.32641	1.11516	0.58	0.10947	1.58
		10064-14.1	0.33069	1.02789	0.94	0.11269	0.39

Notes (see Stern, 1997):

Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number.

Multiple analyses in an individual spot are labelled as x-y.z.z

Uncertainties reported at 1 σ (absolute) and are calculated by numerical propagation of all known sources of error $f(^{206})^{204}$ refers to mole fraction of total ^{206}Pb that is due to common Pb, calculated using the 204Pb-method

common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)

* refers to radiogenic Pb (corrected for common Pb)

Discordance relative to origin = $100 * ((^{207}/^{206} \text{ age} - ^{206}/^{238} \text{ age}) / (^{207}\text{Pb}/^{206}\text{Pb} \text{ age}))$ Error in $^{206}\text{Pb}/^{238}\text{U}$ calibration 1%Th/U calibration: $F = 0.03900 * \text{UO} + 0.85600$

Cont'd.

IP#	Session	Spot Name	Apparent Ages (Ma)				Disc (%)	Conc (%)
			²⁰⁶ Pb	± ²⁰⁶ Pb	²⁰⁷ Pb	± ²⁰⁷ Pb		
08JP-52A, Lone Gull granite			²³⁸ U	²³⁸ U	²⁰⁶ Pb	²⁰⁶ Pb		
565	1	10068-21.1	1808	18.46	1812	13	0.27	100
		10068-22.2	1819	20.70	1799	9	-1.25	101
		10068-27.1	1811	19.22	1813	13	0.13	100
		10068-29.1	1854	18.01	1833	8	-1.31	101
		10068-33.1	1866	18.41	1820	11	-2.90	103
		10068-34.1	1869	25.18	1829	11	-2.48	102
		10068-37.1	1800	18.76	1813	12	0.79	99
		10068-37.2	1814	21.82	1844	13	1.86	98
		Inherited Zircon						
		10068-25.1	2628	34.98	2695	24	3.04	98
		10068-40.1	2118	23.35	2492	14	17.59	85
09PHA-J39, Lone Gull granite	2	10064-7.1	1780	20.41	1794	13	0.89	99
		10064-9.1	1826	20.84	1780	10	-2.94	103
		10064-15.1	1766	20.52	1741	10	-1.62	101
		10064-29.1	1814	21.60	1835	16	1.30	99
		10064-35.1	1790	23.08	1831	36	2.57	98
		10064-42.1	1848	21.23	1760	12	-5.73	105
		10064-50.1	1835	21.55	1857	16	1.38	99
		10064-54.1	1729	19.66	1749	9	1.36	99
		10064-79.1	1861	21.42	1748	8	-7.44	106
		Inherited Zircon						
		10064-39.1	2611	31.66	2656	12	2.09	98
		10064-70.1	2475	28.11	2575	8	4.68	96
		rerun						
607	3	10064-3.1	1834	17.51	1796	13	-2.48	102
		10064-4.1	1862	17.18	1849	10	-0.81	101
		10064-8.1	1822	17.07	1807	11	-0.94	101
		10064-12.1	1834	18.57	1838	10	0.21	100
		10064-13.1	1821	17.69	1789	29	-2.08	102
		10064-14.1	1842	16.47	1841	7	-0.03	100

Notes (see Stern, 1997):

Spot name follows the convention x-y.z; where x = sample number, y = grain number and z = spot number.

Multiple analyses in an individual spot are labelled as x-y.z.z

Uncertainties reported at 1σ (absolute) and are calculated by numerical propagation of all known sources of error

 $f(206)^{204}$ refers to mole fraction of total ²⁰⁶Pb that is due to common Pb, calculated using the ²⁰⁴Pb-method

common Pb composition used is the surface blank (4/6: 0.05770; 7/6: 0.89500; 8/6: 2.13840)

* refers to radiogenic Pb (corrected for common Pb)

Discordance relative to origin = 100 * ((207/206 age - 206/238 age)/(207Pb/206Pb age))

Error in ²⁰⁶Pb/²³⁸U calibration 1%

Th/U calibration: F = 0.03900*UO + 0.85600

Figure 4.18. Images of zircon in backscattered electron mode showing the general quality of zircon, analyzed where indicated: a) 08JP-52A, a large pyramidal, embayed Proterozoic zircon analyzed twice, the second analysis failed due to an inclusion in the target located in reflected light; b) 08JP-52A, an inherited Archean core showing very little subsequent rim growth; c) 09PHA-J39, a zircon showing recrystallization at the core revealed Nuelin-age analysis from an area of fine oscillatory zoning closer to the rim; d) 09PHA-J39, a Hudson age was inferred from the result of analysis over the unzoned core; e) 09PHA-J39, xenocrystic Archean zircon with diffuse zoning and very fine oscillatory zoning along the rim; f) 09PHA-J39, (not analyzed) a conspicuously large embayed and fractured pyramidal zircon, morphologically similar to those from 08JP-52A.

CHAPTER 5: INTERPRETATION AND DISCUSSION

5.1. Characteristics of the Nueltin suite

5.1.1. Nueltin suite Geochemistry

The Nueltin granite and Pitz Formation rhyolite have many mineralogical and geochemical similarities, and these characteristics are generally consistent despite an extensive range of emplacement conditions. Nueltin granite and Pitz rhyolite are most distinguishable by an often strongly potassium feldspar-phyc texture. All Nueltin granites and related rocks are strongly potassic which increases with silica content (Fig. 16), and felsic rocks ($\text{SiO}_2 > 65$ wt. %) have abundant alkalies (granites average 8.26 and rhyolites average 8.19) but are still subalkaline (Figure 5.1) according to Irvine and Baragar (1971). It was noted by Anderson (1983) that alkali-enrichment, but lack of a peralkaline nature is typical of many rapakivi-related granite associations. Most other major elements (wt. %) show reasonably continuous curves as rocks evolve towards silicic extrusive rocks. Also, alumina saturation correlates with silica increases, from strongly metaluminous, e.g. the McRae Lake diabase and Mallery basalt, through Nueltin granites, which straddle the metaluminous-peraluminous boundary, to the most evolved Pitz rhyolite, which is the most strongly peraluminous (Figure 5.2).

The Nueltin granites generally have high total iron contents, expressed by the presence of primary mafic minerals (biotite $\pm$ hornblende) rich in iron. Common in Nueltin granites, the presence of primary Fe-rich mafic silicate minerals, frequently as

anhedral, interstitial grains between quartz, indicate the felsic intrusive phases crystallized from initially H₂O-undersaturated magmas (Rämö and Haapala, 1995).

The relatively high modal abundance of fluorite is consistent with the high F (ppm) content, and would suggest a fluorine-rich melt source. This association is inferred due to distortions caused by F within aluminosilicate frameworks, which provides bonding sites for the above HFSE (Whalen et al., 1987) and also allows for the formation of complexes with other abundant elements Nb, Ga, and Zn. In the Nueltn suite, the exception to this the abundance of Nb, as all rocks appears to have a weak negative anomaly and could reflect a previously Nb depleted source region. High Zr (ppm) content is also apparent within the silicic units and is reflected in the general abundance of zircon. With the notable exception of the Lone Gull granite, Nueltn pluton contain zircon that typically show magmatic zoning, have low Th (ppm) and U (ppm) contents and generally lack any inheritance (van Breemen et al., 2005). The typically low values of Th and U during the early crystallization of zircon would suggest that fractionation is also responsible for the enrichment of Th and U to the higher values observed in the whole rock analyses.

