THE GEOCHEMICAL AND MINERALOGICAL EVOLUTION OF THE MOUNT ROSA COMPLEX, EL PASO COUNTY, COLORADO, USA

by

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ABSTRACT

The ~1.08 Ga Pikes Peak Batholith is a type example of an anorogenic (A)-type granite and hosts numerous late-stage sodic and potassic plutons, including the peraluminous to peralkaline Mount Rosa Complex (MRC), located ~15 km west of the City of Colorado Springs in Central Colorado. The MRC is composed of Pikes Peak biotite granite, fayalite-bearing quartz syenite, granitic dikes, Mount Rosa Na-Fe amphibole granite, mafic dikes ranging from diabase to diorite, and numerous rare earth (REE) and other high field strength element (HFSE; e.g. Th, Zr, Nb) rich Niobium-Yttrium-Fluorine (NYF)-type pegmatites. The aim of this study is to trace the magmatic evolution of the Mount Rosa Complex in order to understand the relationship between peraluminous and peralkaline rock units and concomitant HFSE enrichment and mineralization processes. Field work, petrography, SEM-based methods, whole rock geochemistry, and electron probe micro-analysis (EPMA) of micas was performed on all rock units to determine their textural, mineralogical and geochemical characteristics.

Early peraluminous units such as the Pikes Peak biotite granite and fayalite-bearing quartz syenite contain annite-siderophyllite micas with high Fe/(Fe + Mg) ratios, and show relatively minor enrichments in REE and other HFSE compared to primitive mantle. Granitic dikes show a crude radial geometry around the inferred intrusive center, contain Al-rich siderophyllite mica, and are relatively depleted in HFSE. The Mount Rosa Na-Fe amphibole granite is very heterogeneous in outcrop (sill-like bodies, blobs, and dikes), as well as mineralogically and texturally variable, and includes large pegmatitic areas as well as coarse and fine-grained enclaves. End-member annite is enriched in Na, K, and F, and the Mount Rosa granite is enriched in REE and other HFSE compared to primitive mantle and earlier rock units. Mafic dikes contain biotite which is more Mg-rich versus other MRC biotites, and show a strong whole rock geochemical enrichment in REE, Zr and other HFSE. While Type-(I) pegmatites are mineralogically simple with irregular contacts, Type-(II) pegmatites are mineralogically complex and have sharp intrusive contacts. A sample transect across an ~80 cm thick, well-zoned Type-(II) pegmatite shows low F + Cl contents in biotite from the contact zone (0.30 a.p.f.u. Cl + F) with increasing concentration of F + Cl in micas towards the core. Biotite from the wall zone, which is mineralogically similar to the Mount Rosa granite, contains 2.5 a.p.f.u. F + Cl, whereas polylithionite-trilithionite from the core zone, which which is associated with cryolite (Na₃AlF₆)
pods, fluorite, and arfvedsonite pseudomorphs, shows extremely high concentrations of F + Cl (up to 7.0 a.p.f.u.)

While early units such as the Pikes Peak granite, fayalite-bearing quartz syenite, and granitic dikes appear to have a similar parental melt to the overall Pikes Peak Batholith, the Mount Rosa granite, with its peralkaline chemistry and strong outcrop, textural and mineralogical heterogeneity may be related to the separation of an immiscible peralkaline melt late in the magmatic evolution of the complex. Mafic dikes show characteristics of both early peraluminous units (whole rock enrichment in and mineralogy reflecting compatible elements such as P and Ba), as well as later peralkaline units (strong enrichment in incompatible elements such as Zr and REE; abundant fluorite and albite), and have a less clear magmatic relationship to either the Pikes Peak or Mount Rosa granite parental melts. Pegmatites are interpreted to represent the very last melt after the crystallization of Mount Rosa Na-Fe amphibole granite.

Three major alteration events were identified in the MRC: (I) early albitization, which is most intense around the Mount Rosa granite, (II) an acidic aegirinization/hematization stage which replaced Na-Fe amphibole with aegirine and all mafic silicates by Fe oxides, and (III) a hydrothermal Ca-F metasomatism stage in which magmatic minerals were altered and replaced by Ca-F minerals. Incompatible elements were likely re-mobilized by late F-rich hydrothermal fluids in stage III which precipitated minerals such as REE fluorcarbonates and hydrothermal zircon in Type-(II) pegmatites.
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CHAPTER 1
INTRODUCTION

Located ~15 km west of the city of Colorado Springs, the ~1.04 Ga Mount Rosa Complex (MRC) consists of a diverse group of peraluminous to peralkaline granitic rocks hosted in the ~1.08 Ga Pikes Peak Batholith. The MRC is enriched in rare earth elements (REE) and other high field strength elements (HFSE, e.g. Th, Ti, Zr, Nb and Ta) and is host to numerous Niobium-Yttrium-Fluorine (NYF)-type pegmatites. One of the most diverse Proterozoic igneous complexes in Colorado, the MRC is an excellent study object for investigating both the petrogenesis of the Pikes Peak Batholith as well as how REE and other HFSE are concentrated in peralkaline igneous systems. The genesis of the MRC is debated (Gross 1962, Gross and Heinrich 1965; 1966, Smith et al. 1999), and the transition from primary magmatic crystallization products to highly-complex pegmatites which show evidence of fluid-driven processes is poorly-understood. Through field work, petrography, SEM-based automated mineralogy, and SEM-EDX and EPMA analysis of select minerals, this study seeks to petrogenetically relate the diverse association of rock types in the Mount Rosa Complex and propose a mechanism for REE and HFSE enrichment in the NYF pegmatites.

1.1 Overview of anorogenic (A-Type) granites

Anorogenic granitic intrusions and their associated peralkaline granite complexes are some of the most highly geochemically fractionated and geologically enigmatic rocks on Earth. These rocks can shed light on processes of rare element enrichment from magmatic to hydrothermal, as well as concentrate such elements to the point of creating an economic deposit (e.g.; Strange Lake Canada, Pitinga Brazil, Khaldzan-Buregtey Mongolia). With the rise in demand for mobile electronics, sustainable energy alternatives, and high-tech ‘super-materials’, agencies including the European Commission and the US Department of Energy have designated certain elements as ‘critical elements’ of ‘strategic resources’ (Linnen et al. 2012). Of the 7 elements defined as ‘critical’ by the US Department of Energy, 5 are sufficiently enriched in some peralkaline
granites to be of economic interest (CMI factsheet; accessed 1-9-2017). As such, the study of A-type granites and associated peralkaline rocks, while still arguably in its infancy, is important for both economic and scientific reasons.

Anorogenic granites, also known as ‘A-type granites’ occur through the geologic record, though most known granites hosting significant rare element deposits occur in Precambrian terranes (Bonin 2007). A-type granites are typically found in quiescent, intra-plate tectonic environments subject to extension, though some also occur at plate boundaries (Bonin 2007). They are believed to have been formed by extreme fractionation of mantle-derived mafic to intermediate magmas (Fig. 1.1), with or without crustal assimilation (Frost and Frost 1997; Smith et al. 1999). Compared to much more abundant sedimentary (S)- and igneous (I)-type granites, A-type granites have very minor surficial expression and represent the result of ‘primitive’ processes which likely occurred in the deep Earth (Bonin 2007). Rare felsic igneous rocks found in meteorites and on other planets such as Mars show strong geochemical characteristics of A-type granites and further support the link between primitive magmas and A-type granites (Dann et al. 2001).

Geochemically, A-type granites are typically enriched in incompatible elements (e.g. HFSE and LILE) and show relative depletion in certain elements which are compatible in most granitic rocks (e.g. Ni, Cr, Co, Ba, Sr, Eu; Frost and Frost 1997). Mineralogically, A-type granites are characterized by Fe-rich mafic phases such as biotite, amphibole and pyroxene, and OH and F-rich fluids can lead to abundant late-stage fluorite and hydrated phases such as Fe-micas and amphiboles (Bonin 2007). Alkali feldspar textures often show a transition from hypersolvus to subsolvus features, and complex textural heterogeneity resulting from mingling of different magmas and localized volatile enrichment are very common (Bonin 2007, Smith et al. 1997, 1999). Likewise, A-type granite intrusions often show an impressive diversity of rock types, including but not limited to syenites, fayalite-bearing granite, Na-Fe amphibole /pyroxene granites, albitites, mafic dikes ranging from diabase to lamprophyre, pegmatites, gabbros, anorthosites, and various eruptive equivalents where an intrusion breaches the surface and is associated with volcanism (Eby 1992, Bonin 2007).

The origin of A-type granites is still a topic of contentious debate, however, several accepted theories have emerged. The first is that crustal thinning in intra-plate extensional regimes, potentially related to post-orogenic crustal relaxation, leads to upwelling of magma in the upper
mantle (Clemens et al. 1986, Smith et al. 1999; Bonin 2007). This upwelled mantle magma in turn melts and incorporates metasomatized lower crustal material as it ascends through the crust, ultimately to undergo prolonged assimilation-fractionation-crystallization (AFC) processes (Fig. 1.1) at mesozonal to epizonal crustal levels (Smith et al. 1999, Bonin 2007). While some argue that A-type granites are completely anhydrous (A-type magmas have the lowest water fugacities observed in silicate magmas), this first petrogenetic model argues this is somewhat misleading, as while the parental magmas may have been anhydrous, the result of highly-efficient AFC processes are smaller volumes of melt which crystallize lithologies rich in hydrous phases (e.g. riebeckite-biotite granites; Bonin 2007).

Figure 1.1 Structural levels and fractional crystallization scheme of A-type granite magmatism (after Bonin 2007; 1986, 1996)
1.2 Overview of peralkaline granites

While volumetrically far outnumbered by metaluminous to peraluminous rocks, peralkaline granites (defined as having a wholerock Na$_2$O/K$_2$O > Al$_2$O$_3$) represent an important lithology in A-type granites, as they often show the most pronounced enrichment in HFSE and other incompatible elements, and reflect complex processes of magma evolution and geochemical fractionation. Peralkaline granite complexes often exhibit a diverse array of rock types and textures, including late-stage lithologies such as pegmatite dikes in which incompatible elements can reach enrichment levels otherwise unheard of in crustal rocks (Chakhmouradian and Zaitsev 2005). In terms of their relationship to larger plutons and batholiths, it generally agreed that peralkaline granites postdate their metaluminous to peraluminous host rocks, often forming the cores of ring complex (e.g; Lake George Ring complex Colorado, Nigerian Younger granites; Erongo and Brandberg complexes, Namibia; Smith et al. 1999, Sakoma et al. 2000, Trumbull et al. 2004). However, peralkaline granite dikes and tabular, sheet-like sills can also intrude metaluminous to peraluminous host rocks (Tulloch 1992).

While some peralkaline granites occur as discrete plutons or dikes which have intruded epigenetic rocks, most are spatially and temporally related to a larger volume of metaluminous to peraluminous magma (Smith Et al. 1997, 1999; Costi et al 2009). The origin of such peralkaline granites is poorly understood, and present theories involve 1.) input of a separate, mantle-derived magma batch which ascends through the crust, like the proposed genesis of A-type granites, or 2.) the development of a peralkaline melt from a much larger volume of metaluminous to peraluminous melt through immiscibility processes in the last stages of magma evolution in a pluton (Smith et al. 1997, 1999; Thomas et al. 2006). The first theory is like that previously discussed for the generation of A-type magmas. The second theory is based on evidence from melt and fluid inclusions as well as limited experimental data and proposes that at the end stages of magmatism in a peraluminous pluton, when most melt has crystallized and the residual liquid (possibly in a ‘crystal mush’) is enriched in volatiles (e.g. F, OH), two different melts, one peralkaline and the other peraluminous, can be generated through immiscibility processes (Thomas et al. 2006). Thomas et al. (2006) argue that one reason peralkaline granites are rare is that the peralkaline melt fraction is rarely preserved in nature because the extreme volatile enrichment of this melt catalyzes reactions between surrounding peraluminous magma as well as
wallrocks such as loss of alkalis by vapor-phase separation and intense greisenization. When preserved, the peralkaline melt phase is highly enriched in Na, K, Cl, P, B, F, Ca, Fe, and HFSE and depleted in Si and Al compared to the peraluminous phase (Thomas 2005). Reactivity of this peralkaline melt phase as well as the prolonged fractionation needed to achieve separation of such a phase may explain the textural and mineralogical heterogeneity of many peralkaline granite complexes (e.g.; Strange Lake Canada, Pitinga Brazil, Ilimaussaq Complex, Greenland; Thomas et al. 2006). Additionally, melt inclusions in minerals in studied peralkaline granites often contain rare minerals such as cryolite, kalsilite, Y-fluorite, REE fluorocarbonates, and zirconosilicate minerals which suggest that the immiscibility processes which may have driven the phase separation of a peralkaline melt continued to concentrate rare and volatile elements into the final residual melt (Thomas et al. 2006).

Many peralkaline granites are associated with coeval volcanic rocks (e.g.; Moat Mountain volcanics New Hampshire, Nigerian Younger granites, Amis Complex Namibia) and may even occur as subvolcanic intrusions which have ‘ponded’ under a thick volcanic pile (e.g.; Wichita Mountains Oklahoma; Hogan et al. 2000). This coeval volcanic association and epizonal (~5-10 km) emplacement depth of many peralkaline granites has also been used to partly explain the diversity of lithologies and mineralogical heterogeneity of peralkaline granites. Some workers have also proposed that mixing of a mafic (gabbroic to basaltic) and felsic melt either at depth or at shallower emplacement levels is critical to explaining observed geochemistry and field relationships (Zanvilevich et al. 1995). Others have proposed that aeromagnetic and gravity anomaly data of some peralkaline granite complexes suggests gabbroic cumulates at depth which may represent residual mafic material after fractionation of a mantle-derived melt which yielded a peralkaline granite (Barker et al. 1975). Some have even compared evolved peralkaline granites such as the Strange Lake Complex in Canada to highly-differentiated peralkaline rhyolites such as the Pico Alto volcano, Azores Islands, and proposed that magmatic processes in a pantelleritic magma combined with separation of an orthomagmatic fluid could generate similar enrichment to early granitic phases at Strange Lake (Mungall and Martin 1996). Regardless of the mechanism invoked, it is clear that the origin of peralkaline granites will require further research to properly elucidate.
As previously mentioned for A-type granites in general, peralkaline granites typically contain a diversity of alkali feldspar textures, related to degrees of melt polymerization, undercooling, and volatile saturation (Nassif 1993). Typically, the least evolved unit, an early hypersolvus granite, crystallizes first, and contains only one perthitic alkali feldspar, in addition to biotite ± Na-Fe amphibole/pyroxene. Next, a so-called ‘transolvus’ granite with more pronounced K-feldspar phenocrysts with perthite and possibly albite, though this albite is typically metasomatic and not primary (Greenberg 1981, Nassif 1993). Finally, a subsolvus granite crystallizes, typically under conditions of elevated water fugacity and enrichment in other volatiles such as F and CO$_2$. This subsolvus granite is the most evolved primary magmatic unit and is characterized by the presence of two primary feldspars: albite and K-feldspar (Greenberg 1981, Nassif 1993).

Another common feature of peralkaline granites are large areas of granophyre or pegmatitic textured granite, which may be surrounded by or in contact with fine-grained microgranite or aplite: these textural features are thought to represent a late-stage residual, undercooled melt which may locally become supersaturated in volatiles and, at lower pressure epizonal levels, expel these volatiles into surrounding melt, triggering rapid crystallization which may produce marked textural differences over small areas (Miller 1996; London 2005). These same concepts of late-stage melt evolution and fluid separation have been applied to granitic pegmatites by London (1992; 2005). Such features may also be seen at the thin section scale in more evolved units in peralkaline granites and likely represent a similar interplay of fluid-rock interaction, volatile saturation, and localized melt undercooling (Vernon 2004).

1.3 Overview of pegmatites

A special feature of A-type granites and their associated peralkaline granites is the incredible diversity of pegmatites they often contain. The term pegmatite, while generally accepted as a textural descriptor only, generally applies to coarse-grained felsic igneous rocks which have been subjected to some degree of melt undercooling driven by volatile content (London 1992; 2005, Cerny and Ercit 2005). Pegmatites occur in diverse tectonic and petrogenetic settings, and pegmatitic textures and processes can occur in mafic and ultramafic intrusive rocks as well (London 2005).
The geochemistry and mineralogy of pegmatites is now understood to be related to tectonic setting and in turn specific types of granite (Martin and De Vito 2005; Cerny and Ercit 2005). Epigenetic pegmatites, or those that intrude unrelated host rocks, have been grouped by their P-T environment as well as chemistry (Cerny and Ercit 2005). The classification scheme proposed for epigenetic pegmatites by Cerny and Ercit (2005) has 5 subdivisions: abyssal, muscovite, muscovite-rare element, rare-element, and miarolitic. These 5 classes of pegmatites have been further broken down by chemistry and mineralogy into subclasses, types, and finally subtypes, all based on observations of natural pegmatites. While the 5 classes of epigenetic pegmatites are partly based on the P-T fields of regional host rocks, it is important to note that these P-T conditions do not necessarily reflect that of the pegmatites themselves upon emplacement and cooling (Cerny and Ercit 2005). For pegmatites hosted in and related to larger granitic plutons, a primary subdivision into NYF (Niobium-Yttrium-Fluorine) and LCT (Lithium-Cesium-Tantalum) pegmatites exists (Martin and De Vito 2005). LCT pegmatites are typically associated with peraluminous to metaluminous I (igneous protolith) and S (sedimentary protolith) type granites and are characterized by Ta > Nb and elevated Li, Rb, Cs, Be, Sn, B, and P (Martin and De Vito 2005). NYF pegmatites, in contrast, are typically associated with peraluminous to peralkaline A-type granites, and are characterized by Nb > Ta and elevated Ti, Y, Sc, REE, Zr, U, Th, and F (Cerny and Ercit 2005). Some pegmatites, however, especially highly evolved bodies exhibit attributes of both NYF and LCT pegmatites (e.g.; Sahatany Valley Madagascar, Malkhan field, Russia) and thus, some workers argue that this classification system is in need of revision or at least caution (London 2005).

The study of pegmatitic textures has been ongoing for 100+ years and is still a matter of debate. Brogger (1890) was the first to propose that pegmatitic textures are due to the interaction of coexisting water vapor and silicate melt. This idea was further refined by Jahns (1955) and later Jahns and Burnham (1969) who used experimental evidence to support their idea that alkali elements can be preferentially extracted into a vapor phase once phase separation has occurred in a volatile-enriched silicate melt, thus triggering crystallization of one mineral (e.g., albite in the case of loss of K into the vapor phase). Only the ‘failed experiments’, however, in which the capsule leaked and Si, Na and K were redistributed throughout the capsule by vapor-phase transfer, actually reflected observations of pegmatite texture and mineralogy in nature (London 2005). While Jahns’ model did not account for all textural and mineralogical heterogeneities
typically seen in pegmatite dikes, it was the first significant, experimentally-based approach at decoding pegmatite formation.

The ‘new generation’ of pegmatite research, led by David London, William ‘Skip’ Simmons, and others, has focused largely on constraining the P-T conditions of typical pegmatite formation, closed vs open-system behavior, and the role of volatiles and incompatible elements in explaining textural and mineralogical heterogeneity as well as patterns of zonation in pegmatites. Traditionally, it was thought that pegmatites crystallize near the water-saturated granite solidus of ~650-700 ºC (London 2005). Jahns (1955) was the first to suggest that pegmatites may actually crystallize below 550 ºC, and London later revised that estimate downwards to ~450 ºC for highly evolved pegmatites, such as the Li-enriched Tanco pegmatite in Canada (London 2005). Later work on feldspar geothermometry and stable isotopes by Morgan and London (1999) and Sirbescu and Nabelek (2003) proposed crystallization at highly evolved LCT-type pegmatites down to ~340 ºC, and Thomas (1994) even suggested that at temperatures as low as 262 ºC in the Tanco pegmatite, silicate melt existed. This extreme ‘undercooling’ or subsolidus crystallization in pegmatites, unheard of in other silicate melts, reflects the highly fluxed parental melts which form pegmatites (London 2005).

Using inferred liquidus and host-rock temperatures, Chakoumakos and Lumpkin (1990) found that the ~20 m thick Harding LCT pegmatite in New Mexico likely cooled in 3-5 months from emplacement. Morgan and London (1999) found that the ~2 m thick Little Three pegmatite in California may have cooled in just ~25 days. Compared to traditional thinking of slow pegmatite crystallization and time needed for growth of huge crystals, this is remarkable. This new thinking about crystallization timescales, combined with laboratory-based capsule experiments, have also led to new thinking about how mineral zonation and textural patterns develop in pegmatites. The delay in crystal nucleation and growth caused by significant melt undercooling leads to pegmatitic textures, but it one of several factors in the creation of such textures (London 2005). Other factors include a change from crystal nucleation in the melt to sidewall (e.g., already crystallized border zone) nucleation only, as well as significant temperature gradients between wallrocks, crystallized melt, and still liquid melt (London 2005). Constitutional zone refining, a metallurgical term in which an advancing crystallization front creates a thin ‘boundary layer’ enriched in volatile and incompatible elements ahead of the front, has also been shown to be very important in pegmatite crystallization (London 2005).
development of this flux-rich front may be key in explaining how pegmatites whose bulk composition is the same as their host granite contain zones which are highly-enriched in rare element minerals and other exotic phases (Morgan and London 1999). Finally, the importance of disequilibrium processes in pegmatite crystallization cannot be overemphasized. Abrupt localized changes in temperature, pressure, volatile content, and viscosity all exhibit a complex interplay in pegmatite melts and likely explain the incongruities we see in pegmatite zonation (London 2005).

1.4 Geological setting and previous work

The following section describes the geological setting of the Mount Rosa Complex, beginning with an overview of the Pikes Peak Batholith, and continuing with its relationship to the MRC. An overview of the pegmatites of the MRC and the previous work on these pegmatites follows this.

1.4.1 Overview of the Pikes Peak Batholith

The Pikes Peak batholith (PPB) is one of several recognized A-type ~1.1 Ga granites in western North America. These include the Red Bluff Suite and Franklin Mountains granite in western Texas and the Llano uplift of central Texas (Smith et al. 1999). The 1.08 Ga Pikes Peak batholith is a ‘type example’ of an A-type granite and shows the classic geochemical, tectonic and petrologic attributes of this family of granites such as high Fe/(Fe + Mg) and K₂O, and evidence of having originated from dry, reduced magmas (Loiselle and Wones 1978; Smith et al. 1999). The PPB has ~3300 km² of surface exposure, which has been aided greatly by Laramide (≤70 Ma) uplift/exhumation and subsequent glaciation (Smith et al. 1999). The PPB (Fig. 1.2) is a composite pluton and while the dominant lithologies are coarse grained biotite ± amphibole syenogranite and monzogranite, minor lithologies include gabbro, diabase, syenite, fayalite and Na-Fe amphibole granite (sodic series) and biotite granite and quartz monzonite (potassic series; Smith et al. 1999).

Numerous small, late-stage plutons which collectively comprise <10% of the exposure area have been divided into two geochemical groups, a potassic series, and a sodic series (Barker
et al. 1975, Wobus 1976, Smith et al. 1999). Recognized late-stage intrusions of the sodic series include the Sugarloaf, West Creek, Taryall and Spring Creek plutons, the Mount Rosa complex, and part of the Lake George ring complex (Fig 1.2; Smith et al. 1999). Late-stage intrusions of the potassic series include the Redskin Stock, part of the Lake George Ring Complex, Devil’s Head, Pikes Peak, and Wigwam Creek (Smith et al. 1999). Mapping of primary flow structures and fracture systems in the PPB has led to the identification of three main ~20-25 km wide intrusive centers in the batholith; the Buffalo Park, Lost Park, and Pikes Peak centers (Hutchinson 1960, 1976; Fig. 1.2). Barker et al. (1975) characterized the PPB as having been emplaced at shallow depth (<10 km) at ~980-1080 Ma, with sharp intrusive contacts with older metamorphic and igneous rocks. While these intrusive characteristics are consistent with A-type granite magmatism, the general ~980-1080 Ma age range proposed by Barker et al. (1975) and others (Aldrich et al. 1957; Gross and Heinrich 1965) has been a matter of debate.

While the general age for the PPB is ~1080 Ma (Smith et al. 1999), Marshall and DePaolo (1982) published an age of 1041±32 Ma for the main phase syenogranite of the PPB using the K/Ca geochronometer. Unruh (pers. communication 2015) obtained U/Pb zircon ages ranging from ~1070 Ma for late-stage potassic series plutons, and ~1087 Ma for sodic plutons. However, quality of the zircon crystals was a concern in some of these analyses (Unruh; personal communication 2015). Most recently, Guitreau et al. (2016) analyzed zircons from different lithologies in the potassic to sodic Lake George Ring Complex and found that the core syenomonzonite, which they interpreted as a roof pendant lens, had an age of 1428±8 Ma and was isotopically identical to PPB wallrocks of the ~1.4 Ga Berthoud Plutonic suite. Other ages obtained in this study for Lake George Ring Complex rocks ranged from ~1066-1115 Ma, and thus the authors call for a reinvestigation of so-called ‘late-stage’ plutons in the PPB as well as the currently accepted emplacement model. It must be emphasized however that the data from this and other studies on PPB geochronology comprise a very limited dataset and the difficulty of using PPB zircon for U/Pb geochronology has been well-demonstrated (Dan Unruh and Wayne Premo personal communication 2016).
Figure 1.2. Overview of the Pikes Peak Batholith showing late-stage potassic and sodic plutons, with Mount Rosa Complex shown (modified from Guitreau et al. 2016; after Wobus and Anderson 1978).
The Pikes Peak batholith and its temporal equivalents in Texas were emplaced at ~1.08 Ga into the Yavapai province, a ~500 km wide belt which was accreted onto the southern edge of the Archean Wyoming craton (Fig. 1.3) around ~1.7 Ga (Smith et al. 1999). Some workers have proposed that PPB evolved from extreme fractionation of an upper mantle-derived basaltic melt which ascended through underplated crust in an intra-plate extensional tectonic setting, which may have been due to post-orogenic collapse and ‘localized’ trans-tension and extension following the proposed ~1.4 Ga Berthoud orogeny (DePaulo 1981). Perhaps a more plausible tectonic setting is the onset of the extension and crustal thinning associated with the mid-continent rift (Fig. 1.3) around ~1.1 Ga (Smith et al. 1999). The geometry of ~1.1-1.2 Ga igneous activity of A-type affinity in Colorado, New Mexico and Texas supports this interpretation, as does the modern geometry of A-type magmatism occurring near the African Rift (Smith et al. 1999).

