Stratigraphy, Geochemistry, and Origin of the Fish Creek Mountains Tuff,
Battle Mountain Area, North-Central Nevada, U.S.A.

by

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ABSTRACT

The Fish Creek Mountains Tuff (FCMT) is a 25 Ma composite rhyolitic ash-flow tuff mostly restricted to its source caldera in the Fish Creek Mountains, north-central Nevada. The deposition of the FCMT followed a 9 Ma hiatus in local silicic to intermediate volcanism of the Mid-Tertiary Ignimbrite Flare-Up. The FCMT is geochemically distinct from local units, and is mineralogically and chemically zoned. The lower compound cooling unit of the FCMT is crystal-bearing with aphyric pumice, and has cyclical geochemical variations due to magma recharge. The upper compound cooling unit is crystal-rich with crystal-bearing pumice; eruption of the upper cooling unit was rapid, resulting in more linearly variable geochemistry and a generally less evolved character than the lower cooling unit. The parental magma of the FCMT was derived from moderately depleted mantle or asthenosphere, underwent fractional crystallization, incorporated a significant (<30%) crustal component, and erupted in a within-plate setting.

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LIST OF ABBREVIATIONS

$(a/r)_{ss}$	Steady state assimilation/recharge ratio
CHUR	Chondritic uniform reservoir
FCMT	Fish Creek Mountains Tuff
HFSE	High field strength elements
LCU	Lower cooling unit
LILE	Large-ion lithophile elements
MOHO	Mohorovičić discontinuity
MORB	Mid-ocean ridge basalt
NCI	Neodymium Crustal Index
NHRL	Northern Hemisphere Reference Line
RAFC	Recharge, assimilation, and fractional crystallization
REE(s)	Rare earth element(s)
UCU	Upper cooling unit

CHAPTER 1: INTRODUCTION

1.1: Introduction

The Fish Creek Mountains Tuff (FCMT) is an early Miocene composite ash-flow tuff mostly restricted to its source caldera in the Fish Creek Mountains of Lander County, north-central Nevada. The Fish Creek Mountains are a circular range covering more than 500 km², and are located approximately 55 km southwest of the town of Battle Mountain. The location of the Fish Creek Mountains and the study area is indicated in Figure 1.1.

Figure 1.1: Location of the Fish Creek Mountains and other neighbouring ranges, with local highways and towns. The index map is of the state of Nevada and the location of the study area. Figure modified from McKee, 1970.

The FCMT is an areally restricted unit, and is mostly confined to the southern Fish Creek Mountains; rare outcrops of the FCMT occur within 10 km of the western boundary of the Fish Creek Mountains as small exposures in the Tobin Range. The FCMT is very well exposed throughout the southern part of the range. The thickness of the FCMT varies widely, from 30 m at the margins to a maximum of 920 m in thickness. The FCMT is considered to be a medium-sized unit compared to other welded ash-flow sheets in Nevada, with a total estimated eruptive volume of 310 km^3 .

The rhyolitic FCMT is composed of two flat-lying compound cooling units separated in some areas by a cooling break; each compound cooling unit is composed of multiple thin flows. The lower compound cooling unit has a thick nonwelded base containing angular fragments of frothy white and/or grey pumice in a lithic clast-rich and crystal-poor matrix. The degree of welding increases upwards to the top of the unit, which is densely welded with a eutaxitic texture. Abundant fiamme are mostly black and hydrated, locally containing cm-scale spherulites, but include rare two-toned white and grey fiamme. Lithic fragments are abundant in this part of the section and crystals are rare or absent. The base of the upper unit is a thin (approx. 3 m) nonwelded tuff. The midsection of the upper unit is moderately welded with small, variably flattened grey or purple pumice with thin glassy rims. Abundant phenocrysts of quartz (smokey and clear varieties) and plagioclase are present in the matrix and pumice throughout the unit, while lithic fragments are mostly absent.

Three new high-precision Ar-Ar ages of the FCMT ($24.91 \pm 0.05 \text{ Ma}$, $24.95 \pm 0.08 \text{ Ma}$, $24.88 \pm 0.05 \text{ Ma}$) indicate that the tuff was deposited over a short period of time. This event followed a ca. 9 Ma hiatus in local silicic to intermediate volcanism, which included several nearby ca. 34 Ma calderas and a $33.8 \pm 0.14 \text{ Ma}$ dacite and $33.3 \pm 0.3 \text{ Ma}$ andesite.

A neighbouring caldera produced the ca. 34 Ma rhyolitic Caetano Tuff, which extends approximately 90 km along a west-trending belt through the Fish Creek Mountains. The Caetano caldera has undergone significant extensional faulting and tilting during the Miocene and is exposed from the

caldera floor to the overlying cap rocks. In contrast, the FCMT is distinct in that it has remained untilted and largely undeformed, with minimal out-flow tuff. This pristine confined system is ideal for reconstructing the volcanic history of the caldera.

XXXXX Samples of the FCMT are rhyolitic in composition, and high-potassium calc-alkaline in character, similar to local ca. 34 Ma Tertiary rocks. Major and trace element geochemistry varies within and between the compound cooling units. The FCMT is enriched in light rare earth elements relative to heavy rare earth elements, with negative phosphorus, europium, and zircon anomalies.

XXXXX The geochemistry of the FCMT is unique in the region, and is distinct from pre-FCMT Tertiary rocks in the northern Fish Creek Mountains, and also from the neighbouring Caetano Tuff and the nearby Buffalo Valley volcanic centre. Trace element geochemistry suggests that the FCMT was likely largely sourced from the moderately depleted mantle, with magma contribution from the upper continental crust; isotopic data supports this model. According to the trace element geochemistry of pumice in the FCMT, the tuff was erupted in a within-plate tectonic setting.

XXXXX The FCMT and other neighbouring calderas formed during the Mid-Tertiary Ignimbrite Flare-Up, a period of intense caldera-forming igneous activity lasting from ca. 55 Ma to ca. 20 Ma, during which an astounding volume of more than 5,500,000 km³ of material was erupted as ash-flow tuffs and felsic to intermediate lava. The Mid-Tertiary Ignimbrite Flare-Up was the result of rollback of the subducting Farallon plate from beneath the North American plate, resulting in upwelling of asthenospheric mantle along the base of the North American continent. The region later underwent widespread extension and faulting during the formation of the Basin and Range Province, producing the unique topography of the region today, in which long and narrow mountain ranges are separated by wide, flat basins.

XXXXX XXXXX

1.2: Study Overview

1.2.1: Scientific Goals

This study had several scientific goals: 1) to establish volcanological differences between and within the major cooling units of the FCMT; 2) to examine the geochemical variations within and between the major cooling units; 3) to determine the source of the melt which produced the FCMT; and 4) to compare the FCMT to other nearby ash-flow tuffs and to the local pre-FCMT Tertiary rocks.

1.2.2: Geological Problems

The FCMT is an anomalous unit in north-central Nevada, having formed ca. 9 Ma after all other nearby calderas. Although its geochemical signature is similar to some neighbouring units, the FCMT is not directly related to any of the neighbouring ignimbrite bodies. The unit has largely escaped deformation during the faulting related to the formation of the Basin and Range Province; the preservation of the original structure of the FCMT and its near-total confinement to within its own caldera presents a unique opportunity to examine the evolution of the magma which produced the FCMT as the eruption of this tuff progressed.

1.2.3: Significance

This study was intended to provide information on a geological event which was part of a broader series of events across California and Nevada which related to the subduction of the Farallon plate beneath the North American plate during the Mesozoic-early Tertiary, the occurrence of the Mid-Tertiary Ignimbrite Flare-Up, and regional extension during the formation of the Basin and Range Province beginning during the Mesozoic.

1.2.4: Methods

In order to pursue the goals of this study, 23 samples were collected by the author from the study area. Petrography and microscopy were performed on the samples in order to determine mineralogical and textural changes in the FCMT. Fused discs were prepared for all samples in order to perform a chemical analysis by X-ray fluorescence (XRF); solution-mode chemical analysis by inductively coupled plasma mass spectrometry (ICP-MS) was also performed on all samples. Select samples were analysed by thermal ionization mass spectrometry (TIMS) to measure their isotopic composition.

1.2.5: Major Findings

The FCMT is clearly mineralogically zoned. The thick basal zone of the lower cooling unit contains no crystals in either the ash matrix or within the pumices. In the lower cooling unit, the variably welded zone above the basal tuff contains an increasing amount of quartz and feldspar in the ash matrix with increasing height in the volcanic pile, while the pumice and fiamme remain aphyric. Across the major cooling break, pumice and fiamme inclusions contain quartz and feldspar crystals as well as the ash matrix.

The eruption of the FCMT occurred in two phases. The lower cooling unit was erupted in sporadic episodes which allowed for the magma chamber to experience differentiation and evolution. The eruption of the upper cooling unit occurred more quickly, with rapid withdrawal of magma from the chamber. Two models are proposed for the eruption of the FCMT: 1) the FCMT was produced by moderately-depleted mantle-sourced magma which assimilated a significant crustal component, and changes in the geochemistry of the FCMT are due to mafic recharge and variable eruption rates; or 2) the FCMT was produced by a middle-crustal melt of mafic rocks similar in isotopic composition to the Buffalo Valley alkali basalts, which assimilated a significant crustal component. Trace element data suggests that the FCMT was erupted in a within-plate setting. The FCMT is geochemically distinct from

all neighbouring units and did not originate from the same source as any other nearby unit; however, the FCMT is similar to local ca. 34 Ma Tertiary rocks in that it is highly potassic and calc-alkaline in character.

CHAPTER 2: REGIONAL PROCESSES

2.1: Subduction of the Farallon Plate

During the Mesozoic and early Tertiary, the oceanic Farallon plate was being subducted beneath the continental North American plate, creating a convergent plate boundary along the entire western margin of North America (Atwater, 1970; Engebretson et al., 1984; Stock and Molnar, 1988). Subduction of the Farallon plate was partially driven by sea-floor spreading along the East Pacific Rise, a spreading ridge separating the Farallon plate from the oceanic Pacific plate to the west.

The Farallon plate split into two smaller plates at ca. 55 Ma, resulting in the Farallon (south) and Vancouver (north) plates which were separated by the Murray fracture zone (Menard, 1978). By ca. 50 Ma, the Vancouver and Farallon plates were moving independently of one another (Menard, 1978; Rosa and Molnar, 1988). At ca. 28 Ma, the Farallon plate continued to fracture into smaller plates between 27° and 31° N (modern latitude; Atwater and Stock, 1998; Lonsdale, 1991; Severinghaus and Atwater, 1990).

The half-spreading rate of the East Pacific Rise was exceeded by the subduction rate, resulting in the migration of the East Pacific Rise and the Pacific plate towards the North American continent. The East Pacific Rise contacted the North American plate during the Early Miocene, and was largely subducted beneath the North American plate by the Middle Miocene. The East Pacific Rise then divided into several domains; the central portion of the East Pacific Rise, which was subducted beneath the modern western United States, ceased to drive subduction along the western Pacific margin. Instead, subduction was replaced with right-lateral transform faulting (Atwater, 1970; Severinghaus and Atwater, 1970) which resulted in the San Andreas Fault (Wallace, 1990). Subduction continued to the north and south of the subducted section of the East Pacific Rise. The northern remnants of the East Pacific Rise include the modern Juan de Fuca Ridge and Gorda Ridge, which are still active and driving the

subduction of the Juan de Fuca and Gorda plates. The East Pacific Rise continues south of the San Andreas Fault, where it drives the motion of the Cocos plate. The progressive subduction of the Farallon plate and the East Pacific Rise beneath the North American plate is shown in Figure 2.1.

Figure 2.1: Progressive subduction of the Farallon plate and the East Pacific Rise (the spreading center visible in the 30 Ma panel) beneath the North American plate and the formation of the San Andreas Fault. Figure modified from Wallace, 1990, and from Dickinson, 1981.

With the subduction of the East Pacific Rise and the formation of the San Andreas fault, the Vancouver and Farallon plates fragmented into smaller plates. The modern remnants of the Vancouver plate are the Juan de Fuca plate, which is still subducting beneath Washington and Oregon, and the Explorer and Gorda plates. The Rivera and Cocos plates are modern remnants of the Farallon plate, and are still subducting beneath Central America (Sonder and Jones, 1999). The modern remnants of the East Pacific Rise and the remaining lithospheric plates in the northeastern Pacific Basin are shown in Figure 2.2.

2.2: Volcanism of the Mid-Tertiary Ignimbrite Flare-Up

The Farallon plate began subducting beneath the North American plate at a high angle, which gradually shallowed with time. By ca. 80 Ma., the Farallon plate was subducting at a rate of 12 cm/yr; this rapid convergence propelled the Farallon plate at a near-horizontal angle along the base of the North American plate, causing it to act as a thermal buffer between the continental crust and the underlying mantle. The Farallon plate continued to subduct at this rate until ca. 45 Ma, at which point convergence slowed and the dip angle increased (Coney and Reynolds, 1977).

Prior to ca. 55 Ma, volcanism in the western United States consisted of low-frequency, low-volume magmatism involving melts derived nearly entirely from cool continental crust through the addition of heat to the base of the lithosphere by the mantle (Patino-Douce et al., 1990). Magmatism changed in style during the mid-Tertiary, at which point two magmatic fronts began to develop and sweep across the southwestern United States (Coney and Reynolds, 1977) in a prolonged event which produced more than 500,000 cubic kilometers of ash-flow tuffs and more than 5,000,000 cubic kilometers of felsic to intermediate lavas (Johnson, 1991). This event, termed the Mid-Tertiary Ignimbrite Flare-Up, spanned from ca. 55 Ma to ca. 20 Ma and was dominated by caldera-forming explosive magmatism (Lipman et al., 1972; Best et al., 1989; Christiansen and Yeats, 1992).

Figure 2.2: Modern relationship of lithospheric plates, spreading centers, and major faults in the northeastern Pacific Basin, including the Pacific plate, North American plate, and various remnants of the Farallon plate and the East Pacific Rise, including the San Andreas Fault. Figure modified from Wallace, 1990, and from Drummond, 1981.1

The volcanism associated with the Mid-Tertiary Ignimbrite Flare-Up was concentrated in two magmatic fronts which swept from east to west across the western and southwestern United States. The northern front initiated in northern Washington and central Idaho at ca. 54 Ma, and swept southwest at a rate of 15 km/Ma. The southern front formed at ca. 40 Ma in western Texas, and swept

west-northwest at a rate of 30 km/Ma (Christiansen and Yeats, 1992). The two fronts converged in southern Nevada at ca. 20 Ma (Armstrong and Ward, 1991; Christensen and Yeats, 1992). The paths of the two magmatic fronts and their positions at various times are indicated in Figure 2.3.

Figure 2.3: Position and direction of east-to-west migrating magmatic fronts in the western United States during the Mid-Tertiary Igneous Flare-Up. Ages are in million years. The double dotted lines mark the position of the Basin and Range Province (discussed below). Figure is from Humphreys, 1995.

The dramatic change in magmatism at ca. 55 Ma is attributed to the initiation of the removal of the Farallon plate from beneath the North American plate at that time. The Farallon plate had been previously rapidly subducting beneath the North American continent at a sub-horizontal configuration, protecting the base of the continental crust from the heat of the asthenosphere. As the subduction rate began to slow, the Farallon plate began to buckle downward into the asthenosphere along an east-northeast axis at ca. 50 Ma. This buckling tore the slab along its northern and southern boundaries, with the separated edges dipping into the asthenosphere. The positioning of the separated edges allowed asthenospheric material to well up against the North American mantle lithosphere, injecting large quantities of both heat and basaltic melt across the MOHO (Johnson, 1991), and thus initiating the Mid-Tertiary Igneous Flare-Up. The model described above was developed by E.D. Humphreys (1995) to

account for the timing, position, and migratory path of the two magmatic fronts observed across the western United States, and is represented in Figure 2.4.

Figure 2.4: The preferred model of E.D. Humphrey's (1995) to account for the sweep of magmatism across the western United States along two converging magmatic fronts. The view is from below eastern Canada, looking up towards the surface, at ca. 35 Ma. The front of the figure is at the current longitude 105°W, or near Denver, Colorado. Arrows indicate the velocity of the slab relative North America. The buckled slab in the foreground is due to necking or tearing and is descending beneath the western United States. Figure is from Humphreys, 1995.

2.3: Formation of the Basin and Range Province

The Basin and Range Province of western North America is a geological province of unique character. At the surface, the province is a vast arid territory renowned for its evenly spaced, parallel north-south mountain ranges which are separated by flat desert basins (Eaton, 1982). The topography is an expression of the intense extension that the region experienced beginning in the Miocene. The surface area of the Basin and Range Province has more than doubled due to extension of the continental crust (Snow and Wernicke, 2000). It is unclear whether the entire column of lithosphere to the convecting mantle was also affected by the extension, or if the upper crust decoupled from the deeper layers (Jones et al., 1992; Wernicke et al., 2008). The region today is an area of high relief, high heat flow, and thin lithosphere.

2.3.1: Extent

The Basin and Range Province covers more than 1,000,000 km² of western North America, and over 10% of the western United States. It encompasses nearly all of Nevada, approximately half of California, and extends to varying degrees into Oregon, Idaho, Utah, Arizona, New Mexico, and Texas.

Within the United States, the province is bounded to the north by the Columbia Plateau and to the south by the Trans-Mexican Volcanic Belt. From north to south, the eastern border of the province connects with the Wasatch Fault, the Colorado Plateau, and the Rio Grande Rift. To the west of the province, the eastern fault scarp of the Sierra Nevada Mountains extends for over 800 kilometers.

The southern extension of the Basin and Range Province into Mexico is bounded by the Gulf of California and the Baja Peninsula to the west, and by the Laramide thrust front of the Sierra Madre Oriental to the east.

2.3.2: Physiographic Subdivisions of the Basin and Range Province

The Basin and Range Province can be subdivided into distinct subsections. According to physiographic character, Fenneman (1931) separated the province into five regions: the Great Basin, the Salton Trough, the Sonoran Desert, the Mexican Highland, and the Sacramento section. The Basin and Range Province and its physiographic subdivisions are shown in Figure 2.5.

The physiographic region of interest to this study is the Great Basin, within which lies the study area. Fenneman (1931) distinguished the Great Basin according to its hydrological properties; within this region, all drainage systems are confined to the interior, with no outlet to the sea. With an average elevation of over 1500 metres, the Great Basin is the highest of the five subdivisions. Its mountain ranges are roughly parallel in alignment, with most striking north or northeast; Figure 2.5B is a shaded relief map of the western United States, in which the alignment of the mountain ranges and the separation of these ranges by broad flat valleys is observed, particularly in the state of Nevada. High

seismicity, abundant thermal springs, and numerous faults and fault scarps indicate that the Great Basin is the most tectonically active subsection of the Basin and Range Province (Eaton, 1982).

Figure 2.5: The Basin and Range Province, including its physiographic subdivisions. A) Distribution of the Basin and Range Province and its physiographic subdivisions in the western United States, showing relationships with neighbouring geologic domains. B) Shaded relief map of the western United States, showing the boundaries of the geologic provinces of Figure 2.5A. Figure modified from U.S. Geological Survey, 2004.

2.3.3: Formation and Extensional Evolution of the Great Basin

The classic topography of the Great Basin is attributed to movement along broadly distributed normal faults during two periods of deformation. The earlier period, beginning locally at ca. 37 Ma but mostly occurring at 30 Ma to 29 Ma (Gans, 1981), saw movement along normal faults with low dips (first described by Longwell, 1933) and shallow listric faults (first mapped by Hamblin, 1965). The later period of deformation, beginning at ca. 17 Ma, is responsible for the current Basin and Range structures which define the region. Extension at this time occurred due to general block faulting along high-angle faults which deeply dissected the crust, or along the steep upper portions of deep-penetrating listric faults. Volcanism associated with extension evolved in character from that of a continental volcanic arc to that of a back-arc setting (Scholz et al., 1971; Eaton et al., 1978; Stewart, 1978).

Extension of the crust was accommodated in different ways within the upper and lower portions of the lithosphere. Eaton (1980) suggested that extension via brittle failure was confined to a thin surface skin of crust only ca. 15 km thick and up to 950 km wide. Other authors (Thompson, 1959; Hamilton and Myers, 1966) pioneered the idea that pervasive flow accommodated extension at depth, possibly through cataclasis, intragranular gliding, and recrystallization.

Eaton (1980) found that 98.1% of local earthquakes occurred within the upper 15 km of the crust, with 80.3% occurring at depths of less than 10 km. Thus, faulting likely is limited to the upper 15 km of the ca. 65 km-thick lithosphere. Furthermore, on the basis of the limiting nature of fault spacing within the Basin and Range Province, Hamilton and Myers (1966) suggest that most of the faults do not dissect the entire lithosphere, and possibly not even the entirety of the continental crust.

Isostatic observations have indicated that isostatic compensation has not occurred for any but a few of the largest ranges within the Basin and Range Province (Eaton et al., 1978). Even features as large as the western Snake River Plain have only been compensated for within the upper 20 km of the crust (Mabey, 1976).

Seismic refraction measurements indicate that variations in the crustal thickness of the Basin and Range Province are very smooth, with thicknesses of only 20-30 km in areas which have undergone the greatest extension. Regions immediately adjacent to areas of extension have crustal thicknesses of 40-50 km. When regional topography is considered with crustal thickness, a cross section of the plate appears ductilely "necked" (Eaton, 1982).

2.3.4: Extensional Deformation in the Great Basin

Within the region defined as the Great Basin, extensional deformation was likely driven by a combination of basal normal forces and internal body forces internal to the crust. Crustal thickening during the Laramide orogeny, asthenospheric upwelling, removal of mantle lithosphere, possible

removal of an attached subducting slab, and shallowing of the dip of the slab all contributed potential energy to the system, which drove the extension (Sonder and Jones, 1999).

During the crustal thickening of the Laramide orogeny, the mantle lithosphere was thickened and cooled, resulting in convective destabilization of the lithospheric mantle beneath what was to become the Great Basin. As the Laramide orogeny ended, the mantle lithosphere was made buoyant, causing the crustal lithosphere to warm and rise, increasing the mean elevation of the entire region (Humphreys, 1995). This created a regime favourable to explosive volcanism (see Section 2.2 above, relating to the Mid-Tertiary Ignimbrite Flare-Up), as the lithospheric mantle was subjected to higher temperatures (Sonder and Jones, 1999).

The thinning and weakening of the lithosphere and the buoyancy forces related to the increase in elevation combined to drive extension in the uplifted region of the Great Basin. These forces also acted on the Sierran lithosphere to the west, forcing it away from North America, and drove extension south of the elevated region but to a lesser degree (Sonder and Jones, 1999).

CHAPTER 3: THE FISH CREEK MOUNTAINS AND LOCAL GEOLOGY

3.1: The Fish Creek Mountains

The Fish Creek Mountains were mapped by E.H. McKee in 1970 for the United States Geological Survey. The Fish Creek Mountains have remained largely unexplored since McKee's initial efforts, with attention focused instead on other nearby ignimbrites and calderas, such as the neighbouring Caetano Tuff and its associated caldera. McKee's observations of the Fish Creek Mountains are detailed below in the remainder of this section. The observations detailed below were made by McKee (McKee, 1970) unless otherwise noted.

3.1.1: Physiography

The Fish Creek Mountains are an approximately circular range with an areal extent of 520 km² and an average relief of 460 m. The range is saucer-shaped in profile, with a shallow central basin several hundred metres lower in elevation than the surrounding outer rim. The rim is broadest and highest in elevation in the western and southern parts of the range. Broad alluvial valleys surround the range on all sides except to the southwest, where the Fish Creek Mountains and the Augusta Mountains are separated by low hills and a canyon. A shaded relief map of the Fish Creek Mountains is presented in Figure 3.1.

3.1.2: Lithology of the Fish Creek Mountains

The Fish Creek Mountains Tuff

The Fish Creek Mountains are nearly entirely composed of an early Miocene rhyolitic ash-flow tuff (ca. 25 Ma) with an eruptive volume of 310 km³. This unit was named the Fish Creek Mountains Tuff (FCMT) by E.H. McKee in his 1970 report. The FCMT is described in detail in Chapter 4. The distribution of the FCMT within the Fish Creek Mountains is shown in Figure 3.2.

Figure 3.1: Shaded relief map of the Fish Creek Mountains. Figure edited from McKee, 1970.

Basement and Pre-FCMT Rocks

The FCMT is underlain by a variety of rock types of Pennsylvanian to Oligocene age. In most places where the basal contact is exposed, an angular unconformity separates the FCMT from the basement. This contact is exposed in the northern and western parts of the range, where the FCMT is thinnest. In the central, eastern, and southern parts of the Fish Creek Mountains, the basal contact is not exposed (McKee, 1970).

Through most of the southern part of the Fish Creek Mountains, no basal contact exists and the underlying rock is not exposed; McKee assumes that the basement consists of Paleozoic sedimentary rock. Beyond the southern border of the range, isolated outcrops of FCMT occur locally on the Triassic Augusta Mountain Formation in the northernmost Augusta Mountains (McKee, 1970).

Along the northeast margin of the Fish Creek Mountains, the FCMT overlies older Tertiary mafic lava flows and lenticular tuffaceous sedimentary rocks. McKee (1970) associates these lava flows and

sedimentary rocks with similar units in the Shoshone Range, 8 km to the east, which are ca. 10 million years older than the FCMT. In the north-central portion of the Fish Creek Mountains, the FCMT is underlain by (presumably) Mid-Jurassic coarse-grained granite (McKee, 1970).

The Caetano Tuff (Gilluly and Masursky, 1965), a flat-lying welded tuff of Oligocene age, underlies the FCMT in a small area in the northern part of the range; although the two units appear conformable, the age difference between the two units is nearly 10 million years (33.8 Ma for the Caetano tuff; John et al., 2008).