Comparisons of published geochemistry data from Nueltn suite rocks and this study are presented in Figure 5.3. Nueltn granite shows negative anomalies in Ba and more strongly for Sr, which could indicate fractionation of potassium feldspar and plagioclase, respectively, and/or a Ba-depleted and plagioclase-rich source region. Pitz Formation rhyolite, which have undergone a greater degree of fractionation has similar, but more strongly negative anomalies, which would support that fractionation is a factor

in Nueltin and Pitz melt evolution. A potential relationship between the McRae Lake diabase and Pitz basalt is inferred based on strong correlations with respect to bulk crust normalized depletions and enrichments, e.g. positive Ba anomaly and negative Sr- and Ti-anomalies. Both appear to have undergone a low degrees of plagioclase fractionation. Negative anomalies for phosphorus (not shown) indicates that apatite is a fractionating phase in the felsic units. Additional fractionation of Ti-oxides and silicates such as ilmenite and titanite could account for the negative Ti anomaly and the decreasing trend observed for TiO_2 with increasing SiO_2 , but may also indicate a source region with residual Ti-bearing phases. The degree of secondary alteration within the Mallery basalt causes mobile LILE elements to have positive anomalies, but the immobile REE and HFSE show relatively MORB-like patterns.

5.1.2. Nueltin suite REE Geochemistry

Nueltin granite are REE enriched, relative to bulk continental crust within Pitz Formation rhyolites and the McRae Lake diabase (Figure 5.4). REE trends are typically parallel to sub-parallel and could suggest a comagmatic relationship between the felsic and possibly some of the mafic units. Relative to chondrite normalization, the LREE have higher abundances, with moderately negative fractionation trends, as HREE trends tend to flatten. The abundance of allanite and apatite has a strong influence on LREE patterns, especially at Nueltin Lake. The Lone Gull pluton appears to have the strongest REE fractionation, with HREE abundances strongly depleted relative to the other units. The

cause of this strong HREE depletion may reflect an instance of source region heterogeneity (possibly garnetiferous), relative to the other Nueltin granites.

All rock types have a negative Eu-anomaly, except the Mallery basalt. Despite a major difference in silica content REE patterns, except for Eu, of the Pitz basalt appear similar to the subvolcanic granite samples collected at Mallery Lake. A strong contrast in Eu-anomalies for the Mallery Lake subvolcanic Nueltin granite samples (average 0.18) and Pitz basalt (0.74) are observed. The Eu-anomaly values for the Pitz basalt are similar to averaged values for previously documented McRae Lake diabase (0.70; $n = 2$), with overall Eu abundances (3.0 ppm) for the basalt slightly lower than the diabasic dyke (3.8 ppm). Except for the Lone Gull pluton, Eu-anomalies becomes more pronounced with increasing silica content (Figure 5.5), which would further support that plagioclase fractionation has affected the melt evolution for much of the Nueltin and related rocks. Using the Eu-anomaly as a proxy for fractionation, the concentration of HFSE and REE elements increases with greater fractionation (i.e. a stronger Eu-anomaly). A weaker Eu-anomaly for the Lone Gull granite would indicate that it has not been affected by a high degree of plagioclase fractionation. The Mallery basalt is significantly different with nearly flat REE patterns. This flat pattern would suggest an asthenospheric melt origin, with a small degree of crustal contamination during ascent, most likely accommodated by crustal thinning.

5.1.3. Nueltin suite Isotope Geochemistry

Scattering within the $\epsilon_{\text{Nd},1.75 \text{ Ga}}$ values is likely a result of source heterogeneities of the Nueltin granite and related rocks. Even locally, differences are apparent, e.g. at Nueltin Lake, where the more silicic and possibly more fractionated southern pluton has higher $\epsilon_{\text{Nd},1.75 \text{ Ga}}$ values. Regardless, the spread of $\epsilon_{\text{Nd},1.75 \text{ Ga}}$ values between -4.8 and -9.8 for the granitoids is interpreted to indicate that they share similar Archean source materials and that little, if any, Paleoproterozoic mantle component contributed directly to the formation of these rocks. Isotopic values from Nueltin suite rocks within the Hearne and Rae domains do not show distinct differences and may have similar formation histories. The positive value for the Mallery basalt ($\epsilon_{\text{Nd},1750 \text{ Ma}} = +2.4$) is consistent with strong input from a depleted mantle source with mild crustal contamination or lithospheric mantle input. This is supported by the weak degree of REE fractionation ($\text{La}_\text{N}/\text{Yb}_\text{N}$) at approximately 10x chondritic values.

Peterson et al. (2002) and Dudás et al. (1991) reported $\epsilon_{\text{Nd},1.75 \text{ Ga}}$ values ranging from -10.7 to -7.1 for Nueltin granites, -8.6 for a Pitz rhyolite sample and two McRae Lake diabase samples averaged -7.0 (Figure 5.6). The $\epsilon_{\text{Nd},1750 \text{ Ma}}$ values determined for the Nueltin suite samples in this study generally overlap previously determined values. The $\epsilon_{\text{Nd},1.75 \text{ Ga}}$ value for the Pitz basalt (-6.7) is markedly similar to the McRae Lake diabase. In general, the Nd-isotopic evolution of Nueltin granites shows a predominant input by melting of a heterogeneous Archean crust.

5.2. Nueltin suite - Anorogenic (A-type) suite

Based on petrographic evidence and field relationships, Peterson et al. (2002) showed that Nueltin granite and Pitz Formation rhyolite were products of an anorogenic event and share similar emplacement style to rapakivi granites, even when rapakivi texture (*sensu stricto*) is not apparent. Geochemical discrimination diagrams have been developed and can also be used to discriminate geotectonic settings. Use of discrimination diagrams comes with a caveat, as granites are typically crystallized from highly evolved sources or the product of melting of relatively differentiated material (i.e. crustally-derived). It is unlikely that granite geochemistry will adhere strictly to any rigid classification scheme, but general trends may be useful in elucidating source material and comparable analogs. Most of these diagrams rely on ratios of relatively alteration insensitive HFSE, which are commonly abundant in anorogenic granites and in the Nueltin granitic rocks, as discussed above.

Using the geotectonic discrimination diagrams proposed by Pearce et al. (1984), Nueltin granites and Pitz rhyolites generally plot in the within plate granite (WPG) field for Rb-Y+Nb and Nb-Y plots with some overlap into syn-collisional granite (syn-COLG) field (Figure 5.7). A-type granites typically plot within the WPG field but Pearce et al. (1984) stresses caution when using these diagrams where there is a high content of volatiles, which could lead to fluxing, even in water-undersaturated granites. This is exemplified in the F-rich pegmatite at Nueltin Lake, which plots in the ocean ridge granite (ORG) field.

