**Petrography, Geochemistry and Isotopic Analysis of Paleogene Volcanism in the
Fish Creek Mountains, Great Basin, North-Central Nevada**

by

Christopher Stevens

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Carleton University
Ottawa, Ontario

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Christopher Stevens

ABSTRACT

The Fish Creek Mountains, located in north-central Nevada, is a site of multiple igneous events ranging from 35Ma to 1Ma, covering most of the igneous history of the Great Basin where my goal of this project is to investigate the Paleogene volcanism within the FCM. Samples collected from the FCM and surrounding Tobin and Shoshone Ranges are Eocene-Oligocene in age (33-34 Ma), calc-alkaline basaltic andesites through to rhyolites. Incompatible elements and isotopic data suggest continental margin subduction-like trace element signatures with highly radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$, consistent with an old, metasomatized lithospheric mantle source. Compared to other areas of the Ancestral Cascades and Western Great Basin, elevated $^{87}\text{Sr}/^{86}\text{Sr}$ found in the FCM lavas are herein interpreted to be the result of a purely lithospheric component, suggesting a thicker crust in the central Great Basin. These intermediate, mantle-derived magmas ascended and mixed with partial melts of the mafic crust to form andesites and dacite.

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LIST OF SYMBOLS, NOMENCLATURE OR ABBREVIATIONS

FCM = Fish Creek Mountains

CAB = Calc-alkaline basalts

OIB = Ocean Island Basalt

MORB = Mid-Ocean Ridge Basalt

LREE = Light Rare Earth Elements

MREE = Middle Rare Earth Elements

HREE = Heavy Rare Earth Elements

HFSE = High Field Strength Elements (Zr, Hf, Nb, Ta)

LILE = Large Ion Lithophile Elements (K, Rb, Cs, Ba)

Chapter 1: Introduction

1.1 Introduction:

The Great Basin of Western United States is a region of Cenozoic lithospheric extension and volcanism that includes the state of Nevada and parts of southeastern California and western Utah. The goal of this thesis is to investigate Paleogene volcanism in the Great Basin focusing on the Fish Creek Mountains (FCM) and surrounding area. The FCM, located in north-central Nevada, is a site of multiple igneous events ranging from 35 Ma to 1 Ma, encompassing most of the igneous history of the Great Basin. The mountains rise up to 6,512 feet above sea level and lie south of the Battle Mountain mining district, Nevada (Figure 1). The area of study is primarily composed of late Paleogene volcanoclastic and volcanic rocks that overly Paleozoic metasedimentary rocks, that also cover a large portion of the Sierra Nevada of northeastern California as well as western Nevada (Stewart and Carlson, 1976). These Paleogene intermediate to felsic lava flows are related to the westward sweep of volcanism through Nevada due to rollback of subduction of the ancient oceanic Farallon plate under North America. The late Paleogene mafic to felsic volcanic rocks that are found throughout the FCM have yet to be thoroughly examined to determine their origin. The suite of Paleogene volcanic samples were collected in the summer of 2010 along with felsic tuffaceous rocks and younger mafic volcanics as part of a larger transect across central Nevada to eastern California. Throughout this thesis I will present the results of field work, petrographic study, geochemical and isotopic analyses of the volcanic rocks and describe a tectonic model and petrologic processes responsible for their origin.

Figure 1: Location of Fish Creek Mountains study area (enclosed in square) and surrounding volcanic areas of the western United States (John, 2001). NNR = Northern Nevada Rift, CRB = Columbia River basalts, SM = Steens Mountains, M = McDermitt caldera. Dark grey area = Great Basin.

1.2 Justification or purpose for this study

Few detailed and modern studies have been done on the volcanology and geochemistry of late Paleogene (33-34 Ma) rhyolites to basalts in north-central Nevada. Ongoing chemical analyses are being done on younger basalts (16-10 Ma and 4-1 Ma) (Cousens) and the 24.9 Ma FCM rhyolitic tuff from the Fish Creek Mountains (Varve, 2013). An undergraduate thesis examined younger 4-1 Ma Pliocene to Quaternary flows and cinder cones exposed in the northwestern margin of the FCM (Wetmore, 2011). By studying these rocks, I will be able to document how much petrological and chemical variation exists between the samples collected and determine whether or not they are likely to be of similar age and volcanic origin. By doing so, I will have filled a portion of the gap in our understanding of the volcanic history and tectonic evolution of the Western United States.

The purposes of this study are to:

- 1) Focus primarily on Paleogene felsic rhyolites to mafic basaltic andesites to determine how and when they may have erupted within the Fish Creek Mountains. This combined with other projects from the area will help determine the volcanic history from California to Nevada in order to understand whether magmatic activity in this area is linked to slab rollback and extensional tectonics.
- 2) To determine the mineralogical, textural, geochemical and isotopic characteristics of the felsic to mafic Paleogene volcanic rocks within the Fish Creek Mountains and to determine what petrologic processes are responsible for their origin.

- 3) To determine the degree of crustal contamination in the more evolved felsic rocks.
- 4) To document Paleogene mantle sources and petrogenetic processes to reconstruct evolution of magma sources through time.
- 5) To determine the tectonic model which reflects the interplay of lithospheric extension and magma generation in the mantle (asthenosphere and lithosphere) and the crust.

1.3 Geology and tectonic setting:

1.3.1 Brief geochronology of the Great Basin

The Great Basin is a zone of Cenozoic lithospheric extension and volcanism that includes Nevada and parts of southeastern California and western Utah forming the widest segment of Basin and Range topography (Figure 2; Dickinson, 2006). The thinning of the lithosphere is a result of extension which created normal faults, expressed at the surface by sub-parallel horst and graben mountain ranges separated by down dropped blocks which form the valleys and basins (i.e. Basin and Range). These regional fault-bounded structural blocks are associated with extensive Paleogene volcanism and mineral deposits. The detachment of structural blocks involving pull apart and/or rotation during extension may have formed widespread “plumbing systems” and allowed for magmas and hydrothermal fluids to migrate through south-central Nevada (Kepper, 1989).

Figure 2: Position of the Great Basin in the Western Cordillera boxed in by the eastern Colorado Plateau, western Sierra Nevada (SN), northern Snake River Plane (SRP) and the southern Garlock fault (Gf) and the Mojave block (From Dickinson, 2006). CRP = Columbia River Plateau, BM = Blue Mountains and KM = Klamath Mountains, Great Basin Segment = Basin and Range topography. Black star represents study location.

The Great Basin preserves a diverse history ranging from Archean to recent. The oldest rocks include the mantle derived Archean Wyoming Province and the Paleoproterozoic Mojave Province which form part of the craton. Neoproterozoic rifting of the craton created basement faults that were reactivated to accommodate subsequent tectonism and act as conduits for later igneous intrusive and/or hydrothermal fluids (Cline et al., 2005). Rifting culminated in development of a west-facing passive margin at the edge of the Precambrian continental crust characterized by westward-thickening sequence of pre-mid Late Devonian sediments (John, 2001). In early Paleozoic, the overall the subsidence rate exceeded that of sedimentation and the margin of shallow water marine environments shifted progressively continent-ward (Stevens et al., 1997). Late Devonian to early Mississippian emplacement of the Roberts Mountains allochthon over the miogeoclinal sediments is the earliest Phanerozoic contractional event to have affected east-central California (Antler orogeny). Late Mississippian to Early Pennsylvanian faulting created a series of fault bounded uplifts throughout the Inyo Mountains and other areas of east-central California. The Late Mississippian faulting also led to the development of a NW-trending transform fault zone that truncated the continental margin (Dickinson, 2006). Permian to earliest Triassic contraction deformation in east-central California occurred as a result of convergent plate motion that previously truncated the continental margin (Stevens et al., 1997). During the early Triassic, sedimentation occurred during a time of relative tectonic dormancy. Middle Triassic contraction (Sonoma orogeny) of deformed oceanic facies was thrust over the Antler orogeny termed the Golconda thrust (Stevens et al., 1997). By early Late Triassic time, east-central California east-dipping subduction zone was established beneath the

continental margin in east-central California producing arc plutonism, volcanism and east-vergent contractional deformation, including the emplacement of the Sierra Nevada Batholith and the development of the East Sierran thrust system. This heightened period of magmatic activity may have been caused by increased seafloor spreading rates that produced a hotter, less dense oceanic plate that slid under the North American plate at an increasingly lower angle. Alternatively, the rate at which the upper plate overrides the lower plate could have also resulted in flat-slab subduction where the upper continental lithosphere in the Western United States was moving faster than the lower oceanic Farallon slab resulting in flat-slab subduction (Jarrard, 1986). As the younger Farallon Plate decreased its angle of descent, it came in contact with the underside of the continental lithosphere where heat and pressure of the underlying plate are thought to have created an easterly shift of volcanism and compressional stresses continuing throughout the rest of the Mesozoic and therefore thickening of the continent producing the Laramide orogeny (Figure 3A; Fiero, 1986). Magmatic front migrated east into Colorado and did not return to Nevada until Mid-Cenozoic time (John, 2001). As the velocity of the previously sub-horizontal subducting Farallon slab decreased, it allowed for more time to cool, contract and sink as a result of its higher density into the hot asthenosphere. This slab rollback is thought to have shifted volcanism in the eastern Colorado plateau in a southwesterly direction back towards the western margin of North America (Figure 3B; Fiero, 1986). The arrival of the Pacific plate replaced compression and subduction with shear stress and a transform plate boundary which will later become the San Andreas Fault where as the Farallon Plate continues to sink into the asthenosphere.

Figure 3: East-dipping continental margin subduction zone with associated periods of heightened magmatic activity and deformation related to the subduction of the Farallon plate under North America. A) Latest Mesozoic-earliest Cenozoic; top=80-70 Ma, middle=75-65 Ma, bottom=60-50 Ma. B) Middle to late Cenozoic; top=45-25 Ma, middle=20-15 Ma, bottom=5-0 Ma (Fiero, 1986; Zhang et al., 2009).

Cenozoic crustal extension and associated normal faulting formed first in northeastern Nevada and northwestern Utah as previous compressional stresses of the Mesozoic and Early Cenozoic are removed allowing the continent to spread laterally. The beginning of extensional faulting at any given latitude in the Basin and Range generally coincided with or immediately postdated voluminous eruptions of intermediate to silicic volcanic rocks (Best and Christiansen, 1991). The actively extended lower crust and mantle lithosphere, combined with crustal weakening by mantle upwelling in the wake of the subducted Farallon Plate, is thought to have produced widespread Basin and Range extension. Bartley et al (1988) found crustal extension to be episodic during four periods of time; prevolcanic before 32 Ma, early syn-volcanic (30-27 Ma), immediately post-volcanic (16-14 Ma), and post-post volcanic Pliocene to Quaternary. Typically mantle-derived basaltic volcanism was associated with faulting during periods of slow extension whereas volcanism ceased during periods of rapid extension. Such mantle upwelling also has a number of effects on the Basin and Range including: bimodal volcanism, decompression melting, thermal expansion and softening of the continental crust (John, 2001).

1.3.2 Great Basin igneous assemblages and volcanism

Great Basin Cenozoic igneous rocks can be divided into three assemblages: 1) interior andesite-rhyolite (Eocene to early Miocene), 2) western andesite (early Miocene to early Pliocene) and 3) bimodal basalt-rhyolite (middle Miocene to Holocene) (Figure 4; John, 2001).

1) The interior andesite-rhyolite assemblage mainly contains dacite to rhyolite ash flow tuffs and flow dome complexes with small volumes of andesitic and dacitic lava

flows. They are generally calc-alkaline; however much more potassic and silicic than typical arc-related magmas (John, 2001).

Eocene magmatic activity began at around 43 Ma with the eruption of silicic ash-flow tuffs and intermediate lava flows in northeast Nevada representative of the interior andesite-rhyolite assemblage (Figure 4A; John, 2001). By the Oligocene, magmatic activity of the assemblage migrated southwestward (27 Ma) from northeastern Nevada. The youngest rocks of this assemblage erupted at about 19 Ma. Caldera complexes are common largely in the Oligocene to early Miocene, and more than 50 calderas have been identified in central Nevada and western Utah. Plutonic rocks, primarily granodiorite, are exposed locally and include porphyry copper and skarn related intrusions in the Battle Mountain mining district, Nevada (John, 2001).

Heating during this period is related to the subduction of the spreading centre between the Farallon and Kula oceanic plates that produces a slab window allowing the upwelling of the asthenospheric mantle to impinge on the base of the lithosphere (Cline et al., 2005). Progressive rollback and separation of the Farallon plate from the base of the lithosphere was responsible for both the onset of extension and high potassium calc-alkaline magmatism in northern Nevada during the Late Eocene and eventually south-central Nevada during the Oligocene to Miocene that lead to the formation of basaltic melts partly mixed with lower crustal melts (Farmer et al., 2002). This assemblage is most similar to the mafic to felsic Paleogene samples collected in the FCM.

2) The western andesite assemblage is a highly potassic calc-alkaline assemblage composed mostly of intermediate lava flows, breccia and hypabyssal intrusions formed

along the northwestern edge of the Great Basin in western Nevada and eastern California (Figure 4B; John, 2001). Small rhyolite intrusions and volumetrically minor basalt are also widely distributed. This assemblage was part of the continental margin arc that was active in and west of the modern Cascade Range. It formed in response to the subduction of Farallon oceanic crust beneath the continental margin of North America during the mid- to late Miocene. Rocks of the Western andesite assemblage range in age from 22-4 Ma, the youngest of which are along the northwestern edge of the Great Basin (John, 2001).

3) The bimodal basalt rhyolite assemblage is the youngest rock assemblage in the Great Basin which include olivine basalt, pyroxene andesite and basaltic andesite, as well as both sub-alkaline and peralkaline rhyolite located east of the western andesite assemblage (Figure 4C; John, 2001). Volcanism began in the northern Great Basin at around 16.5 Ma and continues locally to the present day. Bimodal volcanism in the north-central Great Basin may have initially been concentrated at the north end of the Northern Nevada Rift. Volcanism from this rift lasted from 16.5 Ma to 15 Ma and extends 500km from the Oregon-Nevada border to south-central Nevada. Generally, intermediate rocks (siliceous andesite, dacite and trachydacite) are common in the central and northern parts of the Northern Nevada Rift but are generally uncommon everywhere else. The older phases of this assemblage formed in a back-arc environment related either to back-arc extension or to the encroachment of the Yellowstone mantle plume on the crust at 16.5 Ma along the Nevada-Oregon border. The younger phases formed during continental extension unrelated to subduction, perhaps as a result of lithospheric extension over a mantle plume (John, 2001).

Figure 4: Maps showing the general distribution of Cenozoic volcanic assemblages outlined in dark grey. A) Interior andesite-rhyolite assemblage showing 27 Ma timeline illustrating a southwestward sweep of magmatism. B) Western andesite assemblage. C) Bimodal basalt-rhyolite assemblage. Black stars represent study location. (Figure modified from John, 2001)

1.3.3 Great Basin Tectonics

The subduction of the oceanic Farallon plate and associated rift (the East Pacific Rise) played an important role in the formation of the Great Basin. Tectonics in the western United States can be broken up into two main time periods; (1) latest Mesozoic to earliest Cenozoic and (2) middle-late Cenozoic.

1) During the latest Mesozoic to earliest Cenozoic the creation of the subduction zone that consumed the oceanic Farallon plate lead to an easterly migration of arc volcanism to the Colorado Plateau.

By early Late Triassic time, east-central California was influenced by an east-dipping continental margin subduction zone. As a result, the Mesozoic was a period of heightened magmatic activity as the arc evolved from Late Triassic to Late Cretaceous time. Abundant Triassic (220-210 Ma) and larger Jurassic (152 Ma) plutons and mostly Cretaceous (148 Ma) major dyke swarms intruded the region, where as volcanic complexes (222-98 Ma) accumulated on the surface (Stevens et al., 1997). However, Sierra Nevada arc activity ended as a significant increase in spreading rates produced a hotter and therefore less dense oceanic plate that slid under the North American plate at an increasingly lower angle producing an easterly wave of volcanism until flat-slab subduction. Flat-slab subduction compressed the lithosphere and created the Laramide orogeny (Figure 3A; Fiero, 1986). Magmatic activity migrated east into Colorado and did not return to Nevada until Mid-Cenozoic time, possibly due to crustal thickening during the Late Cretaceous Sevier and Laramide orogenies associated with low-angle

subduction, accretion of island-arc terranes and progressive contraction of the miogeocline from west to east.

2) Middle to Late Cenozoic crustal extension dominated the tectonic history of the Great Basin during the development of several styles of faulting. Beginning in the late Eocene, stress relaxation and mild extension, characterized by multiple sets of normal faults and detachment faults in more deeply exposed terranes, affected much of the Great Basin whereas more extreme extension occurred in the Miocene. Generally, areas of extension formed first in northeastern Nevada and northwestern Utah. This rapid extensional period was followed by an increased angle of the subduction of the oceanic plate leading to a shift of volcanism from easterly back to a westerly direction. As the impact velocity of the previously sub-horizontal subducting Farallon slab decreased, it allowed for more time to cool, and sank as a result of higher density into the hot asthenosphere (slab rollback). This triggered melting of the overlying lithosphere due to previously released volatiles and caused magmatism at the surface (Best and Christiansen, 1991). As the plate continued to steepen this locus of volcanism would shift westerly to its most recent expression as late Paleogene volcanism in the Sierra Nevada in eastern California (Cousens et al., 2008). As the plate sank deeper it eventually broke off causing the upwelling of the asthenosphere (Figure 3B; Fiero, 1986)

At about 16 Ma, more widespread Basin and Range extension began, producing alternating basins and ranges spaced 20-50 km apart (John, 2001). The extension that created the Northern Nevada rift and continued to form the present day Basin and Range topography probably began in the mid-Miocene (Ressel and Henry, 2006). Local, rapid large magnitude extension continued after the initiation of Basin and Range faulting.

Recent estimates for the total amount of Cenozoic extension in the Great Basin are between 100 to 250 percent. More extensive Basin and Range extension occurred in the mid-Miocene in contrast to the Eocene when the North American continent came into greater contact with the Pacific plate creating the San Andreas Fault in between remnants of the subducted Farallon plate termed the Gorda and Cocos plates. This is thought to have removed compression between the Farallon plate and the continent, giving the continent room to spread (John, 2001).

1.4 Local geology (Nevada)

Paleogene igneous activity in Nevada has been suggested to be the result of magmatism due to the subduction of the Farallon Plate and associated spreading centre under North America (Best and Christiansen, 1991). Volcanism can be broken up into three periods on the basis of K-Ar dating (McKee and Silberman, 1970). At about 50 Ma, a magmatic front began migrating southwestward across southern Idaho, central Oregon and into northern Nevada and Utah. Intermediate, arc like basaltic andesites through dacites dominated volcanic activity in northeastern Nevada in between 45 to 36 Ma. From 34 to 24 Ma, the extrusions of ash flow sheets covered large parts of southern and eastern Nevada, termed the ignimbrite flare up (Fiero, 1986). After about 24 Ma magmatism changed gradually in composition and style of eruption. Mafic cinder cones, tuff cones, low shield volcanoes, isolated lava flows, and viscous rhyolitic domes became more characteristic and smaller calderas and ash-flow sheets became less common (Christiansen, 2010). The final period of Paleogene magmatism lasted from 16 to 10 Ma, including basalt and basaltic andesite flows along with intrusive rhyolite flow dome complexes found around the FCM and are related to the volcanism of the Snake River

plain province to the north (McKee and Silberman, 1970). After approximately 10 Ma, the magmatic rocks became increasingly bimodal, with the appearance of basalt and the disappearing of intermediate magmas. Volcanism then migrated westwards towards into Sierra Nevada as a result of progressive sinking and southwestward rollback of a shallowly dipping subducting slab. This resulted in widespread dehydration of the subducted lithosphere and generated voluminous mantle-derived magma which intruded and differentiated in the crust (Christiansen, 2010). The inflow of asthenospheric mantle beneath the Great Basin combined with the development of a transform boundary and regional extension resulted in decompression melting of asthenospheric and lithospheric mantle. A portion of this magma stagnated in the lower crust, then re-melted and differentiated to create subalkaline and peralkaline magmas of anorogenic affinity in contrast to the predominantly calc-alkaline affinity of the central Great Basin. These mafic lavas are typically alkali basalts (lack negative Nb-anomalies) and for the most part, initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios are lower where as $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratios are higher than the older arc magmatism (Christiansen, 2010).

Tuffaceous volcanics in central Nevada can be distinguished in three different units; the upper Bates Mountain tuff (23.1 +/- 1.7 Ma), the Caetano tuff (31.2-33.3 +/- 2.7 Ma) and the FCM tuff (24.9 Ma). The upper Bates Mountain tuff is concentrated northwestward with only a few remote outcrops recognized in the northern part of the FCM. The other two are welded tuffs that make up the Paleogene igneous section in north-central Nevada. The Caetano tuff is concentrated in a west-trending belt from the northernmost Toiyabe Range to the west edge of the central part of the Shoshone Range but also crops out at the northern end of the FCM. The youngest FCM tuff is exposed

throughout the FCM but the outflow sheet is limited to a few miles away from the FCM (McKee and Silberman, 1970).