The youngest known local basement rock is the crystal-poor Bates Mountain Tuff which, at ca. 26 Ma, is almost coeval with the FCMT (McKee and Stewart, 1968; John et al., 2008). Within the Fish Creek Mountains, the Bates Mountain Tuff is restricted to the northern regions and may have spread unconformably as a 15 m thick veneer (McKee, 1970) along a topographic feature which later kept it separated from the FCMT deposited to the south. The distribution of the basement rocks of the Fish Creek Mountains is shown in Figure 3.2, with no distinction made in the various types of basement rock.

A suite of pre-FCMT volcanic rocks, ranging in composition from rhyolite to basaltic andesite, occur in the eastern and western Fish Creek Mountains, and also in the westernmost Tobin range and within the Shoshone Range where it extends due south of the Fish Creek Mountains (Christopher Stevens, personal communication, 2012). Recent Ar-Ar ages for these units (dacite: 33.8 ± 0.14 Ma, andesite: 33.3 ± 0.3 Ma; Christopher Henry, personal communication, 2012) indicate that their formation marked the end of a period of local caldera-forming silicic to intermediate volcanism.

The ca. 34 Ma rocks of the Fish Creek Mountains are calc-alkaline and highly potassic in character. As with other calc-alkaline rocks of the Great Basin, the Fish Creek Mountains rocks have negative Nb anomalies and a high ratio of LILE to HFSE. The basaltic andesites were derived from partial melts of the lithospheric mantle which were contaminated during ascension by rhyolitic magma. The felsic rocks of this suite, including rhyolites and dacites, were derived by partial melting of basalts and by

fractional crystallization of andesitic magma. Isotopic data indicate that the ca. 34 Ma melts which produced the silicic to intermediate samples from the Fish Creek Mountains were contaminated by large volumes of ancient continental crust (Christopher Stevens, personal communication, 2012).

Cap Rocks

The FCMT is overlain by younger rocks in the north, north-central, and western parts of the Fish Creek Mountains, as illustrated in Figure 3.2. Within the broad, shallow basin that constitutes the interior of the range, the FCMT is separated from overlying sedimentary strata by an unconformity. These strata consist of tuffaceous sandstone and mudstone which dip gently towards the centre of the basin, where they reach several hundred metres in thickness. McKee (1970) suggests that these sedimentary units may be as much as 10 million years younger than the FCMT, and that they are late Miocene or early Pliocene in age. The sedimentary rocks of the central basin are overlain in turn by a very thin flow of 4.02 Ma olivine basalt (Cousens et al., 2012), which is only 15 m thick and covers nearly 4 km².

The youngest rocks within the Fish Creek Mountains are other olivine basalt units, which occur along the northern border of the range and trend northeast. The olivine basalt occurs as thin flows, cinder cones, and as dikes, which lie on or intrude into all older rock types. These olivine basalts range from 4.02 Ma to 0.96 Ma in age (Cousens et al., 2012).

The only rock within the Fish Creek Mountains whose age is uncertain is a band of dark intrusive to extrusive rocks ranging in composition from andesite to quartz latite. These rocks lie along the western margin of the Fish Creek Mountains, and may be related to the development of the Fish Creek Mountains volcanic centre or to later Basin and Range faulting. These rocks are presumed to be of a similar age to those studied by Christopher Stevens (personal communication), which are ca. 34 Ma in age.

Figure 3.2: Generalized geology of the Fish Creek Mountains, including the Fish Creek Mountains Tuff, the undifferentiated basement rocks, and the cap rocks. The approximate location of the Fish Creek Mountains caldera is indicated on the map. Basalts (Quaternary to Miocene) along the northwest side of the Fish Creek Mountains are the rocks of the Buffalo Valley field. Figure modified from John et al., 2008.

3.1.3: Structures and Tectonism

The homogeneity of the FCMT works to disguise all but the most obvious faults and other tectonic structures within the Fish Creek Mountains. Few faults are recognized in the Fish Creek Mountains, but McKee (1970) draws attention to a series of sub-parallel arcuate linear faults in the southernmost Fish Creek Mountains which are clearly visible in aerial photographs. These faults extend for more than 15 km and are positioned perpendicular to the regional Basin and Range faulting in central Nevada, which is oriented northeast. McKee attributes these arcuate faults to the volcanic evolution of the Fish Creek Mountains caldera rather than the subsequent Basin and Range faulting.

Basin and Range faults have influenced the Fish Creek Mountains along the western margin of the range. Well-defined faults in this region have uplifted the Fish Creek Mountains Tuff and the underlying Paleozoic basement by 300 m or more, exposing the basal contact of the tuff in that area. McKee suggests that this fault zone may have acted as a conduit for the magmas which produced the

linear array of Pleistocene basalt cones and flows along the northwestern margin of the Fish Creek Mountains. Faulting related to the Basin and Range extension is less obvious throughout the rest of the range.

3.2: Local Geology

The Fish Creek Mountains are separated from neighbouring long, narrow, north-northeast-trending mountain ranges by broad, flat valleys, forming the landscape typical to the Great Basin. The neighbouring ranges include the Shoshone, Cortez, Toiyabe, and Tobin Ranges, the positions of which are indicated in Figure 3.3.

3.2.1: Caetano Caldera, Shoshone, Cortez, and Toiyabe Ranges

Directly to the east of the Fish Creek Mountains, the Shoshone Range is largely made up of basalts and pre-Cenozoic basement rock in the northern half of the range, with minor occurrences of the rhyolitic Tuff of Cove Mine. The southern half of the Shoshone Range, the northern expression of the Cortez Range, and the northern Toiyabe Range consist of the Caetano caldera, a structurally dismembered zone mostly consisting of 33.8 Ma Caetano Tuff, and well as the Bates Mountain Tuff and undifferentiated Pliocene to Oligocene sedimentary rocks and Miocene to Eocene volcanic rocks (John et al., 2008). The Caetano caldera was intensely fractured during the Miocene, and tilted 40°-50° to the east as a series of blocks; the caldera is now exposed from the basement rocks through to the sedimentary cap rocks, a stratigraphic thickness of more than 5 km. The caldera has undergone nearly 100% extension from east to west, accommodated by slip along west-dipping normal faults formed at 16 Ma (John et al., 2008).

3.2.2: Tobin Range

To the west of the Fish Creek Mountains, the long and narrow Tobin Range is mostly composed of undifferentiated pre-Cenozoic rocks. The southern part of the range, in line with the Caetano caldera and the Fish Creek Mountains, is composed of outflow units of Caetano Tuff and FCMT, as well as minor Quaternary to Miocene basalts, Pliocene to Oligocene sedimentary rocks, and undivided Miocene to Eocene volcanic rocks (John et al., 2008).

3.2.3: Local Volcano-Tectonic Trough

A local fault-bounded east-west structural trough extends from the Tobin Range, through the Fish Creek Mountains and the Shoshone Range, to the Toiyabe Range (Burke and McKee, 1979). The trough is widest in the Tobin Range, where it is at least 25 km wide, and narrowest in the Toiyabe Range, where it is reduced to 10 km in width; the trough is over 90 km in length. More than 4,000 km³ of material filled the trough, with an additional 2,000 km³ of material spread out of the confines of the trough; the material originated from source vents in the Shoshone Range. Most of the fill is Caetano Tuff, with minor andesite and basalt in the western portion of the trough. The volcanism which produced the material to fill the trough was short-lived; the Caetano Tuff that fills most of the trough is ca. 34 Ma, and overlying andesites in the Tobin Range are ca. 32 Ma. After a period of ca. 10 Ma, during which erosion of the trough took place, the eruption of the FCMT overlapped the trough fill in the Fish Creek Mountains (Burke and McKee, 1979). The distribution of the volcano-tectonic trough is shown in Figure 3.3.

Figure 3.3: The Fish Creek Mountains and neighbouring ranges. The position of the volcano-tectonic trough is approximated by the caldera margins in the Shoshone and Toiyabe Ranges, and by the distribution of tuffs in the Cortez, Fish Creek Mountains, and Tobin Ranges. TYR = Toiyabe Range. Note: sample locations plotted on this figure indicate samples taken by John et al. (2008), not by the author. Figure from John et al., 2008.

CHAPTER 4: THE FISH CREEK MOUNTAINS TUFF

4.1: Previous Work

The Fish Creek Mountains were first described in some detail by E.H. McKee (McKee, 1970), who had performed reconnaissance-level mapping of the region. McKee (1970) identified the rhyolitic ash-flow tuff which comprises much of the southern part of the range, and named it the Fish Creek Mountains Tuff (FCMT). Fission-track and K-Ar dates for samples from the FCMT and other nearby ash-flow tuffs were published simultaneously (McKee and Naeser, 1970). Burke and McKee (1979) identified a volcano-tectonic trough which intersects the Fish Creek Mountains (see Chapter 3.2.3). Minor sampling (John et al., 2010) and dating (Christopher Henry, personal correspondence) have been performed on the FCMT in the recent past.

4.2: New Work

4.2.1: Field Work

Field work was completed during two visits to the Fish Creek Mountains. The first visit took place from July 12, 2010 to July 16, 2010. The author was accompanied by Dr. Brian Cousens and Christopher Stevens of Carleton University, and intermittently by Dr. Christopher D. Henry of the Nevada Bureau of Mines and Geology, University of Nevada, Reno. A second visit took place on May 05, 2011, and the author was again accompanied by the same individuals.

Potential sample locations were scouted in advance using the Fish Creek Mountains topographic map (U.S. Geological Survey, 1982) and a geologic map of north-central Nevada (Stewart and Carlson, 1976). Samples locations were chosen on the merit of placement within the range, by elevation, and by amount of cliffside exposure. This strategy ensured that the Fish Creek Mountains Tuff would be analysed throughout its expanse, both laterally and vertically.

The field area was accessed by vehicle along Nevada State Route 305 and local roads, some of which are unnamed. Samples were collected as intact blocks (rather than as chips) when possible to preserve the chemical integrity of the sample by not exposing the interior of the blocks to contamination. Where possible, samples were collected from multiple elevations in a single exposure in order to directly examine changes in the FCMT with stratigraphic position. Sampling along traverses led to an increased sample density in some areas and a sampling bias between the major cooling units (see Chapter 4.3.4, below); more samples were collected from the lower cooling unit (10 whole-rock samples, 6 pumice/fiamme samples) than the upper cooling unit (6 whole-rock samples, 1 fiamme sample).

Sample locations were recorded using a handheld Magellan Global Positioning System (GPS) unit. The geodetic reference system used was the North American Datum of 1927 (NAD 27). The State of Nevada is located entirely within Zone 11. The positions of the samples collected for this study are indicated in Figure 4.1 and documented in Table 4.1.

Figure 4.1: Locations of samples collected from the FCMT for this study (red markers). The base map is a 1:100 000 metric topographic map of the Fish Creek Mountains (U.S. Geological Survey, 1982).

Table 4.1: Sample Locations and Elevations for Pumices and Whole-Rock FCMT

Sample	Latitude	Longitude	NAD	Zone	Easting	Northing	Elevation (m)
10-BV-04	40° 18.758N	117° 21.670W	NAD 27	Zone 11	469311 E	4462310 N	1472
10-BV-17	40° 11.781 N	117° 24.749 W	NAD 27	Zone 11	464887 E	4449440 N	2542
10-BV-18	40° 11.656 N	117° 24.740 W	NAD 27	Zone 11	464904 E	4449194 N	2538
10-BV-19	40° 10.664N	117° 23.021W	NAD 27	Zone 11	467333 E	4447344 N	2395
10-BV-20	40° 10.895N	117° 22.851W	NAD 27	Zone 11	467577 E	4447769N	2257
10-BV-21	40° 11.191 N	117° 14.762 W	NAD 27	Zone 11	479481 E	4448275 N	1573
10-BV-22	40° 11.240 N	117° 14.443 W	NAD 27	Zone 11	479509E	4448368N	1612
10-BV-23	40° 11.323N	117° 14.434W	NAD 27	Zone 11	479520E	4448519N	1643
10-BV-24	40° 11.385 N	117° 14.337 W	NAD 27	Zone 11	479658E	4448635N	1700
10-BV-25	40° 11.441 N	117° 14.338 W	NAD 27	Zone 11	479654 E	4448737 N	1694
10-BV-26	40° 11.488 N	117° 14.334 W	NAD 27	Zone 11	479665 E	4448826 N	1768
10-BV-27	40° 11.582N	117° 14.289W	NAD 27	Zone 11	479731E	4449017N	1792
10-BV-28	40° 11.604N	117° 14.275W	NAD 27	Zone 11	479752 E	4449040N	1812
10-BV-33	40° 17.323 N	117° 21.572 W	NAD 27	Zone 11	469439 E	4459655 N	1548
10-BV-35	40° 17.306 N	117° 21.640 W	NAD 27	Zone 11	469312 E	4459592 N	1670
10-BV-36	40° 17.267 N	117° 21.737 W	NAD 27	Zone 11	469219 E	4459544 N	1646
10-BV-37	40° 17.256 N	117° 21.678 W	NAD 27	Zone 11	469318 E	4459523 N	1585
10-BV-38	40° 15.876 N	117° 31.746 W	NAD 27	Zone 11	455007 E	4457071 N	1603
11-BV-01*	40° 17.323 N	117° 21.572 W	NAD 27	Zone 11	469439 E	4459655 N	1548
11-BV-03*	40° 17.306 N	117° 21.640 W	NAD 27	Zone 11	469312 E	4459592 N	1670

*Sample locations and elevations are approximated using samples collected nearby within the same exposure as no GPS device was available during collection of these samples (11-BV-01 approximated using sample 10-BV-33; 11-BV-03 approximated using sample 10-BV-35).

4.2.2: Field Descriptions

Exposure of the FCMT within the Fish Creek Mountains is excellent, with outcrops ranging in style from small rubble outcrops with in-situ blocks to large multi-tiered cliffs. The various outcrop types are pictured in Figure 4.2. The style of outcrop varies with position in the volcanic pile; the unwelded base of the lower cooling unit forms exposures which are visually distinct from exposures higher in the sequence. Figures 4.2A and 4.2B are of the thick unwelded base of the lower cooling unit, which is structureless and lacks differentiation and zoning. Erosion of this zone has resulted in a smooth, rounded outcrop.

In areas where the FCMT was deposited in thinner individual flows, outcrops occur as multi-tiered cliff exposures, as in Figure 4.2C. The cliff faces in the tiers are composed of individual flows of variably welded FCMT, separated by cooling breaks. Where only one thin flow is exposed, smaller rubble outcrops occur, as in Figure 4.2D. Outcrops of the more densely welded FCMT in these two exposure types are typically intensely jointed, forming tabular blocks as in Figures 4.2D.

Several outcrops of FCMT are structureless, as in the unwelded tuff in Figures 4.2A and 4.2B. In more densely welded zones which contain fiamme or flattened pumices, a eutaxitic texture is observed. Figure 4.2E is of a jointed cliff exposure in which flattened pumices define the eutaxitic texture.

Due to the climate of the study area, the FCMT has undergone only minor surficial weathering. Weathering rinds are typically very thin and do not extend in to the core of the tabular blocks of FCMT, as in Figure 4.2F. In outcrops which are intensely jointed, blocks are considerably smaller and thus fresh samples are more difficult to collect.

A)

B)

C)

D)

E)

F)

Figure 4.2: Exposure styles of the FCMT. A) Rounded exposure of thick, unwelded, undifferentiated, structureless basal tuff. B) Erosion pattern of the unwelded basal tuff. C) Multi-tiered cliff exposure of several welded individual flow units separated by cooling breaks. D) Rubble outcrop with tabular blocks. E) Jointed exposure with a eutaxitic texture defined by flattened pumices; orientation of pumices is parallel to the pen in the centre of the photograph. F) Thin weathering rind on a tabular block of FCMT.

4.2.3: Petrography

Samples of FCMT consist of an ash matrix, typically with phenocrysts of quartz and feldspar. The ash matrix varies in colour, occurring as pale grey, various shades of pinkish-grey, or purple-grey. Quartz phenocrysts occur in two varieties: clear, and smokey (discussed below). Smokey quartz occurs in a range of shades of grey, from light grey to nearly black in colour. Feldspar phenocrysts are often weathered and chalky in appearance; plagioclase feldspar is more abundant in the samples, with sanidine also present in lower quantities. Pyroxene phenocrysts are rare, occurring in only a few samples. Detailed descriptions of every sample collected for this study are included in Appendix I.

Also present in most samples of FCMT are pumices, which occur as frothy, uncompacted blocks, or as variably flattened fiamme. Frothy pumice occurs as interspersed angular fragments which are white and/or grey in colour, with very elongated vesicles. Fiamme are variably flattened. Generally, fiamme develop thicker glassy rims and darker cores with progressive flattening; completely flattened fiamme are black, rehydrated, and devitrified, occasionally with cm-scale visible spherulites, and define a eutaxitic texture in the FCMT. Progressive flattening of the pumices is related to the degree of welding of the FCMT at that location, with increased welding resulting in flatter and more elongated fiamme. Notably, samples of frothy pumice and rehydrated black fiamme tend to be aphyric, while the variably flattened pumices with glassy rims contain phenocrysts of similar composition and abundance to the matrix.

Lithic fragments are present in some samples of the FCMT, and occur as variably rounded (angular to sub-rounded) clasts of varying composition. These xenoliths range in size from millimeters in size to several centimeters in diameter. The xenoliths derive from a variety of sources, and range in composition.

Several samples of FCMT are presented in Figure 4.3, highlighting the diversity in the FCMT. Figure 4.3A is of frothy pumice in the unwelded basal zone of the lower cooling unit. Figure 4.3B is of a

Figure 4.3: Samples of FCMT highlighting the variable character of the tuff. A) Unwelded tuff with randomly oriented crystal-free frothy pumices (location for 10-BV-04-series samples); B) moderately welded tuff with flattened pumice, which has a thin glassy rim and a darker purple core (sample 10-BV-37); C) highly welded tuff with eutaxitic texture defined by aphyric black fiamme (large black fiamme at hammer grip is sample 10-BV-33); D) sample of FCMT with light grey matrix (sample 10-BV-27); E) sample of FCMT with pink-grey matrix (sample 10-BV-17, from the type locality at Mt. Moses); F) sample of FCMT with purple-grey matrix (sample 10-BV-19).

moderately flattened pumice with a glassy rim and a purple core. Figure 4.3C is of flattened black fiamme defining a eutaxitic texture, occurring in a more densely welded zone of the FCMT. Figure 4.3D through 4.3F showcase the variety in colour for the ash matrix of the FCMT, with Figure 4.3E representing a typical sample of FCMT, taken at the type locality for the unit.

Smokey Quartz

Quartz crystals in the matrix of the Fish Creek Mountains Tuff occurs as both clear and smokey (light to dark grey) varieties in varying proportion throughout the pile.

Smokey quartz is generated through the irradiation of clear quartz which contains aluminum (Al^{3+}) cations substituting for silicon (Si^{4+}) at several hundred parts per million in the crystal structure. Also present in the crystal are monovalent cations, such as H^+ , Na^+ , or Li^+ , which act to maintain electronegativity (Nassau and Prescott, 1975). Radiation causes the ejection of one electron from an electron pair on an oxygen atom in the crystal, leaving a 'hole colour center'. The hole is then at an excited energy state, as the remaining electron is no longer paired. The energy state affects the light absorption, producing the smokey colour. Thus, it is the radiation and formation of the hole colour center which produces the grey colour of the smokey quartz; the occurrence of aluminum as an impurity is insufficient to produce the colour change (Nassau and Prescott, 1975). The structure of normal quartz and smokey quartz is presented in Figure 4.4.

Radiation occurs via x-ray, gamma ray, or neutron bombardment, and long-term exposure (on the order of several million years) to radioactive materials emitting low levels of radiation may have a significant effect on the transformation of quartz to smokey quartz. Progressive exposure to radiation may eventually lead to the expulsion of an electron from every oxygen atom adjacent to an aluminum atom, darkening the colour of the crystal (Nassau and Prescott, 1975).

Figure 4.4: Structure for normal quartz and irradiated smokey quartz. A) Normal quartz, without substitutions. B) Smokey quartz, with Al^{3+} substituting for Si^{4+} . The H^+ charge maintains charge neutrality after one electron from an electron pair on an O^{2-} anion is ejected due to radiation, leaving a hole colour center in the smokey quartz. Figure modified from Nassau, 1978.

It is important to note that the electron ejected via radiation is not expelled from the crystal; it is simply relocated elsewhere in the crystal structure. Heating of the crystal allows the electrons to reorganize themselves to the original configuration, effectively ‘bleaching’ the crystal and returning it to its original clear colour. The temperature at which this occurs is approximately 60°C (Nassau and Prescott, 1977), indicating that hole colour centers cannot develop above this temperature. This indicates that the development of smokey quartz occurs after formation and cooling of the rock (Nassau and Prescott, 1975).

Quartz crystals may be shielded from radiation by any thin film of material. The pervasive occurrence of smokey quartz in the FCMT requires the presence of radioactive material throughout the matrix of the tuff. The abundance of uranium in the FCMT varies between 4.9 ppm and 25.9 ppm, but given the duration over which the quartz crystals were exposed to radiation, the abundance of smokey quartz is not unexpected.

The occurrence of smokey quartz and clear quartz in the same sample is unusual, and must indicate that the clear quartz xenocrysts lacked the substitutional aluminum required for hole colour center formation. Both types of crystals must have been present during the transformation of the smokey quartz, as the unit must have been deposited and cooled significantly prior to any colour change occurring. The uniform distribution of the crystals rules out the possibility of zoned exposure to

radiation, indicating that both crystal types were exposed to the same amount of radiation, albeit to different effect. The difference must therefore lie in the composition of the quartz, and may indicate different sources for the quartz crystals.

4.2.4: Microscopy

Thin-sections were prepared of whole-rock samples of FCMT, and are detailed in Appendix II. Common minerals observed in all or most samples include quartz, plagioclase, and sanidine. Other mineral phases include zircon, pyroxene, sericite, and Fe-oxides, which occur as minor or accessory phases. Phenocrysts typically appear fractured, with quartz crystals commonly missing fragments of the crystal or as rounded and embayed crystals. Euhedral crystals are rare; most crystals are anhedral to subhedral. Most samples appear to lack flow textures, although flow banding and orientation of fiamme are observed in some samples. In some fiamme, mm-scale spherulites were observed which were not visible the hand samples.

Several differences exist between samples of the lower cooling unit and samples of the upper cooling unit. In samples of the lower cooling unit which contain pumice or fiamme, crystals are present in the ash matrix but the pumice and fiamme are aphyric. Samples of from the upper cooling unit have crystals of quartz and/or feldspar in the pumices/fiamme as well as in the ash matrix. Samples from the upper cooling unit also consistently contain accessory zircon, whereas zircon is less common in the lower cooling unit samples. Pyroxene is also present in samples from the upper cooling unit, but not in samples of the lower cooling unit. Small lithic fragments are also observed in samples from the lower cooling unit, and are very rare in the upper cooling unit samples.

Several microphotographs of samples from the FCMT are presented in Figure 4.5. Figure 4.5A is included to present an example of a typical sample of FCMT, and is of a sample collected from the type locality for the FCMT (10-BV-17). Almost all samples of whole-rock FCMT collected for this study are very

similar mineralogically and texturally to sample 10-BV-17, with minor variations in mineral abundances. The suite of minerals occurring in the FCMT are common from one sample to another, with accessory minerals such as pyroxene and zircon occurring in only a few samples. Figure 4.5B is of spherulites in fiamme in 10-BV-37, Figure 4.5C is of zircon crystals and clusters of sericite in the matrix of sample 10-BV-22.

Figure 4.5: Photomicrographs of thin sections of whole-rock samples of FCMT, taken under cross-polarized light. Pl = plagioclase feldspar; Qtz = quartz; Sa = sanidine; Ser = sericite; Sph = spherulite; Zrn = zircon. A) Representative sample of FCMT. Photomicrograph of sample 10-BV-17 (from the type locality). B) Spherulites in fiamme. Photomicrograph of sample 10-BV-37. C) Zircon phenocrysts and sericite clusters in the ash matrix. Split photomicrograph of sample 10-BV-22, with both panels taken at the same scale.

4.3: The Fish Creek Mountains Tuff

The Fish Creek Mountains Tuff (FCMT) is a composite ash-flow sheet which shows little lithological variation other than the local occurrence of lithic debris (McKee, 1970). It is considered to be a medium-sized unit, relative to other welded ash-flow sheets in Nevada. The type locality of the FCMT is located at the U.S. Government radio facility on Mt. Moses, which is located in the southern Fish Creek Mountains at approximately 4449200N, 464900E.

4.3.1: Eruption of the Fish Creek Mountains Tuff

The distribution of the Fish Creek Mountains Tuff within a circular area and the confinement of the tuff almost entirely to within the Fish Creek Mountains suggest that the eruption of the tuff occurred concurrently with caldera collapse. The walls of the caldera are not visible in the Fish Creek Mountains, but it can be assumed that blocks of the wall occur in the FCMT where they are in contact. The collapse of the caldera during eruption formed a basin in the south-central part of the range in which the ash-flow tuff was trapped (McKee, 1970); at the caldera margins, the ash-flow tuff changes from flat-lying to gently dipping inwards towards the basin centre.

The material which formed the tuff was ejected during a low-energy eruption, as indicated by the short distance the ash-flow material was transported. The tuff spread radially for approximately 15 km in all directions from an assumed vent in the central Fish Creek Mountains. The eruption of the tuff was completed in less than 1 million years, as supported by the dates detailed below. The tuff was likely mostly erupted in one major event, and cooled at the same time (McKee, 1970).

4.3.2: Age of the Fish Creek Mountains Tuff

McKee (1970) reported an age of 24.3 ± 0.7 Ma for the FCMT on the basis of three K-Ar ages of sanidine (McKee and Silberman, 1970) and two fission-track ages of zircons (McKee and Naeser, 1970).