Figure 1.3 Simplified map of Precambrian provinces of North America, modified from Smith et al. (1999). Black objects are related to ~1100 Ma A-type magmatism and include: Mid-Continent rift (MRC), Llano uplift, Pecos Mountains Complex (PMC), Franklin Mountains (FM), and the Pikes Peak Batholith (PPB). SOA = Southern Oklahoma Aulacogen.
The melt generated during this crustal extension likely interacted with metasomatised middle to lower crust which had been affected by a preceding alkaline fluid which ‘fertilized’ it, a process that may explain the mixed geochemical patterns seen in some studies of the Mount Rosa/Pikes Peak granites and other ‘A-type’ granites such as the ~1.1 Ga Red Bluff and Llano granites in Texas (Smith et al. 1999; Martin and DeVito 2005). Barker et al. (1976) proposed that gabbroic cumulates left from the extreme fractionation that yielded the Pikes Peak batholith magma may exist at depth below the Mount Rosa granite on the basis gravity anomaly data and analogous mafic rocks in the Mount Rosa area. The most recent isotopic data from Guitreau et al. (2016) suggest the main source for the PPB were shallow mantle-derived melts with input of Proterozoic Colorado Front Range crust, with the crustal signature prevailing over time. This is in general agreeance with the data of DePaulo (1981) and Smith (1999) who proposed that crustal thinning beginning at ~1.1 Ga triggered upper mantle partial melting, which trapped basaltic magmas in the lower crust. This further catalyzed anataxis of the tonalitic lower crust, which in turn led to a more felsic melt which, through extensive and long-lived ACF processes (Fig. 1.1), produced large volumes of epizonal potassic granite which characterize the PPB (Smith et al. 1999).

The Pikes Peak Batholith is well-known for its diversity and abundance of pegmatites (Simmons and Heinrich 1980, Kile and Foord 1998, Hanson and Zito 2014). These pegmatites are typically related to late-stage plutons of the aforementioned sodic or potassic series (Smith et al. 1999). They include several distinct populations: miarolitic-type pegmatites related to late-stage potassic plutons (e.g.; Lake George Ring Complex, Redskin stock), the South Platte pegmatite district in the northern part of the batholith, and the pegmatites of the Mount Rosa Complex. The miarolitic-type pegmatites consist of numerous small, generally poorly-zoned bodies characterized by open pockets hosting euhedral crystals of rock-forming minerals (e.g.; quartz, amazonitic microcline, albite, fluorite) as well as accessory phases (e.g.; zircon, columbite, phenakite, topaz) and are often associated with late-stage potassic plutons (Hanson and Zito 2014). These pegmatites are believed to be related to residual, volatile-enriched melt in the cupola or exocontact zones of potassic plutons, such as the F and Al-rich Redskin stock, which is known for miarolitic pockets containing gem-quality topaz crystals (Kile and Foord 1998; Desborough et al. 1980). The bulk chemistry of the pegmatites is similar to that of their
host rocks, and the process of miarolitic pocket formation which has been modeled elsewhere can be applied to these pegmatites (Kile and Foord 1998).

The South Platte Pegmatite District comprises ~50 mapped NYF-type pegmatites which are generally large (up to ~100 x 75 m.), well-zoned, and contain significant REE mineralization (both in allanite-monazite subtype and euxenite-gadolinite subtype pegmatites) as well as F, Ti, Nb, and Fe (Simmons and Heinrich 1980; Simmons et al. 1987). The South Platte pegmatite District in located in the northern portion of the Pikes Peak batholith and is not believed to be associated with late-stage plutons (Simmons and Heinrich 1980; Gagnon et al. 2004). Similar to the miarolitic-type pegmatites, their genesis is believed to be related to volatile-rich residual melt trapped at epizonal levels in the batholith which was undercooled and significantly affected by both autometasomatism and well as late fluid interaction with wallrocks (Simmons and Heinrich 1980, Gagnon et al. 2004). The pegmatites of the Mount Rosa Complex show similarities to both zoned South Platte District NYF pegmatites as well as the miarolitic-type pegmatites, and includes a distinct group of pegmatites showing alkali mineralogy and geochemistry which is unique in the Pikes Peak Batholith and will be discussed in more detail below (Gross and Heinrich 1966, Hanson and Zito 2014).

1.4.2 Overview of the Mount Rosa Complex

The ~1.08 Ga Mount Rosa Complex (MRC) is located in the southeastern corner of the PPB, ~15 km West of the city of Colorado Springs (Figs. 1.2; 2.1). The MRC consists of a diverse group of peralkaline to peraluminous granitic rocks associated with the Mount Rosa intrusive center; the youngest intrusive center in the PPB. The Mount Rosa Complex is hosted in peraluminous biotite granite of the Pikes Peak intrusive center and consists of, from oldest to youngest, fayalite-bearing syenite, dikes and aplites of biotite syeno- to monzogranite, Na-Fe amphibole granite (Mount Rosa granite), mafic dikes ranging from diabase to diorite, and two different types of pegmatites (Fig. 2.1). The Mount Rosa Complex is enriched in rare earth elements (REE) and other high field strength elements (HFSE, e.g. Th, Ti, Zr, Nb and Ta) and is host to numerous Niobium-Yttrium-Fluorine (NYF)-type pegmatites.

Gross (1962) obtained several K$^{40}$/Ar$^{40}$ dates on rocks of the Mount Rosa granite ranging from 1020-1045 Ma, making it distinctly younger (but temporally-related to) the Pikes Peak
Batholith. Unruh (personal communication 2015) also obtained several Pb/U zircon dates on the Mount Rosa granite of ~1040 Ma. The Mount Rosa granite is one of the youngest dated units in the PPB, but the adjacent Windy Point granite as well as the Almagre Mountain pluton, both peraluminous granites, were actually found to be slightly younger than the Mount Rosa granite (Barker et al. 1975; Unruh and Primo personal communication 2015). However, their petrogenetic relationship to the Mount Rosa Complex is unclear and they were not included in this study. While Unruh and Primo (personal communication 2015) emphasized the variability of zircon quality for U/Pb geochronology in their samples, field relationships also suggest that the Windy Point and Almagre Mountains granites may be younger than the Mount Rosa granite (Gross 1962; Barker et al. 1975).

Gross (1962) characterized the Mount Rosa complex as a series of many generally-small sill and dike-like bodies of various igneous rocks occurring in an arc-like ~15 x 10 km area. He described the different units, from oldest to youngest, as fayalite granite, porphyritic granite, granitic/aplithic dikes, Mount Rosa Na-Fe amphibole granite, mafic dikes, ‘interior’-type pegmatites, and ‘exterior’-type pegmatites, and concluded that all units were comagmatic. Gross (1962) also concluded that ‘interior’-type pegmatites were related to Pikes Peak granite magma, whereas the ‘exterior’-type pegmatites were related to the Mount Rosa granite magma. Barker et al. (1975) echoed Gross and Heinrich (1965)’s observations on the general geology of the MRC, but proposed a more specific model for the fayalite granite, suggesting a mantle–derived alkali olivine magma reacted with K and Na-poor granulite rocks in the lower crust, in turn creating a syenite magma. This syenite magma then ascended into the intermediate and upper crust, evolving through AFC processes into a fayalite granite magma now enriched in Na, K, Fe, F, and HFSE (Barker et al. 1975). Finally, Barker et al. (1975) propose that through additional fractional crystallization and differentiation of this syenite magma, the riebeckite granite (Mount Rosa granite) formed. Giambalvo (1993) and Goldman et al. (2012) examined the fayalite-bearing quartz syenite and found it to be relatively chemically homogeneous, with SiO2 ranging from 67-74 wt. % and high overall REE with light rare earth element (LREE) enrichment and a negative Eu anomaly. The fayalite-bearing granites also contain amphiboles that plot in a distinctly different chemical field (ferro-edenite) than those from the main Mount Rosa Granite (riebeckite-arfvedsonite), a compositional gap which raises questions on whether the rocks of the Mount Rosa area evolved from a single parental melt (Giambalvo, 1993). Smith et al. (1999)
and Douglass and Smith (1993) conclude that the sodic late-stage plutons such as the MRC formed mainly by the fractionation of mantle-derived magmas, compared to the late-stage potassic plutons which formed mainly by crustal anataxis and crystallization from the main peraluminous magma which comprises >90% of the batholith.

A 2012 study by Goldman et al. on zircon crystals from the Mount Rosa granite suggested near-liquidus temperatures ranging from 865°C to 966°C, and formation pressures of at least 8 kilobar, indicative of fairly shallow (<10 km) crystallization depth. This is in agreement with the estimated relatively high crystallization temperatures (up to ~950 ºC) for the Mount Rosa granite inferred by Smith et al (1999). Mineralogically, the main phase of the granite is dominated by quartz, microcline/perthite, oligoclase (albite-dominant), annite-biotite (Fe-mica), and Na-Fe amphiboles/pyroxenes, with Na-Fe amphibole >> Na-pyroxene (Gross 1962). Other phases such as the fayalite granite contain the aforementioned major minerals in addition to subordinate amounts of fayalite, muscovite, astrophyllite, zircon, aegirine, fluorite, and accessory phases (including rare element minerals; Gross 1962, Giambalvo 1993; Goldman et al. 2012).

Previous workers (Gross and Heinrich 1966, Foord et al. 1984, Hanson and Zito 2014) have also noted evidence that rocks of the MRC as well as the surrounding Pikes Peak biotite granite have undergone extensive fluid-driven alteration. In the earliest stage of observable granite alteration, Gross (1962) found that in both the Mount Rosa granite and pegmatites, microcline and oligoclase have been extensively replaced by albite ranging from veinlets to areas where only small ‘islands’ of primary feldspar remain. He also noted that primary crystals of Na-Fe amphibole in the Mount Rosa granite were often rimmed by Na-pyroxene (aegirine) which had replaced it, and that primary astrophyllite was often altered to a white clay-like mineral. Further alteration of aegirine and Na-Fe amphibole led to the formation of magnetite and Fe-oxides (e.g.; limonite, hematite; Gross 1962). Additionally, in some pegmatites crystals of Na-Fe amphibole often show ragged edges and replacement by Na-pyroxene (aegirine) which is in turn replaced by biotite and magnetite and further alter to Fe-oxides and clays (Foord et al. 1984). This is the reverse of the amphibole replacement reaction observed in the surrounding granite by Gross (1962). Micas (biotite, annite and zinnwaldite) were also found to be commonly replaced or overgrown by other minerals, with corroded grain boundaries (Gross 1962). More recently, quartz-rutile pseudomorphs after narsarsukite crystals have been found in pegmatites of the St. Peter’s Dome area of the MRC (Pavel Kartashov, personal comm. 2014). Such pseudomorphs
occur in similar alkali granites in Western Mongolia, Kazakhstan, and Greenland in rocks affected by late fluid alteration (Pavel Kartashov, personal comm. 2014).

Figure 1.4. Overview of key places in the Mount Rosa Complex referenced in the text; photo taken June 2016 from the summit of Mount Rosa (el. 3504 m) looking southeast towards Colorado Springs (el. ~1840 m).

1.4.3 Overview of Pegmatites of the Mount Rosa Complex

The Mount Rosa Complex is noteworthy for its large number of pegmatites, particularly in the St. Peter’s Dome and Stove Mountain area (Figs. 1.3, 1.4), ranging from small unzoned miarolitic cavities >1 m across to large zoned Niobium-Yttrium-Fluorine (NYF)-type bodies over 40 meters long (Gross and Heinrich 1966). Cerny and Ercit (2005) classified the pegmatites of the MRC as belonging to the Peralkaline NYF-type and Na-Fe amphibole subtype, similar to pegmatites on Hurricane Mountain, New Hampshire, or the Franklin Mountains of western Texas. Gross (1962) and Gross and Heinrich (1966) classified the pegmatites of the MRC into two groups,
one occurring on the margins of the Mount Rosa granite enclosed in Pikes Peak granite (‘exterior’ type), and the other related to the Mount Rosa granite and occurring in contact with or fully enclosed in Mount Rosa granite (‘interior’ type). Many of the ‘exterior’ type pegmatites show mineralogical and geochemical characteristics of the Mount Rosa peralkaline granite and thus were inferred by earlier workers to represent pegmatites overprinted by fluids associated with the intrusion of the Mount Rosa granite (Gross and Heinrich 1966; Foord et al. 1984). The third type of pegmatites in the Mount Rosa area are the ‘Pikes Peak-type’ pegmatite bodies (Fig. 2.1), which occur in Pikes Peak granite and are generally small (<2 meters), irregular, poorly-zoned bodies, and often contain miarolitic pockets with euhedral crystals (Gross and Heinrich 1966; Hanson and Zito 2014). These are thought to be genetically-similar to other miarolitic pegmatites of the Pikes Peak granite, but differ in their accessory mineral content, including euhedral crystals of rare element species such as the Zn-Be silicate genthelvite, REE-F carbonates (e.g. bastnäsite, fluocerite), and REE phosphates (e.g., xenotime-(Y) and monazite-(Ce); Hanson and Zito 2014). Many of these minerals replace or overgrow primary minerals such as feldspars, pyroxenes/amphiboles, and micas (Hanson and Zito 2014). In this regard, these miarolitic-type pegmatites show geochemical similarities to those of the Mount Rosa-type, and it has been proposed that they may have been affected by fluids associated with the crystallization of the Mount Rosa granite (Gross and Heinrich 1966, Foord et al. 1984; Hanson and Zito 2014).

According to Gross (1962) and Gross and Heinrich (1966), pegmatites showing chemical and mineralogical attributes of the Mount Rosa Granite though hosted by the Pikes Peak Batholith (“exterior type”) are larger, more structurally-uniform, and more mineralogically-complex than those pegmatites hosted by the Mount Rosa granite. These pegmatites contain significant Be, Rb, Y, F, Zr, and Ce, with Ta >Nb, and sometimes contain ‘exotic’ minerals such as Al-Fluorides (e.g. cryolite, weberite, prosopite) found only in a few other localities in the world such as Ivigtut, Greenland and the Pitinga Mine, Madeira Granite, Brazil (Gross and Heinrich 1966, Costi et al. 2009). Other unusual accessory minerals containing REE, HFSE, F, Na, and OH occur in the ‘exterior-type’ pegmatites, including murataite-(Y), Zn and Y-bearing senaite, pyrochlore (Pb-rich), cerianite-(Ce), and bastnäsite-(Ce) (Foord et al. 1984). Some of these pegmatites show evidence for intense metasomatic alteration and their mineralogy may have been overprinted by late fluids associated with the intrusion of the Mount Rosa granite (Gross and Heinrich 1966; Foord
et al. 1984). Most of these mapped pegmatites are in the St. Peter’s Dome area, a prominent ~2950-meter-high rocky peak ~5 km east of Mount Rosa.

Pegmatites of the ‘interior’ type, hosted fully within the Mount Rosa Granite, are generally smaller in size, more poorly-zoned, and less mineralogically-complex than the ‘exterior’ type, with considerable Zr, Th, Ti, Nb>Ta, and Ce>Y (Gross 1962; Gross and Heinrich 1966). Their mineralogy and chemistry is close to that of their host rocks, and their genetic connection to the Mount Rosa granite appears less ambiguous than the ‘exterior’ type bodies (Foord et al. 1984). A somewhat distinct sub-class of the ‘interior type’ pegmatites are the hydrothermal thorite veins (Gross 1962) which occur near Rosemont on the eastern slopes of Mount Rosa proper. These pegmatites, up to 50 m long and several meters wide, consist of ‘altered quartz, feldspar and Na-Fe amphibole which contain abundant zircon, thorite, thorogummite, uranothorite, as well as possible unidentified REE phases (Gross and Heinrich 1966). The surrounding granite has been bleached and shows evidence for strong metasomatic alteration such as conversion of feldspars to chlorite-biotite and clays as well as solution cavities in quartz with crystals of zircon and REE species (Gross 1962). These deposits may more strongly related to late faults and fractures than other Mount Rosa-type pegmatites, but their exact structural relationship is unknown (Gross and Heinrich 1966).
2.1 Abstract

A Petrologic and geochemical study of the Mount Rosa Complex (MRC), El Paso County Colorado (USA) was performed to investigate the magmatic and hydrothermal processes that led to the diverse lithologies and the concentration of rare earth elements (REE) and other high field strength elements (HFSE; e.g. Zr, Nb) in processes similar to those observed in other peralkaline granite complexes hosting economic rare element mineralization (e.g. Strange Lake Canada, Pitinga mine Brazil). The MRC consists of a diverse association of peraluminous to peralkaline granitic rocks located in the southeastern corner of the ~1.08 Ga anorogenic (A)-type Pikes Peak Batholith near the City of Colorado Springs. Field work characterized the spatial distribution and cross-cutting relationships of various rock units, and thin section petrography established crystallization sequence and major modal mineralogy of each unit. SEM-based methods (automated mineralogy, EDS analyses, and BSE imaging) characterized quantitative modal and accessory mineralogy as well as alteration. Electron probe micro-analysis (EPMA) was performed on micas from each unit, as well as whole rock geochemical analyses.

The MRC comprises, from oldest to youngest, peraluminous Pikes Peak biotite granite, fayalite-bearing quartz syenite, granitic to aplitic dikes, peralkaline Na-Fe amphibole granite (Mount Rosa granite), mafic dikes ranging from diabase to syenite, and a diverse suite of Niobium-Yttrium-Fluorine (NYF)-type pegmatites. These early peraluminous units have similar whole rock REE patterns and EPMA biotite chemistry which is similar to analyzed biotites from other Pikes Peak Batholith and ~1.1 Ga Texas sodic granitoids. They are believed to be derived from extended fractionation of mantle-derived alkali basalt during crustal thinning at ~1.1 Ga, coincident with initiation of Mid-Continent Rift activity and similar A-type magmatism in Texas. Fayalite-bearing quartz syenite contains clots of mafic minerals (e.g. fayalite, ilmenite) associated with fluorite, aegirine, and REE minerals which suggest early high-temperature crystallization at low $f(\text{O}_2)$ and $f(\text{H}_2\text{O})$ from a melt already enriched in F and REE. Granitic dikes
have a radial geometry around the complex and crystallized from residual peraluminous magma relatively early in the magmatic evolution of the complex, as evidenced by high Al and low F + Cl contents in biotite and low REE in whole rock geochemistry relative to other units. The remaining volatile (e.g. F, OH) enriched peraluminous melt potentially yielded an immiscible peralkaline melt fraction which crystallized irregular blobs, sills, and dikes of Na-Fe amphibole bearing Mount Rosa granite, whose strong mineralogical, textural, and contact heterogeneity reflects the incomplete separation of this peralkaline melt from the peraluminous magma. The Mount Rosa granite is enriched in REE and other incompatible elements and contains biotite with very low Al content and generally high Na + K concentrations which reflect the peralkaline melt chemistry. Emplacement of the Mount Rosa granite was accompanied by strong autometasomatism and melt heterogeneity, as evidenced by an albitization halo, aegirinization/hematization during later acidic alteration, and entrapment of fluorite, REE Fluorocarbonates, and albite in late-magmatic zircon grains. Late mafic dikes are analogous to lamprophyres seen in other rare metal granites and show a strong enrichment in REE, Zr, and other HFSE, with a biotite chemistry which plots in distinctly different domains than that from other MRC units. Pegmatites include poorly-zoned, irregular bodies (Type-(I) pegmatites) which are mineralogically-simple and likely represent residual Pikes Peak and Mount Rosa granite melt, and well-zoned, intrusive, mineralogically-complex Type-(II) pegmatites. A detailed transect study of a Type-(II) pegmatite reveled systematic changes in mineralogy and mica chemistry related to inward magmatic crystallization and the development of a F-rich fluid which altered and replaced magmatic minerals (e.g. arfvedsonite) with hydrothermal minerals, including REE Fluorocarbonates and other HFSE minerals.

2.1.2 Introduction

Peralkaline (Na$_2$O/K$_2$O > Al$_2$O) granite complexes are associated with A-type magmatism in regions of crustal extension and occur throughout geologic time (Bonin 2007, Costi et al. 2009). They typically include a diverse association of granitic as well as coeval volcanic rocks including gabbro, diabase, syenite, lamprophyre, subsolvus to hypersolvus granites, fayalite-bearing granitoids and syenite, and rhyolite (Hogan et al. 2000, Marks et al. 2003). The peralkaline magmas which form such complexes are typically related to larger volumes of
metaluminous to peraluminous magma in large composite batholiths in which their timing is late relative to main emplacement of the batholith (Smith et al. 1999, Sakoma et al. 2000). While generally occurring as roughly circular, epizonal plutons, they can also include ring-dike complexes, sheet-like sills, multiple overprinting intrusive phases, dike swarms, and pegmatites (Tulloch 1992, Trumbull et al. 2004). From an economic standpoint, peralkaline granite complexes often concentrate valuable rare earth elements (REE) and other high field strength elements (HFSE) to mineable levels, and from a scientific perspective they represent some of the most compelling evidence for a link between magmatic and hydrothermal and/or fluid-driven processes in mobilizing and concentrating rare elements (Linnen et al. 2012).

Geochemically, A-type granites are typically enriched in incompatible elements (e.g. HFSE and LILE) and show relative depletion in certain compatible elements (e.g. Ni, Cr, Co, Ba, Sr, Eu; Frost and Frost 1997). Mineralogically, A-type granitoids and their related peralkaline rocks are characterized by Fe-rich mafic phases such as biotite, amphibole and pyroxene, and dissolved OH- and F-rich fluids can lead to abundant late-stage fluorite and hydrated phases such as Fe-micas and amphiboles (Bonin 2007). Alkali feldspar textures often show a transition from hypersolvus to subsolvus, and complex textural heterogeneity resulting from mingling of different magmas and localized volatile enrichment are very common (Bonin 2007; Smith et al. 1997, 1999). Likewise, A-type granite intrusions often show an impressive diversity of rock types, including but not limited to syenites, fayalite-bearing granite, Na-Fe amphibole/pyroxene granites, albitites, mafic dikes ranging from diabase to lamprophyre, pegmatites, gabbros, anorthosites, and various eruptive equivalents where an intrusion breaches the surface and is associated with volcanism (Eby 1992, Bonin 2007).

While debate over the origin of peralkaline granites exists, there are several accepted petrogenetic theories. The first is that enriched mantle-derived basaltic magmas generated during lithospheric rifting can undergo prolonged fractionation and magmatic enrichment in Na, K, volatiles, and incompatible elements to yield a peralkaline melt (Mungall and Martin 1996; Smith et al. 1997). Workers have compared peralkaline granite complexes (e.g., Strange Lake Canada) to pantellerite (peralkaline rhyolite) at Terceira, Azores Islands, which contains similar magmatic enrichment in HFSE, as well as rhyolite laccoliths in West Texas, and concluded that fractional crystallization could yield such a peralkaline melt (Mungall and Martin 1996; Rubin et al. 1993). Another theory, based largely on fluid and melt inclusion work, is that during the
end-stages of crystallization of a volatile (e.g. F, OH) enriched peraluminous melt, a peralkaline melt fraction can separate and continue on a separate evolutionary path of increasing peralkalinity (Thomas et al. 2005; 2006). Finally, others have suggested that particular magma source is less important than tectonic setting (lithospheric rifting) and high temperatures which can melt dehydrated lower crust which is then ‘fertilized’ by assimilation of volatile-rich crustal material during its ascent to yield a peralkaline granite through fractional crystallization (Collins et al. 1982; Trumbull et al. 2004).

Of the 7 elements defined as ‘critical’ by the US Department of Energy and the European Commission, 5 are sufficiently enriched in some peralkaline granites and associated pegmatites to be of economic interest (Cerny and Ecrit 2005, Linnen et al. 2012). Peralkaline granite complexes which have been considered economic REE and rare element targets for mining include the Madeira granite/Pitinga mine, Amazon, Brazil (Costi et al. 2009), the Strange Lake deposit in Quebec, Canada (Gysi and Williams-Jones 2013), and the Katugin deposit, eastern Russia (Levashova et al. 2015). The particular enrichment of many peralkaline granites in the HREE (e.g., \((\text{La/Yb})_N = 2-10\) and \(Y/\text{Nd} \geq 1\)) as well as Sn, Ta, and Nb has made them attractive targets for exploration as well as research (Chakhmouradian and Zaitsev 2005). In all of these deposits, the importance of late-magmatic F-rich fluids has been emphasized in transporting and concentrating REE and other HFSE (Boily and Williams-Jones 1994, Neto et al. 2009).

This study will trace the magmatic and hydrothermal evolution of the Mount Rosa Complex and decipher the significance of various proposed mechanisms in REE and other HFSE mobilization and concentration.

2.2 Geological setting and previous work

The Mount Rosa Complex (MRC) is part of the ~1.08 Ga, A-type Pikes Peak batholith, and is located ~15 km West of the city of Colorado Springs in the central Front Range of Colorado (Figs. 1.2, 2.1). The diverse rocks comprising the MRC range from peraluminous to peralkaline and form an arc-like complex ~10 x 15 km in size associated with the Mount Rosa intrusive center, the youngest intrusive center in the Pikes Peak Batholith (Fig. 2.1; Smith et al. 1999). Peraluminous Pikes Peak biotite granite hosts the following units, from oldest to youngest: fayalite-bearing syenite, granitic dikes, Na-Fe amphibole granite (Mount Rosa
granite), mafic dikes ranging from diabase to diorite, and numerous REE and other HFSE rich Niobium-Yttrium-Fluorine (NYF)-type pegmatites (Fig. 2.1, Gross 1962; Keller et al. 2005).