Mesozoic and Paleogene intrusive suites are much less extensive and most common in the northern part of Nevada (McKee and Silberman, 1970). Based on K-Ar dating, the intrusive rocks range in age from Oligocene to Cretaceous (McKee and Silberman, 1970). Most of the intrusive suites are composed of fine-grained porphyritic quartz diorite to quartz monzonite. Granite Mountain (49-30 Ma) is the largest Paleogene intrusive body in the study area being more than twice to size of any others nearby. It is composed of medium to coarse grained hypidiomorphic granular rock similar to the large Mesozoic plutons of central Nevada (McKee and Silberman, 1970).

1.5 Components in continental subduction zones

The subduction of the oceanic Farallon plate and associated mid-ocean rift played an important role in the formation of the Great Basin. Paleogene Nevada igneous activity has been suggested to be the result of arc volcanism (Best and Christiansen, 1991). There are four main components in subduction zones; role of the subducting slab, the subducted sediments, mantle wedge and fluids overlying the subducting oceanic crust, and the overriding plate (mantle and crust).

1) Role of the subducting slab

The asthenosphere is 1-2% less dense than the lithosphere, and is what drives plate motion and subduction. As the lithosphere ages, the density increases and the subducting slab thickens and cools plunging beneath the continent at a higher angle. The differences between lithospheric age and dip of the subducting slab are reflected in

different subduction styles flanking the Pacific plate. Alternatively to the Pacific Rim, the eastern younger, warmer and less dense plate relates to a shallow angle and often failed subduction zones (Davies, 1999). Fluids driven from the subducting slab are added to the mantle wedge before they give rise to primary magmas.

2) Role of sediment in subduction

Variations among mobile LILE trace elements (K, Sr, Ba) and more specifically Be^{10} have been interpreted to indicate the presence of sediment addition to the mantle wedge (Plank and Langmuir, 1993). The limited range in compositions present in lithospheric mantle and oceanic crust differs with the compositional variation of the sedimentary packages being subducted and often provides a unique geochemical input into the zone of subduction. For instance, Peru-Chile subduct carbonates whereas the Aleutians subduct terrigenous sediment (Rea and Ruff, 1996). Lead isotopes can also provide a possible insight into the role of sediment in magma geochemistry within the mantle wedge and may show steep arrays on the $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ plots between MORB and/or overlapping with local sediments from the subducting plate (Hamilton, 1994).

3) Mantle wedge and fluids:

The mantle wedge is the part of the asthenosphere that is located directly above the subducting slab and below the overriding plate. This is where components from the descending slab are commonly mixed with the convecting mantle to generate magmas and eventually produce new continental crust. Melts generated in the mantle wedge are produced due to influx of slab-derived fluids and lowering of the solidus of the mantle,

as oppose to decompression melting associated with mantle plumes and mid-ocean ridges (Brenan et al., 1995). Aqueous and carbonic fluids are continually released from the subducting slab, from the crust, subducted sediments, and by mineral dehydration leading to metasomatization of the overlying mantle wedge. Some elements, including the large ion lithophile elements (LILE; K, Rb, Cs, Sr, Ba, Pb, and U) are transported in hydrous fluids, where as high field strength elements (HFSE; Y, Zr, Hf, Nb, Ta), the rare earth elements (REE), Ti and the transition elements remain relatively immobile (Brenan et al., 1995). Distinct enrichments of LILE relative to HFSE are observed on trace element plots of arc lavas, which reflect the mixing of LILE carried in the aqueous fluid with the mantle wedge supply of HFSE.

4) Role of the lithospheric mantle in overriding plate

The composition of the continental mantle lithosphere varies with age (i.e. density, thickness and fertility). Archean mantle lithosphere contains relatively low FeO abundances linked to komatiite extraction making it less dense. Typically, post-Archean mantle lithosphere is more enriched in FeO, and hence dense enough to be delaminated and incorporated into asthenospheric source regions of oceanic basalts and is thought to be similar to spinel peridotite inclusions in alkali basalts. The above process is therefore more likely to produce continental flood basalts such as the Columbia River Flood basalts (Hawkesworth et al., 1990). Basalts erupted at active continental margins have Nb, Ta, Zr and Hf components that are not seen in oceanic island arcs and assumed to be from enriched metasomatized subcontinental lithosphere (Pearce, 1983). The subduction process will generally provide Rb, K, Ba, Th and Sr where as the lithosphere contributes portions of LREE and P and all of its Ta, Nb, Zr, Hf, Ti, and HREE (Pearce, 1983).

During flat-slab subduction of the Farallon plate under North America (70-45 Ma), water may have been released from the down-going slab resulting in the hydration of the North American mantle lithosphere. The cooling as a result of the displacement of asthenospheric in flow and the passage of cold oceanic lithosphere beneath the trapped overriding continental lithosphere resulted in a period of ceased magmatism in Western U.S because the mantle material above the flat slab was not hot enough to form melt (Humphreys et al., 2003). As flat-slab subduction ceased, the removal of the Farallon slab would have exposed the hydrated mantle lithosphere to the underlying hot asthenosphere, possibly melting this upper mantle lithosphere and resulting in the rhyolitic ignimbrite flare-up (Humphreys et al., 2003).

1.6 Degree of crustal contamination

The average crustal compositions above Mesozoic and Cenozoic subduction zones are primarily intermediate (andesitic) with high Rb/Sr ratios relative to Bulk Earth (Ellam & Hawkesworth, 1988). As mantle-derived basaltic melts enter the crust, they undergo fractional crystallization or solidify in deeper regions of the crust until they are remobilized (Ellam & Hawkesworth, 1988). Therefore the mantle to crust flux is differentiated into more intermediate to felsic material that rises into the upper crust where as more mafic cumulates and/or remains at lower crustal depths. Andesites are therefore more likely to have been derived from fractional crystallization and crustal contamination processes rather than as primary melts of the mantle (Ellam & Hawkesworth, 1988).

In continental subduction zones, continental crust up to 70 km thick may be present. Consequently the geochemistry of the Paleogene mafic magmas is expected to have been modified as they ascended to the surface, especially through Proterozoic and Paleozoic sedimentary rocks. The crust is generally composed of low density felsic rocks enriched in incompatible elements. As the denser yet thermally buoyant mantle derived mafic magma works its way up through the crust it often undergoes differentiation in areas of magma stagnation. Denser minerals precipitate out of the melt and sink, the residual melt becomes more siliceous and heat latent of crystallization is released into the country rock. This, combined with the low melting temperature of the crust, results in partial melting of the crust which incorporates or contaminates the magma, altering the geochemical and isotopic components towards the crustal composition (Ida, 1983). The felsic minerals of the continental crust can include large amounts of incompatible elements that will preferentially partition out of the crust and into the melt (Rea and Ruff, 1996).

Chapter 2: Methodology

2.1 Field Methods

From July 12 – July 18, 2010, samples were collected from the Fish Creek Mountains field area, with the guidance of the geologic map of North-Central Nevada (Stewart and Carlson, 1976), geologic map of McCoy Mining District (Emmons and Eng, 1995), geology of Golconda Canyon of the Southern Tobin Range (Gonsior, 2006) and Nevada state maps. Samples were collected from outcrops that generally represented a lava flow; however, where flows were poorly exposed, samples were collected from voluminous float determined to be part of the flow. For the most part, all samples were weathered to a brown-red surface: float samples tend to be more weathered. Weathered surfaces were removed for thin section and geochemical analysis by rock hammer in the field. GPS locations were recorded using a Magellan handheld unit. A map of the area with sample locations is presented in Figure 5 and rock descriptions with sample number and locations are presented in Appendix A.

2.2 Petrography

All but one of the 29 samples collected in the Fish Creek Mountains were cut into thin sections and powdered for chemical analysis. H10-53 was the only sample not made into thin section. It was collected by Chris Henry at the Nevada Bureau of Mines and Geology to be dated by Ar-Ar methods, and only the crushed powder was available. To prepare thin sections, each sample was loaded into the diamond rock saw and cross-sectionally cut into 1 cm thick disks. Three disks were cut per sample, one for a thin section and the other two were to be crushed for geochemical analysis. The disc cut for

thin section preparation was made into a 3x5 cm puck using a smaller hand rock saw. To obtain a smooth surface, the pucks were then polished on a spinning rock polisher using a series of fine grit sands. A total 28 thin sections were produced for microscopy. Thin sections were studied using 4-10x magnification both in plain and cross polarized light on a Mel Sobel light microscope. A petrographic summary is in Appendix B.

2.3 Geochemistry powder preparation

The remaining two disks that were prepared on the diamond rock saw were then trimmed on the smaller wet saw to remove the weathered edges. The discs were then broken into smaller fragments using a hammer in order to fit them into the steel plated Braun Chipmunk crusher. The crusher yielded chips ranging in size from 1 mm to 2 cm which were pulverized in a Rocklabs steel ring mill for approximately 45 seconds or until they were an ultrafine powder. The steel head was cleaned for each sample using a run of silica sand to prevent cross-contamination. Soft steel milling and crushing may have added small amounts of iron to the rocks, however the expected contamination of 0.2-0.3 wt % is considered to be insignificant with respect to this study (Iwansson and Landstrom, 2000). Minor amounts of Cr were added to each sample from the steel head, but tests of samples crushed in agate versus steel showed that no other elements were added during crushing. Two vials of powder were produced; one sent for ICP-MS (trace element) at the Ontario Geological Survey, and the other used for XRF analysis of major element oxides and selected trace elements at the University of Ottawa as well as for radiogenic isotope geochemistry at Carleton University.

2.4 Major element geochemistry – X-Ray Fluorescence (XRF)

The analyses of major and trace elements in geological materials by X-ray fluorescence (XRF) is made possible by the behavior of atoms when they interact with x-radiation. An XRF spectrometer works for the reason that a sample is illuminated by an intense incident x-ray beam, causing excitation of atoms in the sample and release of x-ray spectra that varies depending on the composition of the sample (La Tour, 1989).

Samples were analyzed at the University of Ottawa X-Ray Fluorescence facility where the concentrations of SiO_2 , TiO_2 , Fe_2O_3^t , MnO , MgO , CaO , Na_2O , K_2O and P_2O_5 were determined. In addition, Ba, Co, Ga, La, Ni, Pb, Rb, Sr, Th, U, V, Y, Zr, Nb, Cr, Ce and Nd were also determined. The XRF at the University of Ottawa uses a fused disc technique. Loss on ignition (LOI) was calculated from the weight difference before and after fusion of sample and flux at 1000°C . The 28 samples were run with the internal standard 00-LT-2 and average deviation to ensure precise results. Data and standards are listed in the Appendix C.

2.5 Trace element geochemistry – Inductively Coupled Plasma Mass Spectrometry (ICP-MS)

Samples were sent to the Ontario Geological Survey's Geo Labs in Sudbury for acid-dissolution ICP-MS analysis where the 29 samples were run along with internal standard of (00-LT-2) and average deviation to ensure precise results. Concentrations of Ba, Be, Bi, Cd, Ce, Co, Cr, Cs, Cu, Dy, Er, Eu, Ga, Gd, Hf, Ho, La, Li, Lu, Mo, Nb, Nd, Ni, Pb, Pr, Rb, Sb, Sc, Sm, Sn, Sr, Ta, Tb, Th, Ti, Tl, Tm, U, V, W, Y, Yb, Zn and Zr

were determined by ICP-MS. Geochemical results are presented in appendix D and data plotted in GCDKit 3.0 (Janousek et al., 2006).

In general, the samples used for ICP-MS are introduced into an argon plasma as aerosol droplets where the plasma dries the aerosol, dissociates the molecules, and then removes an electron from the components forming singly charged ions. These charged ions are then directed into a mass filtering device known as the mass spectrometer. Only one mass to charge ratio will be allowed to pass through the mass spectrometer from entrance to exit at any given time. When exiting the mass spectrometer, ions strike the first dynode of an electron multiplier serving as a detector. The impact of the ions releases a surge of electrons that are amplified until they become a measurable pulse. The software compares the intensities of the measured pulses to those from known standards, which make up the calibration curve, in order to determine the concentration of the element.

It is typically only necessary to measure one isotope for each element measured since the ratio of the isotopes, or natural abundance is fixed in nature. On the other hand, naturally occurring lead originated from two sources; some was placed here when the Earth was formed and some is the result of the decay of radioactive materials. Therefore, lead isotope ratios may vary depending on the source of the lead. In order to accurately measure the concentration of lead in a sample, it is necessary to sum several of the isotopes available.

2.6 Isotope Geochemistry – Thermal Ionization Mass Spectrometry (TIMS)

TIMS was conducted at the Isotope Geochemistry and Geochronology Research Centre (IGGRC) at Carleton University. Strontium (Sr) and Neodymium (Nd) isotopic geochemistry were done on 24 samples. These samples showing the least alteration based on LOI values less than three. Lead (Pb) isotope geochemistry was then done on 12 samples chosen based on location within the FCM and unique petrographic characteristics. CHUR values used for Nd = 0.512638.

For each sample, approximately 100 mg were weighed and put into Teflon screw-cap beakers. The samples were dissolved in 50% HF-12N HNO₃ for two days before being dried down to a moist paste. 7N HNO₃ and subsequently 6N HCl were added, with dry down after each step to ensure that there was no undissolved residue remaining on the sample. Sr-Nd samples were finally dissolved in 2.5N HCl, and Pb samples in 1N HBr.

Lead was separated out of the sample in polyethylene columns containing anion resin using 1N HBr flush the other elements, then 6N HCl to remove the lead. The collected lead solution was then dried and dissolved in 1N HBr. A second pass of this procedure was then done to ensure purity of the collected lead.

Each Sr-Nd sample was pipetted into a column containing cation resin. Strontium was removed using 2.5N HCl and the rare earth elements were then removed using 6N HCl. The rare earth element solution was dried down and then dissolved in 0.26 HCl. Next, the solution was pipetted into columns containing coated Teflon powder, and the neodymium was removed using 0.26N HCl.

Samples were analyzed for their Pb, Sr, and Nd isotopic composition utilizing a ThermoFisher Triton T1 thermal ionization mass spectrometer at Carleton University (techniques of Cousens, 1996). All Pb mass spectrometer runs are corrected for fractionation using NIST SRM981. The average ratios measured for SRM981 are $^{206}\text{Pb}/^{204}\text{Pb} = 16.890 \pm 0.009$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.426 \pm 0.009$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.494 \pm 0.031$, based on 25 runs between September 2008 and May 2012. The fractionation correction is +0.13%/amu (based on the values of Todt et al., 1984). Sr isotope ratios are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.11940$. Two Sr standards are run at Carleton University, NIST SRM987 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710239 \pm 14$, n=30, Sept. 2008 - May 2012) and the Eimer and Amend (E&A) SrCO_3 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.708012 \pm 15$, n=14, Sept. 2008-May 2012). Nd isotope ratios are normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. Thirty runs of an internal Nd metal standard yield $^{143}\text{Nd}/^{144}\text{Nd} = 0.511823 \pm 12$ (Sept. 2008-May 2012), equivalent to a value for the La Jolla standard of 0.511852. All quoted uncertainties are 2-sigma standard deviations of the mean.

Age-corrected isotopic data can be found in the Appendix E. The samples were run with lab standards to check for precision. The error listed in the table along with the samples is the internal which is equal to 2 standard deviations of the mean. Precision* is the internal error of the run, the variability caused by temperatures and voltages used to excite the sample. The external error is the reproducibility of the run, which is checked by running a standard with each wheel run on the mass spectrometer. This standard is added to, and checked against, the average of standards run in the past year to assure the

machine is running accurately. The standard for each isotope ratio is listed below the data table in the Appendix E.

ϵNd values were calculated using the following formula. $T = 34 \text{ Ma}$.

$$\epsilon\text{Nd}^T = \left(\frac{{}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{sample}} - {}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{CHUR}}}{{}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{CHUR}}} \right) * 10,000$$

Chapter 3: Geology and Field Relationships in the Fish Creek Mountains

3.1 Physical Observations

Paleogene flows of the FCM and surrounding area were either overlain or surrounded by the 24.9 Ma FCMT, 33.8 Ma Caetano tuff, 28.8 Ma Campbell Creek tuff or younger 16-10 Ma basalt and basaltic andesite flows related to the volcanism of the Northern Nevada Rift to the North. Flows have also filled paleocanyons in at least two of the sample locations.

Photographs of representative samples are included from Figure 6–14. Appendix A summarizes the basic geological information from the field for the 29 samples used in this study where as the thin section petrography of the volcanic rocks is summarized in Appendix B for all samples.

The lavas range in composition from basaltic andesites to rhyolites. Paleogene lavas from the FCM occur mainly in 4 different geographical locations; furthest west within the Golconda Canyon of the Tobin range, a western FCM suite, an eastern FCM suite within the Horseshoe Basin, and a southern suite taken from the Shoshone range immediately south of the FCM (Figure 5). Generally, the FCM western flow trends north-south, where as the flows of the Tobin Range, Shoshone range, and eastern outcrops occur in clusters intermingled with tuff and younger syn to post-extensional basalt.

Paleogene volcanic samples taken from the Tobin Range to the west of the FCM (Figure 5) (samples 10-BV-39 through 42) are all classified as basaltic andesites (Le Bas et al., 1986) except for sample 10-BV-41 which is an andesite.

Figure 5: Rock sample locations for samples collected during the 2010 field season plotted as blue crosses on the topographic map of the FCM and surrounding area. They are divided into four main geographical areas (encircled) Red crosses represent 24.9 Ma rhyolitic tuff. (Modified from Fish Creek Mountains and Edwards Creek Valley USGS topographic maps, U.S. Geological Survey, 1:100 000 scale).

They were all sampled from within the Golconda Canyon area of the southern Tobin Range which consists of a pre-Paleogene east-west trending paleo-valley in filled and overtopped by Paleogene volcanic and sedimentary units (Gonsior, 2006). Sample 10-BV-42, located at the west end of the Golconda Canyon is overlain by 33.8 Ma Caetano tuff. This sample comprises a vertically jointed outcrop typical of a thick mafic flows that are common in the western FCM. Other Paleogene units found in the Golconda Canyon but not sampled include a biotite rhyolitic ignimbrite (33.28 $\pm$ 0.22 Ma) and a thick sequence of andesitic rocks overlying the Caetano tuff in most exposures (32 Ma) (Gonsior, 2006). Basaltic andesite samples, excluding sample 42, were all vesicular and exposed in weathered outcrops. In general these samples all have a glassy aphanitic groundmass with up to 20-25% euhedral plagioclase phenocrysts 1-5mm and 10-15% up to 7mm hornblende phenocrysts with minor pyroxene (Figure 6). Sample 41, the hornblende andesite, taken from an outcrop that shows vertical joints, lithic clasts and boulders on top yet has a volcanic matrix filled with crystals (Figure 7). This flow is probably volcanoclastic (collapsed dome or debris flow). Most samples contained weathered mafic minerals and iron oxide stained fractures. Phenocrysts range in size from millimeters to no longer than 1 centimeter.

Lavas from the western FCM found in Jersey Canyon (Figure 5) range from andesite through to rhyolite and can be further subdivided into vesicular and non-vesicular samples. Samples 10-BV-07 through 10-BV-11 are all non-vesicular, massive, and mostly porphyritic andesites taken from columnar jointed outcrops typical of thicker flows (Figure 8). Sample 10-BV-09 strikes 294 with a dip of 70° N and was taken in close proximity to samples 7 to 11. The remainder of the samples were taken from

different types of vesicular flows and tended to be more evolved dacites to rhyolites. Sample 10-BV-12A and B taken from the upper portion of Jersey Canyon are most likely rhyolitic volcanic breccias where as sample 10-BV-34A and B were embedded in a section of FCMT as rounded dacite lithic clasts. Sample 10-BV-06 located just south of sample 34 is suspected to have been part of a lava flow top or a 'a' (Figure 9). Most samples exhibited trachytic texture. Similar to the Tobin Range samples, all western FCM samples also tend to have a glassy aphanitic groundmass with plagioclase and hornblende microphenocrysts.

Southern samples were collected just south of FCM part of the Shoshone Mountain range and next to the Red Butte area (Figure 5). These flows range in composition from andesite through to rhyolite; however this region typically includes more evolved rocks, commonly dacites through rhyolites, and can be subdivided into samples taken from non-tilted columnar jointed outcrops versus those from tilted fissile and platy outcrops. The andesites are more commonly found in jointed outcrops in contrast to the dacitic rocks usually found in platy ones. Sample 10-BV-45 showed crude columnar jointing of fresh lava and contains trace anhedral olivine with minor plagioclase (Figure 10). Sample 10-BV-48A, taken from a dacitic cone, contains shear zones typical of more viscous flows and a very glassy matrix similar to the jointed andesites. The rest of the samples were taken from platy outcrops commonly containing porphyritic hornblende (1mm) and 15 % euhedral plagioclase up to 5mm long. Other samples contain trace amounts of biotite, olivine and pyroxene.

