**Geochemical Transition from Miocene – Pliocene to Quaternary Arc Volcanism in
the Sierra Nevada, Northern California**

Albert Jacob Stoffers
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Department of Earth Sciences
Ottawa Carleton Geoscience Centre
Carleton University
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DEDICATION

For Dad (03.04.46 - 04.27.09).

“Dad, your guiding hand on my shoulder will remain with me
forever. ~Author Unknown

ABSTRACT

The Northern Sierra region separates, by some 250km, the geochemically distinct lavas of the Southern Ancestral Cascades (Miocene-Pliocene) of the Lake Tahoe-Reno area, and lavas of the Southern Modern Cascades (Quaternary) of the Lassen Volcanic Centre (LVC). While minor geochemical differences in both major and trace elements exist, radiogenic isotope geochemistry provides the most information about the possible sources of the lavas. Isotopic values from the Lake Tahoe-Reno area are indicative of an older, subduction modified lithospheric mantle. Compared to Southern Modern Cascades lavas of the LVC and the Southern Ancestral Cascades lavas, the N.Sierra lavas have slightly more elevated $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ than do the LVC lavas, while exhibiting less enriched values compared to the Lake Tahoe area lavas. These geochemical signatures suggest that the older, more enriched lithospheric mantle may be thinning from south (Lake Tahoe area) to the north (LVC).

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List of Symbols, Nomenclature or Abbreviations

CAB = Calc-alkaline basalts

WIP = Within-plate basalts

LKT = Low – potassium (K) tholeiites

HAOT = High Alumina Olivine Tholeiites

LVC = Lassen Volcanic Centre

MORB = Mid-Ocean Ridge Basalt

OIB = Ocean Island Basalt

LREE = Light Rare Earth Elements

MREE = Middle Rare Earth Elements

HREE = Heavy Rare Earth Elements

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

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

CHAPTER 1 - INTRODUCTION

Roughly 65 percent of subaerial volcanoes worldwide that have erupted over the past 10,000 years have done so in the region aptly known as the Ring of Fire. This assembly of volcanoes can be traced almost entirely around the Pacific Ocean, with volcanic arcs predominantly found around the Americas, Russia, Japan, and Southeast Asia.

Although the Ring of Fire is a direct result of the tectonic phenomenon of subduction, separate arcs (e.g. Marianas vs Chilean arc) may differ from one another based on unique characteristics of each locale. Such characteristics include rate of plate movement, angle at which the two plates collide, and nature of the overriding crust, each of which is critical in differentiating one subduction zone from another. In addition to volcanism, subduction zones also return surface and crustal materials to the mantle, such as oceanic lithosphere, sediments, and seawater, which re-equilibrate with surrounding mantle. According to (Stern et al., 2002), an established and steady subduction zone causes magmatic and tectonic occurrences in the overlying lithosphere, which can be acknowledged as “arc-trench complexes,” often shortened to “island arcs” or “arcs”.

The Modern Cascades Subduction Zone (CSZ), shown in Figure 1, lies off the west coast of North America between northern California and the northern tip of Vancouver Island in Canada. The Cascade Range has been an active arc for approximately 36 million years as a result of plate convergence. It includes nearly 20 major active volcanoes, with over 4,000 separate volcanic vents that include stratovolcanoes, shield volcanoes, lava domes, and cinder cones (Hildreth, 2007).

Figure 1: The Modern Cascades. MM=Mt. Meagher, MR=Mt. Rainier, MH=Mt. Hood, TS=Three Sisters, MLV= Medicine Lake Volcanics, MS= Mt. Shasta and the LVC = Lassen Volcanic Centre, which is the most southern active volcano of the Modern Cascades. Figure is from (Hildreth, 2007).

Holocene volcanism in the Cascades extends from the Garibaldi Volcanic Belt in southern British Columbia to the Lassen volcanic complex in northern California (Madsen et al., 2006) and is the direct product of the segmented Juan de Fuca plate being obliquely subducted at a rate of 4 cm/year beneath North America (Bacon, 1990; Rogers, 1985).

The Cascades Arc is commonly divided into two discrete complexes based on age: The Ancestral Cascades (Figure 2), refers to volcanism that is older than five million years (Dickinson, 1997); and the Modern Cascades Arc which refers to volcanism that has occurred from less than five million years ago and includes present day volcanism (Hildreth, 2007).

