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**Abstract**

The recently discovered Money Knob Au deposit represents a bulk tonnage deposit with a measured, indicated, and inferred resource of 20.6 million ounces of gold at an average grade of 0.52 g/t and a cut-off grade of 0.22 g/t. The deposit is located in the central Alaskan portion of the Tintina Gold Province. The low-grade, high tonnage resource of the Money Knob deposit is hosted in a tectonically complex package of Devonian sedimentary and volcanic rocks that have been overthrust by Cambrian ophiolitic rocks. Gold mineralization formed in close association with a mid-Cretaceous felsic dike complex that intruded into this package of sedimentary and volcanic rocks and is thought to be part of the Tombstone Plutonic Suite. A major phase of the felsic dike complex is a biotite syenite, which occurs late in the deposit development and has a U/Pb zircon age of 92.04±0.14 Ma. The dikes account for about <10 percent of the total rock volume in the deposit, with individual dikes ranging from several centimeters to over 10 meters in width. The biotite syenite crosscuts preexisting ore zones and associated altered rocks, but has clearly also been affected by strong hydrothermal alteration and can be mineralized (0.25 to 1.0 g/t Au). The observed field relationships indicate that deposit formation at Money Knob occurred in a tectonically complex environment characterized by tectonic relaxation along reactivated thrust faults.

Utilizing cathodoluminescence microscopy coupled with fluid inclusion petrography and microthermometry, four different quartz types were recognized in the quartz veins that formed at progressively decreasing temperatures. Homogenization temperatures gathered from secondary fluid inclusion assemblages suggest that early quartz formed at temperatures of at least 305-310°C from hydrothermal fluids that contained a high amount of CO₂ while the next quartz type formed at temperatures below 200°C at the same pressure conditions. This low-temperature quartz and subsequent quartz types formed at even lower temperatures also contain significant amounts of CO₂. It is estimated that the vein quartz formed at a minimum depth of 3 kilometers. Lithostatic pressure conditions were maintained throughout the formation of the different quartz
types. The gold in the quartz veinlets occurs as fracture-controlled free gold within the earliest
type of quartz. Gold formation clearly post-dated the formation of the high-temperature quartz.
Outside the veins, much of the gold occurs as fine-grained native gold disseminations within the
intensely altered host rock.

Optical microscopy, cathodoluminescence microscopy, and scanning electron microscopy
were used to identify the different alteration mineral associations related to quartz veining.
Hydrothermal alteration occurred in a generally retrograde environment. The earliest alteration
product observed is hydrothermal biotite. Formation of biotite alteration was followed by
sericite alteration and then albite alteration. Ankerite and lesser dolomite overprint biotite and
sericite alteration and can overprint or accompany albite alteration. At lower temperatures,
these alteration mineral association were overprinted by clay alteration, involving the formation
of smectite and kaolinite as well as ankerite and lesser dolomite. Geochemical data gathered
from the different alteration mineral associations suggest that biotite and sericite alteration
were accompanied by slight enrichment in potassium while albite alteration was typified in a
pronounced enrichment of sodium and silica. Albite alteration appears to have coincided with
gold enrichment while igneous rocks affected by the low-temperature clay alteration have the
lowest gold grades.

Although the biotite syenite at Money Knob does not represent the direct source of the
mineralizing hydrothermal fluids, the close spatial and temporal relationship between igneous
activity resulting in the formation of the dike complex and mineralization suggests that Money
Knob can be classified as an intrusion-related gold deposit.
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## List of Abbreviations

### Mineral Abbreviations:
- Ab: Albite
- Ank: Ankerite
- Ap: Apatite
- Apy: Arsenopyrite
- Bt: Biotite
- Cal: Calcite
- Chl: Chlorite
- Dol: Dolomite
- Ilt: Illite
- Ilm: Ilmenite
- Kln: Kaolinite
- Kfs: K-feldspar
- Mag: Magnetite
- Ms: Muscovite
- Pl: Plagioclase
- Py: Pyrite
- Qz: Quartz
- Sa: Sanidine
- Sme: Smectite
- Spn: Sphene
- Zrn: Zircon

### Other Abbreviations:
- BSE: Back scatter-electron
- CL: Cathodoluminescence
- CPL: Crossed polarized light
- EDX: Energy-dispersive X-ray
- ICP-MS: Induced coupled plasma-mass spectrometry
- PPL: Plane polarized light
- RL: Reflected light
- SEM: Scanning electron microscope
- ID-TIMS: Isotope dilution-thermal ionization mass spectrometry
- XRD: X-ray diffraction
- XRF: X-ray fluorescence
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CHAPTER ONE:
INTRODUCTION

1.1 Money Knob Deposit

The Money Knob gold deposit, discovered in 2003, is located in the central Alaskan portion of the Tintina gold province (Pontius et al., 2010). Currently, the measured, indicated, and inferred resource is 20.6 million ounces of gold putting it among the world’s largest undeveloped gold deposits discovered in the last 20 years. The deposit is hosted in a tectonically complex package of Devonian sedimentary and volcanic rocks overthrust by Cambrian ophiolitic rocks, all of which are cut by a mid-Cretaceous dike swarm. Individual intrusive bodies range from 2 centimeters to 10 meters in width and cross-cut preexisting ore zones and associated altered rocks. However, they clearly also have been affected by hydrothermal alteration and can be mineralized (0.25 to 1.0 g/t Au). Observed cross-cutting relationships suggest that dike emplacement occurred during a prolonged igneous event that was broadly contemporaneous with hydrothermal alteration and gold mineralization.

The deposit characteristics suggest that Money Knob is similar to other distally disseminated intrusion-related gold deposits found in the Tintina gold province. Intrusion-related gold deposits are generally related to Phanerozoic, moderately reduced, I-type, felsic intrusions emplaced inboard of convergent plate margins with low sulfide ore assemblages (Baker, 2002). The most important features that define this deposit type are (1) association with metaluminous subalkalic intrusions of intermediate to felsic composition that lie near the boundary between the ilmenite – magnetite series, (2) gold deposition by carbonic hydrothermal fluids, (3) metal assemblages that variably combine Au with elevated Bi, W, As, Mo, Te, and Sb levels, but low base metal concentrations, (4) low sulfide mineral contents (< 5 vol%), typically with arsenopyrite, pyrrhotite, and pyrite being the principal sulfide minerals, and little to no magnetite or hematite,
(5) areally restricted and commonly weak hydrothermal alteration except with deposits formed
at shallow depths, (6) a tectonic setting well inward from inferred or recognized convergent plate
margin, and (7) a location in magmatic provinces otherwise best known for tungsten and/or tin
deposits (Lang and Baker, 2001).

Early work on the Money Knob gold deposit suggests that albite and sericite are the dominant
hydrothermal alteration minerals. Albite alteration is most strongly developed in the volcanic
host units, while strong sericite alteration is found throughout the Devonian and overlying
Cambrian ophiolite rocks. The intrusive rocks can contain both strong sericite and/or albite
alteration. These alteration styles are associated with the highest gold grades (Pontius et al.,
2010). Alteration occurs both as vein selvages and pervasive fronts through the rock. Bulk
rock and trace element geochemical analysis on samples consisting of 30 - >100 cm long core
intervals have been used to classify alteration styles, but the sampling strategy did not allow
detailed geochemical patterns to be recognized.

1.2 Research Aims

The present study focuses on the most widespread intrusive phase of the dike swarm at
Money Knob, a biotite syenite. Changes in mineralogy, texture, and chemistry of the biotite
syenite were examined to reconstruct the processes of hydrothermal alteration that occurred in
association with the mineralizing event.

Unraveling the relative timing and nature of hydrothermal alteration processes at Money
Knob aids in the identification of physicochemical controls on mineralization. Recognition of
these controls may offer a more detailed understanding of the genetic link between emplacement
of the biotite syenite dikes, hydrothermal alteration, and mineralization. Specific questions that
were addressed in this project include:

1. What are the mineralogical and textural changes associate with the different alteration
styles?

2. What are the chemical changes associated with the different alteration styles?
3. Are there multiple overprinting alteration styles and, if so, what is the relative timing of different alteration processes?
4. How are different alteration styles related to gold mineralization?
5. What is the absolute age of the hydrothermally altered biotite syenite?

1.3 Methods of Study and Thesis Outline

Fieldwork at and around the Money Knob deposit was conducted in May–October 2009 and June–August 2010. Chapters 2 and 3 detail the regional and deposit geology determined through examination of outcrop, drill samples, and an extensive literature review. Much of the fieldwork focused on examination of exploration drill core. In accordance with the goals of this thesis, emphasis was placed on the study of the variably altered biotite syenite. To constrain its emplacement age, the U-Pb ages of zircons were determined by ID-TIMS. The results of the age dating are summarized in Chapter 4.

Selected vein samples were studied by optical microscopy to determine the mineralogy of the ore-bearing veins and the relationships between the ore minerals and the gangue phases (Chapter 5). A combination of fluid inclusion petrography and CL microscopy has been employed to distinguish different types of quartz within the veins and their relationships with the ore minerals. Microthermometric investigations on the fluid inclusions have been performed to constrain the physicochemical conditions under which the different types of quartz must have formed.

Petrographic work on variably altered biotite syenite was conducted to understand mineralogical and textural changes associated with the quartz vein formation and related hydrothermal alteration. Microtextural relationships were examined by SEM. Semi-quantitative EDX spectrometry was used to constrain the composition of key minerals, especially those that were too fine-grained to be identified by optical microscopy. BSE imaging was used to examine small-scale replacement textures, in conjunction with optical microscopy and scanning electron microscopy, CL microscopy was utilized to study replacement textures affecting minerals that show luminescence. The results of the alteration study are summarized in Chapter 6.
Chapter 7 links quantitative chemical changes to observed textural and mineralogical changes. Major element abundances were determined by whole-rock XRF analysis while ICP-MS was employed for trace element analysis. In addition, fire assay for Au, total carbon and sulfur analysis, as well as loss on ignition analysis was conducted on the variably altered samples. A density measurement was also taken to determine alteration-induced density changes.

The results of the various analytical methods are discussed and summarized in Chapters 8 and 9, respectively. In Chapter 8 conclusions about the formation environment of the deposit and its development through time are discussed, including the spatial and temporal relationship to the biotite syenite, and its possible genetic connection to the formation of the deposit. Comparisons between Money Knob and similar deposits types found throughout the Tintina gold province (Hart et al., 2002) were also made to better understand how Money Knob fits into the broader metallogenic province. Finally, in Chapter 8 exploration implications are addressed based on the data gathered during this project. Chapter 9 presents a summary of the research results and provides recommendations for future work.
2.1 Geology of the Livengood Area

The Money Knob gold deposit is located in the Livengood area of east-central Alaska (Fig. 2.1). The geology of the Livengood area consists of a series of lithotectonic terranes that have accreted to the continental margin of the North American craton during the Jurassic to early Tertiary (Jones et al., 1983; Plafker and Berg, 1994; Silberling et al., 1994; Monger and Nokleberg, 1996).

The Livengood area is bounded to the northeast by the Tintina strike-slip fault. The Tozitna and Victoria Creek-Kaltag strike-slip faults limit it to the northwest. In the southwest, the area is covered by Cenozoic poorly consolidated sedimentary rocks and unconsolidated alluvial sediments. The Livengood area abuts the Yukon-Tanana terrane to the southeast (Churkin et al., 1982; Silberling and Jones, 1984; Weber et al., 1985, 1992; Dover, 1994; Athey et al., 2004a).

The central part of the Livengood area is made up of the Schwatka, Livengood, Manley, White Mountain, and Wikersham terranes. Rocks belonging to the Preacher, Crazy Mountain, Minook, and Baldry terranes occur in the periphery of the Livengood area (Fig. 2.1 and 2.2).

2.1.1 Schwatka Terrane

The Schwatka terrane is the northernmost terrane in the Livengood area and is bound to the north by the Tozitna fault and to the south by the Victory Creek – Kaltag strike-slip fault, both of which are splays off the Tintina fault (Dover, 1994).

The basement of the Schwatka terrane is formed by the Precambrian Wickersham assemblage that is primarily composed of grit and quartzite with dark grey arenaceous limestone units and maroon and green argillite units (Weber et al., 1992; Dover, 1994). The Wickersham assemblage
in the Schwatka terrane is indistinguishable from the Wickersham assemblage of the Wickersham terrane in the southern part of the Livengood area. The Wickersham assemblage is overlain by clastic, volcanic, and carbonate rocks (Dover, 1994).

2.1.2 Livengood Terrane

The Livengood terrane in the central portion of the Livengood area is composed of two sub-terranes; the northwestern Livengood Dome sub-terrane and the southeastern Amy Creek sub-terrane. To the north the Livengood Dome sub-terrane is separated from the Schwatka terrane by the major Victoria Creek – Kaltag strike slip fault. To the south, a shallow to moderate south dipping thrust fault separates the Livengood Dome sub-terrane from the Amy Creek sub-terrane (Weber et al., 1992, 1997; Athey and Craw, 2004). The southern contact between the Amy Creek sub-terrane and the Manley terrane is believed to be a south dipping thrust contact, however, the angle of the contact is unknown.

The Livengood Dome sub-terrane contains the Ordovician age Livengood Dome Formation and the Silurian Lost Creek Formation. The Livengood Dome Formation is a mix of chert and other sedimentary rocks dated between middle to late Ordovician via biostratigraphy (Weber et al., 1994). The Lost Creek Formation is composed of limestone, sandstone, siltstone, shale, and conglomerate. The Lost Creek Formation is laterally discontinuous and unconformably overlies the Livengood Dome Formation. Fauna contained in limestone suggests a carbonate platform that subsequently shed debris downslope to be deposited on the Livengood Dome Chert (Athey and Craw, 2004). Both formations contain at least two distinct fold sets, one sub-parallel to regional strike-slip faulting (Weber et al., 1992, Athey and Craw, 2004).

The Amy Creek sub-terrane contains three distinct assemblages separated by low angle thrust contacts (Athey and Craw, 2004). The Amy Creek assemblage is the structurally lowest group of rocks in the sub-terrane. It is composed of siliceous mudstone and chert, with minor interlayered conglomerate and metabasalt. Fossil, radiometric, and interpretive stratigraphic ages have yielded poor age constrains for the Amy Creek assemblage, presently putting the assemblage somewhere between latest Proterozoic to early Devonian (Athey and Craw, 2004). The units
Figure 2.1: Map showing the location of the Livengood area in Alaska and a generalized terrane map of the Livengood area. Major faults and fault zones are labelled. Colored terranes are detailed in the text, grey terranes are bordering terranes and not discussed in the text.
within the Amy Creek assemblage are broadly folded with fold axes subparallel to the regional strike-slip faulting. Tighter and occasionally overturned folds are found closer to the thrust contacts with interpreted fold axial planes dipping to the south (Athey and Craw, 2004). An obducted Early Cambrian ophiolite sequence overlies the Amy Creek assemblage. The ophiolite sequence consists of a thrust stacked sequence of greenstone and metagabbro, Amy Creek sedimentary rocks, and thin interleaved serpentinite. \(^{40}\text{Ar}^{39}\text{Ar}\) dating of a hornblende sample from a metagabbro within the ophiolite sequence yielded an age of 535.3 ± 2.7 Ma (Athey et al., 2004b)

The last assemblage in the Amy Creek sub-terrane is the Cascaden Ridge Formation, which has been overthrust by a sliver of ophiolitic rocks. It is composed of Devonian sedimentary and volcanic flow and clastic rocks that have recumbent folds with south-southeast dipping axial planes. The Cascaden Ridge Formation is the host for the Money Knob gold deposit. The contact between the ophiolite sequence and Cascaden Ridge was originally interpreted as unconformable (Weber et al., 1992), but evidence around the Money Knob area suggests a low-angle south dipping thrust contact (Athey and Craw, 2004).

2.1.3 Manley Terrane

The Manley terrane is located in the central portion of the Livengood area. It is 400 kilometers long and as narrow as 10-20 kilometers, suggesting significant compression compared to the original basin size. The terrane is affected by extensive thrust faulting and local isoclinal folding (Weber et al., 1985, 1992; Athey and Craws, 2004). A moderate- to steep-angle south dipping thrust contact, referred to as the Beaver Creek fault, separates the Manley terrane from the White Mountain terrane in the southeast (Cady and Morin, 1988; Long and Miyaoka, 1988; Dover, 1994).

The Manley terrane consists of three main flysch units. The Jurassic to Early Cretaceous Vrain unit occurs in the eastern part of the terrane while the late Jurassic to early Cretaceous Wilbert Creek unit dominates in the central and western part of the terrane. The Middle Devonian Cascaden Ridge unit occurs in the central part. The Vrain unit comprises carbonaceous, fine-
Figure 2.2: Generalized geologic map of the Livengood area. The Cascadian Ridge Formation and the Cambrian Ophiolite host rocks for the Money Knob deposit are highlighted. Major faults are labeled along with the Money Knob deposit (modified after Weber et al., 1997).
grained to gritty clastic rocks, tuffs, and conglomerates that consist mostly of chert and quartzite. Chert and basalt are minor components. Based on lithological similarities, it appears possible that the Vrain unit and the Cascaden unit are correlative. The Wilbert Creek unit probably unconformably overlies the Vrain and Cascaden Ridge units, although its basal contact is poorly constrained. This unit is primarily composed of shale, greywacke, quartzite, and polymict conglomerate (Weber et al., 1992; Dover, 1994). In addition to the three flysch units, lesser amounts of Triassic black shale, chert, limestone, and calcareous phosphatic shale exists within the Manley terrane (Weber et al., 1992; Dover, 1994).

2.1.4 White Mountains Terrane

The White Mountain terrane is located between the Manley and Wickersham terranes in the central portion of the Livengood area. It is bound to the northwest by the steep- to moderate-angle Beaver Creek thrust that records ~10 kilometers of movement (Long and Miyaoka, 1988; Dover, 1994).

The base of the White Mountain terrane is formed by the Globe Formation. That formation consists of quartzite that is characterized by vitreous quartz. It contains sheared argillite interbeds and abundant hornblende-bearing quartz diorite and quartz gabbro sills. The remainder of the White Mountains terrane is composed of the Ordovician Fossil Creek Formation volcanic and intercalated sedimentary rocks as well as the Silurian Tolovana Limestone. Both units are separated from the Globe Formation by a thrust.

The Fossil Creek Formation consists of alkali basalt, agglomerate, volcaniclastic conglomerate, and minor limestone. It has been demonstrated that volcanic rocks of the Fossil Creek Formation were deposited on the upper Wickersham assemblage during the Ordovician. Agglomeratic volcanic rocks of the Fossil Creek Formation contain pebbles and cobbles that resemble rocks of the upper Wickersham assemblage (Dover, 1994). The Fossil Creek Formation volcanic rocks are unconformably overlain by the Tolovana Limestone. These limestones reach a stratigraphic thickness of up to 1,200 meters. They can be divided into a lower Silurian unit and an upper middle Devonian unit.
2.1.5 Wickersham Terrane

The Wickersham terrane is the structurally highest terrane, located in the southeastern portion of the Livengood area. It is composed of the Wickersham assemblage that includes grit, interlayered argillite, limestone, greywacke, quartzite, green to purple slate and slaty argillite, and phyllite units. It is believed to be Neoproterozoic to early Cambrian in age (Weber et al., 1992; Dover, 1994). The Wickersham assemblage is overthrust from the southeast by greenschist facies metamorphic rocks of the Yukon-Tanana terrane. The overlying Yukon-Tanana terrane rocks are believed to be the metamorphic equivalents of the Wickersham assemblage rocks (Weber et al., 1985).