Figure 6: Typical basaltic andesite representative sample (10-BV-39) from the Tobin Range in outcrop (A) and hand sample (B). The outcrop is fairly weathered and in hand sample is vesicular with a glassy aphanitic groundmass with up to 25% plagioclase (1-5mm) and 10% hornblende (7mm) phenocrysts with minor pyroxene.

Figure 7: Sample 10-BV-41 in outcrop (A) and hand sample (B). This andesite was taken from an outcrop that could have been a collapsed dome or a debris flow that shows vertical joints, lithic clasts and boulders on top yet has a volcanic matrix filled with crystals.

Figure 8: Sample 10-BV-10 in outcrop (A) and in hand sample (B) characteristic of non-vesicular flows from the western FCM suite. They are mostly porphyritic andesites taken from columnar jointed outcrops.

Figure 9: Sample 10-BV-06 in outcrop (A) and hand sample (B). This is the most northern expression of Paleogene volcanics in the western FCM suite and is suspected to have been part of a lava flow top or a'a. Characteristically it is more vesicular and weathered compared the porphyritic andesite samples taken further south.

Dacitic sample 10-BV-47 was taken from a fissile, west-dipping bedded outcrop, and the float down the hill from the outcrop has clasts of dacite mixed with glass and other clasts and sediment that may have originally been flow top material (Figure 11). Sample 10-BV-47, located on the ridge north of Red Butte includes euhedral plagioclase and biotite in a crystalline matrix. Sample number 10-BV-48 A and B were taken from a dacitic to rhyolitic cone which appeared to have lava tunnels or splash up the side, determined to be 28.9 Ma Campbell Creek Tuff. Here the tuff appears to flow up along the margin of the dacite cone with a near-vertical flow attitude, indicating that the cone is older than the Campbell Creek Tuff (Figure 12). Sample 10-BV-48B has spherulites which are indicative of devitrification (formation of minerals from glass). The sample contains abundant plagioclase and biotite. In hand specimen southern Shoshone Mountain Range samples all have a glassy aphanitic groundmass with trachytic texture. Therefore southern flows are typically platy in outcrop and seem to be more evolved than western and eastern flows.

Eastern flows are found on the east side of the FCM enclosed in the area known as Horseshoe Basin (Figure 5). All rock samples are andesites except for 10-BV-16, which is a basaltic andesite. 10-BV-16 has trachytic texture around olivine phenocrysts and it is restricted to this part of the basin, flanked on both sides by glassy flows of Pinnacle Mountain rhyolite flow dome complex (32.9 ± 0.9 to 34.1 ± 0.9) (Emmons and Eng, 1995) (Figure 13). Since this basaltic andesitic large flow is clearly overlain by a vitreous flow, it is most likely that the glassy unit is basal vitrophyre of Pinnacle Mountain rhyolite flow and the upper peak is devitrified Pinnacle Mountain tuff.

Figure 10: Sample 10-BV-45 in outcrop (A) shows crude columnar jointing of fresh lava and in hand sample contains trace anhedral olivine with minor plagioclase (B) sampled from the Shoshone Range.

Figure 11: Sample 10-BV-47 in outcrop (A) and hand sample (B) characteristic of samples taken from platy and fissile outcrops of the Shoshone range. These samples commonly contain porphyritic hornblende (1mm), plagioclase up to 5 mm long and biotite in a crystalline matrix. The fissile dacitic beds of sample 10-BV-47 dip west and are located on the ridge north of Red Butte. Hand sample also shows flow texture of plagioclase laths.

Figure 12: Dacitic-rhyolitic cone of sample 10-BV-48 from the Shoshone range which appeared to have lava tunnels up the side determined possibly to be weathered out Campbell Creek Tuff with a near-vertical flow attitude (enclosed in border).

In general, eastern flow rocks are typically highly vesicular (more so than any other) and have a glassy aphanitic groundmass with xenocrysts and megacrysts of euhedral plagioclase (25%) and hornblende (5%) up to 6 mm. Sample 10-BV-29 and 10-BV-30 overlie each other and are located at the southern end of Horseshoe Basin. Sample 10-BV-29 is bedded and very sheared, striking easterly and dipping 10°N (Figure 14). Samples 31 and 32 were also sampled in close proximity. In sample 31, there is no observed bedding, is more vesicular (round to elongate) and less crystalline than any other in this group, whereas sample 32 appears to be non-vesicular occurring in a jointed outcrop similar to more viscous flows of the western and southern sample locations. Therefore, with the exception of sample 32 and the localized basalt of sample 16, easterly samples are mainly hornblende andesitic flows, which are less viscous or thick due to the abundance of vesicles in this assemblage.

To conclude, sampled flows from the Tobin Range (Golconda Canyon) are generally vesicular basaltic andesites; however more evolved rocks also exist but were not sampled. Flows from the western Jersey Canyon are generally thicker and possibly more viscous due to their lower abundance of vesicles typically found in jointed outcrops, and more evolved dacite to rhyolite were also observed but are less extensive than the other areas of FCM. Southern flows of the Shoshone Mountains contained the most evolved rocks, typically dacite to rhyolite, often found as platy and fissile outcrops. Lastly, the eastern flows within Horseshoe Basin were mostly hornblende andesites and contained the greatest proportion of vesicles implying a less viscous flow or gas-rich magmas.

Figure 13: Outcrop (A) and hand sample (B) picture of 10-BV-16. The outcrop of this glassy basaltic andesite is flanked on both sides by glassy flows of felsic ignimbrite. It has flow textures around olivine phenocrysts and is restricted to this part of the assemblage.

Figure 14: Sample 10-BV-29 in outcrop (A) and hand sample (B) part of the Eastern suite of FCM. (A) Shows the outcrop as bedded and very sheared, striking easterly and dipping 10°N whereas (B) shows an average abundance of vesicles present in most samples of this suite.

Chapter 4: Geochronology

4.1 Introduction

A number of Ar-Ar ages have been established from labs at the USGS sampled from the FCM area in addition to a few from the surrounding rock formations of the Buffalo Valley area and are presented below.

4.2 Previous dating

Paleogene andesitic ages from the surrounding rock units of the FCM have been determined and are as follows; hornblende andesite sample 08-DJ-126 from Mt Caetano (35.2 Ma), andesite sample H05-52 from the Toiyabe Range (35.2 Ma), glassy clinopyroxene-plagioclase andesite porphyry sample 09-DJ-85 from the Fye Canyon Volcanics (35.5 Ma), and 09-DJ-87 porphyry hornblende-plagioclase andesite also from the Fye Canyon Volcanics with an age of 35.5 Ma (John et al., 2008).

Four ages were determined in the Golconda Canyon, part of the Tobin Range by Gonsior (2006); biotite rhyolitic ignimbrite sample TR05-26 (33.28 +/- 0.22 Ma), ignimbrite TR-9 (33.03 +/- 0.25 Ma), rhyolitic ignimbrite TR05-21 (24.95 +/- 0.17 Ma) and a younger basalt sequence TR-77 (14.10 +/- 0.12 Ma). Surrounding tuffaceous units include the Caetano and Campbell Creek tuffs with ages of 33.8 Ma and 28.8 Ma respectively. The thick sequence of andesitic rocks overlying the Caetano tuff in most exposures in the Tobin Range has been dated by McKee et al., (1971) with a K-Ar age of approximately 32 Ma. There also exist younger 16-10 Ma basalts and basaltic andesite flows related to the volcanism of the Snake River Province and the Northern Nevada Rift

to the North which include the TR-77 sample above with an age of 14.10 ± 0.12 Ma (Gonsior, 2006).

Four ages from the Sulphur Springs Range in central Nevada, east of the FCM have been determined and further discussed in Ryskamp et al. (2008). The U-Pb zircon ages of these units are as follows; an andesite ($31.4 \pm 1.3/-0.5$ Ma), plagioclase dacite dome (35 ± 0.5 Ma), biotite dacite tuff (35.5 ± 0.4 Ma) and biotite porphyry intrusion (35.9 ± 0.5 Ma)

A number of tuffaceous units have also been dated from the FCM and Shoshone ranges presented in McKee et al., (1971). FCM K-Ar and fission track dated samples are as follows; FCM rhyolitic welded tuff (24.4 Ma), Bates Mountain rhyolite welded tuff (23.1 ± 1.7), and Caetano rhyolitic welded tuff ($31.2 - 33.3 \pm 2.7$ Ma). Shoshone Range K-Ar and fission track dated samples include; Rhyolitic Caetano welded tuff (31.3 Ma), Bates Mountain rhyolitic welded tuff (24.7 ± 1.0 Ma), and rhyolitic welded Tuff of McCoy Mine (26.3 ± 1.6 Ma). However, in contrast to the dated tuffaceous units, Paleogene rhyolitic to basaltic lava flows from the FCM and surrounding areas have been studied far less. Dates determined by K-Ar method for other Paleogene flows include the Clan Alpine Mountains Andesitic flow (35.0 ± 1.2 Ma), and the Simpson Park Mountain flows; Dacite flow (34.5 Ma), glassy rhyolite flow (30.9 ± 0.7 Ma) and an andesite-dacite flow (35.4 Ma) (McKee et al., 1971).

Four tuffs dated from Bates Mountain, which is located southwest of the FCM, include the Nine Hill tuff (25.27 Ma), Tuff of Campbell creek (28.6 Ma), Tuff of Sutcliffe (30.48 Ma), and Tuff of Rattlesnake Canyon (31.03 Ma) (John et al., 2008).

4.3 New Ar-Ar ages

Of the 29 Paleogene mafic to felsic samples analyzed in this study, three samples have been dated using Ar-Ar methods by Chris Henry at the Nevada Bureau of Mines. Samples H-10-53 (Rhyolite dome), 10-BV-07 (Andesite) and 10-BV-47 (Dacite), were each sampled from different geographical areas of the FCM and surrounding area and yielded ages of 34.24 ± 0.05 Ma, 33.3 ± 0.3 Ma and 33.82 ± 0.14 Ma respectively (C. Henry, pers. Comm., 2011). Since they all overlap with or close to their uncertainty ranges we can deduce that the eruption of these Paleogene lavas seems to have occurred within a relatively short time period of 1 Ma between 33.3 and 34.3 Ma.

In addition to these dated samples, younger Fish Creek Mountain rhyolite ash flow tuff (FCMT) has also been dated using sanidine and further discussed in Varve (2013). The ages and locations are as follows; 10-BV-38 sampled from the western Golconda Canyon of the Tobin Range (24.95 ± 0.08 Ma), 10-BV-17 sampled from the Horseshoe Basin of the Fish Creek Mountains (24.91 ± 0.05 Ma), and H03-73 sampled from the southern Fish Creek Mountains (24.88 ± 0.05 Ma).

Chapter 5: Petrography of Paleogene volcanic rocks of the FCM

5.1: Introduction

The purpose of this chapter is to describe the petrographic characteristics of the Paleogene lava flows in the FCM. Textural differences were observed in both matrix and phenocrysts/xenocrysts. The mineralogy of samples described in this section is done in order of abundance in thin section. The petrographic descriptions of each slide are presented in Appendix B, and photographs of representative thin section slides are included in figures 15 to 23. Plagioclase phenocrysts and xenocrysts are discussed in terms of degrees of disequilibrium; low degree describes phenocrysts with only a slight appearance of sieve texture, which is a sign of disequilibrium, moderate degree describes phenocrysts with abundant sieve texture which looks as if it is eating away at the crystal from outwards to in (Figure 19 and 20); high degree describes xenocrysts which are highly sieved and are anhedral with a precipitated rim which is potentially in equilibrium with the surrounding melt (Figure 15).

5.2 Basaltic Andesites

Basaltic andesites were mostly samples from the Tobin Range within Golconda Canyon; however one sample (16) is from the eastern Horseshoe Basin. They typically contain phenocrysts of plagioclase, pyroxene, olivine, amphibole, biotite and K-feldspar. In these rocks the abundance of plagioclase ranges from 15-20% where plagioclase is present as both the most common phenocrystic and groundmass phase and where trachytic texture of plagioclase laths is common. Plagioclase phenocrysts vary from 0.5 to 5 mm long and typically show minor seritization. Compared to other rock types,

plagioclase phenocrysts seem to show little to no disequilibrium with the melt. Plagioclase xenocrysts were found in sample 42, indicative of some crustal contamination, and display high disequilibrium with the melt (Figure 15). Olivine phenocrysts often show iddingsite weathering along fractures and vary in size from 0.5 to 1 mm (Figure 16). Porphyritic hornblende and biotite (0.5-1.5mm) are more apparent in sample 40 with reaction rims which represent that they were in disequilibrium with the melt during crystallization either due to decompression, degassing or assimilation. Pyroxenes also occur as a phenocrystic phase (0.5-5mm) with no apparent reaction rims. Opaque minerals vary from moderate to abundant throughout the samples. For the most part, surrounding the phenocrysts is a glassy to microcrystalline groundmass of very fine grained feldspars and opaques. Samples 39 and 40 (taken in proximity of each other) are vesicular and less fresh where as samples 16 and 42 are the most mafic samples and fresh. Sample 39 and 40 also contain K-feldspar (2-3%) where as samples 16 and 42 do not.

5.3 Andesites

Andesites can be divided in terms of trachytic and non-trachytic. Andesites contain phenocrysts of plagioclase, pyroxene, K-feldspar (Carlsbad twinning), biotite and hornblende. Alteration is much more common in the andesites in contrast to the basaltic andesites. Chloritization present in the andesites is often associated with biotite, pyroxene or amphiboles and sericite with plagioclase and K-feldspar (Figure 17). Plagioclase is typically euhedral with moderate disequilibrium with the melt, and is usually zoned. Plagioclase phenocrysts also display sieve texture, indicative of magma mixing and disequilibrium, which is very common in the western andesites.

Figure 15: Basaltic andesite sample 10-BV-42 shows a plagioclase xenocryst displaying a high degree of disequilibrium with the melt. It is sieved and almost completely rounded (anhedral) and has a precipitated rim which is possibly in equilibrium with the surrounding melt. Its presence is also indicative of some crustal contamination.

Figure 16: Basaltic andesite sample 10-BV-42 shows olivine phenocrysts with iddingsite weathering in its fractures varying in size from 1-2 mm. The surrounding groundmass consists of glassy very fine grained to microcrystalline composed of finer grained phases plus opaques.

Figure 17: Highly altered andesite sample 10-BV-05 showing chloritization (light green in ppl) on the left and seritization of plagioclase phenocrysts in xpl on the right.

Figure 18: Andesitic sample 10-BV-41 under xpl shows a hornblende opaque reaction rim indicating some disequilibrium with the melt either due to decompression or degassing. Plagioclase seems to be in moderate disequilibrium with the melt.

Biotite and hornblende show evidence of being in disequilibrium as they have opaque reaction rims that sometimes extend into the core. This suggests either shallow degassing, or slow magma ascent across the amphibole stability field limit (Rutherford et al., 1998) (Figure 18). Sample 12B is the most altered, which was collected from a volcanic breccia and has both seritization of plagioclase and Mg rich chloritization of pyroxenes, biotite and amphiboles based on its first order interference colours. Samples 10-BV-46A and 10-BV-41 contain abundant lithic clasts of limestone, based on the presence of carbonate minerals in thin section. Vesicles vary from 1-2%. The groundmass is glassy to microcrystalline.

The trachyandesites can be examined in terms of vesicular versus non-vesicular. Commonly both contain a trachytic texture of plagioclase laths in a glassy to microcrystalline groundmass. The non vesicular samples (7-10, 45) were taken from columnar jointed outcrops typical of thicker flows and contain phenocrysts of plagioclase, pyroxenes, olivine, K-feldspar, hornblende and biotite. Megacrysts of plagioclase are also present in samples 07 and 09 up to 1 cm and also display sieve texture (Figure 19). Plagioclase minerals are also sometimes zoned with moderate disequilibrium. Zonation is typically found on the larger plagioclase phenocrysts and megacrysts, which indicates a possible recharge by magmas from below as the recharged magmas may have different Ca/Na ratio. Large reaction rims are present on biotite and hornblende in sample 07 and trace epidote is found in sample 45. Vesicular samples (29-32, 34A, 49) typically contain phenocrysts of plagioclase, clinopyroxene, K-feldspar, olivine, biotite and hornblende. Plagioclase and K-feldspar are often seritized and sometimes zoned with evidence for moderate disequilibrium. Sample 31 is the most

vesicular containing as much as 5% vesicles (Figure 20). There is a lack of reaction rims around biotite and hornblende in the vesicular samples.

5.4 Dacites

Dacites, classified as trachy-dacites according to the TAS discrimination diagram, vary from alkaline (11, 34B) to subalkaline (6, 44, 46B, 48A). Alkaline and subalkaline samples contain porphyritic plagioclase (1-3 mm) with minor biotite and pyroxene phenocrysts. In both alkaline and subalkaline samples, plagioclase phenocrysts are sometimes seritized, display sieve texture, are partly resorbed and are thus in moderate disequilibrium with the melt (Figure 21). Reaction rims around porphyritic plagioclase are only present in sample 34B whereas biotite reaction rims are more common in the rest of the samples. Biotite is chloritized in sample 11. Vesicle percent varies, but generally the dacites contain 1-2 % vesicles for both alkaline and subalkaline types. Subalkaline sample 46B contains plagioclase megacrysts (12-15mm) and spherulites are apparent in thin section under plane polarized light (ppl) (Figure 22). Sample 10-BV-48A contains rounded xenocrysts of sericitized K-feldspar indicative of crustal contamination. Other phenocrystic phases in subalkaline dacites include K-feldspar, biotite and pyroxene. The groundmass is glassy to microcrystalline.

5.5 Rhyolites

All rhyolitic samples are subalkaline. Four out of 29 samples were classified as rhyolite, 2 of which rest on the dividing line in between dacites and rhyolites. They typically contain phenocrysts of plagioclase, K-feldspar, biotite, pyroxene and minor quartz; however, rhyolites tend to be the least porphyritic out of all rock types.

Figure 19: Andesitic sample 10-BV-09 under xpl displaying a large zoned plagioclase megacrysts with sieve texture.

Figure 20: Andesitic sample 10-BV-31. This sample is highly vesicular and plagioclase phenocrysts display sieve texture and seem to be in moderate disequilibrium with the melt.

Figure 21: Dacitic sample 10-BV-34B under xpl. Plagioclase phenocrysts are seritized, show sieve texture, are partly resorbed and display reaction rims in moderate disequilibrium.

Figure 22: Dacitic sample 10-BV-46B under ppl. Sample shows spherulites (Sph) as rounded structures which indicates divitrification and plagioclase phenocrysts are in moderate disequilibrium.

Plagioclase phenocrysts typically vary from 1 to 5 mm and are commonly in moderate disequilibrium with the melt. Zoned plagioclase is common in sample 12A classified as a volcanic breccia in outcrop. Rounded plagioclase and K-feldspar are also common in sample 12A, interpreted to be xenocrysts from crustal contamination. Reaction rims are found mostly around biotite and less so around plagioclase. Spherulites are also present in plane polarized light of sample 48B as rounded structures (Figure 23). Biotite, like plagioclase, appears to be in moderate disequilibrium in sample 47 based on opaque reaction rims. All rhyolites are vesicular and contain a glassy to microcrystalline groundmass.

5.6 Summary

All rocks types include lavas with <35% phenocrysts, <10% vesicles, although most samples have <25% phenocrysts. Amongst all rock types, plagioclase exists as both the most common phenocrystic phase and is often found in the trachytic groundmass as laths. Microporphyritic is a common texture of the basaltic andesites where most of the weakly porphyritic samples exhibited trachytic texture. Typical microscopic features seen in most samples include sieve texture in phenocrystic plagioclase, oscillatory zoned plagioclase, amphibole phenocrysts with reaction rims, plagioclase microlites and the presence of glass in the groundmass of several of the more weakly porphyritic rocks. These textures either suggest shallow magmatic process or that these flows cooled extremely rapidly. Other phenocrysts phases typically include pyroxene (5-10%), K-feldspar (1-10%), biotite (1-8%), olivine (1-5%), hornblende (1-3%), and quartz (1-2%) with most phenocrysts being euhedral to subhedral. Phenocrysts range from millimeters to up to 1.5 cm; plagioclase and hornblende are commonly the largest. In andesitic and

dacitic rock types, reaction rims are commonly found around biotite and hornblende in varying intensities either the result of decompression and/or degassing and H₂O loss from the magma. In terms of alteration, biotite is commonly altered to chlorite, plagioclase and K-feldspar to sericite, and olivine to iddingsite. The visible matrix grains are notably small compared to the phenocrysts and consists of the same phases as are present as phenocrysts, as well as glass and opaque's. Vesicles are common in all rock suites, especially in eastern Horseshoe Basin samples, except for samples taken from columnar jointed outcrops typical of thicker flows.