Intra-Arc Differences

Pronounced differences in the nature of volcanism occur along the Modern Cascades Arc. In Washington, there are five generally large, widely spaced stratovolcanoes (Mt. Baker, Glacier Peak, Mt. Rainier, Mt. St Helens and Mt. Adams) with only one, Mt. Adams, having significant abundant basaltic lavas (White, 1992). In contrast, Oregon has six generally smaller stratovolcanoes, with a 40-50 kilometer-wide band of basaltic, basaltic-andesite and andesitic lava vents cutting across the entire state. This band of primarily mafic volcanism is composed of lava shields, cinder cones, fissure fed lava fields and smaller stratovolcanoes. South of Crater Lake in Oregon, the Cascade arc curves appreciably to the southeast, towards Lassen Volcanic Centre.

The Cascade Range in Oregon is divided into two physiographic sub-provinces: Western Cascades and High Cascades. The High Cascades sub-province is mainly composed of rocks younger than 3.5 million years and is considered part of the Modern Cascade Range volcanic arc (White, 1992). This province is distinguished by a north-trending belt of upper Miocene - Pliocene to Quaternary volcanic rocks that were erupted on the east margin of the Western Cascades (Dickinson, 1997). Mount Hood, Mount Jefferson, Three Sisters-Broken Top, and Crater Lake (Mount Mazma) are the four major Quaternary volcanic centers along this platform. These major volcanic centers have erupted lava flows and pyroclastic material that varies in composition from basalt to dacite (Cribb and Barton, 1997).

In contrast, the Western Cascades province is an older, deeply eroded volcanic platform lying west of the more recent snow-covered High Cascade range (Conrey et al., 2001). The Western Cascades consist of predominantly Oligocene to Pliocene volcanic and volcanoclastic rocks and are apart of the Ancestral Cascade Range (Noble, 1972) (Dickinson, 1997). Volcanic activity gradually shifted to the east during the Miocene and Pliocene towards the present location of the Modern Cascade volcanoes. (Dickinson, 1997) determined that the Western Cascades began to form 40 million years ago, but have been heavily dissected by erosion and glaciation, with the only evidence remaining being scattered remnants of volcanic necks or plugs which mark former vents. The Modern Cascades Arc cannot be divided into sub-provinces in the state of Washington because the younger volcanic vents are scattered across the more dominant Miocene and older volcanics of Washington, thus leaving little room to discern between the two unique volcanic sequences (Brophy and Dreher, 2000).

Overall, the Western Cascades province roughly coincides, both geologically and geographically, with the Modern Cascade Arc. Most of the lavas situated within the geographical region of the arc, older than 5 million years, are deemed part of the Ancestral Cascades, including all of the Western Cascades. In the Modern Cascade arc, a large variance of tectonic and volcanic settings is present, most of which seem to overlap with one another. Basin and Range extension (Henry and Perkins, 2001), arc volcanism resulting from the subducting Farallon (Juan de Fuca) plate (Dickinson, 1997), contamination by Sierra Nevada granites (Figure 3) and possibly, plume volcanism related to the Columbia River flood basalts (Coe et al., 2005) can all be argued as having an influence on Ancestral Cascade volcanism. The area around Lake Tahoe and Reno, Nevada, are referred to as the Southern Ancestral Cascades, (Cousens et al., 2008) are an extension of the Western Cascades of Oregon. Rocks of the Southern Ancestral Cascades have been extensively studied by using geochemistry, geology, and, to a lesser extent geochronology (Cousens et al., 2008; Henry and Perkins, 2001). These studies have determined that the volcanic rocks from the Southern Ancestral Cascades are petrographically similar to Ancestral Cascade volcanic rocks found to the northwest, along the arc (Cousens et al., 2008). Additionally, there are no Modern Cascade volcanic rocks found in the area from the Lake Tahoe – Reno district to northwest of Susanville, California, the area that is the focus of this study.

Figure 3: Simplified map of the geology of California (Observatory, 2009). Area outlined in black represents the 2008 field area of the N. Sierra.

Figure 4: Proposed slab boundaries for the Juan de Fuca slab over the past 30 Ma. As expected, the age of samples selected shows that the boundary of the southern edge of the subducted JDF slab is progressing northwards. Figure from (Cousens et al., 2008).

Cessation of the Ancestral Cascade Arc around the Lake Tahoe area occurred about 3 Ma as the Mendocino triple junction migrated to the north of the region (Atwater and Stock, 1998). The arc resumed its volcanic activity 150 kilometers east and 200 kilometers north of Lake Tahoe, at the Lassen Volcanic Centre (Hildreth, 2007), which still continues today. This progression of volcanism may be traced geochronologically back to the southern edge of the subducted Farallon plate migrating northwards to its present location under the Lassen Volcanic centre (Figure 4).