The geochemical discrimination plots (Figure 5.8) developed by Whalen et al. (1987) use ratios in abundances of highly charged cations (Ga, Nb, Y, Zr, and REE) and Al, Mg and Ca, to differentiate A-type granites from extensively fractionated granites (FG) or unfractionated granites (OGT) of either I-type and S-type. These diagrams place the Nuelin granites in fields that have characteristics of A-type granites. The aplite dykes from Amer plutonic complex tend to plot within the highly fractionated field, and the Amer plutonic complex rapakivi samples, some of which plot on the cusp or slightly within the I & S field. Both samples were collected at the periphery of the pluton, and therefore crustal contamination may have played a more significant role influencing the local chemistry. This overlap outside of the A-type field is documented with other rapakivi granites such as those noted by Rämö and Haapala (2005). Whalen et al. (1987) and others (e.g. Clemens et al., 1986; Collins et al., 1982; Eby, 1990) attribute the petrogenesis of these A-type granites to the melting of a water-deficient, melt-depleted (granulitic) residue left over by previous extraction of an orogenic granite melt.

Additionally, U-Pb geochronology indicates that the Nuelin event did occur *ca.* 50 Ma after the post-orogenic Hudson granitoids (van Breemen et al., 2005), consistent with a non-orogenic emplacement. A possible conclusion therefore is that the source region for the Nuelin granites was affected by a previous depletion by the late to post-collisional Hudson event. This supports the Nd-isotopic overlap between Hudson suite granites and Nuelin granites (Peterson et al., 2002).

Eby (1992) sub-divided the A-type granites based on elemental ratios Y/Nb and incompatible elements to contrast melt sources between sub-types and highlight source

affinities. The sub-division of granitoids should only be used for granitoids that plot both in the WPG of Pearce et al. (1984) and A-type field of the Ga/Al plots of Whalen et al. (1987), as such non-plutonic units have been removed from the diagram. The A₁ group element ratios have affinities with intraplate and rift-zone magmas leading to the inference that they represent mantle differentiates, variably contaminated by continental crust, while the more diverse A₂ group has geochemical affinities ranging from island-arc basalts to average continental crust (Eby, 1992). The Nuelin granites generally plot within the A₂-type fields (Figure 5.9) which suggests a dominantly crustal origin. Exceptions to this trend, with overlap into the A₁ by some of the Amer plutonic complex rapakivi samples and the Lone Gull granite. This may indicate a degree of fractionation and/or restitic Nb- or Y-rich minerals (e.g. amphibole, ilmenite, allanite, garnet) for the Amer plutonic complex and Lone Gull samples, subsequently altering the ratio.

Work by Frost et al. (2001) instead relies on major element compositions, namely Fe and alkali elements versus silica content, to differentiate between granites, but draws only broad geotectonic conclusions as chemistry is primarily a function of their source rocks, rather than geotectonic setting. Frost et al. (2001) use changes in total FeO content relative to MgO content (Fe^* ; $\text{FeO}^{\text{total}}/(\text{FeO}^{\text{total}} + \text{MgO})$) as a method to determine differentiation history and the modified alkali-lime index (MALI; $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$) as a proxy for feldspar composition and abundance. According to these classification schemes, and with the exception of visibly altered samples, all the Nuelin granites sampled are ferroan and dominantly follow alkali-calcic trends (Figure 5.10). Along with the metaluminous to peraluminous character the Nuelin granite has similar

characteristics as other A-type granites, including Finnish rapakivi granites (Rämö and Haapala, 2005) and other metaluminous to peraluminous A-type granites like the Sherman batholith, in southeastern Wyoming (Frost and Frost, 2011). Suites with A-type affinities are noticeably iron-enriched, which Frost and Frost (1997) attribute to derivation from reduced basaltic rocks, like those common to extensional environments, by either fractional crystallization or partial melting, driving compositions towards higher iron- and alkali-enriched compositions. This interpretation is applicable to the Nueltin granites as normal block-faulting and extension occurs spatially and generally contemporaneous with the emplacement of the Nueltin plutons (Rainbird et al., 2003).

Some consistent geochemical similarities between the Nueltin granite (Pitz rhyolite) and A-type (rapakivi) granites that are not represented in the above discrimination diagrams include, high F content, enriched REE (especially LREE relative to HREE), strong Eu-anomalies, low Ca- and P contents, and aluminum saturation ranging from metaluminous (early intrusive) to mildly peraluminous (late intrusive) (Rämö and Haapala, 2005). A-type granites are often enriched in incompatible elements, base metals (such as Cu, Sn, and Pb) and can have significant concentrations of Th, U and Au, due to their highly evolved nature (Gill, 2010).

The affinity of Nueltin granites with characteristic A-type and rapakivi granitoids of similar mineralogy and chemistry could indicate a similar model for generation, e.g. Finnish rapakivi granites (Rämö and Haapala, 2005). Rämö and Haapala (2005) suggest magmatic underplating during an extensional regime, resulting from partial melting of the upper mantle. Mafic magma intrusion and ponding in the lower crust can lead to partial

melting of the lower continental crust to form granitic magma. Mafic magmas can also intrude along deep fissures and can subsequently mingle/mix locally with felsic magmas.

5.3. Nueltin (A-type) suite and uranium prospectivity

The extent to which source characteristics or fractional crystallization are responsible for observed geochemical characteristics for the Nueltin suite, evidence would favour fractionation as the primary factor. Source heterogeneities are clearly evident within Nd-isotopic data and would suggest they do affect sample geochemistry, but the extent is obscured by more dominant fractionation. Those elements that show little to no increased concentration with fractionation can possibly be used as indicative of source chemistry. Magma mixing, as documented by Booth (1983), and the entrainment/resorption of basaltic fragments, as documented by Peterson et al. (2002), can further complicate interpretation by affecting mineralogy and geochemistry on a local scale.

Mineralization associated with A-type and rapakivi granites has been documented by many (e.g. Haapala (1995) and references therein). Significant examples of metallogenic enrichments include Sn-Be-W-Zn-Cu, observed in the Eurajoki and Kymi stocks of Finland (Rämö and Haapala, 1995), Sn-W-Nb-Ta-Be, associated with the tin granites of Abuna-Rio Caripunas (Bettencourt and Dall'Agnol, 1987), and Cu-Au-Ag-U, associated with the Roxby Downs granite at Olympic Dam (Creaser, 1996; Roberts and Hudson, 1983). The granites associated with these deposits are highly potassic, Fe-rich

and typically shallowly emplaced plutons. The geochemical character of the granites has been linked as causative to their enrichments.