Figure 23: Rhyolitic sample 10-BV-48B under ppl. The rounded structures and spherulites (sph) and the elongated brown-black minerals are biotite with reaction rims in a glassy microcrystalline groundmass.

**Chapter 6: Geochemistry; Geochemical and Radiogenic Isotope Systematics of
Paleogene felsic to mafic rocks of the Fish Creek Mountains, north-central Nevada,
Western United States.**

6.1: Introduction

Whole rock, trace element and isotopic data from the study area are presented in Appendices C, D and E. The purpose of this chapter is to use the geochemical data to describe the chemical characteristics of these rocks. For plotting all major and trace element analyses have been recalculated on a volatile-free basis. Geochemical data from 29 collected rock samples are plotted on the conventional total alkalis silica (TAS) plot. Of the 29 samples analyzed, 4 are classified as basaltic andesites, 15 are andesites, 6 are dacites, and 4 are rhyolites (Figure 24; Le Bas et al., 1986). With the exception of 4 samples that are slightly alkaline, the samples are sub-alkaline. On the AFM diagram (Figure 25) the samples mostly plot in the calc-alkaline field of Irvine and Baragar (1971). Magnesium (Mg) numbers ($Mg\# = 100 \times (Mg / (Mg + Fe^{2+}))$ where $Fe^{2+} = 0.9 \text{ Fe total}$) range from about 13-56 with an average of 40. Whereas the basaltic andesites have restricted $Mg\#$'s in between 48 and 57, the andesites display much more scatter with values between 30 and 55.

6.2 Whole rock major geochemistry

The volatile free major element data from all 29 samples are plotted on SiO_2 variation diagrams (Figure 26). The purpose of these diagrams is to present the data visually and to display the variation in abundance of the elements in the 4 rock types found in each of the areas within the Fish Creek Mountains described above.

Figure 24: Total alkalis vs. silica diagram of Le Bas et al. (1986), using recalculated analyses all in wt (%). Dotted line represents the division in between alkaline and subalkaline from Irvine and Baragar (1971).

Figure 25: Paleogene mafic to felsic samples plotted on a MgO, FeO^t , and $\text{Na}_2\text{O}+\text{K}_2\text{O}$ (wt%) ternary diagram (Irvine and Baragar 1971).

Some of the major element oxides, particularly the alkalis (K_2O and Na_2O) are known to be mobile during alteration (e.g., Hughes 1973). As a result, major element oxide discrimination diagrams that use these elements are likely to be less reliable than diagrams involving the less mobile, high-field strength elements such as Ti, Zr, Nb and Y (Winchester and Floyd 1977). Hence the latter elements carry more weight in the interpretations presented here.

Variations in the major element chemistry, taking into account the observed phenocrysts phases can help constrain the differentiation history of these volcanic rocks. The decrease of FeO^t and TiO_2 with increasing SiO_2 and the presence of opaque minerals in most rock samples, especially the most mafic samples, illustrate the importance of titano-magnetite as a fractionating phase. MnO vs. SiO_2 diagram shows a similar tendency to the Fe plot which reflects the fact that Mn readily substitutes for Fe in iron-bearing minerals (Figure 26).

Al_2O_3 varies widely amongst rock types with the exception of rhyolites and consistent with the abundance of plagioclase in thin section (Figure 26). The decrease of CaO with increased SiO_2 , and small negative Eu anomaly present in more evolved samples (especially the most evolved dacites and rhyolites) indicate the fractionation of plagioclase and possibly clinopyroxene, which is the major phenocrystic phase found in all rock types (Figure 26).

The wide range of Na_2O concentrations may reflect sodium mobility and the differences in the extent of albitization among the samples. The subtle decrease of P_2O_5 with respect to increasing SiO_2 reflects the fractionation of apatite (Figure 26).

Figure 26: Silica variation diagrams for TiO_2 , Al_2O_3 , MgO , CaO , Na_2O , P_2O_5 , FeO^t , MnO and $\text{CaO}/\text{Al}_2\text{O}_3$ all in Wt (%). Symbols are as in Figure 25.

Figure 27: K_2O vs. SiO_2 plot of Peccerillo and Taylor (1976) where the majority of FCM samples plot in the high-K field to shoshonite series with the exception of one basaltic andesite that plots in the med-K field (10-BV-42). Symbols are as in Figure 25.

When samples were plotted on a $K_2O\%$ vs. $SiO_2\%$ graph from Peccerillo and Taylor, 1976 (Figure 27), the majority of the samples plotted in the high-K field to shoshonite series. However, there is one sample (sample 10-BV-42), a basaltic andesite that plots in the med-K field. The more mafic samples generally have lower concentrations of K_2O relative to samples with high silica concentrations. The highest values were mostly observed in southern flow rocks, followed by eastern and western areas.

6.3 Whole rock trace element geochemistry

Combined with major element trends, trace elements can also be used to help constrain the differentiation history. Similar to Al_2O_3 , Sr also varies widely amongst rock types consistent with the variable amount of plagioclase in most samples where Sr often substitutes for Ca in plagioclase (Figure 28).

The observed decreases in CaO/Al_2O_3 , CaO , and Sc with increasing SiO_2 are consistent with clinopyroxene fractionation. The decrease of MgO and Co with respect to SiO_2 in the more mafic rocks probably reflects olivine fractionation (Figure 28).

Combined with the decreases of FeO^t and TiO_2 with increasing SiO_2 , V also supports titanomagnetite as a fractionating phase. When plotting V vs. Fe_2O_3 in Figure 29, a positive correlation exists which reflects the ability of V to substitute for Ti in titanium iron oxides. Cr and Ni are also compatible with ferromagnesian minerals and are therefore relatively abundant in most mafic rocks, especially samples 10-BV-42 and 10-BV-16 where as the more evolved southern samples tend to have the lowest Ni , Cr and V .

The occurrence of xenocrysts and the large range in Pb, Sr and Nd concentrations in most rock types regardless of silica content, suggest that crustal contamination may also be a process affecting the chemistry (Figure 28). Th concentrations may be derived from the continental crust and display the widest ranges within the dacites with sample 10-BV-11 containing as much as 50 ppm Th (Figure 28). The positive correlation that Pb and Th have with SiO₂ increasing by a factor of five indicates that FCM samples experienced some Pb and Th contamination in contrast to other incompatible elements such as Ba or Zr that vary irregularly.

Turning to incompatible element diagrams, there exists a wide range of fluid mobile (large ion lithophile elements or LILE) to fluid immobile (high field strength elements or HFSE and REE) elements in these rock types. The incompatible element diagrams in Figure 30a-d are normalized to primitive mantle after Sun and McDonough (1988) and are strongly enriched in incompatible elements (e.g. Rb, Ba, Th, U) relative to primitive mantle and are relatively homogeneous. Negative troughs occur at Nb and Ti in all samples. Peaks occur in all samples at Pb and most at Sr and K (LILE). Therefore, fluid mobile elements such as Rb, Ba, U, K and Sr are enriched compared to less fluid mobile elements such as Nb, Ta, Zr, Ti and the REE (Figure 30a). The incompatible element diagrams for the basaltic andesites and andesites are relatively homogeneous with the exception of two samples from the basaltic andesites (10-BV-16 and 10-BV-42) and two from the andesites (10-BV-05 and 10-BV-41). Basaltic andesite sample 10-BV-16 shows an overall higher abundance of most incompatible elements, especially Pb, whereas 10-BV-42 shows an overall depletion of incompatible elements whereas still reflecting the same pattern as seen in the other basaltic andesites (Figure 30b).

Figure 28: Silica variation (wt%) diagrams for Ce, Sr, Sc, Co, V, Pb, Zr, Ba, Th, Cr, Ni and Nb (all in ppm). Symbols are as in Figure 25.

Figure 29: Plot of total iron as Fe₂O₃ vs. V. Symbols are as in Figure 25.

Figure 30: Incompatible element diagrams (c) andesites and (d) dacites and rhyolites. All values are normalized to primitive mantle (Sun and McDonough, 1988). The lowest sample of the andesite plot is sample 10-BV-41.

The andesite sample 10-BV-05 shows depletion in barium compared to the overall enrichment and a more prominent depletion of phosphorous, and sample 10-BV-41 shows an overall slight depletion of all incompatible elements compared to the rest of the andesites (Figure 30c). Dacites show an overall enrichment in LREE compared to the rest of the rock compositions and sample 34B is strongly enriched in Pb, possibly as a result of crustal contamination. Lastly, the rhyolites plot at the lowest concentrations of Sr and P, possibly as a result of minor fractional crystallization due to apatite and some plagioclase fractionation (Figure 30d).

Trace element ratio plots such as Zr/Nb vs. Ce/Pb (Figure 31) show a pronounced overlap among basaltic andesites and andesites. Although trends are evident in some ratio plots when reviewing the data as a whole, grouping among individual rock types is less common. The plot of Ce/Pb vs. Cs/Rb (Figure 32) distinguish subduction-related fluid addition characteristics (high Pb) from mantle characteristics (high Ce). The common features of these plots are that the basaltic andesites are located at both extremes of the plots as well as in between them.

6.4 Isotope Geochemistry

The initial $^{87}\text{Sr}/^{86}\text{Sr}$ of the basaltic andesites range from 0.705907 to 0.707873. The ϵNd_{34} values calculated at 34 Ma of the basaltic andesites range from -1.9 to -6.5. The andesites range in $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.705833 to 0.707098, where as their ϵNd_{34} values range from -2.5 to -5 (Figure 33a). Overall, an increase in $^{87}\text{Sr}/^{86}\text{Sr}$ does correlate with an increase in silica content (Figure 33c), whereas a very slight negative correlation can be found when looking at the $^{143}\text{Nd}/^{144}\text{Nd}$ vs. SiO_2 plot (Figure 33e). Overall, the Sr and Nd

isotopic values overlap between all rock groups, however the dacites and rhyolites do show a more restrictive range compared to the basaltic andesites and andesites. When plotting ϵNd_{34} versus $^{87}\text{Sr}/^{86}\text{Sr}$ initial on Figure 33a, the isotopes shows a negative correlation and the basaltic andesites plot at both high Sr and low Nd, and at high Nd and low Sr where as more evolved rock types plot in between.

Pb isotopic ratios also plot similarly to the Sr and Nd isotopic ratios. Individually, the basaltic andesites plot at both high and low $^{206}\text{Pb}/^{204}\text{Pb}$ values where as other rock types plot in between them. Basaltic andesites range in $^{206}\text{Pb}/^{204}\text{Pb}$ from 19.25 to 19.40 where as the andesites range from 19.32 to 19.4 (Figure 33b). Also shown in Figure 33d on the $^{206}\text{Pb}/^{204}\text{Pb}$ vs. $^{208}\text{Pb}/^{204}\text{Pb}$ plot is a slight positive correlation with an observable overlap amongst the more evolved rock types.

Figure 31: Trace element ratio plot of Ce/Pb vs. Zr/Nb. X-axis indicates an increase of partial melting percent and Y-axis indicates an increase of mantle or decrease of slab fluid components. Modern South Cascades Basalts data from Borg (1997) and BV (Buffalo Valley) data from B. Cousens pers. Comm. 2013. Symbols are as in Figure 25.

Figure 32: Trace element plot of Ce/Pb vs. Cs/Rb. X-axis indicates an increase of sediments and Y-axis indicates an increase of mantle or increase of fluid components. Cascadia sediments from Prytulak et al. 2006. Symbols are as in Figure 25.

Figure 33: Initial Nd, Sr and Pb isotopic ratios in the FCM and surrounding areas (a, b, d), (c) $^{87}\text{Sr}/^{86}\text{Sr}$ vs. SiO_2 and (e) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. SiO_2 content.

Chapter 7: Discussion

7.1 Introduction

This chapter includes discussion of the source of the melt, defines the parental magma, determines the degree of crustal contamination and defines the rock samples based on their petrography and geochemistry in an attempt to determine the geotectonic setting for the emplacement of the Paleogene mafic to felsic samples of the Fish Creek Mountains and the neighbouring Tobin and Shoshone Ranges.

7.2 Petrological comparisons

Calc-alkaline basaltic andesites and andesites are the most abundant rock types from this area and are common throughout the Great Basin, which includes widespread calc-alkaline rocks of intermediate to silicic composition capped locally by younger basaltic and rhyolitic rocks (Best and Christiansen, 1991). The calc-alkaline samples from the FCM show similarities in both the petrographic and geochemical signatures to the calc-alkaline samples found in the Sulphur Springs Range in central Nevada (Ryskamp et al., 2008), the North Clan Alpine and Stillwater Range of central Nevada (A. Timmermans, pers. comm. 2012), Western Great Basin (WGB) (Ormerod et al., 1988, 1991) and study areas of the Ancestral Cascades Arc (Cousens et al., 2008; Stoffers 2010; Clark 2011).

Petrographic features in common with all areas include sieve texture in phenocrystic plagioclase, oscillatory zoned and partly re-sorbed plagioclase, non-equilibrated hornblende and biotite phenocrysts, plagioclase microlites and the presence of glass in the groundmass. Besides abundant plagioclase phenocrysts, generally other

phases include pyroxenes, K-feldspar, biotite, hornblende, olivine (intermediate samples), and quartz (felsic samples) in decreasing abundances with most phenocrysts are euhedral to subhedral ranging from millimeters to up to 1.5 cm.

Petrographically, the FCM lava flows compare favourably to the Miocene-Pliocene porphyritic lava suite of the Ancestral Cascades defined by Cousens et al. (2008). The porphyritic suite is characterized by abundant, highly zoned, blocky plagioclase megacrysts as large as 1 cm which are commonly reverse or oscillatory zoned and also commonly have sieve texture. This zoning and texture are also commonly found in FCM andesites (Figure 19). Euhedral to subhedral pyroxenes are the next most abundant phenocrysts in both the FCM and Ancestral Cascades typically making up 5-15% of the rock. Amphibole crystals range in size from 1-5 mm and constitute 1-10% of the rock (Cousens et al., 2008). Typically, amphibole crystals have opaque reaction rims that may extend to the core of the grain, where the hornblende has broken down to an anhydrous assemblage of plagioclase, pyroxene, and magnetite. This texture is also found in FCM andesite and dacite samples (Figure 18). Biotite occurs as subhedral phenocrysts (1-3 mm) and can be deep red in thin section due to oxidation and iron oxide inclusions (Cousens et al., 2008). Biotite grains present in FCM samples are typically this colour; however, also contain opaque reaction rims commonly found around amphiboles. The porphyritic lava suite groundmass is dominated by plagioclase, accompanied by Fe-Ti oxides and high SiO₂ glass, typically the common groundmass found in all rock types of the FCM.

The most mafic basaltic andesites of the FCM typically consist of pyroxene, olivine and some plagioclase with a matrix of plagioclase, Fe-Ti oxides, pyroxene and minor glass. Trachytic flow texture is common and defined by elongate plagioclase phenocrysts all of which is commonly also found in the basaltic andesite dikes of the Sulphur Springs Range (Ryskamp et al., 2008). Calc alkaline basalts of the N. Sierra (Stoffers, 2010) are somewhat comparable to the FCM basaltic andesites as well, however N. Sierra basaltic andesites contain greater abundances of olivine and fewer pyroxene phenocrysts, much like the alkali olivine basalts of the WGB of Ormerod et al. (1988).

Andesitic flows of the Sulphur Springs and North Clan Alpine and Stillwater Ranges demonstrate textures disequilibrium textures that could be indicative of magma mixing, including sieve texture of phenocrystic plagioclase, re-sorbed felsic phases, thick reaction rims and coexisting mafic olivine with felsic quartz (Ryskamp et al., 2008; A. Timmermans, pers. comm. 2012). The Sulphur Springs andesite flows; however, contain unstrained quartz and sanidine megacrysts along with olivine and pyroxene which are not present together in the andesitic rocks of the FCM. The quartz phenocrysts are also extensively resorbed and many have reaction rims of clinopyroxene (Ryskamp et al., 2008). Sieve texture in phenocrystic plagioclase, which is present in andesites (Ryskamp et al (2008), A. Timmermans (pers. comm. 2012), Cousens et al (2008) and Ormerod et al (1988, 1991)), is indicative of non-equilibrium crystallization or re-melting and was likely triggered by magma mixing within the system which is very typical of Western Great Basin andesites (Cousens et al., 2008). Resorption is observed predominately in plagioclase. Amphibole and biotite show evidence of being in disequilibrium as they

have thick opaque reaction rims, suggesting that either the melt degassed beneath the volcano or the magma ascended slowly across the amphibole and biotite stability fields (Rutherford et al., 1998). Andesitic sample 10-BV-08 contains about 5% olivine, atypical for andesitic lavas of the western USA and could represent mixing of a more mafic magma (containing olivine and pyroxene) with a more felsic magma (containing K-feldspar). Zonation was also commonly found on many of the large plagioclase phenocrysts indicating possible recharge by magmas from below as the recharged magmas may have different Ca/Na ratios.

The dacitic lava domes, basaltic andesite dikes, andesite dikes and lava flows, and flow banded rhyolite flows of the Sulphur Springs Range of central Nevada are the most petrographically comparable to the samples of the FCM. The Dacitic lava domes resemble the dacitic-rhyolite dome of sample 10-BV-48 (Figure 12) sampled from the Shoshone Range. They both have distinctively large plagioclase phenocrysts with lesser amounts of quartz, K-feldspar and pyroxene, along with amphibole, biotite, and Fe-Ti oxides. Alteration is present in both types including chlorite, clay minerals, and iron stains where biotite is commonly altered to chlorite and feldspars to sericite (Ryskamp et al., 2008). Flow banded rhyolites of East Sulphur Springs are similar to the crystal poor rhyolites of the FCM. They both contain extremely small phenocrysts of quartz and plagioclase in a glassy matrix (Ryskamp et al., 2008).

Chlorite alteration is found throughout most samples as an alteration product of mafic minerals such as pyroxene and amphibole; however it is most commonly found around biotite. In this environment chlorite may be present as a metasomatism product via the addition of Fe, Mg or other compounds into the rock mass. In thin section,

chlorite is predominately Mg-rich based on its first order interference colours in contrast to anomalous blue in the case of Fe-rich chlorite. Sericitic alteration often found on plagioclase phenocrysts implies acidic conditions which may be associated by the passage of hydrothermal fluids.

Whereas the petrography may be insightful regarding crustal residency time of the magmas it cannot be relied upon for a means of determining the source of the basaltic andesites found in the FCM.

7.3 Geochemical comparisons

By analyzing geochemical trends we can begin to deduce what the main lava type is within the FCM. As noted in chapter 6, lavas from the FCM and surrounding Tobin and Shoshone Ranges are all calc-alkaline and generally more evolved than Paleogene lavas from the Sierra Nevada (Cousens et al., 2008; Stoffers, 2010, Clark, 2011), and most resemble the andesite, dacite and rhyolite flows from the Sulphur Springs Range in central Nevada (Ryskamp et al., 2008).

The majority of the FCM samples are all highly potassic plotting in the high-K to shoshonite series on the K_2O vs. SiO_2 diagram of Peccerillo and Taylor (1976) (Figure 34). FCM samples are more potassic than the Sierra Nevada samples (Stoffers 2010 and Cousens et al., 2008) and generally plot in similar ranges as the lava flows from the Sulphur Springs Range in central Nevada. Calc-alkaline affinity, known to be widespread in the central Great Basin (Best et al., 1989) is shown on tectonic discrimination Figure 35 from Wood (1980) where all of FCM, Sulphur Springs and Ancestral Cascades samples plot within the CAB field.

Figure 34: K_2O vs. SiO_2 diagram of Peccerillo and Taylor (1976). FCM samples plot higher than the Ancestral Cascades into similar and higher potassic values than the Sulphur Springs.

Figure 35: Ternary diagram of Th, Hf and Ta discriminating between calc-alkalic basalts (CAB), within plate alkali (WPA), within plate tholeiitic (WPT) and E-MORB, N-MORB and island arc tholeiitic (IAT) (Wood, 1980).

FCM samples exhibit relatively low Mg #'s (48-57 for basaltic andesites) consistent with the Sulphur Springs area. Both volcanic suites have compositions that are generally consistent with a subduction zone origin including low Mg #'s, high K₂O, and similar “spiky” trace element patterns with LILE enrichment and HFSE depletion (Ryskamp et al., 2008). Both regions were also affected by the same Paleogene detachment or roll back of the Farallon plate, which contributed to continental arc magmatism over a wide area of western North America (Ryskamp et al., 2008).

The chemical variations of all rock types are consistent with at least some fractional crystallization of the observed phases; feldspars and mafic silicates and oxides. The FCM volcanic rocks range in composition from olivine bearing basaltic andesite (5% MgO, 112 ppm Ni, and 263 ppm Cr) to rhyolite (with <1% MgO, 3 ppm Ni, and 29 ppm Cr). These steep declines in compatible element concentrations suggest fractional crystallization of mafic mineral phases that have high partition coefficients for these elements. Other compatible elements, such as TiO₂, Fe₂O₃, MgO, CaO, Sc, and V, decrease in concentration as SiO₂ increases (Figure 26). Therefore, fractional crystallization appears to have played a role in the evolution of FCM samples; however, the lack of significant Eu anomaly (Figure 44 and 47) combined with the relative homogeneous incompatible element patterns from basaltic andesite to rhyolite suggest fractional crystallization was not the dominant process in forming magmas of evolved compositions.