In an area roughly 200 kilometers long, by 150 kilometers wide, extending from Lake Tahoe to Susanville (Figure 3, outlined area) at an azimuth of roughly 335- 345°, is a geographical gap that contains no Modern Cascade volcanic rocks, while Ancestral Cascade rocks are abundant. This gap is referred to as the North Sierra (N.Sierra) and it is the fundamental theme of this thesis to investigate this gap, as little detailed work has been done in this area. With the aid of extensive data from the Ancestral Cascade range around the Lake Tahoe area (Dickinson, 1997; Henry and Perkins, 2001) (Cousens et al., 2008) and from Southern Modern Cascade lavas (Borg et al., 2002; Hildreth, 2007; White, 1992), this study has the primary goal of examining the volcanic rocks petrographically and geochemically. A strong focus on comparing rocks of the N.Sierra to both the Ancestral (south of the study area) and Modern (north of the study area) Cascade lavas will be used in determining whether or not this area may be considered a spatial or temporal transition zone between the two arcs. Sr, Pb, Nd, and O isotopes will be used in determining the source characteristics of the lavas found, which will be compared to surrounding arc lavas.

Although subtle, geochemical differences occur between the Ancestral Arc lavas from the Lake Tahoe area and the Modern Cascades lavas from Lassen, which may either suggest a simple temporal difference or a geographic gradation within the mantle.

CHAPTER 2 – SUBDUCTION ZONE PROCESSES

As subduction zones worldwide have become more comprehensively studied, two end member types of subduction have been classified by type locality: the Marianas-type (Ocean-ocean Island Arc) and the Chilean-type (Ocean-Continent Arc) (Figure 5). While comparing the Marianas subduction zone in the West Pacific to the Chilean subduction zone in the east, (Uyeda and Kanamori, 1979) and (Hamilton, 1994) articulated the geological and tectonic differences between the two. A steeply dipping slab characterizes the Mariana type, and a shallow-dipping slab distinguishes the Chilean type. The Marianas is situated in an ocean setting and is known for, but certainly not limited to, its deep open trench (up to 11 kilometers deep) that subducts old, cold Jurassic crust; a very steep Benioff Zone; extensive volcanism which is mainly basaltic with only minor andesites; little or no sedimentary accretion at the trench; and typically a quiet eruptive style.

The Chilean type is situated in a continental setting and may be classified by a shallower trench (only up to 6 kilometers) that subducts younger, warmer, oceanic lithosphere; a Benioff Zone with a very shallow dip down to about 200 km, followed by a steeper deeper portion below a seismic gap; plutonism being dominant over volcanism; volcanism that is dominantly of andesite-dacite-rhyolite type with basalts being much rarer; violent eruptive style, high viscosity lavas and extensive volcanoclastic dispersal fans; and has thick (~70km) continental crust as the overriding plate.

Figure 5: Diagram showing the general differences between a continental arc setting (Chilean type) and an Island Arc setting (Marianas type). From (Elliott et al., 1997).

Magmas directly related to subduction processes (Figure 6) ('arc magmas') are commonly chemically fractionated and porphyritic, (Tamura et al., 2002; Turner et al., 2000), particularly when compared to hot spot or mid-ocean ridge magmas. Andesites have long been identified as the main rock-type related to subduction zone magmatism. The term was first used to describe intermediate volcanic rocks of the Andes Mountains in Bolivia and Chile. Continental and oceanic arcs have become better understood in the last two decades, and it has been discovered that andesites are erupted from more than 80 percent of arc volcanoes (Gunn, 1974; Hamilton, 1994; Tatsumi, 2005). While andesite is the most dominant rock-type of mature continental arcs, it is the least voluminous rock-type of immature oceanic arcs (Tamura et al., 2002).

Describing arc lavas geochemically may be the most complicated aspect of subduction zones, although the geochemistry is insightful. Arc lavas are unlike Mid-Ocean Ridge Basalts (MORBs) or Ocean Island Basalts (OIBs), where partial melting of the mantle produces dominantly basaltic lavas, which may be modified by limited amounts of fractionation, mixing and contamination by oceanic lithosphere only. In contrast, arc lavas may be influenced by a number of additional geological aspects, such as the amount and type of sediment on the subducting plate, the composition and age of the subducting plate (i.e. fresh/altered basaltic oceanic crust), composition of the mantle wedge, interaction with and the thickness of the overriding lithosphere.

Figure 6: Generic cross section of a subduction zone. Figure from (Hamilton, 1994).

Subduction zone lavas are geochemically distinguished from ocean island lavas and mid-ocean ridge lavas by their elevated abundances of large ion lithophile elements (LILE) and their depletion of the high field strength elements (HFSE), relative to the rare earth elements (REE).