With respect to U-enrichment, Cuney and Kyser (2008) cite alkaline granites, enriched in K (high-K or shoshonitic series), and equivalent to A₂-type classification of Eby (1992) as prospective characteristics for exploration. Enrichment is due to the similarly incompatible nature of K₂O with Th and U and can be observed geochemically within Nueltin granites and Pitz rhyolite (Figure 5.11). The characteristically high melting temperature and fluorine-rich nature of A-type granites is also considered a favourable 'ingredient' in the generation of U-rich igneous rocks (Collins et al., 1982; Plant et al., 1999), exemplified by F-rich ore at Olympic Dam (Haapala, 1995). Cuney and Kyser (2008) also consider the metaluminous nature significant, as it can have an intermediate effect on the behaviour of elements Th, Zr, and REE, leading to possible enrichment. Uranium concentrations in Nueltin granites and Pitz rhyolite appeared to favour those samples closer to the peraluminous-metaluminous cusp (Figure 5.12). Nueltin granites and comagmatic Pitz rhyolites generally comply with the characterizations outlined above and consequently deserve attention as sources for U (Th and REE).

Significant U mineralization in the Kiggavik area, associated with the hypabyssal textured Lone Gull granite demonstrates economic potential. Mineralization occurs elsewhere, but tends to be highly localized. At Nueltin Lake, significant enrichment of U, Th and REE, occurs within extremely fractionated rock, e.g. late-crystallizing pegmatitic pods and aplites, of the Nueltin granite (Scott et al., 2012). In relation to the Pitz rhyolite,

LeCheminant et al. (1987) noted overall elevated U, but only within high-temperature glassy, rhyolitic domes. Local enrichments related to the granites appear to be caused by the incorporation of U, Th, and REE into refractory minerals such as titanite, allanite and, if Th/REE ratio is high enough, (urano-)thorite. Consequently, U can only be mobilized and concentrated if these minerals are radioactive enough to undergo a high degree of metamictization, thereby allowing hydrothermal fluids to scavenge the metals. In the case of the U-rich Pitz rhyolite domes, devitrification is necessary to liberate any metals. In general, Nuelin granite and Pitz rhyolite exceed average crustal contents of U only modestly and do not appear enriched on a wide or exploitable scale, especially relative to the *ca.* 80 Ma older, spatially related Paleoproterozoic Hudson suite and contemporaneous rocks. Similarities in trace element geochemistry and Nd-isotopes have led Peterson et al. (2002) to suggest a related source for Hudson and Nuelin melt generation, therefore the older Hudson suite, inferred to be mid crustal, may have formed deeper in the crust and depleted much of the source regions with respect to U.

Since the Kiggavik U-deposit represents the most prospective occurrence of mineralization known in relation to Paleoproterozoic igneous rocks within the area of the Baker Lake - Thelon Basin region, resolving the nature of the U-hosting Lone Gull granite is important for determining suite affinity of uranium, if any. Geochemical results for the Lone Gull granite indicate great geological complexity, which makes interpretation difficult. The drill core sample (08JP-52A) collected from 75.5 m depth, indicates an age of 1820 ± 12 Ma, and is close in age with the local outcrop of porphyry (Scott et al., in prep). In the shallow (17.3 m) sample (09PHA-J39), the presence of two

populations is suggested based on discrete clustering of samples based on Th/U ratios which correlate with the bimodal ages. The older population *ca.* 1806, is broadly consistent with Hudson emplacement, while the younger population 1748 ± 9.4 Ma, matches the timing for intrusion of Nueltin plutons and eruption of Pitz rhyolite (Rainbird and Davis, 2007; van Breemen et al., 2005). Both of the Lone Gull granite drill core samples are strongly porphyritic. Nowhere is this texture documented for granitic Hudson rocks, which are inferred to have migrated only a short vertical distance from their source region before crystallizing within the mid-crust (Peterson et al., 2002). However, a texturally similar porphyritic character is well documented in hypabyssal stocks of Nueltin granite, associated quartz feldspar porphyries and Pitz Formation rhyolite (Booth, 1983). Petrographic analysis of the Lone Gull granite further supports a volcanic or hypabyssal origin, including resorbed and embayed phenocryst phases.

Analysis of zircon in sample 09PHA-J39 which record Nueltin-ages, took place over zoned areas which had no apparent signs of disturbance (e.g., recrystallization, alteration) in BSE, CL or transmitted light. This would suggest areas of zoning are magmatic in origin, thus supporting hypabyssal Nueltin emplacement. The presence of an older population, seemingly inherited from Paleoproterozoic Hudson-aged sources contrasts with the nature of Nueltin rocks, which crystallized from hot, F-rich melts, and zircon inheritance is not documented elsewhere (Peterson et al., 2002; van Breemen et al., 2005). The reproducibility of the bimodal population was evident when reanalyzing the sample, but the reasons were apparent in the high ^{204}Pb contamination, observed for Nueltin-aged results. In both sessions, Hudson-ages in 09PHA-J39 suffered from low

precision and high MSWD, this could speculatively indicate another subset population. The bimodal results of 09PHA-J39 contrast with those from the deeper Lone Gull sample, within 08JP-52A there is no indication of multiple zircon populations, yet sampling bias cannot be dismissed. If Nuelin age zircons or overgrowths are of quality too poor or rims are too fine to analyze, the zircon may be separated out early in the selection process.

Trace element geochemical concentrations differ between Lone Gull granite samples, likely due to the stronger degree of alteration for the more shallow 09PHA-J39, nevertheless, trends for the two Lone Gull granite samples are sub-parallel. Because hydrothermal alteration is apparent, including fluorine within the fracture system, geochemistry is limited to the most alteration insensitive elements. Patterns show a strong negative Ta- and Nb-anomaly, steep REE fractionation and a weak negative Eu-anomaly, more commonly observed in Hudson geochemistry. This would in fact indicate the strong influence of a melt source similar to that for Hudson granitoids.

If the Lone Gull granite was emplaced at the time of the Hudson plutonic episode, the presence of a hypabyssal unit juxtaposed with mid-crustal plutonic facies must be addressed. Requiring *ca.* 15 - 20 km of exhumation, which could reasonably occur in a 20 Ma span. If the Lone Gull granite represents a Nuelin-aged emplacement, as petrography might suggest, then a similar exhumation is required, but over a 70 Ma span of time. The strong Hudson-like chemistry and presence of Hudson zircon inheritance would suggest a similar melt source and/or strong interaction with Hudson material. Either scenario

represents a level of geological complexity not previously documented for these Paleoproterozoic igneous units.

5.4. Sources of Uncertainty

Due to the coarse-grained nature of most Nueltin granites, very large, representative-sized samples were required, to mitigate any small-scale sample homogeneity, and with some samples run in duplicate.

Geochemical results could be biased by the incomplete dissolution of resistive minerals like zircon during sample preparation. Zr content (ppm) determined via XRF was compared to ICP-MS results, which revealed similar or higher values using ICP-MS. This result cannot be qualified, but would at least suggest that Zr contents are not underreported. Additionally, Ta values in the study granites are relatively close to the reported limit of detection and therefore the determined values may differ highly. Samples in duplicate, marked by (*), show good precision for Ta values.

Sources of analytical error for Nd-isotopic work could involve the incomplete dissolution of titanite, common in Nueltin granite, which had detectable traces of Nd, up to 0.4 wt. % was detected and this could have an effect on the Sm/Nd ratio. Additionally during the filament mounting step, if rapid heating, to evaporate the phosphoric acid, creates a crust over the sample, thermal ionization will be impeded, leading to poor signal.