Sr-Nd-Pb isotopic studies consistently show that intermediate to silicic rocks contain large proportions of ancient continental crust represented by high ⁸⁷Sr/⁸⁶Sr and low ¹⁴⁴Nd/¹⁴³Nd ratios (Figure 33a). When comparing FCM radiogenic isotopic ratios of

Sr and Nd to the WGB suites and Ancestral Cascades, the FCM flows cannot be related by crystal fractionation due to their large Sr isotopic variance and the difference in incompatible element patterns among the three locations. Crystal fractionation of a melt will not alter the isotopic signature of the residual liquid because isotopes cannot fractionate, yet it can increase the overall incompatible element abundances of the liquid as the silica content increases (Figure 36).

In terms of incompatible elements, all samples have negative Nb anomalies and high concentrations of LILE relative to HFSE, all characteristic of calc-alkaline basalt of the central Great Basin (Figure 30) (Best et al., 1989). This suggests a calc-alkaline magma as our primary lava end member. The enrichment or peaks of the mobile elements Rb, U, K, Pb and Sr compared to the depletions or troughs of the immobile elements (HFSE) could reflect melts of the mantle sources that are metasomatized by hydrous fluids from the subducting slab, as hydrous fluids released off the slab would be enriched in the mobile elements relative to the immobile elements (Best et al., 1989).

Middle to heavy REE patterns do not show a steep garnet stability field pattern and instead show a relatively flat pattern with no significant depletions suggesting that they were generated at depths within the spinel peridotite field. Garnet is known to retain the HREE in its chemical structure so we would see an appreciable depletion in HREE if the melts were generated within the garnet lherzolite field (Borg et al., 1997) (Figure 30). In addition, Cr abundances with the basaltic andesites suggests that a spinel peridotite is the likely source of partial melts given the compatible nature of Cr in garnet.

Figure 36: Radiogenic isotopic $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$ plot showing isotopic Sr and Nd ranges for known studies in the Great Basin. FCM samples are shown and plot within the yellow field, BV = Buffalo Valley, and WGB = Western Great Basin. BV data from Wetmore (2011). Modified from Cousens et al. (2008).

Lavas from the Ancestral Cascades area (N. Sierra, Lassen Volcanic Centre and the Lake Taheo-Reno area) and Sulphur Springs (Ryskamp et al., 2008) also display no noticeable HREE depletions suggesting that they are also melts from the spinel lherzolite field (Stoffers, 2010) (Figure 39).

7.4 Sources of the Parental Melts

Let us first evaluate the possible source of partial melts that could potentially generate the basaltic andesites in the FCM. They are 1) intraplate mantle, 2) mantle wedge, and 3) lithospheric mantle.

Basalts erupted in an intraplate setting, such as late Cenozoic Mojave Desert basalts, are enriched in most incompatible elements and LILE, do not have a Nb anomaly, and have REE patterns decreasing from light to heavy when compared to primitive mantle, consistent with a garnet-bearing lherzolite source (Pearce, 1983; Schmidt et al., 2008). Intraplate Mojave Desert basalts also have higher Nd and lower Sr radiogenic isotopic ratios, similar to those of the Buffalo Valley Pliocene – Quaternary cones found in the FCM (Wetmore, 2011) (Figure 36). The Buffalo Valley cinder cones, located on the northwestern margin of the FCM, are thought to be derived from low degree partial melting of the upper asthenospheric mantle, that also erupted as intraplate basalts (Wetmore, 2011). The lack of an Nb anomaly in both the Mojave Desert and Buffalo Valley cinder cones, consistency with a garnet-bearing lherzolite source in contrast to spinel lherzolite combined with isotopically lower Sr and higher Nd ratios relative to the higher Sr and lower Nd ratios of the FCM, suggest that the FCM lavas

were not derived from an intraplate setting like that of the Mojave Desert basalts and Buffalo Valley cinder cones.

Fluid metasomatized peridotites of the mantle wedge have been proposed to be the main sources of modern Cascade Range melts such as those at Lassen Volcanic centre in the Modern South Cascades (Borg et al., 1997) and in the Ancestral Cascades Arc (Cousens et al., 2008). Figure 36 shows that mantle wedge sources like the Modern South Cascades plot at high $^{143}\text{Nd}/^{144}\text{Nd}$ and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Whereas the Ancestral Cascades also plot in the mantle wedge area, they also extend and overlap into lithospheric mantle sources indicating that the source of the Ancestral Cascades contains both mantle wedge and lithospheric mantle components (Cousens et al., 2008). Borg et al., (1997, 2002) showed that Lassen region calc-alkaline basalts can be defined by the degree of Sr enrichment over other incompatible elements. $\text{Sr}/\text{P}_{\text{pmn}}$ (primitive mantle normalized) can be used instead of LILE/REEs to show the relative enrichment of Sr over middle REE in arc rocks due to Sr addition via fluids from the subducting slab at the time of melting (Borg et al., 1997). Variations in Sr and P are not simply due to fractional crystallization of plagioclase and apatite which can be seen by the lack of Eu or P anomalies in intermediate FCM samples. In Figure 37, FCM basaltic andesites plot in similar ranges as the WGB basalts and Cascadia sediments but at relatively higher initial Sr isotopic ratios and the same relatively low Sr/P values. Borg et al. (1997) state that low Sr/P and high Sr ratios indicates a source from a lithospheric mantle whereas high Sr/P and low Sr ratios indicate a mantle wedge source (South Cascades).

Figure 37: Primitive mantle normalized plot of $\text{Sr}/\text{P}_{\text{pmn}}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ of FCM samples compared WGB and Sierra Nevada field areas. Block A represents Precambrian lithospheric mantle and block B represents mantle wedge with a high fluid component. The increase of $^{87}\text{Sr}/^{86}\text{Sr}$ indicates an increase of lithospheric mantle age. Modified from Cousens et al., 2008.

Borg et al. (2002) also state that a high Sr isotope ratio and low Sr/P component is not a fluid component and must either be an enriched component in the mantle wedge or a lithospheric mantle component. The increase of $^{87}\text{Sr}/^{86}\text{Sr}$ also indicates an increase of lithospheric mantle age. In contrast, the Ancestral Cascades plot at mid ranges for both Sr/P and $^{87}\text{Sr}/^{86}\text{Sr}$ along a mixing line between lithospheric mantle (A) and mantle wedge (B) suggesting a source from both components; however, at lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios suggesting a younger lithosphere component such as Phanerozoic mantle lithosphere (Figure 37; Cousens et al., 2008). The fact that FCM samples plot at much more restricted Sr and Nd isotopic values characteristic of the lithosphere and crust compared to the Modern and Ancestral Cascades combined with lower Sr/P values and higher >0.706 initial $^{87}\text{Sr}/^{86}\text{Sr}$ suggests that FCM sources are not derived from the mantle wedge like those of the Modern South Cascades and a component in the Ancestral Cascades. Instead they are inferred to be derived from enriched components in a ca. 1 Ga Precambrian lithospheric mantle (Figure 37).

The alkali olivine basalts of the WGB do show some similarities to the lavas of the FCM. WGB basalt incompatible element patterns include subduction signatures much like those of the FCM. WGB mantle sources also include low Nd ratios and high initial Sr ratios compared to the Ancestral and Modern Cascades, plotting in similar isotopic ranges of the FCM on the $^{144}\text{Nd}/^{143}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ ratio plot (Figure 36). The WGB has a distinctive subcontinental lithospheric mantle source which was metasomatized by earlier subduction episodes beneath the western margin of the south-western U.S. (Ormerod et al., 1991). Thus magmas derived from this source have incompatible element patterns that resemble modern subduction-related magmas but have much higher $^{87}\text{Sr}/^{86}\text{Sr}$ and

lower $^{143}\text{Nd}/^{144}\text{Nd}$ that reflect the ancient age of the lithosphere much like isotopic ratios found in the FCM. In western North America, the western margin of Precambrian North America is most clearly resolved along the western Idaho Shear zone which is a subvertical mylonitic structural boundary coincident with the abrupt change in initial $^{87}\text{Sr}/^{86}\text{Sr}$ of Mesozoic and Cenozoic magmatic rocks (Figure 38). Here, initial $^{87}\text{Sr}/^{86}\text{Sr}$ increases rapidly from <0.706 in the west to >0.706 in the east across the western Idaho shear zone (Armstrong et al., 1977; Manduca et al., 1992). Continental material west of the cratonic margin in westernmost Idaho and central Nevada consists of Paleozoic-Mesozoic oceanic volcanic arcs, accretionary prism complexes and associated basinal successions accreted to western North America during Middle to Late Jurassic arc-continent collision (Dickinson, 1979). In terms of subduction and primitive mantle signatures (Figure 37), the FCM lava flows are very similar to the WGB enriched lithospheric melts which suggests the primitive mantle source of the FCM has also been metasomatized by earlier subduction episodes. However, in Figure 36, the FCM samples are more restricted and more “crustal”, crossing to the east side of the 0.706 initial Sr line that defines the western margin of Precambrian lithospheric mantle, they are calc-alkaline high-K lavas compared to lower-K WGB alkali lavas, and show evidence of larger volcanic edifices with rare dome collapse and debris flow deposits that are lacking in the WGB (Ormerod et al., 1991).

In terms of REE patterns, the FCM suite and the Sulphur Springs Range are similar to igneous rocks from continental margin subduction zones (Ewart, 1979) with relatively steep slopes and small negative Eu anomalies. Like volcanic rocks erupted in

other subduction settings, all of the rocks have negative Nb and Ti anomalies and positive anomalies for Pb and other LILE on primitive-mantle normalized diagrams (Figure 39). According to Figure 40 (basaltic andesites only), the FCM and Ancestral Cascades both follow similar incompatible element patterns, hinting at similar origin; however overall the FCM flows are more enriched in incompatible elements and the Ancestral Cascades have greater Nb, Ti anomalies and higher MREE to LREE (Sr/P) ratios that represent a mantle wedge component that is not present in FCM samples. Sample 10-BV-42 seems to be more depleted and plots at similar concentrations to the Ancestral Cascades. Variance in the incompatible element patterns for the calc-alkaline samples may be the result of: 1) the degree of partial melting, 2) the varying degree of depletion or enrichment of the mantle wedge from which the above partial melts are possibly generated from and 3) the geochemical variability of the fluids released from the slab.

Figures 31 to 32 illustrate that the samples from the FCM show that fluid from the subducting slab is affecting their geochemistry to a certain degree where as sediment added from the subducted slab are not as important as noted by large range of values for Ce/Pb (1-9) and small ranges for Cs/Rb (0.01-0.08). Pb is enriched in slab fluid therefore low Ce/Pb values represent an increase in fluid component where as Cs is enriched in sediments relative to Rb, therefore low Cs/Rb represent a low sediment component in the FCM.

Tectonic discrimination diagrams of Pearce et al., (1977) which discriminate between different geotectonic settings, intermediate samples from the FCM, Sulphur Springs Range of central Nevada, and the Ancestral Cascades plot within the an orogenic setting (island arc and active continental margin) (Figure 41).

Figure 38: Shows the division of western North America defined by the 0.706 line (large dot-dash line). Precambrian North America lies on the east side of the dashed line (>0.706 Sr) and Paleozoic-Mesozoic oceanic volcanic arcs, accretionary prism complexes and associated basinal successions accreted to western North America during Middle to Late Jurassic arc-continent collision lies on the western side of the dashed line (<0.706). Figure from Streck et al., 1999.

Figure 39: Incompatible element diagrams for the FCM (blue), Sulphur Springs (red), and Ancestral Cascades (green). All values are normalized to primitive mantle (Sun and McDonough, 1988)

Figure 40: Incompatible element patterns for FCM basaltic andesites vs. Ancestral Cascades basaltic andesites normalized to primitive mantle (Sun and McDonough, 1989).

On the discrimination diagrams of Pearce et al., (1984), which are based on Rb, Nb, and Y abundances, the felsic rhyolite and dacite lavas are similar to subduction related volcanic arc granites in central Nevada, the volcanic rocks from the East Sulphur Springs and Eocene volcanic suites from Bingham, Utah, and along the Carlin Trend, Nevada (Figure 42A) (Ryskamp et al., 2008). Based on Zr-Ti-Ce-P systematics, mafic rocks in the FCM are also of a continental arc-type (Figure 42B). The most primitive basaltic andesites of the FCM have relatively low TiO_2 (<1.4 wt%) similar to mafic rocks found in arcs (Figure 26) (Pearce et al., 1977). I conclude that, similar to the Sulphur Springs Range of central Nevada, the FCM suite is related to subduction.

Having established that intraplate and mantle wedge components are not found in the FCM, the basaltic source of the melt was likely derived from partial melting of lithospheric mantle. However, with the large ranges of isotopic compositions observed in the basaltic andesites and andesites, not all intermediate rocks from the FCM can be produced by crystal fractionation of a calc-alkaline basaltic source from the same area so it is clear that the FCM rock types tap a mantle source with variable isotopic composition or are variably contaminated.

Sediments add radiogenic Sr and Pb but non-radiogenic Nd to the overlying mantle wedge, but in different concentrations that ultimately affect the isotopic ratio of these elements in partial melts of the wedge (Elliott et al., 1997). If sediments did not play an important role in magmas of the FCM (low Cs/Rb on Figure 32), they are most likely not the source of the high $^{207}\text{Pb}/^{206}\text{Pb}$ component in the FCM. Figure 32 also suggest an enriched mantle source for the basaltic andesites where Ce/Pb ratios vary from 4 to 10 in contrast to MORB or OIB ratios of 25.

Figure 41: Ternary diagram of FeO^T, MgO and Al₂O₃ discriminating between different geotectonic settings. All samples plot within an orogenic setting (island arc and active continental margin). Fields from Pearce et al (1977).

Figure 42: Tectonic discrimination diagrams for rocks from the FCM compared to the volcanic suits from East Sulphur Spring, other Eocene volcanic suites from Bingham, Utah, and along the Carlin trend, Nevada. (A) Silicic rocks (Pearce et al., 1984). (B) Mafic rocks (Müller and Groves, 2000). Compositions of East Sulphur Springs samples, Bingham, Carlin, Tuscarora and Emigrant Pass are from Ryskamp et al. (2008) and sources within.

In this case, subduction-related metasomatism added Pb but not very much Ce to the lithospheric mantle. Figure 43 from White (1997) plots Pb isotopic ratios in major terrestrial reservoirs on the $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ (Figure 43a) and $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ (Figure 43b) plots. In both plots, FCM Pb isotopic ratios plot within the marine sediments to upper crustal fields suggesting high radiogenic lead from crustal contamination of crustal sediments. Sediments have concentrations of Sr and Nd similar to igneous rocks, but tend to be high in Pb since the sediments are derived from older crustal rocks. Older crustal rocks tend to have higher Sr, Pb and lower Nd isotopic ratios than the mantle. Crustal rocks also have high U/Pb ratios, which is why crustal rocks tend to have high Pb isotope ratios. Adding U as a fluid component does not help, because $^{207}\text{Pb}/^{204}\text{Pb}$ ratios are low in modern subducting slabs, and the characteristics of FCM rocks is higher $^{207}\text{Pb}/^{204}\text{Pb}$ than typical upper mantle. Thus the source of high ^{207}Pb must be old such as upper crustal rocks or possibly an older and therefore more enriched 1 Ga Precambrian lithospheric mantle source.

In the Great Basin, $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr ratios are lower for high-silica alkali rhyolite and Cascadia sediments than for calc-alkalic rocks (Scott et al., 1971). This suggests a high radiogenic Sr origin of calc-alkalic rocks by partial melting of the enriched mantle in contrast to crustal contamination of high silica crust (Yoder, 1973). If sediments and crustal contaminants are not responsible for the higher Sr, Pb and lower Nd isotopic values found in basaltic andesite samples from the FCM, then there must be some source within the mantle that is responsible for the shift in isotopic values. The importance of a lithospheric mantle source is supported by the observation that FCM region mafic lavas commonly resemble alkaline to potassic, Miocene-Pliocene lavas from

Figure 43: Pb isotope ratios in major terrestrial reservoirs plots for A) $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and B) $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$. Typical lower continental crust and upper continental crust are represented by lower crustal xenoliths and modern marine sediments respectively (these somewhat underestimate the total variance in these reservoirs). MORB and oceanic islands represent the isotopic composition of upper mantle and deep mantle respectively. FCM symbols same as in Figure 25. Figure modified from White (1997).

the Western Great Basin in terms of trace elements and radiogenic isotope ratios (Figure 36 and 37). Cousens et al. (2008) noted that the elevated $^{87}\text{Sr}/^{86}\text{Sr}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ values of the Ancestral Cascades compared to Southern Modern Cascades mafic lavas is consistent with an enriched ancient lithospheric component beneath the Lake Tahoe-Reno area. Higher $^{87}\text{Sr}/^{86}\text{Sr}$ compared to lavas from the Ancestral Cascades areas may represent an older mantle due to higher ratios of Rb/Sr, or it may represent a melt that is more modified by a lithospheric mantle component (Cousens et al., 2008). Ancestral Cascade lavas also have mantle wedge components that are not found in the FCM that add less radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and more radiogenic $^{144}\text{Nd}/^{143}\text{Nd}$ to Ancestral Cascade sources. Therefore, the addition of the mantle wedge component to the Ancestral Cascades mantle source would have brought the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio down. It is also plausible that the lithosphere may have had a thicker metasomatized mantle or that the crustal component may have been thicker in the central Great Basin and may be responsible for the higher Sr isotopic values seen in the FCM Paleogene lavas (Figure 36). A thicker crust beneath the FCM would account for the more radiogenic Sr and less elevated Nd isotopic values compared to the Ancestral Cascades and WGB, as it may allow for greater periods of interaction between the rising magma and may also explain why andesitic lavas are the more dominant lava type. A thicker lithosphere may force basaltic magmas to pond at deeper crustal levels, allowing the magmas to crystal fractionate or mix to more evolved compositions. The increase in thickness would also allow longer periods of interaction between the melt and crust, possibly undergoing additional crustal contamination during magma ascent. The anomalous sample with higher than average radiogenic Sr (10-BV-16) may be derived from a slightly more enriched overall mantle

source and is the only basaltic andesite found in the east portion of the FCM. Over time, the $^{87}\text{Sr}/^{86}\text{Sr}$ in the mantle will increase due to radioactive decay of ^{87}Rb , which may suggest that a reason behind sample 10-BV-16 increased radiogenic Sr compared to the rest is that the mantle source of Horseshoe Basin may be slightly older.

Based on typical subduction signatures, high $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios and lower $\text{Sr}/\text{P}_{\text{pmn}}$ ratios, the calc-alkaline primitive sources for the FCM samples were likely derived from partial melting of an enriched lithospheric mantle source to generate basalt and basaltic andesite. One way to initiate melting in the lithospheric mantle is the addition of hydrous fluids possibly from the subducting Farallon plate, or any mantle wedge derived magmas could cause partial melting or melt-rock reaction with lithospheric mantle peridotite along conduit walls (Yogodzinski et al., 1996). Since FCM isotopic geochemistry does not represent mantle wedge derived magmas typical of the Modern South Cascades flows (Figure 36), heat and fluids necessary to partially melt the lithospheric mantle were derived from (1) hot asthenospheric in flow impinging on the base of the previously metasomatized lithospheric mantle as a result of slab rollback and (2) possible hydrous fluids released from the subducting Farallon Slab. Slab derived fluids could either be derived from the dehydration of the Farallon plate as hot asthenosphere came into contact with the slab due to rollback or during flat-slab subduction where hydrous fluids were added to the base of the lithospheric mantle causing metasomatism. Even though no mantle wedge evidence was found in the FCM, it is still possible that fluids released from the subducting slab, lowered the melting temperature of the overlying mantle wedge. The mantle wedge melt then could have aided to melt the lithosphere but never made it past that.

As hot asthenosphere in-flow impinged on the base of the previously metasomatized lithosphere, it caused partial melting of lithospheric mantle, generating calc-alkalic arc-like magmas that have incompatible element patterns that resemble modern subduction-related magmas but have much higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ that reflect the ancient age of the lithosphere much like isotopic ratios found in the WGB. The relatively mafic basaltic andesitic samples that contain olivine probably evolved from their basaltic parents mainly by fractional crystallization, as indicated by equilibrium textures and by relatively low MgO, CaO (Figure 26) and high abundances of SiO_2 compared with typical basalts. There is also no textural evidence for magma mixing such as disequilibrium textures in the more mafic basaltic andesites and the presence of olivine makes it unlikely that they formed by melting in the crust (Askren et al., 1997). However, basaltic andesite samples 39 and 40 have greater disequilibrium textures and lack olivine, possibly as a result of a greater fluid component or a lower degree of partial melting. Sample 39 and 40 also contain much lower abundances of Cr and Ni. The basaltic parent magmas probably had the same arc-type trace element signature as the andesitic lavas because fractionation of crystals typically found in basalts is not capable of depleting the magmas in Nb and Ta or producing elevated La/Nb and Ba/La ratios (Askren et al., 1997).