Radiogenic isotopes are especially insightful when dealing with mantle wedge and subducting oceanic lithosphere contributions to arc magmas. There are many different stable isotope systems used by geochemists (i.e B, Hf, O, Be etc), to analyze the origin of magmas and the processes affecting them. Nd, Pb and Sr are useful decay schemes in subduction zones as they exhibit well defined contrasts among the various contributing components of subduction related magmas, such as the mantle, subducted crust and sediments and crustal contamination from the overriding plate. When using these tracers, it is essential that the isotopic compositions of the original components be known.

Components in Subduction Zones

Mantle Wedge and Fluids

The mantle wedge is the part of the asthenosphere directly above the subducting slab and below the overriding plate. Often, the mantle wedge is the area where components from the descending slab are mixed with the convecting mantle to generate magmas, and ultimately to produce new continental crust (Ida, 1983). The convection of the asthenosphere is critical as this is the part of the mantle that interacts with the slab-derived fluids and melts, generating arc magmas. This system is fundamentally different from the other two magma-producing systems, hot spots (mantle plume) and mid-ocean

ridges. Hot spot and ridge melts are associated with upwelling mantle, while melts generated in subduction zones are produced in an environment where the mantle descends due to its loss of heat and viscous drag from descending plate (Brenan et al., 1995). The addition of fluids results in the lowering of the melting temperature of the mantle, which allows for melt generation to take place. It is generally accepted that the asthenospheric mantle (the mantle wedge) overlying the subducted slab moves with the descending plate, resulting in induced convection (Tatsumi, 2005). The aqueous and carbonic fluids released from subducted sediments, crust and mineral dehydration from the slab are constantly released from the subducting slab and leads to metasomatization of the overlying mantle wedge (Figure 7). Movement of fluids away from the slab and toward the corner may be accomplished by several cycles of mantle metasomatism and ascent (Tamura et al., 2002). The addition of aqueous fluids to typical asthenosphere can lead to melting at temperatures several hundred degrees lower than in dry mantle (Elkins-Tanton et al., 2001; Tamura et al., 2002; Tatsumi, 2005). Melt and fluid generation will be more widespread if uncommonly hot asthenosphere is transported into the melting region (Hamilton, 1994). Some elements, including the large-ion lithophile elements (LILE)

Figure 7: A more detailed cross section of a subduction zone, focusing more on the processes that occur as the slab "sinks" beneath the overriding plate. Figure from (Tatsumi, 2005).

(K, Rb, Cs, Sr, Ba, Pb, and U), are transported in hydrous fluids, while others are relatively immobile, such as the high field strength elements (HFSE) (Y, Zr, Hf, Nb, Ta), the heavy rare earth elements (HREE), Ti and the transition elements (Brenan et al., 1995). Distinct enrichments of LILE relative to HFSE are observed on trace element plots of arc lavas, which reflect the mixing of LILE carried in the aqueous fluid with the mantle wedge inventory of HFSE.

Role of Sediment in Subduction Zones

The transfer of sediment to arc lava via slab subduction is a key factor in the discussion of convergent margin volcanism. Vital clues as to magma generation and possible contamination are found in the signatures of arc lavas when geochemically compared to the subducting sediment. Sediments may be the definitive source of many of the unusual chemical enrichments, depletions and other signatures found in arc lavas. In order to demonstrate this, good control on sediment input, in addition to magmatic output is required. The above controls were understood by (Fryer et al., 1999) who worked on Marianas arc lavas, and noticed that the subducting sediment package could not exactly account for the enriched component found in the lavas. In the case of the Marianas (Elliott et al., 1997) showed that the most enriched Mariana lava might be modeled by the addition of 0.3-wt % sediment to an appropriate mantle source. It should be noted that the subducted sediment might only be a 500 m thick layer, whereas there is potentially a 100 km column of higher density mantle above it (Rea and Ruff, 1996). Variations among mobile large ion lithophile (LILE) trace elements (such as K, Sr and Ba) have been interpreted to signify the presence of sediment addition to the mantle

wedge (Plank and Langmuir, 1993)]. The relatively limited range in compositions present, both in lithospheric mantle and oceanic crust, contrasts with the compositional variation of the sedimentary packages being subducted and often provides a unique geochemical input into the zone of subduction. For example, the subduction zones of Peru-Chile and Central America are known to be subducting carbonate sediments, compared to those dominated by terrigenous sediments (e.g. the Aleutians, Cascades, and Nankai) or siliceous sediments (e.g. Kamchatka) (Rea and Ruff, 1996).