Geochronology results can be affected by sample bias during the selection process. Typically, the higher quality zircon are separated from the least magnetic

fractions, as these generally suffer less from matrix effects, i.e. differences in normal zircon crystal structure, which can affect calibration. Multiple sessions on the SHRIMP can also add a degree of error, as analytical conditions have changed relative to other sessions. Additionally, sample populations are too low, approaching the limit of statistical validity. A larger sample set may improve precision in the reporting of ages.

Figure 5.1. Total alkalinity versus silica for Nueltin suite rocks after Cox et al. (1979). Data from previous work from Peterson et al. (2002) and references therein.

Figure 5.2. Alumina saturation versus silica content shows Nueltn granite and related rocks progress from strongly metaluminous to strongly peraluminous as rocks become more felsic.

Figure 5.3. Continental crust-normalized (Taylor and McLennan, 1995) extended trace element diagram for a) Nueltin granite and b) related rocks.

Figure 5.4. Chondrite-normalized (Nakamura, 1974) rare earth element diagram for a) Nueltin granite and b) related rocks.

Figure 5.5. Increased silica generally correlates with a more pronounced Eu-anomaly, likely due to continued plagioclase fractionation as the melt evolves.

Figure 5.6. Diagram showing epsilon Nd vs. age (Ma) for rock suites throughout the western Churchill Province. Nueltin suite rock from this study appear cross-hatched, overlap with Nueltin rocks (Dudás et al., 1991; Peterson et al., 2002). Epsilon Nd values from other suites found in western Churchill Province suites are plotted for comparison (Byrne, 2005; Cousens et al., 2001; Peterson et al., 2002). Depleted Mantle evolution curve is calculated assuming $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.51315 and $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.215 for modern depleted mantle.

Figure 5.7. Geotectonic diagrams after Pearce et al. (1984) which indicates melt generation environment.

Figure 5.8. Geotectonic diagrams after Whalen et al. (1987), differentiating A-type granites (A denotes calculated average) from: a) and b) fractionated felsic granites (FG) and unfractionated I- and S-type granites (OTG); c) to f) I-type and S-type granites.

Figure 5.9. Geotectonic diagrams after Eby (1992) differentiating types of anorogenic granites.

Figure 5.10. Geotectonic diagram after Frost et al. (2001), outlines indicate established fields for A-type granites, including rapakivi granites.

Figure 5.11. Bivariate plots show the covariation of Th and U with increasing potassic character.

Figure 5.12. Bivariate plot shows U content (ppm) is typically highest when the alumina index (A/CNK) approaches 1.

CHAPTER 6: CONCLUSIONS

The emplacement of Paleoproterozoic Nueltin suite igneous rocks characterize a major anorogenic event, signifying the last region-wide pulse of igneous activity across the Rae and Hearne domains of the western Churchill Province which followed a period of relative quiescence after the THO. The *ca.* 1750 Ma Nueltin event is marked by the bimodal extrusion of commonly porphyritic rhyolite and basalt lavas, and the intrusion of megacrystic and rapakivi granite plutons. Major and trace element geochemistry of the silicic rocks indicate strong fractionation, especially for Nueltin granite and Pitz Formation rhyolite which are strongly potassic, have pronounced negative Sr, Eu, and Ti anomalies and steep REE fractionation patterns. These features are consistently observed for Nueltin suite rocks, yet Nd-isotopic data suggests that most lithologies are derived from heterogeneous, crustal sources across a broad range of emplacement. Therefore, observed characteristics of Nueltin suite rocks a of strong fractionation. Where noticeable variations do exist, extreme crystal fractionation or mingling with basaltic magma is observed.

Basaltic rocks associated with the suite indicate partial melting of the mantle occurred coeval to the emplacement of the Nueltin granite plutons. Based on enrichment of REE and strongly radiogenic ϵ Nd-isotope signature of associated dykes, like the McRae Lake diabase and Pitz basalt, a predominant derivation from the lithospheric mantle is inferred. Additionally, these basalts have undergone significant crustal assimilation after a period of prolonged ponding in the crust before ascending. Basalt collected at Mallery Lake further indicates tapping of deeper asthenospheric mantle

during this event if the basalt is coeval. Partial melting of the mantle could have provided a heat source for the partial melting of the crust to generate Nueltin granite and Pitz rhyolite. Partial melting of the mantle, leading to melting of the crust, could be due to extensional thinning of the crust via normal block faulting within the Baker Lake Basin (Hadlari and Rainbird, 2011). Magmatic underplating of the crust is also a possibility in response to delamination because of rapid erosion of the THO or related to a mantle plume. The generation of anorogenic, rapakivi granite due to two stage partial melting of lower crust is a model cited by Rämö and Haapala (2005). Coeval subparallel diabase dyke swarms and normal faulting are indicative of an extensional regime. The low ϵ_{Nd} value for the rapakivi granite is indicative of lower crustal melting and are observed mingling with diabase dykes which overlap with the higher radiogenic values of the granite (Ramo, 1991). Rämö and Haapala (2005) suggest magmatic underplating of the crust accounts for the extensional tectonic setting and bimodal magmatism. The rapakivi granite of Finland has similar characteristics and possibly an analogous mode of petrogenesis to the Nueltin suite (Figure 6.1).

Nueltin suite timing, approximately 80 Ma after the peak of magmatism related to the THO, bimodal emplacement, and context within an extensional regime are all factors common to anorogenic suites. Potassium feldspar dominated phenocrysts assemblages, giving a strongly potassic, yet sub-alkaline character is typical of A-type granites. Rapakivi texture is strongly indicative of an anorogenic granite. Also, the presence of Fe-rich mafic minerals, attributed to an Fe-rich source region is a feature of anorogenic granites according to Frost et al. (2001). Overall elevated contents of HFSE, enrichments

in REE and pronounced negative Eu-anomalies are also characteristic of anorogenic granites. In general the Nueltin.

Crustally derived magmas that subsequently underwent higher degrees of fractionation are responsible for volumetrically small occurrences of primary U and REE mineralization within Nueltin suite rocks. This uranium is hosted in refractory silicates, but can be mobilized with metamictization alteration, favouring the Nueltin suite as more prospective for secondary sources of uranium mineralization. More significant uranium mineralization seems to be favoured by more geologically complex Nueltin emplacements, for example at Kiggavik where the Nueltin phase of granite appears to have been influenced by a geochemically different source.

The relationship of the Paleoproterozoic igneous suites at Kiggavik represents strong geological complexity, as the Nueltin and Hudson granitic plutons more typically occur in mutual exclusion. Nueltin Lake is one of the few locales where the two units are mapped in contact. Speculatively, if U-mineralization at Kiggavik is related to Nueltin granite which has been geochemically influenced by Hudson granitoid as observed in the Lone Gull granite, possibly by incorporation during ascent, then perhaps additional occurrences can be located in the Nueltin Lake area. Recent regional lake sediment and water geochemical data (McCurdy et al., 2012) from the area could also be utilized for targeting anomalous U.