2.1.6 Crazy Mountain Terrane

The Crazy Mountain terrane is located in the northeastern edge of the Livengood area and bound to the south by the Tintina fault and to the north by a north dipping thrust fault (Dover, 1994). The western boundary is the Tozitna strike-slip fault, which is a splay off the Tintina fault. The Crazy Mountain terrane is composed of three units. The basement of the Crazy Mountain terrane is formed by the upper section of the Wickersham assemblage, which is composed of maroon and green slate and slaty argillite, grit, and black limestone. Structurally (?) overlying the Wickersham assemblage are fossiliferous Lower Devonian limestone and mafic igneous rocks, possibly analogous to the rocks found in the Schwatka terrane. A chert-pebble conglomerate forms the top unit in the Crazy Mountain terrane and is indistinguishable from middle and upper Paleozoic conglomerates found in the White Mountain, Manley, and Livengood terranes (Dover, 1994).

2.1.7 Preacher Terrane

The Preacher terrane is located within the Tintina strike-slip fault zone in the far east of the Livengood area. It is mainly composed of highly deformed rocks that may be correlatives to those found in the Livengood terrane south of the Victoria Creek-Kaltag fault, including rocks of the upper part of the Wickersham assemblage and possibly the Amy Creek assemblage (Dover,
2.1.8 Minook Terrane

The Minook terrane is located in the southwestern portion of the Livengood area. It is separated from the adjacent Manley terrane by a north dipping thrust contacts. The contact between the Minook terrane and the Baldry terrane to the north is also marked by a thrust contact. The Minook terrane is dissected into an eastern and western portion by a northwest trending splay off the Victoria Creek-Kaltag fault (Dover, 1994). It is primarily composed of rocks that may be correlative with the lower units of the Manley terrane, including Pennsylvanian to Permian slate, quartzite, chert, argillite, siltstone, sandstone, and minor conglomerate and greenstone (Weber et al., 1992).

2.1.9 Baldry Terrane

The Baldry terrane is bound to the northwest by the Victoria Creek-Kaltag fault and to the southeast by a north-dipping thrust fault. A northwest trending splay off the Victoria Creek-Kaltag fault cuts across the terrane. The Baldry terrane is a highly deformed and structurally complex terrane that is composed of chert, marble, greenschist, and mica schist (Dover, 1994). The low-grade metamorphic rocks of the Baldry terrane are tightly folded, thrusted, and sheared under low-grade ductile conditions. Protolith of appropriate compositions that may be correlative can be found in the White Mountain, Livengood, and Schwatka terranes. It is possible that the metamorphic recrystallization and ductile deformation in the Baldry terrane represent the deeper-crustal manifestations of the more brittle imbrication affecting the potential protolith rocks in the adjacent terranes (Dover, 1994).

2.2 Intrusive Rocks of the Livengood District

The Livengood district is host to numerous mid-Cretaceous to early-Tertiary igneous intrusions. The oldest dated intrusions are mid-Cretaceous and have ages of approximately 92 to
Figure 2.3: Generalized stratigraphic columns for the Livengood area (modified after Dover, 1994).
87 Ma (Chapman et al., 1971, 1982; Reifenstuhl et al., 1997; Hart et al., 2004). These magnetite-series intrusions (Hart et al., 2004) include the Roughtop and Sawtooth Mountain plutons in the Manley terrane and the small syenite intrusion in the Wickersham terrane. All other intrusions in the Livengood area have been dated at 65 to 55 Ma (Chapman et al., 1971, 1982).

2.3 Structural Geology of the Livengood Area

The individual terranes in the Livengood area are bound by either thrust faults or strike-slip faults. The structural segmentation and lithological complexity of the disrupted stratigraphy make correlation between individual terranes difficult (Dover, 1994).

Many of the terrane boundaries in the Livengood district are recognized as thrust contacts (Cady and Morin, 1988; Long and Miyaoka, 1988; Dover, 1994; Athey and Craw, 2004). The thrust contacts typically are narrow and well defined. There is a common association with large, occasionally overturned, folds in the Wickersham, White Mountain, Manely, and Livengood terranes. The axial planes of these large folds dominantly dip to the southeast. Older-on-younger stratigraphic relationships and stratigraphic truncations occur from southeast to northwest, suggesting a northwest direction of transport (Dover, 1994).

The largest thrust in the Livengood district is the contact between the White Mountain terrane and the Manley terranes, which is referred to as the Beaver Creek thrust (Dover, 1994). Geophysical resistivity data suggests that the Beaver Creek thrust records approximately 10 kilometers of movement from southeast to northwest (Long and Miyaoka, 1988). Similar northwest-directed thrusts imbricate much of the stratigraphy in the Wickersham, White Mountain, and Livengood terranes (Dover, 1994).

In the western part of the Livengood area, thrust traces wrap around a large-scale, west-plunging anticline or anticlinal duplex. In this area, fold axial planes dip to the northwest and older-on-younger stratigraphic relationships occur from the northwest to the southeast. This mappable oroclinal bending pattern in the Livengood fold-and-thrust trend is similar to the oroclinal bend in the Charley River area ~450 kilometers to the east-southeast along the Tintina
fault system. Based on the similarity in structural pattern, the fold-and-thrust sequence in the Livengood area is believed to be a portion of the Charley River fold-and-thrust sequence that was offset by approximately 400 to 450 kilometers of strike-slip motion along the Tintina fault system (Tempelman-Kluit, 1979; Dover 1985, 1994; Murphy and Mortensen, 2003).

The Livengood area is transected by the Victory Creek and Tozitna faults, two major dextral strike-slip faults that are splays of the Tintina fault system (Wheeler and Weber, 1988; Dover, 1994; Athey and Craw, 2004; Till et al., 2007). The Victory Creek fault connects to the Tintina fault system in the east with the Kaltag fault system in the west (Dover, 1994; Athey and Craw, 2004). It separates the Schwatka terrane from the majority of the other terranes occurring in the Livengood area. Where exposed, the Victory Creek fault is manifested by an approximately 1.5 kilometer wide zone of disconnected tectonic blocks with intense and pervasively sheared, mylonitic matrixes. The Victory Creek fault is interpreted as a steeply dipping fault that developed after the folding and thrusting, as is evident by its low-angle truncations of fold-and-thrust trends (Dover, 1994).

The Tozitna fault represents the northern boundary of the eastern part of the Livengood area. The fault is not exposed, but physiographic expressions indicate that it connects to the Tintina fault system in the east. The Tozitna fault is thought to record an offset of approximately 55 kilometers (Dover, 1994).

The geometrical relationships indicate that thrusting in the Livengood area predated strike-slip faulting along Victory Creek and Tozitna faults. Thrusting in the Livengood area involved rocks that are as young as early Cretaceous. The principal movement along the Tintina fault system appears to predate deposition of poorly dated Tertiary rocks (Dover, 1994).
Chapter Three: Deposit Geology

3.1 Introduction

The present chapter summarizes the geology of the Money Knob deposit as described in previous publications and internal company reports. The present understanding of the deposit geology is based on outcrop examination and targeted mapping of exposed road cuts and trenches as well as extensive diamond and RC drilling. The availability of oriented drill core permitted orientational measurements on contacts, veins, fractures, and joint sets to unravel relationships between lithologies and structures. Previous petrographic examination (Klipfel, 2009) helped establishing the relative timing and mineralogy of the different alteration styles recorded at Money Knob.

3.2 Deposit Stratigraphy

The Money Knob gold deposit is hosted within the Cascaden Ridge Formation and overthrust ultramafic and mafic rocks (Figs. 3.1 and 3.2). The Cascaden Ridge Formation is made up of three distinct units that have conformable contacts, referred to as the lower sedimentary rocks, the middle volcanic rocks, and the upper sedimentary rocks. Fossil evidence from the lower and upper sedimentary units suggests an early Middle Devonian age for the Cascaden Ridge Formation (Blodgett, 1992; Athey and Craw, 2004).

The lower sedimentary unit of the Cascaden Ridge Formation comprises carbonaceous mudstone and siltstone, and fine sandstone (Athey and Craw, 2004). The grain size of the sedimentary rocks shows a lateral change from fine-grained mudstone in the north to sandy siltstone and fine sandstone in the south (Brechtel et al., 2011). Mudstone is black to grey,
massive, and shows an upward transition into more fissile shale containing diagenetic pyrite. Minor grey siltstone beds are present and locally soft sediment deformation and slump folding textures are preserved. Siltstone and sandstone in the lower sedimentary unit are moderately to strongly carbonaceous and fine-grained. Clusters of crinoids up to 20 centimeters wide can be observed (Athey and Craw, 2004; Brechtel et al., 2011). Numerous intervals of upward-fining sandstone-siltstone-argillite are visible in core. The true thickness of this unit is not known because the lower contact of the lower sedimentary unit has not been intercepted in drill core, nor is it outcropping anywhere in the Livengood area. The known thickness of the unit exceeds 200 meters. The upper contact between the lower sedimentary unit and middle volcanic unit is a conformable contact as suggested by extensive drilling in the deposit area. However, the units have been offset laterally by shallow and/or steep thrust and normal faulting which juxtapose the two units (Klipfel et al., 2009; Brechtel et al., 2011).

The middle volcanic unit is composed of coherent intermediate to felsic lavas which locally have flow banding (Klipfel et al., 2009). Coherent volcanic rocks are aphanitic to porphyritic and grey to grey-brown. Locally, amygdales filled by secondary quartz, albite, and/or carbonate minerals are present (Klipfel et al., 2009). The coherent volcanic rocks are overlain by volcaniclastic rocks, mostly brown to grey massive, intensely altered tuffs and matrix- to clast-supported breccia with lithic and volcanic fragments (Klipfel et al., 2009). The upper contact between the middle volcanic unit and the overlying upper sedimentary unit is conformable. The thickness of the volcanic unit is generally between 50 and 75 meters. Due to later offset by shallow and/or steep thrust and normal faulting, the two units are juxtaposed (Klipfel et al., 2009; Brechtel et al., 2011).

The upper sedimentary unit consists mostly of sandstone and siltstone with lesser conglomerate, and rare limestone and dolostone. Fine- to medium-grained silty sandstone and sandy siltstone and lesser coarse-grained pebbly sandstone are brown, brown-grey, yellow-grey, and grey. The sandstones are composed of subangular to round quartz, feldspar, chert, and lithic fragments, some of which have a calcareous component in the matrix. They are dominantly clast-
Figure 3.1: Generalized log of diamond drill hole MK-09-34. The log shows the distribution of lithological units, alteration facies, and gold grade.
Figure 3.2: Geology of the Money Knob area. (A) Plan map of the geology of the Money Knob deposit area (modified from Brechtel et al., 2011). The AA’ line indicates the location of the cross-section shown below. (B) Geologic cross section of the Money Knob deposit (modified after Myers et al., 2011). The Money Knob chert is incorporated with the Cambrian ophiolite unit in the cross-section, and Cretaceous intrusive rocks are shown in orange.
supported with local matrix-supported siltstone and mudstone (Klipfel et al., 2009; Brechtel et al., 2011). Crinoids are visible in the mudstone and have been observed in clusters up to ~10 centimeters wide. The upper sedimentary unit has a known thickness of greater than 150 meters. However, as the top contact of the unit is a structural contact, the true thickness of the upper sedimentary unit is unknown.

Overlying the upper sedimentary unit are overthrust Cambrian ophiolitic rocks (Plafker and Berg, 1994; Athey and Craw, 2004). The contact between the Devonian Cascaden Ridge Formation and the structurally overlying ophiolitic rocks is a sub-horizontal to shallowly south dipping thrust contact that is usually marked by strongly-altered serpentinite. The ophiolitic rocks are dominated by serpentinite with lesser metagabbro and rare greenstone. Locally, siliceous mudstone, conglomerate, sandstone, and chert are tectonically interleaved with the ultramafic and mafic rocks. The cherty unit was initially interpreted as intercalated parts of the Amy Creek Assemblage, but has subsequently been separated into its own unit referred to as the Money Knob Chert (Athey and Craw, 2004; Brechtel et al., 2011).

Serpentinite in the ophiolite unit is waxy and ranges in color from green, light green, to black, and is fine-grained and massive. Carbonate-altered serpentinite can be orange, white, cream, brown, pale gray-brown, and pale green (Athey and Craw, 2004; Klipfel et al., 2009; Brechtel et al., 2011). Although mostly massive, the serpentinite can be highly sheared and fissile in zones affected by intense shearing. Metagabbro in the ophiolite unit is black, dark gray, gray, green, and white. It is medium-grained, equigranular, and massive with recognizable plagioclase, hornblende, and pyroxene (Klipfel et al., 2009; Brechtel et al., 2011). Metagabbro is more competent and less altered then the surrounding serpentinite and displays well-defined contacts in drill core. Greenstone is dark green, green-grey, and black. The greenstone is mostly fine-grained and massive, but rare clast-supported breccia with angular clasts of gabbro, volcaniclastic material, and serpentinite have been recognized (Athey and Craw, 2004).

A minimum age of 535.3 ± 2.7 Ma for the ophiolite unit of the Amy Creek Assemblage has been established through ⁴⁰Ar/³⁹Ar dating of hornblende contained in the gabbro (Athey et al.,
3.3 Intrusive Rocks

Igneous rocks forming a dike swarm cross-cutting all other lithological units are the youngest rocks found at the Money Knob deposit (Fig. 3.2). Hundreds of individual dikes have been observed. They have variable thicknesses, ranging from less than 0.1 meter to over 10 meters. Dikes include feldspar porphyry dikes which have an aphanitic groundmass, biotite porphyry dikes, which can have a phaneritic or aphanitic groundmass, and biotite-feldspar porphyry dikes. Dike compositions are moderately alkalic and vary widely from felsic to mafic (Myers et al., 2011).

Many of the dikes forming the Cretaceous dike swarm at Money Knob appear to have been emplaced along preexisting faults. Structural measurements on oriented core revealed the presence of two structurally distinct dike sets. One dike set has a general east-west strike and a shallow to moderate dip to the south. These dikes include nearly all the feldspar porphyry dikes and some biotite porphyry dikes. A second dike set has a northwest strike and dips steeply to the southwest, and these dikes are biotite porphyry and biotite-feldspar porphyry dikes (Myers et al., 2011). Based on cross-cutting relationships, it appears that the east-west striking dikes were emplaced prior to the northwest striking dikes (Myers, 2009).

3.4 Structure

The structural development of the deposit area began with the formation of east-west striking recumbant fold and thrust structures related to the Lillian Fault. The Lillian fault strikes west-northwest and dips steeply to the south (Brechtel et al., 2011; Myers et al., 2011). The surface exposure of the Lillian fault is defined by a steep valley where sheared and juxtaposed units of the Cascaden Ridge Formation and the Cretaceous dike swarm are exposed (Fig. 3.2).

Following the development of the recumbant fold, northeast trending faults offset the fold
hinge prior to thrust emplacement of the Cambrian ultramafic rocks over the Cascaden Ridge Formation (Myers et al., 2011). A subhorizontal thrust fault marks the contact between the Cascaden Ridge Formation and the structurally overlying Cambrian ultramafic rocks (Fig. 3.2). Thrust faulting is clearly identifiable in drill core and is marked by the occurrence of highly sheared serpentinite, which may locally reach a thickness of several meters (Brechtel et al., 2011; Myers et al., 2011).

The last recognized deformation event is a normal offset of the stratigraphy. Stratigraphic patterns suggest that the Lillian fault experienced some of the greatest normal and oblique offset in the deposit area. It appears that normal faulting down-dropped the south limb and recumbent fold nose and juxtaposed it against the fold limb in the north. The subhorizontal fault between the Cascaden Ridge and Cambrian rocks is also normally offset by shallow to moderately (30-50 degrees) south-dipping faults (Fig. 3.2; Brechtel et al., 2011; Myers et al., 2011).

The timing of the structural development is poorly constrained. However, it is clear that dike emplacement occurred sometime after the overthrust Cambrian rocks were emplaced. The east-west striking and moderately south dipping dikes are parallel to the orientation of the fold axis and related faults which occurred early in the structural history, further indicating that the feldspar porphyry dikes represent the earliest intrusions. The biotite porphyry and biotite-feldspar porphyry dikes, which can have a northwest strike and dip steeply to the southwest appear to have formed during the transition from compression to oblique offset in the deposit area (Myers, 2009). The timing between dike emplacement and normal faulting is not well constrained.

Less well understood, but possibly affecting the structural history of the deposit area, is the north-south striking Myrtle Creek fault west of the deposit area (Brechtel et al., 2011). The Myrtle Creek fault is interpreted to have a dextral oblique-slip, west side down movement and truncates the Lillian fault to the northwest (Athey et al., 2004a). The Myrtle Creek fault marks the western extent of the deposit area (Brechtel et al., 2011).
3.5 Hydrothermal Alteration

The earliest stage hydrothermal alteration at Money Knob is represented by a hornfels event, which resulted in the widespread formation of secondary biotite. The secondary biotite has been recognized in all lithologies, including the igneous rocks of the Cretaceous dike swarm (Klipfel, 2009). In hand specimen, secondary biotite causes a color change of the altered rocks, which have a characteristic dark brown hue. While secondary biotite has been observed in all rock types, this style of alteration is commonly only found locally as remnant patches that have been overprinted by the younger albite and/or sericite alteration (Brechtel et al., 2011; Myers et al., 2011).

Albite alteration is widespread at Money Knob and overprints the earlier biotite hornfels. Pervasive albite alteration is frequently associated with the development of black cherty silica. Albite alteration may also occur as selvages surrounding quartz veins or localized shear structures. Albite alteration is strongest in the middle volcanic unit where it is pervasive and can be completely texturally destructive. The white or creamy white massive albite is accompanied by disseminated arsenopyrite and pyrite. Ankerite and dolomite may be associated with albite alteration as these minerals occur with albite alteration in vein selvages (Brechtel et al., 2011; Myers et al., 2011).

Both the biotite and albite alterations are overprinted by sericite alteration which occurs in all rock types. It is generally pervasive but can occur as vein and shear/fracture zone selvages. Disseminated pyrite is the most abundant sulfide mineral associated with the sericite alteration although arsenopyrite may also be present. In the sedimentary rocks the sericite alteration destroys organic carbon resulting in a profound bleaching of the rocks (Brechtel et al., 2011; Myers et al., 2011). In the Cambrian mafic and ultramafic rocks, fuchsite alteration represents the analogue to the sericite alteration recognized elsewhere in the deposit area (Klipfel et al., 2009). An $^{40}$Ar/$^{39}$Ar isotope age date was collected from secondary white mica in a sericite- and clay-altered biotite porphyry, yielding an age of 88.9 ± 0.3 Ma (Athey et al., 2004b).