Three excellent new Ar-Ar dates of the Fish Creek Mountains Tuff have been acquired by Dr. Christopher D. Henry of the Nevada Bureau of Mines and Geology, University of Nevada, Reno. These dates of 24.91 ± 0.05 Ma, 24.95 ± 0.08 Ma, and 24.88 ± 0.05 Ma (Christopher D. Henry, personal communication, 2012) show remarkable homogeneity in age, with very little error; the dates also confirm that the eruption of the FCMT occurred very rapidly, requiring less than one million years. Furthermore, the dates by McKee and Henry confirm that the deposition of the FCMT followed a ca. 9 m.y. hiatus in local silicic to intermediate volcanism which produced several nearby ca. 34 Ma calderas and the local pre-FCMT rocks discussed in Chapter 3.1.2.

4.3.3: General Properties

The Fish Creek Mountains Tuff is fairly homogenous in character. The ash matrix varies from light grey to pink or purple in colour, and the tuff is crystal rich (generally more than 25% crystals by volume, but may be less than 25% in some localities). Mafic phenocrysts are minor, and are usually completely altered. Major phenocrysts are feldspars (plagioclase feldspar and sanidine) and quartz (clear and smokey varieties) of varying proportion, occurring as crystals and crystal fragments up to 3 mm in diameter. Zircon and apatite constitute less than 1% of the tuff (McKee, 1970).

The degree of welding is variable throughout the stack of the FCMT. The extent to which the rock has been welded affects the volume percentage of crystals in the tuff, where the least welded tuff contains 25% crystals and 75% ash matrix and voids. The most densely welded tuff is 50% crystals and 50% devitrified glass matrix. A typical sample of FCMT contains 35% crystals and 65% matrix (McKee, 1970).

Pumice is present throughout the FCMT volcanic pile. Where the tuff is non-welded or lightly welded, the pumices are angular and frothy, may be white and/or grey in colour, and reach 10 cm in size. Moderately welded tuff contains pumices that have been slightly flattened, and are reddish-purple

in colour with dark 1-3 mm rims. These pumices have been observed to be as much as 12 cm in length and 3.5 cm in thickness. Densely welded tuff contains fiamme which are typically only a few centimeters in length and millimeters thick, but reach more than 25 cm in length in some localities. Fiamme may occur as black aphyric bodies, rarely with cm-scale spherulites.

The FCMT is typically structureless, and lacks evidence of vapour phase alteration (McKee, 1970). Eutaxitic structure is rare, but is very well developed at some localities where the fiamme are extremely flattened and elongated.

Quartz occurs as crystals which range in colour from clear to dark grey (smokey quartz). The presence of the clear and smokey varieties, and the ratio of one to the other, does not appear to relate to position within the FCMT.

4.3.4: Cooling Units

The Fish Creek Mountains Tuff is a composite ash-flow sheet comprised of two major compound cooling units, each of which includes several flow units that commonly exhibit a thin cooling break. The two major cooling units are also separated by a thicker cooling break, which is observed in Dacies Canyon (McKee, 1970). The lower and upper compound cooling units are distinct from one another in that the lower compound cooling unit contains abundant lithic fragments, whereas the upper compound cooling unit does not. Each compound cooling unit possesses a basal zone of non-welded tuff, with the degree of welding gradationally increasing upward through the unit to a densely welded top (McKee, 1970). For the remainder of this report, the upper and lower compound cooling units will be referred to simply as the lower cooling unit and the upper cooling unit.

There are only a few locations in the Fish Creek Mountains in which a significant portion of the FCMT is exposed, particularly when both the lower and upper cooling units are present. One such exposure is pictured in Figure 4.6, in which the top of the lower cooling unit and the base of the upper

cooling unit are exposed in a single hillside. In Figure 4.6, several positions are marked: 1) a cliff exposure of the lower cooling unit, where it is moderately welded and contains abundant xenoliths and aphyric pumice; 2) a transition zone where the lower cooling unit becomes densely welded; xenoliths are smaller and less abundant, and fiamme are black and devitrified; 3) the base of the upper cooling unit, where 3 m of non-welded tuff is exposed; 4) a cliff exposure of lithic-free, lightly- to moderately-welded tuff which is pumice-free; and 5) the top of the exposure, with the highest abundance of phenocrysts and particularly of smokey quartz in the entire sequence.

Figure 4.6: Lower and upper cooling units, separated by a cooling break. 1: Cliff exposure of the lower cooling unit, where it is moderately welded and contains abundant xenoliths and aphyric pumice. 2: Transition zone where the lower cooling unit becomes densely welded; xenoliths are smaller and less abundant, and fiamme are black and devitrified. 3: The base of the upper cooling unit, where 3 m of non-welded tuff is exposed. 4: Cliff exposure of lithic-free, lightly- to moderately-welded tuff which is pumice-free. 5: The top of the exposure, with the highest abundance of phenocrysts and particularly of smokey quartz in the entire sequence.

Lower Cooling Unit

The lower cooling unit of the Fish Creek Mountains Tuff is characterized by the presence of lithic fragments within the matrix (McKee, 1970). These xenoliths are volcanic rock of intermediate composition, and vary from angular fragments up to 10 cm in diameter to well-rounded boulders up to

25 cm in diameter. McKee (1970) speculates that the lithics may represent a contamination on eruption of the flow which produced the lower cooling unit, and that the abundance reflects local concentrations due to pre-eruption topography.

The pre-eruption topography influenced the basal portion of the lower cooling unit. The basal zone is non-welded to weakly welded, and variably thick; the thickness of this portion of the unit reaches more than 150 m where the ash-flow tuff filled small shallow paleobasins, such as at the mouth of Dacies Canyon, and is dependent on relief at the time of deposition (McKee, 1970).

The non-welded basal portion and the overlying moderately welded tuff constitute two-thirds of the lower cooling unit. The upper one-third of the unit is highly welded. The size and abundance of xenoliths generally decreases moving upsection in the lower cooling unit from the lithic-rich non-welded base to the more welded top, where lithics are rare.

Also present in the basal zone of the lower cooling unit are angular clasts of frothy white and/or grey pumice ranging in size from 1-10 cm. The matrix and the pumice fragments are crystal-poor, or crystals are absent. The pumice fragments become increasingly flattened as the degree of welding increases; in the densely welded top of the lower cooling unit, the pumice have developed into fiamme which define a eutaxitic texture. These fiamme are mostly black and aphyric, and locally contain cm-scale spherulites. Rare two-toned white and grey fiamme also occur.

Upper Cooling Unit

The upper cooling unit of the Fish Creek Mountains Tuff is characterized by its lack of xenoliths. It contains abundant phenocrysts of quartz (smokey and clear varieties) and sanidine in the matrix and pumice throughout the unit.

The basal portion of the upper cooling unit is appreciably thinner than that of the lower cooling unit; due to the lack of significant changes in topography, the upper cooling unit has a uniform basal

zone of approximately 3 m of non-welded tuff. The basal zone does not show any evidence of vapour phase alteration, indicating that the lower cooling unit was completely cooled before the deposition of the upper cooling unit occurred.

The midsection of the unit is moderately welded, with small, variably flattened grey or purple pumice with thin (1-3 mm) glassy rims. Unlike the pumice and fiamme in the unwelded base and densely welded zones, the moderately welded pumices with glassy rims often contain phenocrysts of quartz and feldspar in similar proportions to the matrix in which the pumices occur.

The upper cooling unit is structureless throughout most of the central part of the Fish Creek Mountains, where it is thick (915 m) and massive. Towards the margins of the range, where the tuff is only 30 m thick, well-formed columnar joints developed normal to sub-horizontal layering in the tuff (McKee, 1970).

Cooling Breaks

McKee (1970) recognized that the upper and lower cooling units are separated locally by a complete cooling break which evolves into a partial cooling break before disappearing. This cooling break can be traced for nearly 1.5 km in the west exposure of Dacies Canyon, but is absent in the thicker tuff of the southern Fish Creek Mountains. Thin cooling breaks separating individual flows in the compound cooling units can be observed in some locations where exposure is excellent.

CHAPTER 5: GEOCHEMISTRY OF THE FISH CREEK MOUNTAINS TUFF

Whole-rock major and trace element geochemistry of the FCMT was determined using x-ray fluorescence (XRF), performed at University of Ottawa, and by inductively coupled plasma mass spectrometry (ICP-MS), performed by Geoscience Laboratories in Sudbury, Ontario. Isotope analysis was performed using the thermal ionization mass spectrometer (TIMS) of the Isotope Geochemistry and Geochronology Research Centre at Carleton University. Complete results are listed in Appendix III.

5.1: Major Element Geochemistry

5.1.1: Lithological Classification by Major Element Geochemistry

Samples of the Fish Creek Mountains Tuff plot almost exclusively in the rhyolite field of the total alkali versus silica (TAS) diagram in Figure 5.1, with one pumice sample (10-BV-04C, a frothy white pumice in the basal zone of the lower cooling unit) plotting above the other fields. The white and grey pumices are less abundant in SiO_2 and more abundant in the alkalis than the aphyric fiamme. Overall, the whole-rock samples of FCMT are more abundant in SiO_2 and less abundant in the alkalis than the pumice and fiamme; the whole-rock samples also display less variety in composition than the pumice and fiamme samples. Within the whole-rock samples of FCMT, samples of the lower cooling unit display a wider range in SiO_2 and alkali abundances than samples from the upper cooling unit.

The whole-rock and pumice samples from the FCMT are high in silica and potassium. The whole-rock samples generally classify as high-potassium calc-alkaline while the pumice and fiamme samples typically classify as shoshonitic in character, according to the SiO_2 vs. K_2O plot in Figure 5.2. The high potassium content of the FCMT is a trait shared with other ca. 35 Ma felsic rocks of the western Great Basin (Christopher Stevens, personal communication; Christopher Henry, personal communication, 2012).

Figure 5.1: Felsic portion of a total alkali-silica (TAS) diagram, classifying samples of whole-rock FCMT and pumice and fiamme of the FCMT. Sample 10-BV-04C (white pumice) plots above the trachyte field. Also plotted are whole-rock samples pre-FCMT Tertiary rocks in the northern and western Fish Creek Mountains (Christopher Stevens, personal communication, 2012), samples of whole-rock Caetano Tuff and pumices of the Caetano Tuff (plotted as fields; Christopher Henry, personal communication, 2012), and whole-rock samples of ignimbrites from the Tobin Range (Gonsior, 2006). Figure modified from Le Maitre et al., 1989.

Whole-rock and pumice samples from the FCMT are potassic; only one sample of white frothy pumice is sodic in composition. In Figure 5.3, whole-rock samples from the upper and lower cooling units have a similar range in potassic rock index values, where samples from the upper cooling unit are more potassic than samples from the lower cooling unit. Samples of pumice and fiamme from the FCMT demonstrate a greater range of values than the whole-rock samples of FCMT. Overall, the pumice and fiamme samples from the FCMT nearly encompass the range in values of other nearby units in the western Great Basin.

Figure 5.2: SiO_2 vs. K_2O plot classifying samples of whole-rock FCMT and pumice and fiamme of the FCMT. Sample 10-BV-04C (white pumice) plots in the shoshonite series. Also plotted are whole-rock samples pre-FCMT Tertiary rocks in the northern and western Fish Creek Mountains (Christopher Stevens, personal communication, 2012), samples of whole-rock Caetano Tuff and pumices in the Caetano Tuff (plotted as fields; Christopher Henry, personal communication, 2012), and whole-rock samples of ignimbrites from the Tobin Range (Gonsior, 2006). Figure modified from Peccerillo and Taylor, 1976.

Figure 5.3: Potassic rock index values of whole-rock, pumice, and fiamme samples from the FCMT, pre-FCMT Tertiary rocks (Christopher Stevens, personal communication, 2012), whole-rock and pumice samples from the Caetano Tuff (Christopher Henry, personal communication, 2012), and ignimbrite units from the Tobin Range (Gonsior, 2006). The minimum, average, and maximum potassic rock index value for each sample group is plotted. Sample 10-BV-04C (white pumice) is the only sodic sample from the FCMT.

5.1.2: Variations and Trends in Major Element Geochemistry

Variations between Whole-Rock FCMT and Pumice or Fiamme

Samples of whole-rock FCMT contain variable amounts of pumice or fiamme as well as xenoliths of country rock. These contaminants influence the geochemistry of the samples. The pumice and fiamme, however, are contaminant-free and therefore represent the true composition of the magma. Provided in Table 5.1 is a summary of the major oxides present in frothy pumice (4 samples) and rehydrated aphyric fiamme (3 samples) in the lower and upper cooling units. The frothy pumice occurs in the basal portion of the lower cooling unit and the aphyric fiamme occur in the upper welded portion of the lower cooling unit (2 samples) and within the upper welded portion of the upper cooling unit (1 sample). Oxides with the greatest variability between sample types include Na_2O , CaO , TiO_2 , MgO , and P_2O_5 .

Table 5.1: Average Concentrations of Major Oxides in Samples of Whole-Rock Fish Creek Mountains Tuff and Pumice and Fiamme of the FCMT

Oxide	Fish Creek Mountains Tuff			Pumice and Fiamme of the FCMT		
	Average LCU FCMT	Average UCU FCMT	Average FCMT	Average Frothy Pumice	Average Aphyric Fiamme	Average Pumice
SiO_2 (%)	74.87	75.67	75.14	70.19	75.88	72.63
Al_2O_3 (%)	13.52	13.41	13.49	13.75	12.99	13.43
K_2O (%)	5.21	5.31	5.24	5.65	6.10	5.84
Na_2O (%)	3.30	3.21	3.27	5.87	3.04	4.66
$\text{Fe}_2\text{O}_{3(\text{t})}$ (%)	1.56	1.05	1.39	1.41	1.14	1.29
CaO (%)	1.06	0.93	1.02	2.16	0.60	1.50
TiO_2 (%)	0.23	0.23	0.23	0.08	0.14	0.10
MgO (%)	0.18	0.14	0.17	0.85	0.07	0.52
P_2O_5 (%)	0.06	0.05	0.05	0.01	0.03	0.02
MnO (%)	0.02	0.00	0.01	0.02	0.01	0.02
Total Oxides:	100.00	100.00	100.00	100.00	100.00	100.00
No. of Samples:	10	6	16	4	3	7

Note: Major element abundances have been recalculated to exclude volatiles. LCU = lower cooling unit, UCU = upper cooling unit.

Table 5.1 also provides a summary of the average abundances of major oxides in the lower cooling unit, upper cooling unit, and the overall average abundance in the FCMT. The average values in Table 5.1 indicate that compared to the pumice and fiamme, the whole-rock FCMT is enriched in SiO_2 ,

Al_2O_3 , $\text{Fe}_2\text{O}_{3(\text{t})}$, TiO_2 , and P_2O_5 . The whole-rock FCMT is depleted in K_2O , Na_2O , CaO , MgO , and MnO compared to the pumice and fiamme. The relative enrichment and depletion of the major element oxides in the averaged whole-rock samples of FCMT compared to the averaged pumice and fiamme are presented in Figure 5.4. Enrichment-depletion charts compare the abundances of select elements or oxides in two data sets; data sets used in this study include sample type (i.e. whole-rock vs. pumice) or stratigraphic position (i.e. upper cooling unit vs. lower cooling unit). The ratios presented in these charts indicate whether the abundances of the oxides are higher in the numerator index (become enriched; ratio is greater than 1) or lower (become depleted; ratio is less than 1).

Figure 5.4: Enrichment-depletion diagram of major element oxides in averaged whole-rock samples of FCMT compared to averaged samples of pumice and fiamme. Whole-rock FCMT is enriched in SiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_{3(\text{t})}$, TiO_2 , and P_2O_5 and depleted in K_2O , Na_2O , CaO , MgO , and MnO relative to the pumice and fiamme of the FCMT, which represent true magma.

Variations Within and Between the Upper and Lower Cooling Units

Comparing the major elemental abundances of the FCMT, it is clear that chemical variations do exist between the base of the lower cooling unit and the top of the uppermost exposed unit (not necessarily the last material expelled from the magma chamber). Figure 5.5 plots the change in select major element abundances with elevation for whole-rock samples of FCMT. In Figure 5.5, samples from the lower cooling unit vary greatly in major element abundances over a minor elevation change; samples from the upper cooling unit are less variable over a greater elevation change.

lower cooling unit also display a wider range of values when major oxides are plotted against SiO_2 concentrations. Despite the differing variability of the major element oxides in each cooling unit, each cooling unit has similar averaged values, as shown in Table 5.1.

Figure 5.5: Variations in select major element oxide abundances of whole-rock samples of FCMT. Major element oxide concentrations are normalized to volatile-free composition. Solid points were collected from the lower cooling unit of the FCMT, hollow points were collected from the upper cooling unit of the FCMT. A) Variations in the concentrations (weight %) of $\text{Fe}_2\text{O}_{3(t)}$, TiO_2 , K_2O , and P_2O_5 with elevation (m); B) Variations in the concentrations (weight %) of $\text{Fe}_2\text{O}_{3(t)}$, TiO_2 , K_2O , and P_2O_5 with SiO_2 content (weight %).

In order to clearly highlight the differences in major element oxide abundances within and between the cooling units, enrichment-depletion charts are presented in Figure 5.6. Figure 5.6 includes four enrichment-depletion charts to illustrate the changes in major oxide abundances across various intervals: Figure 5.6A compares the oxide abundances at the top of the upper cooling unit to the abundances at the base of the upper cooling unit; Figure 5.6B compares the oxide abundances at the top of the lower cooling unit to the base of the lower cooling unit; Figure 5.6C compares the average oxide abundance of the upper cooling unit to the average oxide abundance of the lower cooling unit; Figure 5.6D compares the oxide abundances of the base of the upper cooling unit to the top of the lower cooling unit, thereby presenting the change in major element oxide abundances in material erupted before and after the cooling break. In each of these charts, material from higher in the volcanic pile is compared to material lower in the pile, indicating the change in major element oxide abundances with time as the eruption progressed. It is important to note that in Figure 5.6, the oxides occur in very different abundances and thus a minor change in an oxide which occurs in small concentrations creates a large scale bar in Figure 5.6 while a larger change in the concentration of an abundant oxide may produce a small scale bar. For example, the average abundance of P_2O_5 changes from 0.06% to 0.05% between the lower cooling unit and the upper cooling unit. The abundance of P_2O_5 drops by 0.01% between the two cooling units, and is thereby depleted by 16.67% overall. The average abundance of SiO_2 changes from 74.87% to 75.67% between the lower cooling unit and the upper cooling unit. The abundance of SiO_2 increases by 0.80%, and is thereby enriched by 1.07% overall. Although the overall change in abundance in SiO_2 is much greater than the overall change in abundance of P_2O_5 , the scale bar produced by change in the abundance of P_2O_5 is much larger due to the lower overall concentration of P_2O_5 . The effect of this is to exaggerate the appearance of enrichment or depletion in less-abundant oxides while diminishing the appearance of enrichment or depletion of more-abundant oxides.

Figure 5.6: Enrichment-depletion diagrams of major element oxide concentrations in the Fish Creek Mountains Tuff. All charts compare samples higher in the stratigraphy to samples lower in the stratigraphy, illustrating the change in major oxide concentrations with increasing height in the volcanic pile. A) The upper cooling unit; B) the lower cooling unit; C) the average composition of the upper cooling unit compared to the average composition of the lower cooling unit; and D) the base of the upper cooling unit relative to the top of the lower cooling unit.

The charts in Figure 5.6 indicate that several changes occurred in the major element geochemistry of the FCMT as the eruption proceeded. With increasing height in the lower cooling unit, the FCMT is depleted in CaO, Fe₂O_{3(t)}, MgO, MnO, P₂O₅, and TiO₂ compared to the base of the cooling unit. A compositional change occurred in the magma chamber during the cooling break between the eruption of the lower and upper cooling units; the base of the upper cooling unit is enriched in all major oxides except K₂O, Na₂O, and SiO₂, compared to the top of the lower cooling unit. The average composition of the upper cooling unit differs from the average composition of the lower cooling unit, with the upper cooling unit enriched in Al₂O₃, CaO, SiO₂, and TiO₂ relative to the lower cooling unit. The

composition of the upper cooling unit is also variable with height, with the top of the cooling unit enriched in Na_2O , P_2O_5 , SiO_2 , and TiO_2 relative to the base of the unit.

With regards to the major element oxide trends of the pumices analyzed in this study, results are very basic; only seven pumice samples were collected, representing three stratigraphic levels of the FCMT. Samples were collected of the frothy grey or white pumices in the basal zone of the lower cooling unit, of the aphyric black fiamme in the densely welded top of the lower cooling unit, and of aphyric black fiamme in a densely welded zone of the upper cooling unit. These samples indicate that the abundance of SiO_2 , P_2O_5 , TiO_2 , and K_2O increase with height in the FCMT, while the abundances of all other oxides decrease. Pumices represent original magma, providing a true snapshot of composition of magma in the chamber.

5.2: Trace Element Geochemistry

Trace elements are elements which comprise less than 0.01% of the rock. The trace elements of interest in this study occur in concentrations which are expressed as parts per million (ppm; $1 \text{ ppm} = 1 \times 10^6$). The trace elements analysed for this study are identified in Appendix III; trace elements of significance to this study are discussed below.

Figure 5.7 plots the change in select trace element (Th, Sr, La, and Zr) abundances with elevation for whole-rock samples of FCMT. In Figure 5.7, samples from the lower cooling unit vary greatly in major element abundances over a minor elevation change; samples from the upper cooling unit generally are less variable over a greater elevation change. Samples from the lower cooling unit also display a wider range of values when these trace elements are plotted against SiO_2 concentrations.

Figure 5.7: Variations in select trace element abundances of whole-rock samples of FCMT. Solid points were collected from the lower cooling unit of the FCMT, hollow points were collected from the upper cooling unit of the FCMT. A) Variations in the concentrations (ppm) of Th, Sr, La, and Zr with elevation (m); B) Variations in the concentrations (ppm) of Th, Sr, La, and Zr with SiO₂ content (weight %, normalized to volatile-free composition).

5.2.1: The Significance of Rare Earth Elements

The rare earth elements (REEs) are a series of trace elements which have similar physical and chemical properties. The REEs, also referred to as lanthanides, form the series of metals with atomic numbers 57 through 71 in the periodic table of the elements, and include La, Ce, Pr, Nd, Pm, Sm, Eu, Gd,

Tb, Dy, Ho, Er, Tm, Yb, and Lu. The REEs generally occur as cations with a 3+ charge; an exception to this is Eu, which occurs as Eu^{3+} or Eu^{2+} . The REEs are generally incompatible during felsic magma evolution; the exception again is Eu.

The REEs are amongst the most useful trace elements in geochemistry; their very low solubility renders them practically immobile during hydrothermal alteration, weathering, and low-grade metamorphism. Thus, the REE patterns of the collected samples accurately represent the original composition of the source material and the crystal-melt equilibria which have occurred throughout its evolution, even if the samples have undergone slight weathering (Rollinson, 1993; Michard, 1989). It should be noted that high-grade metamorphism increases the mobility of REEs (Humphries, 1984), but metamorphism is not a factor in considering the Fish Creek Mountains Tuff or other nearby calderas, tuffs, and units discussed in this study.

5.2.2: Variations and Trends in Rare Earth Element Geochemistry

The REE composition of an igneous rock is influenced by the original REE chemistry of the source material, and also by the crystal-melt equilibria of the magma during its evolution. Due to the immobility of the REEs, a plot of the REE abundances normalized to a reservoir such as the primitive (primordial) mantle produces a pattern with meaningful features. The slope of the line and any anomalous peaks or troughs (referred to as anomalies) in the line are a direct result of the evolution of the magma, and as such, provide insight into the magmatic history of the rock.

Rare Earth Element Geochemistry of Pumice and Fiamme in the Fish Creek Mountains Tuff

Rare earth element patterns of the pumice and fiamme in the FCMT are plotted in Figure 5.8, where they are normalized to the primitive mantle (Sun and McDonough, 1989). The REE patterns of these pumice samples indicate that there is a very minor depletion of heavy REEs relative to light REEs.

All pumice samples display a strong negative Eu anomaly. Samples of frothy white and grey pumice from the same stratigraphic height within the basal portion of the lower cooling unit are plotted in Figure 5.8A. The pumice fragments show similar trends, except in the REEs heavier than Gd, where the behaviour of 10-BV-04D deviates from the other frothy pumices. Samples 10-BV-04A, 10-BV-04B, and 10-BV-04C show slight but increasing depletion of heavy REEs beyond Gd, while 10-BV-04D shows no progressive depletion of these elements.

The rehydrated, glassy, black, aphyric fiamme of the welded portions of the lower (10-BV-33 and 11-BV-01) and upper (11-BV-03) cooling units are plotted in Figure 5.8B, where a clear difference is demonstrated in the REE patterns of the two cooling units. The samples from the lower cooling unit are nearly perfectly identical, but contrast sharply with the sample from the upper cooling unit. The fiamme of the lower cooling unit show slight depletion of the REEs with increasing atomic weight, but have a strong negative Eu anomaly. The fiamme of the upper cooling unit is enriched in light REEs relative to heavy REEs, and shows a much less pronounced Eu anomaly.

Overall, the pumice and fiamme of the lower cooling unit do not demonstrate significant changes in the concentrations of the REEs with position in the volcanic pile, except for the concentration of Eu, which decreases with height in the lower cooling unit. The upper cooling unit demonstrates an enrichment in the light REEs relative to heavy REEs, and a less pronounced Eu anomaly. However, only one sample of fiamme was recovered from the upper cooling unit, and no conclusions may be drawn regarding REE trends in the upper cooling unit.

A)

B)

Figure 5.8: Rare earth element patterns of pumice and fiamme in the FCMT, normalized to the primitive mantle. A) REE patterns of frothy pumice in the lower cooling unit of the FCMT (same stratigraphic position); B) REE patterns of aphyric fiamme in the lower cooling unit (10-BV-33, 11-BV-01) and the upper cooling unit (11-BV-03). Primitive mantle chemistry from Sun and McDonough, 1989.