The PPB is one of several recognized A-type ~1.1 Ga granites in western North America, including the Red Bluff Suite and Franklin Mountains granite in western Texas and the Llano uplift of central Texas (Smith et al. 1999). The PPB is believed to have been emplaced at ~1080-1040 Ma from mantle-derived magmas associated regional transtension-extension at ~1.1 Ga (Smith et al. 1999). The ~1.4 Ga Berthoud orogeny, itself associated with A-type granites such as the Longs-Peak St. Vrain batholith, preceded the emplacement of the PPB (Frost and Frost 1997; Allaz et al. 2015). The geometry of ~1.1-1.2 Ga mafic to felsic igneous activity of A-type affinity in Colorado, New Mexico and Texas, as well as the modern geometry of A-type magmatism occurring near the African Rift suggest that the onset of Mid-Continent rift-related crustal thinning in central North America at ~1.2 Ga may be linked to the emplacement of the PPB (Fig. 1.3; Smith et al. 1999). Assimilation of continental crust into this ascending mantle melt and subsequent assimilation, fractionation and crystallization (AFC) processes have been used to explain the sodic and potassic differentiation trends as well as the mixed mantle and crustal isotopic signatures seen in the PPB (DePaolo 1981, Smith et al. 1999). The most recent isotopic data (Lu/Hf data from zircon) on the PPB suggesting the mantle-derived magmas dominated at first with a progressively larger input of Proterozoic Colorado Front Range crust over time (Guitreau et al. 2016).

The PPB is a ‘type example’ of an A-type granite and shows the classic geochemical, tectonic and petrologic attributes of this family of granites such as high Fe/(Fe + Mg) ratios and K$_2$O, HFSE, and LILE enrichment, and isotope values consistent with a large mantle input (Smith et al. 1999, Bonin 2007). The PPB has ~3300 km$^2$ of surface exposure, due to Laramide (~70 Ma) uplift, exhumation and associated glaciation (Barker et al. 1975, Smith et al. 1999). Numerous small, late stage-plutons which collectively comprise <10% of the exposure area have been divided into two geochemical groups, a potassic series, and a sodic series (Barker et al. 1975, Wobus 1976, Smith et al. 1999). The PPB is a composite pluton and while the dominant lithologies are coarse grained biotite ± amphibole syenogranite and monzogranite, minor lithologies include gabbro, diabase, syenite, fayalite and Na-Fe amphibole granite (sodic series) and biotite granite and quartz monzonite (potassic series; Smith et al. 1999).
Recognized late-stage intrusions of the sodic series include the Sugarloaf, West Creek, Taryall and Spring Creek plutons, the Mount Rosa Complex, and part of the Lake George ring complex (Smith et al. 1999). Late-stage intrusions of the potassic series include the Redskin Stock, part of the Lake George Ring Complex, Devil’s Head, Pikes Peak, and Wigwam Creek (Smith et al. 1999). Gross (1962) obtained several K\(^{40}/Ar^{40}\) dates on rocks of the Mount Rosa granite in the MRC ranging from 1020-1045 Ma, making it distinctly younger (but temporally-related to) the Pikes Peak Batholith. Unruh and Premo of the U.S Geological Survey (personal communication, 2015) obtained comparable Pb/U zircon age dates on the Mount Rosa granite of ~1040 Ma.

Rocks related to the MRC have been studied for the past 135 years, starting with the description of cryolite and other unusual fluoride minerals from pegmatites in the St. Peter’s Dome area by Cross et al. (1882). Coinciding with the discovery and exploitation of the famous Ivigtut cryolite deposit (pegmatite associated with a peralkaline granite) in Greenland, this study paved the way for future research on the MRC. Finlay (1916) was the first to recognize the Mount Rosa granite as a different unit within the Pikes Peak batholith on the basis of its Na-Fe amphibole content. Numerous mineralogical studies from ~1920-1960 focused on the rare and unusual minerals found in many MRC pegmatites, including Na-Fe amphibole (Coleman 1951), genthelvite (Glass and Adams 1953; Scott 1957), and weberite and related Al-fluorides (Pauly 1954). The first comprehensive study on the MRC was the PhD thesis of Gross (1962) which resulted in two papers in the American Mineralogist describing the petrology of the complex (Gross and Heinrich 1965) and its pegmatites (Gross and Heinrich 1966) of the MRC.

Gross (1962) and Gross and Heinrich (1966) characterized the Mount Rosa complex as a series of many generally small sill and dike-like bodies of various igneous rocks occurring in an arc-like ~15 x 10 km area (Fig. 1.3). He described the different units, from oldest to youngest, as fayalite granite, porphyritic granite, granitic/aplitic dikes, Mount Rosa Na-Fe amphibole granite, mafic dikes, ‘interior’-type pegmatites, and complex-type pegmatites, and concluded that all units were comagmatic. Barker et al. (1975) proposed a more specific model for the fayalite granite, suggesting a mantle source created an alkali olivine magma, which reacted with K and Na-poor granulite rocks in the lower crust, in turn creating a syenite magma. This syenite magma then ascends into the intermediate and upper crust, evolving through AFC processes into a fayalite granite magma now enriched in Na, K, Fe, F et cetera (Barker et al. 1975). Finally,
Barker et al. (1975) propose that through additional fractional crystallization and differentiation, the Mount Rosa granite formed.

Douglass and Smith (1993) and Smith et al. (1999) concluded that the sodic series rocks, including the MRC, formed via fractionation of mantle-derived magmas, versus crustal anataxis which they viewed as the source of the potassic-series rocks. Pegmatites of the MRC were first described by Landes (1935), who studied several complex Al-fluoride mineral bearing pegmatites in the St. Peter’s Dome and suggested that Al-fluorides were associated with a later hydrothermal fluid, which replaced earlier magmatic minerals, as well as suggesting the importance of F and Zr in these late fluids. Gross (1962) and Gross and Heinrich (1966) classified pegmatites of the MRC into several types: ‘Exterior’ pegmatites are hosted in Pikes Peak granite, are often large and well-zoned, and contain unusual minerals such as Al-Fluorides (e.g.; cryolite, weberite, prosopite) and REE fluorocarbonate phases (Gross and Heinrich 1966). Geochemically, they contain significant Be, Rb, Y, F, Zr, and Ce, with Ta>Nb (Gross and Heinrich 1966). ‘Interior’ pegmatites, in contrast, are either fully-enclosed by or in contact with Mount Rosa granite and are generally small, poorly zoned, and mineralogically-simpler than exterior pegmatites, with Zr, Th, Ti, Nb>Ta, and Ce>Y (Gross 1962, Gross and Heinrich 1966). Foord et al. (1984) suggested that the mineralogy and chemistry of the ‘interior’ pegmatites is close to that of their host rocks, and that their genetic connection to the Mount Rosa granite appears less ambiguous than that for the ‘exterior’ type pegmatites. Finally, Cerny and Ercit (2005) classified the pegmatites of the MRC as belonging to the peralkaline Niobium-Yttrium-Fluorine (NYF)-type and Na-Fe amphibole subtype, similar to pegmatites on Hurricane Mountain, New Hampshire, or the Franklin Mountains of western Texas.

### 2.2.1 Field Relations

Field relationships are summarized in Figure 2.1; a synthesis of mapping done by Gross (1962), Gross and Heinrich (1965, 1966), Keller (2005), and the author of this thesis (2014-2016), as well as in an idealized cross-section in Figure 2.2. While most of the contacts and occurrences of various rock types discussed in this text and shown in Figure 2.1 are the result of mapping done by the author from 2014-2016, some areas reflect mapping done by Gross (1962) and Keller et al. (2005). While contact relationships are often complex as well as poorly-exposed...
in the MRC, mapped contact relationships have allowed the establishment of a relative age relationship between each major lithology. In this scheme, *hypersolvus Pikes Peak biotite granite* is the oldest and volumetrically most abundant unit exposed in the MRC. Outcrops range from highly-weathered, coarse-grained ‘grusy granite’ in which feldspar alteration to clays and mafic silicate alteration to Fe-oxides has obscured primary textures, to relatively fresh, typically fine-grained white to pink biotite granite which forms larger erosion-resistant outcrops (Fig. 2.3A; C, Fig. 2.4A; H). Sometimes these outcrops are quite large, forming sub-vertical cliffs up to 300 m across and ~100 m tall on the southern and eastern flanks of Stove Mountain and St. Peter’s Dome (Fig. 1.4). The next oldest unit is *subsolvus Pikes Peak biotite granite* (Fig. 2.3B), which is visually similar to Pikes Peak hypersolvus biotite granite and not easily distinguishable without the aid of thin section petrography. The occurrence of subsolvus Pikes Peak biotite granite is restricted to areas generally <500 m from the contact with either the main body of Mount Rosa granite or one of the smaller bodies in the Gold Camp Road and St. Peter’s Dome areas, though it was noted in upper Cheyenne Canyon (sample MR34) between Gold Camp Road and Wye Summit Road, ~1 km from the contact with the main body of Mount Rosa Granite.

The next oldest unit is the *fayalite-bearing quartz syenite*, formerly described as ‘fayalite granite.’ While volumetrically much less significant than either the Pikes Peak hypersolvus or the Mount Rosa granite, the fayalite-bearing quartz syenite does occur in rounded to oblong bodies up to ~1 x .8 km in size, as well as more frequent elongate bodies measuring from ~2 x 2 m to ~200 x 500 m in size (Fig. 2.3D). The largest body of fayalite-bearing quartz syenite mapped was noted in contact with the main body of Mount Rosa granite in the north-central part of the MRC. Close to the contact with the Mount Rosa Granite, the fayalite-bearing quartz syenite is bleached and altered (sample MR14). Most fayalite-bearing quartz syenite, however, occurs in the St. Peter’s Dome area as well as below Gold Camp Road to the northeast of St. Peter’s Dome. Fayalite-bearing quartz syenite forms erosion-resistant outcrops of Fe-stained dark greenish-brown color which are characteristic of this unit (Fig. 2.3C). Contacts with older hypersolvus and subsolvus Pikes Peak biotite granite are generally sharp (Fig. 2.3D).
Figure 2.1. Geologic Map of the Mount Rosa Complex (mapping by Gross (1962), Keller et al. (2005), and Persson 2014-2016).

Granitic dikes range from monzogranite to syenogranite in composition and locally exhibit aplitic textures as well as pegmatitic enclaves, occurring throughout the MRC. However, larger and more continuous dikes occur in the St. Peter’s Dome and Gold Camp Road areas (Fig. 2.3E, F). The two most prominent of these dikes each measure ~1.5-2.5 m in thickness and have similar orientations; ~280/90 and ~310/90. The smaller of these two dikes (sample MR30) outcrops at the St. Peter’s Dome overlook at the juncture of Gold Camp and Old Stage Roads, and cuts fayalite-bearing quartz syenite and hypersolvus Pikes Peak biotite granite. The larger dike (sample MR37) is traceable in outcrop over ~1.5 km and cuts fayalite-bearing quartz syenite and hypersolvus Pikes Peak biotite granite, though it is cut by a cluster of Type-(I) pegmatites close to its southern terminus below Gold Camp Road (Fig. 2.3F). Large outcrops of granitic dikes are largely confined to aplitic or fine-grained dikes which are generally harder and will preferentially weather out of surrounding rocks (Fig. 2.3E). While most granitic dikes exhibit a N/NW strike and vertical to subvertical dip, several on the southwest side of the main body of
Mount Rosa granite show a S/SWs strike, exhibiting a crudely radial geometry around the core of the MRC (Fig. 2.1).

Figure 2.2 Field Images of Mount Rosa granite: A) oikocrystic Na-Fe amphibole granite, B) coarse-grained enclave in fine-grained Na-Fe amphibole granite, C) fine-grained Mount Rosa granite with randomly-oriented Na-Fe amphibole and astrophyllite crystals, D) irregular contact between Pikes Peak biotite granite (left) and Mount Rosa granite.
Mount Rosa granite occurs in two main areas of the MRC: as a large oblong body measuring ~6 x 2.5 km centered around Mount Rosa, and as numerous smaller oblong bodies as well as intrusive dikes which are focused in the St. Peter’s Dome and Stove Mountain area, generally below the elevation of Gold Camp Road. Field relations of the Mount Rosa granite are the most complex and heterogeneous of any studied lithology. In general, the Mount Rosa granite does not occur as continuous bodies, but as a series of sheet-like tabular dikes and sill-like bodies. The Mount Rosa granite occurs as several different facies (Fig. 2.2), which exhibit both gradational and abrupt cross-cutting contact relationships to each other. These include coarse-grained oikocrystic Na-Fe amphibole granite (Figs. 2.2, 2.4A), fine-grained oikocrystic Na-Fe amphibole granite, fine-grained prismatic randomly-oriented Na-Fe amphibole granite (Fig. 2.2C), fine-grained granite with relatively oriented Na-Fe amphibole crystals, porphyritic K-feldspar Na-Fe amphibole granite, and both fine and coarse-grained enclaves and inliers of Na-Fe amphibole granite up to several meters in diameter (Figs. 2.2B, 2.4B). Additionally, large pegmatitic of Mount Rosa granite occur, particularly on the margins of the main body of Mount Rosa granite (Fig. 2.6A). While contacts, where exposed, between the Mount Rosa and Pikes Peak granite are generally fairly sharp but irregular (Fig. 2.2D), however diffuse contacts marked by an increase in albite, Na-Fe amphibole, and a general decrease in grain size from Pikes Peak to Mount Rosa granite have also been noted. Finally, xenoliths of Mount Rosa granite in Pikes Peak biotite granite have been observed close to the border of the main body of the Mount Rosa granite (Fig. 2.4C). Dikes of Mount Rosa granite ranging from ~30 cm to ~5 m in width are vertical to nearly horizontal in dip and show a general N/NW strike (Figs. 2.4D). They are typically fine-grained and often have a ‘sugary’ texture due to a high amount of albite versus quartz, and consequently are often weathered/ altered with poor outcrop exposure.

Mafic dikes occur throughout the MRC, though most are concentrated in two clusters; one on the northeastern edge of the main body of the Mount Rosa granite and the other above and below Gold Camp Road in the St. Peter’s Dome and Stove Mountain areas. However, mapped mafic dikes occur as far as ~10 km away from the main Mount Rosa granite body near Helen Hunt Falls. Mafic dikes fall into two outcrop populations: silicic, fine-grained gray to black dikes ranging from ~1-4 m in width, which generally strike N/NW, dip vertically to steeply N/NE, and outcrop prominently (Figs. 2.4E; F). The second population is less erosion-resistant, sometimes rich in porphyritic feldspar and quartz-poor with similar dimensions as the silicic
dikes. These dikes outcrop poorly and strike W/SW, with dips ranging from vertical to sub-horizontal (Figs. 2.4D, 2.5E).

Figure 2.3 Contact Relationships and Field photographs of various rocks in the Mount Rosa Complex: A) typical hypersolvus Pikes Peak biotite granite close to the summit of Mount Rosa, B) visually similar subsolvus Pikes Peak biotite granite close to the Mount Rosa granite contact, C) typical fayalite-bearing quartz syenite close to Helen Hunt Falls, D) rounded masses of fayalite-bearing quartz syenite from ~1-15 m in diameter in hypersolvus Pikes Peak granite near Helen Hunt Falls, E) ~2 m wide granitic dike oriented ~340/90 cutting fayalite-bearing quartz syenite along Gold Camp Road near St. Peter’s Dome, F) Several Type-(I) pegmatites cutting extension of same granitic dike shown in (E) downhill from Gold Camp road.
Mafic dikes cut Pikes Peak biotite granite (both hypersolvus and subsolvus), granitic dikes, fayalite-bearing quartz syenite, and Mount Rosa granite. Mafic dikes sometimes occur in contact with and parallel to Type-(I) as well as Type-(II) pegmatites, and some mafic dikes (Fig. 2.6E) show inter-fingering, irregular contacts with Type-(I) pegmatites. Dikes range from ~0.3 x 10 m to ~5 x 50 m in size, and are typically oriented N/NW with vertical to sub-vertical dips,

Figure 2.4 Contact Relationships and Field photographs of various rocks in the Mount Rosa Complex: A) typical oikocrystic Na-Fe amphibole Mount Rosa granite, B) heterogeneous distribution of Na-Fe amphibole in Mount Rosa granite. C) xenolith of hypersolvus biotite granite in fine-grained Mount Rosa granite, D) mafic dike (outlined in blue) parallel to Mount Rosa granite dike (outlined in red) both cutting hypersolvus biotite granite, along Gold Camp Road on Stove Mountain, E) ~1 m thick mafic dike (outlined in red) oriented ~270/80E cutting hypersolvus biotite granite and type-(I) pegmatite (outlined in green), exposed along Gold Camp Road on Stove Mountain, F) erosion-resistant ~1 m thick mafic dike oriented ~330/90 outcropping on lower St. Peter’s Dome.
though some are oriented ~E-W with shallower dips, typically to the south. Some mafic dikes, such as the ~1.5 m wide, ~330/90, fine-grained hornblende diabase dike on St. Peter’s Dome below Gold Camp Road (Fig. 2.4F) can be followed in outcrop for several hundred meters, though their outcrop is discontinuous. Chilled margins and clean intrusive contacts with both hypersolvus and subsolvus Pikes Peak biotite granite indicate mafic dikes are younger.

Figure 2.5 Contact Relationships and Field photographs of various rocks in the Mount Rosa Complex: A) Highly-altered Mount Rosa granite adjacent to ‘Type-(I) pegmatite, with all Na-Fe amphibole altered to Fe-oxides and clays, B) altered Mount Rosa granite dike from area of Fig 2.4D, C) irregular inclusions of Mount Rosa granite in mafic dike near summit of Mount Rosa, D) hydrothermal zircon-bearing Type-(II) pegmatite subcrop in Gold Camp Road on St. Peter’s Dome (sample MR38), E) irregular thin mafic dike stringers intruding hypersolvus biotite granite along Gold Camp Road on Stove Mountain, F) inclusions of hypersolvus biotite granite in mafic dike on lower St. Peter’s Dome.
Mafic dikes also cut fayalite-bearing quartz syenite, and sometimes occur at the contact between fayalite-bearing quartz syenite and Pikes Peak biotite granite, where they appear to have exploited pre-existing weaknesses between these two units. Additionally, mafic dikes often have intruded parallel to pegmatites, particularly in the St. Peter’s Dome and Stove Mountain area, and pegmatites sometimes contain small xenoliths of mafic dikes, and also exhibit anastomosing, irregular contacts with these pegmatites (Figs. 2..5C; 2.6E).

Figure 2.6 Field photos of Type-(I) pegmatites- A) irregular Th-rich pegmatite in main body of Mount Rosa granite, B) narrow pegmatite with quartz core and microcline-albite wall zone cutting fayalite-bearing quartz syenite, C) simple, irregular pegmatite in hypersolvus Pikes Peak granite near Helen Hunt Falls, D) poorly-zoned Na-Fe amphibole pegmatite cutting granitic dike on St. Peter’s Dome, E) irregular contact between Type-(I) pegmatite (left) and mafic dike (right), F) poorly-zoned Na-Fe amphibole pegmatite cutting subsolvus Pikes Peak granite near Mount Rosa contact.

Type-(I) pegmatites occur throughout the MRC and can be subdivided into two populations. The first pegmatite population is hosted in Mount Rosa granite, exhibiting sharp but irregular, anastomosing contacts with their host rocks (Fig. 2.6A; F). These pegmatites are poorly zoned and range from ~1.5 x 1 m to ~8 x 5 m in size. A group of Th-rich Type-(I) pegmatites (Fig. 2.6A) have been exposed at former Th exploration (1950’s) trenches on the southeastern slopes of Mount Rosa (Fig. 1.4), and other similar pegmatites occur in Mount Rosa
granite nearby. Type-(I) pegmatites also occur along or just below Wye Summit Road near the eastern contact of the main body of Mount Rosa granite. The second population of Type-(I) pegmatites occurs in the St. Peter’s Dome area as dikes (Figs. 2.6B, C, D) as well as miarolitic pockets ranging from <10 cm to ~30 m x 2 m. Their mineralogy is simple (albite, K-feldspar, quartz, biotite), poorly zoned, and typically cut all other units except Type-(II) pegmatites.

![Figure 2.7 Field photographs of Type-(II) pegmatites, A) ~1.5 m thick pegmatite cutting fayalite-bearing quartz syenite; note fine-grained border zone with abrupt transition to coarse-grained quartz-albite-microcline with inward-projecting Na-Fe amphibole crystals to ~30 cm, B) ~80 cm pegmatite cutting hypersolvus biotite granite, C) large ~6 m thick Type-(II) pegmatite cutting fayalite syenite on lower St. Peter’s Dome, D) Na-Fe amphibole crystals brecciated by late albite, Al-fluoride minerals, and quartz, E) large area of relatively mono-mineralic brown zircon with large microcline crystals in quartz, F) Large pod of altered Aluminofluoride minerals in quartz core of Eureka Tunnel, with hydrothermal zircon-bearing zone highlighted in blue.](image)

Type-(II) pegmatites (Fig. 2.7) also occur throughout the MRC but are concentrated in the St. Peter’s Dome and Stove Mountain areas (Fig. 1.4), at or below the elevation of Gold Camp.
Road. There is also a cluster of type-(II) pegmatite in upper Cheyenne Canyon near South Cheyenne Creek between Gold Camp Road and Wye Summit Road (Fig. 2.1). Finally, several Type-(II) pegmatites occur near the southeastern edge of the main body of Mount Rosa granite (Fig. 2.1), though these are anomalous and were not examined in detail for this study. They range from ~1 x 8 m to ~8 x 45 m in size and are typically tabular dikes oriented approximately E-W to NW-SE with a shallow (~20-40°) dip to the south. Type-(II) pegmatites cut all other rock types and are typically well-zoned, with strongly defined unidirectional solidification textures (UST’s) such as large Na-Fe amphibole and K-feldspar crystals up to 50 cm and 1.5 m, respectively, which have grown perpendicular to pegmatite walls (Figs. 2.7A, B). Type-(II) pegmatites contain exotic minerals in their core zones, such as aluminofluoride species (Fig. 2.7F), as well as large monomineralic areas of HFSE minerals (e.g. zircon; Fig. 2.7E). Brecciated textures are common, in which magmatic minerals such as Na-Fe amphibole have been disrupted (Fig. 2.7D). Type-(II) pegmatites cut Pikes Peak granite, fayalite syenite (Fig. 2.7C), granitic dikes, mafic dikes, and Mount Rosa granite, and in rare instances cut older Type-(I) pegmatites as well.

2.3 Analytical Methods

The following section will briefly describe various analytical methods used in this study. These methods include automated mineralogy, field-emission SEM, electron probe micro analysis (EPMA) and whole rock geochemical analyses.

2.3.1 Automated Mineralogy

From the total of 47 thin sections and 17 epoxy mounts studied by optical light microscopy (Table 2.1), a representative set of 23 samples, as well as a group of 17 epoxy mounts were selected for SEM-based automated mineralogy analysis at the Colorado School of Mines in Golden, Colorado. The purpose of these analyses was to quantify the modal mineral abundance and textural information, as well as to obtain semi-quantitative mineral chemistry data. Carbon-coated thin sections and 30 mm epoxy mounts were loaded into the VEGA3 tungsten filament SEM from TESCAN. Scan parameters were input using the TESCAN TIMA (Tescan Integrated Mineral Analyzer) software, and pixel size (distance between rasterized
analysis points) varied from 5-20 µm, depending on sample grain size and textural heterogeneity. Out of 23 thin sections analyzed, 13 were scanned entirely at 20 µm resolution, and selected areas of the remaining 10 thin sections were analyzed at a resolution of < 20 µm (Table 2.1.). Epoxy mounts were analyzed at 15 µm resolution. Four energy dispersive X-ray (EDX) spectrometers in the TIMA instrument collected 1,000 total X-ray counts for each analysis point in the established raster grid for each sample measurement. The acceleration voltage was 25 keV and a beam size of ~ 500 nm. The collected EDX and BSE data were compared and matched to a library of mineral definitions and mineral assignments were done by the TIMA software.

2.3.2 Field-Emission SEM

Out of the overall sample set of 47 thin sections and 23 epoxy pucks, 17 thin sections and two epoxy mounts were examined using the TESCAN MIRA3 field-emission SEM instrument. The FE-SEM is equipped with one Bruker XFlash© 6/30 silicon drift Energy Dispersive Spectrometer (EDS) detector, which is operated via the Esprit software. BSE images were acquired for each sample, with overview images taken of pre-selected areas at the approximate scale of reference photomicrographs taken during transmitted light petrography. BSE images were acquired at a working distance of 10 mm and an accelerating voltage of 20 keV. For specific samples containing REE or other HFSE minerals, an accelerating voltage of 25 keV was used to allow better peak differentiation of these elements.

2.3.3 Electron Probe Micro-Analysis (EPMA)

A select group of 17 representative thin sections from all studied rock units were selected for electron probe micro-analysis (EPMA) of biotite at the US Geological Survey in Denver microbeam laboratory at the Denver Federal Center in Lakewood, Colorado. A JEOL 8900 EPMA instrument was used for these analyses. Prior to sample analysis, 9 natural and synthetic standards were used to calibrate the instrument. The standards used were: Tiburon albite (Na), Spring Water olivine (Mg, Si), synthetic CaF₂ (F), synthetic MgCrO₄ (Cr), Miyake Anorthite (Al, Ca), synthetic FeSiO₄ (Fe), synethetic TiO₂ (Ti), Or-1A Orthoclase (USGS-Menlo #5-168; K), and sodalite (Cl). For biotite in the MRC samples, an accelerating voltage of 15 keV, a beam
current of 20 nA, and a beam size of 10 µm was used. Count time on the peak position was 20 seconds, with background counting times of 10 seconds on each side of the peak. Sodium was analyzed first to avoid possible Na migration away from the beam. ZAF matrix corrections were applied to the raw data using the JEOL 8900 superprobe software. Biotite formulae were calculated on the basis of 22 oxygen atoms per formula unit (afpu).

### 2.3.4 Whole Rock Geochemistry

A select group of 5 samples (Table 2.3) representing the major rock types of the MRC were sent to Dr. Sarah Hanson at Adrian College in Adrian, Michigan, where they were processed for submission to ActLabs in Ancaster, Ontario Canada for whole rock geochemical analysis. The samples were crushed to ±200 mesh using rotary cone crushers, and representative ~500 gram samples were sent to ActLabs for analysis via Lithium tetraborate fusion ICP and ICP-MS (inductively coupled plasma mass spectroscopy). Detection limits range from 0.001 - 0.01 mass percent for major elements (Table 2.3) and from 0.10 - 30 ppm for minor and trace elements. Duplicates and blanks were analyzed for each analysis for QA/QC purposes by ActLabs Corporation.