Clay alteration represents the youngest alteration style. Clay alteration resulted in the
widespread, locally pervasive and texturally destructive, formation of illite, smectite, and kaolinite. Clay alteration can be dull grey to yellow-grey and extremely fine-grained in the groundmass of intrusive and coherent volcanic rocks or the matrix of volcaniclastic and sedimentary rocks (Brechtel et al., 2011; Myers et al., 2011). The clay alteration is generally structurally controlled, forming broad zones around fault structures (Klipfel et al., 2009).

Carbonate is found in altered rocks throughout the Money Knob deposit area. Carbonate composition appears to be controlled by the host rocks and ranges from ankerite to dolomite. Secondary carbonate minerals form finely dispersed clusters of crystals around quartz veins (Klipfel et al., 2009). Complete replacement of mafic and ultramafic rocks by carbonate resulted in the characteristic listwanite alteration that can be brown, yellow-brown, orange, tan, white, and/or cream and ranges in composition from siderite to magnesite. Listwanite alteration is pervasive and most intensely developed as broad halos around south dipping faults.

3.6 Ore Zones

The hydrothermal footprint of the Money Knob deposit is large. Mineralized rock, defined to have gold grades above 0.1 g/t, occurs in an area that is approximately 2.5 km² in size and on average 280 meters thick (Klipfel et al., 2009). Gold is present as native gold grains that are typically intergrown with sulfide grains, most commonly arsenopyrite and pyrite. Metallurgical testing on drill core samples has shown that 90 percent of the gold is less than 500 microns in size with more or less evenly distributed grain sizes below 500 microns. Only a small proportion of the gold is tied up in the structure of the sulfide minerals (Myers et al., 2011). Company assay data suggests a strong correlation between Au and As (correlation coefficient of >0.8; Pontius et al., 2010).

A direct correlation between dikes and gold mineralization does not exist. However, the strongest mineralization is commonly found where the most abundant dikes occur. Apart from the Cretaceous dike swarm, the middle volcanic unit, and the upper sedimentary unit host the most significant amounts of gold. Gold is also found in the lower sedimentary unit and the
Cambrian mafic and ultramafic rocks where they are intruded by the Cretaceous dikes (Pontius et al., 2010).

The highest grades occur where hydrothermally altered rocks containing disseminated gold (0.3-0.8 grams/ton) have been overprinted by veins, locally increasing the grades to > 1.0 grams/ton. This style of mineralization and grade enhancement is most notable in the southern portion of the deposit area, termed the Core Zone (Pontius et al., 2010).

Gold in the northern portion of the deposit, termed the Sunshine Zone, is hosted almost completely in the upper sedimentary unit (Klipfel et al., 2009; Myers et al., 2011). Compared with the Core Zone, more abundant visible gold occurs and is found in thin (0.5-40 millimeters) quartz veins (Klipfel et al., 2009). While disseminated sulfides are present in the Sunshine Zone, they appear to be less indicative of mineralized rocks than in the Core Zone (Brechtel et al., 2011). Mineralized rock in the Sunshine Zone has a close spatial relationship to the Cretaceous dikes, however some dikes are noticeably barren (Myers et al., 2011).

In both parts of the deposit, the highest gold grades are typically associated with quartz veins and associated hydrothermal alteration. In hand specimen, different types of quartz veins can be distinguished based on quartz textures and sulfide minerals and abundance (Pontius et al., 2010). Early veins are stibnite-rich veins containing little to no quartz, followed by quartz-sulfide veins, and quartz-carbonate-sulfide veins (Pontius et al., 2010; Myers et al., 2011).

It is important to recognize that the emplacement of dikes, stibnite-rich veins, quartz-sulfide veins, and quartz-carbonate-sulfide veins all had different structural controls. This is evident from the east-west strike and shallow- to moderate-dip of the early dike set, the north strike and steep dip of stibnite-rich veins which cross-cut the early dike set, and the northwest strike and steep southwest dip of the later cross-cutting dike set which host quartz-sulfide and quartz-carbonate-sulfide veins having the same general orientation (Myers et al., 2011).
Chapter Four:
Petrology of the Biotite Syenite

4.1 Introduction

The Money Knob gold deposit formed in close spatial and temporal association with a dike complex that is primarily composed of biotite syenite. The present study focuses on the biotite syenite dikes as they represent an important temporal marker in the development of the deposit and provide an ideal framework in which to describe quartz veining and alteration associated with the mineralizing event. The biotite syenite dikes crosscut early stages of mineralization, but are also altered and mineralized. The present chapter describes the petrographic characteristics of least-altered samples and establishes the original geochemical characteristics of the dikes for later comparison with altered samples.

4.2 Methodology

Petrographic analysis was conducted on seven macroscopically least-altered samples collected from the biotite syenite intrusive complex. Thin sections of the least-altered samples were inspected by optical microscopy in transmitted and reflected light to study primary textures and to identify the mineralogy of the least-altered biotite syenite. In addition, optical cathodoluminescence microscopy was performed to study primary zoning patterns of feldspars. Cathodoluminescence microscopy was conducted with a HC5-LM hot-cathode CL microscope (Lumic Special Microscopes, Germany) operated at 14 kV with a current density of ca. 10 μA mm\(^{-2}\) (Neuser, 1995). CL photomicrographs were taken using a high sensitivity, double-stage Peltier cooled Kappa DX40C CCD camera with exposure times ranging from 5 to 10 seconds. In addition to the cathodoluminescence microscopy, the least-altered syenite samples were studied by scanning electron microscopy. Microtextural relationships were mostly studied in backscatter.
Table 4.1: Least-altered biotite syenite samples.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Drill Hole</th>
<th>Depth (m)</th>
<th>Dike Type</th>
<th>Dike Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>147905A</td>
<td>MK-09-36</td>
<td>299.82</td>
<td>Biotite porphyritic syenite</td>
<td>Coarser-grained, biotite phenocryst abundant</td>
</tr>
<tr>
<td>148175A</td>
<td>MK-09-37</td>
<td>357.74</td>
<td>Biotite porphyritic syenite</td>
<td>Coarser-grained, biotite phenocryst abundant</td>
</tr>
<tr>
<td>148724A</td>
<td>MK-09-37</td>
<td>369.05</td>
<td>Biotite porphyritic syenite</td>
<td>Coarser-grained, biotite phenocryst abundant</td>
</tr>
<tr>
<td>150141A</td>
<td>MK-09-41</td>
<td>164.79</td>
<td>Equigranular biotite syenite</td>
<td>Equigranular, evenly distributed biotite and K-feldspar</td>
</tr>
<tr>
<td>150176A</td>
<td>MK-09-41</td>
<td>204.06</td>
<td>Equigranular biotite</td>
<td>Equigranular, evenly distributed biotite and K-feldspar</td>
</tr>
<tr>
<td>151112A</td>
<td>MK-09-43</td>
<td>372.78</td>
<td>Biotite-feldspar porphyritic syenite</td>
<td>Coarser-grained, biotite-feldspar phenocryst abundant</td>
</tr>
<tr>
<td>151114A</td>
<td>MK-09-43</td>
<td>374.17</td>
<td>Biotite-feldspar porphyritic syenite</td>
<td>Coarser-grained, biotite-feldspar phenocryst abundant</td>
</tr>
</tbody>
</table>
electron images, with phase identification being based on semi-quantitative energy-dispersive X-ray spectrometry. An FEI Quanta 600i environmental scanning electron microscope with a PGT Prism energy-dispersive X-ray spectrometer was used. All backscatter electron images were taken at operating conditions of 20.0 kV, with a working distance of 10 millimeters.

Geochemical analysis of the least-altered biotite syenite samples was performed by ALS Chemex in Vancouver, British Columbia. The samples were crushed in a steel jaw crusher and subsequently milling to obtain a sample pulp. Major element analysis initially involved fusing of the sample powders with a lithium metaborate flux. The fused samples were then poured into a platinum mould to obtain a glass disk to use in XRF analysis. Minor and trace element analysis involved a near-total digestion of the sample in a four acid solution, followed by ICP-MS analysis. The detection limits of the various analytical methods as well as the analytical results are given in Appendix A.

4.3 Petrography of Least-Altered Biotite Syenite

The groundmass of the least-altered samples investigated are composed of about 50-70 modal% K-feldspar, 10-50 modal% biotite, <10 modal% plagioclase, and <10 modal% quartz. Based on the modal proportions, the igneous rock samples can be classified as biotite syenite.

Based on their textures, the biotite syenite samples can be classified as equigranular or porphyritic. The three different biotite syenite dike types distinguished include biotite-feldspar porphyry samples which contains both biotite and sanidine (sanidine contains 1-8% Na) phenocrysts in a fine- to coarse-grained groundmass of K-feldspar and biotite, equigranular biotite syenite samples composed of fine-grained subhedral K-feldspar and 10-15 modal% biotite, and biotite porphyry samples which contains 20-50 modal% biotite and medium- to coarse-grained feldspar in the groundmass (Fig. 4.1; Table 4.1).

In biotite and biotite-feldspar porphyritic samples, biotite phenocrysts account for approximately 10-15 modal% of the rock. The biotite phenocrysts typically ranges from 0.1 to 1 centimeters in size, but some phenocrysts can be up to 2 cm in size. Biotite phenocrysts show
a well-developed cleavage but have no preferred orientation. The biotite phenocrysts are evenly distributed throughout the rock. Approximately half of the biotite phenocrysts are Mg-rich (Fig. 4.2A).

In biotite-feldspar porphyritic syenite samples, K-feldspar phenocrysts are lath-shaped sanidine and account for approximately 5 modal% or less and range from 0.25 to 2 cm in length. The sanidine phenocrysts show distinct oscillatory growth zoning (Fig 4.2B). The phenocrysts are randomly distributed throughout the biotite syenite, but tend to occur in clusters. Cathodoluminescence images show that some of the larger phenocrysts display distinct resorption textures.

The groundmass of the biotite syenite samples is composed of K-feldspar and rare plagioclase which are fine-grained, 20 to 100 micrometers in size, lath-shaped, and randomly oriented (Fig. 4.2C). The K-feldspar grains have slightly variable Na concentrations, ranging from 0 to ~5 weight percent Na$_2$O. Core-rim growth zoning is observed in cathodoluminescence and backscatter electron images in some of the larger groundmass feldspars.

Groundmass biotite occurs as randomly oriented euhedral to subhedral blocky grains (Fig. 4.2D). They show a well-developed cleavage and range from 0.1 to 2 mm in size. Almost all fine-grained groundmass biotite is Fe-rich, with a molar ratio of Fe/(Fe+Mg) $\geq$ 0.75.

Quartz is a minor constituent of the groundmass and generally fine-grained. Grain morphology is defined by surrounding grains and only rare subhedral to euhedral grains are observed.

Accessory minerals in the biotite syenite include apatite, titanite, and rare zircon. All accessory minerals are found as fine-grained euhedral to subhedral grains in the groundmass. Primary apatite is evenly distributed but accounts for <1 modal% of the rock. It typically occurs as elongate crystals that have a hexagonal shape in cross-section. Titanite is rare but evenly distributed throughout the rock. This mineral is fine-grained and typically occurs as euhedral diamond- or wedge-shaped grains. Zircon is a trace mineral and typically forms inclusions in biotite grains or is intergrown with biotite. The zircon grains are blocky or diamond-shaped and
Figure 4.1: Hand samples of biotite syenite. (A) Coarse-grained biotite porphyritic syenite sample containing ~10 modal% biotite phenocrysts. Sample 148724A. (B) Medium-grained biotite porphyritic sample containing abundant biotite. A cross-cutting quartz vein is surrounded by an inner selvage of albite, ankerite, and quartz and an outer selvage of clay minerals, quartz, and minor ankerite. The least-altered rock contains ~5 modal% biotite phenocrysts. Sample 147905. (C) Equigranular biotite syenite, medium- to fine-grained sample. The total biotite content is ~10 modal%. Sample 150141A. (D) Biotite-feldspar porphyritic sample containing abundant sanidine phenocrysts and some biotite phenocrysts. Cross-cutting veins are surrounded by selvages of albite, quartz, and minor ankerite. Sample 151112.
Figure 4.2: Photomicrographs of least-altered biotite syenite. (A) Biotite phenocryst surrounded by fine-grained groundmass biotite. Sample 147905. PPL. (B) Sanidine phenocryst showing resorption and oscillatory growth zoning. Sample 151112A. CL. (C) Sanidine phenocryst surrounded by partially altered groundmass K-feldspar. Sample 151112B. BSE. (D) Groundmass biotite in a groundmass of partially altered K-feldspar. Sample 150141A. PPL.
show a high relief. Some zircon inclusions in biotite are surrounded by pleochroitic radiation damage halos.

4.4 Geochemistry of Least-Altered Biotite Syenite

The major element data collected on the least-altered biotite syenite samples reveal that the intrusive rocks are indeed alkalic (Fig. 4.3; Le Bas et al., 1986). Based on Shand’s Index, the samples are metaluminous to weakly peralkaline (Maniar and Piccoli, 1989), with the most peralkaline sample being a fine- to medium-grained equigranular biotite dike with abundant biotite.

The trace element geochemical signature also indicates that all the least-altered rocks sampled are differentiated and alkaline (Winchester and Floyd, 1977). They have moderate to high Zr/Ti ratios ranging from 0.006 to 0.022. The Y/Nb ratios range from 0.586 to 0.897.

4.5 Age of Dike Emplacement

The age of formation of the Money Knob deposit can be effectively constrained through absolute age dating of the biotite syenite dike complex. Early feldspar porphyry dike emplacement occurred along faults dipping between 30-45 degrees and contemporaneously with regional compression responsible for the folding and thrusting in the deposit area. Thrusting is believed to have terminated sometime in the Late Cretaceous (Dover, 1994; Johnsson, 2000; Pontius et al., 2010). Relative age relationships suggest that emplacement of the biotite syenite dike complex occurred after thrusting and broadly synchronous with mineralization along steep dipping NW-trending faults and fracture zones. The steeply dipping mineralized veins and dikes are approximately parallel to the Lillian fault, suggesting that biotite syenite dike emplacement and mineralization possibly continued through the regional structural reorientation to strike-slip (Myers, 2009).

An absolute age date for the biotite syenite was established using U-Pb zircon geochronology.
Figure 4.3: Total alkali-silica plot for least-altered biotite syenite samples. The plot indicates that the igneous rocks at Money Knob are strongly to moderately alkalic, which agrees with the petrographic observations (diagram modified after Maniar and Piccoli, 1989). Field names of volcanic rocks were omitted.
Selected single zircon grains were dated by single-collector ID-TIMS following chemical abrasion at the Pacific Centre for Isotopic and Geochemical Research at the University of British Columbia, Vancouver (procedure modified from Mundil et al., 2004; Mattinson, 2005; Scoates and Friedman, 2008). Analytical blanks were 0.2 pg for U and up to 2.0 pg for Pb. Data reduction was conducted using the Excel-based program of Schmitz and Schoene (2007). The program Isoplot 3.0 was used for the calculation of the regression intercepts and weighted averages (Ludwig, 2003). All errors are quoted at the 2 sigma (95%) level of confidence. Full sample preparation and analysis details are provided in Appendix A.

The results of U-Pb zircon analysis are presented in a Wetherill Concordia diagram (Fig. 4.4). One of five zircon grains analyzed was concordant within analytical uncertainty, with a $^{206}\text{Pb}/^{238}\text{U}$ age of 92.04±0.14 Ma (see Appendix A). The remaining four analyses are slightly discordant and lie along a chord projecting to older ages. A regression through all five data points produced a lower intercept with the Concordia of 91.5±0.9 Ma and an upper intercept of 667 ± 330 Ma (MSWD = 2.3). The more precise age of the single concordant analysis is interpreted to be the age of crystallization of the biotite syenite intrusion. The four discordant analyses are interpreted to reflect minor inheritance of older material within igneous zircon grains from the biotite syenite. However, the poor constraint on the upper intercept prevents any link to the age of inherited material to be established.
Concordant age of 92.04 ± 0.14 Ma

Regression for all 5 zircon grains:
lower intercept concordia 91.5±0.9 Ma
upper intercept concordia 667 ± 330 M
MSWD = 2.3

Figure 4.4: Wetherill Concordia diagram presenting the U-Pb data from the biotite syenite. For details on the analytical technique and data used for construction see Appendix. A
5.1 Introduction

Quartz veins are widespread at the Money Knob deposit and many high-grade zones are the product of gold hosted in these veins. The present chapter presents the findings of a detailed petrographic study of the quartz veins. Using CL microscopy, different quartz types have been distinguished within the veins. Fluid inclusion petrography revealed that these different types of quartz can also be distinguished based on their fluid inclusion inventory. The study of fluid inclusion petrography and microthermometry yielded important constraints on the temperature and pressure conditions under which the different types of quartz formed. Recognition of changes in physicochemical conditions of quartz formation provided a framework, allowing the reconstruction of the environment of gold deposition. In addition to the fluid inclusion data, some important constraints can be derived from the vein mineralogy described in this chapter.

5.2 Methodology

A total of eight representative samples were collected to study the petrography of the quartz veins crosscutting the biotite syenite (Table 5.1). The samples were collected in a way that all alteration styles macroscopically associated with the quartz veins in drill core (albite, sericite, clay, and carbonate alteration) were represented. The selected quartz vein samples originated from core intervals having variable gold grades, but did not typically contain visible gold. For that reason, three quartz veins hosted by other host rocks were analyzed to study the textural setting of gold. In addition, one sample of a stibnite-rich vein from sedimentary rocks was inspected.

Polished thin sections were prepared for the microscopic investigations. In addition, 60
micrometer-thick sections were obtained from selected samples for additional fluid inclusion investigations. Optical microscopy using transmitted and reflected light was performed using an Olympus BX51 research microscope. CL investigations were conducted with a HC5-LM hot-cathode CL microscope (Lumic Special Microscopes, Germany) operated at 14 kV with a current density of ca. 10 μA mm\(^{-2}\) (Neuser, 1995). CL photomicrographs were taken using a high sensitivity, double-stage Peltier cooled Kappa DX40C CCD camera with exposure times ranging from 5 to 10 seconds.

Fluid inclusion petrography was conducted with an Olympus BX51 polarizing light microscope. Microthermometric data on selected fluid inclusion assemblages were collected with a Fluid Inc.-adapted U.S. Geological Survey gas-flow heating and freezing stage (Werre et al., 1979) that was calibrated using synthetic fluid inclusions. Freezing temperatures were accurate to ±0.1°C. The accuracy of heating measurements ranged from approximately ±2°C at 200°C to approximately ±10°C at temperatures above 600°C. Imaging of representative fluid inclusions was performed using an Optronics Microfire A/R digital camera.

5.3 Vein Petrography

A total of 12 vein samples were collected from the biotite syenite, the upper sedimentary unit, and the serpentine unit found at the Money Knob deposit (Table 5.1).