Individual subduction zones may demonstrate steep arrays on $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ plots between MORB and/or overlapping with local sediments from the subducting plate (Hamilton, 1994). This allows for the use of Pb isotopes as a possible insight into the role of sediment in magma geochemistry within the mantle wedge. However, the above conclusion is uncertain where the volcanic arc crust may contain a more radiogenic Pb constituent, as is the case of continental arcs (Plank and Langmuir, 1993).

Role of Subducting Slab

The asthenosphere is approximately 1–2% less dense than the overlying lithospheric mantle; it is this disparity that drives plate motion and consequent subduction (Davies, 1999). The density increase in the mantle develops as the lithosphere ages, thickens and cools. This is readily observable in the Pacific Plate, where the observed difference between the lithosphere age and dip of the subducting slab is reflected in the different subduction zones flanking the plate. In the eastern region, subduction of a younger, warmer and less dense plate occurs which relates to a shallow dip, whereas in

the western Pacific basin, a much older, colder and denser oceanic plate subducts steeply, exemplifying the vitally different subduction zones. Oceanic crust, about to descend at Pacific convergent margins, range in age from 170 Ma (east of the Mariana Trench) (Stern et al., 2002) to < 10 Ma (e.g. spreading ridges being subducted off the coast of North America and parts of South America). The variation in subduction zones lead (Jarrard, 1986) to subdivide these zones into seven strain classes; class 1 being strongly extensional and class 7, being strongly compressional. Jarrard (1986) noted that the strain regime present in the overlying plate is strongly linked with the age of the subducted lithosphere and the absolute motion of the overriding plate. Thus, scientists can now classify subduction zones, based on the above criteria, as those subducting old lithosphere (Mariana-type) from those subducting young lithosphere (Chilean-type). Consequently, Jarrard (1986) classifies the Mariana-type subduction zones as strongly extensional (strain class 1) with weak coupling between the upper and lower plates, and Chilean-type subduction zones as strongly compressional (strain class 7) with strong coupling between the two plates.

The density of the subducted crust can influence whether the subduction zone operates normally or fails. Older (> 5 Ma) oceanic crust is invariably subductable. Conversely, subduction of continental crust or (less likely) a young (< 5 Ma), warm oceanic slab, leads to subduction zone failure. Oceanic crust averages 6 km in thickness and is mainly composed of MORB basalt, gabbro and peridotite (Cloos, 1993; Hamilton, 1994; Richter et al., 2002). Fresh MORB is depleted in incompatible trace elements, such as the high field strength elements (HFSE) Nb, Ta, Zr, Hf, and the rare earth elements (REE) and contains almost no water. Subsequent hydrothermal alteration

significantly increases water content, along with K, U and other cations in seawater, as well as the volatile CO₂. It is the subduction of this hydrated crust that is responsible for a large proportion of the water and water-soluble trace elements transported into subduction zones (Wallmann, 2001).

Crustal Contamination in Subduction Zones

The major difference between continental and island arc volcanism is the presence of continental crust in continental subduction zones, which may be up to 70 km thick. The geochemistry of the newly generated mafic magma may be modified as it ascends to the surface. The crust is generally composed of low-density felsic minerals that are enriched in incompatible elements. As the denser, yet thermally buoyant, mantle-derived mafic magma works its way up through the crust, it often undergoes differentiation in areas of stagnation. As heavier minerals precipitate out of the melt and sink, the residual melt becomes more siliceous and heat is released into the country rock. The low melting point of the continental crust allows it to partially melt due to the heat being transmitted from the mafic magma. As partial melting progresses, the incorporation of, or contamination by, the crust alters the geochemistry of the magma body towards it being more siliceous (Ida, 1983). The felsic minerals of the continental crust can sequester large amounts of incompatible elements that will partition out of the minerals and into the magma (Rea and Ruff, 1996). In combination, these properties have led to a diverse assortment of lavas being erupted within continental arcs (Hildreth and Moorbath, 1988).

Figure 8: Abbreviations of Andean volcanic edifices and figure from (Hildreth and Moorbath, 1988) showing how crustal contamination may be linked to the geochemical modification of lavas. In this example, an increase in the K₂O content is positively correlated to an increase in thickness of the continental crust.