### 2.4 Results

The following section covers the results of petrography, mineralogy and geochemistry of all major lithological units in the MRC. The results of whole rock geochemistry, SEM-based automated mineralogy, imaging and mineral characterization, and EPMA analysis of biotite are also summarized.

#### 2.4.1 Petrography

A total of 47 polished thin sections and 17 epoxy mounts were investigated in this study. All samples were collected in the field from 2014 to 2016, and represent all major rock units in the MRC, as well as significant textural, mineralogical, and petrologic variations within individual rock units. A summary of the samples used in this study can be found in Table 2.1.
Detailed transmitted light petrography and automated mineralogy allowed the establishment of a crystallization sequence and the identification and textural characterization of select minerals and grains for subsequent SEM and EPMA analyses. Fine-grained or rare accessory minerals were mainly characterized by SEM-based automated mineralogy (TIMA) and imaging (FE-SEM).

_Hypersolvus Pikes Peak Biotite Granite_

Two facies of Pikes Peak biotite granite were identified in the MRC: _hypersolvus_ and _subsolvus_ granite. The oldest unit in the MRC is the hypersolvus Pikes Peak granite, represented by sample MR09, MRT-08, and MR2016-18. Hypersolvus Pikes Peak biotite granite is peraluminous and typically medium to coarse-grained, allotriomorphic biotite syenogranite. Perthitic alkali feldspar is slightly more abundant than quartz, comprising ~50% of the rock, with perthite grains typically 5-20 mm in size (Figs. 2.8A, B). Quartz comprises ~21% of the rock, and later albite (albitization) is visually estimated to comprise half of the measured albite total (albite measured in automated mineralogy includes both albite lamellae in perthite and albitization), or ~20% of the total area. Biotite is the major mafic mineral and comprises about 2-5 modal percent of the rock, and is interstitial to larger perthitic alkali feldspar grains and quartz. Biotite is associated with lesser amounts (<1%) of fluorite, zircon, Fe and Ti-oxides, apatite, and REE minerals; both monazite/xenotime and REE fluorocarbonates (Figs. 2.9A, B). On the basis of textural relationships, zircon and Fe-Ti oxides are interpreted to have crystallized first (Fig. 2.11), followed by large crystals of perthitic alkali feldspar. This was followed by quartz, which partly overlapped with feldspar crystallization (Fig. 2.11) as evidenced by rounded quartz inclusions to ~1 mm in the rims of feldspar grains (Fig. 2.8A). Biotite crystallization followed closely, with biotite grains incorporating inclusions of quartz and feldspar and typically nucleating on and surrounding older zircon and opaque Fe-Ti oxide grains (Fig. 2.8B). Fluorite is typically associated with biotite and forms subhedral grains generally <500 µm (Fig. 2.9A, B). REE minerals occur in two populations: early REE minerals which typically occur as inclusions in zircon or biotite are characterized by small (<50 µm) grains of monazite/xenotime.
Table 2.1: Select sample set used in this study from the Mount Rosa Complex.

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>UTM Northing</th>
<th>UTM Easting</th>
<th>Rock Type</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>PP-MR01</td>
<td>38.7661</td>
<td>104.9436</td>
<td>Subvolus Pikes Peak biotite granite</td>
<td>Coarse-grained white subvolus Pikes Peak biotite granite <em>1 m from Mount Rosa granite contact.</em></td>
</tr>
<tr>
<td>PP-MR09</td>
<td>38.7717</td>
<td>104.9326</td>
<td>Hypersolvus Pikes Peak biotite granite</td>
<td>Coarse-grained pink hypersolvus Pikes Peak biotite granite <em>1 km from MRG contact.</em></td>
</tr>
<tr>
<td>PP-MR34</td>
<td>38.743</td>
<td>104.9268</td>
<td>Subvolus Pikes Peak biotite granite</td>
<td>White subvolus Pikes Peak biotite granite <em>500 m from MRG contact.</em></td>
</tr>
<tr>
<td>PP-MR16</td>
<td>38.75972</td>
<td>104.9508</td>
<td>Fyallite-bearing quartz syenite</td>
<td>Birchedge and altered fayalite-bearing quartz syenite close to MRG contact; No Amph replacing Biotite</td>
</tr>
<tr>
<td>PP-MR31</td>
<td>38.7425</td>
<td>104.9128</td>
<td>Fyallite-bearing quartz syenite</td>
<td>Fresh fayalite-bearing quartz syenite; St. Peter's Dome overlook.</td>
</tr>
<tr>
<td>PP-MR30</td>
<td>38.7425</td>
<td>104.9128</td>
<td>Granitic Dike</td>
<td>*150 m long, E-W subvertical 2-meter wide aplite to medium-grained granitic dikes cutting MR31.</td>
</tr>
<tr>
<td>PP-MR65</td>
<td>38.7606</td>
<td>104.9442</td>
<td>Mount Rosa Granite</td>
<td>Olivicrestic Na-mphibole Mount Rosa Granite; large oolites in main MRG body.</td>
</tr>
<tr>
<td>PP-MR04</td>
<td>38.74444</td>
<td>104.9458</td>
<td>Mount Rosa Granite</td>
<td>Dilike bodies of fine-grained MRG with perphyritic sills of forsterite crystals to 5 mm and perphyritic Na-amphibole prisms oriented *050/48.</td>
</tr>
<tr>
<td>PP-MR066</td>
<td>38.7667</td>
<td>104.9364</td>
<td>Mount Rosa Granite</td>
<td>Contact between coarse-grained subvolus MRG enclave and surrounding fine-grained MRG.</td>
</tr>
<tr>
<td>PP-MR16</td>
<td>38.76639</td>
<td>104.9394</td>
<td>Mount Rosa Granite</td>
<td>Fine-grained white albitite-rich Mount Rosa Granite with Na-amphibole &amp; mafic plagioclase; biotite.</td>
</tr>
<tr>
<td>PP-MR26</td>
<td>38.75833</td>
<td>104.9056</td>
<td>Mount Rosa Granite</td>
<td>*2 m dikes of fine-grained a-bite-rich Mount Rosa granite oriented *280/40.</td>
</tr>
<tr>
<td>PP-MR26</td>
<td>38.75833</td>
<td>104.9056</td>
<td>Mafic Dike</td>
<td>*1 m fine-grained mafic dike parallel to MR26; MRG dikes oriented *280/40; both cut PPG.</td>
</tr>
<tr>
<td>PP-MR11</td>
<td>38.7622</td>
<td>104.9369</td>
<td>Mafic Dike</td>
<td>*1.5 m wide *285/90 Perphyritic mafic dike cutting MR11 near Wye Summit Road in main MRG body.</td>
</tr>
<tr>
<td>MRS-11</td>
<td>38.7522</td>
<td>104.9064</td>
<td>Type-(I) pegmatite</td>
<td>Sample of mafic pegmatite in Gold Camp Road above Eureka Tunnel; cross-cutting vein assemblage of hydrothermal origin.</td>
</tr>
<tr>
<td>MRS-12</td>
<td>38.7522</td>
<td>104.9061</td>
<td>Eureka Tunnel complex Al-Fluoride peg in PPG</td>
<td>Sample of mafic pegmatite with distinctive <em>azitic</em> clinopyroxene crystals from *40 x 50 cm; pod associated with Al-fluorides in quartz core of Eureka Tunnel pegmatite.</td>
</tr>
<tr>
<td>PP-MR2016-17</td>
<td>38.75212</td>
<td>104.9128</td>
<td>Mafic Dike/Hypersolvus Pikes Peak Granite</td>
<td>Contact between *1 m fine-grained mafic dike oriented *230/27 E-W cutting pink hypersolvus biotite PPG in Gold Camp Road cut.</td>
</tr>
<tr>
<td>MRT-01</td>
<td>38.752</td>
<td>104.9113</td>
<td>Fyallite-bearing quartz syenite; upper contact of pegmatite; 0-10 cm</td>
<td>Typical green-bluish-grey coarse-grained fayalite-bearing quartz syenite; upper pegmatite contact.</td>
</tr>
<tr>
<td>MRT-02</td>
<td>38.752</td>
<td>104.9119</td>
<td>upper contact of pegmatite; 10-20 cm</td>
<td>Finer-grained fayalite-bearing quartz syenite; upper pegmatite contact, more altered, just above pegmatite contact.</td>
</tr>
<tr>
<td>MRT-03A</td>
<td>38.752</td>
<td>104.9119</td>
<td>upper contact of pegmatite; 20-30 cm</td>
<td>Pegmatite wallzone; contact with fayalite granite, altered wallrock of pegmatite (hanging wall side).</td>
</tr>
<tr>
<td>MRT-03B</td>
<td>38.752</td>
<td>104.9119</td>
<td>upper wall zone of pegmatite; 40-60 cm</td>
<td>Pegmatite wallzone (hanging wall side).</td>
</tr>
<tr>
<td>MRT-03C</td>
<td>38.752</td>
<td>104.9119</td>
<td>upper wall zone of pegmatite; 40-60 cm</td>
<td>Pegmatite wallzone (hanging wall side).</td>
</tr>
<tr>
<td>MRT-06</td>
<td>38.752</td>
<td>104.9119</td>
<td>upper core zone of pegmatite; 40-60 cm</td>
<td>Pegmatite upper core zone.</td>
</tr>
<tr>
<td>MRT-07</td>
<td>38.752</td>
<td>104.9119</td>
<td>central core zone of pegmatite; 40-60 cm</td>
<td>Pegmatite central core zone</td>
</tr>
<tr>
<td>MRT-07</td>
<td>38.752</td>
<td>104.9119</td>
<td>central core zone of pegmatite; 40-60 cm</td>
<td>Pegmatite central core zone</td>
</tr>
<tr>
<td>MRT-07</td>
<td>38.752</td>
<td>104.9119</td>
<td>central core zone of pegmatite; 40-60 cm</td>
<td>Pegmatite central core zone</td>
</tr>
<tr>
<td>MRT-08</td>
<td>38.752</td>
<td>104.9119</td>
<td>lower zone of pegmatite; 90-200 cm</td>
<td>Pegmatite lower wall zone.</td>
</tr>
<tr>
<td>MRT-09</td>
<td>38.752</td>
<td>104.9119</td>
<td>lower wall zone of pegmatite; 90-200 cm</td>
<td>Pegmatite lower wall zone.</td>
</tr>
</tbody>
</table>
Later REE minerals in hypersolvus Pikes Peak granite are characterized by small (<20 µm) grains of REE fluorocarbonates (e.g.; bastnasite, synchysite, parisite) and typically rim biotite and fluorite. Albititization is ubiquitous in all samples and typically invades grain boundaries between alkali feldspar grains, where it blurs perthitic textures (Fig. 2.9A, B). Minor late clay alteration minerals (kaolinite, sericite) occur mainly as an alteration of biotite and along feldspar grain boundaries.

Zircon in hypersolvus Pikes Peak biotite granite occurs as subhedral to euhedral crystals from ~50-500 µm, comprising ~0.13% of the rock. They exhibit fairly well-defined oscillatory zoning in their rims, and have cores with patchy zonation. Inclusions are sparse, throughout the grains, and are typically phases like monazite/xenotime and Fe-Ti oxides.

Figure 2.8 Photomicrographs of hypersolvus Pikes Peak biotite granite: A.) large perthitic alkali feldspar grains with interstitial quartz, biotite and small zircon grains, B.) Fe-Ti oxides and zircon in biotite with large perthitic alkali feldspar grains, C.) fluorite grains in biotite with quartz and perthitic alkali feldspar, D.) hypersolvus biotite granite at Type-(II) pegmatite contact with secondary albite and Na-Fe amphibole with fluorite and biotite.
Subsolvus Pikes Peak Biotite Granite

Subsolvus Pikes Peak biotite granite is represented by samples MR01 and MR34, which are ~1 m and ~800 m from the contact with the main body of Mount Rosa granite, respectively (Table 2.1, Fig. 2.1). Visually, it is a white to pink, generally medium-grained biotite granite, though MR34 exhibits a somewhat porphyritic texture in which large alkali feldspar grains are set in a fine-grained quartz-albite matrix.
Figure 2.10 Summary of mineral assemblages and modal abundance in the Mount Rosa Complex
<table>
<thead>
<tr>
<th>MINERAL</th>
<th>Hypersolvus Pikes Peak Biotite Granite</th>
<th>Subsolvus Pikes Peak Biotite Granite</th>
<th>Fayalite-bearing Quartz Syenite</th>
<th>Granitic Dikes</th>
<th>Mount Rosa Granite</th>
<th>Mafic Dikes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ti/Fe-oxides</td>
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<tr>
<td>Zircon</td>
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<tr>
<td>Monazite</td>
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<tr>
<td>Thorite</td>
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<tr>
<td>Fayalite</td>
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<tr>
<td>Apatite</td>
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<tr>
<td>Ba-Ti silicate</td>
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<tr>
<td>Plagioclase</td>
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<tr>
<td>Alkali Feldspar</td>
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<tr>
<td>K-Feldspar</td>
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<td>Quartz</td>
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<tr>
<td>Biotite</td>
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<td>Na-Amphibole</td>
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<tr>
<td>Na-Pyroxene</td>
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<tr>
<td>Aegirine-Augite</td>
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<tr>
<td>Astrophylite</td>
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<tr>
<td>Fluorite</td>
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<tr>
<td>Y-Fluorite</td>
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<tr>
<td>REE Fluorocarbonates</td>
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<tr>
<td>Clays</td>
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</tbody>
</table>

Figure 2.11 Crystallization Sequence of Major Units in the MRC. Note pegmatite crystallization sequence is shown in figure 2.24.
Subsolvus Pikes Peak biotite granite is characterized by roughly equal modal abundances of albite (27-44%) K-feldspar (20-25%) and quartz (25-35%; Figs. 2.9C, D, 2.12). Magmatic albite crystallized early, followed by perthitic alkali feldspar (Fig. 2.11). Only quartz inclusions can be found in alkali feldspar grains (Fig. 2.12). Fe-Ti oxides, apatite, zircon, and monazite/xenotime crystallized early and typically form inclusions in biotite, which crystallized after quartz and feldspars. Fluorite is noticeably more abundant than in hypersolvus Pikes Peak granite, and is more abundant in sample close to the Mount Rosa granite contact (Fig. 2.11C).

Figure 2.12 Photomicrographs of subsolidus Pikes Peak biotite granite, showing: A) albite grains with perthitic alkali feldspar and slightly later biotite from sample MR01 close to the Mount Rosa granite contact, B) late biotite with quartz blebs in albite and alkali feldspar, C) sericitized alkali feldspar grains with large albite crystals and biotite with fluorite and zircon inclusions from subsolvus biotite granite ~1 km from Mount Rosa granite contact, D) altered subsolvus biotite granite.
REE minerals in subsolvus Pikes Peak granite are typically associated with fluorite, and fall into two populations. Early, less abundant monazite/xenotime forms inclusions in zircon and biotite and is subordinate to REE fluorocarbonate minerals, which rim and locally replace biotite. Late albitization is significant, and occurs as lath-like, intergrown albite crystals up to 500 µm, invading grains boundaries and locally replace perthitic alkali feldspar (Fig. 2.9C, D). Minor late clay alteration minerals (kaolinite, sericite) occur mainly as an alteration product of biotite and along feldspar grain boundaries. Zircon in subsolvus Pikes Peak biotite granite is similar to that of hypersolvus granite, with no noticeable differences.

Fayalite-bearing Quartz Syenite

Fayalite-bearing quartz syenite is represented by samples MR31, MR14, MRT01, and MRT02 (Table 2.1). Fayalite-bearing quartz syenite is dominated by large perthitic alkali feldspar grains comprising about 35 modal percent of the rock, with lesser primary albite (~20 modal percent), microcline (~15 modal percent), and quartz (~12 modal percent; Fig. 2.14). Fayalite (~0.5%) crystallized early, forming rounded blebbly grains around 50-300 µm in size (Fig. 2.13C, D). Fayalite is typically associated with Fe-Ti oxides (ilmenite > rutile or hematite), fluorite, aegirine, and biotite. Early Fe-Ti oxides crystallized with fayalite and form inclusions in biotite, while later Fe-Ti oxides replace altered fayalite and biotite (Fig. 2.14). Fayalite is typically altered to serpentine-group minerals and Fe oxides. Zircon crystallized early and occurs as euhedral to subhedral crystals and grains from 50-300 µm in size. Albite in fayalite-bearing quartz syenite crystallized next, followed by alkali feldspar, microcline, and quartz (Fig. 2.11). Biotite forms ragged interstitial grains between quartz and feldspar grains. Aegirine forms small radiating groups of thin acicular green needles which have nucleated on biotite and crystallized late (Fig. 2.13A, B).

Fluorite is associated with biotite and forms interstitial blebs of up to 500 µm. Monazite/Xenotime and apatite occur as early inclusions in zircon, as well as small (<50 µm) inclusions in fayalite in sample MRT01. REE fluorocarbonate minerals are associated with fluorite and biotite, and also occur as inclusions in altered fayalite. Minor allanite is associated with fluorite, altered fayalite and zircon (Fig. 2.13D). Apatite grains exhibit xenotime rims in sample MRT01.
Minor allanite is associated with fluorite, altered fayalite and zircon (Fig. 2.13D). Apatite grains exhibit xenotime rims in sample MRT01. Close to the contact of the Mount Rosa granite, fayalite-bearing quartz syenite contains biotite which has been largely replaced by Na-Fe amphibole. Albite alteration is ubiquitous in all samples (Fig. 2.14) and is most intense in samples adjacent to pegmatites (MRT01, MRT02) or the Mount Rosa granite (MR14). Biotite, fayalite and other mafic minerals display Fe-oxide and clay alteration.

Figure 2.13 Photomicrographs of fayalite-bearing quartz syenite: A) Early fayalite with elongated ilmenite grains, fluorite, and minor biotite and later aegirine sprays, in coarse-grained perthitic alkali feldspar, albite and quartz, B) twinned albite grains in quartz and perthitic alkali feldspar, with small zircon grains associated with aegirine and biotite, C) large altered fayalite grains with characteristic fractures and fluorite, biotite and zircon from fayalite-bearing quartz syenite adjacent to pegmatite contact, D) rounded fayalite grains with fluorite, allanite (aln), and zircon from fayalite-bearing quartz syenite adjacent to pegmatite contact.
Figure 2.14 False-color TIMA images of fayalite-bearing quartz syenite: A.) MRT02: sample from upper contact of pegmatite transect showing altered perthitic alkali feldspar being replaced by albite, with interstitial fayalite associated with fluorite, biotite, and minor REE minerals, B.) MRT01: upper contact of pegmatite transect, with minor arfvedsonite associated with biotite, fayalite, and fluorite. Ilmenite and Fe-oxides replace fayalite.

Granitic Dikes

Granitic dikes occur throughout the MRC and range from medium-grained, relatively homogeneous granite to aplite with small (20-50 cm) enclaves of coarse-grained granite, to locally pegmatitic material. Granitic dikes are represented by samples MR30 and MR37 (Table 2.1). MR30 was studied in detail in thin section and is a sample of a prominent 1.5-2 m wide dike oriented ~280/90 which cuts both hypersolvus Pikes Peak biotite granite and fayalite-bearing quartz syenite (Fig. 2.1). Fe-Ti oxides, zircon, and monazite/xenotime crystallized first (Fig. 2.11), forming small subhedral to roughly euhedral grains between 10-100 μm.
Zircon in granitic dikes is one of the first minerals to crystallize, forming subhedral, distorted grains between 100-250 µm in size with irregular rims and minor ‘sieve’ texture, with holes inside grains (Fig. 2.15D). Inclusions are sparse, occur throughout grains, and tend to be phases like Pb-rich uraninite, thorite, and monazite/xenotime (Fig. 2.15D). REE fluorocarbonate minerals form minor overgrowths on some zircon grains (Fig. 2.11D). Albite crystallized next, forming lath-like subhedral to somewhat ragged crystals up to ~5 mm, in a matrix of fine-grained rounded quartz blebs, followed by alkali feldspar as well as microcline grains (Figs. 2.15A, B).
Sample MR30 contains a greenish-brown to golden-yellow mica, which occurs as ragged clusters of grains up to ~200 µm (Fig. 2.16). Mica is associated with minor REE fluorocarbonate minerals, which surround and possibly replace it (Fig 2.15C). Fine-grained portions of granitic dikes (samples MR30 and MR37) are relatively equigranular, with approximately equal modal K-feldspar and albite (~26% each) and more abundant quartz (~42%) (Fig. 2.16). Mica minerals and clay alteration of micas account for ~2.5% of sample MR30, with remaining accessory minerals totaling <0.50 modal percent.

Figure 2.16 False-color TIMA automated mineralogy image of Granitic Dike MR30.

*Mount Rosa Granite*

The peralkaline Mount Rosa granite is texturally the most complex and variable unit studied and occurs throughout the MRC. It occurs in the main body forming the eastern and northern slopes of Mount Rosa itself, or as isolated dikes, blobs and sheets up to 500 x 500 m in
the Stove Mountain and St. Peter’s Dome areas below the elevation of Gold Camp Road. Mount Rosa granite (Fig. 2.1) is represented by the following samples (Table 2.1): MR03 (oikocrystic Na-Fe amphibole granite), MR04 (fine-grained K-feldspar-phyric Na-Fe amphibole granite), MR06A (fine-grained Na-Fe amphibole granite), MR06B (contact between coarse-grained enclave and fine-grained Na-Fe amphibole granite), MR25 (Mount Rosa granite dike), MR-2016-06 (elongate K-feldspar crystal-phyric Na-Fe amphibole granite), MR-2016-07 (Na-Fe amphibole biotite granite inlier at contact with Windy Point granite) MR-2016-08 (fine-grained equigranular Na-Fe amphibole granite) MR-2016-13 (Mount Rosa granite dike) MR-2016-10 (altered aegirinized Na-Fe amphibole granite).

The mineralogy of the Mount Rosa granite is dominated by albite, microcline, quartz, and Na-Fe amphibole, with minor accessory biotite, astrophyllite, fluorite, REE minerals, zircon, clays/sericite, and Fe-Ti oxides (Figs. 2.17; 2.18). The most important facies of the Mount Rosa granite is the oikocrystic Na-Fe amphibole granite (sample MR03), consisting of oikocrystic bundles of elongate prismatic to ragged stubby Na-Fe amphibole crystals up to 4 cm wide, but more often 1-2 cm in size (Fig. 2.4A). Pegmatitic areas of this Mount Rosa granite facies and coarse-grained enclaves feature rare, typically randomly oriented Na-Fe amphibole crystals up to ~20 cm in length. Na-Fe amphibole also forms large (up to 10 cm) areas of poikolytic intergrowth with albite, microcline, and alkali feldspar. Fine-grained Na-Fe amphibole granite is the second most important Mount Rosa granite facies, and is characterized by elongated poikolytic Na-Fe amphibole grains up to ~1 cm (though most are <5 mm) which are generally randomly oriented in albite, quartz and microcline matrix. Finally, dikes of Mount Rosa granite ranging from <1 m to ~10 m in width and 10-100+ m in length are characterized by abundant albite with quartz, Na-Fe amphibole, minor microcline, and significant zircon and REE minerals (Fig. 2.17D).

In all Mount Rosa granite facies, large inclusion-poor or inclusion-free microcline and K-feldspar crystals grew first, followed by co-precipitation of quartz and microcline, leading to extensive development of micrographic textures and granophyre (Figs. 2.18A, D). Albite crystallized after quartz and K-feldspar, filling interstices between micrographic intergrowths and earlier microcline crystals.
Figure 2.17 False-color automated mineralogy image of Mount Rosa granite samples, A.) oikocrystic Na-Fe amphibole granite with microcline and albite inclusions in Na-Fe amphibole, B.) fine-grained Mount Rosa granite with large albitized microcline crystals and small poikolytic Na-Fe amphibole grains, C.) MR16: fine-grained Na-Fe amphibole granite with accessory astrophyllite and biotite, D.) Mount Rosa granite dike from St. Peter’s Dome area, with abundant albite, zircon, and REE minerals.

Bimodal distribution of quartz and feldspar crystal sizes is typical, in which larger quartz-feldspar intergrowths form rounded grains or domains which are surrounded by smaller ‘matrix filling’ quartz and feldspar (Fig. 2.18A). Na-Fe amphibole crystallized after quartz and feldspars, incorporating abundant inclusions of quartz, microcline, and albite. Na-Fe amphibole forms skeletal or complex cuneiform textures with fine-grained albite and quartz (Fig. 2.18C).
Minor biotite (typically <1 modal percent), astrophyllite and aegirine, forming small, typically poikilotic grains up to ~2 cm (typically <5 mm) crystallized with Na-Fe amphibole.

Zircon occurs in several populations in the Mount Rosa granite. Early magmatic, subhedral to euhedral zircon occurs relatively sparingly, in grains ranging from <50 to ~500 µm. These are similar to early magmatic zircon observed in other units, with oscillatory zoning typically confined to rims of grains and cores showing patchy zonation and occasional inclusions of thorite, uraninite, monazite, and Fe-Ti oxides. These zircons are typically associated with Na-Fe amphibole, biotite, or other mafic silicates (e.g. astrophyllite, aegirine) and appear to have
been some of the first minerals to crystallize. Early magmatic zircon in the Mount Rosa Granite, however, is minor in comparison to a later zircon population, which can be divided into two sub-populations. The first population shows fairly euhedral to subhedral rounded crystals generally <300 µm in size which exhibit the patchy, irregularly-zoned cores typical of early magmatic zircon but contain sparse to moderately-abundant rounded inclusions throughout them. These inclusions represent all minerals observed in the Mount Rosa granite. Some of these zircon grains also exhibit a poikolytic to almost skeletal morphology in which crudely tetragonal zircon grains have included and grown around earlier prismatic Na-Fe amphibole and albite grains.

Figure 2.19 Backscatter electron images of Mount Rosa granite, A) Y-rich fluorite replacing Na-Fe amphibole, B) Y-fluorite and REE fluorocarbonate minerals replacing biotite and astrophyllite in quartz and albite, C) REE fluorocarbonate grains in astrophyllite grain, D) Y-fluorite and REE fluorocarbonates replacing Na-Fe amphibole in albite.
The second later zircon population appears to be one of the last minerals to crystallize. This late zircon population is characterized by generally large (500 µm to 5 mm) grains with poikolytic to subhedral crystal. Inclusions are abundant, occur throughout grains, and are typically euhedral crystals of albite, K-feldspar and Na-Fe amphibole.