5.3.1 Hand Samples

The collected quartz-sulfide and quartz-carbonate-sulfide veins range from 0.2 to 2 centimeters in width. Veins are composed of at least 90 modal% quartz. Other minerals identified include calcite, pyrite, arsenopyrite, stibnite, biotite, tourmaline, and/or kaolinite. Quartz can be translucent or milky, coarse- to fine-grained, and euhedral to anhedral. The majority of quartz veins are planar, but some veins are highly irregular, possibly due to deformation postdating vein formation. Euhedral quartz is uncommon but visible in at least one sample (147905) and in a few samples vuggy open space exists. Planar quartz veins tend to be associated with the most distinct
<table>
<thead>
<tr>
<th>Sample</th>
<th>Vein Type</th>
<th>Quartz Type</th>
<th>Host Rock</th>
<th>Visible Gold</th>
</tr>
</thead>
<tbody>
<tr>
<td>140219</td>
<td>stibnite-rich</td>
<td>No quartz</td>
<td>upper sedimentary unit</td>
<td>No</td>
</tr>
<tr>
<td>147905</td>
<td>quartz–carbonate–sulfide</td>
<td>Q1, Q2, Q3</td>
<td>biotite syenite, coarser-grained, biotite phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>148175B</td>
<td>quartz–sulfide</td>
<td>Q1, Q2</td>
<td>biotite syenite, coarser-grained, biotite phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>149942-2</td>
<td>Stibnite-rich</td>
<td>Q1</td>
<td>biotite syenite, equigranular, evenly distributed biotite and K-feldspar</td>
<td>No</td>
</tr>
<tr>
<td>151112A</td>
<td>quartz–carbonate–sulfides</td>
<td>Q1, Q2</td>
<td>biotite syenite, coarser-grained, biotite–feldspar phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>151112B</td>
<td>quartz–carbonate–sulfides</td>
<td>Q1, Q3</td>
<td>biotite syenite, coarser-grained, biotite–feldspar phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>151114</td>
<td>quartz–carbonate–sulfides</td>
<td>Q1, Q2</td>
<td>biotite syenite, coarser-grained, biotite–feldspar phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>MK-07-08</td>
<td>quartz–carbonate–sulfides</td>
<td>Q1</td>
<td>biotite syenite, coarser-grained, biotite phenocryst abundant</td>
<td>No</td>
</tr>
<tr>
<td>142.2m</td>
<td>quartz–carbonate–sulfides</td>
<td>Unknown</td>
<td>upper sedimentary unit</td>
<td>Yes</td>
</tr>
<tr>
<td>Mk-10-53-1</td>
<td>quartz–carbonate–sulfide</td>
<td>Unknown</td>
<td>upper sedimentary unit</td>
<td>Yes</td>
</tr>
<tr>
<td>Mk-10-53-2</td>
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<td>Unknown</td>
<td>upper sedimentary unit</td>
<td>Yes</td>
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<tr>
<td>Mk-10-63</td>
<td>quartz–sulfide</td>
<td>Q1</td>
<td>biotite syenite, equigranular, evenly distributed biotite and K-feldspar</td>
<td>Yes</td>
</tr>
<tr>
<td>Mk-10-67-2</td>
<td>quartz–sulfide</td>
<td>Unknown</td>
<td>serpentinite</td>
<td>Yes</td>
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Figure 5.1: Photographs of different vein types. (A) Hand specimen of discontinuous early stage stibnite-rich veins. The pen cap is ~1 inch wide. Sample 140219. (B) Quartz-carbonate-sulfide vein with stibnite being the most abundant sulfide mineral. Sample 149942-2. RL. (C) Hand specimen of quartz-carbonate-sulfide vein with distinct alteration selvages cutting biotite syenite. The igneous host rock is least-altered away from the vein. Sample 151112A. (D) Quartz-carbonate-sulfide vein with coarse-grained quartz with planar edges hosted in altered biotite syenite. Sample 147905. CPL. (E) Vein with subhedral fine-grained quartz with serrate and irregular edges. Sample 151114. CPL. (F) Quartz-carbonate-sulfide vein with coarse-grained quartz and hydrothermal biotite at the edges. Sample 151112A. PPL.
alteration halos (Fig. 5.1).

5.3.2 Quartz Types and Related Fluid Inclusion Assemblages

Based on their CL characteristics, four different types of quartz have been identified in the samples crosscutting the biotite syenite at Money Knob (the same quartz types occur in the three quartz vein samples not hosted by biotite syenite).

The first type of quartz, referred to as Q1, consists of generally medium to coarse-grained quartz with planar or serrate grain boundaries. In some vein sample (in particular 147905), quartz Q1 is euhedral, forming large crystals growing into open space. In cross-polarized light, the quartz Q1 shows undulose extinction, either throughout the entire crystal or within subgrains. Q1 is characterized by its blue to blue-grey short-lived CL emission which changes to a long-lived blue-grey-pink or blue-grey-purple emission after tens of seconds of electron bombardment (Fig. 5.2A-F). Q1 shows complex zoning patterns, which includes oscillatory growth zoning and sector zoning. These types of zoning are most distinct after short exposure to the electron beam although the long-lived CL emission still shows these zoning patterns (Fig. 5.2F). Other zoning patterns include distinct changes in CL color that are not parallel to crystal faces. In many cases, the zones of different CL color truncate each other at variable angles. It is possible that these CL textures are related to deformation of the quartz.

In plane polarized light, Q1 is transparent and contains both abundant primary and secondary fluid inclusion assemblages (Table 5.2). Primary fluid inclusions in Q1 are abundant in some vein samples, allowing microthermometric investigations while other samples of Q1 do not contain fluid inclusion assemblages that can be unequivocally identified as being primary. Three-phase liquid-vapor inclusions coexist with vapor-rich inclusions. The inclusions form negative crystal shapes and range from 5-25 µm in size. Microthermometric measurements were performed on two fluid inclusion assemblages that showed consistent volumetric liquid to vapor ratios of about 60/40. These measurements showed that the three-phase H₂O-CO₂ inclusions have homogenization temperatures between 300-310°C (Fig 5.3.A-B; eight microthermometric
Figure 5.2: Photomicrographs of Q1 in a quartz-carbonate-sulfide vein (sample 151112A). (A) Q1 with hydrothermal biotite intergrowths at edge of vein. PPL. (B) Coarse-grained quartz with planar and serrate grain boundaries. CPL. (C) CL image of Q1 quartz during initial electron irradiation (blue) along with apatite and ankerite (yellow-pink and bright red, respectively). Notice the difference between the oscillatory growth zones and the irregular growth zones. (D to F) CL images taken every ~ 7 seconds illustrating the change in the short-lived CL emission of Q1. Note consistency of the CL emission for both the apatite and ankerite grains. The lower right portion of the field of view has been previously irradiated. For that reason, this area of the quartz has no short-lived CL in (C) and (D).
The presence of CO₂ is confirmed by visible double bubbles (CO₂ vapor, CO₂ liquid, and H₂O liquid) and the nucleation of a clathrate when the inclusions are cooled to ~5°C. The CO₂ homogenization temperature to liquid is 24°C. The presence of three phases at room temperature suggest that the primary inclusions have a CO₂ content that exceeds ~10 mol%, however some volumetric measurements suggest >20 mol%. Based on the homogenization temperatures of the carbonic phase and the homogenization temperature of the fluid inclusion assemblages there must be other constituents in the inclusions apart from H₂O and CO₂ (Bodnar et al., 1985; Bakker and Diamond, 2000).

In addition to the high-temperature primary fluid inclusions, Q1 contains abundant secondary fluid inclusion assemblages of three-phase CO₂-H₂O inclusions and vapor-rich inclusions formed through fluid immiscibility. The secondary inclusions form trails that crosscut primary quartz growth zones. These fluid inclusions are subhedral to euhedral and 5-15 µm in size and

Table 5.2: Microthermometric data of fluid inclusions from Money Knob.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Analyzed Inclusions in Assemblage</th>
<th>Homogenization Temperature Ranges</th>
<th>Melting Temperature</th>
<th>Quartz Generation Host</th>
<th>Fluid Inclusion Assemblage Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>MK-07-08 142.2m</td>
<td>4</td>
<td>305°C to 310°C</td>
<td>N/A</td>
<td>Coarse-Grained Q1 quartz</td>
<td>High-temperature primary inclusions</td>
</tr>
<tr>
<td>MK-07-08 142.2m</td>
<td>4</td>
<td>300°C to 310°C</td>
<td>N/A</td>
<td>Coarse-Grained Q1 quartz</td>
<td>High-temperature primary inclusions</td>
</tr>
<tr>
<td>147905</td>
<td>4</td>
<td>161°C to 166°C</td>
<td>N/A</td>
<td>Euhedral Q1 quartz</td>
<td>Low-temperature secondary inclusions</td>
</tr>
<tr>
<td>149942</td>
<td>5</td>
<td>154°C to 165°C</td>
<td>10.1°C ± 0.2°C</td>
<td>Euhedral Q1 quartz</td>
<td>Low-temperature secondary (?) inclusion</td>
</tr>
</tbody>
</table>
Figure 5.3: Photomicrographs of fluid inclusions contained in Q1. (A) Area of analyzed primary fluid inclusion assemblage. Sample MK-07-18_142.2. PPL. (B) Area of analyzed primary fluid inclusion assemblage hosted by Q1 quartz. Sample MK-07-18_142.2. CL. (C) Analyzed high-temperature fluid inclusion assemblage (see Table 5.2 for microthermometric data). Sample MK-07-08-142.2. PPL. (D) Secondary fluid inclusion trail containing both three-phase CO$_2$-H$_2$O inclusions and vapor-rich inclusions indicating entrapment during fluid immiscibility. Sample 147905. PPL. (E) Euhedral Q1 quartz hosting secondary (?) fluid inclusions. Sample 1499942-2. CL. (F) Analyzed euhedral fine-grained secondary (?) low-temperature inclusions. Sample 1499942-2. PPL.
commonly have inconsistent liquid to vapor ratios, suggesting heterogeneous entrapment. In one case, a secondary trail containing three-phase CO$_2$-H$_2$O inclusions with a consistent liquid to vapor ratio of 80/20 was noted (Fig. 5.3D-E). Four inclusions in this assemblage homogenized at temperatures between 161-166°C. Based on the evidence of fluid immiscibility, it is concluded that this homogenization temperature equals the temperature of entrapment, with no pressure correction required.

Another secondary (?) fluid inclusion assemblage has been analyzed in a euhedral Q1 quartz grain intergrown with carbonate and stibnite in a stibnite-rich vein. Five inclusions in this assemblage homogenized at temperatures between 154 and 165°C. The melting point was determined to be at 10.1°C, which indicates the presence of CO$_2$ in the inclusion. As the fluid inclusion assemblage shows no evidence for fluid immiscibility, the homogenization temperature would need to be pressure corrected to obtain the temperature of entrapment.

In some vein samples, native gold is hosted by quartz Q1. In those samples, gold grains always occur along fractures that are not healed, implying that gold formation postdated the deposition of quartz Q1 (Fig. 5.3F). Although no conclusive evidence was found, it appears that the gold formed from low-temperature hydrothermal fluids as abundant necked secondary inclusions occur in the quartz Q1 around the gold grains. Bodnar et al. (1985) showed that these types of necked inclusions typically form at temperatures less than 200°C. In terms of fluid inclusion petrography, the necked secondary inclusions resemble those associated with the quartz types Q2 and Q3 described below. However, there is no evidence in the investigated samples to link gold formation to either of these two quartz types.

A second type of quartz, referred to as Q2, occurs typically as very fine-grained net-textured interstitial quartz between Q1 grain boundaries or fills micro-fractures crosscutting Q1. The fine-grained Q2 in this textural context shows no growth patterns such as oscillatory growth zoning. However, in open space, Q2 forms euhedral crystals that show oscillatory growth zoning. Q2 has a distinct bright yellow short-lived CL emission, which changes to a long-lived dark yellow or purplish yellow CL emission with exposure to the electron beam (Fig. 5.5A-F). The crosscutting
relationships unequivocally show that Q2 postdated the previously formed Q1. Within the quartz veins investigated, Q2 is generally less abundant than Q1. However, locally the net-textured Q2 can be very abundant, comprising more than 50 modal% of the vein quartz. Most quartz veins investigated contain quartz Q2.

The distinct cloudiness of Q2 is caused by the abundant presence of small fluid inclusions. Trails of fluid inclusions associated with the yellow CL haloes occur in planes of wispy or stringy overprints of Q1 (Fig. 5.5.D-F). Individual inclusions are subhedral to anhedral and/or anastomosing and inter-connected, ranging in size from 2-30 µm (Fig 5.6A-D). The fluid inclusions have inconsistent liquid to vapor volumetric phase ratios, ranging from 90 percent liquid and 10 percent vapor to 100 percent vapor. The inconsistent phase ratios are most likely a result of necking of the inclusions. The occurrence of necking precluded microthermometric investigations on these inclusions. However, the necking characteristics for fluid inclusion assemblages found in Q2 point to formation at temperatures below 200°C. Ratty, irregular shaped inclusions are typical of immature inclusions formed at temperatures less than 200°C in response to sluggish maturation of the quartz (Bodnar et al., 1985).

In rare case, euhedral crystals of Q2 formed in open space contain primary fluid inclusions. These inclusions are also strongly affected by necking, preventing microthermometric measurements. In rare inclusions that are the least necked, three phases are present and double bubble inclusions are visible.

A third type of quartz, referred to as Q3, is recognized as a late-stage overgrowth on euhedral Q1 crystals. This type of quartz is rare and has only been observed in one sample (sample 147905). Although no direct contact relationships with Q2 have been observed, fluid inclusion petrography suggests that Q3 is younger than Q2. In one location, Q3 overgrowth truncated a secondary fluid inclusion trail hosted by Q1. This inclusion trail forms part of the network-like distribution of Q2 inclusions (Fig. 5.7A-D). In plane polarized light, Q3 is distinctly cloudy, allowing distinction from Q1. Q3 has a short-lived blue-purple or blue-pink CL emission which changes to a longer-lived dark purple or dark blue-red emission during electron bombardment.
Figure 5.4: Photomicrographs of gold grains in Q1 quartz. All samples are from sample MK-10-63. (A-B) Fine-grained gold grains hosted in microfractures in Q1 quartz. RL. (C-D) Plane light image of the gold locations in image (a). PPL. (E-F) Gold is clearly hosted in fractures that cross-cut Q1. No visible Q2 is present.
Figure 5.5: Photomicrographs of Q2. (A) Fine-grained Q2 quartz formed in a cavity within coarse-grained Q1 quartz. Numerous small fluid inclusions related to Q2 make the quartz Q1 look dirty. Sample MK-07-18_142.2. CPL. (B) The quartz Q2 shows a bright yellow luminescence after ~5 seconds of exposure. The quartz Q1 is crosscut by abundant inclusion trails surrounded by yellow CL. Sample MK-07-18_142.2. CL. (C) Bright yellow luminescence of Q2 quartz illustrating individual grains. The exposure time is ~ 1 sec. Sample MK-07-18_142.2. CL. (D) Quartz-carbonate-sulfide vein with kaolinite replacing biotite. The ‘dirty’ look is caused by wispy inclusions related to Q2 quartz. Sample 151112B. PPL. (E) Image (D) after ~7 second exposure showing the net textured Q2 CL signature overprinting Q1 quartz. (F) Image (D) after ~21 seconds of exposure showing the decreased CL response.
The CL signature of Q3 is very similar to the emission of Q1. However, no evidence for growth zoning has been observed (Fig. 5.7B-D). Q3 contains abundant primary fluid inclusions that have irregular shapes. These fluid inclusions have inconsistent liquid to vapor volumetric ratios, probably due to necking at temperatures below 200°C (Fig. 5.7E-F). No visible petrographic evidence exists for CO₂. Ice melting temperature of 0.0 to -0.5°C were observed, suggesting that the inclusions have a very low salinity.

In one thin section (sample 147905), formation of Q3 was followed by deposition of fine-grained, possibly originally amorphous, silica that shows colloform textures. Deposition of this type of silica occurred within open space created by large euhedral Q1 grains (Fig. 5.7A).

A summary diagram depicting the textural relationships between the different quartz types (Q1 through Q3) and their related fluid inclusion inventories is shown in Figure 5.8.

5.3.3 Accessory Minerals

The quartz veins investigated contain variable proportions of accessory minerals. These phases are particularly common along the margins of the quartz veins. Accessory phases identified include biotite, tourmaline, apatite, calcite, pyrite, arsenopyrite, stibnite, and kaolinite.

Biotite and tourmaline are associated with veins that contain more than 95 modal% Q1. Biotite is found exclusively at the margins of the quartz veins as fine-grained euhedral crystals. In many cases, biotite is entirely hosted by Q1. Tourmaline is less common and occurs with biotite along the margins of the quartz veins. It is found as fine-grained and subhedral grains. Both biotite and tourmaline show no CL emission (Fig. 5.9A). Calcite is the most common accessory mineral. It is found in veins which contain Q1 and Q2 (also minor Q3). Calcite universally shows a bright orange CL emission that is long-lived. In some cases, concentric growth zoning can be observed. Irregular zoning patterns also occur. In veins containing only Q1, calcite is fine-grained. The calcite is usually found along the margins of the quartz veins, in contact with the wall rock. In veins dominated by Q1, coarse-grained calcite grains can be associated with stibnite. Calcite makes up approximately 5 modal% in veins that contain both Q1 and Q2. The calcite in these veins is fine- to medium-grained and occurs as individual interstitial
Figure 5.6: Photomicrographs of secondary fluid inclusion assemblages related to the formation of Q2. All images are from sample 151112B. (A) Q2 quartz in a quartz-carbonate-sulfide vein overprinting and replacing Q1. CL. (B) Irregular secondary fluid inclusion assemblages, mostly occurring around larger, partially overprinted Q1 quartz grains. PPL. (C) Individual secondary fluid inclusions with irregular shapes and inconsistent liquid-vapor ratios hosted in Q2. PPL. (D) Sets of wispy secondary fluid inclusion assemblages related to Q2 overprinting Q1 quartz. PPL.
Figure 5.7: Photomicrographs of Q3 and contained fluid inclusion assemblages. All images are of sample 147905. (A) Euhedral coarse-grained Q1 rimmed by anhedral, fluid inclusion-rich Q3 quartz (dashed lines). PPL. (B) Euhedral coarse-grained Q1 with overgrowth of Q3. CPL. (C) Different CL signatures of Q1 and Q3. Exposure time is ~6 sec. CL. (D) Decreased CL response after ~12 sec. exposure time. Also, truncation of fluid inclusion trails related to Q2 quartz by Q3 quartz. (E) Primary fluid inclusion assemblages contained in Q3. Note the distinct boundary between fluid inclusion dense and fluid inclusion sparse areas which marks the boundary between Q1 and Q3 quartz. PPL. (F) Example of primary inclusions in Q3 that have irregular shapes and inconsistent liquid-vapor ratios. PPL.
Figure 5.8: Schematic diagram illustrating textural relationships between different quartz types identified by CL microscopy and related fluid inclusion types. Q1 can show oscillatory growth zoning, sector zoning, and zoning patterns that may have developed during deformation. It contains both high-temperature (310-315°C) primary fluid inclusions and secondary subhedral to euhedral low-temperature (155-165°C) fluid inclusions. Q2 can be both fine-grained and euhedral rimming Q1 quartz or very fine-grained ‘stringers’ and ‘trails’ overprinting Q1. Q2 contains anhedral anastomosing fluid inclusions with inconsistent liquid to vapor ratios. Q3 rims Q1. It also truncates trails of Q2 which overprint Q1. The Q3 quartz contains abundant medium-grained anhedral liquid only, vapor only, and liquid+vapor inclusions that are interconnected.
Figure 5.9: Accessory minerals found in quartz-sulfide and quartz-carbonate-sulfide veins. (A) Accessory biotite intergrown with Q1 quartz. Sample 151112B. CPL. (B) Fine-grained needle-shaped stibnite and coarser grained arsenopyrite hosted in Q1 quartz. Sample 148175B. PPL. (C) Kaolinite in Q1 quartz which has been overprinted by trails of quartz Q2. Kaolinite is typified by a bright blue CL emission. The bright red CL emission is K-feldspar altered to albite and ankerite. Sample 151114. CL. (D) Apatite is identified by the distinct yellow/pink CL emission. Apatite is found in veins with all quartz types, however, this vein has only Q1 quartz. Sample 148175B. CL.
grains or as clusters of grains overprinting Q1. Clusters of calcite aggregates in veins consisting of Q1 and Q2 occur together with kaolinite as well as pyrite and arsenopyrite. In the vein that also contains Q3 (sample 147905), medium- to coarse-grained calcite can be found as a late open-space filling in contact with Q3.