Future work to further characterize Nueltin suite mineral chemistry using EMP, would help differentiate Paleoproterozoic plutons where mineral assemblages and geochemistry between Hudson granite and Nueltin granite rocks overlap, especially in the

Kiggavik area, for example both units show common titanite at this location and are within the granite. Mineral chemistry may "fingerprint" the granites that can be compared to titanite at from the Lone Gull granite. Additionally, chemical characterization of resistive minerals which can be separated via till sampling methods, e.g. titanite, could assist in locating unmapped Nuelin plutons using glacial drift modeling.

Basaltic rocks, initially identified by (A. LeCheminant, pers. comm., 2011) outcrop along the Thelon river and are speculated to be part of the Nuelin suite. The locating and sampling of the mafic component of the Nuelin event will better allow characterization and contribute to developing a better petrogenetic model for the Nuelin suite and thus the geotectonic history of the western Churchill Province.

Figure 6.1. Adapted model after (Rämö and Haapala, 2005) for generation of rapakivi granite, with analogous units from this study at approximate levels of exposure.

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APPENDIX A: LATITUDE AND LONGITUDE CO-ORDINATES FOR SAMPLES

Station	Latitude	Longitude
09PHA-J03	60.21025	-99.96575
09PHA-J04	60.21102	-99.96765
09PHA-J05	60.21169	-99.96723
09PHA-J06	60.21222	-99.96789
09PHA-J08	60.20775	-99.96483
09PHA-J19	60.20822	-99.96723
09PHA-J11	60.09195	-100.13496
09PHA-J13	60.09311	-100.13944
09PHA-J14	60.09539	-100.11749
09PHA-J18	60.09036	-100.13814
09PHA-J21	60.17512	-100.12920
09PHA-J22	60.17068	-100.13015
09PHA-J23	60.16903	-100.13271
09PHA-J39	64.44137	-97.65598
10PHA-J06	63.61127	-98.36908
10PHA-J07A2	63.83851	-98.32633
10PHA-J08	63.85575	-98.26431
10PHA-J09	63.85666	-98.26799
10PHA-J24	65.71778	-98.01116
10PHA-J25	65.72114	-98.01561
10PHA-J26	65.72073	-98.01865
10PHA-J28	65.71059	-98.14397
10PHA-J29	65.70662	-98.14609
10PHA-J33	65.70449	-98.08957
82TX-S188	65.69783	-98.15591

APPENDIX B: WHOLE ROCK ANALYSIS

Sample	10PHA-J09	10PHA-J08	10PHA-J06A*	10PHA-J07A2	09PHA-J03*
Locale	Mallery Lake	Mallery Lake	Ruby Lake	Mallery Lake	Nueltin Lake
Lithology	granite	granite	basalt	basalt	granite
SiO ₂	74.3	71.7	55.6	46.8	66.4
Al ₂ O ₃	12.22	12.45	16.52	14.41	14.39
CaO	0.50	0.54	2.10	11.63	2.44
K ₂ O	5.44	5.73	5.14	0.89	5.16
MgO	0.14	0.35	2.52	6.78	0.51
MnO	0.05	0.05	0.08	0.24	0.07
Na ₂ O	3.38	3.31	5.27	2.99	3.00
P ₂ O ₅	0.02	0.03	0.56	0.07	0.19
Fe ₂ O _{3t}	2.24	2.83	8.05	11.72	4.93
TiO ₂	0.34	0.36	1.40	1.09	0.61
Ba	127	230	1922	145	1038
Be	6	8	3	b.d.l.	4
Cs	2	2.8	1.2	21.9	7.6
Ga	20.5	21.2	17.4	14.7	25.9
Hf	10.9	11.5	9.5	1.8	12.6
Nb	41.7	42.2	46.3	2.9	33.4
Rb	314.5	337.8	132.9	77.4	315.5
Sn	4	3	6	b.d.l.	6
Sr	27.6	41.7	426	318.5	229.6
Ta	2.8	2.7	2.3	0.1	1.7
Th	25.8	26.4	14.0	0.2	20.9
U	3.8	4.4	2.7	0.1	5.2
W	1.1	1.3	2.1	1.4	1.8
Zr	354.9	392.6	393.4	51.7	446.4
Y	36.2	35.6	35.4	18.2	51.8
La	106.5	106.8	96.6	3.6	40.5
Ce	227.4	220.9	202.7	9.3	85.7
Pr	23.95	23.58	23.13	1.42	11.95
Nd	84.0	83.7	89.5	7.5	54.1
Sm	12.46	12.67	14.00	2.35	11.24
Eu	0.57	0.76	3.08	0.83	1.84
Gd	8.73	8.82	10.60	3.01	10.00
Tb	1.25	1.25	1.38	0.53	1.61
Dy	6.92	6.66	7.23	3.50	8.98
Ho	1.24	1.24	1.24	0.68	1.76
Er	3.44	3.6	3.16	1.95	5.11
Tm	0.57	0.55	0.49	0.31	0.79
Yb	3.66	3.54	3.12	1.87	5.14
Lu	0.56	0.53	0.45	0.29	0.76
Mo	1.1	3.0	0.4	0.9	0.9
Cu	0.7	1.5	1.8	114.5	6.8
Pb	10	28.9	9.6	3.1	13.3
Zn	75	96	57	92	111
Ni	1.1	1.7	1.2	84.4	4.6
As	b.d.l.	0.6	0.9	2.7	0.9
Ag	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Au	9	6.6	5.2	b.d.l.	b.d.l.
F	2461	2480	1383	580	3032

b.d.l. = below detection limit

* indicates an averaged duplicate

oxides are reported in weight percent, trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

Sample	09PHA-J04	09PHA-J08	09PHA-J10*	09PHA-J11	09PHA-J13
Locale	Nueltin Lake	Nueltin Lake	Nueltin Lake	Nueltin Lake	Nueltin Lake
Lithology	granite	granite	granite	granite	granite
SiO ₂	67.1	67.4	65.9	69.6	71.8
Al ₂ O ₃	13.55	13.92	14.52	13.30	12.75
CaO	2.40	2.39	2.45	1.22	1.21
K ₂ O	4.66	5.16	5.26	6.03	5.85
MgO	0.64	0.52	0.58	0.26	0.32
MnO	0.09	0.07	0.07	0.03	0.04
Na ₂ O	2.78	2.79	2.94	2.43	2.24
P ₂ O ₅	0.23	0.19	0.26	0.14	0.14
Fe ₂ O _{3t}	6.25	4.52	5.20	3.17	3.47
TiO ₂	0.72	0.56	0.67	0.32	0.36
Ba	890	1124	1144	512	633
Be	4	5	4	2	2
Cs	11.9	10.2	4.2	5.5	6.7
Ga	26.2	24.4	24	25	24
Hf	13.8	11.9	12.9	11.1	10.6
Nb	40.6	36.4	29.4	30.5	30.1
Rb	345.4	337.5	261.5	381.3	381.7
Sn	10	9	5	3	5
Sr	208.6	233.7	249.7	100.6	95.4
Ta	2.1	3.4	1.9	1.4	1.7
Th	15.5	23.4	41.3	126.2	109.5
U	5.3	5.7	4.6	6.5	9.2
W	3.1	2.8	1.8	1.7	2.2
Zr	512.6	409.1	467.8	338.6	352
Y	56.2	66.2	49.6	37.5	43.0
La	36.3	45	72.2	189.7	172.8
Ce	84.9	100.9	172.4	396.7	377
Pr	12.22	15.7	23.54	42.88	41.21
Nd	55.7	65.8	91.5	147.0	143.8
Sm	12.13	13.27	15.44	20.26	20.77
Eu	1.70	1.90	2.20	1.25	1.19
Gd	10.42	11.02	11.60	11.9	12.97
Tb	1.67	1.92	1.74	1.54	1.84
Dy	9.01	10.97	9.24	6.79	8.19
Ho	1.8	2.19	1.76	1.18	1.43
Er	5.26	6.68	5.02	3.16	3.72
Tm	0.83	1.03	0.77	0.44	0.56
Yb	5.58	6.82	5.07	2.88	3.49
Lu	0.88	1.05	0.76	0.42	0.51
Mo	1.3	1.2	1.3	1.5	1.2
Cu	9.5	5.5	6.4	6.0	7.3
Pb	13.9	10.2	17.85	23.9	18.9
Zn	140	98	92	88	96
Ni	6.3	5.3	4.0	4.6	4.1
As	0.5	0.5	1.1	0.8	0.7
Ag	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Au	0.6	b.d.l.	0.6	b.d.l.	1.6
F	3230	3470	2414	2190	2360