An excellent example of crustal contamination is the Andes in the Southern Volcanic Zone (SVZ). As described by Hildreth and Moorbath (1988), the SVZ extends from 33°S to 37°S, between the Andean Cordillera volcanic front and the Pacific Coast of Chile. The SVZ is a unique continental arc subduction zone in that the Nazca oceanic plate subducts everywhere beneath the SVZ at an equal angle and rate. The composition of the continental crust is also fairly constant, while the thickness of the continental crust increases from south to north. Also, regional river discharge transfers a relatively uniform quantity and composition of sediment to the slab prior to subduction (Hildreth and Moorbath, 1988). Geochemical discrepancies attributed to the distance traveled by the magma through continental crust, are easily identified due to the other constant variables mentioned above.

Hildreth and Moorbath (1988) illustrated that potassium oxide (K_2O) has a pronounced increasing concentration moving northward along the volcanic front, while most of the other major oxides such as calcium (CaO), aluminum (Al_2O_3), phosphorous (P_2O_5), and sodium (Na_2O) do not show any variation. Geophysics along the SVZ reveals a Bouguer Anomaly, which positively correlates with K_2O suggesting an increase in crustal thickness moving northwards (Figure 8). This theory on thickening is also supported by an increase in silica (SiO_2) abundance in erupted lavas, from 50-60% in the south to 58-65% in the north. A plot of K_2O vs. SiO_2 shows increasing concentrations of both K_2O and SiO_2 , however K_2O is increasing slightly more quickly than SiO_2 . This K_2O and SiO_2 trend remains constant despite the crystallization of silica-poor phenocrysts of olivine and pyroxene in the south, and silica-rich phenocrysts of biotite and amphibole

in the north that Hildreth and Moorbath (1988) observed. The authors also show that both the LILE's and LREE's increase with an increasing thickness of the continental crust. Their position is supported by isotope geochemistry results: an increasing ratio of $^{87}\text{Sr}/^{86}\text{Sr}$, and a decreasing ratio of $^{143}\text{Nd}/^{144}\text{Nd}$, from south to north, which suggests an increasing continental crust contribution to magmas.

CHAPTER 3 – THE MIOCENE TO MODERN CASCADIA SUBDUCTION ZONE

Geology of the Modern Cascade Arc

The Modern Cascadia Subduction Zone is a chain of active volcanoes, which are late Tertiary to present in age. It extends from Northern California, through Oregon and Washington states, to southern British Columbia and the Garibaldi Volcanic Centre. The southern boundary of the modern arc is currently located at the Lassen Volcanic Centre in Northern California. The Modern Cascade Arc contains roughly 20 composite centers (Figure 1) erupting a range of lavas from basalts to high silica rhyolites (Bacon, 1990; Borg et al., 2002; Brophy and Dreher, 2000). The Cascades Arc is a direct result of the easterly dipping subduction zone where the Juan de Fuca, Gorda and Explorer plates are driven beneath the North American plate (Riddihough, 1984). Along the entire arc front, there are typical major composite volcanoes that have erupted andesites, dacites and some rhyolites, and approximately 2500 smaller monogenetic vents and satellite vents, such as cinder cones, fissure zones and shield volcanoes. The Modern Cascade Arc is currently active with known historic eruptions having occurred at major composite centres such as Lassen Peak (1914-1917), Mt St. Helens (mid 1800's, 1980-86, 2001-2007), Mt. Shasta (1786), Mt. Baker (mid 1800's), Mt. Hood (mid 1800's) and Mt. Rainier (mid 1800's) (Observatory, 2009).

The geological setting of both the Miocene (Dickinson, 1997) and the Modern (Hildreth, 2007) Cascade Arc have been extensively documented, and while there are many similarities between the two, both the Quaternary and Miocene arc show some

geological variations. The Cascade Range, as noted earlier, is normally separated into two parts based on the foundation of age and manner of volcanism, however, over the past 5 million years, noticeable changes have occurred within the modern arc. Hildreth (2007) recorded that north of Mount Rainier young volcanism is restricted to only a few solitary andesitic and dacitic composite cones (notably Glacier Peak, Mount Baker, and the volcanoes of the Garibaldi belt in British Columbia). South of Mount Hood, average-sized andesitic and dacitic composite cones are relatively trivial features of a landscape dominated by small basaltic andesite and basalt vents. The area between Mount Rainier and Mount Hood is intermediate in terms of the style of volcanism; large andesite and dacite composite cones (Rainier, Adams, St. Helens and Hood) occur together with fields and scattered vents of olivine basalt (Indian Heaven and the Simcoe Mountains).