Fluorite in the Mount Rosa Granite crystallized late, commonly surrounding (or surrounded by) Na-Fe amphibole (Fig. 2.17). While early fluorite is Y- and REE-poor, a later generation of Y- and REE-rich fluorite associated with alteration forms veinlets and blebs which replace Na-Fe amphibole (Fig. 2.19A). Associated with this later Y-REE fluorite as well as rimming and locally replacing mafic silicates (Na-Fe amphibole, biotite, astrophyllite) are REE fluorocarbonate and carbonate minerals (e.g.; bastnasite-(Ce), parasite-(Ce), synchysite-(Ce); Figs. 2.19B, D). Rare earth element fluorocarbonates form elongate lath-like grains up to ~100 µm included in Y-REE fluorite. Both replace mafic silicates such as Na-Fe amphibole and biotite (Figs. 2.19B, D). Albitization is pervasive and is intensified in fine-grained areas between larger K-feldspar grains, which are also locally replaced by albite (Fig. 2.17). Finally, late alteration is characterized by alteration of feldspars to clay minerals (kaolinite/sericite) as well as mafic silicate alteration to Fe-oxides and clays.

_Mafic Dikes_

_Mafic dikes_ occur throughout the MRC but are generally clustered into two main areas - Stove Mountain and St. Peter’s Dome at or below the elevation of Gold Camp Road, and the northern and eastern slopes of Mount Rosa (Fig. 2.1). Mafic dikes are represented by samples MR10, MR11, MR26, MR58, MR-2016-15, and MR-2016-17. Mafic dikes in the MRC were originally described as ‘lamprophyres’ by Gross (1962) and Gross and Heinrich (1966) but in modern petrogenetic classification they include hornblende diabase, clinopyroxene diabase, quartz diorite, syenite, and quartz syenite.
Clinopyroxene syenite dikes (MR11) from the northeastern slopes of Mount Rosa are characterized by large euhedral plagioclase phenocrysts up to ~3 x 1.5 cm set in a fine-grained equigranular matrix of pyroxene, biotite, quartz, and feldspar (Figs. 2.20A, 2.22A). In thin section, Fe-Ti oxides, fluorapatite (~2 modal percent), and zircon (0.05-0.08 modal percent) crystallized first. Ilmenite (~3 modal percent) is the main opaque phase and occurs as subhedral to euhedral clusters of grains up to ~500 µm associated with prismatic crystals of fluorapatite (~2 modal percent; Fig. 2.21D). Fluorapatite also occurs as numerous small (~1 x 5 µm) clear prismatic crystals which are scattered throughout the groundmass. Zircon in mafic dikes forms small (generally <50 µm) rounded subhedral grains with patchy zonation (Fig. 2.21D). Large albitic plagioclase (~An20) phenocrysts (~20 modal percent) crystallized next.
Quartz is negligible in clinopyroxene syenite dikes. K-feldspar (~10 modal percent) crystallized next and forms part of the fine-grained groundmass (Fig. 2.20A). Pyroxene (aegirine-augite; ~15 modal percent) occurs in irregular clots and masses up to ~3 mm inter-grown with biotite (~10 modal percent; Fig. 2.22A). Fluorite (~0.6 modal percent) is scattered through the fine-grained groundmass but is often associated with mafic silicates as small scattered grains and masses (Fig. 2.20A). Clay minerals (sericite and kaolinite; ~3.7 modal percent) are fairly abundant and are mainly associated with large altered plagioclase phenocrysts (Fig. 2.20A). Other minor accessory minerals include monazite and other REE minerals. Early plagioclase and K-feldspar in clinopyroxene syenite dikes has been extensively alibitzed (Fig. 2.20A).
Fine-grained equigranular biotite diabase dikes are represented by sample MR26 from a small, ~30 cm wide, vertically-dipping mafic dike with a strike of ~285 which is adjacent to a ~1 m wide dike of Mount Rosa granite (MR25). It is a very fine-grained biotite diabase, with average grain size of <1 mm and visually homogeneous textural appearance (Fig. 2.20B). It differs from clinopyroxene syenite dikes in its higher modal abundance of albitic plagioclase (~51 modal percent) and biotite (~22 modal percent; Fig. 2.20B). Fluorite is also much more abundant (~3.9 modal percent), occurring as small rounded blebs up ~50 µm associated with mafic silicates (Fig. 2.20B). Aegirine-augite in minor (~3 modal percent), as is quartz (~1.7 modal percent). An unidentified Ba-Ti silicate species (~0.5 modal percent) forms bundles of acicular prismatic crystals with a distinctive ‘bowtie’ habit up to ~500 µm in diameter (Figs. 2.21B, 2.22B). K-feldspar in biotite diabase dikes is Ba-rich (Barian Orthoclase). REE minerals are notably more abundant in MR26 than MR11 (.27 modal percent vs. .01 percent, respectively) and occur both as early monazite/xenotime grains associated with zircon, ilmenite andapatite as well as late overgrowths of REE fluorocarbonates on mafic silicates.

**Feldspar-phyric biotite syenite** dikes are represented by sample MR-2016-17, a ~1 m wide dike oriented ~310/75E which cuts pink hypersolvus Pikes Peak biotite granite and is exposed in Gold Camp Road below Stove Mountain (Fig. 2.4E). Albitic plagioclase (~An30) comprises ~40 modal percent, K-feldspar comprises ~15 modal percent, quartz comprises ~10%, biotite comprises ~20 modal percent, and aegirine-augite comprises ~10 modal percent, with the remaining 5 modal percent comprised of apatite, ilmenite, fluorite, REE minerals, and clays (Figs. 2.22C, D). Aegirine-Augite occurs both as fine-grained aggregates (~20 µm clusters) associated with biotite in the groundmass as well as large (up to ~800 µm) rounded megacrysts composed of intergrown prismatic pyroxene grains associated with apatite and minor zircon (Fig. 2.22D). Minor REE minerals occur as allanite and REE fluorocarbonates, typically associated with mafic silicates such as biotite (Fig. 2.21C)
Figure 2.22 Photomicrographs of mafic dikes: A) clinopyroxene syenite dike with large plagioclase (pl) phenocrysts showing sericitisation (sct) in fine-grained albite, biotite, and aegirine-augite (aeg-aug) matrix, B) fine-grained biotite diabase dike with Ba-Ti silicate blades and small apatite (ap) prisms, C) biotite syenite dike with large plagioclase phenocrysts in fine-grained equigranular matrix, D) biotite syenite dike with aggregate of aegirine-augite and biotite with large apatite grains in fine-grained biotite, albite and quartz.

Type-(I) Pegmatites

Type-(I) pegmatites occur throughout the MRC in all rock units described above and is the more abundant of the two pegmatite groups observed in the MRC. Overall, their bulk mineralogy is similar to that of their host rocks. While Type-(I) pegmatites occur in all parts of the MRC, they are more common closer to the main body of Mount Rosa granite as well as on Stove Mountain above Gold Camp Road (Fig. 2.1). Type-(I) pegmatites are represented by epoxy mount samples MR49A, MR49B, MR49C, and (Table 2.1). Type-(I) pegmatites are differentiated from Type-(II) pegmatites based on their host rock contacts and mineralogy (Fig.
Type-(I) pegmatites typically have irregular contacts with their host rocks and lack well-defined zonation (Figs. 2.6A, C). Some Type-(I) pegmatites, however, are clearly intrusive, and form dikes ranging from <10 cm to 2 m thick and 5-100 m long exhibiting crude zonation (Figs. 2.6B, D, F). Type-(I) pegmatites are dominated by quartz, albite, K-feldspar, biotite ± Na-Fe amphibole, with minor accessory zircon, fluorite (Y- and REE-poor), thorite, and REE minerals. Crude zoning is characterized by a wall zone of coarsely intergrown albite, perthitic alkali feldspar, and biotite ± Na-Fe amphibole and a quartz-dominant core sometimes containing large microcline crystals. Na-Fe amphibole can occur in Type-(I) pegmatites hosted in both Pikes Peak and Mount Rosa granite, but is not always present. Accessory rare element minerals (e.g. zircon, thorite, pyrochlore-group phases, REE minerals) are associated with fluorite in irregular zones and pods and are not in any particular zone of the pegmatite.

Small (2.5 x 3 m) irregularly-shaped and poorly-zoned Type-(I) pegmatites hosted in Mount Rosa granite are most abundant around several thorium exploration trenches from the 1950’s (Fig. 1.4), and have been studied using automated mineralogy and FE-SEM imaging and analysis. The major mineralogy of the sample is dominated by albite (~35 modal percent) with lesser K-feldspar (~25 modal percent) and quartz (~25 modal percent). Altered Na-Fe amphibole which has largely been replaced by Na-pyroxene (aegirine), accounts for ~5 modal percent of the sample, and is associated with minor biotite. The remaining ~10 modal percent of the sample are comprised of fluorite (~4 modal percent), Fe-Ti oxides (~2 modal percent), zircon (~2 modal percent), and HFSE minerals, which include thorite, columbite, cassiterite, Y-Nb-Ti oxides, and REE fluorocarbonates.

In FE-SEM images and analyses, a complex paragenesis of HFSE minerals can be seen (Fig. 2.23) highlighting the fact that while in general, Type-(I) pegmatites are mineralogically-simpler than Type-(II) pegmatites, strong variability and local complexity can be seen. Rutile/brookite forms rounded grains up to ~200 µm which are often rimmed by an altered Y-Nb-Ti oxide mineral (Fig. 2.23B). This pyrochlore-like mineral contains irregular domains which are either Y or Pb-rich and occurs in grains up to ~500 µm and is associated with fluorite and biotite (Fig. 2.23D). Rutile/brookite grains are frequently brecciated by late quartz and biotite veinlets, and are frequently rimmed by cassiterite (Fig. 2.23B).
Figure 2.23 Backscatter electron images of MR49A, a Type-(I) pegmatite in Mount Rosa granite: (A): large zircon with inclusions of albite and K-feldspar with columbite, (B): rutile with cassiterite overgrowth cut by late quartz, (C): “snowball” type zircon grain with inclusion-rich core and inclusion-poor rim, overgrown by Y-Nb-Ti oxide mineral, (D): Y-Nb-Ti oxide, locally Pb-rich, replacing rutile, (E): hexagonal REE fluorocarbonate (possibly bastnasite) in quartz, cut by late fluorite veinlets, (F): Large zircon grain with quartz and albite inclusions oriented along growth planes.
Columbite occurs as rounded grains up to ~200 µm which contain irregular domains rich in U and Th. Thorite occurs as isolated rounded grains and clusters of grains up to ~10 cm in size in Type-(I) pegmatites. Zircon in Type-(I) pegmatites occurs as large (up to ~800 µm) euhedral grains with irregular inclusions of quartz, feldspar and other minerals which are oriented along growth domains in crystals (Fig. 2.23F). Some zircon grains contain cores, which are rich in small inclusions of thorite and monazite/xenotime, surrounded by inclusion-poor rims (Fig. 2.23C). The zircon grains are overgrown by thin rims of Hf-rich zircon and REE fluorocarbonates. Other zircon grains in Type-(I) pegmatites are small, rounded, contain few inclusions, and appear to have crystallized early.

In Type-(I) pegmatites, zircon, thorite, and Fe-Ti oxides crystallized first, followed by perthitic alkali feldspar. These minerals are followed by albite, quartz, and later co-precipitating Na-Fe amphibole and biotite. Low-REE fluorite is often associated with biotite and other mafic silicates, including astrophyllite. REE minerals, including the pyrochlore-like Y-Nb-Ti oxide, columbite and a second zircon generation are late in the crystallization sequence. Albitization of early perthitic alkali feldspar occurred, as well as aegirinization of Na-Fe amphibole and hematization of mafic silicates and Fe-Ti oxides can be observed. Both aegirinization and hematization are especially pervasive in Type-(I) pegmatite hosted in the main body of Mount Rosa granite (Fig 2.6A). Hydrothermal quartz with minor secondary biotite/chlorite fractured earlier minerals like zircon and fluorite, and feldspar breakdown to clays occurred (Fig. 2.23).

**Type-(II) Pegmatites**

Type-(II) pegmatites (Fig. 2.7) are subordinate to Type-(I) pegmatites but are also widely distributed in the MRC. Generally, they are hosted in hypersolvus Pikes Peak granite (subordinately in fayalite-bearing quartz syenite and co-occuring with mafic dikes), at least 800-1000 m away from the contact with the main body of Mount Rosa granite.
Figure 2.24 Summary of pegmatite mineral paragenesis; Type-(I) pegmatites contain only minerals highlighted in yellow, Type-(II) pegmatites contain all minerals (yellow and orange).
Type-(II) pegmatites differ from Type-(I) pegmatite chiefly in their mineralogy (Fig. 2.24): in addition to the aforementioned Type-(I) pegmatite minerals and crystallization sequence, they contain astrophyllite, sulfides (galena >> sphalerite), pyrochlore-group minerals, Y-REE-rich fluorite, genthelvite/helvite, monazite/xenotime, REE fluorocarbonates (synchysite-(Y), bastnasite-(Ce)), and aluminofluoride minerals (e.g. cryolite, weberite, prosopite, pachnolite, elpasoite, ralstonite/thomsenolite; Fig. 2.7F).

Additionally, Type-(II) pegmatites have sharp, intrusive contacts with their host rocks, and typically contain large crystals oriented perpendicular to the pegmatite contacts (e.g.; Na-Fe amphibole; Figs. 2.7A, B). Minerals, which are typically present as minor accessory phases can form large mono-mineralic masses in some Type-(II) pegmatites (e.g. zircon; Fig. 2.7E). Late brecciation of hmhearely magmatic minerals is common (Fig. 2.7D).

Type-(II) pegmatites are principally represented in this study by a transect taken of a ~1m thick pegmatite oriented ~250/30S which is well-exposed in Gold Camp Road just west of Tunnel 4 (MR41, Fig. 2.1). From this pegmatite, a series of 10 thin sections were taken: MRT01, MRT02, MRT03A, MRT03B, MRT03C, MRT04, MRT05, MRT06, MRT07, and MRT08. These samples represent a progression from fayalite-bearing quartz syenite at the pegmatite hanging wall contact (MRT01 and MRT02) into the upper wall zone of the pegmatite (MRT03A, MRT03B, MRT03C, and MRT04), the core zone (MRT05 and MRT06), the lower wall zone (MRT07) and finally the footwall contact with hypersolvus Pikes Peak biotite granite (MRT08). This sample set was systematically examined for changes in mineralogy, geochemistry and texture using transmitted light petrography, automated mineralogy, BSE imaging, EDS analyses, and EPMA analysis of mica chemistry.

Fayalite-bearing quartz syenite at the pegmatite hanging wall contact (10-20 cm from pegmatite contact, MRT01) is typical of fayalite-bearing quartz syenite- in hand sample it is a medium to coarse-grained greenish-gray equigranular rock with large perthitic alkali feldspar and microcline crystals with interstitial clots of biotite and mafic minerals. In thin section, it is also typical of fayalite-bearing quartz syenite, with major perthitic alkali feldspar, microcline, and quartz (Figs. 2.25A, B). Quartz comprises ~30 modal percent of the sample, with K-feldspar and albite each comprising approximately 20 modal percent (Fig. 2.25B).
Figure 2.25 (A) Type-(II) pegmatite transect showing changes in magmatic mineralogy as seen in false-color automated mineralogy images of select areas of representative samples. (A) overview of ~80 cm thick pegmatite outcrop (B), hanging wall fayalite syenite showing fayalite with fluorite and biotite in quartz and perthitic alkali feldspar, (C) same fayalite syenite as in A, (D), biotite with albite and arfvedsonite rim in pegmatite contact, (E), albite and quartz with altered arfvedsonite in core zone, (F) altered arfvedsonite, quartz in core zone, (G) biotite, arfvedsonite, and perthitic alkali feldspar in wall zone, (H) hypersolvus biotite granite in footwall.
Figure 2.26 Type-(II) pegmatite transect showing changes in hydrothermal mineralogy across sample set as seen in backscatter electron images and cross-polarized light photomicrographs of select samples: A.) overview image of pegmatite, showing wallrock contacts and zonation, B.) magmatic allanite grain altered to REE fluorocarbonates and fluorite in hanging wall fayalite-bearing quartz syenite, C.) fayalite grains associated with allanite and fluorite in fayalite-bearing quartz syenite, D.) biotite grains with overgrowth of albite and Na-Fe amphibole in pegmatite wall zone, E.) weberite and prosopite in fluorite in pegmatite core zone, F.) large Na-Fe amphibole crystal altered to Al-fluorides, fluorite and clays in pegmatite core zone, G.) large REE fluorocarbonate grain with Na-Fe amphibole and biotite in pegmatite footwall hypersolvus biotite granite.
Biotite (~11.7 modal percent) is associated with early magmatic zircon (~0.5 modal percent), fluorite (~2 modal percent), Fe-Ti oxides (~3.7 modal percent; Fe-oxides > ilmenite), and altered fayalite grains (~3.5 modal percent; Fig. 2.25B) in sample MRT01. Na-Fe amphibole comprises ~1.6 modal percent of the sample and overgrows as well as locally replaces biotite. Allanite is the most abundant REE mineral in MRT01, and appears to be an early magmatic phase. In BSE imaging, it can be observed that later REE fluorocarbonates overgrow and replace allanite (Fig. 2.26B). Additionally, apatite inclusions in altered fayalite in MRT01 show rims of xenotime which are distinctive in BSE. Pethitic alkali feldspar and microcline has been extensively albitized (Fig. 2.25B), but biotite is largely unaltered.

Fayalite-bearing quartz syenite at the pegmatite hanging wall contact (0-10 cm from pegmatite contact, MRT02) is similar in grain size and mineralogy to sample MRT01, but contains more fayalite (~6.5 modal percent) and also has approximately twice as much albite (~40 modal percent) as MRT01, reflecting more intense albitization close to the pegmatite contact (Fig. 2.25C). Additionally, REE minerals (REE-F minerals, allanite) are more abundant accessory phases in this sample compared to MRT01 (Fig. 2.26C).

Upper wall zone of the pegmatite: Sample MRT03A is the first of three thin sections, which cover a continuous span of ~15 cm across the pegmatite hanging wall contact with fayalite-bearing quartz syenite (Fig. 2.27C). Sample MRT03A is from the pegmatite contact and shows an abrupt textural change approximately halfway through the thin section where the pegmatite contact is. While the modal abundance of quartz, K-feldspar and albite is approximately the same, mafic silicate and REE accessory mineralogy is markedly different in the outermost pegmatite wall zone, with large blebs of Na-Fe amphibole up to ~1 cm which has replaced and overgrown biotite (Figs. 2.25D, 2.26D). This Na-Fe amphibole has in turn been replaced by Na-pyroxene (aegirinization) and finally, a mixture of fluorite and clays (ferrosaponite, kaolinite) has overgrown and locally replaced these mafic silicates. Minor fine-grained REE fluorocarbonates are associated with secondary fluorite. Euhedral, early zircon crystals up to ~200 µm are scattered throughout these clots of altered mafic silicates, fluorite and clays. Aegirinization takes the form of replacement by groups of acicular green prisms, or cracks in subhedral to euhedral Na-Fe amphibole grains which are replaced by green aegirine. In Sample MRT03B, small astrophyllite prisms also occur with Na-Fe amphibole and biotite. Mafic mineral clots in this sample are associated with fine-grained areas of quartz and albite.
Additionally, intergrowths of biotite, Na-Fe amphibole, and astrophyllite occur which appear to have all crystallized contemporaneously. Fluorite is an increasingly abundant accessory mineral and is associated with mafic silicates and REE minerals, comprising ~10 modal percent of MRT03A, versus <1 modal percent in MRT02. Additionally, sample MRT03B marks the first appearance of a distinctive texture in which biotite grains have a thin rim of fine-grained albite separating them from both Na-Fe amphibole overgrowths as well as inclusions of Na-Fe amphibole rimmed by albite. This texture (Figs. 2.25D, 2.26D) is abundant in sample MRT03C, where large biotite grains contain Fe-oxides and clay minerals along cleavage planes and are surrounded by a thin rim of fine-grained albite with lesser K-felspar, in turn surrounded by Na-Fe amphibole.

Sample MRT04 is from the upper wall zone of the pegmatite (Figs. 2.25A, E), and is composed dominantly of albite (~52 modal percent), quartz (~33 modal percent) and fluorite (~5 modal percent). The remaining ~10 modal percent of the sample is very mineralogically-complex and fine-grained and will require further characterization, but can be summarized as mafic silicates (dominantly Na-Fe amphibole with lesser biotite) in various stages of replacement by fluorite, quartz, clay minerals, and minor REE fluorocarbonate phases (Fig. 2.25E). Sample MRT04 has large (up to ~1 cm) grains of Na-Fe amphibole which have been almost completely replaced by a mixture of fluorite, quartz, and clays (Fig. 2.25E). Relict areas on Na-Fe amphibole show the approximate outline of former cleavage planes in the amphibole and demonstrate that it was a continuous crystal before alteration.

Sample MRT05 is from the pegmatite core zone (Fig. 2.27A, B) and is different from MRT04 in that it is composed dominantly of microcline (~50 modal percent) with lesser quartz (~30 modal percent), fluorite (~10 modal percent), sericite/clays (~5 modal percent), and altered aluminofluoride minerals (~5 modal percent). Sericite and clay alteration cuts grid twinning in large microcline crystals in MRT05 and is moderate to pervasive. Fluorite is associated with altered aluminofluoride minerals and forms irregular but continuous veinlets and blebs which cut quartz and microcline. These veinlets are late relative to primary magmatic minerals in the pegmatite core. Zircon occurs both large inclusion-rich grains and smaller euhedral to subhedral crystals which are isolated in quartz and show oscillatory zoning.
Figure 2.27 Transect Pegmatite Field Photos A.) Large pod of pinkish-white cryolite (Cry) in pegmatite core zone with large Na-Fe amphibole (Arf) and microcline (Mcl) crystals and lesser quartz, albite, and fluorite (Fl), B.) close-up of cryolite showing pseudo-cubic cleavage and alteration around Na-Fe amphibole crystal, C.) upper contact of pegmatite showing coarsening grain size away from fayalite-bearing quartz syenite, D.) pegmatite core zone with leached Al-fluoride minerals (e.g. cryolite).
MRT06 is from the central pegmatite core and was collected on the margins of a large rectangular vug where cryolite and other aluminofluoride minerals have been weathered out of microcline and quartz, leaving a cellular ‘boxwork’ structure (Fig. 2.27D) defined by cryolite cleavage planes and filled by powdery white geiksutite. Relatively fresh cryolite also occurs in this zone (Figs. 2.27A, B) in pods up to ~20 cm and shows the characteristic pseudo-cubic cleavage and light pink to white color. Sample MRT06 is dominated by altered microcline, quartz, and fluorite (Fig. 2.25F). Fluorite, with lesser quartz and clays, has replaced large Na-Fe amphibole grains, which display relict amphibole cleavage planes (Fig. 2.26F). Aluminofluoride minerals have been largely replaced by fluorite as well, but where preserved show the following paragenesis (Fig. 2.26E): weberite \((\text{Na}_2\text{MgAlF}_7)\) crystallizes first, along with fluorite-(I) and is replaced partly by prosopite \((\text{CaAl}_2\text{F}_8\text{OH})\) and fluorite-(II). Euhedral, transparent crystals of zircon up to ~200 µm occur in fluorite in this assemblage. Rare grains of barite up to ~200 µm also are associated with aluminofluoride minerals (Fig. 2.25F). Veinlets of fluorite also extend out into surrounding quartz from aluminofluoride mineral replacement zones. Minor cassiterite was also associated with aluminofluoride minerals, and, while not observed in MRT06, most pegmatites in the MRC containing significant aluminofluoride minerals also contain galena, in sparse but coarse-grained masses up to ~15 cm.

Sample MRT07 is from the lower wall zone of the pegmatite, which is thinner (~20 cm) than the upper wall zone (~40 cm). It is similar to samples MRT03A-MRT03C from the upper wall zone, containing large biotite grains replaced by a fine-grained mixture of Na-Fe amphibole and fluorite, as well as grains of REE minerals associated with fluorite (Fig. 2.25G). Major minerals are albite, K-feldspar and quartz.

Sample MRT08 is hypersolvus Pikes Peak biotite granite immediately adjacent to the pegmatite contact. Large perthitic alkali feldspar crystals and quartz dominate the sample, with significant late albitization of perthite textures (Fig. 2.25H). The distinctive texture of biotite surrounded by a thin albite rim, in turn rimmed by Na-Fe amphibole is seen, but some ‘cores’ of these intergrowths are composed of REE fluorocarbonates (bastnasite) instead of biotite (Figs. 2.25H; 2.26G). It is unclear if the REE fluorocarbonates are a replacement of biotite or a primary late-magmatic mineral. The cores of these REE fluorocarbonate grains are composed for Fe-rich clay minerals, suggesting they are a replacement of biotite and/or Na-Fe amphibole. Similarly,
fine-grained intergrowths of fluorite, albite, and Na-Fe amphibole which are intimately associated with biotite may be a replacement of biotite.

2.4.2 Mineral Chemistry

*EPMA Major and Minor Element Chemistry of Biotite*

Biotite mica (a series between the joins annite-phlogopite and siderophyllite-eastonite) occurs in all studied lithologies in the MRC and typically occurs in the same position in the crystallization sequence. Unaltered biotite grains were chosen for analysis, and approximately 3 analysis points in each of 3-5 grains were chosen for each of the 17 samples representing each major rock type as well as various portions of the MRT pegmatite transect samples. Representative compositions of analyzed biotites are shown in Table 2.2.