Kaolinite is only found in veins that contain Q2 quartz. It occurs in close association with the fine-grained wispy net-textured Q2 quartz and typically appears to be related to the occurrence of calcite in the veins. However, individual small grains of kaolinite can be found at the margins of the quartz veins proximal to the wall rock. Kaolinite occurs as fine-grained crystals that have a bright blue short-lived CL emission that becomes darker over time (Fig. 5.9C). Kaolinite is a product of biotite replacement.

Pyrite, arsenopyrite, and stibnite are the most common sulfide minerals. Pyrite is found in all veins types, but occurs at variable abundances. Pyrite is uncommon in veins that contain only Q1. In these veins, pyrite forms fine-grained, subhedral to euhedral grains and aggregates. Pyrite is present throughout veins that contain both Q1 and Q2 (also minor Q3). In these veins, pyrite occurs at concentrations of 1 to 3 modal%, forming fine- to coarse-grained, subhedral to euhedral grains and aggregates. Pyrite is typically accompanied by arsenopyrite, which forms fine- to medium-grained, subhedral to euhedral crystals. Arsenopyrite is found in clusters with pyrite or forms individual grains hosted by quartz. Stibnite can be coarse-grained and massive in veins that only contain Q1. Bladed or acicular stibnite is present in veins that contain Q1 and Q2 (Fig. 5.9sB). It was attempted to identify fluid inclusions in stibnite contained in the two stibnite-rich veins investigated using the infrared capability of the fluid inclusion microscope, but no inclusions were found.
6.1 Introduction

Hydrothermal alteration of the biotite syenite at the Money Knob deposit generally occurs as selvages around fractures that may or may not have quartz veins filling them. The alteration halos may contain gold, usually occurring as native gold grains associated with sulfide minerals. Gold can also occur within the quartz veins. To better understand the conditions under which gold mineralization occurred, detailed microscopic investigations have been carried out to identify secondary phases and their textural relationships. Recognizing the relationships between different mineral associations and their textural characteristics is paramount to understanding the temporal and spatial relationships between different alteration styles and gold mineralization. The present chapter provides a detailed description of the alteration-induced mineralogical and textural changes in the biotite syenite. Based on replacement relationships, a sequence of alteration events is established.

6.2 Methodology

Petrographic analysis was conducted on 13 different hydrothermally altered samples collected from the biotite syenite dike complex (Table 6.1) that were collected immediately adjacent to the least-altered samples described in the previous chapter. The samples were selected on the basis of macroscopically identifiable variations in alteration styles and intensities. Some of the polished thin sections covered pronounced alteration boundaries between least-altered and altered biotite syenite to allow direct examination of alteration-induced mineralogical and textural changes.

The thin sections were inspected by optical microscopy in transmitted and reflected light.
Table 6.1: List of samples and their petrographic characteristics.

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<th>Dike type</th>
<th>Hand Specimen Alteration Type</th>
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to document alteration-induced changes in mineralogy and texture. In addition, optical cathodoluminescence microscopy was conducted to study primary and secondary zoning patterns of feldspars and to identify secondary carbonate and clay minerals such as ankerite and kaolinite. Operating conditions of the cathodoluminescence microscope were identical to those listed in the previous chapter. In addition to the cathodoluminescence microscopy, replacement relationships at small scales were studied by scanning electron microscopy. Microtextural relationships were mostly studied in backscatter electron images, with phase identification being based on semi-quantitative energy-dispersive X-ray spectrometry. An FEI Quanta 600i environmental scanning electron microscope with a PGT Prism energy-dispersive X-ray spectrometer was used. All backscatter electron images were taken at operating conditions of 20.0 kV, with a working distance of 10 millimeters.

6.3 Alteration Styles

Core logging revealed that the biotite syenite is affected by several distinct styles of hydrothermal alteration. Macroscopically, the quartz veins appeared to be commonly associated with sericite and albite alteration. Widespread clay alteration appears to overprint these two earlier mineral associations (e.g. 148724 see Table 6.1). Petrographic analysis suggested that the overprinting relationships are more complex than initially identified in the field. Thin section inspection revealed that many of the analyzed samples were affected by early secondary biotite formation that has been extensively overprinted by the other, macroscopically identifiable, styles of alteration. Overprinting relationships also affect the other styles of alteration. On the microscopic scale, no sample has only one style of hydrothermal alteration. Carbonate alteration is treated here as a separate style of alteration as no unequivocal textural relationships were found allowing the formation of secondary carbonate minerals to be tied to one of the other alteration styles characterized by the formation of secondary silicate minerals.
6.3.1 Biotite Alteration

Biotite alteration was not recognized in hand specimens during the sampling process. Secondary biotite can only be confidently identified microscopically. In most samples, biotite alteration is not extensive as it has been overprinted by the other styles of alteration. As a consequence, none of the samples analyzed can be classified as biotite-altered only, which prevents chemical mass changes associated with biotite alteration to be quantified (see Chapter 7).

Biotite alteration is observed in vein selvages which have been classified as either sericite- or albite-altered macroscopically due to the abundant presence of overprinting sericite and albite, respectively (Fig. 6.1A). In general, secondary biotite is uncommon in the sericite or albite alteration selvages as it has been destroyed by later stage alteration. However, where biotite is found, it occurs as fine-grained euhedral intergrowths at the margins of quartz veins (Fig. 6.1B) or as very fine-grained replacement minerals along grain boundaries and cleavage planes of primary biotite within the igneous host rock. Semiquantitative EDX analysis suggested that secondary biotite has a higher Mg content than the biotite it is replacing (Fig. 6.1C). Secondary biotite is typically replaced by secondary fine-grained muscovite (Fig. 6.1D). In some cases, no evidence of secondary biotite exists, presumably because it is completely pseudomorphed by fine-grained muscovite and later-stage alteration products.

6.3.2 Sericite Alteration

Sericite alteration is an early style of alteration recognized and occurs as selvages around individual 1-3 cm wide quartz veins. The sericite alteration selvages are between to 2-5 cm in width, although more pervasive alteration completely replacing primary textures of the igneous rock can also occur (Fig. 6.1A). The intensity of the alteration correlates with vein proximity. With increasing distance from the veins, a progression from completely texturally destructive sericite alteration to preferential replacement of primary biotite can be observed. The most
Figure 6.1: Biotite syenite samples affected by sericite alteration. (A) Hand sample of moderately sericite- altered biotite syenite cut by albite selvage. Sample 148175A. (B) Hydrothermal biotite hosted in a Q1 vein. Sample 151114B. PPL. (C) Primary biotite with cleavage controlled Mg-enrichment. Sample 151114B. CPL. (D) Muscovite replacing secondary biotite in and around Q1 quartz vein. Sample 151114B. CPL. (E) Primary biotite partially altered by fine-grained patchy muscovite. Sample 148175A. CPL. (F) Replacement of biotite by muscovite in a matrix of quartz and ankerite. Sample 150175B. BSE.
intense alteration has destroyed all primary textures.

The principal secondary mineral formed during sericitization is muscovite. Secondary muscovite is more widespread than secondary biotite, but also only occurs proximal to the quartz veins. In addition to the secondary biotite, muscovite also replaces primary biotite (Fig. 6.1E). Muscovite replaces the primary and secondary biotite along cleavage planes to ultimately pseudomorph the entire biotite grains, or as fine-grained patchy replacement (Fig. 6.1E). Muscovite also replaces K-feldspar. In zones of intense sericite alteration, muscovite completely pseudomorphs all K-feldspar and plagioclase grains as very fine-grained, subhedral to anhedral grains, controlled by remnant grain boundaries and/or cleavage planes. Where feldspar replacement by muscovite is less prevalent, muscovite is very fine-grained, occurring along microfractures and grain boundaries (Fig. 6.1F). Complete replacement of all feldspars in a given sample has not been observed. However, muscovite has been observed to completely replace biotite in some samples.

In strong muscovite alteration where primary and secondary biotite is replaced by muscovite, pyrite and arsenopyrite can accompany the replacement. Pyrite and arsenopyrite occurring in biotite generally form grains that are aligned along cleavage planes or are found at the margins of biotite grains. The least-altered zones show only partial replacement of biotite, whereas in the most intensely altered zones, biotite grains are completely replaced by muscovite plus pyrite and arsenopyrite.

Sericite alteration of the biotite syenite was also accompanied by alteration of primary titanite to ilmenite. The secondary ilmenite forms fine-grained anhedral and/or wispy irregular grains. Complete replacement of titanite is seen in the most intensely altered areas, however, ilmenite only accounts for 50 to 70 modal% of the replaced titanite, the remainder typically being quartz. Ilmenite is also found as small grains along cleavage planes in the biotite.

6.3.3 Albite Alteration

Albite alteration followed primary mineral replacement by muscovite. It occurs as vein selvages and less commonly as pervasive fronts (Fig. 6.2A). Albite alteration is manifested by
Figure 6.2: Biotite syenite samples affected by albite alteration. (A) Hand sample of biotite syenite affected by albite alteration. Notice the vein selvage and the pervasive alteration front. Sample 151114A. (B) Partially replaced K-feldspar set in a groundmass of quartz and albite. Patches of muscovite and carbonate minerals. Sample 151112B. CPL. (C) Image of microcline grains in the groundmass that show Na-enrichment along grain boundaries. The microcline microlites are surrounded by anhedral albite and quartz. Sample 151112B. BSE. (D) Sanidine phenocryst showing partial alteration by albite. Complete edge replacement by albite. Groundmass largely replaced by albite and lesser ankerite. Sample 151114B. PPL. (E) CL image of the same sanidine phenocryst shown in previous image. Growth zoning and resorption textures with ratty and irregular edges where albite has completely replaced the sanidine. Sample 151114B. CL. (F) Arsenopyrite, pyrite, and stibnite. Stibnite occurs as irregular inclusions in arsenopyrite and less commonly in pyrite. Sample 147905C. BSE.
the flooding of the groundmass by albite and quartz and replacement of K-feldspar phenocrysts and K-feldspar and plagioclase microlites (Fig. 6.2B). The effects of albite alteration extend furthest away from the quartz veins when compared to biotite and sericite alteration.

Backscatter electron images and semiquantitative EDX analyses show that partial replacement of sanidine phenocrysts and microcline microlites by albite occurs along grain boundaries and microfractures. Replacement progressed from the grain boundaries inward and along fractures, and in some cases only islands of original K-feldspar remain (Fig. 6.2C). Where moderate to weak albite alteration replaces phenocrysts and microlites, albite and quartz occur only along grain boundaries, causing embayed edges.

In the most intensely altered albite alteration selvages, all primary groundmass textures are destroyed and replaced by anhedral albite and quartz, while primary biotite remains relatively unaltered or weakly altered by quartz and pyrite. Moderate to weak albite alteration in the groundmass occurs along microfractures and grain boundaries, creating anhedral patches and irregular and embayed edges (Fig. 6.2D). The amount of altered primary groundmass feldspar decreases to 10 to 40 percent, and the Na-metasomatism at the edges of feldspar grains penetrates less deeply.

Secondary pyrite and arsenopyrite also occur as disseminated grains throughout albite alteration selvages surrounding the quartz veins. The morphology and abundance of the sulfide minerals varies with vein proximity. Pyrite and arsenopyrite make up >99 modal% of the total sulfide minerals. They occur as euhedral to subhedral grains, ranging from 0.1 to 2 millimeters in size proximal to the veins. Lesser amounts of anhedral pyrite and arsenopyrite are found as irregular masses within the albite-altered groundmass of the biotite syenite farther away from the veins. The As content in arsenopyrite is variable, causing the arsenopyrite to have different hues in reflected light, including yellow, yellow-grey, pale yellow, and pale grey. Both arsenopyrite and pyrite appear to replace secondary ilmenite. Stibnite inclusions occur as anhedral inclusions within arsenopyrite or in intergrowth with arsenopyrite and/or pyrite (Fig. 6.2F).

Albite alteration has been observed overprinting sericite alteration on both, the hand sample
(Fig. 6.1A) and microscopic scales. Secondary muscovite is partially replaced at the edges by albite and quartz and in the most intense alteration albite overprints and replaces muscovite (Fig. 6.3A). Albite alteration creates embayed edges in secondary muscovite where it has replaced biotite (Fig. 6.3B). Where albite and sericite alteration are observed together, muscovite occurs as island and partially replaced fine-grained laths surrounded by albite and quartz, and sometimes with ankerite (Fig. 6.3C).

Observations of albite after sericite alteration in sanidine phenocrysts and K-feldspar groundmass are less clear (Fig. 6.3D). Only a few direct replacement textures were observed and it is possible that due to the fine-grained nature of sericite any secondary sericite has been destroyed by albite and quartz flooding. Based on clear replacement textures in residual primary biotite it appears that sericite alteration formed prior to albite alteration in the biotite syenite.

6.3.4 Carbonate Alteration

Carbonate alteration is composed of ankerite and lesser dolomite and can be prevalent in the biotite syenite. These carbonate minerals commonly occur in association with the products of albite and clay alteration. Ankerite and dolomite may also be abundant where albite and quartz flooding is strong. Ankerite dominates the carbonate alteration mineralogy and is most copious proximal to quartz veins, however it has been observed distal to the veins. Ankerite and dolomite form subhedral interstitial grains in the altered igneous rocks immediately adjacent to the veins (Fig. 6.4A). Carbonate alteration is the strongest directly proximal to veins, and can make up >75 modal% of the replacement minerals. Coupled with quartz and sometime albite flooding, carbonate alteration can be completely texturally destructive, but limited to within 1-5 cm of the quartz veins. More distal from veins, very fine-grained ankerite and dolomite replace primary biotite and secondary muscovite and occur as interstitial fine-grained patches that replace the groundmass (Fig. 6.4B). Where other alteration types are less abundant, ankerite can also occur as irregular stringers between groundmass flooding quartz and primary K-feldspar and/or secondary albite (Fig. 6.4C). Ankerite and dolomite in clay-altered samples are fine-grained and patchy. They occur as replacement minerals in biotite with some clay and/or quartz.
Figure 6.3: Microtextural evidence for albite alteration overprinting sericite alteration. (A) Muscovite replaced by fine-grained patchy ankerite and quartz. The groundmass consists of residual K-feldspar with secondary albite and quartz. The albite and quartz flooding partially alter the edges of residual muscovite. Sample 151112B. CPL. (B) Secondary biotite hosted in Q1 vein replaced by secondary muscovite. Ankerite and minor dolomite overprint both the biotite and muscovite. Surrounding the biotite and muscovite is anhedral albite. Islands of residual biotite remain in the secondary albite. Secondary albite grains truncate secondary muscovite, indicating formation after muscovite formation. Sample 151112A. BSE. (C) Residual biotite replaced along cleavage planes by muscovite. Overgrowing the residual biotite and secondary muscovite is quartz and albite. Accompanying albite and quartz replacement is ankerite which replaces both muscovite and biotite. Sample 151112B. BSE. (D) Sanidine phenocryst surrounded by quartz and albite flooding which completely destroyed the groundmass K-feldspar. The edges of the sanidine phenocryst are diffuse and enriched in Na. Within the sanidine a irregular patch of muscovite is partially replaced by quartz. Sample 148175A. BSE.
Figure 6.4: Biotite syenite samples affected by carbonate alteration. (A) Distinct vein selvages around a quartz-carbonate vein. The inner selvage contains >75 modal percent carbonate, the outer selvage is primarily composed of clay minerals. Sample 147905A. (B) Altered edge of a primary biotite grain by ankerite and quartz. Minor amounts of quartz and albite are overprinted by ankerite alteration. 150175B. BSE. (C) Ankerite and lesser dolomite replacing residual muscovite surrounded by secondary albite and quartz groundmass. Sample 147266B. CPL. (D) Ankerite overprints and occurs between residual K-feldspar and lesser plagioclase groundmass. Minor secondary quartz accompanies carbonate alteration. Sample 147905C. CPL.
6.3.5 Clay Alteration

Clay alteration is the most volumetrically abundant alteration style recognized in the analyzed samples. It can occur as both the most distal alteration selvages around quartz veins and pervasive alteration fronts. The effects of clay alteration can be detected quite distal to fluid pathways and generally overprint primary textures and earlier products of biotite, sericite, albite, and carbonate alteration. However, some sericite and/or albite vein selvages have been observed in the same location as pervasive clay alteration. These possible cross-cutting relationships may be a result of focused higher-temperature alteration related to quartz veining proceeded at least locally while more pervasive and widespread lower temperature alteration developed.

Clay alteration is characterized by the presence of kaolinite, smectite, illite, quartz, and lesser ankerite (Fig. 6.5A). The most intense clay alteration is found where hydrothermal quartz replacement of the groundmass is most prevalent.

Where clay alteration affects previously unaltered or weakly-altered biotite syenite, primary biotite is the most readily affected mineral (Fig. 6.5B). It can be completely replaced by clay minerals like kaolinite, Na-Fe-rich and less often Ca-bearing smectite, ankerite, and quartz (Fig. 6.5C). Fine-grained chlorite can also occur together with kaolinite, smectite, and carbonate minerals, generally forming the outermost zone of replaced primary biotite. The rims of chlorite define the remnant grain boundary and at least partially retain the original biotite grain shape. In less intense clay alteration, partial replacement by the same association of clay minerals, ankerite, and quartz occurs along grain boundaries and cleavage planes (Fig. 6.5D).

Clay alteration of feldspar ranges from complete destruction and/or replacement to weak alteration along the grain margins. Only in the most intense clay alteration are K-feldspar gains completely replaced by illite, kaolinite, and smectite, with lesser quartz. Where strong quartz flooding accompanies clay minerals, remnant feldspar grains are ratty and only faint grain boundaries are visible, although the cores of primary K-feldspar phenocrysts may remain (Fig. 6.5C). In less intensely affected samples, feldspar phenocrysts have diffuse edges and
Figure 6.5: Biotite syenite samples affected by clay alteration. (A) Hand samples of least-altered and clay-altered biotite syenite. Samples 150141A and 150141B. (B) Groundmass of clay-altered biotite syenite with remnant biotite surrounded by secondary clay minerals and chlorite. Sample 150175B. PPL. (C) Biotite grain that is completely replaced by smectite, illite, and quartz. Below the biotite is a remnant feldspar grain that is mostly replaced by illite and smectite. Sample 150141B. PPL. (D) Smectite and quartz partially replacing a biotite grain. Sample 150141B. CPL. (E) Sanidine phenocryst with clay alteration proceeding from the grain boundary inward. Sample 150175B. PPL. (F) Remnant groundmass feldspar laths are replaced by very fine-grained clay minerals. Sample 150175B. PPL.
clay minerals, quartz and variable albite inclusions may occur along cleavage planes and/or microfractures. Abundant very fine-grained inclusions of illite, kaolinite, and/or smectite in feldspar grains makes the feldspar look dirty in plane polarized light (Fig. 6.5E). Moderate clay alteration also causes embayments along the grain boundaries where quartz flooding of the groundmass impinges on the feldspar grains (Fig. 6.5F).