Rare Earth Element Geochemistry of the Fish Creek Mountains Tuff

Rare earth element abundances for whole-rock samples of FCMT were normalized relative to the primitive mantle and plotted in Figure 5.9. Samples of the upper cooling unit (Figure 5.9A) and the lower cooling unit (Figure 5.9B) of the FCMT show an enrichment of the light REEs (La to Eu) relative to the heavy REEs (Gd to Lu), and also a negative Eu anomaly (Eu anomalies are discussed below). The samples of the upper cooling unit demonstrate two trends: 1) the heavy REEs become increasingly depleted with height in the stratigraphic column; and 2) the light REEs become increasingly enriched with height in the stratigraphic column. It should be noted that height in the stratigraphic column represents depth in the original magma chamber. Thus, the results plotted in Figure 5.9A indicate that the magma chamber was zoned with respect to REE abundances, with the lower portion of the magma chamber progressively depleted in heavy REEs and more enriched in light REEs relative to the upper portion of the magma chamber.

As the lower cooling unit does not demonstrate clear trends in the abundances of REEs relative to position in the volcanic pile (Figure 5.9B), it can be inferred that the upper portion of the magma chamber was not zoned to the same extent as the lower parts of the magma chamber. Samples of the lower cooling unit of the FCMT follow the same general patterns as samples of the upper cooling unit; the only exception is sample 10-BV-37, which has the largest Eu anomaly of all samples, and is also significantly more enriched in the light REEs than the other samples. This may be due to the presence of lithic fragments in this sample, which have contaminated the sample and influenced its whole-rock geochemistry, or to a greater pumice component.

A)

B)

Figure 5.9: Rare earth element patterns for whole-rock samples of the FCMT normalized to the primitive mantle. The order of the samples in each legend reflects the relative position in the stratigraphic column for each grouping of samples. A) REE patterns of the upper cooling unit of the FCMT. B) REE patterns of the lower cooling unit of the FCMT. Primitive mantle chemistry from Sun and McDonough, 1989.

Enrichment or Depletion of Rare Earth Elements Between Whole-Rock FCMT and Pumice or Fiamme

The whole-rock samples of FCMT contain xenoliths of country rock which contribute to the geochemical properties of the sample, thereby altering the chemical composition from the original magma. Pumice and fiamme samples represent true magma. Comparing the average whole-rock FCMT composition to the average composition of the pumice and fiamme of the FCMT indicates how the addition of foreign material has influenced the geochemistry of the samples. Figure 5.10 is an enrichment-depletion chart comparing the REE abundances in average whole-rock FCMT to the average pumice or fiamme sample. According to Figure 5.10, the whole-rock FCMT is enriched in most light REE and depleted in the heavy REEs, relative to the original magma, and is less Eu-depleted.

Figure 5.10: Enrichment-depletion diagram of REEs in averaged whole-rock samples of FCMT compared to averaged samples of pumice and fiamme. Whole-rock FCMT is enriched in most light REEs and depleted in the heavy REEs relative to the pumice and fiamme in the FCMT, which represent true magma.

Enrichment or Depletion of Rare Earth Elements Within and Between the Upper and Lower Cooling Units

The enrichment and depletion of REE abundances with respect to position in the volcanic pile is charted in Figure 5.11. Within the lower cooling unit, the top of the cooling unit is more enriched in all REEs except for Ce and Eu. Across the cooling break, a distinct change in REE abundances occurs, with the base of the upper cooling unit depleted in light REEs and enriched in heavy REEs. Within the upper cooling unit, the top of the unit is enriched in light REEs and depleted in heavy REEs relative to the base

of the unit. Overall, the average composition of the upper cooling unit is depleted in all REEs except Ce and Eu compared to the average composition of the lower cooling unit.

Figure 5.11: Enrichment-depletion diagrams of rare earth element concentrations in the Fish Creek Mountains Tuff. All charts compare samples higher in the stratigraphy to samples lower in the stratigraphy, illustrating the change in REE concentrations with increasing height in the volcanic pile. A) The upper cooling unit; B) the lower cooling unit; C) the average composition of the upper cooling unit compared to the average composition of the lower cooling unit; and D) the base of the upper cooling unit relative to the top of the lower cooling unit.

Variations in Major and Trace Element Geochemistry Along a Traverse Line

Several whole-rock samples from the lower and upper cooling units of the FCMT were collected from a single traverse along an exposure in order to directly examine the changes in geochemistry with stratigraphic position. The variations in the concentrations of several trace and major elements through the cooling units and with position in the volcanic pile are presented in Figure 5.12. Samples from the lower cooling unit exhibit highly variable concentration of La, Ba, Y, SiO₂, K₂O, and P₂O₅, while

concentrations of Nb, Th, and U change with more regularity. In almost all instances, the concentrations of the elements in the upper cooling unit differ greatly from the concentrations in the lower cooling unit, suggesting that processes within the magma chamber changed during the cooling break between the deposition of the lower cooling unit and the formation of the upper cooling unit.

Figure 5.12: Variation of select element abundances with increasing height in the stratigraphy. Samples were collected over a single exposure. Solid points: lower cooling unit of the FCMT; hollow points: upper cooling unit of the FCMT. Position in the stratigraphic column is not to scale. A) La (ppm); B) Nb (ppm); C) Th (ppm); D) Ba (ppm); E) U (ppm); F) Y (ppm); G) SiO₂ (weight %, volatile-free); H) K₂O (weight %, volatile-free); I) P₂O₅ (weight %, volatile-free).

5.2.3: Incompatible Element Anomalies: Europium (Eu), Zirconium (Zr), and Phosphorous (P)

A deviation in the slope of the REE patterns creates peaks or troughs in the pattern, where peaks are considered to be positive anomalies or relative enrichments, and troughs are negative anomalies or relative depletions. The REE patterns in Figure 5.8 and Figure 5.9 all exhibit negative Eu

anomalies, albeit of different magnitudes. Europium anomalies are the result of the compatibility of Eu^{2+} in feldspars; in felsic melts, Eu^{2+} prefers to exist in the crystal phases of plagioclase and potassium feldspar, while the other REEs prefer to exist in the melt (Rollinson, 1993).

A negative Eu anomaly then suggests that the Eu was bound up in feldspars and removed or excluded from the system in the following ways: 1) the material was the product of partial melting in which the feldspars of the parent material were not completely melted, thus retaining Eu in the source; or 2) the melt underwent fractional crystallization, where feldspars crystallized and were removed from the system, taking Eu with them (Rollinson, 1993).

The Eu anomaly can be quantified. The following equation was suggested by Taylor and McLennan (1985), who recommended using the geometric mean to express the anomaly:

$$\frac{\text{Eu}}{\text{Eu}^*} = \frac{\text{Eu}_N}{\sqrt{(\text{Sm}_N) \cdot (\text{Gd}_N)}}$$

where

$$\frac{\text{Eu}}{\text{Eu}^*} = \text{the Eu anomaly}$$

Eu^* = expected value of Eu if feldspar did not retain or remove Eu

Eu_N = measured concentration of Eu (normalized)

Sm_N = measured concentration of Sm (normalized)

Gd_N = measured concentration of Gd (normalized)

The value of Eu^* is found by drawing a line between the normalized measured values of Sm and Gd on the REE pattern, with Eu^* taken as the position on the slope where it crosses the Eu marker; Sm and Gd are the neighbouring REEs to Eu in a REE pattern organized according to atomic number, and are thus used to calculate the Eu anomaly. Figure 5.13 illustrates the position of Eu^* relative to the normalized measured values of Eu, Sm, and Gd.

Figure 5.13: Determining the value of Eu* for use in the calculation of the Eu anomaly.

The Eu anomalies of all of the samples of whole-rock FCMT and the pumice and fiamme were calculated using the method described above. Zirconium (Zr) and phosphorous (P) also display anomalous abundances relative to other elements with similar atomic numbers (Figure 5.14) due to partitioning between solid and liquid phases. The value of the Zr anomaly (Zr/Zr^*) is calculated in the same way as the Eu anomaly, with respect to Sm and Gd concentrations. Zr is normalized relative to Sm and Gd because they are the neighbouring REEs to Zr on a REE plot which is organized according to atomic number; although Eu has an atomic number closer to Zr than Gd, Eu is not used because it too has anomalous concentrations relative to the other REEs. The P anomaly (P/P^*) is calculated using Pr and Nd concentrations. The minimum, maximum, and average values of the Eu, Zr, and P anomalies for each sample type (whole-rock FCMT, pumice or fiamme) and sample location (upper cooling unit, lower cooling unit) are compared in Table 5.2.

Table 5.2: Minimum, Maximum, and Average Eu, Zr, and P Anomalies of Pumice and Whole-Rock FCMT

Anomaly		Whole-Rock FCMT			Pumice or Fiamme in FCMT		
		UCU	LCU	All	Pumice	Fiamme	All
Eu/Eu*	Minimum	0.104	0.031	0.031	0.013	0.002	0.002
	Maximum	0.175	0.172	0.175	0.031	0.127	0.127
	Average	0.142	0.119	0.128	0.020	0.044	0.030
Zr/Zr*	Minimum	1.108	0.663	0.663	0.443	0.426	0.426
	Maximum	1.533	2.187	2.187	0.753	1.125	1.125
	Average	1.298	1.249	1.268	0.540	0.668	0.595
P/P*	Minimum	0.018	0.018	0.018	0.010	0.004	0.004
	Maximum	0.052	0.060	0.060	0.015	0.059	0.059
	Average	0.034	0.032	0.033	0.013	0.023	0.017

UCU = Upper Cooling Unit, LCU = Lower Cooling Unit

Extended trace element patterns of samples of pumice and fiamme in the FCMT and of whole-rock samples of FCMT are presented in Figure 5.14. These extended trace element patterns include several elements (Eu, Zr, and P) for which a trough, or negative anomaly, appears on the pattern.

Variations in the Eu, Zr, and P anomalies with stratigraphic height of all whole-rock samples of FCMT are plotted in Figure 5.15. For Eu/Eu*, the values of the anomaly repeatedly fluctuate from high to low in the lower cooling unit, and increase steadily in the upper cooling unit. The Zr/Zr* values follow a similar pattern, but the fluctuations are less regular and less pronounced in the lower cooling unit. The values of P/P* also follow this pattern in the lower cooling unit, but are more variable within the upper cooling unit than Eu/Eu* and Zr/Zr*. These patterns support a change in the processes affecting the magma chamber during the cooling break between the deposition of the lower and upper cooling units of the FCMT.

Figure 5.14: Extended trace element patterns for samples of pumice and fiamme from the FCMT, normalized to the primitive mantle (Sun and McDonough, 1989), including elements used for anomaly calculations (Eu, Zr, P). A) Trace element patterns of frothy pumice in the lower cooling unit of the FCMT (same stratigraphic position); B) Trace element patterns of aphyric fiamme in the lower cooling unit (10-BV-33, 11-BV-01) and the upper cooling unit (11-BV-03).

C)

D)

Figure 5.14 (continued): Extended trace element patterns for whole-rock samples of the FCMT normalized to the primitive mantle (Sun and McDonough, 1989), including elements used for anomaly calculations (Eu, Zr, P). The order of the samples in each legend reflects the relative position in the stratigraphic column for each grouping of samples. C) Trace element patterns of the upper cooling unit of the FCMT. D) Trace element patterns of the lower cooling unit of the FCMT.

A)

B)

C)

Figure 5.15: Select rare earth element anomalies of the FCMT, plotted according to stratigraphic height (not to scale). A) Europium anomaly (Eu/Eu^* , with Eu^* calculated with respect to Sm and Gd); B) Zirconium anomaly (Zr/Zr^* , with Zr^* calculated with respect to Sm and Gd); C) Phosphorus anomaly (P/P^* , with P^* calculated with respect to Pr and Nd). Closed symbols = lower cooling unit; open symbols = upper cooling unit.

5.2.4: Trace Element Discrimination Diagrams and Tectonic Environments

Samples of whole-rock FCMT and of pumice and fiamme within the FCMT plot within a variety of fields in the tectonic discrimination diagrams of Figure 5.16 and Figure 5.17. The bivariate plots of Figure 5.16, with incompatible trace elements plotted against dry SiO_2 weight percent values or other incompatible elements, are based on granitic rocks but are applicable for felsic extrusive rocks. Figure 5.16 highlights the differences between the compositions of the whole-rock samples and the pumice samples; pumice and fiamme of the LCU have higher Nb and Y concentrations and lower SiO_2 than all other plotted samples, and plot consistently in the within-plate field. The single pumice sample of the UCU plots at or near the boundary between the within-plate and volcanic-arc fields, and is chemically distinct from the LCU pumices due to lower Nb and Y concentrations. Whole-rock samples of FCMT from the LCU and UCU generally plot in the volcanic-arc field, with minor overlap into the within-plate field. Samples of whole-rock FCMT from the LCU have slightly higher SiO_2 , Nb, and Y concentrations than samples from the UCU, but occupy similar regions of the plots in Figure 5.16. In Figure 5.16C, ocean-ridge granites have a distinct field, in which none of the samples plot.

Figure 5.17 consists of ternary plots which were developed to indicate the tectonic setting of basalts. Discrimination diagrams for basalts may be used to infer the tectonic origin of felsic extrusive rocks if the diagrams are derived using truly incompatible elements, and if the abundance of these elements in the melt is not affected by fractional crystallization and are therefore representative of the parent source. The diagrams of Figure 5.17 do not provide a definitive tectonic setting for the FCMT. In Figure 5.17A, all samples except one pumice outlier plot within the volcanic arc or island arc field. Samples plot in two distinct clusters in Figure 5.17B, with whole-rock samples from both cooling units and the pumice sample from the UCU plotting within the field for volcanic-arcs which produce calc-alkalic rocks (Field 1A); pumice and fiamme from the LCU plot within the field for intercontinental rifts which produce alkalic rocks (Field 3A). One sample of whole-rock FCMT from the LCU plots outside of

the two main clusters, near the convergent point for three tectonic settings (intercontinental rifts producing alkali rocks, enriched E-type MORB, and continental basalts). In Figure 5.17B, as in the diagrams of Figure 5.16, there is a distinction between the fields in which pumice and fiamme tend to plot and the fields in which the whole-rock samples of FCMT plot.

Figure 5.16: Bivariate tectonic discrimination diagrams for granitic rocks and felsic extrusive rocks, with incompatible elements plotted against SiO₂ (dry weight percent values) or other incompatible trace elements. (syn-)COLG = (syn-)collisional granite; ORG = ocean-ridge granite; VAG = volcanic-arc granite; WPG = within-plate granite. Whole-rock samples of FCMT plot within the volcanic-arc or syn-collisional setting fields, pumice and fiamme samples from the FCMT plot within the within-plate field. A) Nb (ppm) vs. SiO₂ (dry, weight %); B) Y (ppm) vs. SiO₂ (dry, weight %); C) Nb (ppm) vs. Y (ppm). Figure modified from Pearce et al., 1984.

Figure 5.17: Ternary tectonic discrimination diagrams for basaltic rocks. Whole-rock samples of FCMT and samples of pumice and fiamme in the FCMT do not plot consistently within the same tectonic setting. A) Th-(Hf÷3)-Ta diagram with fields A (N-type MORB), B (E-type MORB and within-plate tholeiites), C (alkaline within-plate basalts), and D (volcanic-arc basalts and island-arc tholeiites). Figure modified from Wood, 1980. B) (La÷10)-(Y÷15)-(Nb÷8) diagram with Field 1 (volcanic-arc basalts; A: calc-alkali basalt, C: volcanic-arc tholeiite, B: overlap between A and C), Field 2 (continental basalts; A: continental basalts, B: back-arc basalts), and Field 3 (oceanic basalts; A: alkali basalts from intercontinental rifts, B: enriched E-type MORB, C: weakly enriched E-type MORB, D: N-type MORB). Figure modified from Cabanis and Lecomte, 1989.

5.3: Isotope Geochemistry

In order to classify the isotopic character of the FCMT, radiogenic isotopes were analysed in samples of whole-rock FCMT and samples of pumice and fiamme. The isotopes of interest include $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$, $^{208}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{206}\text{Pb}/^{204}\text{Pb}$. The isotopes of Sr, Nd, and Pb are useful in determining the source of the magma which produced the FCMT, and also in describing the extent of crustal assimilation into the magma.

Samples were run for Pb, Sr, and Nd isotopic analysis 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.

5.3.1: Variations in Isotope Geochemistry

Figure 5.18 plots the initial (at time of formation, ca. 24.9 Ma) isotopic ratios with elevation for select samples of whole-rock FCMT and pumice and fiamme in the FCMT. Samples with anomalously high initial $^{87}\text{Sr}/^{86}\text{Sr}$ values ($^{87}\text{Sr}/^{86}\text{Sr} > 0.765$) and/or extremely high $^{87}\text{Rb}/^{86}\text{Sr}$ ratios ($^{87}\text{Rb}/^{86}\text{Sr} > 10$) were excluded. These anomalous values are due, in part, to low Sr content in the samples; low Sr concentrations are considered unreliable due to the mobility of Sr (Cousens et al., 1993).

The initial values for $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ show the greatest variation within and between sample types, with the greatest variation occurring in the Nd isotopes. Variation in the Sr isotopic ratios is less obvious due to the exclusion of certain samples, as discussed above. The initial values for $^{206}\text{Pb}/^{204}\text{Pb}$ show less variation, which may be a result of the smaller sample population.

there is a clear distinction between pumice/fiamme samples, where the $^{143}\text{Nd}/^{144}\text{Nd}$ values decrease and the $^{87}\text{Sr}/^{86}\text{Sr}$ values increase from the aphyric fiamme, through the grey pumice, to the white pumice samples.

Figure 5.19: Initial Nd-Sr isotope plot for samples of whole-rock FCMT and of grey pumice, white pumice, and aphyric fiamme in the FCMT. Select samples are excluded from the plot due to anomalously high initial $^{87}\text{Sr}/^{86}\text{Sr}$ values (10-BV-33, 10-BV-36, 11-BV-01; initial $^{87}\text{Sr}/^{86}\text{Sr}$ values range from 0.765059 to 0.945268) and/or suspicious $^{87}\text{Rb}/^{86}\text{Sr}$ ratios (10-BV-04D, 10-BV-23, 10-BV-27, 10-BV-33, 10-BV-36, 10-BV-37, 11-BV-01; $^{87}\text{Rb}/^{86}\text{Sr} > 10$ are considered suspicious).

5.3.2: Epsilon Values of Nd and Sr

One method to express isotopic ratios for Nd and Sr is through epsilon notation. The epsilon value calculates the deviation between a sample and a standard isotopic ratio. The standard commonly used for epsilon values is CHUR (chondritic uniform reservoir; Rollinson, 1993). The epsilon value of Nd (ϵ_{Nd}) is calculated using the ratio $^{143}\text{Nd}/^{144}\text{Nd}$, and the epsilon value of Sr (ϵ_{Sr}) is calculated using $^{87}\text{Sr}/^{86}\text{Sr}$. Isotopic ratios normally cannot be directly compared unless the samples are the same age. The epsilon

value accounts for the age of a sample, thereby allowing for meaningful comparison of samples of varying age.

The ϵ_{Nd} value of a sample at the time of its formation is calculated using the following equation:

$$\epsilon_{Nd}^t = \left[\frac{(^{143}Nd/^{144}Nd)_{Sample,t}}{(^{143}Nd/^{144}Nd)_{CHUR,t}} - 1 \right] \times 10^4$$

where t is the time of formation for the sample. The composition of CHUR at time t is calculated using the following equation:

$$\left[\frac{^{143}Nd}{^{144}Nd} \right]_{CHUR,t} = \left[\frac{^{143}Nd}{^{144}Nd} \right]_{CHUR,today} - \left[\frac{^{147}Sm}{^{144}Nd} \right]_{CHUR,today} \times (e^{\lambda t} - 1)$$

where λ is the decay constant. The composition of the sample at time t is calculated in the same manner as the composition of CHUR in the equation above.

The ϵ_{Sr} value is calculated in the same way as the ϵ_{Nd} value, but is far less commonly used due to the larger range of $^{87}Sr/^{86}Sr$ in rocks and the general uncertainty in the values of $^{87}Sr/^{86}Sr$ for the CHUR. Many authors prefer to use the age-corrected initial $^{87}Sr/^{86}Sr$ ratio instead to eliminate the uncertainty of the ϵ_{Sr} values. The ϵ_{Sr} values of samples from the FCMT are used with caution, due to high Rb/Sr ratios and low Sr contents for some samples. The ϵ_{Nd} and ϵ_{Sr} values which have been deemed reliable are included in Table 5.3; the epsilon values of all samples are included in Appendix III. Note that the epsilon values referenced in the text below are always the initial epsilon value (epsilon value at time of formation) for ϵ_{Nd} and ϵ_{Sr} , corrected using an age of 24.9 Ma.

Table 5.3: Initial ϵ_{Nd} and ϵ_{Sr} Values for Samples from the Fish Creek Mountains Tuff

Sample	Type	ϵ_{Sr}	$^{87}Sr/^{86}Sr$	ϵ_{Nd}	$^{143}Nd/^{144}Nd$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
10-BV-04A	Pumice	23.29	0.706128	-1.26	0.512542	-	-	-
10-BV-04B	Pumice	21.16	0.705977	-1.13	0.512548	-	-	-
10-BV-04C	Pumice	-	-	-1.44	0.512532	19.214	15.679	38.900
10-BV-04D	Pumice	-	-	-1.16	0.512546	19.218	15.721	38.997
10-BV-18	FCMT	20.89	0.705959	-0.91	0.512559	19.187	15.681	38.927

10-BV-19	FCMT	18.31	0.705777	-0.98	0.512555	-	-	-
10-BV-23	FCMT	-	-	-0.91	0.512559	-	-	-
10-BV-25	FCMT	18.39	0.705782	-1.05	0.512552	19.209	15.673	38.891
10-BV-27	FCMT	-	-	-1.12	0.512549	-	-	-
10-BV-33	Fiamme	-	-	-0.83	0.512564	19.174	15.672	38.891
10-BV-36	FCMT	-	-	-0.46	0.512583	-	-	-
10-BV-37	FCMT	-	-	-0.96	0.512557	-	-	-
10-BV-38	FCMT	21.69	0.706015	-1.17	0.512546	19.187	15.658	38.851
11-BV-01	Fiamme	-	-	-0.82	0.512564	19.170	15.663	38.851
11-BV-03	Fiamme	18.38	0.705781	-0.89	0.512560	-	-	-

Note: All initial values are age-corrected using an age of 24.9 Ma.

5.3.3: Pb Isotopic Data

Samples of whole-rock FCMT and of pumice and fiamme in the FCMT are plotted with several other nearby volcanic units in the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ isotope correlation diagram of Figure 5.20. The Northern Hemisphere Reference Line (NHRL; Hart, 1984) plotted in Figure 5.20 reflects the Pb-isotopic values of the modern convecting mantle of the northern hemisphere, and has a slope with an age significance of 1.77 Ga. Basalt samples from the Pliocene-Quaternary Buffalo Valley volcanic field plot closest to this line, followed by samples of whole-rock FCMT and the pumice and fiamme of the FCMT, samples of pre-FCMT Tertiary volcanic rocks, and samples of whole-rock Caetano Tuff. The nearness of each unit to the NHRL correlates with the age of the unit, with the Buffalo Valley samples (mostly ca. <4 Ma) plotting nearest and the Caetano Tuff and Pre-FCMT samples (both ca. 35 Ma) plotting furthest from the line. The isotopic composition of the average upper continental crust is also plotted in Figure 5.20; the average upper continental crust has a higher $^{207}\text{Pb}/^{204}\text{Pb}$ ratio than most of the samples, but has a similar $^{206}\text{Pb}/^{204}\text{Pb}$ ratio to most samples plotted in Figure 5.20.

Pumice and fiamme in the FCMT have greater diversity in Pb isotope ratio values than whole-rock samples of FCMT; pumice and fiamme from the lower cooling unit are more radiogenic than the fiamme from the upper cooling unit, as well as all of the whole-rock samples of FCMT from both cooling units. The ca. 24 Ma pumice and whole-rock samples from the FCMT have a broader range in values

than the ca. 35 Ma pre-FCMT Tertiary rocks and Caetano Tuff, which plot nearer to the isotopic composition of the average upper continental crust.

Figure 5.20: $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ isotope correlation diagram plotting the isotopic values of whole-rock FCMT, samples of pumice and fiamme from the FCMT, pre-FCMT Tertiary rock from the Fish Creek Mountains (Christopher Stevens, personal communication, 2012), whole-rock samples of Caetano Tuff (Christopher Henry, personal communication, 2012), whole-rock samples of alkali basalt from the Buffalo Valley volcanic field (Wetmore, 2011), and the average isotopic composition of the upper continental crust (Zartman and Haines, 1988). Also shown is the position of the northern hemisphere reference line (NHRL); the slope of the NHRL has an age significance of 1.77 Ga (Hart, 1984). Error ellipses are excluded from the plot to improve visibility of sample points.

CHAPTER 6: DISCUSSION

6.1: Compositional Zonation of the FCMT

Many ignimbrite complexes are chemically and mineralogically zoned (Hildreth, 1981) or compositionally heterogeneous (Wilson et al., 2006). This may be due to changing conditions in the magma chamber over the duration of the eruption, such as variable magma recharge rates, or the result of a magma chamber which was compositionally zoned since its formation due to the sequential stacking of melts according to their buoyancies (Hildreth and Wilson, 2007). Both mineralogical and compositional zonation is observed in the FCMT.