7.5 Dacite and Rhyolite genesis

Dacitic and rhyolitic rocks may be formed by either the crystal fractionation of an andesitic melt, which is the dominant process for dacite and rhyolitic production in the Modern Cascades (Hildreth, 2007), or by partially melting a basaltic rock. In order to distinguish between both processes, REE plots and incompatible element patterns for

both dacite and rhyolite are used. Negative Eu, P, Sr and Ba anomalies (Figure 30d and 44) would represent significant crystal fractionation, as plagioclase fractionation will lead to the Eu and to some of the Sr anomaly, whereas fractionation of apatite would produce all four anomalies. Consequently, fractional crystallization appears to have not played a major role in forming magmas of evolved compositions represented by the lack of anomalies in FCM lavas although dacites do have a small Eu anomaly (Figure 44). Rhyolites from the study area were also phenocryst poor relative to the other rock types, which according to Hildreth (2007) favours melt extraction from a pluton sized crystal mush reservoir.

Partial melting of a basaltic parent would produce a higher yet similar incompatible element pattern and would also produce a melt with the same isotopic value as the parental rock. Evidence which supports partial melting of crustal basaltic material to form dacitic magma include a lack of significant Eu anomaly and similar incompatible element pattern the dacites have compared to the basaltic andesites besides higher LREE enrichment found in dacites (Figure 30, 44). Rhyolite REE plots (Figure 44) also show a lack of negative Eu anomaly which suggest partial melting as well; however, from a more evolved source such as basaltic andesite in order to generate a rhyolitic melt. The volcanic dome of sample 10-BV-48 varies from dacite (sample A) to rhyolite (sample B) in composition, which suggests dacitic and rhyolitic magmas were formed around the same time.

Therefore, felsic magmas in the FCM are formed primarily by partial melting of middle crustal metabasaltic rocks which forms dacitic magma and by partial melting of basaltic andesite to form rhyolitic magma (Rapp and Watson, 1995). The heat and fluids

required to melt the crust must have been provided by the arc-like magmas derived from the partial melting of the lithospheric mantle. Given the lower isotopic Pb ratios for dacites and rhyolites (Figure 33b,d) and a different isotopic trend on Figure 33a compared to the basaltic andesites, the felsic rocks were most likely derived from a younger possibly less enriched source such as Phanerozoic lithospheric mantle that was previously emplaced causing partial melting of crustal rocks. This is made possible in the FCM since the FCM lie very close to the western margin of Precambrian lithospheric mantle or the 0.706 Sr line. This hot magma rose buoyantly into the crust, stagnated because of density differences in the crust, and promoted partial melting of the continental crust to form metaluminous silicic magma. Figure 45 and 46 suggest that the felsic FCM samples evolved through young accreted oceanic geosynclinal terranes west of the 0.706 Sr line suggesting metaluminous silicic magma as opposed to peraluminous silicic magma characteristic of Precambrian continental crust to the east. FCM felsic samples plot with Eugeocline and Sierra Nevada batholiths of DePaolo and Farmer (1984) on the Epsilon Nd vs. Epsilon Sr plot (Figure 45) and as metaluminous on Figure 46, both characteristic of the west side of the 0.706 Sr line. Partial melting of middle continental metabasaltic and metabasaltic andesite crust will produce a silicic melt that may have impeded the ascent of more mafic magmas due to density differences resulting in magma mixing.

Figure 44: REE diagrams for dacite and rhyolite FCM samples. All values are normalized to primitive mantle (Sun and McDonough, 1988).

Figure 45: Initial ϵ_{Nd} and ϵ_{Sr} values of granitic rocks in the northern Great Basin divided by the groupings eugeocline, miogeocline and craton. Eugeocline corresponds to the accreted area west of the 0.706 line and Miogeocline and craton correspond to areas above the start of the Precambrian crystalline basement and the Archean Craton itself. FCM symbols are as in Figure 25. Figure modified from DePaolo and Farmer, 1984.

Figure 46: Shands index showing mainly metaluminous compositions for FCM samples. Fields from Shand (1943).

7.6 Andesites

Similarities in the geochemistry between the basaltic andesites and andesites indicate that both are most likely derived from the same mantle source. Comparing the trace element plots for the basaltic andesites and andesites from the FCM (Figure 30b,c), the basaltic andesites show a trace element pattern that is very similar to the andesites; however, with a few exceptions where the overall abundances of LILE and HFSE are lower than the andesites. Compared to the basaltic andesites, the andesites appear to be higher degree partial melts as they generally exhibit lower abundances of MREE and HREE, which behave incompatibly during melting but are diluted as the percentage of melting increases. Hornblende fractionation might also be the cause, however it would have to be a lot of fractionation which does not appear to be the case in FCM andesites represented by the lack of hornblende phenocrysts ($< 3\%$).

Andesites may be produced by crustal fractionation from the basaltic andesites melt in which case their REE patterns (Figure 47) would show a negative Eu anomaly due to plagioclase fractionation, which is not evident, and a higher abundance of other REE relative to basaltic andesites. Due to the lack of significant Eu anomaly, the andesites may be formed primarily by magma mixing between a basaltic andesite melt with mantle-like characteristics and a felsic dacitic melt derived from partial melting of middle crust and/or with a rhyolitic melt derived from previously emplaced mantle-derived melt. On a Nd and Sr isotopic ratio plot (Figure 33a), the Sr isotopic values of the andesites from the FCM show overall more radiogenic Sr values than those of the basaltic andesites.

Figure 47: REE patterns for Andesites normalized to Sun and McDonough (1989).

This leads to the proposal that the andesites may have formed by mixing a more primitive melt that has a more depleted mantle-like isotopic signature (higher $^{143}\text{Nd}/^{144}\text{Nd}$, lower $^{87}\text{Sr}/^{86}\text{Sr}$) with a source that has higher Sr and lower Nd isotopic ratios such as the crust (Cousens et al., 2008). A mixing line between these two end members would pass through most samples in Figure 33a and produce a trace element pattern that exhibits no large negative Eu anomaly. Mixing between a rhyolitic magma and basalt is characteristic of samples found in the Sulphur Springs Range where quartz and olivine phenocrysts coexist together in andesite. No such mineral phases have been found together in the andesitic rocks of the FCM; however in terms of disequilibrium textures, dacitic rocks similar to andesitic rocks display sieve texture and reaction rims indicating disequilibrium mixing or pressure-temperature change rather than equilibrium textures characteristic of fractional crystallization.

Textural evidence has been found to support magma mixing between mafic and felsic magmas. Sieve texture is common on andesitic plagioclase phenocrysts (Figure 19) of the FCM which is indicative of non-equilibrium crystallization and was likely triggered by magma mixing within the system. This texture is also very typical of Ancestral Cascade andesites (Cousens et al., 2008). Other disequilibrium textures include reaction rims and extensive resorption of phases that may have come from a different magma source. Oscillatory zoning of plagioclase present in many samples may also have been related to the injection of a hotter, more juvenile magma into a cooling and crystallizing chamber. The common occurrence of corroded or remelted embayments of the crystal rims accompany many reversals supports this conclusion (Winter, 2001). Mineral assemblages, olivine presence (in andesitic sample 10-BV-08), and trends on

most silica variation diagrams (Ti, Al, Fe, Mg, Ca, Na, and K) suggest it was a member of the basaltic andesite suite. The majority of linear trends observed on two element variation diagrams of compatible and trace element concentrations are more typical of magma mixing as oppose to curved trends and sharp decreases more typical of fractional crystallization (Ryskamp et al., 2008) (Figure 26). When compared to trace element mixing plots of Ryskamp et al. (2008), the lava flows of the FCM plot in similar mixing relationships as the samples from Sulphur Springs for incompatible Ba vs. SiO_2 suggesting they may have formed in a similar fashion; by mixing with silicic crustal magma (Figure 48).

Therefore, it is possible that mainly hot asthenospheric in-flow on the base of the previously metasomatized lithospheric mantle, lowered the melting temperature of the lithospheric mantle, generating calc-alkaline magmas as a result of hydrous partial melting. This hot basaltic magma rose buoyantly into the crust as it experienced rapid decreases in pressure and fractionated olivine but no plagioclase to form basaltic andesites. The basaltic andesites magma then intercepted and mixed with metaluminous silicic melt that was derived from partial melting of the crust from previously emplaced Phanerozoic mantle-derived melt to form mainly biotite-hornblende andesites. As the overlying felsic magma either crystallized at depth or erupted, it allowed more mafic magmas to do the same.

The crustal silicic magma may have prevented the ascent of andesitic lavas because the density of felsic magma (2.45 g/cm^3 for dacitic magma; Whitney & Stormer, 1985) is less than that of typical olivine or hornblende andesites (2.66 g/cm^3 at 54.6% SiO_2 and 2.56 g/cm^3 at 59.4% SiO_2 ; Gill, 1981) .

Figure 48: Ba vs. SiO₂ variation diagram comparing compositions of FCM samples to those of East Sulphur Spring, Carlin, and Bingham volcanic suites. Blue lines show the results of mixing basaltic andesite with rhyolite, red arrows are schematic paths for fractional crystallization of basaltic andesite, and yellow line represents the andesite mixing trend. Figures modified from Ryskamp et al. (2008) and sources within.

Subsequent eruption of felsic magmas then provided unobstructed conduits for the eruption of andesite magmas, as felsic magma chamber emptied or remaining magma crystallized. If emptied, then felsic magma no longer remained to trap andesitic magmas. If crystallized, then the increased density of crystallized dacitic magma (to 2.65 g/cm^3 for a typical granodiorite) allowed hornblende andesite magma to rise and erupt (Askren et al., 1997).

7.7 Evidence of crustal contamination

There exists evidence that the basaltic andesites and andesites may have been affected by crustal contamination due to the presence of plagioclase crustal xenocrysts; however no quartz xenocrysts were found suggesting little to no interaction with Paleozoic metaluminous metasedimentary and granitic crustal rocks that are widespread in the Great Basin (Kistler et al., 1981; Barnes et al., 2001). Assimilation of metaluminous and peraluminous granitoids of North-central Nevada is a possible cause of the distinctions between FCM and WGB $^{87}\text{Sr}/^{86}\text{Sr}$ values. Isotopically, central Nevada granitoids have a high $^{87}\text{Sr}/^{86}\text{Sr}$ component similar to FCM lavas. Figure 49 shows that Rb/Sr increase as magmas become more evolved and show a notable increase in $^{87}\text{Sr}/^{86}\text{Sr}$. Unlike the Ancestral (Tahoe-Reno) and Modern South Cascades, a rough linear mixing relationship is found between FCM basaltic andesite and a Sierran Nevada batholiths-like source in central Nevada. Lower degree partial melting, of central Nevada granitoids would produce liquids potentially with even higher Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ than bulk granitoid, since one of the first phases to melt is high Rb/Sr biotite (Kaczor et al., 1988; Knesel and Davidson, 1996).

Figure 49: Initial Sr isotope ratios vs. Rb/Sr plot for FCM samples plotted along with Sierran Granites, Ancestral Cascades Tahoe-Reno area and Modern South Cascades. FCM symbols are the same as in Figure 25. Sierran Granites, Tahoe-Reno and South Cascades geochemistry from Cousens et al. (2008).

However, lower $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr than granitoids combined with no negative Eu anomaly from the melting of feldspar suggest no assimilation of central Nevada granitoids by FCM magmas (Figure 49; Cousens et al., 2008).

The $^{87}\text{Sr}/^{86}\text{Sr}$ from the FCM are more radiogenic than the Ancestral Cascades and overlap with the most radiogenic lavas of the Western Great Basin suite (Omerod, 1991). The more evolved rocks of the FCM approach crustal $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, suggesting Sr contamination from crustal sources (Figure 36). SiO_2 concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ are useful geochemical tools in helping to identify continental crust as a possible contaminant of ascending magmas. On the SiO_2 vs. $^{87}\text{Sr}/^{86}\text{Sr}$ plot (Figure 50), there is a notable increase in radiogenic Sr with increasing SiO_2 and a notable decrease in radiogenic Nd with increasing SiO_2 (Figure 33e) which may suggest contamination. However, the positive correlation of Figure 50 is more likely to represent mixing of rhyolitic magma with basaltic andesite in order to generate dacites and rhyolites. The positive correlation that Pb and Th have with SiO_2 may indicate that FCM samples experienced some Pb and Th contamination in contrast to other incompatible elements such as Ba or Zr which vary irregularly (Figure 28).

Figure 50: SiO_2 vs. radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ plot. L mantle = Lithospheric mantle. Symbols are as in Figure 25.

7.8 Mantle beneath the FCM and tectonic model

Tectonic setting discrimination diagrams clearly show that FCM lava flows are calc-alkaline on Figure 35 from Wood (1980) which are widespread in the central Great Basin (Best et al., 1989). Figures 43A (felsic) and B (mafic) show that samples are related to a continental volcanic arc suite presumably related to the rollback of the subducting Farallon plate under North America.

Late Cretaceous subduction related magmatism ended near Lake Tahoe around 90 Ma as the magmatic activity of the Sierra Nevada batholith shifted eastward, most likely as a result of shallowing of the dip of the subducting Farallon slab (Dickinson and Snyder, 1978; Lipman, 1992). During flat slab subduction, metasomatism changed the base of the lithosphere and added components enriched in LILE, LREE, Zr and depleted in Nb and Ti, which are characteristics expected in components derived from the subducting slab (Rivalenti et al., 2007). Given the Ar-Ar ages determined from the basaltic andesite to rhyolites of the FCM, they span approximately 1 m.y. from 33.3 to 34.3 Ma during the Paleogene time period. At this time the Great Basin experienced a significant change in magmatism when the Farallon plate detached from the lithosphere in a southwestward sweeping fashion allowing hot asthenospheric mantle to flow between the subducting slab and the lithospheric mantle (Best and Christiansen, 1991). This created a continental magmatic arc that also swept to the SW producing magma compositions ranging from basaltic andesite to rhyolite (or their intrusive equivalents) (Severinghaus and Atwater, 1990). Middle Paleogene magmatism may have resulted from (1) decompression melting of hot mantle as a result of small scale convection above the subducting Farallon plate, (2) dehydration of the Farallon plate, which caused

hydrous melting of the mantle wedge, and (3) heating of metasomatized lithospheric mantle by hot asthenospheric mantle or by wedge derived magma. These mafic, mantle-derived magmas ascended and powered crustal magma systems in which more felsic magmas evolved by partial melting continental crust and magma mixing (Ressel and Henry, 2006).

From studying the most primitive basaltic andesites and andesites, it is clear that there is a distinctive geochemical domain in the mantle. These distinctive signatures are identifiable when looking at the incompatible element plots and radiogenic isotopes (Figure 30, 34a,b,d). There are mainly two geochemical components present in the mantle beneath the FCM. These are an enriched ancient Precambrian lithospheric mantle source metasomatized by previous subduction episodes to generate the basaltic andesites combined with mixing of crustal magma which has parental melts derived from a less enriched younger Phanerozoic mantle lithosphere.

FCM Eocene-Oligocene lava flows form a dominantly high K, calc-alkaline suite with low Mg #'s similar to those found in subduction settings worldwide. The volcanic and subvolcanic rocks range from olivine-bearing, high-MgO basaltic andesite to high silica rhyolite. The intermediate to silicic rocks have subduction related trace element signatures with Nb-Ti depletions and enrichment of Pb, also similar to those formed at convergent margins, and have highly radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$.

This leads to the assumption that melts are being stored as batch melts in the middle crust after partially melted from enriched ancient lithospheric mantle and

undergoing fractional crystallization processes from basalt to basaltic andesite given the presence of olivine, Cr-spinel and equilibrium textures in basaltic andesites.

These characteristics suggest the magma system was rooted in a subduction zone setting that formed as the Farallon plate steepened during the Paleogene where heating and partial melting of lithospheric mantle by hot asthenospheric mantle and possibly wedge derived magmas was the likely cause of middle Paleogene magmatism in the FCM (Figure 51). As hot asthenosphere came in contact with the base of the previously thick metasomatized lithospheric mantle, arc-like magmas with elevated $^{87}\text{Sr}/^{86}\text{Sr}$, low $^{143}\text{Nd}/^{144}\text{Nd}$, and high $^{207}\text{Pb}/^{204}\text{Pb}$ were generated as a result of hydrous partial melting of Precambrian lithospheric mantle. Small proportions of mantle wedge derived magma could have been generated from slab fluids caused by the dehydration of the subducting Farallon slab however; FCM geochemistry does not show evidence to support this. It is possible that this mantle wedge derived magma helped to partially melt the lithospheric mantle but did not make it past the base of the lithosphere. The more mafic sub-alkaline samples 16 and 42 are most likely derived from higher degrees of partial melts given their more subalkaline nature where as samples 39 and 40 are alkaline.

This hot magma ascended buoyantly into the crust as it experienced rapid decreases in pressure, intercepted and mixed with previously emplaced rhyolitic magma derived by partial melting of crustal basaltic andesites to form mainly biotite-hornblende dacites and andesites. Petrographic evidence, linear trends on variation diagrams, homogeneous incompatible element plots and overlapping isotope plots support this conclusion. In addition, evidence of magma mixing involving alkaline olivine bearing magmas is widespread in the central Great Basin. During the Eocene, mafic alkaline

magmas intercepted and mixed with evolved calc-alkaline magmas in shallow subvolcanic settings (Ryskamp et al., 2008). Structural boundaries including Proterozoic age basement-penetrating faults, Paleozoic and Mesozoic thrust faults, and magma-generated fractures, guided magma emplacement routes through thick central Great Basin continental crust and controlled levels of stagnation. As the more evolved magmas crystallized at depth or erupted, the increased density of crystallized dacitic magma or eruption of overlying felsic magma allowed andesite and basaltic andesite magma to rise and erupt. As a result of crustal trapping, only a small fraction of the mafic magma was able to erupt (Figure 51).

Figure 51: Plate tectonic reconstruction of the western United States modified from Ryskamp et al. (2008) during the Paleogene Period. Gt—Golconda thrust; RMT—Roberts Mountains thrust. (A) A period of low-angle subduction during the Cretaceous and early Paleogene metasomatises the lithospheric mantle, thickens the crust and magma generation is stalled. (B) The subducting oceanic lithosphere rolled back inducing asthenosphere counterflow. As it heated, the slab dehydrated and initiated the production of hydrous, oxidized fluids in the overlying mantle wedge over a broad area (green). Hot asthenospheric mantle impinges on the base of the enriched lithospheric mantle near the edge of the Precambrian continental crust inducing partial melting of lithospheric mantle. The melts ascend and undergo fractional crystallization to form basaltic andesite from basalt. Basaltic andesite magma intercepts and mixes with rhyolitic magma derived from partial melting of crustal metabasaltic andesites to form dacites and andesites. Structural boundaries including faults and magma generated fractures guided magma emplacement routes and controlled levels of stagnation.

Conclusions

Paleogene volcanic rocks collected from the Fish Creek Mountains and surrounding Tobin and Shoshone Range area in north-central Nevada, are calc-alkaline basaltic andesites through to rhyolites and range in age from 33.3-34.3 Ma. Volcanic edifices include lava flows, domes and volcanoclastic flows which typically include phenocrysts of plagioclase, pyroxene, potassium feldspar, olivine (intermediate samples) and quartz (felsic samples), with less common phenocrysts of hornblende, biotite, and an unidentified opaque mineral, probably a Fe-Ti oxide.

FCM Paleogene magmatism is related to the southwestward sweep of magmatism due to the increased angle of subduction of the ancient oceanic Farallon plate under North America. There are mainly two geochemical components present in the mantle beneath the FCM, which are an enriched ancient Precambrian lithospheric mantle source and a younger less enriched Phanerozoic lithospheric mantle responsible for the partial melts of rhyolitic crustal magma.

Major and trace element geochemistry suggest that FCM volcanic compositions are generally consistent with a subduction zone origin including low Mg #'s, high K₂O, negative Nb and Ti anomalies and high concentrations of LILE to HFSE. The enrichment of mobile elements (LILE) to depletions of immobile elements (HFSE) reflect melts of the mantle sources that are metasomatized by hydrous fluids from the subducting slab. Middle to heavy REE patterns suggest the melt was generated at depths within the spinel lherzolite field (<75 km depth).