![Figure 2.28 Compositional variation in analyzed MRC biotites after Deer et al. (1962).](image_url)
Compositionally (Fig. 2.28), biotite from Pikes Peak granite samples MR01 (subsolvus), MR09 (hypersolvus), MRT-2016-17 (hypersolvus) and MRT08 (hypersolvus) are Mg-poor with \( \text{Fe/(Fe + Mg)} \) p.f.u. close to 1.0 and \( \text{Al} \) p.f.u. ranging from 1.60-2.60. Sample MR34 (subsolvus) forms a distinct cluster around \( \text{Fe/(Fe + Mg)} = 0.70 \) and \( \text{Al} \) p.f.u. ~2.45 (Fig. 2.28). Two data points from biotite adjacent (~1 cm away) from a mafic dike contact in MR-2016-17 plot by themselves below this cluster with \( \text{Fe/(Fe + Mg)} \) of ~0.40 (Fig. 2.28). Fayalite-bearing quartz syenite clusters with most biotite from the pegmatite transect and Pikes Peak granite samples, representing biotite-siderophyllite. Biotite in sample MR30 (granitic dike) is Mg-poor with \( \text{Fe/(Fe + Mg)} \) ~1.0, but very Al-rich and is close to siderophyllite end-member composition (Fig. 2.28). Biotite in some grains of sample MR03 (oikocryst Na-Fe amphibole Mount Rosa granite) and MR04 (fine-grained Mount Rosa granite) approaches annite end-member composition, and is very Al-poor with \( \text{Fe/(Fe + Mg)} \) p.f.u. remarkably close to 1.0, reflecting the very low Mg content of the Mount Rosa granite. Biotite in samples MR16 (fine-grained Mount Rosa granite), MR03, and MR14 (altered Mount Rosa granite near contact) are also very Mg-poor with \( \text{Fe}/(\text{Fe + Mg}) \) ~1.00, but have more variable Al p.f.u., ranging from 1.47-2.17, and thus are annite-siderophyllite (Fig. 2.28). Biotite in mafic dike MR26 has approximately the same \( \text{Fe/(Fe + Mg)} \) value as mafic dike MR11 (~0.70) but plots in a distinct field due to lower Al p.f.u. (~1.5 vs. ~2.5 respectively). Biotite from the pegmatite transect shows fairly consistent \( \text{Fe/(Fe + Mg)} \) values of 0.90-1.00, but Al p.f.u. varies from ~1.4 in the outer wall zone to ~2.0 in the inner wall zone, to ~2.5 in the core zone (Fig. 2.28).

Peralkalinity index (\( \text{Al}/(\text{Ca} + \text{Na} + \text{K}) \)) vs. Si a.p.f.u (A/CNK) shows a negative trend across different units, with A/CNK between 1.0 and 1.40 for hypersolvus Pikes Peak granite, fayalite-bearing quartz syenite, and mafic dike MR11 (Fig. 2.29). The peralkalinity index for biotite from the Mount Rosa granite varies from close to zero for Al-poor grains from samples MR03 and MR04 (fine-grained Mount Rosa granite facies) to ~0.65-1.2 for more Al-rich grains in MR03 (oikocrystic Mount Rosa granite facies) and samples MR16 and MR14 (fine-grained Mount Rosa granite facies Fig. 2.29). In samples of Pikes Peak granite, (\( \text{Na} + \text{K} \))/(\( \text{Al} + \text{Ca} \)) versus Si p.f.u values are mainly clustered between 0.8 and 1.0, and samples from the pegmatite core show substantial variability (Fig. 2.29). K and Ca show a negative correlation in all analyzed biotite, though most are clustered around K ~1.8-2.2 p.f.u and Ca <0.20 p.f.u.
Ca content of MRC biotites is low, though some fayalite-bearing quartz syenite and Mount Rosa granite samples contain Ca up to .18 p.f.u (Fig. 2.31). Biotite from Mount Rosa granite, mafic dikes, and Type-(II) pegmatites shows the strongest enrichment in F, though values are variable (Fig. 2.30). Biotite has a more consistent range of Ti across different units (Fig. 2.30).

Figure 2.30 Diagram Illustrating the compositional variety in Mg/(Fe + Mg), F, and Ti biotite from various MRC rocks (diagram after Marks et al. 2008).
<table>
<thead>
<tr>
<th>Lithology</th>
<th>Hypersolvus Pikes Peak Biotite Granite</th>
<th>Subsolvus Pikes Peak Biotite Granite</th>
<th>Fayalite-bearing Quartz Syenite</th>
<th>Mount Rosa Granite</th>
<th>Mafic Dikes</th>
<th>Type-(II) pegmatites</th>
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<td>MR09_A3 MR09_C3 MR17_C2</td>
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* bdl = below detection limits.
*: MRT04 H3 is an F-rich polyolithionite-trilithionite mica from the pegmatite core zone.
Biotite is generally very poor in Mn (<0.10 a.p.f.u), though some Mount Rosa granite biotites contain up to 0.20 a.p.f.u Mn (Fig. 2.31). Biotite from granitic dikes is most enriched in Al (~4.0 a.p.f.u; Fig. 2.31), with Mount Rosa granite biotite being poorest in Al p.f.u.
Halogens (Cl + F) versus Si p.f.u in micas show a systematic increase from fayalite-bearing quartz syenite (Cl + F = 0.50) to the pegmatite core zone (Cl + F ~6.0), though Mount Rosa granite samples show considerable variability in Σ Cl + F (0.39 – 2.40, Fig. 2.32). F-rich mica in the pegmatite core zone (MRT04), transparent/colorless in thin section, is polythionite-trililionite, with a composition (analysis MRT04 K1; Table 2.2) of $K_{1.1}Li_{2.23}Al_{1.22}Si_{4.25}O_{20}F_{3.21}$, with Li content* estimated from Tindle and Webb (1990). This analysis point is not shown in Figure 2.32.

Fig. 2.33 Ti vs. (Mg, Fe, Mn) a.p.f.u in MRC biotite.

Ti and Al show a broad correlation in MRC biotite, with more Ti-rich biotites having more Al, according to the coupled substitution $MgSi_2Ti^{4+}Al^{3-}$-$2$. Additionally, siderophyllite from MR30 (granitic dike; representative composition from MR30) do not fit this trend and has very low Ti and high Al. A slight negative correlation in Ti vs. (Mg + Fe + Mn) can also be seen in Mount Rosa granite biotite (Fig. 2.33).
2.4.3 Whole Rock Geochemistry

Five samples (MR01, MR16, MR26, MR30, and MR31) representing major MRC lithologies which were studied in detail via petrography and EPMA biotite chemistry were submitted for whole rock geochemical analysis by Lithium tetraborate fusion ICP-MS analysis for major, minor and trace elements at ActLabs in Ancaster, Ontario Canada. The results of these analyses are summarized in figures 2.34, 2.35 and 2.36 below and can also be found in Table 2.3 below. All whole rock geochemical data are listed in Appendix B.

Table 2.3 Major Element Whole Rock Geochemistry of analyzed MRC samples.

<table>
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<tbody>
<tr>
<td>Lithology</td>
<td>subsolvus PPG</td>
<td>Mount Rosa Granite</td>
<td>Mafic Dike</td>
<td>Granitic Dike</td>
<td>Fayalite Syenite</td>
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<td>Major Element (wt%)</td>
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<tr>
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<td>1.16</td>
<td>1.43</td>
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<td>Peralkalinity (Al/(Na + K))</td>
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<td>1.27</td>
<td>1.35</td>
<td>1.46</td>
<td>1.41</td>
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</table>

In terms of major element geochemistry (Table 2.3), the 5 submitted samples vary in SiO₂ from 61.75 wt% (diabase dike) and 78.97 wt% (granitic dike). The Mg # (100* Mg/(Mg + Fe³⁺)) varies from 0.51 in Mount Rosa granite sample MR16 to 3.48 in subsolvus biotite granite sample MR01 (Table 2.3). Peralkalinity (Al/(Na + K)) varies from 1.27 (Mount Rosa granite sample MR16) to 1.46 (Granitic dike MR30). Titanium varies from 0.04 wt% in granitic dike MR30 to 0.51 wt% in diabase dike MR26 (Table 2.3). Mn is minor and varies from 0.01 wt%
(MR30) to 0.26 wt% (MR26). Calcium is also minor and varies from 0.16 wt% (MR30) to 1.80 wt% (MR26). Al/(Ca + Na + K) varies from 1.16 (MR26) to 1.43 (MR30). When plotted on a TAS diagram (Fig. 2.34), analyses range from alkali syenite (mafic dike) to alkali granite (subsolvus Pikes Peak granite, Mount Rosa granite, and fayalite granite) to granite (granitic dikes). The slope ([Na$_2$O + K$_2$O] vs. SiO$_2$) is very close to 1.0 between analyses (Fig. 2.34), which shows a linear decrease in Na$_2$O + K$_2$O with increasing SiO$_2$.

![Figure 2.34 MRC Whole Rock analyses plotted on TAS diagram after Cox et al. (1979) with red curve delineation alkali and sub-alkalic fields, and dashed line showing slope of ~1.0 between analyses.](image)

Rare earth elements, normalized to primitive mantle data from Palme and O’Neill (2004), show a general enrichment in the LREE, with a fairly flat HREE profile (Fig. 2.35). This is followed by a relative depletion in Yb and then an enrichment in Lu. The Eu anomaly is present in all analyzed samples, being most significant for granitic dike MR30 and least significant for diabase dike MR26 (Fig. 2.35).
Figure 2.35 Rare Earth Element spider diagram for MRC whole rock analyses, normalized to CI chondrite data from McDonough and Sun (1995).

In terms of other trace element data, samples MR16 (Mount Rosa granite) and MR26 (diabase dike) show the strongest general enrichment in incompatible elements compared to chondrite. MR16 is especially enriched in LREE (e.g.; La, Ce, Nd), Sm, Zr, Y, and HREE (e.g. Yb, Lu). These HFSE are also notably enriched in diabase dike MR26 (Fig. 2.36), which also shows a relative enrichment in certain compatible elements (e.g., P, Ba, Zn; Appendix B). Rb/Sr varies from 6.07 (MR31; fayalite-bearing quartz syenite) to 35.30 in Mount Rosa granite sample MR16, closely followed by granitic dike MR30 (34.22).

In terms of trace element data, samples MR16 (Mount Rosa granite) and MR26 (diabase dike) show the strongest general enrichment in incompatible elements compared to primitive mantle. MR16 is especially enriched in LREE (e.g.; La, Ce, Nd), Sm, Zr, Y, and HREE (e.g. Yb, Lu). These HFSE are also notably enriched in diabase dike MR26 (Fig. 2.36), which also shows a relative enrichment in certain compatible elements (e.g., P, Ba, Zn; Appendix B). Rb/Sr varies from 6.07 (MR31; fayalite-bearing quartz syenite) to 35.30 in Mount Rosa granite sample MR16, closely followed by granitic dike MR30 (34.22).
2.5 Discussion

The results of field work, petrography, automated mineralogy, EDS point analyses, BSE imaging, EPMA analysis of biotite and whole rock geochemistry help elucidate the complex magmatic to hydrothermal history of the MRC.

2.5.1 Zircon as a recorder of magmatic to hydrothermal processes in the MRC

While traditionally considered an ‘immobile’ element, the mobility of Zr and other HFSE in peralkaline systems has been widely documented (e.g.; Rubin et al. 1992, Salvi and Williams-Jones 1996, Hoshino et al. 2012, Gysi and Williams-Jones 2013). As such, zircon paragenesis in peralkaline complexes can be very complex (e.g.; 5 types of zircon at Nechalacho, Canada; Hoshino et al. 2012). Through field work, transmitted and reflected light petrography, automated mineralogy and BSE imaging, the following zircon generation were defined (Fig. 2.33): early-magmatic zircon-(I), late-magmatic zircon-(IIa) and -(IIb), hydrothermal zircon-(A) and -(B).
Zircon-(I) occurs in all units and is one of the first minerals to crystallize (Fig. 2.37A, B, C, D, H, I). Zircon-(I) generally occurs as subhedral to euhedral crystals between <20 up to 500 µm, and exhibit well-defined oscillatory zoning in the rim zone, with cores exhibiting patchy zonation. Inclusions are sparse but occur throughout grains and are typically phases like monazite/xenotime, thorite, uraninite, and Fe-Ti oxides. Subhedral, distorted zircon-(I) grains (100-250 µm in size) with irregular rims and minor ‘sieve’ texture occur in granitic dikes (Fig. 2.38D). Zircon-(I) in mafic dikes forms small (typically <50 µm) rounded grains, with minor inclusions of monazite and thorite (Fig. 2.37H). Interestingly, while biotite diabase dike MR26 contains only small, scattered zircon grains, its shows more than twice as much Zr (1703 ppm; appendix B) compared to any other unit analyzed from the MRC. Zircon-(I) is typically associated with biotite, Fe-Ti oxides, apatite, REE minerals, and Na-Fe amphibole. In many cases, these later minerals appear to have nucleated on early zircon-(I) grains. Zircon-(I) morphology is the typical tetragonal zircon dipyramid with typical aspect ratios of ~1:1 to 1:4. Zircon-(I) in pink hypersolvus Pikes Peak biotite granite adjacent to a mafic dike in sample MR-2016-17 is an exception, with highly elongated crystals showing a ~1:8 aspect ratio. Abundance of zircon-(I) in MRC lithologies ranges from ~0.02-0.13 modal percent (Appendix A), which is typical of peraluminous granitic rocks (Rubin et al. 1993). An interesting feature of many zircon-(I) grains, especially those in early peraluminous units (e.g.; Pikes Peak biotite granite) is disrupted oscillatory zoning and the presence of patchy, irregularly-zoned cores (Figs. 2.37A, B, C; 2.38A). At the Nechelacho REE-Zr-Nb deposit in Northwest Territories, Canada, hosted in a layered magmatic intrusion of peralkaline rocks including aegirine syenite, nepheline syenite, and cumulates, Hoshino et al. (2012) proposed that early magmatic “Type-1” zircon formed in miaskitic syenite and was then altered by an F and CO$_3$-rich hydrothermal fluid which stripped Zr and other HFSE and precipitated hydrothermal rims on these “Type-1” grains (zircon-(I) in the MRC), resulting in the creation of “Type-2” zircon (zircon-(II) in the MRC). Soman et al. (2010) examined zircon from the well-studied Zomba peralkaline granite complex in Malawi, and found zircon with two texturally and chemically different domains: cores which are homogeneous in BSE and are enriched in trace elements, especially HFSE, and rims or outer zones of zircon grains containing abundant pores and inclusions of HFSE minerals, but significantly lower trace elements and non-formula elements such as Ca, Fe, and Al in the zircon itself. They attributed this to an alkaline, CO$_3$-rich hydrothermal fluid which altered the zircon
rims and drove coupled precipitation-dissolution event in which an inward-moving alteration front dissolved trace element-rich zircon and precipitated trace-element poor zircon and HFSE minerals behind it. The occurrence of pores/holes and inclusions in disrupted rims of Type-(I) zircon in MRC rocks, as well as overgrowths of minerals associated with later alteration (e.g.; REE fluorocarbonates; Fig. 2.38D) suggest that some zircon-(I) in the MRC may have been affected by this type of fluid alteration (Fig. 2.39). Finally, while the abundance of Zr versus observed zircon in some rocks such as mafic dikes may be due to the very small grain size and scattered nature of zircon in these rocks, it is also possible that the elevated Zr is due to later hydrothermal zircon which precipitated in these dikes. The high abundance of fluorite (~4 modal percent; appendix A) in mafic dike MR26, which contains 1703 ppm Zr, supports the possibility that the high whole rock Zr in this dike is because of extremely small (<5 µm) hydrothermal zircon grains included in fluorite, which was observed in peralkaline rhyolite laccoliths in the Christmas Mountains in Texas (Rubin et al. 1993). While inclusions in Type-(I) zircon are early magmatic minerals which crystallized along with zircon, the patchy cores (Figs. 2.38A, D) are similar to what are interpreted as unaltered domains by Corfu et al. (2003) and Soman et al. (2010), which are surrounded by altered rims, and finally, thin hydrothermal overgrowths which show oscillatory zoning. However, these patchy cores in zircon-(I) may also represent minor metamict alteration and/or irregular incorporation of trace elements during crystal growth, and are not necessarily indicative of alteration (Corfu et al. 2003).

Zircon-(II) occurs in Mount Rosa granite and pegmatites (both Type-(I) and Type-(II)) in the MRC and has been further subdivided into zircon-(IIA) and zircon-(IIB). Zircon-(IIa) grains are euhedral to subhedral, typically <300 µm in size, and exhibit the patchy, irregularly-zoned cores typical of zircon-(I) but contain sparse to moderately-abundant rounded inclusions throughout (Figs. 2.37E, G). These inclusions represent all minerals observed in their host rocks, and some small, rounded inclusions are very chemically and texturally-complex and may represent melt inclusions (Thomas et al. 2006). Some zircon-(IIa) grains also exhibit a poikolytic to almost skeletal morphology in which zircon grains have included and grown around earlier prismatic Na-Fe amphibole and albite grains (Fig. 2.37G). In Type-(I) pegmatites, zircon-(IIa) occurs as large (up to 800 µm), fairly euhedral grains with irregular inclusions of quartz, feldspar and other minerals which are oriented along growth domains in the crystals (Fig. 2.23A).
Figure 2.37 Backscatter Electron images of zircon in the Mount Rosa Complex: (A, B): early magmatic zircon-(I) in subsolvus biotite Pikes Peak granite showing weak oscillatory zoning (rim) and patchy zonation (core). (C): zircon-(I) from fayalite-bearing quartz syenite showing patchy zoned core and inclusions of monazite and thorite (D): zircon-(I) from granitic dike with inclusions of uraninite/thorite and later REE fluorcarbonate overgrowth (E, G): late-magmatic zircon-(IIa) in Mount Rosa granite dike with rounded inclusions of all other minerals (F): late-magmatic zircon-(IIb) with euhedral crystal inclusions of albite and other minerals (H): small, subhedral zircon-(I) crystals with thorite/monazite inclusions in mafic dike (I): magmatic zircon-(I) with hydrothermal zircon in Type-II pegmatite. (J): Hydrothermal zircon vein brecciated by late fluorite with zircon-free quartz (K): Hf/Sn-rich overgrowths on ‘spongy’ hydrothermal zircon (L): Hydrothermal zircon veinlets in quartz; intimately associated with fluorite.
‘Snowball’ textured zircon-(IIa) grains also occur, with inclusion-rich cores and zoned rims which may be hydrothermal overgrowths (Fig. 2.23C). Some zircon-(II) rims are overgrown by thin rims of Hf-rich zircon and REE fluorocarbonates. Zircon-(IIb) is characterized by generally large (>500 µm up to 5 mm) grains with poikolytic to subhedral crystal form (Fig. 2.37F; 2.38C, D). Inclusions are abundant, occur throughout grains, and generally take the form of euhedral crystals of other minerals, particularly albite, K-feldspar and Na-Fe amphibole. In samples in which both zircon-(IIa) and zircon-(IIb) occur, zircon-(IIb) is typically more abundant. Complex inclusions (Fig. 2.38D) which are represented by multiple minerals (including major rock-forming minerals, e.g. albite, K-feldspar, quartz) as well as HFSE accessory phases (e.g. REE fluorocarbonates, Y/REE fluorite, columbite-tantalite, pyrochlore) also occur in zircon-(IIb). These complex inclusions are similar to what were interpreted as melt inclusions in zircon and other minerals studied by Thomas et al. (2006) and Levashova et al. (2014). Zircon-(IIb) was one of the last minerals to crystallize in the Mount Rosa granite and pegmatites, but overlapped with crystallization of zircon-(IIa).

Van Lichtervelde et al. (2010) found that in hydrous, F-rich pegmatitic and granitic melts, zircon solubility increased dramatically with increasing peralkalinity ((Na + K)/Al). This increased zircon solubility in the peralkaline rocks of the MRC (e.g.; Mount Rosa granite and pegmatites) may explain the late crystallization of zircon-(II) in these units. However, the occurrence of subordinate early-magmatic zircon-(I) in these rocks suggests zircon crystallization may be more complex, and perhaps related to the higher overall abundance of Zr in the Mount Rosa granite melt. Zircon-(IIa) and (IIb) only occurs in peralkaline rocks of the MRC whose whole rock chemistry and biotite composition correlates to very low Al and Ca and elevated Na, K, and HFSE, but their distribution is also highly variable (e.g. .04 modal percent zircon in fine-grained Na-Fe amphibole granite sample MR16, and 2.57 modal percent zircon in Na-Fe amphibole granite dike MR25; Appendix A), which may be related to the apparent strong Mount Rosa granite melt heterogeneity. At the Nechelacho deposit in Canada, Hoshino et al. (2012) characterized late magmatic, inclusion-rich zircon (“type-3”, similar to MRC zircon-(II)) as a late hydrothermal alteration of magmatic zircon and propose that the inclusions/intergrowth minerals are therefore hydrothermal. However, at the Katugin peralkaline-granite hosted REE deposit in Russia, Levashova et al. (2014) studied the trace element chemistry of inclusion-rich zircon and concluded that similar REE patterns of zircon cores and rims and a depletion in trace
elements in rims suggests that they are magmatic zircon grains with ‘postmagmatic’ rims. The mineralogy of the inclusions in Katugin zircon (thorite, REE fluorocarbonates, arfvedsonite, aegirine, Y-fluorite) strongly mirrors that observed in MRC samples and suggests that these zircons formed from a late, evolved melt, which formed the Mount Rosa granite and pegmatites. Levashova et al. (2014) interpret the inclusion-rich cores of Katugin zircon as magmatic, but suggest that rims of these zircons, which are depleted in trace elements (except for Hf) compared to the cores, have a hydrothermal or metamorphic origin.

**Figure 2.38 Zircon in Mount Rosa granite:** A) zircon-(I) with patchy zonation in quartz and albite, B) irregular zircon-(IIA) grain with arfvedsonite, biotite, and quartz, C) zircon-(IIA) grain in albite and arfvedsonite-riebeckite with possible melt inclusions of Y-fluorite, quartz, and other minerals, D) large zircon-(IIB) grain with albite, fluorite, quartz, and REE-F mineral inclusions.

*Hydrothermal zircon* occurs in some Type-(II) pegmatites of the MRC. It occurs in fine-grained veins in blob-like masses of up to ~1 m, which are confined to pegmatite cores and are
chiefly composed of quartz with lesser zircon, fluorite, and accessory minerals. The possible presence of hydrothermal zircon in the MRC was first noted by mineral collector Dean Allum (Allum pers. comm.). Samples were used in an undergraduate research project (Emphraim 2013), investigating the hydrothermal zircon-bearing rocks using optical microscopy, BSE and CL imaging, and EDS analyses.

Hydrothermal zircon-bearing pegmatites are confined to a small group of known pegmatites at or below the elevation of Gold Camp Road on St. Peter’s Dome near Tunnel 4, though their occurrence may be more widespread. Hydrothermal zircon from two pegmatites was examined for this study, the ‘Eureka Tunnel’ pegmatite and a small ~1.5-2 m wide pegmatite, MR38, exposed in Gold Camp Road ~300 m from the Eureka Tunnel (Fig. 2.1). The Eureka Tunnel pegmatite is one of the largest and most strongly-zoned Type-(II) pegmatites in the MRC and features a nearly mono-minerallic quartz core of about 8 m width. This core is surrounded by an intermediate zone containing irregularly distributed veins and blob-like masses of up to 1 m of hydrothermal quartz. This hydrothermal quartz contains abundant (10-50 modal percent) zircon, and minor fluorite (Fig. 2.7F). Adjacent to this outcrop are irregular masses of aluminofluoride minerals up to 1.5 m (Fig. 2.7F). Micaceous green chlorite and clays (e.g. sericite, kaolinite) in masses of up to 20 cm is associated with these zones. The outer - or wall zone – of the pegmatite is characterized by large (up to 1.5 m) pink microcline crystals, albite, and quartz in coarse-grained intergrowths, which grades into a border zone which is mineralogically similar to the surrounding hypersolvus Pikes Peak biotite granite.

Hydrothermal zircon was sub-divided into two populations, hydrothermal zircon-(A) and hydrothermal zircon-(B). Hydrothermal zircon-(A) occurs as small (generally <50 µm) subhedral to euhedral scattered grains with irregular, patchy zonation (Fig. 2.37J, K). Semi-quantitative EDS spectra on zircon-(A) in sample MR53 and MR38 indicate appreciable amounts of Hf and REE (Ce, La). Equant to elongated prismatic crystals of cassiterite of up to 80 µm occur scattered in quartz and as rare inclusions in zircon-(A). In sample MR53, hydrothermal zircon-(A) occurs as small scattered grains in quartz, along with large (>1 mm) euhedral magmatic zircon-(I) crystals with strong oscillatory zoning, which are always surrounded by a ‘halo’ of zircon-free quartz (Fig. 2.37I). Thin veinlets of quartz and fluorite locally cut all minerals in sample MR53. Overgrowths of possible hydrothermal zircon-(A) on spongy, REE-poor zircon occur in MR38 (Fig. 2.39). Hydrothermal zircon in rhyolite at Sierra Blanca, Texas, an extrusive
equivalent of a peralkaline granite was found to occur as similar Hf-rich overgrowths on some magmatic zircon, but a large degree of variability was noted in terms of HFSE content of hydrothermal versus magmatic zircon (Rubin et al. 1989; 1993). Sierra Blanca hydrothermal zircon was also found to be associated with cryolite (Na$_3$AlF$_6$), an indicator of strong enrichment in F (Rubin et al. 1993). In the Christmas Mountains of Texas, metasomatic fluorite deposits in limestone contain up to 14000 ppm Zr (Rubin et al. 1989). At both these occurrences, Zr was interpreted to have been transported as F-and-OH complexes, possibly also in a vapor phase (Rubin et al. 1993).