6.1.1: Mineralogical Zonation

The FCMT is mineralogically zoned: it varies from crystal-free ash matrix and pumice at the base of the lower cooling unit, to (ca. 20%) crystals in the ash-matrix at the top of the lower cooling unit, in which the pumice and fiamme remain aphyric. In the upper cooling unit, crystal abundances in the ash-matrix remain high (>20%) and pumices include quartz and feldspar crystals. In the ash matrix, pyroxene is absent in the lower cooling unit but occurs in some samples of the upper cooling unit. Zircon crystals are common and consistently occur in whole-rock samples of the upper cooling unit, but not in the lower cooling unit whole-rock samples.

6.1.2: Chemical Zonation

The FCMT is also geochemically zoned. Changes within and between cooling units for major element oxide and trace element compositions are presented in Figure 5.6 and Figure 5.11, respectively. The first material erupted, forming the base of the lower cooling unit, and the last remaining material erupted, forming the top of the upper cooling unit, are geochemically distinct. With respect to major element oxides, the last material erupted is enriched in TiO_2 , Al_2O_3 , CaO , K_2O , and P_2O_5 and depleted in

SiO₂, Fe₂O_{3(t)}, MnO, MgO, and Na₂O; the depletion of these oxides is minor, and no other oxide changed in abundance by more than twofold. The enrichment factors of the trivalent REEs steadily increase from La to Yb, with the exception of Eu, with a crossover from relative depletion to enrichment between the top of the volcanic pile and the base of the pile occurring between Gd and Tb. The pattern of increase in the enrichment factors ends at Lu, which decreases; this is a characteristic of most high-silica systems (Hildreth, 1981). With respect to ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁷Pb/²⁰⁴Pb, the whole-rock sample from the upper cooling unit is more radiogenic than the whole-rock samples from the lower cooling unit. The opposite is true for pumice and fiamme samples of the FCMT; pumice samples from the lower cooling unit are more radiogenic with more of a crustal signature than the pumice sample from the upper cooling unit.

6.1.3: Chemical Evolution of the Magma

Relationship between Position in the Volcanic Pile and the Magma Chamber

Chemical variations in the stratigraphic column of the FCMT reflect vertical variations in the magma chamber which produced the FCMT. This is due to the eruption process, wherein the top of the magma chamber was evacuated first, forming the base of the FCMT. As the eruption proceeded, lower portions of the magma chamber were progressively tapped as magma moved up to replace magma that was erupted. The top portion of the FCMT then represents material from lower in the magma chamber. Thus, analysing the chemical variation of the FCMT from bottom to top provides a first order analysis of the chemical variations in the magma chamber from the top of the chamber downwards.

Cooling Rates of the Lower Cooling Unit of the FCMT

Each major cooling unit of the FCMT is composed of several distinct flow units, some of which have poorly welded bases when separated from the underlying flow by a short cooling break. The upper and lower compound cooling units are separated by a complete cooling break in Dacies Canyon, which

transitions to a partial cooling break before disappearing. The presence of an unwelded base for the upper cooling unit suggests that the lower cooling unit had cooled completely prior to the deposition of the upper cooling unit. Cooling rates for pyroclastic-flow deposits are very slow, ranging from ca. $10^{-6}^{\circ}\text{C s}^{-1}$ to ca. $10^{-8}^{\circ}\text{C s}^{-1}$ for the interior of the deposit and as high as ca. $10^{-1}^{\circ}\text{C s}^{-1}$ to ca. $10^{-4}^{\circ}\text{C s}^{-1}$ for material cooled against the cold ground (Wallace et al., 2003). Using these cooling rates, the lower cooling unit may have decreased in temperature by 300°C after 9.5 to 950 years, with the base of the cooling unit cooling after 50 minutes to 35 days. The upper constraint on the cooling time assumes a cooling rate of $10^{-8}^{\circ}\text{C s}^{-1}$, which is modelled after the Bishop Tuff (Wallace et al., 2003). The FCMT is a considerably smaller and thinner unit, and may have cooled more rapidly than the Bishop Tuff, perhaps in less than 100 years. During this cooling break volcanism was suspended, and the magma chamber underwent several changes over the course of this hiatus.

Recharge of the Magma Chamber

During the eruption of the lower cooling unit, the magma chamber experienced repeated recharge. This is indicated by the SiO_2 concentrations of the rocks of the lower cooling unit, in which the SiO_2 content is cyclic; in Figure 6.1, the concentration of SiO_2 fluctuates from higher (75-78 weight %) to lower (73-75 weight %) values over very small changes in elevation. In this way, the magma varies from more evolved to less evolved frequently within the lower cooling unit, suggesting that pulses of mafic magma were recharging the magma chamber. The concentrations of several incompatible elements in Figure 6.1, including Nb, and Th, and total alkalis ($\text{K}_2\text{O} + \text{Na}_2\text{O}$, not included in Figure 6.1) behave in the same manner as SiO_2 , with La concentrations varying opposite of SiO_2 concentrations. The Fe/Mg and Na/Ca ratios mirror the behaviour of the other elements discussed, with more evolved magma presenting higher Fe/Mg and Na/Ca ratios; the observed ratios support the influx of mafic magma into the reservoir and the resulting changes in the degree of evolution of the magma. Furthermore, in the

examination of a traverse in Chapter 5.2.2, several elements and oxides, including Nb, Th, Ba, U, Y, and SiO_2 , show a shift within the top of the lower cooling unit towards the concentrations observed in the upper cooling unit. The trends of other elements and oxides along the traverse are less distinct, with La, K_2O , and P_2O_5 concentrations fluctuating more widely. This may suggest the intrusion of new magma as the upper part of the lower cooling unit was deposited. There is insufficient evidence of mingling, however, indicating that input occurred at deeper levels than those tapped by the eruption, with mixing occurring below the evacuation zone.

Figure 6.1: Variations in SiO_2 (weight %), La (ppm), Nb (ppm), and Th (ppm) abundances of whole-rock samples of FCMT with elevation. Solid points were collected from the lower cooling unit of the FCMT, hollow points were collected from the upper cooling unit of the FCMT. Samples from the lower cooling unit have fluctuating concentrations of these elements or oxide, while the abundances in the samples from the upper cooling unit generally increase or decrease in a more linear fashion.

Across the major cooling break, a change in whole-rock geochemistry occurs. The first whole-rock material erupted after the break is lower in SiO_2 , La, and total alkalis, higher in Th, and considerably higher in Nb. This composition suggests that during the cooling period, while the reservoir was sitting,

the magma chamber was recharged with mafic material, producing a less evolved magma. The ratios of Fe/Mg and Na/Ca are also lower after the cooling break. Samples collected immediately adjacent to the cooling break have very similar REE patterns in Figure 5.9, and are also isotopically similar in $^{143}\text{Nd}/^{144}\text{Nd}$, suggesting that the recharging magma was derived from a similar parental magma as the FCMT and is not from a different source. The available isotopic data are not ideal for this comparison, as no data are available for Pb isotopes and no $^{87}\text{Sr}/^{86}\text{Sr}$ values can be used reliably for samples at the cooling break.

The material of the upper cooling unit is much less variable chemically than the material of the lower cooling unit. As discussed above, the base of the upper cooling unit is less evolved than the top of the lower cooling unit; the base of the upper cooling unit is poorer in SiO_2 , La, and total alkalis, and enriched in Nb and Th. The Fe/Mg and Na/Ca ratios are also lower at the base of the upper cooling unit. Within the body of the upper cooling unit, the concentration of SiO_2 increases slightly with height, as do La and the total alkalis. The abundances of Nb and Th decrease. The Fe/Mg and Na/Ca ratios also increase slightly. The chemistry of the upper cooling unit suggests that mafic recharge did not play as significant a role in the magma chamber as it did during the formation of the lower cooling unit. This may only be an apparent effect and not a true effect, as a bias in the field sampling allowed for a larger sample base for the lower cooling unit and a much smaller sample base for the upper cooling unit, resulting in poorer resolution in the upper cooling unit.

6.2: Origin of the FCMT Pumice

The REE patterns of the pumices from the FCMT change with position in the volcanic pile, reflecting changes with depth in the magma chamber. The frothy pumices occur in the base of the lower cooling unit, representing original magma from the top of the magma chamber; these pumices are some of the first material erupted from the magma chamber. In Figure 5.8, the REE patterns for these frothy pumices show a slight enrichment in light REEs compared to heavy REEs. The aphyric fiamme which

occur in the upper portions of the lower cooling unit have similar REE patterns, except with a much more pronounced negative Eu anomaly.

The aphyric fiamme sample from the upper cooling unit (11-BV-03) has a significantly different REE pattern, in which the light REEs are much more enriched than the heavy REEs, and the negative Eu anomaly is much less pronounced. The fiamme sample also has a negative Nb anomaly that is not present in pumice from the lower cooling unit. The REE pattern for the aphyric fiamme is similar to the patterns for the whole-rock samples, in that the fiamme is enriched in light REEs relative to heavy REEs and has a similar Eu and Nb anomaly. Thus, the magma which was tapped to produce the pumice of the upper cooling unit now had a chemistry which was very similar to that of the ash matrix.

In Figure 5.11D, the enrichment-depletion diagram comparing the REE concentrations of whole-rocks immediately adjacent across the major cooling break indicates that while the magma chamber was resting, it became less evolved. Incompatible elements, being excluded from uptake into crystallizing mineral structures, become more concentrated in the melt as the melt evolves. In many mineral phases, the heavy REEs are more compatible than the light REEs, and thus are present in lower concentrations in evolved magmas, and in higher concentrations in magmas which are less evolved. The partition coefficients of REEs for mineral phases typically associated with mafic melts are presented in Table 6.1; the partition coefficients of most minerals are generally <1 , with partition coefficients of heavy REEs marginally higher than light REE, resulting in a slight deficiency of heavy REEs relative to light REEs in the evolving melt over time. In Figure 5.8, the REE patterns of the pumices and fiamme in the lower cooling unit are much flatter than the pattern of the fiamme in the upper cooling unit. The fiamme of the upper cooling unit has higher light REE concentrations and lower heavy REE concentrations than the samples of the lower cooling unit, indicating that the first material erupted out of the chamber during the formation of the upper cooling unit was more evolved than the last material erupted for the formation of the lower cooling unit.

Alternatively, the variations in pumice and whole-rock geochemistry may be interpreted as the occurrence of two distinct magmas co-existing in the chamber which are being simultaneously evacuated, one of which formed the pumices of the lower cooling unit while the other produced the ash matrix.

Table 6.1: Rare Earth Element Mineral/Melt Partition Coefficients of Select Mineral Phases for Basalts

	Olivine	Orthopyroxene	Clinopyroxene	Hornblende	Plagioclase	Apatite
La	-	-	-	-	-	8.6
Ce	0.0069	0.02	0.15	0.20	0.120	11.2
Pr	-	-	-	-	-	-
Nd	0.0066	0.03	0.31	0.33	0.081	14
Sm	0.0066	0.05	0.50	0.52	0.067	14.6
Eu	0.0068	0.05	0.51	0.40	0.340	9.6
Gd	0.0077	0.09	0.61	0.63	0.063	15.8
Tb	-	-	-	-	-	15.4
Dy	0.0096	0.15	0.68	0.64	0.55	-
Ho	-	-	-	-	-	13.3
Er	0.0110	0.23	0.65	0.55	0.063	-
Tm	-	-	-	-	-	-
Yb	0.0140	0.34	0.62	0.49	0.067	8.1
Lu	0.0160	0.42	0.56	0.43	0.06	-

Note: Olivine, orthopyroxene, clinopyroxene, and plagioclase partition coefficients from Fujikami et al. (1984), hornblende partition coefficients from Green and Pearson (1985), apatite partition coefficients from Paster et al. (1974).

6.2.1: Magma Evolution Model

Within the magma chamber, the areas which experience the most rapid cooling are along the chamber walls; the core of the chamber remains insulated. As the magma cools along the walls, minerals crystallize out of solution according to the magma chemistry and the Bowen's reaction series. The resulting magma is chemically 'evolved' from the parent, and less dense. Due to a buoyancy difference, this magma then travels upwards along the chamber walls and accumulates at the top of the reservoir (Christiansen et al., 1988). Therefore, the top of the magma chamber contains the most highly

evolved magma, with less evolved magma at depth. This evolved magma is typically aphyric, as the nucleation sites for mineral growth are depressed by the high volatile content (Christiansen et al., 1988).

The most notable change in the FCMT is the progression from a crystal-free ash matrix and aphyric pumices at the base of the lower cooling unit to a crystal-bearing ash matrix with aphyric pumices and fiamme at the top of the lower cooling unit, to crystal-rich ash matrix and crystal-bearing pumices throughout the upper cooling unit. This change suggests that during the eruption of the lower cooling unit, the magma chamber was erupting material at a rate that did not exceed the rate of accumulation of the aphyric layer at the top of the magma chamber. During the early eruptions of the upper cooling unit, the aphyric layer was exhausted and the eruption process drew material primarily from the deeper crystal-bearing zone of the magma chamber, as in Figure 6.2.

Figure 6.2: Progressive evacuation of the magma chamber and withdrawal of the aphyric magma layer. Hot magma cools and begins to crystallize along the chamber walls, driving evolution of the magma in that zone of the chamber. The less-dense, aphyric, evolved magma travels up the chamber walls and accumulates at the top of the chamber (Hildreth, 1979). Magma mixing occurs below the evacuation zone (EVAC). Figures are not to scale. A) Evacuation of the magma chamber during the eruption of the lower cooling unit, in which only material from the aphyric zone is tapped. B) Evacuation of the magma chamber during the eruption of the upper cooling unit, in which evacuation rates exceed accumulation rates of aphyric evolved magma, and less-evolved crystal-bearing magma is tapped.

In Figure 5.15, the values of the Eu, Zr, and P anomalies vary widely within the lower cooling unit, and generally decrease in size upsection within the upper cooling unit. Although Figure 5.15 is not to scale, it provides an estimation of the processes occurring in the magma chamber. During the eruption of the lower cooling unit, the magma chamber had sufficient time between eruptions to form differentiated layers and to become fractionated. The regular evacuations of the upper cooling unit prevented such layers from forming, and the resulting rocks are less fractionated.

Fractionation of Elements into Select Mineral Phases

During partial melting or fractional crystallization, elements will partition into the melt phase or mineral phase according to the element's compatibility in the melt; incompatible elements remain in the melt while compatible elements are accepted into the crystal structure of the mineral. Elements which partition into mineral crystals are then removed from the melt, resulting in lower concentrations of those elements compared to more incompatible elements in the final product. The negative Eu, P, and Zr anomalies discussed above are an expression of this partitioning between phases. The major phase which fractionates Zr out of the melt is zircon; P fractionates into apatite, and Eu fractionates into feldspar and apatite more so than any other rare earth element (Rollinson, 1993).

Accessory zircon and apatite are observed in the matrix of the FCMT, and although they occur as accessory phases (<1%), their high partition coefficients can result in a significant influence on the REE patterns of the samples. Zircon crystallization and removal works to deplete the heavy REEs (Green, 1994), while apatite partitions the middle REEs preferentially to the light and heavy REEs (Rollinson, 1993). The fractionation of garnet from the melt produces the same effect on the REE chemistry as the fractionation of zircon, but is dismissed from consideration due to the flat trends of the REE patterns for pumice samples from the FCMT as well as the curving-to-flat, middle to heavy REE patterns in all samples.

In order to produce the REE patterns observed for samples of the FCMT, the fractional crystallization of several mineral phases is required. Great Basin basaltic magmas are enriched in light REEs relative to heavy REEs, resulting in a primitive mantle-normalized REE pattern with a steep negative slope (Fitton et al., 1991; Rogers et al., 1995). The REE patterns for the pumices and fiamme of the lower cooling unit of the FCMT are nearly flat, however, where normalized heavy REE concentrations are only slightly lower than normalized light REE concentrations but there is a very slight increase in the normalized middle REE concentrations, with large negative Eu anomalies. The REE pattern from the single fiamme sample of the upper cooling unit is more similar to that of local Tertiary basalt, in that the light REEs are enriched relative to the heavy REEs, with a smaller negative Eu anomaly and a negative Nb anomaly (Christopher Stevens, personal communication, 2012).

6.2.2: Multiple Magma Model

Several properties of the FCMT suggest that multiple magmas may have coexisted in the magma chamber, at least during the eruption of the lower cooling unit. Notably, the presence of aphyric fiamme hosted in a crystal-bearing ash-matrix in the lower cooling unit suggests that two liquids were being erupted simultaneously: a crystal-bearing liquid which produced the ash matrix, and a crystal-free liquid which produced the fiamme. Geochemically, the aphyric pumice and fiamme are distinct from the whole-rock samples of the lower cooling unit. The whole-rock samples have sloped REE patterns with moderate negative Eu and Nb anomalies, while the pumice samples have flatter patterns with more pronounced negative Eu anomalies but no Nb anomaly. Isotopically, the pumice samples of the lower cooling unit have higher $^{208}\text{Pb}/^{204}\text{Pb}$ ratios, higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than the whole-rock samples from the lower cooling unit.

In the upper cooling unit, the pumices in the FCMT begin to bear crystals of the same composition and relative abundance as the ash-matrix. The fiamme of the upper cooling unit is also

markedly different geochemically than the fiamme and pumice of the lower cooling unit, with a sloped REE pattern and negative Nb anomaly in Figure 5.14 and Sr- and Nd-isotopic compositions similar to the whole-rock FCMT samples of the upper cooling unit in Figure 5.19. These factors combined suggest that during the eruption of the upper cooling unit, only one liquid was tapped during the eruption. It is unknown whether the second magma was locally or entirely exhausted by the eruption of the lower cooling unit and thus absent in the upper cooling unit.

6.2.3: Magma Mingling

The only feature of the FCMT which may represent mingling is the two-toned white-and-grey frothy pumice present in the basal zone of the lower cooling unit (a two-toned pumice is visible in the upper left section of Figure 4.3A). Variation exists within the single-toned frothy pumices of the lower cooling unit. The two grey pumice samples have similar geochemical and isotopic properties. The major element and trace element geochemistry as well as the isotope ratios of the two white pumice samples are markedly different from one another, with one sample (10-BV-04C) plotting above the trachyte field of Figure 5.1 and well into the shoshonite field of Figure 5.2; sample 10-BV-04C is also the only sodic sample collected from the FCMT. The REE pattern for the other white pumice (10-BV-04D) is distinct from the other white and grey pumice samples in Figure 5.8, wherein the pattern for 10-BV-04D is slightly more sloped for the light REEs and more flat for the heavy REEs, and the Eu anomaly is slightly more pronounced; the REE pattern for sample 10-BV-04D crosses the patterns for the other three frothy pumice samples at Er and Tm. The differences within the single-toned pumices suggest that more than one liquid may have been tapped to produce the white and grey pumices, and that the two-toned pumices may represent mingling of these two liquids. However, no samples of two-toned pumice were collected, and therefore no data exists to prove or dismiss this pumice as evidence of magma mingling.

6.2.4: Summary of the Origin of the FCMT Pumices

The pumices in the lower cooling unit of the FCMT have distinct geochemical and petrographic characteristics. The REE patterns of these pumices are distinct from the whole-rock samples of the lower and upper cooling units and also of the single fiamme sample collected from the upper cooling unit in that the lower cooling unit pumices have flat patterns while all others have negative slopes. The pumices are aphyric while the welded portion of the lower cooling unit consists of crystal-bearing ash-matrix. The lower cooling unit pumices are isotopically distinct from all other samples; although there is some overlap in the data sets, the lower cooling unit pumices generally have higher $^{87}\text{Sr}/^{86}\text{Sr}$ values, lower $^{143}\text{Nd}/^{144}\text{Nd}$ values, higher $^{208}\text{Pb}/^{207}\text{Pb}$ values and lower $^{207}\text{Pb}/^{204}\text{Pb}$ values than all other samples. The fiamme from the upper cooling unit is chemically and isotopically distinct from the pumice samples of the lower cooling unit, and much more similar to the whole-rock of the upper cooling unit.

In the magma evolution model for the pumices of the lower cooling unit, the pumices are presumably derived by crystallization along the chamber walls, producing an evolved liquid which would pond at the top of the magma chamber due to differences in buoyancy. The high volatile content of this evolved liquid could act to suppress nucleation of crystals, creating an aphyric liquid. The presence of aphyric fiamme in the crystal-bearing ash matrix of the lower cooling unit may suggest that the eruption tapped both the aphyric liquid at the top of the magma chamber and the less evolved crystal-bearing material beneath it, and that viscosity differences between the two materials prevented complete mixing, i.e. the magmas mingled. The simultaneous tapping of both liquids would result in the presence of the aphyric fiamme in the crystal-rich matrix of the upper cooling unit.

However, the REE patterns of the pumices of the lower cooling unit suggest that the liquid which produced the lower cooling unit pumices was not derived via crystallization and evolution of the lower rhyolitic melt. Replicating the REE patterns of the lower cooling unit using a basaltic magma or the material of the upper cooling unit (representative of the less evolved material lower in the magma

chamber from which the liquid which produced the pumice was presumably derived) is very difficult due to the enrichment of light REEs relative to heavy REEs. The mineral phases which would be expected to fractionate out of these liquids, including pyroxene, and feldspar, have low partition coefficients and only feldspar preferentially removes light REEs relative to heavy REEs. Fractionation of mineral phases in order to produce the liquid which formed the LCU pumices does not account for the isotopic differences between the pumices of the lower cooling unit and the rest of the FCMT assemblage.

The alternative model is the co-existence of multiple liquids in the magma chamber, which can account for the flat REE patterns on the lower cooling unit pumices and their distinct isotopic ratios. The change in REE and isotope chemistry between the lower and upper cooling unit pumices indicates that the liquid which produced the lower cooling unit pumices was locally or regionally exhausted prior to the eruption of the upper cooling unit; the pumices of the upper cooling unit bear crystals of the same mineral phases and relative abundances as the ash-matrix, the REE pattern of the upper cooling unit fiamme are very similar to those of the whole-rock samples.

6.3: Origin and Source of the FCMT

In order to determine if the FCMT is genetically linked to melting of the mantle, the crust, or to other nearby units, the trace element chemistry (REE and select other trace elements) of the pumices are normalized to the composition of different potential sources. Figure 6.3 shows the extended trace element patterns of the pumice and fiamme normalized to potential magma sources, including the primitive mantle, the upper continental crust, and the lower continental crust. A genetic relationship between a source and the magma that formed the FCMT, represented by uncontaminated pumice and fiamme, would result in a normalized trace element pattern with approximately no slope, provided the magmas have not undergone fractionation of a mineral that fractionates the trace elements. In Figure 6.3, each comparison indicates that the magma which formed the FCMT was significantly enriched in Th,

U, and to a lesser extent Nb, compared to the potential source magma reservoirs. The remainder of the elements occur in concentrations which produce a variably flat line in the normalization patterns.

The extended trace element patterns of the pumice and fiamme inclusions normalized to other nearby volcanic units, including a ca. 34 Ma rhyolite dome of the northern Fish Creek Mountains, a ca. 34 Ma high-magnesium basaltic andesite from the northern Fish Creek Mountains, the ca. 34 Ma pumices in the neighbouring Caetano Tuff, and a ca. 5 Ma basalt flow from the northern Buffalo Valley volcanic centre are presented in Figure 6.4. The rhyolite of the northern Fish Creek Mountains and the pumice of the Caetano Tuff represent units that are chemically similar and may have been sourced from a similar region. The basaltic andesite from the northern Fish Creek Mountains represents a mantle melt that is approximately the same age as the northern Fish Creek Mountains rhyolite and the Caetano Tuff. The young basalt flow from the northern Buffalo Valley represents mantle-derived magma which has no contribution from the continental crust. The normalization patterns of Figure 6.4 using the northern Fish Creek Mountains rhyolite and the Caetano Tuff are virtually identical, suggesting that the latter two units may be related by similar processes. The normalized pattern relative to the high-magnesium basaltic andesite and the basalt from the northern Buffalo Valley volcanic field are also virtually identical after Nb, with less of an arc in the heavy REE patterns for the Buffalo Valley-normalized patterns. In each normalization pattern, the magma of the FCMT is enriched in U, Th, and often Nb, depleted or slightly enriched in the light REEs, and progressively enriched in the heavy REEs. The curved slope of the light REEs and heavy REEs in each normalized pattern indicates that the FCMT was not derived from the same source, fractionation, and assimilation as the magmas which produced the northern Fish Creek Mountains, the Caetano Tuff, and the Buffalo Valley volcanic field, and that the FCMT is unique in the region.

6.3.1: Model 1 - Mantle Melting/Crustal Assimilation Model

In Figure 6.3, normalization to the primitive mantle produces a pattern with the flattest line. Normalization to the upper continental crust produces a flatter line than normalization to the lower continental crust. The flatness of the primitive mantle-normalized line suggests that the magma which produced the FCMT may have originated in the mantle. The increased flatness of the upper continental crust-normalized line relative to the lower-continental crust-normalized line, and the lower enrichment of Th and U in the FCMT compared to the upper continental crust, suggest that the mantle-sourced magma later incorporated material from the intermediate or upper continental crust; however, mantle-sourced magma constitutes a larger portion of the melt than crustal-sourced magma. In Figure 6.3, the primitive mantle-normalized abundances of REEs in the pumice samples are only ca. 10x the mantle abundances, which is equivalent to what a 10% partial melt of the mantle would have. If the FCMT parental magma was derived from the mantle that then evolved by fractional crystallization, such that a rhyolite is <10% residual liquid from basalt (e.g., Clague, 1978), then fractional crystallization of an REE-rich phase such as apatite must be significant to account for the low normalized REE patterns. The REE patterns of the pumices normalized to the upper and lower continental crust in Figure 6.3 are only enriched by ca. 1x, suggesting that near whole-scale melting of the crust would be required, which is not likely.

Figure 6.3: Pumice and fiamme inclusions from the FCMT normalized to magma reservoirs. A) Primitive mantle (chemistry from Sun and McDonough, 1989); B) upper continental crust (chemistry from Taylor and McLennan, 1981); C) lower continental crust (chemistry from Weaver and Tarney, 1984).