Crustal contamination processes in the FCM magmas did not play an important role and are not the source of high $^{87}\text{Sr}/^{86}\text{Sr}$, $^{207}\text{Pb}/^{206}\text{Pb}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ in the FCM. Rather, low $\text{Sr}/\text{P}_{\text{pmn}}$ values and high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios suggest the FCM suite to be derived from enriched components in a 1 Ga Precambrian lithospheric mantle which was metasomatized by earlier subduction episodes, similar to the source of the WGB. If an enriched ancient lithospheric mantle source existed beneath the FCM region as in the WGB, magmas derived from this source would have incompatible element patterns that resemble modern subduction-related magmas but have higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ that reflect the ancient age of the lithosphere.

Samples from the FCM exhibit values of $^{87}\text{Sr}/^{86}\text{Sr}$ that are slightly more radiogenic compared to the Ancestral Cascades (Cousens et al., (2008) and most of the Western Great Basin (Ormerod et al., 1991). This is because there are no mantle wedge components in the FCM and possibly an older lithospheric mantle. It is also plausible that the lithospheric component may have been thicker in the central Great Basin as a result of flat-slab subduction and may be responsible for the higher Sr isotopic values seen in the FCM Paleogene lavas. A thicker lithosphere beneath the FCM would account for the more radiogenic Sr and less elevated Nd isotopic values compared to the Ancestral Cascades and WGB, as it may allow for greater periods of interaction between the rising magma and may also explain why andesitic lavas are the more dominant lava type. Therefore, partial melting of an enriched Precambrian lithospheric source with no mantle wedge component could have occurred in order to get the elevated $^{87}\text{Sr}/^{86}\text{Sr}$, low $^{143}\text{Nd}/^{144}\text{Nd}$, and high $^{207}\text{Pb}/^{204}\text{Pb}$ values seen in the FCM basaltic andesites through to rhyolites.

Fractional crystallization appears to have played a minor role due to the lack of significant Eu anomaly in FCM REE plots and suggests it was not the dominant process in forming magmas of evolved compositions. Evidence which supports partial melting of crustal basaltic andesites to form rhyolitic magma followed by subsequent magma mixing between rhyolite and basaltic andesite include the lack of significant Eu anomaly that would result from plagioclase fractionation, the homogeneous incompatible element patterns from basaltic andesite to rhyolite and the overlap between isotopic ratios. The heat and fluids required to melt the crust must have been provided by the arc-like magmas derived from the partial melting of the Precambrian lithospheric mantle. This hot magma rose buoyantly into the crust as it fractionated from basalt to basaltic andesite, intercepted and mixed with a rhyolitic magma chamber derived from previously emplaced Phanerozoic mantle-derived melts to form biotite-hornblende andesites and dacites. The linear trends observed on two element variation diagrams of compatible and trace element concentrations are also more typical of magma mixing as oppose to curved trends and sharp decreases more typical of fractional crystallization. The melt then rose into shallower and eventually erupted as cones (10-BV-48) or flows.

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Appendix A: Rock descriptions and TAS name with sample number and location.

UTM: NAD 27, Zone 11

Sample #	Hand Specimen Description	Easting	Northing
10-BV-05	Andesite. Taken from part of the float below an outcrop. Dark grey-black matrix. Dark green altered mafic minerals 5-8%, feldspar laths 1-2mm crystals 10-15% and olivine (2%). Approximately 1.5km from Jersey summit	463137E	4458312N
10-BV-06	Trachydacite. Taken from a weathered outcrop of plag-phyric or flow top lava. Altered feldspar laths in a flow banded reddish-brown matrix, 5-10% plagioclase laths 2-3 mm, iron stained fractures. Located south of Buffalo Valley cones.	465592E	4459389N
10-BV-07	Trachyandesite. From part of the float beneath an outcrop. Dark grey glassy matrix, plag-phyric with 20% plagioclase laths 1mm with minor	465435E	4452055N

	olivine (1-2%) and pyroxene. Brown-red weathered surface, massive and non-vesicular. Dated at 33.3 +/- 0.05 Ma. Located in Cow Canyon		
10-BV-08	Trachyandesite. Dark grey plagioclase rich matrix with 20-25% plagioclase laths of 1-2 mm, minor pyroxene and rare large hornblende 4-5 mm. Located in Cow Canyon at a higher elevation than 07.	465256E	4451873N
10-BV-09	Trachyandesite taken from a jointed outcrop. Fine grained dark grey to black matrix with pyroxene phenocrysts. Outcrop is massive and non-vesicular with flat plagioclase laths showing trachytic texture. Located in Cow Canyon.	465948E	4451890N
10-BV-10	Trachyandesite exposed in an old river bed. Outcrop is jointed and massive with a grey to dark grey matrix and an oxidized weathered surface. Contains medium grained pyroxenes (3-5mm)	465950E	4451892N

	and some olivine with elongated plagioclase (1mm). Located in Cow Canyon.		
10-BV-11	Trachydacite taken from a jointed outcrop. Light grey massive and non vesicular matrix with green altered mafic minerals (25%), blocky laths of plagioclase (35%) up to 15mm and chloritized biotite (1-2%) up to 2mm. Located in Jersey Canyon.	462113E	4446107N
10-BV-12A	Rhyolite. Taken from a volcanic breccia bedded outcrop with some planar features. Fine grained dark grey matrix with elongated euhedral plagioclase laths (20%) up to 1-2mm. Some fine grained plagioclase, pyrite grains (1%) and pyroxene (8%). Located in upper Jersey Canyon.	463056E	4446260N
10-BV-12B	Trachyandesite. Taken from a volcanic breccia outcrop. Lighter grey partially weathered sample of matrix, plagioclase feldspar rich (40%), quartz	463056E	4446260N

	and accessory red garnet. Also located in the upper Jersey Canyon.		
10-BV-16	Basaltic Andesite taken from an angular outcrop with some float. Black fine grained matrix with pyroxene clusters (glassy dark green 3mm) olivine and blocky plagioclase (1%) up to 1mm. This flow seems to be restricted to this part, flanked by both sides by glassy flows of rhyolitic 10-BV-15.	480523E	4459155N
10-BV-29	Trachyandesite taken from a bedded and sheared outcrop. Beds strike east and dip 10° N. Porphyritic andesite with elongated euhedral plagioclase up to 5 mm (25%) and 6-7 mm hornblende (1-5%) in a dark black matrix. Weathered outcrop with altered mafic minerals and some iron stained fractures. Located at the south end of Horseshoe Basin.	488461E	4457888N

10-BV-30	Trachyandesite taken from the flows overlying those of sample 10-BV-29. Light grey-green matrix with 25% plagioclase, hornblende, 15% pyroxene. Outcrop beds are 2-5 cm thick. Located at the southern end of Horseshoe Basin.	488050E	4457894N
10-BV-31	Trachyandesite taken from a massive outcrop with no apparent bedding. Glassy medium grey matrix, round to elongated vesicles <1mm (5%), and euhedral elongated feldspars (15%) with Carlsbad twinning. Sample is less crystalline than previous sample 10-BV-30. Located in the south end of Horseshoe Basin.	486490E	4459327N
10-BV-32	Trachyandesite taken further up from previous 10-BV-31. Glassy black matrix with 15% euhedral plagioclase up to 5 mm and 5-10% hornblende laths up to 2 mm. Non-vesicular jointed outcrop with little weathering.	486292E	4459257N

	Located in the south end of Horseshoe Basin.		
10-BV-34A	Trachyandesite boulder (20x25cm) embedded in FCMT welded outcrop. 15% plagioclase feldspar phenocrysts in a brown-maroon matrix.	469439E	4459655N
10-BV-34B	Trachydacite boulder. Another volcanic lithic fragment in FCMT, 5cm in diameter, with 15% plagioclase crystals.	469439E	4459655N
10-BV-39	Basaltic trachyandesite. Maroon matrix with trachytic texture of plagioclase euhedral laths (15%), outcrop is very weathered.	454978E	4457058N
10-BV-40	Basaltic trachyandesite. Fine grained maroon matrix with 8% euhedral hornblende phenocrysts (2-5 mm to a cm megacrysts) and 15% euhedral plagioclase laths up to 2 mm. Located in Golconda Canyon, Tobin Range.	453891E	4458420N
10-BV-41	Trachyandesite. Sample is more finely	453276E	4459784N

	<p>porphyritic and lighter in colour than previous with a volcanic matrix.</p> <p>Outcrop is vertically jointed, weathered, volcanoclastic and vesicular. Phenocrysts range from 1-4 mm of 5% euhedral hornblende and 10% euhedral plagioclase. Located further up the Golconda Canyon.</p>		
10-BV-42	<p>Basaltic andesite overlain by 33.8 Ma Caetano Tuff. Dark grey vertically jointed outcrop. Phenocrysts of plagioclase and olivine weathered to iddingsite. Black matrix with fine grained plagioclase, olivine and green altered mafic minerals. Located at the west end of the Golconda road.</p>	11449013E	4463684N
10-BV-44	<p>Trachydacite taken from a fresh platy outcrop. Red-dark grey glassy matrix, nearly aphyric, with 15% euhedral plagioclase laths (3-5mm), pyroxenes and possible hornblende up to 1mm. Minor vesicles. Located off of</p>	11477163E	4424042N

	Highway 305 in the southern Shoshone Mountains and Red Butte area.		
10-BV-45	Trachyandesite taken from poorly exposed, crudely columnar jointed outcrop. Fine grained, very glassy black matrix with 15% plagioclase (1-3mm), trace anhedral olivine <1mm weathered to brown (5%) and fine grained pyroxene and olivine. Located North of Red Butte.	11474768E	4426305N
10-BV-46A	Andesite taken from a platy weathered outcrop. Light grey black matrix with phenocrysts up to 1 mm of euhedral plagioclase (15%) and olivine and 1 mm hornblende. Located on the flank of Red Butte.	474725E	4425612N
10-BV-46B	Trachydacite with more elongated feldspars than 10-BV-46. Phenocrysts are not as well preserved in a dark grey altered matrix. May contain	474725E	4425612N

	plagioclase megacrysts.		
10-BV-47	<p>Rhyolite-trachydacite taken from a very fissile outcrop. Beds dip west.</p> <p>Euhedral plagioclase 1-4 mm (15%), euhedral biotite (10%), some chloritized, and pyroxene phenocrysts present in a medium grey crystalline matrix of plagioclase and fine grained hornblende, biotite and pyroxenes.</p> <p>Dated at 33.82 +/- 0.14 Ma. Located on a ridge north of Red Butte area.</p>	474807E	4425374N
10-BV-48A	<p>Trachydacite similar to jointed basalts from a dacitic dome. Contains shear zones more typical of viscous flows.</p> <p>Very glassy aphanitic matrix with few 1 mm plagioclase phenocrysts (10%) with trace biotite, olivine and pyroxene.</p>	474713E	4424929N
10-BV-48B	<p>Rhyolite taken from a platy moderately weathered outcrop from the dome. Contains spherulites, an indication of divitrification. Also</p>	474713E	4424929N

	contains phenocrysts of plagioclase, quartz and biotite in a fine grained plagioclase groundmass. 28.9 Ma Campbell Creek Tuff is vertical adjacent to the dacite with a near vertical flow attitude making the dacite a paleo-wall rock. Located at a lower outcrop of Red Butte.		
10-BV-49	Trachyandesite taken from a platy moderately weathered outcrop. Light grey matrix with 35% plagioclase, 1-4 mm blocky euhedral plagioclase phenocrysts, 5% hornblende and trace 1-2mm euhedral pyroxene. Located on the East side of Highway 305.	492020E	4452583N
H10-53	Rhyolitic Dome with sanidine. Sampled and dated by Chris Henry. Dated at 34.24 +/- 0.05 Ma.	480460E	4459215N

Appendix B: Thin Section Summary

Sample #	Rock type	Phenocrysts (%)							Groundmass	Xenocrysts k-spar, carbonate lithics
		Ol	Pyx	Plag	Amph	Bio	K-spar	Qtz	Opq	
10-BV-05	Andesite	-	5	10	-	3	10	-	1	too fg to tell, glassy
10-BV-06	Trachy-dacite	-	1	2	-	1	1	-	-	too fg to tell, glassy and vesicular
10-BV-07	Trachy-andesite	5	5	20	2	1	-	-	2	fg plag and pyx
10-BV-08	Trachy-andesite	5	5	15	1	-	1	-	1	vesicular, fg plag
10-BV-09	Trachy-andesite	-	5	15	-	-	8	-	1	fg plag, k-spar, qtz
10-BV-10	Trachy-andesite	2	8	15	-	-	5	-	1	fg plag, k-spar, qtz
10-BV-11	Trachy-dacite	-	2	20	-	5	5	-	1	vesicular, glassy-vfg plag, k-spar
10-BV-12A	Rhyolite	-	-	5	-	5	1	2	2	vesicular, glassy, vfg plag
10-BV-12B	Andesite	-	2	8	1	1	5	1	1	too fine to tell, glassy
10-BV-16	Basaltic andesite	5	-	1	-	-	-	-	3	fg plag, ol
10-BV-29	Trachy-andesite	2	10	15	-	-	3	-	1	too fine to tell, glassy and vesicular
10-BV-30	Trachy-andesite	1	8	15	1	1	3	-	1	vesicular, fg plag, qtz, k-spar
10-BV-31	Trachy-andesite	1	8	10	-	-	2	-	1	vesicular (5%) with glassy vfg plag
10-BV-32	Trachy-andesite	1	10	12	-	-	2	-	1	glassy, vfg plag
10-BV-34A	Trachy-andesite	-	1	8	-	1	1	2	1	vfg plag
10-BV-34B	Trachy-dacite	-	-	10	-	1	2	-	2	vesicular, glassy vfg felsics
10-BV-39	Basaltic andesite	-	8	20	-	-	2	-	2	fg felsics of mostly plag, vesicular

Appendix B continued: Thin Section Summary

Sample #	Rock type	Ol	Pyx	Plag	Amph	Phenocrysts (%)			Qtz	Opq	Groundmass	Xenocrysts
						K-	Bio	spar				
10-BV-40	Basaltic andesite	-	8	15	3	3	3	2	-	2	vfg felsics	rounded k-spar and plag
10-BV-41	Andesite	-	8	15	3	3	3	2	-	2	vesicular, vfg felsics	carbonate lithics
10-BV-42	Basaltic andesite	5	-	10	-	-	-	-	-	-	m.g plag and opq	rounded plag
10-BV-44	Trachy-dacite	-	5	5	-	3	-	-	-	1	vesicular, glassy vfg plag	
10-BV-45	Trachy-andesite	-	2	20	-	-	-	-	-	-	too fg to tell, glassy	
10-BV-46A	Andesite	-	5	15	-	-	-	1	1	1	too fg to tell, glassy	round plag, lithic clasts
10-BV-46B	Trachy-dacite	-	2	5	-	1	-	-	-	1	vfg felsics	
10-BV-47	Rhyolite	-	2	10	-	8	2	2	1	1	plag and too fg to tell, glassy	
10-BV-48A	Trachy-dacite	-	5	2	-	8	2	2	2	1	plag and too fg to tell, glassy	rounded k-spar
10-BV-48B	Rhyolite	-	5	5	-	5	2	2	1	1	plag and too fg to tell, glassy	felsic lithics
10-BV-49	Trachy-andesite	-	10	15	3	3	2	2	-	1	vfg felsics	

Notes: Ol = olivine, pyx = pyroxene, plag = plagioclase, amph = amphibole, bio = biotite, k-spar = potassium feldspar, qtz = quartz, opq = opaques, - = no phenocrysts were present in thin section

Appendix C: X-ray Fluorescence major element data collected in this study

Sample Name	SiO₂	TiO₂	Al₂O₃	Fe₂O₃ Total	MnO	MgO
Units	wt%	wt%	wt%	wt%	wt%	wt%
Method	XRF	XRF	XRF	XRF	XRF	XRF
Detection Limit	0.01	0.01	0.01	0.01	0.01	0.01
10-BV-05	57.1	0.78	13.91	6.18	0.16	2.07
10-BV-06	66.28	0.62	14.97	4.71	0.05	0.41
10-BV-07	58.57	0.78	16.99	6.06	0.1	2.19
10-BV-08	58.58	0.8	17.03	6.08	0.09	2.01
10-BV-09	56.26	0.88	18.71	5.71	0.09	1.41
10-BV-10	55.84	0.98	16.57	6.74	0.13	2.8
10-BV-11	63.29	0.7	15.13	4.62	0.09	1.83
10-BV-12A	67.84	0.42	14.76	3.67	0.02	0.47
10-BV-12B	54.7	0.98	15.59	6.62	0.11	3.64
10-BV-16	54.89	1.42	15.38	8.57	0.15	5.08
10-BV-29	60.67	0.84	15.21	6.47	0.09	2.92
10-BV-30	60.57	0.77	15.56	6.21	0.08	2.23
10-BV-31	59.28	0.83	15.31	6.47	0.11	2.98
10-BV-32	59.78	0.83	15.2	6.65	0.11	3.17
10-BV-32	59.46	0.9	15.12	6.89	0.12	3.24
10-BV-34A	58.19	1	18.65	5.75	0.04	0.55
10-BV-34B	59.46	0.83	16.93	5.91	0.03	0.43
10-BV-39	52.92	1.06	16.37	8.04	0.14	3.69
10-BV-40	51.93	1.04	17.18	8.89	0.14	3.79
10-BV-41	61.02	0.88	14.43	6.69	0.09	1.36
10-BV-42	52.4	1.33	15.23	9.31	0.15	4.85
10-BV-44	61.23	0.6	15.24	5.1	0.11	2.15
10-BV-45	58.42	1.25	16.31	7.74	0.12	1.84
10-BV-46A	56.05	0.96	15.34	6.43	0.12	3.19
10-BV-46B	63.67	0.56	16.04	4.64	0.02	0.7
10-BV-47	68.21	0.41	14.99	3.26	0.03	0.72
10-BV-48A	67.12	0.49	14.84	3.51	0.05	1.06
10-BV-48B	67.7	0.41	14.63	2.89	0.03	0.66
10-BV-49	61.07	0.77	15.27	6.1	0.08	2.63
H-10-53	70.88	0.297	14.06	0.918	0.01	0.17
00-LT-2*	50.95	2.41	13.51	13.33	0.24	3.94
Avg deviation**	0.25	0.02	0.07	0.05	0.00	0.03

*Sample 00-LT-2 is an internal rock standard

**Average deviation for 6 runs of 00-LT-2

Appendix C continued: X-ray Fluorescence major element data collected in this study

Sample Name	CaO	Na2O	K2O	P2O5	LOI	Total
Units	wt%	wt%	wt%	wt%	wt%	
Method	XRF	XRF	XRF	XRF	XRF	
Detection Limit	0.01	0.01	0.01	0.01	0.05	
10-BV-05	7.32	2.54	2.92	0.2	6.81	100
10-BV-06	1.98	3.49	5.25	0.18	1.45	99.38
10-BV-07	5.21	4.16	3.31	0.49	1.71	99.57
10-BV-08	5.37	4.01	3.41	0.53	1.11	99.02
10-BV-09	6.73	3.5	3.33	0.38	3.09	100.08
10-BV-10	6.85	3.64	3.03	0.51	2.64	99.71
10-BV-11	2.38	4.28	4.65	0.3	1.43	98.69
10-BV-12A	2.74	3.53	5	0.22	1.56	100.22
10-BV-12B	5.34	1	4.24	0.39	7.18	99.79
10-BV-16	7.84	2.89	2.19	0.47	1.17	100.04
10-BV-29	5.27	3.03	4.09	0.36	0.69	99.65
10-BV-30	4.98	3.2	4.2	0.36	0.89	99.05
10-BV-31	5.36	2.99	3.87	0.36	1.78	99.33
10-BV-32	5.47	2.91	3.96	0.37	1.05	99.51
10-BV-32	5.55	2.97	3.95	0.39	0.9	99.5
10-BV-34A	5.2	3.24	3.65	0.4	2.17	98.84
10-BV-34B	4.26	3.76	4.6	0.46	2	98.67
10-BV-39	6.99	3.56	2.88	0.62	2.56	98.83
10-BV-40	7.24	3.52	3.09	0.59	2.29	99.7
10-BV-41	5.95	3.4	3.39	0.53	1.59	99.35
10-BV-42	8.98	3.21	0.84	0.24	2.85	99.39
10-BV-44	5.13	3.45	4.3	0.33	1.13	98.78
10-BV-45	5.31	3.74	3.03	0.55	1.17	99.48
10-BV-46A	6.98	1.9	3.34	0.37	4.61	99.29
10-BV-46B	3.68	3.58	4.74	0.3	1.45	99.39
10-BV-47	2.63	3.58	5.11	0.21	1.04	100.2
10-BV-48A	3.6	3.45	4.87	0.25	0.98	100.21
10-BV-48B	3.02	3.42	4.99	0.22	1.54	99.52
10-BV-49	5.29	3.04	4.23	0.36	0.6	99.44
H-10-53	1.91	3.09	4.9	0.09	1.69	98.01
00-LT-2*	7.67	3.24	1.92	1.19	1.06	99.47
Avg deviation**	0.12	0.04	0.03	0.02	0.29	