Contrary to the classic subduction model of Japan (Hamilton, 1994), which is defined by very linear and abrupt arc boundaries, the Modern Cascades arc cannot be easily broken down into the classic fore-, sub- and back-arc tectonic areas. Despite the 60-80 km breadth of the Cascade arc, the locations of the boundaries separating the fore-arc, subarc and backarc are not well defined because of the lack of a clear volcanic front (Uyeda and Kanamori, 1979). Typically, forearc volcanoes are defined by their trenchward location in regards to the volcanic front itself. If the volcanic front is defined by the seaward limit of the Cascade stratovolcanoes, as (Hildreth, 2007) explains, the Cascadia forearc is therefore difficult to define because the volcanic front zig-zags back and forth down the entire arc. Also contrary to the classic Japan model, a discernible systematic cross-arc compositional pattern is lacking. The reason for the irregularities and indefinable boundaries may be the subduction of a warm slab (as opposed to a cold

slab such as in Japan), leading to an overall warmer thermal regime in the forearc of the Cascade subduction zone (Elkins-Tanton et al., 2001).

Petrology and Geochemistry of Cascade lavas

Hildreth (1997) noted that the magmatic vents of the Cascades are polygenetic and erupt a range of rock types. Calc-alkaline basalts (or arc basalts), basaltic andesites, andesites and dacites dominate the Cascade petrology. Alkalic basalts and High Alumina Olivine Tholeiites (HAOT) are also present across the arc's axis. The Cascades have been extensively studied in terms of both trace element geochemistry (Conrey et al., 2001; Hawkesworth et al., 1993; McDade et al., 2003) and isotope geochemistry (Eiler et al., 2000; Hart et al., 2003; Hawkesworth et al., 1997); (Bacon, 1990); (Schmidt et al., 2008). With numerous studies completed on all rock types from all parts of the Modern Cascade arc subduction, numerous end members have been documented. In order to compare and contrast volcanic rocks found in the area of interest for this thesis, the most commonly found end members are discussed below.

High Alumina Olivine Tholeiites

High Alumina Olivine Tholeiites (HAOT (Bacon et al., 1997)), or low K tholeiites (LKT) as (Conrey et al., 2001) and (Schmidt et al., 2008) refer to them, may be petrologically identified as being generally phenocryst-poor to aphyric with minor amounts of augite present in the groundmass. The surfaces of HAOTs commonly have a pahoehoe texture and are capable of flowing up to 45 km away from their source vents. These low viscosity basalts are commonly found as part of some large shield volcanoes, fissure fed eruptions, and as a part of large tube fed lava fields.

Bacon (1997) characterized HAOTs as being almost identical geochemically to MORBs with the exception of a higher Al contents, roughly $> 17\% \text{ Al}_2\text{O}_3$. HAOTs also have high CaO ($>11\%$) and incompatible element compositions very similar to MORBs. The HAOTs have a mild subduction signature with respect to their Ba/Nb and Ba/La ratios. In regards to compatible trace elements, HAOT exhibit higher Co and Sc, but similar Ni compared to typical arc basalts. Their trace element patterns (normalized to Primitive Mantle from (Sun and McDonough, 1989) show a MORB like pattern, with a small but noticeable Nb – Ta trough, a positive Eu anomaly, and flat middle to heavy REEs.

Calc-alkaline basalts

Calc-alkaline basalts (commonly referred to as CAB (Bacon et al., 1997)) are volumetrically more abundant than HAOTs (Schmidt et al., 2008). Petrographically, they have more abundant olivine and plagioclase phenocrysts and may also contain clinopyroxene as a phenocryst phase or of the groundmass. These lavas commonly erupt from monogenetic cinder cones or form larger shield volcanoes, where the flows are usually thick and resemble a'a flows (Rogers, 1985) (Richter et al., 2002). There have been instances where basaltic andesites and some andesites have had Mg, Cr and Ni levels similar to the CAB, which would imply that the basaltic andesites and andesite may have been in equilibrium with a mantle peridotite and therefore may be direct partial melts of the mantle (Hildreth, 2007). Some of the basaltic andesites (and more frequently the andesites) are believed to have evolved from basaltic parents within the crust through typical assimilation and fractional crystallization (AFC) processes, and

possibly batch mixing of melts (Hawkesworth et al., 1993; Hildreth, 2007). Basalts, basaltic andesites and some andesites are fairly phenocryst-poor due to their assumed unimpeded ascent to the surface from the lower crust, suggesting that most of the crystallization that took place occurred at deep reservoir depths.