b.d.l. = below detection limit

* indicates an averaged duplicate

oxides are reported in weight percent

trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

Sample	09PHA-J14	09PHA-J18*	09PHA-J21	09PHA-J22	09PHA-J23*
Locale	Nueltin Lake	Nueltin Lake	Nueltin Lake	Nueltin Lake	Nueltin Lake
Lithology	granite	granite	granite	granite	granite
SiO ₂	68.9	70.2	71.8	67.8	72.4
Al ₂ O ₃	14.10	13.41	12.81	13.93	12.25
CaO	1.49	1.28	1.11	1.62	1.39
K ₂ O	6.10	6.55	6.05	5.82	5.00
MgO	0.37	0.30	0.36	0.38	0.31
MnO	0.04	0.03	0.03	0.05	0.03
Na ₂ O	2.57	2.38	2.20	2.74	2.55
P ₂ O ₅	0.20	0.15	0.14	0.18	0.11
Fe ₂ O _{3t}	3.92	3.36	3.52	4.89	3.64
TiO ₂	0.46	0.37	0.35	0.50	0.37
Ba	573	678	508	668	603
Be	3	3	2	3	2
Cs	6.9	6.1	6.8	9.6	6.8
Ga	26.3	23.6	23.8	27.6	23.2
Hf	13.7	11.9	11.9	16.6	13.6
Nb	36.1	26.9	29.8	39.8	26.7
Rb	393.9	377.8	379.2	423.3	317.8
Sn	5	6	6	7	6
Sr	104.4	110.7	86.8	112.4	121.7
Ta	1.9	1.4	1.6	1.8	1.45
Th	116.7	189.7	123.6	100.6	99.6
U	6.4	6.1	4.7	6.3	11.8
W	2.1	2.15	2.4	2.6	1.75
Zr	451.6	352.6	361.5	541.3	441.2
Y	42.0	50.3	39.0	65.5	51.7
La	204.8	238.2	195.8	157	242.2
Ce	444.6	490.3	412.3	324.2	531.6
Pr	48.96	50.88	45.04	35.97	55.43
Nd	172.9	173.8	156.5	128.7	185.3
Sm	25.33	21.96	21.24	19.85	24.36
Eu	1.36	1.31	1.06	1.29	1.33
Gd	16.64	13.77	13.1	14.76	14.94
Tb	2.12	1.875	1.61	2.3	2.155
Dy	8.64	9.03	6.84	11.55	10.6
Ho	1.39	1.61	1.09	2.14	1.86
Er	3.46	4.36	2.8	6.18	5.19
Tm	0.52	0.65	0.41	0.93	0.76
Yb	3.55	4.24	2.8	5.77	4.73
Lu	0.51	0.6	0.44	0.83	0.69
Mo	2.8	1.2	0.8	3.6	1.9
Cu	10	6.5	5.4	6.1	4.2
Pb	21.9	30.1	14.4	28.8	17.7
Zn	102	98.5	95	133	91
Ni	4.7	4.6	4.7	5.3	3.4
As	0.7	1.2	0.8	0.6	1.1
Ag	b.d.l.	0.1	b.d.l.	b.d.l.	b.d.l.
Au	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
F	2730	3008	2520	3560	3392

b.d.l. = below detection limit

* indicates an averaged duplicate

oxides are reported in weight percent

trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

Sample	09PHA-J39	08JP-52A	10PHA-J24B Amer plutonic complex	10PHA-J25A Amer plutonic complex	10PHA-J28 Amer plutonic complex
Locale	Lone Gull	Lone Gull			
Lithology	granite	granite	granite	granite	granite
SiO ₂	72.4	69.7	70.3	70.7	71.6
Al ₂ O ₃	13.52	13.07	13.84	13.36	13.50
CaO	0.51	1.25	1.78	1.45	1.29
K ₂ O	6.52	5.78	5.62	5.80	5.66
MgO	0.77	1.67	0.63	0.63	0.35
MnO	0.03	0.04	0.04	0.04	0.04
Na ₂ O	2.62	2.71	3.11	2.84	3.26
P ₂ O ₅	0.09	0.23	0.13	0.13	0.09
Fe ₂ O _{3t}	1.40	2.25	2.53	2.59	2.10
TiO ₂	0.20	0.36	0.46	0.50	0.35
Ba	929	1911	1381	1119	908
Be	6	5	3	3	6
Cs	4.8	3.9	4.3	3	3.7
Ga	19.7	20	18.3	17.9	17.7
Hf	6.4	7.4	5.8	8.8	6.8
Nb	12.7	12.5	17.1	23	18.9
Rb	233.3	217.3	265.4	282.7	373.8
Sn	2	2	2	3	1
Sr	295.1	580.3	304	263.3	199.7
Ta	1.1	1	1.3	2	1.4
Th	49.9	46.5	26.2	33.7	53.4
U	7.0	12.1	4.6	6.4	10
W	1.9	2.1	1.5	0.8	1.1
Zr	190.3	252.9	195	281.2	183.8
Y	5.8	8.7	29.3	37.7	20.3
La	49.4	75.7	73.5	59.9	83.8
Ce	87.1	144.9	159	144.7	154.7
Pr	7.91	14.38	17.88	17.35	14.7
Nd	24.2	47.8	65	66	49.5
Sm	3.14	6.27	9.59	10.68	6.34
Eu	0.68	1.37	1.74	1.64	1.16
Gd	1.67	3.79	6.85	7.77	4.57
Tb	0.23	0.42	0.97	1.15	0.62
Dy	1.01	1.93	5.29	6.85	3.48
Ho	0.15	0.25	1.02	1.23	0.67
Er	0.47	0.59	2.81	3.59	1.9
Tm	0.07	0.1	0.46	0.62	0.34
Yb	0.49	0.61	2.94	3.93	2.44
Lu	0.07	0.08	0.45	0.58	0.4
Mo	1.3	2.3	3.4	3.7	5.5
Cu	3.9	15.6	3	2.6	1.6
Pb	11.9	8	13.1	12.3	26
Zn	25	39	43	36	26
Ni	9.9	35.8	2.7	2.5	3
As	0.5	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Ag	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Au	b.d.l.	0.9	6	0.7	2.2
F	340	1644	1653	1796	1511

b.d.l. = below detection limit

* indicates an averaged duplicate

oxides are reported in weight percent, trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