Appendix D: ICP-MS trace element data collected in this study

Sample #	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho
Units	ppm	ppm	ppm	ppm	ppm	ppm	Ppm	ppm	ppm	ppm
Method	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS
10-BV-05	44.98	83.52	9.43	34.38	6.28	1.54	5.24	0.735	4.09	0.76
10-BV-06	54.36	101.67	12.44	46.35	8.49	1.83	6.69	0.956	5.31	1.02
10-BV-07	49.2	94.83	11.3	42.71	7.38	2	5.58	0.718	3.73	0.68
10-BV-08	50.09	97.61	11.74	44.4	7.7	2.08	5.85	0.758	3.86	0.71
10-BV-09	43.96	84.7	10.18	38.68	7.08	1.94	5.83	0.829	4.49	0.87
10-BV-10	40.91	81.61	10.26	41.49	8.12	2.19	6.58	0.88	4.68	0.87
10-BV-11	60.06	110.78	12.42	43.24	7.24	1.59	5.3	0.707	3.61	0.66
10-BV-12A	45.27	78.05	9.19	32.71	5.82	1.34	4.7	0.669	3.76	0.78
10-BV-12B	44.54	86.44	10.38	39.55	7.31	1.76	6.03	0.815	4.45	0.82
10-BV-16	59.54	116.26	13.59	51.88	9.14	2.26	7.19	0.988	5.26	0.96
10-BV-29	49.28	93.19	11.15	42.36	7.57	1.76	6.15	0.835	4.59	0.88
10-BV-30	47.88	84.67	10.95	41.88	7.82	1.79	6.35	0.862	4.72	0.9
10-BV-31	52.32	94.53	11.91	44.88	8.23	1.9	6.52	0.894	4.87	0.95
10-BV-32	47.89	91.69	10.87	40.03	7.38	1.72	5.95	0.815	4.49	0.84
10-BV-34A	48.25	95.07	11.54	44.05	8.73	1.93	7.51	1.053	5.57	1
10-BV-34B	46.16	89.02	11.5	44.04	9.44	1.83	7.85	1.055	5.29	0.89
10-BV-39	44.12	88.25	11.32	44.96	8.49	2.24	6.74	0.89	4.68	0.87
10-BV-40	41.86	82.68	10.62	42.59	8.28	2.25	6.58	0.871	4.55	0.84
10-BV-41	37.75	76.98	9.17	34.93	6.34	1.67	4.99	0.651	3.49	0.64
10-BV-42	27.55	53.91	6.52	25.53	5.3	1.6	5.08	0.763	4.4	0.86
10-BV-44	43.1	80.48	9.3	34.81	6.59	1.7	5.3	0.711	3.86	0.76
10-BV-45	48.11	96.1	11.9	46.94	8.92	2.25	7.65	1.068	5.97	1.16
10-BV-46A	45.32	87.89	10.49	39.74	7.45	1.78	6.17	0.85	4.7	0.89
10-BV-46B	43.48	74.79	8.99	32.21	5.71	1.62	4.46	0.597	3.15	0.58
10-BV-47	44.71	78.46	8.99	31.59	5.47	1.3	4.24	0.565	3.04	0.56
10-BV-48A	42.23	78.12	8.75	31.73	5.69	1.38	4.47	0.613	3.29	0.61
10-BV-48B	43.71	79.23	8.68	30.56	5.32	1.31	4.08	0.566	3.08	0.57
10-BV-49	47.09	88.59	10.61	39.65	7.25	1.7	5.69	0.787	4.27	0.81
H-10-53	55.56	100.7	10.96	38.28	5.91	1.4	3.88	0.513	2.82	0.53
00-LT-2*	27.06	58.31	8.32	39.1	9.42	3.92	10.24	1.467	8.93	1.76
Avg deviation**	0.57	0.77	0.05	0.25	0.03	0.08	0.25	0.01	0.11	0.01

*Sample 00-LT-2 is an internal rock standard

**Average deviation for 6 runs of 00-LT-2

Appendix D continued: ICP-MS trace element data collected in this study

Sample #	Er	Tm	Yb	Lu	Rb	Sr	Nb	Cs	Hf	Ta
Units	ppm	ppm	ppm	ppm	ppm	ppm	Ppm	ppm	ppm	ppm
Method	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS
10-BV-05	2.11	0.295	1.918	0.29	90.96	392	13.65	6.32	4.41	1.1
10-BV-06	2.93	0.439	2.849	0.43	193.96	378	19.51	6.27	8.85	1.3
10-BV-07	1.91	0.271	1.784	0.27	93.96	1204	12.66	3.45	6.21	0.7
10-BV-08	1.98	0.283	1.85	0.28	102.16	1175	12.8	3.16	6.52	0.7
10-BV-09	2.54	0.358	2.363	0.36	111.18	877	14.42	3.39	5.85	0.9
10-BV-10	2.47	0.347	2.268	0.34	90.58	1100	11.47	3.42	5.43	0.7
10-BV-11	1.85	0.269	1.749	0.26	212.76	613	17.79	9.95	3.9	1.4
10-BV-12A	2.39	0.378	2.64	0.43	189.79	559	12.84	6.47	5.31	1.1
10-BV-12B	2.23	0.306	1.979	0.28	133.52	359	15.02	3.77	4.17	1
10-BV-16	2.62	0.357	2.236	0.33	61.85	734	19.27	1.08	6.93	1
10-BV-29	2.47	0.358	2.312	0.35	152.02	660	15.31	5.11	6.81	1
10-BV-30	2.55	0.362	2.327	0.35	156.94	661	14.09	4.55	6.14	1
10-BV-31	2.73	0.393	2.6	0.39	144.59	718	14.98	5.67	6.78	1
10-BV-32	2.4	0.352	2.302	0.35	140.91	663	14.5	5.75	6.63	1
10-BV-34A	2.71	0.37	2.29	0.32	135.31	697	15.34	3.86	6.89	1
10-BV-34B	2.23	0.299	1.792	0.25	160.57	752	14	6.28	6.12	0.9
10-BV-39	2.4	0.341	2.225	0.33	71.64	1027	12.23	0.5	5.71	0.7
10-BV-40	2.33	0.311	2.006	0.29	96.02	963	10.16	1.99	2.6	0.6
10-BV-41	1.8	0.252	1.638	0.24	71.85	842	12.46	0.38	4.63	0.7
10-BV-42	2.4	0.334	2.185	0.32	54.75	514	10.9	4.2	4.02	0.7
10-BV-44	2.14	0.307	2.093	0.32	156.68	760	12.12	6.77	5.58	0.9
10-BV-45	3.3	0.464	3.027	0.45	116.18	722	16.82	2.89	6.85	1
10-BV-46A	2.44	0.342	2.208	0.32	86.05	760	15.28	3.22	4.94	1
10-BV-46B	1.63	0.225	1.533	0.23	171.96	751	12.86	6.12	5.77	1
10-BV-47	1.57	0.228	1.555	0.23	190.2	538	13	6.08	5.39	1.1
10-BV-48A	1.77	0.264	1.795	0.27	178.73	565	12.7	7.39	5.32	1
10-BV-48B	1.65	0.252	1.667	0.26	186.48	552	12.96	7.5	5.38	1.1
10-BV-49	2.34	0.325	2.188	0.33	153.68	661	14.2	4.42	6.15	1
H-10-53	1.52	0.227	1.52	0.23	129.98	339	11.74	2.25	3.32	0.8
00-LT-2*	4.94	0.686	4.393	0.65	42.17	414	7.42	1.73	4.09	0.5
Avg deviation**	0.02	0.01	0.03	0.01	0.33	5.70	0.15	0.03	0.04	0.03

*Samples 00-LT-2 is an internal rock standard

Appendix D continued: ICP-MS trace element data collected in this study

Sample #	Th	U	Y	Zr	Sc	Pb	Ba	Be	Bi	Cd
Units	ppm	ppm	ppm	ppm	ppm	ppm	Ppm	ppm	ppm	ppm
Method	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS
10-BV-05	18.17	5.4	20.81	164	18.1	20.9	1650.9	2.55	0.043	0.13
10-BV-06	25.33	5.48	28.54	358	9.5	21.5	1571.3	2.34	<0.009	0.11
10-BV-07	11.28	3.14	19.21	266	9.1	20.5	1737.9	2.97	0.049	0.2
10-BV-08	11.57	3.21	20.01	277	8.9	17.6	716.8	2.39	0.046	0.07
10-BV-09	14.75	3.96	24.24	236	11	14.1	1525.3	2.59	0.072	0.16
10-BV-10	9.21	2.66	24.25	220	16.6	11.8	1561	2.8	0.044	0.13
10-BV-11	49.98	12.06	18.89	128	9.4	23.1	1033.5	1.61	0.081	0.15
10-BV-12A	22.33	5.91	23.92	208	8.6	24.4	2147.1	2.73	0.057	0.17
10-BV-12B	12.37	3.41	22.46	149	18.2	24.6	1605.2	3.13	0.11	0.13
10-BV-16	11.88	2.1	25.66	295	22	24	1606.4	2.3	0.013	0.23
10-BV-29	20.43	5.16	24.5	272	16.9	18.5	2035.2	2.9	0.028	0.1
10-BV-30	18.3	5.16	25.12	241	15.1	21	1595.5	2.65	0.051	0.21
10-BV-31	20.2	5.63	26.46	270	16.8	19.4	1897	2.38	0.056	0.21
10-BV-32	19.87	5.37	23.74	261	16.6	18.8	1929.7	2.26	0.031	0.2
10-BV-34A	18.1	4.65	26.54	278	13.4	16.9	1593.5	2.58	0.091	0.11
10-BV-34B	16.48	5.82	22.96	238	14	73.7	1567.3	3.52	0.02	0.22
10-BV-39	9.7	2.19	24.44	240	16.9	10.9	614.8	1.37	0.027	0.11
10-BV-40	7.92	1.83	22.72	88	18	9.1	1581	2.07	0.046	0.23
10-BV-41	8.74	1.02	17.78	173	12.9	13.2	1577.6	2.29	0.031	0.15
10-BV-42	6.92	1.39	23.12	169	25	7.5	1889.1	2.81	0.064	0.06
10-BV-44	19.51	5.9	21.35	225	13	19.6	1559.9	2.31	<0.009	0.21
10-BV-45	11.47	2.96	31.73	285	16.6	14.1	1404.7	3.3	0.024	0.16
10-BV-46A	13.7	3.69	24.03	184	18.3	15.3	1824.5	3.05	0.027	0.12
10-BV-46B	21.07	5.94	16.64	232	9.6	20.4	1520.5	3.19	0.087	0.07
10-BV-47	23	6.08	15.77	208	7.5	22.4	1638.2	2.83	0.064	0.1
10-BV-48A	21.18	6.61	16.84	206	10.5	22	1555.6	2.26	0.035	0.16
10-BV-48B	23.63	7.31	16.6	203	8.3	24	1559.4	2.81	0.019	0.12
10-BV-49	19.37	4.39	22.89	243	15.5	20.3	1338	2.22	0.01	0.12
H-10-53	16.17	4.51	15.76	113	4.3	18.4	1388			
00-LT-2*	4.3	1.6	47.01	154	37.6	7	2188.3	1.38	0.053	0.18
Avg deviation**	0.19	0.06	0.45	2.50	0.40	0.07	96	-	-	-

*Samples 00-LT-2 is an internal rock standard

Appendix D continued: ICP-MS trace element data collected in this study

Sample #	Co	Cr	Cu	Ga	Li	Mo	Ni	Sb	Sn	Ti
Units	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm
Method	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS	ICP-MS
10-BV-05	15	89	11	18.67	17.8	2.63	9	0.5	2.35	4677.51
10-BV-06	21.6	33	15	19.84	21.6	1.59	6	0.16	1.31	6306.38
10-BV-07	13.2	113	9	18.65	11.9	3.49	7	0.51	2.33	3758.21
10-BV-08	21.3	172	19	16.54	24	1.07	49	1.8	2.23	4728.11
10-BV-09	15.8	65	11	19.41	22.4	1.99	8	0.42	2.46	5069.52
10-BV-10	7.6	55	6	17.66	31.5	1.61	3	0.46	2.92	2971.89
10-BV-11	29.4	263	37	19.24	8.7	1.74	107	0.08	1.91	8347.65
10-BV-12A	16.2	69	13	19.49	11.9	3.18	7	0.52	2.57	5081.5
10-BV-12B	4.8	48	5	17.77	15.8	2.8	3	0.53	2.93	2610.54
10-BV-16	13.6	59	7	19.87	15.8	2.4	5	0.11	1.54	5985
10-BV-29	5.1	26	5	19.21	32.1	3.12	3	0.76	1.88	3480.6
10-BV-30	16.9	68	11	18.83	11.2	3.15	7	0.48	2.41	4912.56
10-BV-31	9.9	48	5	21.61	16.4	3.39	4	0.4	1.83	4843.31
10-BV-32	9.5	29	4	21.95	13	2.31	3	0.21	1.88	4924.87
10-BV-34A	7.1	45	5	18.08	33.4	2.75	3	0.52	2.75	2681.07
10-BV-34B	8.8	36	9	22.62	29.9	1.76	5	0.71	7.85	5432.34
10-BV-39	39.7	199	50	17.69	7.5	1.17	112	0.09	1.78	8043.49
10-BV-40	12.6	28	8	21.79	9.5	3.51	3	0.15	2.19	7569.1
10-BV-41	15.2	80	9	18.78	9.4	3.24	8	0.5	2.35	5859.39
10-BV-42	6.7	53	6	18.54	33.3	2.49	4	0.37	2.36	3379.66
10-BV-44	15.8	58	7	20.96	38.1	2.03	6	0.17	0.57	6513.93
10-BV-45	10.4	62	14	20.12	23.5	2.06	9	0.7	2.72	4120.85
10-BV-46A	7.7	59	7	20.6	38.9	2.17	4	0.7	4.14	4943.47
10-BV-46B	5.7	29	4	17.96	36.7	2.67	3	0.33	2.76	2488.78
10-BV-47	17	68	12	18.67	19	3.43	7	0.6	2.24	5643.06
10-BV-48A	10.2	47	7	21.78	21.4	3.11	4	0.32	1.84	5169.45
10-BV-48B	13.8	53	10	19.3	23	2.61	8	0.34	2.29	4765.01
10-BV-49	12.1	28	11	17.71	22.6	0.35	4	0.55	6.68	5403.99
H-10-53	3	51					2			
00-LT-2*	30.4	<24	32	19.56	9.8	1.39	12	0.47	1.83	>12000
Avg deviation**	2	4	-	-	-	-	2	-	-	-

*Samples 00-LT-2 is an internal rock standard

Appendix D continued: ICP-MS trace element data collected in this study

Sample #	Tl	V	W	Zn
Units	Ppm	ppm	ppm	ppm
Method	ICP-MS	ICP-MS	ICP-MS	ICP-MS
10-BV-05	0.76	134.41	1.37	83.35
10-BV-06	0.07	218.25	1.11	137.13
10-BV-07	0.84	106.29	1.86	86.67
10-BV-08	0.41	140.64	1.61	53.49
10-BV-09	0.72	149.29	1.22	91.48
10-BV-10	0.95	74.55	1.93	69.08
10-BV-11	0.33	184.61	0.61	118.39
10-BV-12A	0.85	148.85	1.74	89.68
10-BV-12B	1.06	57.19	2.14	61.53
10-BV-16	0.35	165.11	0.9	107.65
10-BV-29	0.7	44.2	1.89	78.6
10-BV-30	0.67	147.51	1.61	88.84
10-BV-31	0.41	97.3	0.74	107.74
10-BV-32	0.49	95.35	0.86	121.8
10-BV-34A	1.07	50.34	1.65	82.37
10-BV-34B	2.24	82.73	2.79	125.96
10-BV-39	1.36	183.88	<0.5	95.3
10-BV-40	0.64	108.48	0.94	122.59
10-BV-41	0.52	170.15	1.4	98.4
10-BV-42	0.91	73.37	1.49	75.58
10-BV-44	0.08	133.41	0.8	134.63
10-BV-45	1.15	89.05	3.41	80.83
10-BV-46A	3.92	107.43	2.82	110.73
10-BV-46B	1.06	48.95	1.66	65.85
10-BV-47	0.76	174.64	1.5	83.78
10-BV-48A	0.52	99.31	0.99	83.91
10-BV-48B	0.71	134.01	1.38	85.17
10-BV-49	0.23	100.07	0.54	89.07
H-10-53		14		23
00-LT-2*	0.21	344.42	0.6	144.62
Avg deviation**	-	10	-	4

*Samples 00-LT-2 is an internal rock standard

Appendix E: Mass Spectrometry isotopic data collected in this study

Sample #	143/144Nd	2 σ error	143/144Nd (i)	87/86Sr	2 σ error	87/86Sr (i)	206/204Pb	2 σ error	207/204Pb	2 σ error	208/204Pb	2 σ error
10-BV-05												
10-BV-06	0.51233859	8.18E-06	0.51231406	0.70752387	8.90E-06	0.706807						
10-BV-07	0.51241803	8.98E-06	0.51239489	0.7063759	1.03E-05	0.706267						
10-BV-08	0.51242100	9.71E-06	0.51239775	0.70638978	8.68E-06	0.706268						
10-BV-09												
10-BV-10	0.51249190	8.44E-06	0.512465691	0.70604332	1.00E-05	0.705928						
10-BV-11	0.51238664	8.06E-06	0.512364217	0.7069039	1.09E-05	0.706418	19.401	8.30E-04	15.683	8.24E-04	39.178	2.95E-03
10-BV-12A	0.51236534	7.36E-05	0.512341512	0.70705662	9.25E-06	0.706582	19.319	7.29E-04	15.672	8.12E-04	39.039	2.85E-03
10-BV-12B												
10-BV-16												
10-BV-29	0.51237420	7.92E-06	0.512350268	0.70787331	9.01E-06	0.707755	19.248	4.65E-04	15.712	4.83E-04	39.359	1.71E-03
10-BV-30	0.51237877	8.66E-06	0.512353764	0.70683211	9.51E-06	0.706510						
10-BV-31	0.51237696	8.67E-06	0.512352402	0.70690839	9.59E-06	0.706577	19.326	1.14E-03	16.690	1.20E-03	39.091	4.20E-03
10-BV-32	0.51236361	8.23E-06	0.512338921	0.70684467	1.00E-05	0.706563						
10-BV-34A	0.51235794	9.32E-06	0.5123314	0.70681873	1.01E-05	0.706524	19.361	1.21E-03	15.709	1.20E-03	39.186	4.14E-03
10-BV-34B	0.51240606	1.01E-05	0.512377355	0.70709845	9.41E-06	0.706828						
10-BV-39	0.51247035	1.08E-05	0.512445062	0.70678282	9.62E-06	0.706484	19.224	5.64E-04	16.678	5.47E-04	38.954	1.94E-03
10-BV-40	0.51252269	9.96E-06	0.512496655	0.7063078	1.53E-05	0.706210						
10-BV-41	0.51249307	9.01E-06	0.512468763	0.70590737	1.26E-05	0.705768						
10-BV-42	0.51229389	9.52E-06	0.512266089	0.70583258	9.20E-06	0.705713	19.397	8.23E-04	15.734	8.26E-04	39.471	2.84E-03
10-BV-44	0.51241856	8.91E-06	0.512393207	0.70649962	7.53E-06	0.706350	19.310	9.90E-04	15.682	1.04E-03	39.063	3.65E-03
10-BV-45	0.51238052	1.05E-05	0.512355072	0.70664651	1.09E-05	0.706358	19.317	7.76E-04	15.694	8.27E-04	39.106	2.70E-03
10-BV-46A				0.70683209	1.24E-05	0.706608	19.363	1.16E-03	15.685	1.23E-03	39.113	4.16E-03
10-BV-46B	0.51240199	8.67E-06	0.51237825	0.70672991	1.03E-05	0.706410						
10-BV-47	0.51238258	9.14E-06	0.512359391	0.70683209	9.22E-06	0.706588						

Appendix E: Mass Spectrometry isotopic data collected in this study con't

Sample #	143/144Nd	2σ error	143/144Nd (i)	87/86Sr	2σ error	87/86Sr (i)	206/204Pb	2σ error	207/204Pb	2σ error	208/204Pb	2σ error
10-BV-48A	0.51239293	1.02E-05	0.512368915	0.70697525	9.65E-06	0.706533						
10-BV-48B	0.51234039	7.90E-06	0.512317077	0.70707142	9.52E-06	0.706601	19.361	2.42E-03	16.708	2.71E-03	39.155	9.73E-03
10-BV-49	0.51238362	9.16E-06	0.512359133	0.70705737	9.30E-06	0.706732						
H-10-53	0.512382		0.512361179	0.707478		0.706951						
Precision*	0.511819			0.710237	1.09E-05		16.888		15.425		36.492	

* Pb stds = NBS 981 measured over 1.5 yrs, Sr std = NBS 9A7 std measured over 2 years and Nd std measured over 4 years