Figure 6.4: Pumice and fiamme inclusions from the FCMT normalized to nearby volcanic units. A) The ca. 34 Ma basaltic andesite in the northern Fish Creek Mountains (Christopher Stevens, personal communication, 2012). B) Young (<4 Ma) basalt from the northern Buffalo Valley volcanic centre (Wetmore, 2011).

C)

D)

Figure 6.4 (continued): Pumice and fiamme inclusions from the FCMT normalized to nearby volcanic units. C) The ca. 34 Ma basaltic andesite in the northern Fish Creek Mountains (Christopher Stevens, personal communication, 2012). D) Averaged pumice inclusions from the ca. 34 Ma Caetano Tuff (Christopher Henry, personal communication, 2012).

Mafic Precursor Magma for the FCMT

One flaw in this mantle-derived magma model is the apparent absence of mafic volcanics which are approximately coeval with the FCMT. However, field relationships and geochemistry suggest that at least one basaltic andesite in the northern Fish Creek Mountains may be related to the FCMT. This basaltic andesite has not been dated, but is a potential basaltic precursor because the vent for this basalt appears to overlie the ca. 25 Ma rhyolite vents of the northern Fish Creek Mountains. Geochemically, it is distinct from other basaltic andesites in the northern Fish Creek Mountains in that it has higher REE, Cr, Ni, and TiO_2 concentrations, and lower K_2O values. The $^{87}\text{Sr}/^{86}\text{Sr}$ values of this unit are higher than the other basaltic andesites, and the $^{206}\text{Pb}/^{207}\text{Pb}$, $^{207}\text{Pb}/^{207}\text{Pb}$, and $^{208}\text{Pb}/^{207}\text{Pb}$ values are lower (Christopher Stevens, personal communication, 2012). The relationship of this basaltic andesite to the FCMT is uncertain and remains to be confirmed.

Mantle-Crustal Mixing Curve Models

In order to develop a first-order approximation of the mixing ratio between mantle-derived magma and crustal-derived material, a mixing model is presented in Figure 6.5. Several mantle and crustal end-members are modelled in Figure 6.5; mantle end-members include basalt flows from the northern Buffalo Valley volcanic field (Wetmore, 2011), ca. 34 Ma basaltic andesite from the northern Fish Creek Mountains (Christopher Stevens, personal communication, 2012), and mantle-derived basalt from the southern Basin and Range (Hedge and Noble, 1971; Gans et al., 1989). Crustal end-members include granitoids of the Great Basin (Farmer and DePaolo, 1983; Gans et al., 1989) and granitoids of the Buffalo Valley volcanic field (Wetmore, 2011). These end-members were selected to approximate the isotopic composition of the crust and the underlying mantle; actual values for the mantle and crust at the Fish Creek Mountains are unknown and these proxies must be used. The basalts of the northern Buffalo Valley (<3 Ma) and the southern Great Basin (mostly <10 Ma) are young and uncontaminated,

and are intended to represent mantle values; the basaltic andesite of the northern Fish Creek Mountains represents a mantle melt which is approximately the same age as the northern Fish Creek Mountains rhyolite and the Caetano Tuff. The granitoids of the Buffalo Valley area were selected because of their proximity to the Fish Creek Mountains, and the granitoids of the southern Great Basin were selected because of the availability of data and because of their spatial relationship to the basalts of the southern Great Basin. Granitoids in particular were selected to model the crust because they were likely derived from the lower continental crust and assimilated components of the upper continental crust, including sedimentary rocks.

In Figure 6.5, samples from the FCMT do not plot near the mixing line for two plots where the crust is modelled using granitoids of the Buffalo Valley area. Samples from the FCMT have higher $^{143}\text{Nd}/^{144}\text{Nd}$ values than the pre-FCMT basaltic andesite and the southern Great Basin basalt, and the samples plot well above the mantle end-member in each of these plots. For the mixing curve involving granitoids of the Buffalo Valley and basalt from the northern Buffalo Valley area, the samples plot below the mixing curve and are closer to crustal values, near the 40% mantle-60% crustal marker. Perry (1993) places crustal assimilation in the Great Basin at generally less than 30% for the time at which the FCMT erupted; it is considered improbable that the mantle-derived magma which produced the FCMT assimilated more than 50% crustal material due to regional crustal assimilation patterns, and thus this scenario is excluded from consideration.

Mixing curves involving granitoids from the Great Basin as the crustal end-member show more potential as good first-order approximations of magma mixing for the FCMT. Again, models which include the pre-FCMT basaltic andesite or the basalt of the southern Great Basin have mantle $^{143}\text{Nd}/^{144}\text{Nd}$ values which are lower than the FCMT samples, indicating that these mantle end-members are not suitable for modelling mantle-crust magma mixing for the origin of the FCMT. The mixing curve which uses the granitoids of the Great Basin as the crustal end-member and the basalts from the

northern Buffalo Valley volcanic field as the mantle component is the most successful mixing model available for the FCMT; in this plot, samples from the FCMT plot along the mixing curve near the 80% mantle, 20% crust value. Although the models in Figure 6.5 are only first-order approximations of possible magma mixing sources for the FCMT, a parental magma which consisted of a ca. 20% crustal component is not unlikely for the FCMT. It should be noted, however, that the Buffalo Valley basalts are Pliocene to Quaternary in age, and no Tertiary mafic rock with Buffalo Valley-like chemistry has been found in the Fish Creek Mountains area (Christopher Stevens, personal communication, 2012).

The Sr and Nd Isotopic Character of the FCMT and Magma Mixing End-Members

According to the magma mixing model proposed above, a mantle end-member best approximated using the Buffalo Valley alkali basalts assimilated a crustal end-member best approximated by the average Great Basin granitoid. The Buffalo Valley basalts plot within the depleted mantle quadrant in Figure 6.6, and the Great Basin granitoids plot within the enriched quadrant. The FCMT, which may represent a mixing of the two magmas, also plots in the enriched quadrant.

Figure 6.5: Magma mixing curves for Sr and Nd isotope ratios modelling multiple mantle and crustal end-members. Small grey diamonds represent 10% increments in source volume contribution; large grey diamonds represent end-members. Mantle end-members include northern basalt flows from the Buffalo Valley volcanic field (Sr: 496 ppm, Nd: 33 ppm; Wetmore, 2011), ca. 34 Ma basaltic andesite of the northern Fish Creek Mountains (Sr: 995 ppm, Nd: 44 ppm; Christopher Stevens, personal communication, 2012), and young (<10 Ma), uncontaminated mantle-derived basalt of the southern Basin and Range (Sr: 400 ppm, Nd: 20 ppm; Hedge and Noble, 1971; Gans et al, 1989). Crustal end-members include granitoids of the Great Basin (Sr: 300 ppm, Nd: 40 ppm; Farmer and DePaolo, 1983; Gans et al., 1989) and granitoids of the Buffalo Valley volcanic field (Sr: 605 ppm, Nd: 37 ppm; Wetmore, 2011).

Figure 6.6: ϵ_{Nd} - ϵ_{Sr} ($^{143}Nd/^{144}Nd$ vs. $^{87}Sr/^{86}Sr$) isotope correlation diagram with crustal and mantle reservoirs. Average values of initial ϵ_{Nd} and initial $^{87}Sr/^{86}Sr$ for whole-rock FCMT and pumice and fiamme plot on the border of the mantle array zone (defined by the range of isotopic ratios of the mantle) and the field for magmas sourced from young upper crust. Samples of felsic to intermediate pre-FCMT Tertiary rocks (Christopher Stevens, personal communication, 2012) plot within and near the end of the mantle array zone (age-corrected to 24.9 Ma). Samples of basalts from the northern Buffalo Valley volcanic field (age-corrected to 24.9 Ma) plot within the mantle array zone. Great Basin granitoids plot in the enriched quadrant of crustal reservoirs (Farmer and DePaolo, 1983; Gans et al., 1989). Figure modified from DePaolo and Wasserburg, 1979, and Brownlow, 1996.

Neodymium Crustal Index Values and Assimilation-Recharge Rates

As discussed above, partial melting of a crustal source may have contributed significant quantities of material to a mantle-derived magma, producing the hybrid magma which formed the FCMT. The crustal contribution to a mantle-derived magma can also be expressed using the Neodymium Crustal Index (NCI), which describes the fraction of crustal Nd in a sample. The NCI value is determined using the ϵ_{Nd} values of the sample, the mantle, and the local crust, and is defined by the following equation (Perry, 1993):

$$NCI = \frac{\epsilon_{Nd}(\text{Sample}) - \epsilon_{Nd}(\text{Mantle})}{\epsilon_{Nd}(\text{Crust}) - \epsilon_{Nd}(\text{Mantle})}$$

If the ϵ_{Nd} value of the sample is equivalent to the ϵ_{Nd} value of the underlying mantle, then all of the Nd in the sample was derived from the mantle and no crustal Nd was contributed to the sample. In this case, the NCI is equal to 0. If the ϵ_{Nd} value of the sample is equivalent to the ϵ_{Nd} value of the local crust, then all Nd in the sample has a crustal origin, and the NCI is equal to 1. The NCI is not a direct representation of the mass of crustal material contributed to the magma, but is a useful approximation of relative crustal contributions and provides an insight into local shifts in source material with time. In Nevada, NCI values of rhyolites consistently decrease with time. Oligocene rhyolites of the Great Basin (35-26 Ma) have NCI values ranging from 1 to 0.4, while all systems which are Miocene in age or younger (<20 Ma) have NCI values of less than 0.3. The shift in NCI values over time suggests that crustal contributions of Nd to Great Basin rhyolitic magma sources have decreased significantly since the Oligocene, during which some systems derived all Nd from a crustal source. The decreasing contribution of crustal Nd since the Oligocene is largely attributed to cooling of the continental crust on a regional scale. The lower crust may have been ca. 300°C during the Miocene, resulting in less contribution of crustal material because: 1) the crust was further from the solidus temperature and thus less melt was produced, and 2) failure of the crust would have occurred as brittle fracturing, which allowed for the rapid ascent of basaltic magma without fractionation and with less crustal assimilation (Perry, 1993).

The NCI values of the samples collected in this study are listed in Table 6.2. The value of ϵ_{Nd} of the crust in the region of the Fish Creek Mountains is unknown; the ϵ_{Nd} value of the neighbouring Kalamazoo system (east of the Fish Creek Mountains) was substituted (Perry, 1993). The value of ϵ_{Nd} of the mantle beneath the Fish Creek Mountains is also unknown; the ϵ_{Nd} value of the mantle was approximated by young (ca. 5 Ma) basalt flows of the northern Buffalo Valley volcanic field (Wetmore, 2011). The NCI values of whole-rock samples of FCMT and of pumice and fiamme range from a minimum of 0.14 to a maximum of 0.24. These values are only a first-order approximation of the NCI value of samples from the Fish Creek Mountains, due to the data substitutions discussed above. Nevertheless,

these values indicate that a significant amount of crustal material was contributed to the mantle-derived magma which produced the FCMT, and that basaltic magmas derived from the mantle had sufficient time to fractionate to a rhyolitic composition.

Table 6.2: Neodymium Crustal Index and Assimilation-Recharge Rate Values for Samples from the Fish Creek Mountains

Class	Sample	ϵ_{Nd} Of Sample	Northern Buffalo Valley Basalt ϵ_{Nd} of Mantle: 2.47 to 3.75				Lithospheric Mantle ϵ_{Nd} of Mantle: -3.27		Pre-FCMT Basaltic Andesite ϵ_{Nd} of Mantle: -2.25 to -6.71			
			NCI		(a/r) _{ss}		NCI	(a/r) _{ss}	NCI		(a/r) _{ss}	
			Min. ^A	Max. ^B	Min. ^A	Max. ^B			Min. ^C	Max. ^D	Min. ^C	Max. ^D
Frothy Pumice	10-BV-04A	-1.44	0.18	0.23	0.22	0.30	-0.14	-0.12	-0.06	-0.48	-0.06	-0.33
	10-BV-04B	-1.16	0.18	0.22	0.21	0.29	-0.15	-0.13	-0.07	-0.49	-0.07	-0.33
	10-BV-04C	-4.90	0.19	0.24	0.24	0.31	-0.12	-0.11	-0.05	-0.47	-0.05	-0.32
	10-BV-04D	-0.92	0.18	0.23	0.22	0.29	-0.14	-0.13	-0.07	-0.49	-0.06	-0.33
Aphyric Fiamme	10-BV-33	-0.99	0.16	0.21	0.19	0.27	-0.17	-0.14	-0.09	-0.52	-0.08	-0.34
	11-BV-01	-0.92	0.16	0.21	0.19	0.27	-0.17	-0.14	-0.09	-0.52	-0.08	-0.34
	11-BV-03	-1.05	0.16	0.21	0.20	0.27	-0.16	-0.14	-0.09	-0.51	-0.08	-0.34
FCMT, UCU	10-BV-18	-1.12	0.17	0.21	0.20	0.27	-0.16	-0.14	-0.08	-0.51	-0.08	-0.34
	10-BV-19	-0.82	0.17	0.22	0.20	0.28	-0.15	-0.13	-0.08	-0.51	-0.07	-0.34
	10-BV-27	-0.47	0.18	0.22	0.21	0.29	-0.15	-0.13	-0.07	-0.49	-0.07	-0.33
FCMT, LCU	10-BV-23	-0.96	0.17	0.21	0.20	0.27	-0.16	-0.14	-0.08	-0.51	-0.08	-0.34
	10-BV-25	-1.18	0.17	0.22	0.21	0.28	-0.15	-0.13	-0.08	-0.50	-0.07	-0.33
	10-BV-36	-0.81	0.14	0.19	0.17	0.24	-0.19	-0.16	-0.11	-0.55	-0.10	-0.36
	10-BV-37	-1.11	0.17	0.22	0.20	0.28	-0.16	-0.14	-0.08	-0.51	-0.08	-0.34
	10-BV-38	-0.90	0.18	0.23	0.22	0.29	-0.14	-0.12	-0.07	-0.49	-0.06	-0.33

Note: The value of ϵ_{Nd} of the crust in the region of the Fish Creek Mountains is unknown; the ϵ_{Nd} value of the neighbouring Kalamazoo system (east of the Fish Creek Mountains; $\epsilon_{Nd}=-18$) was substituted (Perry, 1993). The ϵ_{Nd} value of the mantle beneath the Fish Creek Mountains is also unknown; the ϵ_{Nd} value of the mantle was approximated by the following sources: young (ca. 5 Ma) basalt flows of the northern Buffalo Valley volcanic field (Wetmore, 2011), lithospheric mantle of the Basin and Range (Kempton et al., 1991), and by ca. 34 Ma basaltic andesite of the northern Fish Creek Mountains. A: Calculations correspond to ϵ_{Nd} of Buffalo Valley sample 08-BV-12 (Wetmore, 2011); B: Calculations correspond to ϵ_{Nd} of Buffalo Valley sample 08-BV-09 (Wetmore, 2011). C: Calculations correspond to ϵ_{Nd} of basaltic andesite sample 10-BV-40 (Christopher Stevens, personal communication, 2012); D: Calculations correspond to ϵ_{Nd} of basaltic andesite sample 10-BV-42 (Christopher Stevens, personal communication, 2012).

In order to provide a comparison between models, the NCI values were calculated for two other mantle end-members: 1) lithospheric mantle-derived basalts beneath the western Great Basin (Kempton et al., 1991), and 2) ca. 34 Ma basaltic andesites of the northern Fish Creek Mountains, which represent mantle melts which are the same age as the Caetano Tuff. Each of these end-members produce negative NCI values, which are not possible in reality and indicate that no crustal material was

assimilated in the mantle-derived magma. This is unlikely, as the crust was warm at ca. 24 Ma and easier to melt partially than it is today. Another possibility is that the crust and mantle have the same isotopic ratios, in which case no method exists for measuring how much crustal melt is present in the magma.

In addition to the NCI, Perry (1993) also discusses the recharge, assimilation, and fractional crystallization (RAFC) of rhyolitic systems in the western United States, as originally defined by DePaolo (1985). Perry found that most rhyolites in the western U.S. showed very little variation (less than 3 ϵ_{Nd} units) in their ϵ_{Nd} values, despite being erupted over a period of several million years. Perry interpreted this lack of variation as a quasi-steady state of assimilation and recharge rates in the magmatic systems, producing a characteristic isotopic composition. The steady state assimilation/recharge ratio $(a/r)_{ss}$ can be calculated for a magma chamber undergoing RAFC using the following equation (DePaolo, 1985):

$$\left(\frac{a}{r}\right)_{ss} = \frac{C_r(\epsilon_m - \epsilon_r)}{C_a(\epsilon_a - \epsilon_m)} = \frac{C_r}{C_a} \left(\frac{NCI}{1 - NCI} \right) = \left(\frac{NCI}{1 - NCI} \right)$$

where

- C = concentration
- $\epsilon = \epsilon_{Nd}$
- m = magma
- a = assimilated material
- r = recharging magma
- $C_r = C_a$

In the study by Perry (1993), Miocene-Pleistocene rhyolites of the western United States had NCI values of less than 0.3 and $(a/r)_{ss}$ of less than 0.5. The $(a/r)_{ss}$ values of samples from the Fish Creek Mountains are included in Table 6.2; samples of whole-rock FCMT and pumice or fiamme have $(a/r)_{ss}$ values ranging from 0.17 to 0.31 when the mantle is approximated using basalts of the northern Buffalo Valley flows. The negative $(a/r)_{ss}$ values produced using the lithospheric mantle-derived basalts and the ca. 34 Ma basaltic andesite of the northern Fish Creek Mountains are impossible to achieve in reality. The NCI and $(a/r)_{ss}$ values from the FCMT modelled with the Buffalo Valley basalts are consistent with Perry's (1993) observations of the broader western United States.

6.3.2: Model 2 - Mafic Crustal Melting Model

Felsic magmas may be generated by means other than melting of the mantle and fractional crystallization of that basaltic magma towards a rhyolitic composition. Partial melting of a basaltic source to produce dacitic melt which then fractionates to a rhyolitic composition is a possibility and would require less fractional crystallization overall to produce the rhyolitic end member (Rapp and Watson, 1995).

Melting of Mafic Rocks in the Middle to Upper Crust

Most of the pumice samples of the FCMT share many of the characteristics of the FIIIb class of felsic metavolcanic rocks (rhyolites and high-Si rhyolites) as described by Leshner et al. (1986), in which flat REE patterns with large negative Eu anomalies define the group. The flat REE patterns indicate that garnet was not a fractionation product or a residual phase after melting, as garnet works to retain heavy REEs relative to light REEs, producing a REE pattern with a negative slope. Garnet is stable at high pressures (>10 kbar or 1000 MPa for andesitic magma; Green, 1972), suggesting that the magma which produced the FCMT may have formed via melting of a mafic rock followed by fractionation in the upper portions of the magma chamber which were both above the garnet stability field in the middle to upper crust. The large negative Eu anomalies of the FCMT pumice samples and the FIIIb rocks indicate that plagioclase was either retained in the residue or was fractionally crystallized, or both. The retention of plagioclase in the residue would produce a melt with initially low Eu^{2+} abundances; further fractionation of plagioclase from the melt would act to severely deplete Eu^{2+} in the melt, producing the pronounced negative Eu anomalies observed in the pumice samples of the FCMT.

In order to evolve a basaltic magma towards a rhyolitic composition, several mineral phases are crystallized, including feldspars, pyroxene, hornblende, and olivine. The REE partition coefficients of these minerals for basaltic liquids are compared in Table 6.1.

To achieve the flat REE pattern of the FCMT pumice and fiamme, the dacitic magma would need to undergo fractional crystallization of mineral phases which would preferentially remove the light REEs relative to the heavy REEs, assuming the basaltic source is similar in composition to local Tertiary basalts. Of the minerals included in Table 6.1, only plagioclase preferentially removes light REEs. Orthopyroxene and clinopyroxene prefer to incorporate the heavy REEs into their structures, and fractional crystallization of these minerals would result in a REE pattern which is more negative than the original basalt REE pattern. Hornblende prefers the middle REEs to the heavy REEs and especially to the light REEs; fractionation of hornblende would result in a concave REE pattern with an overall negative slope.

The fractional crystallization of a single mineral phase is insufficient to replicate the REE patterns of the pumice and fiamme of the lower cooling unit; the REE patterns of these pumices and fiamme may be the result of the fractionation of multiple minerals. The large negative Eu anomaly is likely due to the fractionation of a large amount of plagioclase, which would also work to remove light REEs from the system and flatten the REE pattern. A small proportion of pyroxene may have been crystallized, although the very low partition coefficients of the REEs for the pyroxenes would result in little effect on the REE patterns. Apatite also likely fractionated out of the melt in very small quantities; the partition coefficients of apatite are much higher for the REEs than the other minerals listed in Table 6.1, and as such, the fractional crystallization of apatite has a much larger influence on the REE patterns than other minerals for the same volume of crystallization. Thus it is unlikely that the pumices from the lower cooling unit were evolved by fractional crystallization of a lower cooling unit/upper cooling unit matrix parent.

Alternatively, if a steeply-sloped REE pattern of the lower cooling unit/upper cooling unit matrix were to be derived from the flat REE patterns of the FCMT pumice and fiamme, the light REEs would need to become enriched relative to the heavy REEs. The crystallization of hornblende and

clinopyroxene would work to deplete the heavy REEs relative to the light REEs in the residual liquid, but cannot account for the Nb anomaly. In the FCMT system, this progression is the opposite of that seen in the stratigraphy, suggesting that two distinct magmas may have been present in the magma chamber.

Melting of a Middle Crustal Source Similar to the Northern Buffalo Valley Basalts

Several components of the mantle-derived magma model feature the northern Buffalo Valley alkali basalts. Approximating the mantle using the Buffalo Valley basalts produces some of the most fitting explanations for the origin of the FCMT. The NCI calculations using a mantle composition approximated by the Buffalo Valley basalts produce crustal assimilation values which are consistent with measured values in the Great Basin at ca. 24 Ma (Perry, 1993). Magma mixing models are also best approximated using Buffalo Valley alkali basalt as a mantle approximation. Of the normalization patterns in Figure 6.4, the REE patterns of pumices normalized to the Buffalo Valley basalts produce the flattest patterns and the only patterns which do not indicate a significant enrichment in Nb. The enrichment in Th and U in the FCMT pumices is still high (ca. 20x relative to the Buffalo Valley basalts) but considerably lower than the enrichment relative to the primitive mantle or the average lower crust in Figure 6.3. The enrichment in Th and U is assumed to be related to assimilation of granitoids in the upper crust.

It is important to note that the volcanic event which produced the Buffalo Valley alkali basalts did not occur until ca. 20 Ma after the eruption of the FCMT, thus making the actual Buffalo Valley basalts an impossible mafic source rock for the FCMT. However, the source of the Buffalo Valley basalt may have existed in the middle crust by the time of the formation of the FCMT, and some as yet unidentified geological process may have allowed for the melting of this middle crustal source at ca. 25 Ma and again at ca. 5 Ma. However, this is a theoretical situation based on a model, and may have no basis in actuality.

Issues with a Crustal Melting Model

The generation of a dacitic melt from a basaltic source in the middle crust, particularly of a Buffalo Valley type source, which then underwent fractionation and crustal assimilation, provides a model which best approximates the geochemical and isotopic characteristics of the FCMT. However, several issues exist with this model. In order to produce the volume of magma needed to form the FCMT, large-scale melting of the mid-crustal source would be required. A heat source for this type of melting has not been identified; melting of this magnitude would also be expected to produce precursor magmas ranging in composition from basalt to dacite. No precursor mafic rocks have been identified in the Fish Creek Mountains. In the neighbouring Shoshone Range, the Caetano Tuff is underlain by its associated mafic precursor rocks, which are exposed at the surface due to block faulting and tilting of the Caetano Tuff. The FCMT has escaped faulting and remains upright; if precursor mafic rocks underlie the FCMT as they do the Caetano Tuff, they may be obscured by the overlying tuff. The existence of precursor mafic rocks is speculative and has not been proven.

6.3.3: Summary of the Origin of the FCMT

The mantle-derived magma plus crustal assimilation model discussed above provides an excellent approximation for the FCMT system. It provides a very reasonable model for the degree of crustal assimilation, well within the regional observations described by Perry (1993). Samples of pumice from the FCMT plot on a magma mixing curve involving a mantle end-member approximated by the northern Buffalo Valley alkali basalts and a crustal end-member approximated by average Great Basin granitoids.

The mantle-derived magma model is incapable of accounting for the REE geochemistry of the lower cooling unit pumices. If these pumices represent evolved liquid derived from a mantle-derived basaltic magma, the REE patterns are extremely difficult to replicate through fractional crystallization of

the mineral phases expected to crystallize from a typical Great Basin basaltic magma as it evolves towards a rhyolitic composition, namely olivine, pyroxene, and feldspar. The lack of a Nb anomaly is difficult to account for, and the Th and U enrichment is very high in the lower cooling unit pumice samples relative to the both mantle and crustal values, although Th and U concentrations may be due to assimilation of granitoids.

Melting a mafic source in the middle crust is an alternative to the mantle-derived magma model, but a melting mechanism has yet to be identified. If the initial material is basaltic in composition, the resulting melt may be dacitic in composition (Rollinson, 1993). The retention of plagioclase in the source and/or the fractionation of plagioclase would work to produce the pronounced negative Eu anomalies observed in all FCMT samples, particularly the pumices of the lower cooling unit. A melt which is dacitic in composition would also require less fractional crystallization than a basaltic melt, making it easier to crystallize out the amount of feldspar and other accessory minerals necessary to produce the lower cooling unit pumice REE patterns. The production of a dacitic melt from a basaltic source would have to occur at a depth above the garnet stability field, since fractionation of garnet is not a factor for the FCMT magma-

Several features of the FCMT are best modelled using the northern Buffalo Valley alkali basalts as a mantle approximation, and although the ca. 20 Ma hiatus between the deposition of the FCMT and the eruption of the Buffalo Valley basalt cones removes the possibility that the Buffalo Valley basalts played a direct role in the formation of the FCMT, there is no evidence to suggest that the Buffalo Valley source or a very similar source could not have existed at ca. 25 Ma. The formation of the FCMT was an isolated event at ca. 24 Ma, as was the formation of the Buffalo Valley basalts at ca. 4 Ma. Some as yet unknown geological process may have stimulated the melting of this particular source for both of these eruptive events, although no evidence exists to prove a relationship between the two events.