Ankerite and lesser dolomite can also occur with clay alteration and are found as random patches or islands within the replaced biotite or as elongated patches which cross-cut the entire remnant biotite grains. Ankerite and dolomite are more often found as diffuse fine-grained anhedral grains along the grain margins of the replaced biotite. In weakly altered samples, biotite replacement by clay minerals and quartz is restricted to grain boundaries and microfractures. Illite and/or kaolinite are also found between grain boundaries of altered biotite and K-feldspar, typically where secondary quartz is abundant (Fig. 6.5C).

Clay alteration commonly overprints previous higher-temperature alteration mineral associations. Where previously developed strong sericite or albite alteration is replaced by clay alteration, secondary muscovite and/or biotite are replaced by clay minerals (Fig. 6.6A). The most intense overprinting clay alteration completely destroys all sericite alteration textures, and/or anhedral albite is replaced by anhedral quartz flooding, very fine-grained clay minerals, and fine-grained patchy ankerite (Fig. 6.6B). In weakly clay-altered rocks, muscovite is replaced along grain margins often leaving only cores of secondary muscovite surrounded by patches of clay and carbonate minerals set in a matrix of anhedral quartz with residual secondary albite and/or primary K-feldspar.

Where moderate to weak clay alteration overprints moderate to weak sericite and/or albite alteration, coincidence of primary minerals, muscovite, albite, and clay minerals can be observed. Primary biotite best illustrates the development from primary through muscovite and albite alteration to clay alteration. Zoned from the core outward, windows of primary biotite remain surrounded by fine-grained secondary muscovite. Beyond the irregular and patchy muscovite, and partially consuming them at their margins, very fine-grained masses of clay
Figure 6.6: Microtextural evidence for clay alteration overprinting previous alteration styles. (A) Remnant biotite with muscovite in cleavage planes surrounded by albite alteration which was subsequently consumed by ankerite closest to the vein, and clay minerals distal to the vein. The groundmass is remnant albite with strong quartz flooding. Sample 148175C. CPL. (B) Remnant biotite phenocryst with grain boundary still visible replaced by muscovite which is overprinted by ankerite, calcite, clay minerals, and chlorite. The groundmass surrounding the remnant biotite is residual albite with quartz and patchy carbonate. Sample 151112B. PPL. (C) Remnant biotite replaced by muscovite and subsequently overprinted by clay minerals, ankerite, and rimmed by chlorite. Sample 150141B. CPL. (D) Muscovite partially replaces biotite in the core and is surrounded by clay and chlorite. Sample 151112B. PPL.
minerals, quartz, and chlorite make up the outer zone of replaced biotite. Minor amounts of ankerite and/or dolomite are also intergrown with the clays minerals (Fig. 6.6C-D).

Clay alteration is accompanied by little to no sulfide formation. Pyrite is the most abundant sulfide mineral. It is fine-grained, interstitial, irregular, and accounts for <1 modal% of the alteration association. Pyrite occurs with clay minerals, quartz, and carbonate minerals as a replacement product of biotite, commonly forming small grains aligned along cleavage planes.
7.1 Introduction

The present chapter examines the geochemical changes between least-altered rocks and their hydrothermally altered equivalents to determine which types of mass changes occurred in association with the development of the different styles of alteration of the biotite syenite. Graphical interpretation of the data was performed by comparing the least-altered samples with their altered equivalents, employing bivariant plots, molar ratio plots, and other techniques. Utilizing Gresens’ (1967) equation for mass-balance calculations, the mass transfer of individual elements (enrichment and/or depletion) was quantified, permitting important conclusions on the nature of hydrothermal alteration.

7.2 Materials and Methods

Whole-rock geochemical analysis was performed on a total of 20 representative least-altered biotite syenite samples and altered equivalents affected by the different alteration styles described in the previous chapter. The bulk samples were submitted to ALS Chemex in Vancouver, British Columbia, for whole-rock geochemical analysis. As a first step, specific gravity measurements were conducted. This was followed by crushing of the rocks in a steel jaw crusher and milling to obtain a sample pulp. Major element analysis initially involved fusing of the sample powder with a lithium metaborate flux. The fused samples were then poured into a platinum mould to obtain a glass disk that was subsequently used for XRF analysis. Minor and trace element analysis involved a near-total digestion of the sample in a four acid solution, followed by ICP-MS analysis. The gold content of the samples was measured by fire assay and
ICP-AES analysis using a 30 gram aliquot from each sample.

The total carbon and sulfur contents were measured by infrared spectrometry following heating of the samples in a Leco induction furnace. Sulfate sulfur was measured by carbonate leaching and gravimetric analysis, while sulfide sulfur was measured by sulfate dissolution with sodium carbonate and infrared spectrometry in a Leco induction furnace. Carbonate carbon was measured by coulometry, while non-carbonate carbon was measured by way of weak acid digestion and infrared spectrometry in a Leco induction furnace. The water content of the samples was measured by gravimetry after drying at 105°C, followed by heating of the samples in a Leco furnace to release H2O contained in the crystal structure of minerals. Finally, loss on ignition analysis was performed by heating the powdered samples to 1000°C. Analytical results and the detection limits of the various analytical methods can be found in Appendix A.

7.3 Immobile Element Signatures

Although the present study focused on the investigation of only one rock type occurring at Money Knob, it is clear that the biotite syenite comprises more than one intrusive phase. The petrographic investigations showed that the samples investigated vary in terms of texture, biotite abundance, and feldspar composition. This may suggest that the dikes were emplaced from slightly different magma batches. As a consequence, it can be expected that their geochemical characteristics vary slightly.

Due to the variable hydrothermal overprint, inspection of the immobile element signatures of the rocks analyzed probably represents the most reliable way to geochemically distinguish different intrusive phases. MacLean and Barrett (1993) showed that plots of incompatible elements that behave immobile in the hydrothermal environment can be used to identify magmatic affinities and magmatic fractionation trends. For multiple samples derived from a single precursor, elements plotted in bivariant plots should fall on alteration lines as the ratio of immobile elements would not have been modified during alteration (Fig. 7.1A). Data from multiple samples will define an alteration line because addition of other elements will dilute
Figure 7.1: Immobile element plots illustrating that different least- and weakly-altered dike types can be distinguished geochemically. (A) Conceptual diagram showing alteration trends for immobile elements (solid lines). The dashed lines intersect ‘alteration lines’ where different fractionation ratios occur within different intrusive bodies. Arrows indicate possible variations due to alteration either in mobility (away from alteration line) or concentration (along alteration line). (B) Plot of Al versus Ti whole-rock concentrations of biotite syenite samples. Alteration lines can be constructed for two distinct dike types (see text for details). (C) Plot of Zr versus Ti concentrations. The same two dike types can be identified in this plot. (D) Plot of Al versus Zr concentrations, indicating similar differences for the two dike types. (E) Plot of Zr versus Y concentrations. The two dike types can also be distinguished in this plot.
the concentrations of the elements under consideration, causing the samples to plot closer to the origin along the alteration line. In contrast, removal of other elements during alteration will concentrate the elements under consideration and cause the samples to plot farther from the origin along the alteration line (Fig. 7.1A). Immobility can be established if the interelement ratio is consistent. Typically, a correlation coefficient of better than 0.85 is assumed to indicate immobility (MacLean and Kranidiotis, 1987; Russell and Nicholls, 1988; MacLean and Barrett, 1993; Stanley and Madiesky, 1994).

Based on their immobile element geochemistry, three main sample groups of biotite syenite can be distinguished (Fig. 7.1). The three dike types can also be distinguished petrographically (Chapter 4).

The biotite-feldspar porphyritic syenite samples (blue crosses in Fig. 7.1B-E) are distinguished geochemically by higher Al/Ti and Zr/Ti ratios. Elements typically behaving immobile during hydrothermal alteration have correlation coefficients between 0.84 and 0.92 (Table 7.2; Fig. 7.2). Based on these plots, Al, La, Ti, Y, and Zr were chosen for mass-balance calculations for this dike type.

Equigranular biotite syenite samples (green circles in Fig. 7.1B-E) are distinguished by lower Al/Ti and Zr/Ti ratios. Bivariant plots constructed for assumed immobile elements (Fig. 7.3) show that Al, Nb, Ti, Y, and Zr appear to be consistently immobile. Their correlation coefficients range from 0.82 to 0.98 (Table 7.2). Consequently, these elements were chosen for mass-balance calculations in this dike type.

The two biotite porphyritic syenite samples showed a distinctly different geochemical signature than the other two dike types. Since both samples have different geochemical signatures, immobility testing could not be conducted. However, due to petrographic similarities to the biotite-feldspar porphyritic syenite, the mass-balance calculations were based on the immobile element suite established for that dike type.
Table 7.1: Correlation coefficients between immobile element pairs for different dike types.

<table>
<thead>
<tr>
<th>Immobile element used for plots</th>
<th>Biotite–feldspar porphyritic syenite dikes</th>
<th>Equigranular biotite syenite dikes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al/Zr</td>
<td>0.86</td>
<td>0.94</td>
</tr>
<tr>
<td>Al/Ti</td>
<td>0.92</td>
<td>0.98</td>
</tr>
<tr>
<td>Al/Y</td>
<td>0.92</td>
<td>0.96</td>
</tr>
<tr>
<td>Al/Nb</td>
<td>0.76</td>
<td>0.93</td>
</tr>
<tr>
<td>Al/Ta</td>
<td>0.78</td>
<td>0.85</td>
</tr>
<tr>
<td>Al/Th</td>
<td>0.72</td>
<td>0.80</td>
</tr>
<tr>
<td>Al/La</td>
<td>0.88</td>
<td>0.71</td>
</tr>
<tr>
<td>Zr/Ti</td>
<td>0.71</td>
<td>0.92</td>
</tr>
<tr>
<td>Zr/Y</td>
<td>0.81</td>
<td>0.94</td>
</tr>
<tr>
<td>Zr/Nb</td>
<td>0.89</td>
<td>0.82</td>
</tr>
<tr>
<td>Zr/Ta</td>
<td>0.83</td>
<td>0.70</td>
</tr>
<tr>
<td>Zr/Th</td>
<td>0.89</td>
<td>0.91</td>
</tr>
<tr>
<td>Zr/La</td>
<td>0.84</td>
<td>0.82</td>
</tr>
<tr>
<td>Ti/Y</td>
<td>0.94</td>
<td>0.96</td>
</tr>
<tr>
<td>Ti/Nb</td>
<td>0.75</td>
<td>0.94</td>
</tr>
<tr>
<td>Ti/Ta</td>
<td>0.85</td>
<td>0.85</td>
</tr>
<tr>
<td>Ti/Th</td>
<td>0.77</td>
<td>0.77</td>
</tr>
<tr>
<td>Ti/La</td>
<td>0.90</td>
<td>0.67</td>
</tr>
</tbody>
</table>
Figure 7.2: Bivariant plots of immobile elements for biotite-feldspar porphyritic syenite dikes. The high correlation coefficients between the selected elements ($R^2 > 0.85$) indicate that they behaved immobile during hydrothermal alteration. Based on the bivariant plots, Al, La, Ti, Y, and Zr were selected to conduct mass-balance calculations.
Figure 7.3: Bivariant plots of immobile elements for equigranular biotite syenite dikes. The high correlation coefficients between the selected elements ($R^2 > 0.85$) indicate that they behaved immobile during hydrothermal alteration. Based on the bivariant plots, Al, Nb, Ti, Y, and Zr were selected to conduct mass-balance calculations.
7.4 Molar Ratio Plots

Differences in the geochemical composition of the variably altered rocks can be effectively evaluated graphically using molar ratios plots of elements which have been added or subtracted during hydrothermal alteration. Using the technique of Stanley and Madiesky (1994) and modified by Warren et al. (2007), molar K/Al versus molar 2Ca+Na+K/Al were plotted to evaluate the degrees of Ca, Na, and K metasomatism caused by the different styles of hydrothermal alteration. Aluminum was chosen as a denominator for both ratios plotted as this element remained essentially immobile during alteration of the biotite syenite.

Figure 7.4A shows the (2Ca+Na+K)/Al molar ratio plotted against the K/Al molar ratio. Sericite-altered biotite syenite samples show elevated (2Ca+Na+K)/Al molar ratios when compared to least-altered samples (Fig. 7.4). As the K/Al ratio is approximately constant, K gains were probably only minor. Petrographic analysis confirmed that muscovite was largely formed at the expense of biotite although feldspars have also been replaced. Albite alteration is characterized by a loss in K, coupled with a gain in Na and perhaps Ca. Petrographically, this change is observable as the replacement of K-feldspar by albite and the formation of carbonate minerals (Fig. 7.4A). Clay-alteration caused a depletion of K coupled with a decrease in Na and Ca (Fig. 7.4A). This alteration style can be petrographically seen as a breakdown of the host rock, previously altered or not, to clay minerals.

In the molar ratio plot (2Ca+Na+K)/Al versus K/Al, formation of secondary carbonate minerals partially masks chemical changes caused by other alteration processes. Since CO$_2$ analyses have been performed for this study, it is possible to correct for the presence of carbonate minerals by subtracting the molar C content from the molar Ca content, which gives the new molar ratio (2(Ca-C)+Na+K)/Al (Fig. 6.4B). This calculation makes the assumption that all CO$_2$ analyzed is contained in calcite (Madeisky, 1996). Correcting the data for the total molar C content has two effects. Firstly, all data points in the diagram shift to the left as all samples contain a certain amount of carbonate minerals (Fig. 7.3B). Secondly, it should be noted that some of samples have negative (2(Ca-C)+Na+K)/Al values. The values are negative because a
Figure 7.4: Molar ratio plot of all biotite syenite samples affected by different styles of hydrothermal alteration. (A) Molar ratio plot of (2Ca+Na+K)/Al versus K/Al. (B) Molar ratio plot of (2(Ca-C)+Na+K)/Al versus K/Al. Subtracting the C content from Ca corrects for the carbonate content and removes the masking effect of carbonate alteration which provides a clearer picture on alteration processes resulting in the replacement or formation of silicate minerals.
significant proportion of the C content is contained in ankerite, not calcite.

Figure 7.5 shows a comparison of the molar ratio of $\text{CO}_2/(\text{Ca+Mg+}(\text{Fe-0.5S}))$ between the least-altered biotite syenite samples and samples affected by the other alteration styles. This molar ratio was established to measure carbonate alteration intensity in altered rocks associated with orogenic gold deposits (Kishida and Kerrich, 1987; Eilu and Groves, 2001). The subtraction of S from Fe corrects for the presence of pyrite.

As expected, carbonate-altered sample shows the highest $\text{CO}_2/(\text{Ca+Mg+}(\text{Fe-0.5S}))$ ratio, suggesting that nearly 90 percent of the Ca+Mg+(Fe-0.5S) is present as carbonate minerals. The albite- and clay-altered samples show the next highest ratios, which suggests that these samples have also been affected by strong carbonate alteration. The geochemical and petrographic observations are consistent with this observation since most albite alteration is accompanied by moderate to strong carbonate alteration, and carbonate alteration either overprints or accompanies clay alteration. Sericite-altered samples also contain significant amounts of Ca+Mg+(Fe-0.5S) which suggests that carbonate alteration overprinted previously developed sericite alteration and that the high ratio is a result of Fe and Mg depletion caused by sericite formation and enrichment in Ca due to later carbonate alteration. These observations, again, correspond with observations made in petrographic analysis, suggesting that variable amounts of carbonate minerals are present in rocks affected by all alteration styles.

7.5 Mass-Balance Calculations Using Gresens’ Equation

Geochemical mass-balance calculations were first introduced by Gresens (1967) to quantify alteration-induced changes in host rock volume, density, and composition as a result of alteration. The mass change can be quantified using the following equations:

$$\Delta X = a\{F_v(S^B/S^A) \times X^B\} - X^A \quad (1)$$

where:

$\Delta X =$ mass change for component $X$;
Figure 7.5: Molar ratio plot of $\text{CO}_2/(\text{Ca+Mg+}(\text{Fe-0.5S}))$ for all biotite syenite samples affected by different styles of hydrothermal alteration. The vertical lines represent the range of molar ratios for each alteration style. The horizontal lines represent the average value. See text for discussion.
a = starting mass (arbitrarily 100 g for concentrations in weight percent or 1 metric ton for concentrations in parts per million);

\[ F_v = \text{ratio of volume of altered rock to fresh rock (volume factor)}; \]

\[ S^B/S^A = \text{specific gravity ratio of altered rock to least-altered rock}; \]

\[ X^B = \text{concentration of component X in altered rock}; \]

\[ X^A = \text{concentration of component X in least-altered rock}. \]

To solve equation (1), either \( \Delta X \) or \( F_v \) must be known. The volume factor can be derived using an immobile element in Gresens’ equation and setting \( \Delta X = 0 \). This is possible because immobile elements show no alteration-induced mass changes. By solving for \( F_v \) for a selected immobile element, all other \( \Delta X \) can be calculated. Rearranging equation (1) to solve for \( F_v \) leads to:

\[ X^A = F_v(S^B/S^A) \times X^B \quad (2) \]

\[ X^A / X^B = F_v(S^B/S^A) \quad (3) \]

\[ F_v = (X^A S^A) / (X^B S^B) \quad (4) \]

Since each immobile element defines a unique reference frame in which to analyze changes during alteration (unique \( F_v \) value for each immobile element), each element will return different results when used in Gresens’ equation. Thus, each mass-balance calculation will vary depending on which immobile element is used. Under ideal circumstances, the different solutions would converge. However, as this is often not the case in reality (Mukherjee and Gupta, 2008), an average \( F_v \) value is calculated from equation (4). Once an average \( F_v \) is established, mass balance changes can be calculated from equation (1) for all other elements. Each average \( F_v \) is unique to the sample pair it has been calculated for. A list of data used in the calculation of the \( F_v \) values and the results of the mass-balance calculations on the various samples collected from Money Knob is given in the Appendix A.

Figures 7.6 and 7.7 illustrates the results of mass-balance calculations performed for selected
major elements. Sericite-altered rocks show variable enrichment and depletion patterns. These observations may be explained by the fact that sericite-altered rocks are always overprinted by subsequent alteration stages. These later stages of alteration may partially mask the true mass changes that occurred during sericite alteration. Sericite-altered rocks show a slight loss of K and Mg coupled with marked enrichments in Si and Ca (Fig. 7.6), which relates to the destruction of biotite and the formation of secondary muscovite, quartz, and carbonate minerals. Sericite-altered rocks also show gains in S and C, confirming that they have been overprinted by carbonate alteration (Fig. 7.7). The fact that S is gained but Fe is conserved indicates that S is fixed by the sulfidation of Fe. The increase in Si seems to reflect the addition of Si from the hydrothermal fluids.