Sample	10PHA-J29	10PHA-J33A	82TX-S188	10PHA-J24A*	10PHA-J26A
Locale	Amer plutonic complex	Amer plutonic complex	Amer plutonic complex	Amer plutonic complex	Amer plutonic complex
Lithology	granite	granite	granite	aplite	aplite
SiO ₂	72.5	74.4	73.1	76.25	75.4
Al ₂ O ₃	12.75	11.97	12.85	12.155	12.76
CaO	0.97	0.93	1.23	0.595	0.25
K ₂ O	5.77	5.81	5.55	5.785	5.2
MgO	0.39	0.31	0.42	0.08	0.02
MnO	0.04	0.03	0.04	0.01	b.d.l.
Na ₂ O	2.90	2.55	2.95	2.96	3.99
P ₂ O ₅	0.08	0.06	0.09	b.d.l.	b.d.l.
Fe ₂ O _{3t}	2.04	1.73	1.94	0.88	0.87
TiO ₂	0.31	0.33	0.46	0.12	0.05
Ba	634	496	990	141	213
Be	7	4	4	5	4
Cs	3.2	4	4.1	2.8	1.9
Ga	17.4	15.8	18.4	16.3	17.6
Hf	8.3	7.4	5.9	5.4	5.8
Nb	20.3	18.4	20.4	11.7	32.8
Rb	369.4	356.3	362.7	306.5	278.2
Sn	2	2	2	b.d.l.	b.d.l.
Sr	163.6	147.8	207.9	66.3	65.3
Ta	1.5	1.6	1.9	1.1	2.2
Th	72.1	92.9	53.6	56.6	21.6
U	15.2	16.2	6.8	11.1	11.8
W	1.7	2.7	3.0	1.0	1.0
Zr	217.5	199	169.4	121.5	80.8
Y	20.7	24.3	24.9	7.3	18
La	99.4	138.4	91.5	34	5.4
Ce	173.6	236.4	174.6	53.9	13.7
Pr	15.85	20.13	16.75	4.44	0.89
Nd	49.7	62.7	57.7	12.4	3.4
Sm	6.47	7.59	7.32	1.56	1.25
Eu	1.01	0.98	1.26	0.29	0.18
Gd	4.5	5.2	5.26	1.08	1.61
Tb	0.6	0.71	0.71	0.18	0.38
Dy	3.55	4.02	4.11	1.02	2.66
Ho	0.68	0.77	0.77	0.26	0.59
Er	2.05	2.42	2.4	0.86	1.95
Tm	0.35	0.4	0.41	0.17	0.38
Yb	2.55	2.78	2.89	1.34	3.01
Lu	0.41	0.46	0.46	0.27	0.56
Mo	4.7	6.5	4.1	1.2	0.7
Cu	2.3	1.1	1.3	0.6	3.2
Pb	21.6	24.5	18.2	21.9	4
Zn	23	23	27	6	3
Ni	2.1	3.1	2	1.7	1.4
As	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Ag	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.1
Au	3.3	1.1	0.7	4	2.9
F	1606	1520	1368	347	760

b.d.l. = below detection limit

* indicates an averaged duplicate

oxides are reported in weight percent, trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

Sample	10PHA-J33B	09PHA-J05	09PHA-J06
Locale	Amer plutonic complex	Nueltin Lake	Nueltin Lake
Lithology	aplite	aplite	pegmatite
SiO ₂	75.2	70.11	49.94
Al ₂ O ₃	12.5	13.59	11.86
CaO	0.36	1.61	5.88
K ₂ O	6.28	5.926	2.549
MgO	0.01	0.33	0.79
MnO	b.d.l.	0.042	0.199
Na ₂ O	3.2	2.90	0.59
P ₂ O ₅	b.d.l.	0.155	0.008
Fe ₂ O _{3t}	0.77	2.779	9.674
TiO ₂	0.05	0.335	0.750
Ba	173	538	291
Be	5	4	4
Cs	3.9	5.7	3.8
Ga	21.2	20.1	37.4
Hf	8.3	37.5	4.2
Nb	45.1	17.8	12.7
Rb	460.8	330.8	130.7
Sn	b.d.l.	3	43
Sr	63.1	180.9	490.9
Ta	3.4	0.8	1.2
Th	22.3	787.7	8461.4
U	11.2	56.8	589.3
W	4.1	0.8	5.5
Zr	85.5	1107	90.3
Y	23.9	44	886.5
La	7.5	111.9	21949.4
Ce	16.6	169.5	43927.7
Pr	1.36	17.8	4372.15
Nd	5.2	62.7	11559**
Sm	1.58	9.17	1443.31
Eu	0.28	1.39	36.96
Gd	1.93	7.47	686.17
Tb	0.52	1.2	70.19
Dy	3.91	6.49	247.74
Ho	0.93	1.39	31.93
Er	3.2	4.19	74.89
Tm	0.68	0.64	10.26
Yb	5.68	4.47	69.31
Lu	1.01	0.75	10.12
Mo	2.8	0.6	0.4
Cu	1.6	0.4	0.5
Pb	5.3	104.7	1266.4
Zn	2	62	85
Ni	2.2	0.2	0.4
As	b.d.l.	b.d.l.	30
Ag	0.5	0.1	0.3
Au	3.5	b.d.l.	6.1
F	808	2950	9740

b.d.l. = below detection limit

** indicates trace element value determined by XRF

oxides are reported in weight percent

trace elements are reported in parts per million (ppm) and Au reported in parts per billion (ppb)

APPENDIX C: BULK CONTINENTAL CRUST NORMALIZATION VALUES

Element	Bulk Continental Crust (Taylor and McLennan, 1995)
Cs (ppm)	1.0
Rb (ppm)	32
Ba (ppm)	250
Th (ppm)	3.5
U (ppm)	0.91
K (wt. %)	0.91
Nb (ppm)	11
Ta (ppm)	1.0
La (ppm)	16
Ce (ppm)	33
Sr (ppm)	260
Nd (ppm)	16
Hf (ppm)	3.0
Zr (ppm)	100
Sm (ppm)	3.5
Ti (wt. %)	0.54
Tb (ppm)	0.60
Y (ppm)	20
Tm (ppm)	0.32
Yb (ppm)	2.2

APPENDIX D: REE CHONDRITE NORMALIZATION VALUES

Element	Chondrite (Nakamura, 1974)
La	0.329
Ce	0.865
Nd	0.630
Sm	0.203
Eu	0.0770
Gd	0.276
Dy	0.343
Er	0.225
Yb	0.220
Lu	0.0339