In the case that a Buffalo Valley-type source was the parental material for the FCMT, attributes of the mantle-derived magma isotopic mixing model can now also be explained by the mafic-melt model. Due to the very minor fractionation of Sm and Nd between mantle and basaltic melt, a mafic crustal source beneath the FCMT would have similar isotopic ratios as the mantle source. Thus modelling of mixing between a mafic crust-derived dacite and continental crust would be similar to the mantle magma and crustal assimilation mixing modelling discussed earlier in Figure 6.5.

The Buffalo Valley-type source effectively explains all properties of the FCMT under contention:

- 1) *Difficulties in evolving the REE patterns of the parental material into those characteristic of the lower cooling unit pumices:* Evolving the dacitic melt of a Buffalo Valley-type middle crust source would require less fractionation than a basaltic magma in order to approximate the REE patterns of the lower cooling unit pumices. While exact duplication of the lower cooling unit pumice patterns is still extremely difficult, the fractionation of feldspar required is more possible.
- 2) *The identity of the mafic end-member in the magma-mixing models and the degree of crustal assimilation:* The magma-mixing curves and degree of crustal assimilation in this system are modelled using the Nd and Sr isotopic character of the crustal and mafic end-members of the system, namely Great Basin granitoids and Buffalo Valley alkali basalt cones. The Buffalo Valley-type source would have a very similar isotopic character as the Buffalo Valley alkali basalts used to model crustal assimilation in the magma-mixing curves in Figure 6.5 as well as the NCI calculations, and would produce the same assimilation rates as the Buffalo Valley cones. The assimilation rates modeled using the Buffalo Valley basalt cones are consistent with those determined by Perry (1993) for the Great Basin at ca. 24 Ma, as would be those involving the Buffalo Valley-type mafic source.
- 3) *The enrichment of U, Th, and Nb in the FCMT relative to potential source reservoirs:* The lower cooling unit pumices do not have a negative Nb anomaly relative to the Buffalo Valley alkali basalts, and the Nb concentrations of the lower cooling unit pumices of the FCMT are approximately

equivalent to those of the Buffalo Valley alkali basalts; the REE patterns of lower cooling unit pumices normalized to the Buffalo Valley alkali basalt cones are the only patterns which do not show a significant enrichment in Nb. The lower cooling unit pumices are enriched in Th and U by ca. 20x relative to the Buffalo Valley basalts, which is likely due to assimilation of granitoid from the upper crust and not inherited from melting the middle crust mafic source.

6.4: Comparison of the FCMT to the Bishop Tuff

6.4.1: Compositional Zonation of the Bishop Tuff

One of the best-studied high-silica rhyolite magma bodies in the western United States is the large-volume (ca. 600 km³; Hildreth and Wilson, 2007) Bishop Tuff in California, which is mineralogically and chemically zoned (Hildreth, 1977; Hildreth, 1979; Wallace et al., 1999; Hildreth and Wilson, 2007). The top of the magma chamber for the Bishop Tuff was crystal-poor, gas-rich, highly evolved, and lower in temperature compared to lower in the magma chamber, which was crystal-rich and less evolved. The main mineral phases of the Bishop Tuff are quartz and sanidine; plagioclase, biotite, ilmenite, zircon and apatite are important but less abundant. The volcanic pile of the Bishop Tuff is mineralogically zoned in composition and abundance of crystals. The base of the Bishop Tuff is crystal-poor (<10%), and pyroxene crystals are absent. Moving upsection through the Bishop Tuff, the abundance of crystals increases to ca. 25% at the top of the pile, during which pyroxene crystals appear in the mineral assemblage. The earliest ejected material was erupted at a temperature of ca. 720°C while the latest material was erupted at a temperature of ca. 790°C, with the temperature increasing systematically through the eruptive sequence. Allanite occurs only in material erupted at temperatures ranging from 720°C to 763°C. Pyrrhotite, hypersthene, and augite, as well as other dense phenocryst phases, appear in the sequence at temperatures of 737°C and above (Halliday et al., 1984).

Chemical zonation in the Bishop Tuff is variable, with several key features: 1) major-element gradients were moderate, with changes in abundances generally less than twofold; the Bishop Tuff becomes slightly less SiO₂-rich with increased stratigraphic height; 2) several elements were strongly enriched at the base compared to the top, and include Nb, Ta, and Th; 3) the abundance of Zr increases with stratigraphic height while the abundance of Y decreases and Hf remains approximately constant; 4) several elements were enriched by an order of magnitude or more at the top of the pile relative to the bottom, and include Eu, Mg, P, Sr, and Ba; 5) some light REEs (La, Ce, Nd) are more enriched at the top of the pile than the bottom, while the heavier REEs (Sm to Lu, excluding Eu) are more enriched at the bottom of the pile; and 6) all samples had a negative Eu anomaly, with the magnitude of the negative Eu anomaly decreasing with stratigraphic height (Hildreth, 1981).

The work performed on the Bishop Tuff by W. Hildreth (1977, 1979) has shown that crystal-liquid equilibria has little to no effect on compositional zonation of high-silica rhyolite magma bodies; rather, compositional zonation occurs in the silicate liquid itself. Zonation in the Bishop Tuff is mostly due to sequential stacking of melt batches in the magma chamber by release of melt from an underlying granitoid mush or by extraction of melt from batches of recharge magma deeper in the reservoir which have undergone crystallization, creating a magma chamber which was zoned to begin with. Other processes may have contributed to the zoning in a secondary or local capacity: a) the production of buoyant, water-enriched residual melt through crystallization of magma along the chamber walls; b) inward crystallization along the roof and/or walls; c) removal of crystals from suspension in the melt by settling, dropping from the walls, or by convective flow differentiation; d) magma mixing; e) partial melting of roof rocks and possibly the chamber walls; f) rising bubbles in the gas-saturated magma which may have stabilized a density gradient and prevented rehomogenization while stimulating upward transport of halogens alkalis, and a variety of other trace elements; g) addition of CO₂ into the reservoir

roots, lowering the H₂O activity and promoting crystallization in the deeper chamber, allowing for the release of evolved melts into the upper zones (Hildreth and Wilson, 2007).

6.4.2: Comparison of the FCMT to the Bishop Tuff

The mineralogical and chemical zonation of the FCMT follows many of the same patterns as the Bishop Tuff. The Bishop Tuff is mineralogically zoned, with a crystal-poor base which lacks pyroxene crystals. In the upper sections of the Bishop Tuff, crystal abundances increase to ca. 25%, with pyroxene and other minerals, including allanite, pyrrhotite, augite, appearing in the volcanic pile at temperature-controlled intervals (Hildreth, 1981). Although no temperature constraints are currently available for the FCMT, it is also mineralogically zoned in a similar way, with a crystal-free or very crystal-poor base which becomes more crystal-rich with increased height in the volcanic pile. As with the Bishop Tuff, pyroxene occurs later in the unit; zircon is also common and consistent in the upper parts of the FCMT volcanic pile.

Chemical zonation in the FCMT and the Bishop Tuff also follows similar patterns. In both units, changes in major-element abundances were generally small, and less than two-fold. Negative Eu anomalies were common to both tuffs, with the magnitude of the Eu anomaly generally decreasing with stratigraphic height. The enrichment and depletion of REEs in the FCMT occurred at a smaller scale than within the Bishop Tuff, where trace element abundances vary by an order or magnitude or more. Comparing the top of the Bishop Tuff to its base, the enrichment factors of trivalent REEs steadily increase from La to Yb. For the FCMT, the crossover from depletion to enrichment occurs between Gd and Tb, and ends at Lu; for the Bishop Tuff, the crossover occurs between Nd and Sm (Hildreth, 1981). The REE patterns for the Bishop Tuff are not flat, as the patterns for the lower cooling unit pumices of the FCMT are; rather, the Bishop Tuff patterns have a general negative slope which is steeper for the heavy REEs than the light REEs, and a pronounced negative Eu anomaly (Hildreth, 1977).

The cause of the compositional zonation in the Bishop Tuff is somewhat similar to the cause of compositional zonation in the FCMT. Zonation in the Bishop Tuff is largely attributed to the original zonation of the magma chamber, produced by the sequential stacking of melt from magma recharge batches (Hildreth, 1981). Within the FCMT, compositional variations occur in cycles within the lower cooling unit, varying between more- and less-evolved magma compositions with an overall gradual progression to a slightly less-evolved composition by the top of the lower cooling unit. This could be due to sequential stacking of melts, as is the case for the Bishop Tuff. Compositional variations in the upper cooling unit are more gradational and lack the cyclical frequency of the lower cooling unit. This has been attributed to a more rapid eruption rate than that which occurred during the eruption of the lower cooling unit, preventing the accumulation of discrete zones in the magma chamber due to the frequent tapping and disruption of the magma chamber; the magma chamber processes would likely produce the same cyclical variation in the upper cooling unit if eruption rates were slower. It is important to note that the eruptive volume of the Bishop Tuff (ca. 600 km³; Hildreth and Wilson, 2007) is nearly double that of the FCMT (310 km³; McKee, 1970), which may suggest that the magma chamber of the FCMT was smaller than the magma chamber of the Bishop Tuff. The smaller volume of the FCMT chamber may have served to amplify the cyclical variations in the geochemistry of the LCU, as less magma was present in the FCMT chamber in which to dilute the incoming magma and reduce the effect on the overall geochemistry of the system by the new magma additions.

CHAPTER 7: CONCLUSIONS

7.1: Unique Character of the FCMT

The FCMT is a unit of distinct rhyolite ignimbrite in north-central Nevada, having erupted ca. 9 Ma after the volcanic front associated with the Mid-Tertiary Ignimbrite Flare-Up had passed through the region. The FCMT remains undisturbed by tectonism and metamorphism, having suffered only minor weathering. Excellent exposures of the FCMT are available throughout the range, allowing for easy

sampling. These factors combined allow for a singular opportunity to examine the changes in the FCMT over the course of its eruption.

The FCMT is geochemically distinct from its neighbouring units. The REE patterns of the FCMT lower cooling unit pumices and of samples from the Caetano Tuff, rhyolites and basaltic andesites of the northern Fish Creek Mountains, and alkali basalts from the Buffalo Valley volcanic field are different, indicating that the FCMT does not share a direct source with any of the aforementioned units. The REE patterns of the FCMT are best estimated by either: 1) an asthenosphere-derived basaltic magma with an 'intraplate' composition, or 2) a mafic melt from the middle crust which has Buffalo Valley-type isotopic characteristics, which later assimilated a significant proportion of upper crust (<30%). Trace element data suggests that the FCMT was erupted in a within-plate setting.

7.2: Mineralogical and Geochemical Zonation of the FCMT

The mineralogical zonation observed in the FCMT is very similar to that present in the well-studied Bishop Tuff (Hildreth, 1979; Wallace et al., 1999; Hildreth and Wilson, 2007). Significant zonations in the FCMT include the shift from aphyric pumices in the lower cooling unit to crystal-bearing pumices in the upper cooling unit. The abundance and mineralogy of phenocrysts in the ash matrix also change with stratigraphic height; the base of the lower cooling unit is crystal-free, but the overlying welded zones of FCMT in the lower cooling unit possess quartz and feldspar. The upper cooling unit also contains quartz and feldspar, but in a greater proportion than the lower cooling unit. Several minor or trace minerals increase in abundance and occurrence with height in the volcanic pile, and include pyroxene and zircon.

Geochemical zonation is also present in the FCMT. Relative to the lower cooling unit, the upper cooling unit is slightly enriched in TiO_2 , Al_2O_3 , CaO , K_2O , and P_2O_5 and depleted in SiO_2 , $\text{Fe}_2\text{O}_{3(\text{t})}$, MnO , MgO , and Na_2O ; it is also depleted in all REEs except Ce and Eu with proportionally greater depletion in

the heavy REEs. Changes in elemental abundances in the lower cooling unit are much more cyclical than changes in the upper cooling unit, which progress in a more linear fashion. The same is true of the Eu, Zr, and P anomalies. The whole-rock samples from the upper cooling unit are more radiogenic than the whole-rock samples from the lower cooling unit with respect to $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$, and less radiogenic in $^{87}\text{Sr}/^{86}\text{Sr}$. The combined geochemical properties of the lower and upper cooling unit indicate that the magma which produced the lower cooling unit was more evolved than that which formed the upper cooling unit.

7.3: Evolution of Magma Chemistry and Eruption Style

The eruption of the FCMT occurred in two phases. The lower cooling unit was erupted in sporadic episodes which allowed for the magma chamber to experience differentiation and stratification. Mafic recharge contributed to the chemical signature of the lower cooling unit. During a longer hiatus in the evacuation of the chamber, the reservoir received an input of mafic melt which produced a less evolved magma; during this time, the lower cooling unit cooled completely. The eruption of the upper cooling unit occurred more quickly, with rapid withdrawal of magma from the chamber preventing the accumulation of a layer of evolved magma from forming at the top of the reservoir. This rapid eruption style allowed for the evacuation of less-evolved material from deeper in the chamber, drawing in to the crystal-rich zone of the chamber.

7.4: Modelled Source of the FCMT

Combined major and trace element geochemistry and isotopic data indicate that the melt which produced the FCMT was likely derived from a moderately depleted mantle source or possibly a Buffalo Valley-type mid-crustal mafic source if some mechanism for melting this source could be determined. The mantle-derived basaltic magma underwent fractional crystallization of feldspars, and to a lesser

extent, zircon, apatite, and minor pyroxene and hornblende. The magma incorporated a large volume of upper crustal material (up to ca. 30%, as estimated by the NCI values; ca. 20% by magma-mixing models) as it passed through the warm lithosphere beneath the Fish Creek Mountains. The amount of crustal contamination, and the approximated recharge rate for the magma chamber, are consistent with slightly younger (Miocene; ca. 23 Ma to ca. 5 Ma) rhyolites elsewhere in the Great Basin.

The mineralogical and geochemical differences between the pumice and whole-rock FCMT may be explained by the presence of multiple liquids in the magma chamber. Within the lower cooling unit, the crystal-bearing ash-matrix may represent one liquid in the chamber while the chemically distinct aphyric pumices represent a second liquid with unique flat REE patterns and large negative Eu anomalies. The liquid which produced the pumices may have been locally or regionally exhausted after the eruption of the lower cooling unit, as the pumice in the upper cooling unit bear crystals of the same phases and relative abundances as the hosting ash-matrix, and have REE patterns similar to the whole-rocks of the upper cooling unit of the FCMT. The coexistence of two liquids in the magma chamber is further supported by the differences in the REE patterns of the pumices in the lower cooling unit and that of the presumed parental rock, a mantle-derived basalt from the asthenosphere. The generation of the flat pumice REE patterns through fractional crystallization of a basaltic magma is discounted for the FCMT system due to the degree of crystallization required and the inability to approximate the shape of the REE pattern for the lower cooling unit through fractional crystallization of the mineral phases common in a basaltic magma, such as feldspar and pyroxene.

The proposed model for the source of the FCMT then is of an asthenosphere-derived basaltic magma, or possibly a Buffalo Valley-type mafic melt derived from a source in the middle crust, which has undergone fractionation and assimilation of a proportion of crustal material, and which incorporated a smaller volume of immiscible magma which was derived from another source and which was exhausted during the eruption of the lower cooling unit.

7.5: Recommendations for Further Work

The model for the origin of the FCMT presented in this paper would be strengthened by continued work in the Fish Creek Mountains. The potential precursor basaltic andesites in the northern Fish Creek Mountains should be dated, sampled, and mapped in detail to firmly establish a relationship with the FCMT. As the work done in this study was only reconnaissance-level, detailed mapping and sampling to correct for the unintended sampling bias in this study would be very useful for providing a better resolution on the chemical and physical zonations of the FCMT. Also pertinent is to establish the chemical and isotopic composition of the local continental crust in order to better constrain magma-mixing models.

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APPENDIX I: PETROGRAPHY

Table I: Petrographic Descriptions

Sample	Description				Phenocrysts			Fiamme					
	Lithology	Rock Type	Colour	Welding	Phenocryst	Abundance	Size (mm)	Fiamme (Yes/No)	Max. Size (mm)	Min. Size (mm)	Phenocrysts	Abundance	Size (mm)
10-BV-04A	Rhyolite	Pumice	Grey	Poor	-	-	-	No	-	-	-	-	-
10-BV-04B	Rhyolite	Pumice	Grey	Poor	-	-	-	No	-	-	-	-	-
10-BV-04C	Rhyolite	Pumice	White	Poor	-	-	-	No	-	-	-	-	-
10-BV-04D	Rhyolite	Pumice	White	Poor	-	-	-	No	-	-	-	-	-
10-BV-17	Rhyolite	Tuff	Pink-grey	High	Quartz	15	1-2	Yes	5 x 24	5 x 10	Smokey quartz	10	1-4
					Feldspar	5	1-10				Feldspar	5	1-4
					Smokey quartz	15	1-4						
10-BV-18	Rhyolite	Tuff	Light grey	Moderate	Quartz	20	1-2	Yes	35 x 120	1 x 3	Quartz		1-2
					Feldspar	15	1-6				Feldspar	5	1-9
					Smokey quartz	20	1-3				Smokey quartz	20	1-3
10-BV-19	Rhyolite	Tuff	Purple-grey	Moderate	Quartz	5	1-3	Yes	4 x 22	1 x 6	Smokey quartz	5	1
					Feldspar	7	1-5				Feldspar	2	1
					Smokey quartz	30	1-4						
10-BV-20	Rhyolite	Tuff	Light grey	Moderate	Quartz	10	1-3	Yes	24 x 40	2 x 5	Smokey quartz	5	1-2
					Feldspar	5	1				Feldspar	<1	<1
					Smokey quartz	15	1-3						
10-BV-21	Rhyolite	Tuff	Light grey	Moderate-high	Quartz	10	<1	No	-	-	-	-	-
					Feldspar	15	1-6				-	-	-
					Smokey quartz	5	1-2						
10-BV-22	Rhyolite	Tuff	Light grey	Moderate-high	Quartz	20	1	No	-	-	-	-	-
					Feldspar	5	1-4				-	-	-
10-BV-23	Rhyolite	Tuff	Medium grey	Moderate-high	Quartz	35	1-3	No	-	-	-	-	-
					Feldspar	10	1-4				-	-	-
10-BV-24	Rhyolite	Tuff	Medium grey	Moderate	Quartz	20	1-2	No	-	-	-	-	-
					Feldspar	15	1-5				-	-	-
					Smokey quartz	5	1-2						
10-BV-25	Rhyolite	Tuff	Medium grey	Moderate-high	Quartz	20	1-2	No	-	-	-	-	-
					Feldspar	20	1-11				-	-	-
10-BV-26	Rhyolite	Tuff	Light grey	Moderate-high	Quartz	15	1-2	No	-	-	-	-	-
					Feldspar	10	1-4				-	-	-
					Smokey quartz	2	1-2						
10-BV-27	Rhyolite	Tuff	Light grey	Moderate-high	Quartz	20	1-2	Yes	2 x 9	1 x 5	Aphyric	-	-
					Feldspar	5	1-4						
					Smokey quartz	2	1-2						
10-BV-28	Rhyolite	Tuff	Yellow-grey	Moderate-high	Quartz	15	1-3	Yes	16 x 33	1 x 6	Feldspar	1	4
					Feldspar	5	1-2						
					Smokey quartz	7	1-2						
10-BV-33	Rhyolite	Pumice	Black	High	-	-	-	-	-	-	-	-	-

Table I (continued): Petrographic Descriptions

Sample	Description				Phenocrysts			Fiamme					
	Lithology	Rock Type	Colour	Welding	Phenocryst	Abundance	Size (mm)	Fiamme (Yes/No)	Max. Size (mm)	Min. Size (mm)	Phenocrysts	Abundance	Size (mm)
10-BV-35	Rhyolite	Tuff	Pink-grey	Moderate-high	Quartz	5	1-2	Yes	5 x 42	1 x 3	Smokey quartz	3	1-2
					Feldspar	2	1-3				Feldspar	5	1-4
					Smokey quartz	15	1-2						
10-BV-36	Rhyolite	Tuff	Pink-grey	Moderate-high	Quartz	5	1	No	-	-	-	-	-
					Feldspar	35	1-8						
					Smokey quartz	10	1-2						
10-BV-37	Rhyolite	Tuff	Pink-grey	Moderate-high	Quartz	10	1-3	Yes	6 x 30	2 x 6	Feldspar	<1	1
					Feldspar	2	1-2				Quartz	<1	1
					Smokey quartz	15	1-2						
10-BV-38	Rhyolite	Tuff	Pink-grey	Moderate-high	Quartz	10	1-2	No	-	-	-	-	-
					Feldspar	5	1-3						
					Smokey quartz	20	1-2						
11-BV-01	Rhyolite	Pumice	Black	High	-	-	-	-	-	-	-	-	-
11-BV-02	Rhyolite	Xenolith	Purple-grey	Moderate-high	Feldspar	2	1	No	-	-	-	-	-
					Smokey quartz	5	1						
11-BV-03	Rhyolite	Pumice	Black	High	-	-	-	-	-	-	-	-	-
Sample	Notes												
10-BV-04A	Aphyric frothy grey pumice from unwelded tuff. Vesicles are elongate and aligned.												
10-BV-04B	Aphyric frothy grey pumice from unwelded tuff. Vesicles are elongate and aligned.												
10-BV-04C	Aphyric frothy white pumice from unwelded tuff. Vesicles are elongate and aligned.												
10-BV-04D	Aphyric frothy white pumice from unwelded tuff. Vesicles are elongate and aligned.												
10-BV-17	Matrix is dark pink. Pumice inclusions are either completely devitrified (with spherulites) or pale grey with abundant smokey quartz and Feldspar and lack glassy rims. Possible lithic fragments in the sample (≤ 8 mm, $<1\%$). Feldspar are white and appear fresh.												
10-BV-18	Abundant pumice (40% overall) which are mostly small (2x15mm) with a few larger pumice fragments. Small pumices are very flattened with variably thick glassy rims: pumices with thick rims have dark grey to black cores, pumices with thin rims have light grey cores. Large pumices have thick (1-4mm) glassy rims or are glassy throughout with high Feldspar contents. Larger pumices have mottled grey colours.												
10-BV-19	Colouring is mottled. Phenocrysts occur in both colours. Pumice inclusions are small, somewhat flattened, with glassy brown rims (≤ 1 mm thick). Pumice contains few phenocrysts and are roughly tabular in shape, sometimes curved. Feldspar is yellowish. Small specks of reddish mineral throughout ($<1\%$).												
10-BV-20	Matrix is variably grey in colour. Pumice inclusions are not very flattened, and have rounded ends; the rims (1-3mm) of the pumices are not purely glassy but are darker. Pumice appear to lack clear quartz crystals.												
10-BV-21	Feldspar is altered, yellowish/brownish. Oxidized mineral present. Interior is weathered.												
10-BV-22	Matrix is a mottled grey. Feldspar are yellowish with thin white rims. Rare volcanic clasts (subangular, <1 cm, $<1\%$). Smokey quartz is rare or absent. Small specks of reddish mineral throughout ($<1\%$).												
10-BV-23	Matrix has zones of yellowish weathering. Feldspar crystals are yellow, weathered, and weakly aligned. Smokey quartz is rare or absent.												
10-BV-24	The matrix of the sample is various shades of grey. Pumice inclusions have the same phenocryst contents as the whole-rock. Feldspar is yellowish-brown.												

Table I (continued): Petrographic Descriptions

Sample	Notes
10-BV-25	Sample contains large, well-formed feldspar phenocrysts and rare or absent smokey quartz. Feldspar is slightly yellowish.
10-BV-26	Feldspar is very yellow and weathered. Matrix can be yellowish. Smokey quartz is not very abundant.
10-BV-27	Sample matrix is variable in colour, with two shades of grey mixed together. Lithic clasts are brown (angular-subangular, 2-5mm, <1%) or yellowish-grey with white rims (subangular-subrounded, 3-6mm, <1%). Abundant flattened small fiamme occur which are slightly darker grey than the matrix; larger fiamme have paler cores. Fiamme are aligned. Lithic clasts occur in layers.
10-BV-28	Rock is weathered with zones of yellowish discolouration. Minor lithic clasts: 7mm, rounded, with reddish rims or coatings. Feldspar is yellowish and chalky, not very fresh. Fiamme are elongate with pointed ends and usually have pale cores. Small specks of reddish mineral throughout (<<1%).
10-BV-33	Aphyric black rehydrated fiamme.
10-BV-35	Matrix is a mottley of medium grey, light grey, and pink ash. Feldspar are yellowish. Fiamme are abundant (10-15%) and occur in a wide range of sizes. Thinner fiamme are glassy; larger fiamme have thin (<1mm) glassy rims and pale centres, sometimes with 'worming' Feldspar. Phenocrysts are concentrated in the cores of thinner fiamme and are evenly distributed throughout the larger fiamme.
10-BV-36	Matrix of sample is pinker in centre of the sample. Feldspar crystals are pale yellow and aligned. Very thin crescent-shaped fiamme occur throughout. Small specks of reddish mineral throughout (<<1%).
10-BV-37	Sample has reddish-brown lithic clasts (subangular-rounded, 2-5mm, 1%) and abundant (40-45%) flattened fiamme with glassy rims (2-3mm thick). Thinner fiamme are entirely glassy, thicker fiamme have light grey cores with very rare phenocrysts of quartz and Feldspar. The fiamme are aligned, and have rounded ends.
10-BV-38	Feldspar is yellowish. Small specks of reddish mineral throughout (<<1%).
11-BV-01	Aphyric black rehydrated fiamme.
11-BV-02	Sample is clearly layered with mm-scale layers of alternating light grey and pale pink-grey layers. Phenocrysts mostly occur in the light grey layers. Feldspar phenocrysts are rare. Rare voids occur in the pale pink-grey layers. Small specks of reddish mineral occur (<<1%).
11-BV-03	Aphyric black rehydrated fiamme.