Na enrichment is the most notable gain during albite alteration. In three albite-altered samples, Na, Si, Ca, and to a lesser extent Fe are all gained (Figs. 7.6). Au and S are also gained (Fig. 7.7). A fourth albite-altered sample is similar except it shows a loss of Ca (Fig. 7.6) and C (Fig. 7.7) suggesting that it was not overprinted by carbonate alteration (the least-altered sample used for mass balance calculation was affected by weak carbonate alteration). In the fifth albite-altered sample, there is a loss in Ca and a substantial gain in C which may not reflect albite alteration processes, but instead indicate carbonate alteration which has been superimposed on the albite-altered sample. Similar to sericite alteration, albite-altered samples have little Fe gain or loss, even where massive additions of S are seen, suggesting that S is fixed by sulfidation of Fe. As might be expected water is generally lost from the albite alteration as a result of replacing micas with the anhydrous feldspar (Fig. 7.7). Significant addition of quartz to the rock is implied from the addition of Si to the rock (Fig. 7.6).

All of the clay-altered samples show a marked increase in water content consistent with the formation of smectite. In all cases K is conserved. The clay-altered samples show losses of Na, Fe, and Mg consistent with the leaching of these cations (Fig. 7.6). Two of the clay-altered samples show C enrichment while one shows a loss. The sample with C loss also shows a Ca, suggesting that the least-altered rock sample was affected by slightly stronger carbonate alteration
Figure 7.6: Illustrates the results of mass-balance calculations performed for selected major elements.
Figure 7.7: Illustrates the results of mass-balance calculations performed for selected major elements.
Similarly two clay-altered samples show a loss in S but one shows a gain in S. Finally, one clay-altered sample shows a loss of Si while others show a gain (Fig. 7.6). The heterogeneity of the clay-altered samples compared to the albite- and sericite-altered samples suggests that they represent the superposition on previously altered rocks.

The one carbonate-altered sample shows an addition of C with no S or water gain. The gain in C is associated with a loss of Si and a slight gain in Ca. This seems to indicate that in this alteration albite and K-feldspar are stable while Ca, Mg and Fe silicates are destroyed and the cations preserved in the carbonate minerals. This is a very specific, relatively anhydrous, alteration. The superposition of this type of alteration on a previously albite-altered sample could explain the characteristics of the peculiar sample 148724B.

7.6 Metal Enrichment

Additional important constraints can be derived by comparing the Au content of the samples affected by different alteration styles (Figs. 7.8, 7.9). Samples affected by albite alteration have the highest average Au content, but also show pronounced S enrichment. By comparison, sericite- and clay-altered biotite syenite samples show wider variations in Au grades and slightly lower average values (0.25-0.33 ppm). These samples also show a variable changes in S and C. The increase in the C content in some samples correlates with petrographic observations of carbonate alteration. However, the strongest carbonate-altered samples only show slight enrichment in Au. Clay-altered samples all have an increase in H$_2$O while showing no increase in Au. These observations agree with the petrographic observation that carbonate alteration occurs distal to mineralized veins that are surrounded by either albite or sericite alteration halos.

Throughout the deposit, Au is strongly correlated with As and shows an association with Sb (Fig. 7.9A-B). In the samples from this study, a statistically significant correlation between Au and As does not exist, but gold grades are highest where arsenopyrite is found petrographically (Fig 7.9A). The same is true for Au and Sb in the samples of this study (Fig. 7.9B).
Figure 7.8: Gold content of biotite syenite affected by different styles of hydrothermal alteration. The vertical lines represent the range of gold grades for each alteration style. The horizontal lines represent the average value. See text for discussion.

Figure 7.9: Relative mass changes caused by hydrothermal alteration of the biotite syenite for Au, As, and Sb. (A) Au values are multiplied by 10,000 in order to scale the changes for comparison to As. (B) Au values multiplied by 1,000 in order to scale the changes for comparison to Sb. Mass changes in ppm.
CHAPTER EIGHT:

DISCUSSION

8.1 Spatial and Temporal Relationship to Igneous Activity

The Money Knob gold deposit is a low-grade high-tonnage gold deposit (20.6 million ounces of gold at an average grade of 0.52 g/t and a cut-off grade of 0.22 g/t) located in the Tintina Gold Province (Hart et al., 2002). Although the majority of the ore is hosted in a Devonian package of sedimentary and volcanic rocks and structurally overlying Cambrian mafic and ultramafic rocks, mineralization and alteration at Money Know is intimately associated both spatially and temporally with a felsic dike complex, the emplacement of which was structurally controlled. Structural controls are important both on the deposit scale, controlling ore and vein orientations, and on the regional scale controlling the location of the deposit within the Tintina Gold Province.

The focus on the biotite syenite intrusive phase of the dike complex provided a well-constrained framework in which to study the effects of mineralization and alteration in the later stages of deposit formation. The northwest striking and steep southeast dipping orientation of the dikes in the main ore body and the steeply dipping fault-controlled dikes around the Lillian fault indicate that dike emplacement occurred in the later part of the compressional regime of the deposit area. It is possible that dike emplacement continued during the early stages of strike-slip movement in the deposit area based on cross-cutting relationships and the orientation of dikes found in the Lillian fault. Since dike emplacement is synchronous with mineralization, it can be concluded that mineralization continued during the late stages of compression and possibly during the structural shift from compression to strike-slip.

The 92.04±0.14 Ma biotite syenite cross-cuts the earliest stages of mineralization, while also containing gold-bearing quartz veins and related hydrothermal alteration. Muscovite contained in a mineralized sericite- and clay-altered biotite syenite sample exposed at the surface yielded
a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 88.9±0.3 Ma (Athey et al., 2004b). The age difference between the age of intrusion and this age indicates that the mineralizing event lasted at least 2.7 and a maximum of 3.6 million years.

An important difference between Money Knob and other deposits in the Tintina Gold Province is the fact that the dike complex at Money Knob is distinctly alkaline with an intermediate composition, while moderately alkalic intrusions such as monzogranite, monzonite, and locally mozodiorite are more common elsewhere in the Tintina Gold Province (Hart and Goldfarb, 2005). The main K-feldspar minerals found both as phenocrysts and microlites in the groundmass in the biotite syenite are sanidine and microcline. However, like many other intrusions in the Tintina Gold Province, biotite is the dominant mafic phase with little to no amphibole or pyroxene being present. In comparison with the tectonic reconstruction of Pflaker and Berg (1994) the 92.04±0.14 Ma age of the biotite syenite dikes suggests that they were emplaced late in the orogenic cycle and possibly during the regional transition to strike-slip faulting along the western margin of North America. The alkalic nature of the igneous rocks possibly reflects back arc extension, which followed calc-alkaline arc magmatism (Newberry, 2000). It is possible that emplacement continued during the onset of dextral movement along the Tintina Fault in the region. The abundance of biotite suggests that the parental magma was not completely water-deficient and may have been a result of the last stages of dehydration of a subducted slab which began to ‘rollback’, allowing hot mantel material to upwell in the region (Bonin, 1990; Goldfarb et al., 1998).

### 8.2 Reconstruction of Fluid Evolution

The biotite syenite at Money Knob has been affected by subsolidus alteration and mineralization. Frequent quartz veins occur in the intrusion and are surrounded by alteration halos. Hydrothermal alteration was not intense and pervasive enough to completely destroy primary textures in the biotite syenite, making the dikes an ideal rock type to study alteration-induced textural and chemical changes.
The biotite syenite is cross-cut by quartz veins that consist of at least four types of quartz (Q1, Q2, Q3, and Q4). Two of these quartz types occur abundantly (Q1 and Q2) while late-stage quartz has only been observed in veins having open vugs (Q3 and Q4). The different quartz types can be linked to the development of different styles of hydrothermal alteration in a progressively cooling hydrothermal environment. The coarse euhedral quartz Q1 recognized in the quartz veins formed at the highest temperatures. Microthermometric evidence revealed a homogenization temperature of 305-310°C. Along the vein margins, the quartz is intergrown with biotite. This confirms that quartz Q1 formed at relatively high temperatures. In modern geothermal systems, secondary biotite formation is observed at temperatures of at least 270-350°C (Reyes, 1990). Quartz Q1 is typified by a short-lived bluish cathodoluminescence color, which is a common characteristic of hydrothermal quartz (Götze et al., 2001). The different types of zoning observed, including oscillatory growth zoning and zoning that may be related to the deformation of quartz Q1, imply that quartz formation occurred in a tectonically active environment. Fluid inclusions contained in quartz Q1 are carbonic. They contain a minimum of 10 mol% CO₂. Based on rough volumetric estimates, the CO₂ content could be >15 mol%. Entrapment as a single phase fluid suggests that pressures exceeding 1 kbars prevailed during quartz formation (Gehrig, 1980; Fig. 8.1).

High-temperature hydrothermal alteration occurred proximal to veins containing quartz Q1 and early in the evolution of the progressively cooling hydrothermal system. Within the hydrothermal alteration halos around the veins, a progressive outward zonation can be observed from biotite to sericite to albite alteration. This progression is likely related to the cooling and increasing reaction of the hydrothermal fluids with the biotite syenite during outward percolation away from the fluid pathways, represented by the veins (Fig. 8.2). This pattern is also reflected in the geochemical data where muscovite alteration closest to the veins have the highest K gains, and areas farther from veins overprinted by albite alteration have the strongest gains in Na and Si. As a result of the muscovite alteration, primary or secondary biotite are partially to completely replaced by muscovite, while secondary albite infiltrates and replaces previously
Figure 8.1: Pressure–temperature diagram of the H$_2$O-NaCl-CO$_2$ model system (from Gehrig, 1980), illustrating phase boundaries for different mol% CO$_2$ content in a H$_2$O+6.0 mol% NaCl fluid. Point “A” represents 2.2 mol%, the minimum amount of CO$_2$ contained in the system, at ~310°C. A minimum pressure of 0.3 kbars, which equates to ~2 kilometers depth at lithostatic conditions, is required to avoid phase separation. The dashed line extending from point “A” shows the possible path of the fluid through cooling and the “X” marks the approximate temperature (~270°C) where phase separation in the model system would occur. Point “B” represents the location of the phase boundary at 15 mol% CO$_2$ and ~310°C. The pressure needed to avoid phase separation would be ~1.7 kbars, which equates to ~6 kilometers depth.
Figure 8.2: Time and temperature diagram of vein and alteration paragenesis. Secondary biotite forms early in a high-temperature alteration environments and is subsequently overprinted by sericite and albite alteration. Carbonate and clay alteration occur later in the paragenetic sequence. The diagram illustrates that the recorded alteration occurs in a general retrograde environment. The diagram is constructed for a fixed location is space.
developed muscovite during albite alteration. In many cases, selvages around veins dominantly containing Q1 are primarily albite- and/or carbonate-altered because minerals previously developed during biotite and muscovite alteration have been totally replaced (Fig. 8.2).

The early quartz type Q1 is overprinted by quartz Q2. While some Q2 forms euhedral to subhedral quartz grains that show oscillatory growth zoning, Q2 also occurs along wispy microfractures transecting the earlier quartz Q1. The morphology of the fluid inclusions, and the presence of CO$_2$ found in the wispy trails of Q2 suggest formation from a comparably cooler hydrothermal fluid (<200°C). Pressure conditions do not appear to have changed between formation of quartz Q1 and Q2. The petrographic properties of the inclusions resemble those of orogenic gold deposits formed at comparably high lithostatic pressure conditions (Bodnar et al., 1985; Goldfarb et al., 1998; Reynolds, pers. commun.). The presence of kaolinite and smectite associated with quartz Q2 suggests that the maximum temperature of the hydrothermal fluids could not have exceeded approximately 180°C. Kaolinite transforms to pyrophyllite at temperatures of approximately 280°C (Sverjensky et al., 1991). However, smectite is only stable at temperatures below approximately 180°C. At those temperatures, trioctahedral smectite transforms to a chlorite/smectite mixed-layered phase or chlorite. Dioctahedral smectite transforms to illite/smectite or illite at even lower temperatures (Reyes, 1990; Monecke et al., 2007).

Alteration halos associated with quartz veins containing abundant Q2 quartz show losses in all the major elements contained in silicate phases. The low-temperature alteration associated with quartz Q2 was caused by hydrothermal fluids of variable acidity. The presence of smectite suggests that alteration occurred in a relatively neutral environment while kaolinite forms under moderately acidic conditions. This acidity was probably related to the presence of significant amounts of CO$_2$ in the hydrothermal fluids, which becomes reactive at lower temperatures (Giggenbach, 1992; Bischoff and Rosenbauer, 1996). H$_2$O and CO$_2$ react at temperatures between 150-200°C to form H$_2$CO$_3$, which subsequently dissociates to form H$^+$ and HCO$_3^-$ and then reacts with the wall rock to form carbonate minerals such as ankerite and clays stable in
moderately acidic environments such as kaolinite. At temperatures exceeding approximately 300°C, the reaction of \( \text{H}_2\text{O} \) and \( \text{CO}_2 \) is limited and does not cause significant wall rock alteration (Bischoff and Rosenbauer, 1996).

The third type of quartz is indicative of even lower temperatures and occurs only as an overgrowth of Q1. The forth type of quartz Q4 may initially have been formed as chalcedony. The two late-stage types of quartz are uncommon and are not notably associated with alteration of the biotite syenite, presumably because of the low temperature of formation and previous alteration of the biotite syenite in an overall retrograde environment.

The petrographic investigations did not provide unequivocal evidence for the relationship between quartz formation and gold deposition. Gold is hosted by unhealed fractures transecting the high-temperature quartz Q1, clearly indicating that the gold did not form synchronously with Q1. The possible textural association with strongly necked inclusions suggests that the gold was formed as part of the lower temperature evolution of the hydrothermal system. The geochemical analyses suggest that the bulk of the gold was introduced during albite alteration.

8.3 Comparison to Other Deposits of the Tintina Gold Province

Similar to other gold deposits in the Tintina Gold Province, the Money Knob deposit is spatially and temporally associated with a Cretaceous moderately reduced intrusive complex emplaced in an active tectonic environment. Based on the absolute age and location, the intrusive complex at Money Knob forms part of the Tombstone Plutonic Suite, which is a narrow 550 kilometer long belt in central Yukon. The belt is offset by ~400-450 kilometers along the Tintina Fault and continues in the Fairbanks district (Dover, 1994; Hart et al., 2004). The Tombstone Plutonic Suite host numerous gold deposits including most notably Brewery Creek (Diment and Craig, 1999), Dublin Gulch (Maloof et al., 2001), Clear Creek (Marsh et al., 2001), and Scheelite Dome (Mair et al., 2006) in the Selwyn Basin in the Yukon, and Fort Knox (Bakke, 1995), True North (Harrison and Gorton, 1998), and Ryan Load (Hart et al., 2002) in the Fairbanks district in Alaska. Beyond the Tombstone Plutonic Suite, other deposits like Donlin Creek (Goldfarb et al.,
2004), Shotgun (Rombach and Newberry, 2001), and Nixon Fork (Bundtzen and Miller, 1997) in the Kuskokwim district, and Pogo in the Goodpaster district (Smith et al., 1999) are found in the Tintina Gold Province, and all have been classified as intrusion-related gold deposits.

The Money Knob deposit shares structural similarities with many deposits found in the Tintina Gold Province, including a structural control on dike emplacement and the general consistency of mineralized vein orientations. Lateral movement along the Tintina fault system, and the Denali-Farewell fault system to the south, resulted in new and reactivated northeast trending faults found in many of the deposits of central Alaska, including Money Knob. The Money Knob deposit shares a general east-northeast to east-northwest strike orientation of faults and sets of sheeted mineralized veins with other deposits hosted by intrusions of the Tombstone Plutonic Suite such as Brewery Creek, Dublin Gulch, Clear Creek, and Scheelite Dome (Stephens et al., 2004). However much of the detailed structural similarities are obscured by later lateral offset along the Tintina Fault system.

On the Cordillera scale, the tectonic setting for Money Knob and similar deposits found in the Tombstone Plutonic Suite are generally part of the autochthonous or para-allochthonous miogeoclinal rocks formed in the late-Proterozoic or Paleozoic (Dover, 1994; Goldfarb et al., 1998; Monger and Price, 2002; Nelson and Colpron, 2007). The host rocks of the Tombstone Plutonic Suite which were affected by tectonic activity, magmatism, and mineralization, are found on the craton-ward side of the most landward crustal-scale suture created during the formation of Alaska (Plafker and Berg, 1994; Goldfarb et al., 1998), suggesting that igneous activity and associated mineralization occurred further inboard than typical arc-related processes.

Magmatic activity in the Tintina Gold Province is associated with approximately 25 different plutonic suites and belts, ranging from 140 to 74 Ma in age. Each suite or belt is of different metallogenic importance. However, the most volumetric and diverse magmatic activity occurred between 118 and 90 Ma. The highest gold endowment in the Tintina Gold Province is associated with intrusions of this age (Moll-Stalcup, 1994; Hart et al., 2004).

As noted above, an important difference between the Money Knob deposit and many
deposits found in the Tintina Gold Province is the strongly alkalic signature of the intrusive rocks, as opposed to the more common calc-alkaline related magmas occurring elsewhere in the Tintina Gold Province (Hart et al., 2002). The Tombstone Plutonic Suite is a metallogenetically important suite, and is composed of approximately twenty compositionally zoned monzonite, monzodiorite, and lesser syenite, quartz monzonite, and monzogabbro phases (Anderson, 1987, 1988; Hart et al., 2004). Dated intrusive bodies have an age range between 92 and 87 Ma (Anderson, 1987; Reifenstuhl et al., 1997). As a group, the Tombstone Plutonic Suite has been designated as a magnetite-series suite of rocks, although individual bodies can vary in terms of \( \frac{Fe_2O_3}{FeO} \) ratios and magnetic susceptibility values. In the case of Money Knob no magnetite was observed in the intrusive body and magnetic susceptibility reading are low on average (<0.1 SI \( 10^{-3} \) units).

In the Fort Knox district to the south of Money Knob, deposits of the Tombstone Plutonic Suite are associated with granites and granodiorites characterized by I-type, ilmenite-series, calc-alkaline plutons and are considered to be arc-related in origin and emplaced at around 91.4 Ma (McCoy et al., 1997; Hart et al., 2002). In the Yukon extension of the Tombstone Plutonic Suite, deposits are found with small- (<5 km\(^2\)) to medium-sized (~100 km\(^2\)) plutons and stocks that are medium-grained to porphyritic, hornblende-biotite granodiorite and quartz monzonite, and rare quartz-poor alkalic rocks (analogous to the biotite syenite at Money Knob) and mafic-rich to quartz-rich, two-mica granitoids (Hart et al., 2002). They are classified as metaluminous, calc-alkaline, I-type, and possibly arc-related plutons. The intrusions have a low primary oxidation state and high initial Sr ratios. This indicates that the parental magmas have a crustal component (Hart et al., 2002, 2004).