![Figure 2.39 BSE image of Type-(II) pegmatite sample MR38 showing overgrowths of Hf/Sn-rich hydrothermal zircon on spongy REE-poor zircon.](image)

Hydrothermal zircon-(B) occurs as veinlets of up to several mm but typically <500 µm, sometimes forming complex anastomosing networks (Figs. 2.37J, L). In MR53, hydrothermal zircon-(B) forms thicker veinlets (up to 500 µm) which are brecciated and cut by late fluorite veinlets (Fig. 2.37J). In sample MR38, from a Type-(II) pegmatite exposed in Gold Camp Road ~400 m west of the Eureka Tunnel (Fig. 2.1) hydrothermal zircon-(B) forms large areas of anastomosing veinlets which occur in complex networks in quartz and fluorite (Fig. 2.37L).
Hydorthermal zircon-(A) is only slightly less abundant than zircon-(B) in MR38 and occurs as scattered to clustered small euhedral to subhedral grains which lack defined zonation in BSE and sometimes form small (<50 µm) rosette-like clusters of radiating blebs and acicular crystals (Fig. 2.37K). Qualitative EDX analysis indicates that hydrothermal zircon in MR38 is enriched in Sn, Hf, and REE. Small acicular needles of cassiterite to ~25 µm are sparsely distributed throughout the sample. Similar to MR53, fluorite is intimately associated with zircon in MR38, and late veinlets of quartz cut zircon veins. Some zircon-(A) occurs as fine-grained (<30 µm) intergrowths with bladed Fe-oxide minerals, which appear to have overgrown zircon.

Sheard et al. (2012) documented hydrothermal zircon from the Nechalacho deposit and proposed a complete dissolution-re-precipitation mechanism in which Zr and HFSE ‘stripped’ from early peraluminous to peralkaline rocks by a F-rich fluid and re-precipitated zircon in late hydrothermal vein assemblages. In peralkaline granitic to rhyolite of the Trans-Pecos Magmatic Province, Texas USA, Rubin et al. (1993) suggest that formation of hydrothermal zircon (up to 38,000 PPM Zr in whole rock data) in peralkaline rhyolite of the Christmas Mountains is related to remobilization of Zr in Na-Fe amphibole and/or pyroxene, which is easily altered. Such a mechanism (Fig. 2.40) of Zr remobilization into a fluid responsible for hydrothermal zircon formation seems plausible for the MRC and was suggested for float samples of hydrothermal zircon from the St. Peter’s Dome area of the MRC studied by Ephraim (2013). In the MRC, early, euhedral, magmatic zircon (e.g. zircon-(I) in Fig. 2.37I) exhibit strong oscillatory and sector zoning in cathodoluminescence (CL) and later veinlets of hydrothermal zircon show poorly-defined CL zonation (Emphraim 2013). Therefore, hydrothermal zircon in MRC pegmatites may indicate the occurrence of a F-rich hydrothermal fluid associated with pegmatite formation, which stripped Zr from magmatic zircon in surrounding rocks (e.g. hypsersolvus biotite granite; fayalite-bearing quartz syenite) and then re-precipitated hydrothermal zircon in late type-(II) pegmatite assemblages (Fig. 2.40). Another possible source of Zr for hydrothermal zircon observed in the MRC is the alteration of arfvedsonite and aegirine, which can contain minor Zr (Rubin et al. 1999). This was proposed as the main source of remobilized hydrothermal Zr in rocks of the Trans-Pecos Magmatic Province in Texas, with magmatic zircon contributing only minor Zr (Rubin et al. 1993). Given the abundance of both arfvedsonite and aegirine in the MRC, and its frequent alteration, particularly in late-stage rocks (e.g.; Type-(II) pegmatites), this is also a plausible scenario.
Figure 2.40 Schematic cartoon of zircon in the Mount Rosa Complex.
2.5.2 Magmatic Evolution of the MRC

The petrogenetic evolution of the MRC encompasses a magmatic stage and a subsequent hydrothermal stage. The magmatic evolution of the MRC began with upwelling of alkali basalt in the upper mantle due to crustal thinning at ~1.1 Ga, around the time of the Mid-Continent Rift A-type magmatism (Smith et al. 1999). This alkali basalt assimilated Proterozoic Colorado Front Range crust (PCFRC; Fig. 2.41) during its ascent, and through additional assimilation, fractionation, and crystallization (AFC) processes, yielding the peraluminous syenogranitic to monzogranitic magma which characterizes the bulk (~80%) of the Batholith (Smith et al. 1999; Guitreau et al. 2016). At the southern end of the Batholith, magma was focused in the Pikes Peak intrusive center, located ~15 km northwest of the MRC (Barker et al. 1976). This magma crystallized the hypersolvus Pikes Peak biotite granite which forms the volumetric bulk of the MRC. Hypersolvus biotite granite is the oldest unit exposed in the MRC and is texturally and chemically similar to the pink to white monzo-to syeno-granite which comprises the bulk of the Pikes Peak Batholith (Smith et al. 1999). Other than localized changes in grain size, the modal mineralogy and texture is fairly homogeneous across exposures in the MRC, and it is cut by all other lithologies. Biotite Al/(Ca + Na + K) p.f.u in hypersolvus Pikes Peak granite reflects the peraluminous nature of the Pikes Peak hypersolvus biotite granite. Compared to analyses of ‘pink syenogranite’ from other parts of the Pikes Peak Batholith, the hypersolvus biotite granite of the MRC is elevated in REE and other HFSE (Smith et al. 1999). Biotite chemistry of hypersolvus biotite granite is defined by high Fe/(Fe + Mg) ratios and Al contents (1.8-2.4 p.f.u.), similar to biotite reported from other sodic Pikes Peak Batholith rocks (e.g., Lake George Ring Complex, Sugarloaf granite; Smith et al. (1997); Fig. 2.42). In the ternary diagram of Abdel-Rahman (1994), biotite from hypersolvus biotite granite plots mainly in the A-type granite field (Fig. 2.43), though some analyses plot in the calc-alkaline field as well. The biotite analyses which plot in the calc-alkaline field, however, are from hypersolvus granite at the pegmatite transect contact, and may have been affected by late fluid alteration associated with the pegmatite.

Hypersolvus biotite granite in the MRC is similar to early hypersolvus biotite granites seen in other peralkaline granite complexes (e.g., Strange Lake Canada, Madeira granite, Brazil; Gysi et al. 2013, Costi et al. 2009). At the Pitinga Sn-Zr-REE deposit in Brazil, hypersolvus
biotite granite, including porphyritic phases, is believed to be co-magmatic with albite-rich subsolvus biotite granite as well as peralkaline Na-Fe amphibole granite which are hosted in it (Costi et al. 2009). Zircon (TIMS) age dates on hypersolvus Pikes Peak biotite granite (1085.6 ± 4.1) and on Mount Rosa granite (1088.3 ± 6.8) also suggest their co-magmatic nature (Wayne Premo, personal communication). Basically identical $\varepsilon_{Nd}$ values for Pikes Peak biotite granite and Mount Rosa amphibole granite (-0.6 – 0.0) also suggest that both rock units share an enriched mantle as the origin of the parental melt (Wayne Premo, personal communication).

The next unit that intruded the MRC was the fayalite-bearing quartz syenite. Fayalite-bearing quartz syenite occurs as blob-like to elongate bodies, often occurring in clusters, measuring from 2 x 2 m to 500 x 800 m in size (Figs. 2.1; 2.3D). While volumetrically much less significant than either the Pikes Peak hypersolvus or Mount Rosa granite, the fayalite-bearing quartz syenite is locally significant in the MRC, and shows a frequent spatial association with pegmatites, mafic dikes, and the Mount Rosa granite (Fig. 2.2). Contacts with surrounding rocks are sharp and host rocks are not noticeably altered, nor were chilled margins noted. Noteworthy is the drop like nature of all fayalite-bearing quartz syenite outcrops. Whole rock geochemistry of MRC fayalite-bearing quartz syenite sample MR31 (Table 2.3, Figs. 2.35; 2.36) reveals an enrichment in REE and other HFSE (Th, Zr, Hf) and depletion in certain compatible elements (e.g. Sr, Ba, P) compared to early peraluminous units (e.g. Pikes Peak biotite granite and granitic dikes). Biotite from MRC fayalite-bearing quartz syenite is close to annite end-member composition, and reflects the high Fe$^{2+}/(\text{Fe}^{2+} + \text{Mg})$ values typical of fayalite-bearing granitoids (Mücke 2003). Fayalite grains from MRC fayalite-bearing quartz syenite examined by BSE imaging were found to contain inclusions of apatite (sometimes with xenotime rims), REE fluorocarbonates, fluorite, ilmenite, and hematite (Fig. 2.26C), which all suggest a magma already enriched in Fe, Ti, F, HFSE at the time the first magmatic minerals such as fayalite and zircon crystallized from the melt. Mücke (2003) examined fayalite and its alteration products in fayalite granite from the Nigerian Younger granite and found similar inclusions in fayalite as well as similar associated minerals, which he interpreted as an indication of the dry, Fe-rich, reduced magmatic conditions at the time of crystallization. In MRC fayalite syenite, the presence of early magmatic Fe-rich olivine (fayalite), with ilmenite and other Fe-Ti oxides as well as the presence of early magmatic plagioclase with later alkali feldspar, microcline and quartz all suggest a magmatic environment of low $f_O^2$, low $f_H^2O$, and relatively high temperature (Petrella
et al. 2013). At both the Cobquecura pluton in Chile (Vasquez et al. 2009), and fayalite granitoids in Southern China (Huang et al. 2011), it is suggested that fayalite granitoids originated from high temperature (<960 °C) melting of granulitic metasedimentary rocks in the lower crust during crustal extension and mantle upwelling. In the Misery Lake intrusion in northern Quebec, Canada, fayalite-bearing quartz syenite contains zones of magmatic cumulate fayalite associated with abundant zircon, REE minerals, and other HFSE (Petrella et al. 2013). Fayalite-bearing rocks are interpreted to have crystallized early at high temperatures from an immiscible fayalite-rich liquid saturated in FeO and also rich in alkalis and incompatible elements (Petrella 2013). Vasquez et al. (2009) examined fayalite-bearing granitoids from the Cobquecura pluton in Chile and concluded that they were related to mantle-derived magmas which selectively incorporated crustal material and crystallized in the presence of Fe-rich mafic rocks which acted as a chemical buffer to keep oxygen fugacity low, a necessary condition for the formation of such fayalite-bearing granitoids. In the MRC, a frequent spatial association between fayalite-bearing quartz syenite, pegmatites, and mafic dikes has been noted. All 3 of these units show strong enrichment in Fe, Ti, F and HFSE, and, similar to the role of metapelitic lower crustal xenoliths in buffering fayalite granite crystallization in the Cobquecura pluton, these associated rocks could have buffered fO2 and other elements (Vasquez et al. 2009).

Gibalvo (1993) studied the chemistry of amphiboles from the MRC and found that fayalite-bearing quartz syenite contained amphiboles that plot in a distinctly different chemical field (ferro-edenite) than those from the main Mount Rosa Granite (riebeckite-arfvedsonite), a compositional gap which also suggests that the fayalite-bearing quartz syenite may be different from other MRC lithologies. However, similarities in whole rock geochemistry, biotite chemistry, and overlapping ages (1078.5 ± 7.6 Ma for fayalite syenite vs. 1083.3 ± 1.8 Ma; Wayne Premo, personal communication) suggest that the fayalite-bearing quartz syenite and Pikes Peak granite are likely co-magmatic.

Granitic dikes occur throughout the MRC and cut hypersolvus Pikes Peak granite and fayalite-bearing quartz syenite, have sharp intrusive contacts and have a geometry that is roughly radial to the inferred intrusive center of the MRC (Fig. 2.1, Barker et al. 1975). They often intrude along the contacts with fayalite-bearing quartz syenite bodies, and cut fayalite-bearing quartz syenite (Fig. 2.3E, F). Dikes range from <1 m in width and ~10 m in length to over 1 km in strike length and ~4 m width (Figs. 2.3E, F). Grain size in the granitic dikes is highly
heterogeneous with pegmatitic areas, aplite, and coarse to fine granite, suggesting varying volatile content and crystallization rates (Vernon 2004). Whole rock geochemistry of granitic dikes (Figs. 2.35, 2.36) shows the least enrichment in REE and most other trace elements of any analyzed MRC rocks, though an enrichment in Ta vs. Nb is anomalous for peraluminous A-type granite magmas and is noteworthy (Smith et al. 1999). Whole rock data show a Al/(Na + K) ratio of 1.46 for the granitic dike sample, plotting in the peraluminous field and while still low, the Mg # of 3.45 is high compared to other units in the MRC (Table 2.3). This chemistry is similar to aplite and granite dikes associated with other late potassic (e.g. Lake George ring complex) or sodic (e.g. Sugarloaf pluton) platons in the Pikes Peak Batholith (Smith et al. 1999; Guitreau et al. 2016).

The occurrence of coarse-grained pegmatitic enclaves surrounded by microgranite and aplite also suggest a volatile-rich melt (Vernon 2004). These textural aspects of the granitic dikes, along with their whole rock and biotite chemistry suggests that they originated from relatively early residual peraluminous melt that crystallized the Pikes Peak granite and that they intruded the complex in the form of dikes. The occurrence of Al-rich, Mg-poor siderophyllite mica in granitic dike MR30 (Fig. 2.42) is unique in studied MRC rocks and suggest that the residual, but relatively early magma which formed these dikes was still strongly peraluminous. Desborough et al. (1980) studied the Redskin pluton, a late potassic pluton in the Pikes Peak Batholith ~35 km northwest of the MRC and analyzed biotite, which is similarly Al-rich to that in MRC granitic dikes (Fig. 2.42). While the Redskin stock is evolved compared to the overall Batholith (e.g., significantly enriched in volatiles), its peraluminous nature is similar to its Pikes Peak granite host rock, a relationship that may be similar to that of the granitic dikes and Pikes Peak granite in the MRC.

The genesis of the Mount Rosa granite is the most complex and poorly understood petrogenetic question in the magmatic evolution of the MRC. Field relationships of the Mount Rosa granite do not suggest it to be a separate pluton, as was suggested by previous workers (e.g. Gross 1962, Smith et al. 1999). Sharp intrusive contacts with host rocks have not been observed, but rather gradational, wavy, irregular contacts with surroundings rocks are the norm. Xenoliths of Pikes Peak biotite granite (Fig. 2.4C) have been observed on the edges of the main body of the Mount Rosa granite, as well as rare to moderately abundant xenoliths of Mount Rosa granite in subsolvus Pikes Peak granite. Additionally, the Mount Rosa granite does not form a continuous
pluton—rather it is composed of many tabular, sheet-like bodies which are generally <100 m thick and rarely larger than ~500 x 500 m (Fig. 2.1). Even the ‘main body’ of Mount Rosa granite which comprises the eastern and northern slopes of Mount Rosa shows areas of Pikes Peak granite within it which can be interpreted as A) ‘erosional windows’ into underlying Pikes Peak granite, B) large xenoliths of Pikes Peak granite, or C) areas of biotite-facies Mount Rosa granite. Additionally, pegmatitic areas of the Mount Rosa granite are common, as are variations in mafic mineral content, grain size, and albite vs. K-feldspar, all suggesting significant melt heterogeneity.

Whole Rock geochemistry of Mount Rosa granite (sample MR16) is characterized by \( \frac{Al}{(Na + K)} = 1.27 \), and a Mg # of 0.51, both the lowest seen in analyzed MRC samples. Low Ca (0.27 wt%) and P (b.d.l) in whole rock geochemistry are also consistent with the derivation of the Mount Rosa granite from an evolved late-stage melt fraction (Figs. 2.35, 2.36; Thomas et al. 2006). Whole rock REE concentrations (Fig. 2.35) are high compared to other MRC units, especially for Y and the HREE. MR16 also shows strong depletion in compatible elements such as Ba and P, and strong enrichment in incompatible elements (HFSE (e.g. Zr, Hf, Ti)).

Biotite chemistry for the Mount Rosa granite also reflects a peralkaline parental melt, with very low Al p.f.u, high \((Na + K)\) p.f.u, variable but generally low Ca p.f.u, and elevated F + Cl p.f.u. When plotted in the biotite tectonic discriminant ternary diagram of Abdel-Rahman (1994), MRC biotite mainly falls within the expected A-type granite field (Fig. 2.43). Biotite from okiocrystic Na-Fe amphibole granite is almost pure end-member annite with Fe/(Fe + Mg) = 1.0 and Al 0.0-1.5 (Fig. 2.28). No biotite from sodic PPB rocks analyzed by Smith et al. (1999) had this composition, though annite from the most evolved biotite-arfvedsonite granite of the ~1.1 Ga Red Bluff Granite suite in western Texas was very close in composition to some Mount Rosa granite biotites (Fig. 2.42; Smith Et al. 1997. The Red Bluff granite suite is believed to have been derived from extended fractional crystallization of enriched mantle basaltic magmas with minor crustal assimilation (Shannon et al. 1997; Smith et al. 1997). Previous workers (Barker et al. 1975, Douglass 1993, Smith et al. 1999) suggested that enriched mantle-derived magmas were involved in the formation of the Mount Rosa peralkaline granite.
Figure 2.41 Schematic magmatic evolution path of the Mount Rosa Complex (concept after Barker et al. 1975; 1976; Smith et al. 1999; Guitreau et al. 2016).
According to Thomas et al. (2006), fractional crystallization of a peraluminous, volatile-rich granite melt can progressively enrich the remaining melt with volatiles and incompatible elements to the point where an immiscible peralkaline melt fraction may separate. This melt fraction continues to evolve towards increasing peralkalinity, and concentrates HFSE (such as REE; Thomas et al. 2005). Thomas et al. (2006) argue that evidence of this peralkaline melt may exist from melt inclusion scale up to several km$^2$ areas, such as the Sn-W ore mineralization at Zinnwald, Germany, and the Ehrenfriedersdorf pegmatite in the nearby Erzgebirge Mountains. However, this residual, mobile peralkaline melt fraction is highly reactive, meaning that it can act as a metasomatic agent on surrounding peraluminous rocks (Thomas et al. 2005). The separation of this mobile melt fraction from crystal phases is also seen as critical in the peralkaline melt development, as the addition of Na$_2$O to peralkaline levels nearly doubles the solubility of H$_2$O in the melt (Thomas 2006). If the peralkaline melt fraction does not separate
completely from the peraluminous bulk melt, then its effect is analogous to autometasomatism instead (Thomas et al. 2006). In the MRC, the occurrence of Mount Rosa granite as discontinuous sheets, dikes, sills and blobs with irregular contact with host rocks, as well as similar ages to surrounding units, all suggest that such an evolution may have been possible. The TIMS zircon age date of 1088.3 ± 6.8 (Premo personal communication) and microprobe ages (Pb/U in zircon) of 1083 ± 3.5 (Premo Personal Communication) and 1079.9 ± 3 are similar in age to various Pikes Peak biotite granite in the MRC area and suggest a co-magmatic origin of the Mount Rosa granite. While the bulk character and mineralogy of the Mount Rosa granite is peralkaline, the actual distribution of albite, Na-Fe amphibole, and other alkali minerals is highly heterogeneous, ranging from >30 modal percent to almost no Na-Fe amphibole and ~15 to over 50 modal percent albite. The occurrence of abundant micrographic quartz-feldspar intergrowths, granophyre, and ‘quench’ textures in which coarse grained intergrowths are surrounded by microgranite rims all suggest a shallow, volatile-saturated environment for the Mount Rosa granite (Vernon 2004; Thomas et al. 2006). Furthermore, many zircon-(II) grains in the Mount Rosa granite contain complex, blob-like inclusions (Figs. 2.38C, D) which may represent melt inclusions (Thomas et al. 2006). These inclusions contain minerals that strongly reflect a peralkaline melt (e.g. Y-fluorite, REE fluorocarbonate minerals, albite, possible cryolite) and are chemically similar to ‘Type-B’ peralkaline melt inclusions observed in the Sauberg pegmatite in Germany by Thomas et al. (2006). While the Sauberg pegmatite is dominantly peraluminous, it contains strongly peralkaline areas, as evidenced by melt inclusions and a paragenesis of coarse-grained Sn-W minerals (Thomas et al. 2006). Thomas et al. (2005; 2006) also propose that after the separation of the peralkaline melt fraction, the remaining melt evolves on a path of increasingly peraluminosity. They also propose that in shallow systems, such as the MRC, a large quantity of alkalis will leave the system due to magmatic degassing, producing a more peraluminous rock than the original bulk magma (Thomas et al. 2006). Subsolvus Pikes Peak biotite granite contains what appears to be primary magmatic albite in addition to perthitic alkali feldspar and microcline, but this could also be the result of a metasomatic overprint (e.g. extensively albitized hypersolvus Pikes granite close to the Mount Rosa granite). Autometasomatism during the separation of the peralkaline melt fraction could have created such a ‘psuedo-subsolvus’ granite (Thomas et al. 2006). Such heterogeneities in both rock unit scale (e.g. Mount Rosa granite vs. subsolvus biotite granite chemistry) and mineral scale (e.g. complex
inclusions in Type-(II) zircon in Mount Rosa granite which show a peralkaline character; Figs. 2.38C, D) are typical, however, of the peraliminous to peralkaline melt transition (Thomas et al. 2005; 2006). Therefore, a scenario is proposed for the MRC in which a larger volume of peraluminous magma (Pikes Peak granite) fractionates and yields a much smaller volume of relatively immiscible peralkaline melt, which is preserved as the Mount Rosa granite and related pegmatites (Thomas et al. 2006).

Figure 2.43 Fe-Mg-Al biotite discriminant diagram after Abdel-Rahman (1994). A = anorogenic alkaline granite suite, C = calc-alkaline orogenic granite suite, P = peraluminous granite suites (including S-type granites). Data from all units of the Mount Rosa Complex.

The next younger unit are the mafic dikes. These dikes, which range from clinopyroxene diabase to biotite-quartz syenite, are anomalous in several respects. Their whole rock geochemistry (diabase dike MR26) is defined by a strong enrichment in REE compared to both other units in the MRC and primitive mantle, especially in the LREE (Fig. 2.35). MR26 also shows a strong enrichment in Zr (1703 PPM) and Nb (146 PPM) as well as other incompatible
HFSE (e.g. Hf, Ti; Fig. 2.36). While F was not analyzed in whole rock geochemical data, the high modal abundance of fluorite (~3.9 modal percent; Appendix A), suggest a significant F content. This enrichment in REE and other HFSE as well as volatiles such as F is consistent with a strong relationship to a mantle-derived magma as proposed by Goldman (1993) who found that certain mafic dikes from the MRC had the most ‘primitive’ chemistry of any analyzed Pikes Peak Batholith rocks. The biotite chemistry of mafic dikes is defined by distinctly-lower Fe/(Fe + Mg) ratios compared to biotite from most other MRC units, as well as other sodic Pikes Peak Batholith rocks (Smith et al. 1999; Fig. 2.42). While biotite from Mount Rosa granite near mafic dike MR11 has similar Al values (~1.5 p.f.u), it is significantly more Fe-rich and Mg-poor than the mafic dike. Biotite from MR11 also has the highest Cl values (0.12- 0.17 p.f.u) seen in MRC rocks. Similarly, biotite from mafic dike MR26, which cuts hypersolvus Pikes Peak granite >1 km from the nearest significant body of Mount Rosa granite, is Mg-rich, with significant F (~1.4-2.0 p.f.u) and relatively low Fe (Fig. 2.42). These characteristics, combined with cross-cutting field relationships, and primitive whole rock geochemistry (a group of analyzed hornblende and clinopyroxene diabase dikes from Stove Mountain was found to have the most primitive geochemistry of any known Pikes Peak batholith rocks (Goldman 1993) as well as recent whole rock geochemical data suggest that the mafic dikes may have had a different parental magma. While some late dikes associated with other plutons in the Pikes Peak batholith are somewhat similar to MRC mafic dikes (e.g. syenite dikes in the Lake George ring Complex; Smith et al. 1999), the petrologic and chemical variety of dikes seen in the MRC is not observed elsewhere in the batholith. Biotite from the ~1.1 Ga A-type Enchanted Rock Batholith (ERB) in central Texas was analyzed by Smith et al. (1997) and plotted in the same field in a Fe/(Fe + Mg) vs. Al diagram (Fig. 2.42) as MRC mafic dikes, as well as subsolvus Pikes Peak granite sample MR34. The Enchanted Rock Batholith and the related Llano Granites are inferred by Smith et al. (1997) to represent anatetic melts derived from juvenile crust, versus granites of the Red Bluff suite in West Texas, which are characterized by biotite with high Fe/(Fe + Mg) (Fig. 2.42) and are interpreted to have been derived from extended fractionation of mantle-derived basalts.

The presence of Barian orthoclase and unusual Ba-Ti silicate minerals in the studied dike samples, and the peraluminous character of their whole rock geochemistry, reflected in their content of more calcic (e.g.; aegirine-augite and hornblende) vs. sodic (e.g.; Na-Fe amphibole and aegirine) amphiboles and pyroxenes, also suggest a different origin for the mafic dikes.
versus the Mount Rosa granite. However, irregular, anastomosing contacts between some mafic
dikes and adjacent pegmatites (Fig. 2.6E), or Mount Rosa granite (Fig. 2.5C) suggest that dike
emplacement overlapped in time with crystallization of the peralkaline Mount Rosa granite and
related pegmatites (Fig. 2.41). The biotites from the Enchanted Rock Batholith in Texas which
plot very close to those from MRC mafic dikes were from the ‘outer zone’ as well as enclaves
within the main granite (Smith et al. 1997), which were interpreted to be co-magmatic with the
main Enchanted Rock batholith phase. Stemprok and Seifert (2012) studied granite-related rare
metal (Sn, W, Mo) mineralizations associated with lamprophyric mafic dikes. They concluded
that these dikes occur principally in roof zones of highly-evolved granites, where they show
often complex anastomosing contacts with adjacent rare-metal bearing veins or pegmatites and
are typically calc-alkaline to alkaline in composition, with high volatile content (e.g. F, Cl, H2O,
CO2). All these characteristics could be applied to mafic dikes in the MRC, as their biotite
chemistry as well as whole rock geochemistry indicates an enrichment in the volatiles F and Cl.
Their occurrence in the field defines the outermost occurrence of rocks associated with the MRC
(e.g., at Helen Hunt Falls >10 km from the summit of Mount Rosa; Figs. 1.3, 2.1), and mafic
dikes often have substantial strike length continuity (up to ~1 km observed for a ~2 m dike). In
many studied examples (e.g. Transabaikal Russia, Variscan provinces of Central Europe),
lamprophyric mafic dikes are late relative to other associated granitic rocks including rare metal
mineralization, and are believed to be the product of fractionated mantle-derived magmas mixing
with felsic crustal melts (Stemprok and Seifert 2012). The geochemistry of MRC mafic dikes
(e.g., enriched in volatiles and compatible elements such as F, Cl, Zr, Nb, REE but also in
compatible elements such as P and Ba; Fig. 2.36) as well as their irregular contact relationships
and frequent spatial association with Mount Rosa granite and pegmatite dikes suggest such a
scenario of emplacement may be likely.