APPENDIX II: MICROSCOPY

Table II: Microscopy Descriptions

Sample	Description	Phenocrysts in Matrix					Phenocrysts in Fiamme	Notes
		Major			Minor	Accessory		
		Qtz (%)	Pl (%)	Sa (%)	Mineral, %	Mineral, %		
10-BV-01	Rhyolite of unknown affinity	15	20	5	Bt, 3 FeOx, 5 Op, 1 Px, 1	Zrn, <1 Ttn, <1	-	Matrix is dark brown in PPL, with some banding. Bt crystals are frayed and have oxides in cleavages. Some plg crystals are zoned. Small lithic fragments with abundant qtz occur.
10-BV-15	Rhyolite of Pinnacle Mtn (?)	10	15	3	Ttn, 1 Px, 3	Zrn, <1 Op, <1	-	Matrix is very fine ash, banded, and appears dark grey in XPL. Crystals are intensely fractured shards.
10-BV-17	Rhyolitic ash-flow tuff (FCMT)	20	15	5	Px, 5 FeOx, 1 Aln, 3	Zrn, <1 Bt, <1 Op, <1	Qtz	Spherulites in fiamme. Minor Fe-oxides in matrix. Moderate alignment of elongate crystals.
10-BV-18	Rhyolitic ash-flow tuff (FCMT)	40	20	-	FeOx, 1	Zrn, <1 Op, <1 Aln, <1	Qtz Rare Pl	Rare spherulites in fiamme, minor Fe-oxides on fiamme rims. No alignment of elongate crystals. Crystal fragments are very angular.
10-BV-19	Rhyolitic ash-flow tuff (FCMT)	25	10	-	FeOx, 7 Op, 3	Zrn, <1	Qtz Sa Plg UM	Fe-oxides rimming fiamme are more abundant. No alignment of elongate crystals. Plg has minor sericitization. Lithic clast contains abundant small phenocrysts of qtz, sa, plg, and an unidentified mineral which are approx. euhedral. Fine opaques occur throughout.
10-BV-20	Rhyolitic ash-flow tuff (FCMT)	15	10	5	Px, 3	Zrn, <1 Op, <1	-	Matrix consists of abundant phenocrysts of qtz and fsp which are coarser than other samples, but very small (<0.5mm).
10-BV-21	Rhyolitic ash-flow tuff (FCMT)	20	15	-	-	Ser, <1 Zrn, <1 Ttn, <1	-	Plg is intensely fractured. No apparent alignment of elongate minerals.
10-BV-22	Rhyolitic ash-flow tuff (FCMT)	25	2	3	Zrn, 1 Ser, 1	Ttn, <1 Bt, <1	Qtz	Rare, large, altered bt phenocrysts in the matrix. Large clusters of ser flakes occurring as rims on minerals, or as partial to complete replacements. Unusually large zrn occur in this sample.
10-BV-23	Rhyolitic ash-flow tuff (FCMT)	40	3	-	Op, 1	Zrn, <1 Ttn, <1 Ser, <1	Qtz	Opaques rim some voids. Ser occurs as clusters, but far less abundant than in 10-BV-22.
10-BV-24	Rhyolitic ash-flow tuff (FCMT)	25	3	-	Op, 1	Zrn, <1 Ser, <1 Ttn, <1	Qtz Ser Pl Ttn	Crystals are larger and more abundant in the fiamme. Fiamme are abundant in this sample. Fiamme matrix is small equant granular qtz.
10-BV-25	Rhyolitic ash-flow tuff (FCMT)	25	5	-	Ser, 1	Op, <1 Bt, <1 Ttn, <1	Qtz	Very small flakes of bt in the ash matrix. Ser occurs as masses and as flakes (muscovite).
10-BV-26	Rhyolitic ash-flow tuff (FCMT)	35	5	-	FeOx, 1	Zrn, <1 Ser, <1 Op, <1	Qtz	Ash matrix is very fine and banded. Plg phenocrysts are intensely fractured, with fragments missing.
10-BV-27	Rhyolitic ash-flow tuff (FCMT)	25	5	-	FeOx, 1	Zrn, <1 Ser, <1 Op, <1	Qtz	Small spherulites in fiamme. Ser occurs as clusters of flakes in the matrix, often rimming other minerals.
10-BV-28	Rhyolitic ash-flow tuff (FCMT)	25	5	-	Ser, 3 FeOx, 3	Zrn, <1 Am, <1	Qtz	Ash matrix is reddish. Zircon is very small and very rare. Ser occurs throughout the matrix as fine flakes, and also as clusters.
10-BV-35	Rhyolitic ash-flow tuff (FCMT)	25	5	-	Ser, 3 FeOx, 3	Aln, <1 Zrn, <1 Ttn, <1 Op, <1	Qtz Rare Ttn	Ash matrix is banded. Plg is slightly sericitized. Ser also occurs as fine flakes and as clusters.
10-BV-36	Rhyolitic ash-flow tuff (FCMT)	35	5	5	FeOx, 3 Ser, 2	Ttn, <1 Op, <1 Zrn, <1	-	Matrix is banded and brown in PPL. Many crystals are angular shards. Elongate minerals are oriented parallel to the banding. Some plg crystals are zoned with ser along internal fractures.
10-BV-37	Rhyolitic ash-flow tuff (FCMT)	20	15	5	FeOx, 5 Ser, 1	Ttn, <1 Op, <1 Zrn, <1	Rare Qtz	Several thin crystal-poor fiamme with distinct spherulites are preserved in this section. The ash matrix is banded and brown in PPL. Elongate minerals are oriented parallel to the banding. Angular lithic clasts with abundant qtz crystals and fine Fe-oxides occur. All minerals appear distressed.
10-BV-38	Rhyolitic ash-flow tuff (FCMT)	25	15	5	FeOx, 5	Ttn, <1 Ser, <1 Op, <1	-	Small, thin, crystal-poor fiamme. Ser rims some voids and occurs as fine-grained clusters. The ash matrix is dark-brown in PPL and banded, with elongated minerals oriented approx. parallel to banding.
11-BV-02	Rhyolite tuff xenolith	3	1	1	-	FeOx, <1 Ttn, <1	-	Very fine matrix with rare crystals. No banding.

Abbreviations:

Aln: Allanite Bt: Biotite Cpx: Clinopyroxene FeOx: Iron-Oxide Op: Opaque Pl: Plagioclase Px: Pyroxene Qtz: Quartz Sa: Sanidine
Ser: Sericite Ttn: Titanite UM: Unidentified Mineral Zrn: Zircon PPL: Plane Polarized Light XPL: Cross-Polarized Light

APPENDIX III: GEOCHEMISTRY

Table IIIA: Major Element Oxide Abundances

Sample	Notes	Oxide (Weight %) by XRF Analysis										LOI
		Al ₂ O ₃	CaO	Fe ₂ O ₃ (total)	K ₂ O	MgO	MnO	Na ₂ O	P ₂ O ₅	SiO ₂	TiO ₂	
10-BV-04A	Frothy white pumice (FCMT)	12.30	1.39	1.39	5.42	0.59	0.02	3.61	0.01	66.98	0.08	7.50
10-BV-04B	Frothy white pumice (FCMT)	12.33	1.66	1.37	5.10	0.65	0.03	4.20	0.01	65.32	0.08	8.70
10-BV-04C	Frothy grey pumice (FCMT)	10.50	3.64	1.11	3.94	1.09	0.01	8.45	0.01	50.02	0.07	19.20
10-BV-04D	Frothy grey pumice (FCMT)	13.56	0.64	1.11	5.59	0.62	0.02	3.89	0.01	66.73	0.04	8.20
10-BV-17	Rhyolitic ash-flow tuff (FCMT)	12.83	0.90	0.49	5.09	0.07	0.00	3.36	0.07	75.54	0.24	0.80
10-BV-18	Rhyolitic ash-flow tuff (FCMT)	12.97	0.96	0.65	5.29	0.08	0.00	3.13	0.05	74.38	0.23	1.20
10-BV-19	Rhyolitic ash-flow tuff (FCMT)	13.04	0.85	0.72	5.20	0.05	0.01	3.17	0.03	74.51	0.25	0.80
10-BV-20	Rhyolitic ash-flow tuff (FCMT)	13.18	0.85	0.80	5.20	0.11	0.01	3.42	0.04	74.07	0.22	0.70
10-BV-21	Rhyolitic ash-flow tuff (FCMT)	13.31	0.79	2.02	5.35	0.14	0.01	3.14	0.04	76.78	0.22	1.50
10-BV-22	Rhyolitic ash-flow tuff (FCMT)	12.50	0.79	0.98	5.10	0.12	0.00	3.15	0.03	75.18	0.18	1.30
10-BV-23	Rhyolitic ash-flow tuff (FCMT)	12.16	0.61	0.78	4.93	0.10	0.00	2.97	0.03	76.59	0.17	1.00
10-BV-24	Rhyolitic ash-flow tuff (FCMT)	12.98	0.74	0.92	5.26	0.16	0.00	3.15	0.04	74.45	0.19	1.30
10-BV-25	Rhyolitic ash-flow tuff (FCMT)	13.05	0.72	1.03	5.19	0.15	0.00	3.07	0.04	74.63	0.23	N/A
10-BV-26	Rhyolitic ash-flow tuff (FCMT)	13.04	0.68	1.24	5.26	0.13	0.00	3.12	0.04	73.98	0.20	1.40
10-BV-27	Rhyolitic ash-flow tuff (FCMT)	13.29	0.98	1.78	5.14	0.25	0.01	2.99	0.04	72.06	0.20	3.70
10-BV-28	Rhyolitic ash-flow tuff (FCMT)	13.19	0.88	1.69	5.16	0.25	0.00	2.74	0.04	72.34	0.21	2.70
10-BV-33	Aphyric black fiamme (FCMT)	12.35	0.36	0.83	6.25	0.03	0.01	2.80	0.01	73.78	0.08	4.20
10-BV-35	Rhyolitic ash-flow tuff (FCMT)	13.42	0.79	1.02	5.38	0.09	0.02	3.48	0.05	73.61	0.20	0.80
10-BV-36	Rhyolitic ash-flow tuff (FCMT)	13.54	0.91	1.66	5.39	0.11	0.01	3.55	0.05	73.05	0.22	0.70
10-BV-37	Rhyolitic ash-flow tuff (FCMT)	12.47	0.47	1.61	4.62	0.11	0.01	3.63	0.06	75.07	0.13	0.50
10-BV-38	Rhyolitic ash-flow tuff (FCMT)	13.12	1.30	2.12	5.54	0.14	0.04	3.29	0.04	71.84	0.22	1.20
11-BV-01	Aphyric black fiamme (FCMT)	12.21	0.41	0.79	6.24	0.03	0.02	2.54	0.00	73.28	0.07	3.60
11-BV-02	Rhyolite tuff xenolith (pre-FCMT)	13.07	0.73	0.83	4.70	0.08	0.01	3.65	0.02	75.14	0.06	3.70
11-BV-03	Aphyric black fiamme (FCMT)	13.14	0.99	1.69	5.18	0.15	0.01	3.48	0.09	73.02	0.25	0.60

LOI: Loss on ignition

Table IIIB: Geochemical Assay Results

Method Element Units	ICP-MS Ba (ppm)	XRF Ba (ppm)	ICP-MS Ce (ppm)	XRF Ce (ppm)	ICP-MS Dy (ppm)	ICP-MS Er (ppm)	ICP-MS Eu (ppm)	ICP-MS Gd (ppm)	ICP-MS Hf (ppm)	ICP-MS Ho (ppm)	ICP-MS La (ppm)	XRF La (ppm)
10-BV-04A	13.2	31	63.58	25	20.21	9.94	0.08	19.80	9.19	3.70	25.96	34
10-BV-04B	20.2	16	67.28	75	19.60	9.59	0.10	19.09	9.11	3.61	26.96	21
10-BV-04C	26.7	27	47.39	39	14.78	6.95	0.15	15.17	7.70	2.66	23.75	22
10-BV-04D	21.4	35	70.46	88	15.01	9.09	0.08	11.86	16.40	2.96	31.80	51
10-BV-17	483.2	475	187.57	189	4.55	2.25	0.45	6.22	3.63	0.82	97.39	108
10-BV-18	468.2	433	182.56	198	4.34	2.03	0.45	6.21	3.62	0.76	105.38	100
10-BV-19	495.7	470	187.12	215	5.10	2.37	0.46	7.08	3.21	0.90	112.04	135
10-BV-20	389.5	352	232.08	220	5.94	2.95	0.40	7.36	3.70	1.08	95.69	108
10-BV-21	397.3	386	196.45	222	5.12	2.39	0.36	7.08	1.98	0.89	103.88	104
10-BV-22	281.2	263	212.40	216	5.54	2.55	0.28	8.13	2.39	0.97	123.41	113
10-BV-23	281.3	261	68.45	88	2.74	1.41	0.25	3.65	2.39	0.50	46.83	33
10-BV-24	374.4	364	185.62	209	4.49	2.09	0.34	6.46	3.38	0.80	104.20	85
10-BV-25	450.6	422	150.35	138	4.03	1.76	0.42	6.00	2.47	0.68	100.22	97
10-BV-26	330.8	320	189.37	186	5.67	2.54	0.31	8.22	3.27	0.99	117.87	122
10-BV-27	320.7	300	144.36	142	6.88	3.51	0.33	8.20	3.51	1.28	83.68	83
10-BV-28	323.9	302	136.03	144	6.08	2.89	0.34	7.68	3.40	1.08	81.39	91
10-BV-33	7.1	1	72.27	55	20.92	10.66	0.01	19.78	8.98	3.90	25.41	32
10-BV-35	354.1	319	185.08	179	6.15	3.11	0.33	8.12	4.50	1.14	99.46	114
10-BV-36	383.6	358	193.92	192	6.61	3.50	0.48	8.85	4.43	1.25	111.93	145
10-BV-37	139.5	105	142.43	151	15.10	8.25	0.14	13.43	7.37	2.92	56.77	56
10-BV-38	439.4	433	179.68	153	6.09	3.17	0.52	7.54	3.79	1.12	95.79	107
11-BV-01	11.1	14	67.67	109	20.26	10.41	0.01	18.80	8.83	3.84	24.08	35
11-BV-02	14.9	5	69.46	65	15.33	8.82	0.25	12.81	4.95	3.04	31.67	38
11-BV-03	471.2	451	196.36	197	7.33	3.79	0.43	8.92	4.05	1.37	110.28	131

Table IIIB (continued): Geochemical Assay Results

Method Element Units	ICP-MS Lu (ppm)	ICP-MS Nb (ppm)	XRF Nb (ppm)	ICP-MS Nd (ppm)	XRF Nd (ppm)	ICP-MS Pb (ppm)	XRF Pb (ppm)	ICP-MS Pr (ppm)	ICP-MS Rb (ppm)	XRF Rb (ppm)	ICP-MS Sm (ppm)	ICP-MS Sr (ppm)
10-BV-01	0.15	12.20	11	29.27	26	19.3	25	8.44	118.77	115	4.82	534
10-BV-04A	1.06	94.16	90	49.12	33	24.5	33	11.00	>300	471	18.01	211
10-BV-04B	1.02	92.59	87	48.71	49	27.2	55	11.20	>300	475	17.47	215
10-BV-04C	0.73	74.73	69	41.69	36	27.6	45	9.72	>300	372	14.53	727
10-BV-04D	1.42	99.15	100	40.11	38	20.6	49	10.89	>300	1055	12.18	96
10-BV-15	0.23	11.57	14	39.05	35	19.0	21	11.24	123.78	120	5.97	349
10-BV-17	0.29	18.10	17	66.56	58	18.4	34	19.47	155.35	157	9.96	89
10-BV-18	0.27	18.75	18	69.47	62	19.3	40	20.66	164.84	167	10.30	92
10-BV-19	0.29	19.11	18	78.52	73	19.6	25	22.94	156.93	157	11.75	83
10-BV-20	0.36	21.19	21	71.09	61	21.0	35	20.86	179.56	178	11.49	69
10-BV-21	0.29	18.85	19	73.64	69	19.8	27	21.55	161.72	168	11.28	70
10-BV-22	0.30	19.97	18	85.79	69	21.9	46	25.20	170.28	171	13.17	52
10-BV-23	0.21	19.95	19	36.00	39	19.1	36	10.48	166.76	170	6.01	48
10-BV-24	0.27	18.63	18	70.39	72	19.9	51	20.88	171.69	178	10.72	64
10-BV-25	0.21	18.92	16	67.56	56	18.9	33	19.97	158.17	155	10.13	73
10-BV-26	0.30	21.87	19	83.88	67	19.7	29	24.17	174.95	174	13.06	59
10-BV-27	0.42	31.45	29	64.41	59	24.5	38	18.61	255.52	250	11.48	74
10-BV-28	0.36	31.17	30	62.45	57	25.0	50	17.94	257.58	255	10.99	88
10-BV-33	1.14	95.01	87	47.16	27	27.1	38	10.84	>300	539	17.39	2
10-BV-35	0.43	26.20	24	73.65	55	24.4	54	21.29	201.99	204	12.05	65
10-BV-36	0.50	21.92	20	84.88	68	21.5	45	24.38	185.60	186	13.48	72
10-BV-37	1.00	61.28	60	55.67	46	43.8	54	15.05	>300	414	13.85	31
10-BV-38	0.42	19.96	20	70.66	52	18.1	49	20.56	174.32	179	11.36	99
11-BV-01	1.12	94.98	92	44.30	50	34.5	46	10.05	>300	605	16.51	3
11-BV-02	1.12	37.05	37	40.32	34	33.8	44	10.05	>300	295	11.70	12
11-BV-03	0.46	22.47	23	74.94	59	25.2	66	21.89	176.49	184	11.96	81

Table IIIB (continued): Geochemical Assay Results

Method Element Units	XRF Sr (ppm)	ICP-MS Ta (ppm)	ICP-MS Tb (ppm)	ICP-MS Th (ppm)	XRF Th (ppm)	ICP-MS Tm (ppm)	ICP-MS U (ppm)	XRF U (ppm)	ICP-MS Y (ppm)	XRF Y (ppm)	ICP-MS Yb (ppm)	XRF Zr (ppm)
10-BV-04A	203	5.7	3.384	>55	82	1.338	22.35	34	95.88	92	7.945	182
10-BV-04B	207	5.7	3.269	>55	90	1.293	21.64	39	91.85	85	7.676	180
10-BV-04C	669	4.9	2.546	>55	69	0.936	14.67	22	67.13	59	5.563	165
10-BV-04D	97	17.9	2.299	>55	76	1.488	22.37	46	109.28	110	10.212	197
10-BV-17	86	1.3	0.829	27.24	45	0.311	7.29	16	22.05	18	1.989	252
10-BV-18	92	1.4	0.800	28.65	51	0.292	9.23	27	19.66	18	1.862	267
10-BV-19	82	1.3	0.937	28.57	40	0.339	5.98	15	22.69	21	2.107	246
10-BV-20	66	1.6	1.051	33.76	50	0.407	7.02	18	28.95	25	2.581	244
10-BV-21	71	1.4	0.922	29.31	40	0.326	5.21	13	24.05	23	2.046	255
10-BV-22	52	1.4	1.045	32.56	52	0.349	4.69	17	26.89	26	2.129	239
10-BV-23	48	1.4	0.498	30.69	47	0.218	5.02	15	12.61	11	1.422	223
10-BV-24	67	1.5	0.852	31.29	57	0.286	5.91	24	20.83	19	1.854	239
10-BV-25	69	1.3	0.755	29.16	47	0.239	4.79	18	17.92	15	1.529	239
10-BV-26	56	1.5	1.065	32.76	47	0.338	5.94	14	27.08	24	2.053	243
10-BV-27	72	2.7	1.191	37.28	50	0.483	6.55	17	36.39	33	3.022	234
10-BV-28	88	2.5	1.079	36.61	56	0.393	6.80	22	28.22	26	2.521	244
10-BV-33	3	5.9	3.400	>55	84	1.430	25.92	36	100.48	94	8.470	172
10-BV-35	66	1.9	1.111	34.78	61	0.449	8.32	27	30.64	29	2.881	250
10-BV-36	74	1.7	1.197	32.88	57	0.508	7.67	28	34.59	34	3.314	250
10-BV-37	29	5.1	2.369	53.92	60	1.178	14.73	23	89.02	84	7.320	197
10-BV-38	101	1.5	1.053	30.02	57	0.442	6.60	27	31.18	29	2.810	253
11-BV-01	5	5.9	3.274	>55	87	1.393	25.55	41	98.01	99	8.236	174
11-BV-02	14	3.8	2.337	42.86	51	1.269	9.95	20	99.07	97	8.025	119
11-BV-03	84	1.7	1.247	31.92	61	0.519	7.03	28	42.57	41	3.167	253

Table IIIC: Analytical Uncertainty of Major Element Oxides and Trace Elements

Major Element Oxide or Trace Element	00-LT-02 This Study (ppm)	00-LT-02 Average, 6 Samples (ppm)	00-LT-02 Average Deviation (ppm)
Al ₂ O ₃	13.67	13.62	0.07
CaO	7.43	7.51	0.12
Fe ₂ O _{3(t)}	13.47	13.43	0.05
K ₂ O	1.83	1.86	0.03
MgO	4.03	4.01	0.03
MnO	0.24	0.24	0.00
Na ₂ O	3.11	3.16	0.04
P ₂ O ₅	1.16	1.18	0.02
SiO ₂	50.28	50.52	0.25
TiO ₂	2.49	2.46	0.02
LOI	1.70	1.55	0.29
Ba	2352.0	2200.5	96.0
Ce	58.62	59.06	0.77
Dy	8.63	8.85	0.11
Er	4.94	4.96	0.02
Eu	3.72	3.79	0.08
Gd	9.54	9.87	0.25
Hf	4.05	4.11	0.04
Ho	1.76	1.77	0.01
La	26.05	26.78	0.37
Lu	0.62	0.64	0.01
Nb	7.83	7.53	0.15
Nd	39.87	39.41	0.25
Pb	6.7	6.9	0.1
Pr	8.44	8.42	0.05
Rb	41.25	41.86	0.33
Sm	9.54	9.47	0.03
Sr	405.52	415.46	5.70
Ta	0.45	0.45	0.03
Tb	1.43	1.44	0.01
Th	3.71	4.08	0.19
Tm	0.67	0.68	0.01
U	1.40	1.53	0.06
V	341	338	10
Y	46.31	47.11	0.45
Yb	4.27	4.33	0.03
Zr	150.01	154.50	2.50

LOI: Loss on ignition

Table IIID: Sr, Nd, and Pb Isotopic Data

Sample	Sr Isotopes				Nd Isotopes				Pb Isotopes		
	Present $^{87}\text{Sr}/^{86}\text{Sr}$	Initial $^{87}\text{Sr}/^{86}\text{Sr}$	Present ϵ_{Sr}	Initial ϵ_{Sr}	Present $^{143}\text{Nd}/^{144}\text{Nd}$	Initial $^{143}\text{Nd}/^{144}\text{Nd}$	Present ϵ_{Nd}	Initial ϵ_{Nd}	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
10-BV-01	0.706823	0.706517	32.98	28.82	0.512420	0.512398	-4.26	-3.85	19.306	15.680	39.061
10-BV-04A	0.708502	0.706128	56.81	23.29	0.512577	0.512542	-1.18	-1.26	-	-	-
10-BV-04B	0.708326	0.705977	54.30	21.16	0.512583	0.512549	-1.07	-1.13	-	-	-
10-BV-04C	0.706846	0.706277	33.30	25.42	0.512566	0.512533	-1.40	-1.44	19.348	15.685	39.107
10-BV-04D	0.717775	0.706635	188.43	30.50	0.512576	0.512547	-1.21	-1.16	19.492	15.734	39.304
10-BV-15	0.707467	0.706968	42.12	35.22	0.512364	0.512344	-5.35	-4.90	19.351	15.686	39.096
10-BV-18	0.707816	0.705959	47.07	20.89	0.512574	0.512559	-1.26	-0.92	19.307	15.687	39.049
10-BV-19	0.707736	0.705777	45.93	18.31	0.512570	0.512556	-1.32	-0.99	-	-	-
10-BV-23	0.709545	0.705921	71.62	20.36	0.512576	0.512560	-1.22	-0.92	-	-	-
10-BV-25	0.708081	0.705782	50.83	18.39	0.512567	0.512553	-1.38	-1.05	19.272	15.676	39.018
10-BV-27	0.709891	0.706337	76.52	26.26	0.512566	0.512549	-1.40	-1.12	-	-	-
10-BV-33	0.948700	0.760524	3466.29	795.43	0.512600	0.512565	-0.75	-0.82	19.415	15.683	39.147
10-BV-36	0.947837	0.945204	3454.03	3416.92	0.512598	0.512583	-0.78	-0.47	-	-	-
10-BV-37	0.720578	0.705952	228.22	20.80	0.512581	0.512558	-1.11	-0.96	-	-	-
10-BV-38	0.707828	0.706015	47.24	21.69	0.512562	0.512546	-1.49	-1.18	19.278	15.662	38.988
11-BV-01	0.970765	0.843767	3779.49	1977.04	0.512601	0.512565	-0.73	-0.81	19.356	15.672	39.060
11-BV-02	0.732881	0.702498	402.86	-28.24	0.512576	0.512536	-1.21	-1.11	19.238	15.700	39.054
11-BV-03	0.708023	0.705781	50.01	18.38	0.512576	0.512561	-1.21	-0.90	-	-	-