In the Kuskokwim district, which does not form part of the Tombstone Plutonic Suite, gold deposits are associated with younger (77-58 Ma) and more felsic intrusive bodies. At the Donlin Creek deposit, mineralization is associated with a hypabyssal porphyritic rhyolitic to rhyodacitic dike swarm (Goldfarb et al., 2004) while the Shotgun deposit is hosted by a granite porphyry stock (Rombach and Newberry, 2001). In the Yukon-Tanana Uplands Plutonic suite, older (109-
Figure 8.3: Schematic diagram showing metal associations and distance from causative pluton for different intrusion-related gold deposit types (modified after Lang and Baker, 2001)
102 Ma) granite and granodiorite with lesser quartz monzonites host subeconomic and poorly studied gold occurrences along with the world-class Pogo deposit (Hart et al., 2004).

The deposits of the Tintina Gold Province show variable metal associations, which include Au, As, Bi, Sb, Sn, Te, W to Ag, Pb, and Zn (Lang and Baker, 2001; Hart et al., 2002; Groves et al., 2003). In general, ore assemblages vary with proximity to the related intrusive bodies. Ore hosted in or directly adjacent to the plutons or stocks show a general enrichment in Au, Bi, Te, W, or Sn (Clear Creek, Dublin Gulch, Fort Knox, Scheelite Dome; Baker, 2002; Hart et al., 2002). Deposits that are hosted in sedimentary rocks distal to a related intrusive body commonly show enrichment in Au, As, Sb, and sometimes Hg (Brewery Creek, Donlin Creek; Baker, 2002; Hart et al., 2002).

The ores at Money Knob deposit show an element enrichment of Au, As, and Sb. Although the Hg content can be elevated, it is not clear whether Hg enrichment is related to gold mineralization. The association between Au and As is related to the fact that native gold grains commonly occur in association with arsenopyrite while gold and stibnite are less common. Unlike deposits such as Donlin Creek (Goldfarb et al., 2004) and Brewery Creek (Diment and Craig, 1999), gold is not refractory at Money Knob. The Money Knob deposit is largely sedimentary-hosted, which is consistent with the pronounced enrichment in As (Lang and Baker, 2001). It is possible that the deposit is located distal to the causative intrusions. Although there is a clear spatial and temporal association with the biotite syenite at Money Knob, this intrusion could not have been the source of the hydrothermal fluids as it has been affected by subsolidus alteration and mineralization. However, it is likely that the biotite syenite forms part of a much larger and deeper igneous system.

Previous fluid inclusion studies have constrained the nature of the hydrothermal fluids responsible for the formation of intrusion-related gold deposits in the Tintina Gold Province. The most prevalent fluid inclusion types encountered in deposits of the Tintina Gold Province are low-salinity and CO$_2$-rich fluids. This type of fluid appears to have been involved in the ore formation at deposits like Emerald Lake, Dublin Gulch, Scheelite Dome, Mike Lake, and
Mactung in the Tombstone Plutonic Suite (Baker and Lang, 2001); Shotgun (Rombach and Newberry, 2001) and Donlin Creek (Goldfarb et al., 2004) in the Kuskokwim district; and Fort Knox in the Fairbanks district (McCoy et al., 1997). Fluid inclusions of the low-salinity CO$_2$-rich fluids contain variable, but typically minor amounts of CH$_4$, N$_2$, and H$_2$S. Less common, and apparently always post-dating entrapment of the low-salinity CO$_2$-rich fluid inclusions, are liquid-rich H$_2$O inclusions with moderate to high salinities. In some cases, these late fluid inclusions contain halite (Baker and Lang, 2001).

The study of fluid inclusions allows a distinction between deposits formed at different crustal depths. Deposits formed at deeper crustal levels ($\geq$3.5 kilometers) typically only show fluid inclusion assemblages characterized by abundant low-salinity CO$_2$-rich aqueous fluids (Baker, 2002). In deposits formed at more shallow ($\leq$2 kilometers) crustal levels, these low-salinity CO$_2$-rich aqueous fluids have experienced phase separation, producing hypersaline liquid inclusions (>30 wt% NaCl equiv.) that coexist with low-salinity CO$_2$-bearing vapor inclusions.

Fluid inclusions found at the Money Knob deposit have low salinities and are CO$_2$-rich. The homogenization temperatures of the fluid inclusions at Money Knob are comparable to those measured in other deep deposits like Fort Knox (McCoy et al., 1997), Dublin Gulch (Maloof et al., 2001), and Scheelite Dome (Mair et al., 2006), where temperatures range from approximately 140 to 355$^\circ$C (Baker, 2002). Trapping pressures for deposits with low salinity CO$_2$-rich inclusions have been estimated to have been in excess of 1 kbar (Baker and Lang, 2001). As the amounts of CO$_2$ contained in fluid inclusions at Money Knob could not be determined quantitatively, exact pressures for entrapment of the fluids cannot be determined. However, based on visual volumetric estimates and comparisons with similar fluid inclusion assemblages from other deposits, it is not unreasonable to expect the same pressure conditions for the Money Knob deposit. In contrast, deposits formed at more shallow crustal levels like Donlin Creek, Brewery Creek, and Shotgun have homogenization temperatures ranging from 300 to 500$^\circ$C and entrapment pressures of <0.5 kbars (Baker, 2002). Mineralization at Money Knob formed distal to the related intrusion in a relatively deep setting and at low temperatures.
The source of fluids forming intrusion-related gold deposits in the Tintina Gold Province remains controversial. Most workers agree that the fluids are either of magmatic origin or are sourced from dehydration processes occurring in the deep crust, which is the orogenic gold model (Goldfarb et al., 1998; Thompson et al., 1999; Lang et al., 2000; Baker and Lang, 2001; Baker, 2002; Groves et al., 2003). The controversy centers on the question whether gold deposit formation is truly genetically linked to igneous activity as suggested by the close spatial and temporal relationship between intrusions and gold deposits, or if this relationship between intrusions and gold deposits is simply a reflection of larger scale processes occurring in the deeper crustal environment without the intrusion being the direct source of the hydrothermal fluids.

In the magmatic model, the presence of CO$_2$ in the hydrothermal fluids plays only an indirect role in deposit formation although it would influence the exsolution behavior of the fluids from the magma and impact its evolution during fluid decompression and cooling. Exolving CO$_2$ from the generative magma early in the fluid development, which begins at great depths, allows for an increase in relative abundances of volatiles in the magma. Later exsolusion of fluids from the magma with an increased relative abundance of volatiles creates a fluid rich in ore-forming elements. Ore-forming fluids can be generated over a wide range of depths as the magma ascends from its generative depths (Thompson et al., 1999; Lang and Baker, 2001; Baker, 2002). This model accounts for the variable fluid compositions and depths of formation of intrusion-related gold deposits.

The contrasting orogenic-gold model suggests that the ore-forming fluids were derived from dehydration during regional continental margin deformation caused by subduction and/or continental collision (Groves et al., 1998, 2003). Ore metals and fluids would be sourced from subducted crust, supracrustal rocks, or deep granitoids during the collisional processes. Mineralization occurs at higher crustal levels due to fluid ascent along regional- and district-scale fault zones. The orogenic model accounts for the strong structural controls found in many intrusion-related gold deposits and the temporal association between lamprophyre dike
emplacement and gold mineralization.

Although this study does not provide unequivocal evidence for the origin of the hydrothermal fluids, it is important to note that the available age dates suggest that the hydrothermal system at Money Knob was long lived. Deposit formation over a period of 2.7 to 3.6 Ma is inconsistent with the fluids being derived from a single intrusion. However, the association with the dike complex clearly suggests that gold mineralization was associated in both time and space with igneous activity.

8.4 Exploration Implications

In the past, bulk tonnage gold deposits have not been considered economically feasible and little exploration has been carried out for these low-grade deposits. However, with the recent increases in gold prices and improved recovery processes, it may now be feasible to recover gold from these deposits. Money Knob has a measured, indicated, and inferred resource of 20.6 million ounces of gold at an average grade of 0.52 g/t and a cut-off grade of 0.22 g/t. The deposit is currently advancing towards a feasibility study. Two similarly large, bulk tonnage deposits that are currently developed in North America include Osisko’s Canadian Malartic deposit (a proven and probable reserves of 10.71 million ounces of gold at a grade of 0.97 g/t and a cut-off of 0.32 g/t) and Detour Gold’s Detour Lake deposit (a proven and probable reserve of 15.6 million ounces Au at an average grade of 1.03 g/t and a cut-off of 0.5 g/t). These examples demonstrate that the bulk tonnage model today is a viable option for exploration targeting.

The geological relationships at Money Knob suggest a close spatial and temporal association between intrusive activity and gold mineralization. The results of the present study show that the biotite syenite at Money Knob has an age of approximately 92 Ma and appears to be part of the mid-Cretaceous Tombstone Plutonic Suite. The intrusions of the Tombstone Plutonic Suite are calc-alkaline to alkalic, and mostly intermediate to felsic, with calc-alkaline lamprophyres also common in many locations. The Tombstone Plutonic Suite is known for its mineral endowment. Discovery of the Money Knob deposit suggests that intrusions of the Tombstone Plutonic Suite
may be hosts to bulk tonnage gold deposits. Targeting plutons and dike-sill complexes with the same petrographic characteristics and age as the one at Money Knob may, therefore, lead to other discoveries in the district.

The present study suggests that vectoring towards intrusion-related bulk tonnage gold deposits like the Money Knob deposit could be effectively based on a combination of geochemical, mineralogical, and geological criteria. The quartz veins at Money Knob show a distinct enrichment in minerals such as arsenopyrite and stibnite. Paragenetically, the formation of arsenopyrite appears to be closely associated with gold mineralization. Due to the carbonic nature of the mineralizing fluids, carbonate alteration is widespread in the general deposit area, although carbonate alteration appears to have formed at lower temperatures than the gold deposition. Anomalous enrichment of As and Sb in soil or rock samples, coupled with widespread carbonate alteration and the presence of disseminated sulfides in and around mid-Cretaceous felsic intrusive rocks should, therefore, be considered as generally favorable indications during regional exploration. Further constraints could be derived from mapping of alteration mineral associations on the regional to district scales and documentation of alteration mineral associations in core. At Money Knob, the petrographic evidence suggests that gold is associated with strong albite alteration and not necessarily with the more commonly targeted biotite (potassic) or muscovite (phyllic) alteration. Based on the results of the present study, it appears likely that albite alteration could be used as a vector towards mineralization.

The discovery of different quartz types within quartz veins at Money Know has significant exploration implications. The present study demonstrates that the earliest quartz at Money Knob has very distinct cathodoluminescence properties, including the short-lived CL color and the complex zoning patterns. This generation of quartz represents the host of the gold that formed along fractures within the quartz. The presence of this early quartz is considered to be a favorable indication, even if no gold is present in a particular sample. If quartz in other intrusion-related gold deposits has similar cathodoluminescence properties, this analytical technique could be used routinely to distinguish between vein quartz related to an intrusion and other
types of quartz veins. Cathodoluminescence investigations on polished thin sections obtained from vein quartz can be conducted at comparably low cost and fast turn-around times, which would be a prerequisite for its use in mineral exploration. On a district or even a regional scale, cathodoluminescence investigations could potentially even be used to fingerprint hydrothermal quartz in stream sediments. Combined with fluid inclusion petrography, allowing identification of inclusions of carbonic fluids and the study of vein paragenesis, a novel technique for exploration for intrusion-centered deposits could be devised.
CHAPTER NINE:
CONCLUSIONS

The principal aims of the present study on the Money Knob deposit were to (1) determine the nature of the hydrothermal fluids forming the deposit, allowing development of a better deposit model, (2) provide further constraints on the absolute age of mineralization, (3) identify different styles of alteration and constraining their relationships to the ore-forming event, and (4) document and interpret the mineralogical and textural changes that were associated with the development of the different styles of hydrothermal alteration. The following sections summarize key observations and interpretations that are based on the combination of field and laboratory investigations carried out. Recommendations for future work are made that could further improve the understanding of the geological setting and processes involved in the formation of the Money Knob deposit.

9.1 Summary of Findings:

The results of previous research combined with the findings of this study allow the following key conclusions.

1) The Money Knob gold deposit is a low-grade, high-tonnage gold deposit (measured, indicated, and inferred resource of 20.6 million ounce at an average grade of 0.52 g/t with a cut-off grade of 0.22 g/t) located in the Tintina Gold Province of Alaska. The majority of the ore is hosted in a Devonian package of sedimentary and volcanic rocks that is structurally overlain by Cambrian mafic and ultramafic rocks. However, gold mineralization and associated hydrothermal alteration appear to be spatially and temporally associated with the intrusion of a felsic intrusive suite.
2) The Money Knob deposit is located within the Livengood Terrane, which is composed of miogeoclinal rocks that formed in a shelf and slope environment along the ancient North American continent. The Livengood Terrane hosting the deposit is bounded to the north by major splays of the Tintina Fault and to the south by the highly deformed Manley Terrane. The Devonian package of sedimentary and volcanic rocks and the overlying Cambrian mafic and ultramafic rocks hosting the deposit likely initially formed as part of the Selwyn Basin and have been subsequently offset 400-450 kilometers laterally along the Tintina Fault (cf. Dover, 1994; Plafker and Berg, 1994; Hart et al., 2002). This hypothesis is strengthened by recent geochemical investigations by Hart et al. (2004) on the intrusions of the Tombstone Plutonic Suite in the Selwyn Basin and the Livengood area.

3) The felsic intrusive suite interpreted to be associated with the formation of the Money Knob deposit is composed of several distinct igneous phases. Volumetrically most important are biotite syenite dikes that have been the focus of the present study. U/Pb age dating on zircon from the biotite syenite conducted as part of the present study yielded an age of 92.04±0.14 Ma. This age is interpreted to be the intrusive age of the biotite syenite and is slightly older than a previously obtained ⁴⁰Ar/³⁹Ar age of hydrothermal muscovite of 88.9 ± 0.3 Ma (Athey et al., 2004b). Since the biotite syenite cuts early stages of mineralization, the new age dates suggest that the mineralizing event and associated hydrothermal alteration occurred over a period of approximately 2.7 to 3.6 million years. The relative timing for the emplacement of the biotite syenite corresponds to post- or late syn-compression in the deposit area. Offsetting relationships suggest dike emplacement continued during tectonic relaxation along reactivated thrust faults developed during fold-and-thrust activity.

4) Within the biotite syenite, gold is hosted by quartz veins and/or occurs as fine-grained disseminated Au. Gold is most commonly associated with arsenopyrite and stibnite. The correlation coefficient between Au and As is ~0.89, but is statistically insignificant for Sb (Pontius et al., 2010). Au occurs as fine-grained native Au grains that are intergrown with sulfide
grains or forms grains that are located along fractures within the quartz veins.

5) Cathodoluminescence microscopy on quartz veins hosted by the biotite syenite revealed the presence of at least four distinct quartz types (Q1 to Q4). Fluid inclusion petrography and microthermometry on the different quartz types suggest that quartz vein formation occurred from a generally cooling hydrothermal system. Early Q1 quartz formed at temperatures of at least 300-350°C at a pressure of >0.5-1 kbars. Gold is present along fractures in this early type of quartz. Subsequent Q2 quartz formed at distinctly lower temperatures (155-165°C). Based on fluid inclusion petrography, the late Q3 and Q4 also formed at low temperatures. The hydrothermal fluids forming all quartz types were distinctly enriched in CO₂.

6) The quartz veins hosted by the biotite syenite are surrounded by distinct alteration halos. The mineralogical composition of the alteration halos and replacement relationships between secondary minerals point to a general decrease in temperature over time, confirming the results of the fluid inclusion investigations. The earliest hydrothermal alteration associated with quartz veins containing only Q1 occurred at comparably high temperatures (>300°C) and resulted in the formation of secondary biotite. This was followed by sericite alteration and subsequent albite alteration. Quartz veins associated with albite alteration show the highest gold grades. These alteration styles are overprinted by low temperature (<180°C) clay alteration. Quartz veins containing halos of clay alteration contain the quartz types Q2 to Q4 that are interpreted to have formed at low temperatures.

7) Geochemical investigations on the alteration halos surrounding the veins showed that the different styles of alteration resulted in different geochemical signatures. As expected from the petrographic analysis, gold enrichment is associated with gains in Na and to a lesser extent in Ca and Si. The lowest gold grades occur in samples showing intense low-temperature clay alteration.

8) The results of the present study demonstrate that the Money Knob deposit formed
in a comparably deep crustal environment (≥ 3 kilometers at lithostatic conditions) from hydrothermal fluids that progressively cooled from over 300-350°C to less than 155-165°C. The hydrothermal fluids were distinctly carbonic, but not hypersaline in character. Field relationships suggest that mineralization commenced prior to the intrusion of biotite syenite that was dated at 92.04±0.14 Ma by U/Pb geochronology on zircons. Since the biotite syenite cuts early stages of mineralization, the new age dates suggest that the mineralizing event and associated hydrothermal alteration occurred over a period of approximately 2.7 to 3.6 million years, which is inconsistent with a single inclusion being the source of the hydrothermal fluids. However, field relationships suggest that there is a spatial and temporal relationship between multi-phase igneous activity at Money Knob and gold deposit formation, possibly suggesting a genetic link. This inferred genetic link along with the reconstructed conditions of deposit formation are consistent with the Money Knob deposit being a distally disseminated, deep intrusion-related gold deposit.

### 9.2 Recommendations for Future Work

1) Future studies should focus in more detail on the nature of the hydrothermal fluids and their evolution during deposit formation. Such a study should integrate CL microscopy, fluid inclusion petrography, and microthermometry. Studying the evolution of the fluids during the mineralizing processes may provide more insight into the mechanisms for gold transport and the conditions under which gold was precipitated.

2) Potential links between the structural characteristics of different vein sets and the different types of quartz should be clarified. This relationship may offer a greater understanding for the formation of different quartz types through the structural development of the deposits while yielding more detail about the conditions under which gold was deposited.

3) As a next logic step, quartz veins and disseminated gold hosted by the Devonian package
of sedimentary and volcanic rocks should be studied to test whether the same types of quartz veins and alteration styles occur in these host rocks. Special emphasis should be placed on the study of ore zones that appear to be crosscut by the biotite syenite intrusive rocks to clarify under which conditions gold deposition occurred prior to dike emplacement.

4) It is recommended to conduct a more detailed study on the vein accessory phases. The results of the present study and company geochemical data suggest that there is a close association between gold and arsenopyrite. Establishing this link for structurally different quartz veins and veins hosted by the different host rocks could have significant exploration implications as arsenopyrite can be readily recognized in core. Future studies on sulfide minerals contained in the quartz veins should also include sulfur isotopic investigations to better constrain the source of sulfur.

5) It is proposed here to conduct a regional survey of the cathodoluminescence properties of quartz veins in the Tintina Gold Province. At Money Knob, quartz veins associated with the gold event, whether they are mineralized or not, could be easily identified based on their short-lived blue cathodoluminescence colors and complex zoning patterns. If a more comprehensive study would establish that this cathodoluminescence signal is characteristic of intrusion-related gold deposits, this technique could be easily used as a tool in mineral exploration.
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### Appendix A

**Supplemental Electronic Files**

The supplemental electronic files provide the data that was used for the preceding thesis. The files are organized in the order in which they are encountered in the thesis.

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