While type-(I) pegmatites appear to be residual melt from crystallization of either the
Pikes Peak biotite granite or the Mount Rosa Na-Fe amphibole granite, respectively, the genesis
of the Type-(II) pegmatites is more complex. Type-(II) pegmatites occur throughout the MRC,
though in they occur overwhelmingly 3-8 km east of the main body of the Mount Rosa granite,
on St. Peter’s Dome, Stove Mountain, and upper/lower Cheyenne Canyon (Fig 2.1). Type-(II)
pegmatites range from 1 x 8 m to 8 x 45 m in size and are typically tabular dikes oriented
approximately E-W to NW-SE with a shallow (~20-40°) dip to the south (Figs. 2.1, 2.7). Type-
(II) pegmatites were studied in detail through a sample transect of a ~1 m thick pegmatite oriented ~250/30S which is well-exposed in Gold Camp Road just west of Tunnel 4 (MR41, Fig. 2.1). This pegmatite is well-zoned and has fairly sharp intrusive contacts with its wallrocks (Figs. 2.25; 2.26). The magmatic crystallization sequence is defined by early perthitic alkali feldspar, which forms blocky crystals up to 20 cm in length, followed by crystallization of finer-grained albite, microcline, and quartz. Na-Fe amphibole crystallization also started relatively early, as evidenced by large (up to 25 cm) prismatic arfvedsonite crystals which have grown inward from the pegmatite walls. The well-developed zonation and unidirectional solidification textures (UST’s) seen in the MRC transect pegmatite are a product of the delay in crystal nucleation and growth caused by significant melt undercooling (London 2005). Changes from crystal nucleation in the melt to the pegmatite sidewall nucleation only, as well as significant temperature gradients between wallrocks, crystallized melt, and still liquid melt also contribute to mineralogical and chemical zonation in such pegmatites (London 2005). Considering that the pegmatite melt was likely highly fluxed in F, OH, and CO₂ (London 2005; 2017), the growth of these large crystals was likely very rapid, and could be compared to the similar-sized (1-2 m thick) but mineralogically different Little Three pegmatite in California, which may have cooled in just ~25 days (Morgan and London 1999).

Biotite analyzed by EPMA from different zone of the transect pegmatite shows interesting chemical features indicative of changes in chemistry of the melt induced by magmatic crystallization. The F-content of biotite changes abruptly from the upper contact zone (~0.30 p.f.u F) to F-depleted (~0.15 p.f.u F) biotite in the wall zone of the pegmatite, to very F-rich (up to 7.0 p.f.u F) polylithionite-trilithionite mica in the core zone, to F-rich biotite in the lower wall zone (~2.0 p.f.u F) to 0.75 p.f.u F at the lower contact (Fig. 2.25). This sharp change in F-content of biotite across sample intervals of only ~10 cm can likely be attributed to the process of constitutional zone refining discussed for granitic pegmatites by London (2013; 2017) in which the inward (from pegmatite walls) crystallization of major minerals such as quartz and feldspars pushes elements such as F which are incompatible in these minerals into a thin ‘boundary layer’ at the crystallization front. Elements like F have relatively slow diffusion rates away from the crystal-melt interface relative to the rapid crystallization rates of pegmatites, and this imbalance creates a highly fractionated ‘boundary’ layer enriched in HFSE and other incompatible elements such as F. In miarolitic pegmatites in other parts of the Pikes Peak Batholith, zoned mica crystals
of F and Li-poor annite overgrown with F and Li-rich zinnwaldite or lepidolite have been described (Kile and Foord 1998). The modal mineralogy of the pegmatite transect reflects a magmatic process in which the pegmatite melt crystallizes from the walls inward, creating a melt progressively enriched in F, OH, and incompatible elements (e.g., REE, Zr, Nb). Biotite from the pegmatite contact and outer wall zones is commonly rimmed by Na-Fe amphibole and albite (Figs. 2.25, 2.26) and zircon occurs as zircon-(IIa) >> zircon-(I). Albite, K-feldspar, and quartz occur approximately in equal amounts, similar to the magmatic crystallization of the Mount Rosa granite. In the pegmatite core zone, however, an abrupt change in mineralogy is defined by albite >> K-feldspar, with large perthitic alkali feldspar crystals to 20+ cm surrounded by albite and quartz (Figs. 2.27A, B). Biotite is absent and instead micas are F-rich triliothionite-polylithionite, and Na-Fe amphibole is less abundant. Finally, the central core zone is defined by late-magmatic cryolite, associated with quartz, fluorite, and minor REE fluorocarbonate minerals and zircon. This is similar to the crystallization sequence and occurrence of late-magmatic cryolite defined at the Pitinga mine in Brazil (Costi et al. 2009) as well as the Katugin deposit in Russia (Sharygin and Vladykin 2014). Neto et al. (2009) suggest that cryolite forms during the last magmatic stages of crystallization from a low-viscosity, highly-fluxed melt which has become supersaturated in F, Na, and Al, but is Si-deficient. Sharygin and Vladykin (2014) analyzed Na-Fe amphibole associated with late-magmatic cryolite from the Katugin deposit in Russia and found it to be extremely F-rich and representing the new species Fluor-arfvedsonite.

Detailed study of the MRC transect pegmatite suggests that the bulk chemistry of the pegmatite may be similar to the Mount Rosa granite. While the occurrence of significant cryolite, fluorite, and F-rich mica in the core zone indicates a strong enrichment in F, and the absence of F-minerals and F-rich mica in the wall zone (composed essentially of albite, microcline, and quartz) may mean that the ‘average’ F-content across the pegmatite is no different than the F-enriched Mount Rosa granite. To truly understand the connection between the Mount Rosa granite and pegmatites, however, trace element and isotope analyses of select minerals would be necessary.
2.5.3 Hydrothermal Evolution of the MRC

While the magmatic evolution of the MRC is critical in understanding the genesis of the complex and the formation of peralkaline rocks within it, the hydrothermal evolution of the complex is equally complex and responsible for many of the mineral assemblages we observe today, including many REE and other HFSE minerals. Hydrothermal alteration and metasomatic processes have affected all rock units in the MRC.

Albitization, or Na-metasomatism, has affected all rock units in the MRC. Albitization varies from minor in hypersolvus Pikes Peak biotite granite on the outer edges of the MRC to pervasive in pegmatites and Mount Rosa granite. In hypersolvus biotite granite, perthitic alkali feldspar grains are altered to small twinned prismatic albite grains along their edges, which is evident in both transmitted light petrography and automated mineralogy (Figs. 2.10, 2.11). Fayalite-bearing quartz syenite contains both primary magmatic albite (larger crystals which are among the first minerals to crystallize) and later albitization, which commonly replaces large perthitic alkali feldspar grains (Fig. 2.14A). Albitization is more intense in subsolvus biotite granite as well as Mount Rosa granite. In some Mount Rosa granite dikes, almost all primary K-feldspar and alkali feldspar has been replaced by albite (Fig. 2.17D). In both type-(I) and type-(II) pegmatites, albitization is pervasive, and some type-(II) pegmatites have large areas of platy ‘cleavelandite’ albite which has replaced magmatic K-feldspar. Albitization intensity appears to be directly correlated with abundance of and distance to the nearest body or dike of Mount Rosa granite, and the albitization process can be compared to that modeled for the Strange Lake Complex in Quebec, Canada (Nassif 1993; Gysi and Williams-Jones 2013), as well as the Madeira granite in Brazil (Costi et al. 2009). According to these authors, a NaCl-bearing high-temperature orthomagmatic fluid drives Na metasomatism, in which Ca$^{2+}$ and K$^+$ is replaced by Na$^+$. Additionally, K-metasomatism, in which albite lamellae in perthitic alkali feldspar are replaced by microcline, was observed in the pegmatite transect, and is likely similar to K-metasomatism was observed elsewhere (e.g. Strange Lake Quebec; Gysi and Williams-Jones 2016).

Alteration of Na-Fe amphibole to Na-pyroxene (aegirinization) affected Mount Rosa granite and pegmatites, and following albitization. Aegirinization was accompanied by replacement of biotite by Na-Fe amphibole in fayalite-bearing quartz syenite and Pikes Peak
biotite granite bordering Mount Rosa granite and pegmatites (Fig. 2.25A; H). Alteration of mafic silicates (e.g. Na-Fe amphibole and biotite) to Fe-Ti oxides and clays (e.g. goethite, hematite, ferrosaponite, kaolinite/sericite) also followed albitization. Unlike aegirinization, however, this hematization occurred in all units in the MRC. Mount Rosa granite close to clusters of Type-(I) pegmatites has been strongly altered, presumably by fluids associated with the generation of these pegmatites from melt pockets, and show significant volume loss and porosity (e.g. vuggy/” leached” textures), with feldspars altered to late albite and finally clay minerals, complete Na-Fe amphibole replacement by aegirine, and finally aegirine breakdown to Fe-Ti oxides and clays. Ratty texture and reddish Fe-oxide staining is characteristic of the aegirinization/hematization stage of alteration in the Mount Rosa granite (Figs. 2.5A; B) and pegmatites (Fig. 2.6A). Similar to the role of acidic volcanic fluids in ‘ground preparation’ for subsequent hydrothermal mineralization in epithermal systems, this early acidic alteration has been shown to be important in creating open spaces for subsequent HFSE mineralization in peralkaline granites (Salvi and Williams-Jones 1990; Gysi and Williams-Jones 2013).

Albitization and aegirinization/hematization were followed by Ca-F metasomatism, which affected all units in the MRC and is characterized by the occurrence on Y-rich fluorite associated with REE-F minerals. While REE-F minerals form minor magmatic accessory phase in some units (e.g., Mount Rosa granite), their principal occurrence is as late veinlets and pseudomorphs of early magmatic minerals, particularly mafic silicates (e.g. Na-Fe amphibole, biotite; Fig. 2.18). This alteration of mafic silicates can vary from minor (e.g. thin Y-rich fluorite veinlets cutting Na-Fe amphibole; Fig. 2.18A) to pervasive (complete replacement of Na-Fe amphibole grains by fluorite, quartz and clay minerals; Figs. 2.25 E, F). In larger type-(II) pegmatites, it can be seen in hand sample scale, where Na-Fe amphibole crystals up to 5-10 cm have been completely replaced by fluorite, quartz, clays, and minor REE-F minerals. This replacement of mafic silicates (e.g. Na-Fe amphibole) may actually involve both the aegirinization/hematization stage as well as the Ca-F metasomatism stage, as suggested for a similar assemblage at Strange Lake, Canada, where quartz fluorite veinlets cut aegirinized arfvedsonite (Gysi et al. 2016). Other mineralogical indicators of Ca-F metasomatism are varied. In sample MRT01 from the upper contact of the pegmatite transect, large magmatic allanite-(Ce) grains are replaced by a mixture of REE fluorcarbonates, fluorite, and clays, likely involving the reaction of allanite with a late F and REE-rich fluid (Fig. 2.26B). Late-magmatic cryolite
(Na$_3$AlF$_6$) and weberite (Na$_2$MgAlF$_7$) are replaced by Ca-bearing fluoride minerals such as pachnolite (NaCaAlF$_6$ · H$_2$O), and prosopite (CaAl$_2$(F,OH)$_8$; Figs. 2.26E, 2.44). This paragenesis of aluminofluoride minerals has been reported from other localities (e.g., Ivigtut Greenland, Perga granite Ukraine, Katugin Deposit Russia; Bailey 1980, Koehler et al. 2008, Sharygin and Vladyskin 2014). Fluorite content of MRC pegmatites as well as other lithologies is highly variable, and type-(II) pegmatites which contain exotic Al-fluoride minerals (e.g. cryolite, prosopite) are typically relatively fluorite-poor, whereas some type-(I) pegmatites are very rich in fluorite but poor in rare element minerals. Other mineral assemblages observed in MRC pegmatites also suggest the importance of the Ca-F metasomatism stage in mobilizing HFSE. In Type-(II) pegmatites on Stove Mountain, large (up to 5 cm) crystals of pyrochlore are replaced by columbite, and have small bastnasite overgrowths. Neto et al. (2009) observed a similar assemblage at the Pitinga Mine in Brazil, where magmatic pyrochlore was replaced by columbite and small amounts of native Pb and Pb oxides. In a Type-(I) pegmatite in the MRC, magmatic Ti-oxides (e.g. rutile) were overgrown and replaced by Y-Ti-Nb (pyrochlore group) oxides, as well as cassiterite (Fig. 2.23).

While Ca is an integral component of the Ca-F metasomatism observed in MRC rocks, whole-rock geochemistry and EPMA biotite chemistry suggest that most MRC lithologies (e.g., Mount Rosa granite) are very Ca-poor. This ‘Ca imbalance’ has been addressed by other workers, notably Martin and DeVito (2014) who proposed that late acidic orthomagmatic fluids associated with early alteration (e.g., albitization, aegirinization) scavenges Ca from early magmatic Ca-bearing minerals such as plagioclase and amphiboles and pyroxenes from other surrounding rock units (e.g., Pikes Peak granite). They specifically apply this model to previous work by Gagnon et al. (2004) on the Oregon #3 pegmatite in the South Platte Pegmatite District. Based on fluid inclusion work, Gagnon et al. (2004) proposed that Ca-F fluids responsible for fluorite and REE-F mineralization were internally generated in the pegmatites as acidic F-rich fluids leached Ca from perthitic alkali feldspar and mafic silicates in the pegmatite wall zone.
Figure 2.44 Schematic cartoon of Type-(II) pegmatite crystallization and mineralogy.
Martin and DeVito (2014) argue, however, that the lack of Ca-bearing primary minerals in these pegmatites suggests that while these fluids may have been pegmatite-derived, elements like Ca were derived from interaction with surrounding wallrocks, such as the more Ca-rich fayalite-bearing quartz syenite (1.1 wt% Ca) and mafic dikes (1.8 wt% Ca; Table 2.3). Given that many pegmatites are spatially associated with both these units, these may have provided a source of Ca for the observed Ca-F metasomatism. Gysi et al. (2016) also suggest that Ca involved in Ca-F metasomatism at Strange Lake was likely derived from fluids external to the pegmatites, whereas F was derived from the pegmatites themselves.
CHAPTER 3

CONCLUSIONS AND OUTLOOK

While the magmatic and hydrothermal processes governing the creation of peralkaline granite complexes such as the MRC remain poorly-understood, certain themes and patterns in their petrogenesis can be identified. The importance of enriched mantle magmas, having incorporated crustal material and undergone extended assimilation, fractionation, and crystallization, and the development of both immiscible peralkaline melt fractions and late F-rich hydrothermal fluids can be seen as common themes. Evidence from this study suggest all these processes are important in the formation of the Mount Rosa Complex.

3.1 Summary and Conclusions

While the Mount Rosa Complex consists of a diverse association of granitic rocks whose magmatic relationship is difficult to resolve, the results of this study suggest that all units are essentially co-magmatic and likely evolved from a common enriched mantle source. Early units such as Pikes Peak biotite granite and fayalite-bearing quartz syenite are characterized by a mineralogy, whole rock geochemistry, and biotite chemistry that reflects crystallization from the peraluminous melt which composes the bulk of the Pikes Peak Batholith and was focused around the inferred Mount Rosa intrusive center relatively late in the emplacement of the batholith. Granitic dikes show a similar peraluminous chemistry and their radial geometry around the inferred intrusive center suggest they were emplaced along fractures in Pikes Peak biotite granite. Textural and mineralogical features such as pegmatitic areas surrounded by microgranite, aplitic zones, and subsolvus feldspar textures suggest that the melt which crystallized these dikes was more enriched in volatiles (e.g. F, OH) than that which formed the Pikes Peak granite. A trend in biotite composition can be observed from early peraluminous units from siderophyllite-annite composition with intermediate Al, Na, K, and Ca concentrations and relatively minor F + Cl concentrations to later peralkaline units (Mount Rosa Na-Fe amphibole granite, pegmatites) with end-member annite showing low concentrations of Al and Ca and elevated Na, K, and F + Cl concentrations. Peralkaline mineralogy (e.g. Na-Fe amphiboles and
pyroxenes, albite, astrophyllite) and whole rock geochemical enrichment in incompatible elements (e.g. REE, Y, Zr, Hf) in Mount Rosa granite and pegmatites also reflects a peralkaline parental melt for these units. While previous workers had proposed that the Mount Rosa granite represented a separate peralkaline intrusion (Gross 1962, Barker et al. 1976, Smith et al. 1999), the heterogeneity of Mount Rosa granite outcrops (sill-like bodies, blobs, and dikes) as well as variations in mineralogy and texture do not support this interpretation.

Recent research into the genesis of peralkaline granites (e.g. Thomas et al. 2005; 2006) has suggested that during the end stages of crystallization of an enriched mantle-derived peraluminous melt in high-level A-type granites, when the remaining melt is enriched in volatiles, a much smaller peralkaline melt fraction can separate by immiscibility processes. If such a peralkaline melt fraction separated from the Pikes Peak peraluminous melt at the end stages of magmatism and crystallized as a series of spatially-associated sill-like bodies and dikes, this may explain the observed outcrop pattern, as well as the strong geochemical and mineralogical heterogeneity of the Mount Rosa granite. If the Mount Rosa Complex evolved solely through fractional crystallization, continuous trends in biotite and whole rock chemistry would be expected between units, but this is not the case, with differences between the Mount Rosa granite and associated pegmatites and earlier peraluminous units which suggest a conjugate peralkaline melt fraction may have separated and continued on a different evolutionary path.

Late mafic dikes show characteristics of both early peraluminous units (whole rock enrichment in and mineralogy reflecting compatible elements such as P and Ba), as well as later peralkaline units (strong enrichment in incompatible elements such as Zr and REE, abundant fluorite and albite), and have a less clear magmatic relationship to either the Pikes Peak or Mount Rosa granite parental melts. While earlier Type-(I) pegmatites can be compared to pegmatites in other parts of the Pikes Peak Batholith (e.g. South Platte District; Simmons and Heinrich 1980) which are probably related to residual melt pockets left over after crystallization of their host rocks, Type-(II) pegmatites are chemically and mineralogically-complex dikes hosted in unrelated rock units (e.g., hypersolvus Pikes Peak granite). Their mineralogy and biotite chemistry suggest, however, that their bulk composition may be similar to the Mount Rosa granite and that their strong zonation and textural features can be explained using modern models of pegmatite crystallization (London 2005; 2017), including constitutional zone refining.
Alteration in the MRC can be summarized by three stages, starting with albitization or Na-metasomatism (stage-(I)), which forms a halo around the Mount Rosa granite and pegmatites and likely is related to high-temperature orthomagmatic fluids. In stage-(II), aegirinization and hematization affected mafic silicates in the Mount Rosa granite and pegmatites. Finally, in stage-(III), Ca-F metasomatism associated with an F-rich fluid altered and replaced magmatic minerals with hydrothermal minerals, many containing HFSE (e.g. REE Fluorocarbonates, hydrothermal zircon). These alteration processes are best observed in Type-(II) pegmatites and the studied transect pegmatite shows systematic changes in mineralogy and biotite chemistry which can be attributed to these stages.

Compared to other well-studied peralkaline granite complexes (e.g., Strange Lake Quebec, Pitinga mine Brazil, Katugin Russia), the MRC shows many similarities in terms of both magmatic and hydrothermal evolution, but is decidedly less enriched overall in HFSE minerals as well as minerals indicative of exceptional concentrations of F and other volatiles (e.g. cryolite). It is worth noting that the three aforementioned peralkaline granites are small complexes (less than ~25 km$^2$) and appear to have evolved under relatively closed-system conditions compared to the MRC, which is part of a very large (~3300 km$^2$) peraluminous batholith. If the process of peralkaline melt separation is applied to the MRC, it is plausible to consider that overall alkalinity and the concentration of HFSE and volatiles may have never become sufficiently enriched within the complex to form aspects of other peralkaline granites that make them economic rare element deposits (e.g. disseminated zirconosilicate minerals in subsolvus granite at Strange Lake, large massive cryolite bodies at Pitinga, Brazil, broadly disseminated cryolite-REE fluorocarbonate mineralization at Katugin, Russia; Gysi and Williams-Jones 2013, Costi et al. 2009, Levashova et al. 2014). Under the more ‘open system’ behavior of peralkaline melts proposed by Thomas et al. (2005; 2006), the reactivity of the peralkaline melt fraction as well the possible incomplete separation of this melt fraction may impede the concentration of HFSE and volatiles by both magmatic and hydrothermal processes. The result, therefore, may be a less ‘evolved’ peralkaline granite complex, such as the MRC.
3.2 Outlook and Future Work

While the results of this study have shed light on the magmatic to hydrothermal evolution of the MRC and processes which have contributed to REE and rare element enrichment in the complex, additional work is needed to elucidate the complex petrogenesis of the MRC. The magmatic relationship of MRC rocks to the surrounding Pikes Peak Batholith is still somewhat unclear and additional petrography coupled with major and trace-element analysis of select minerals such as zircon and possibly Fe-Ti oxides may help elucidate the relationship between peraluminous and peralkaline rocks in the MRC. LA-ICP-MS trace element analysis on zircon, following characterization of zircon zonation and paragenesis using optical light microscopy, BSE imaging and cathodoluminescence (CL), could be a powerful tool for exploring the petrogenesis of the MRC, as zircon occurs throughout the MRC, spans multiple generations and populations, and is a recorder or both magmatic and hydrothermal processes. New tools for in-situ analysis of zircon via secondary ionization mass spectroscopy (SIMS) allow for trace element analysis and U/Pb dating of zircon whose paragenesis has been well-constrained by petrography. This ‘melt fingerprinting’ by zircon could potentially quantify the transition from a peraluminous to peralkaline melt in ways this study has been unable to do.

Previous research on Pikes Peak Batholith petrogenesis has focused on Nd/Sm and Lu/Hf isotope work (e.g.; DePaolo 1980, Smith et al. 1999, Guitreau et al. 2016), but as of now there is no isotopic data on rocks of the MRC specifically. Such data could potentially clarify whether the Mount Rosa granite and related rocks evolved directly from a fractionated mantle-derived melt or whether they are the product of extended fractionation of the larger peraluminous melt which generated the bulk of the Pikes Peak Batholith. The relationship of mafic dikes and subsolvus biotite granite at the MRC is also somewhat ambiguous based on field work, petrography, and EPMA chemistry of biotite in this study. Isotopic work on these units could clarify their connection to other MRC rocks and support or dispute the hypothesis that the mafic dikes may have formed from a different melt. While some unpublished USGS dates of MRC rocks via microprobe U/Pb and TIMS geochronology exist (Wayne Premo pers. comm.), the analyzed rocks were not well-constrained in terms of either their field location or petrogenesis within a particular unit. Therefore, while issues associated with past geochronology work on
MRC zircon (metamict grains, high Th: Pb ratio) will likely persist, new age data on MRC rocks from well-characterized samples is essential in understanding the petrogenesis of the complex.

On a smaller and more localized scale, the petrogenesis of the pegmatites, particularly Type-(II) pegmatites, remains somewhat ambiguous. While evidence from this study suggests that they are related to the emplacement of the Mount Rosa granite and have been effected by the same late F-rich hydrothermal fluids which have affected the Mount Rosa granite, it is unclear what the relative role of autometasomatism (e.g., fluids internally-derived from the pegmatites) versus externally-derived fluids is on the pegmatite mineralogy. Fluid inclusion petrography followed by thermometry and possibly in-situ LA-ICP-MS analysis of fluid inclusion chemistry could elucidate this, as has been done on a limited scale for South Platte District pegmatites (e.g. Gagnon et al. 2004, ongoing study on the Wellington Lake Pegmatite by Raschke, Stern and Persson). Analysis of the chemistry and thermometry of the presumably F-rich fluid trapped in fluid inclusions in MRC pegmatites compared to fluid inclusions from the Mount Rosa granite could provide valuable insight into the mobilization of REE and other HFSE in the MRC, and could be compared to a fairly robust body of data for both Pikes Peak Batholith pegmatites (e.g. Gagnon et al. 2004), and other peralkaline granite complexes (e.g. Strange Lake Canada; Salvi and Williams-Jones 1996).
REFERENCES


Simmons, W., and Heinrich, E. (1980). Rare-Earth Pegmatites of the South Platte District, Colorado: CGS Resource Series 11, 131. P.


## APPENDIX A

### TIMA AUTOMATED MINERALOGY DATA

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## APPENDIX B

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<td><strong>Sc (PPM)</strong></td>
<td>6</td>
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<td>14</td>
<td>2</td>
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</tr>
<tr>
<td><strong>Be (PPM)</strong></td>
<td>6</td>
<td>3</td>
<td>43</td>
<td>5</td>
<td>7</td>
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<tr>
<td><strong>Zn (PPM)</strong></td>
<td>110</td>
<td>280</td>
<td>280</td>
<td>40</td>
<td>120</td>
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<tr>
<td><strong>Ga (PPM)</strong></td>
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<tr>
<td><strong>Ge (PPM)</strong></td>
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<tr>
<td><strong>Mo (PPM)</strong></td>
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<td>4</td>
<td>b.d.l</td>
<td>b.d.l</td>
<td>4</td>
</tr>
<tr>
<td><strong>Sn (PPM)</strong></td>
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<td>3</td>
<td>37</td>
<td>2</td>
<td>5</td>
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<tr>
<td><strong>Cs (PPM)</strong></td>
<td>1.3</td>
<td>1.5</td>
<td>1.9</td>
<td>1.3</td>
<td>2.4</td>
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<tr>
<td><strong>Tm (PPM)</strong></td>
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<td>2.17</td>
<td>2.36</td>
<td>1.03</td>
<td>2.07</td>
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<tr>
<td><strong>W (PPM)</strong></td>
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<td>1</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td><strong>Pb (PPM)</strong></td>
<td>32</td>
<td>16</td>
<td>37</td>
<td>33</td>
<td>33</td>
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<tr>
<td><strong>U (PPM)</strong></td>
<td>6.4</td>
<td>2.5</td>
<td>3.3</td>
<td>13.2</td>
<td>7.6</td>
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b.d.l = below detection limit.