Calc-alkaline basalts and basaltic andesites are abundant near the Quaternary arc axis and are assumed to have been abundant during the Miocene and Pliocene. Relative to HAOTs, CABs are identified by lower CaO and MgO contents and LILE enrichment relative to the HFSEs. Arc lavas exhibit La/Nb ratios in the 1.5-4 range, which is higher than in MORB or intraplate basalts (Conrey et al., 2001). The easiest way to geochemically identify these rocks is by looking at their incompatible element plot. When normalized to primitive mantle, the pattern of the arc basalts will show the aforementioned LILE enrichment, strong negative Nb-Ta trough, and LREE enrichment relative to the HREEs, with the Nb-Ta trough being the most typical characteristic of most arc basalts (Figure 9). CAB may also show positive spikes at numerous elements (i.e. Ba, Th, Pb, Sr, K, Eu) that give clues as to the amount of fluid or slab component added to the mantle from the subducting slab.

Intraplate Basalts (OIB-like End-Member)

Intraplate basalts occur infrequently in the Modern Cascade arc. They are sometimes referred to as within-plate basalts (WIP) (Conrey et al., 2001), and sometimes, erroneously, ocean-island basalts (OIB). These basalts are commonly erupted from scoria cones and may produce small lava fields (Bacon et al., 1997).

Figure 9: Incompatible element plot of the respective rock types found in the Modern Cascades.

Although rare, WIP-type basalts have been identified in the Cascade arc (Bacon, 1990), and are sometimes referred to as enriched HFSE basalts (Schmidt et al., 2008). These basalts have a low $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio, and when compared to Primitive Mantle values, show a very similar pattern to OIBs as they exhibit a LILE enrichment, lack a negative Nb-Ta anomaly, and exhibit an overall REE level decreasing from light to heavy. The most common occurrence of this type of basalt in the modern Cascade arc is found in the Garibaldi Volcanic Belt of British Columbia (Hildreth, 2007). They have also been documented in two other arc systems, the Banda-Sunda and the Marianas (Schmidt et al., 2008).

Dacites and Rhyolites

Few detailed studies on felsic lavas from the Cascade Range have been conducted. Basaltic lavas are numerous, sparsely crystalline and rarely demonstrate evidence of contamination, making them a more ideal means of studying mantle source characteristics (Borg et al., 2002). There are also a few studies on felsic lavas from individual volcanic centres or segments exist (Brophy and Dreher, 2000; Condie and Swenson, 1973) (Borg et al., 2002; Turner et al., 2000). According to Hildreth (2007), some dacites and rhyolites may be result of partial melting of young or older basaltic crustal rocks. This point is important as it implies that these more felsic end members may show just as “mantle-like” isotopic characteristics as basaltic rocks derived from slab and wedge derived melts that travel tens of km during their ascent. Borg (1997) described felsic lavas in the southernmost Modern Cascades as being derived from partially melting basalt and basaltic andesite rocks of the lower crust. LVC felsic rocks

also contain two end-members: lavas with a single ferromagnesian mineral present (clinopyroxene) and lavas with an assemblage of amphibole $\pm$ biotite $\pm$ orthopyroxene. In both end-members, the groundmass is aphyric, with 95% of the groundmass being comprised of glass. Plagioclase is the dominant phenocryst, with weak zonation, and pyroxene and biotite comprise less than 15% of the phenocrystic assemblage and are generally euhedral and fresh.

Isotopic values of Pb, Sr, Nd and O

$^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios measured for all of the basaltic end members in the Cascade arc are very similar to one another; however overall, the median of $^{87}\text{Sr}/^{86}\text{Sr}$ decreases south to north (Schmidt et al., 2008). Pb isotopic work completed by Bacon (1997) also showed that most samples had a higher $^{207}\text{Pb}/^{204}\text{Pb}$ than the Northern Hemisphere Regression Line, and they also plotted above the Pacific MORB when plotting $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{207}\text{Pb}/^{204}\text{Pb}$, which is indicative of a sediment contribution into the mantle wedge. Pb isotopic values overlap amongst all Cascades volcanic centres, yet values also suggest regional differences (Borg et al., 2002). At Crater Lake in Oregon, HAOT have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.70346–0.70364; basaltic andesite, 0.70349–0.70372, andesite, 0.70324–0.70383, and dacites of Mount Mazama have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.70348–0.70373 (Bacon 1994). Pb isotopes also vary ($^{206}\text{Pb}/^{204}\text{Pb} = 18.838$ –18.967, $^{207}\text{Pb}/^{204}\text{Pb} = 15.556$ –15.616, $^{208}\text{Pb}/^{204}\text{Pb} = 38.405$ –38.619, and generally covary with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Bacon (1994) reported that these radiogenic isotopic data from Crater Lake plot within the range of published data for other Cascade volcanoes.

Oxygen isotopic ratios in the Cascade arc for all rock types show a clear regional trend with the highest values found in the Lassen and Shasta areas and the median values generally decreasing to the north. These features are interpreted as reflecting assimilation of heterogeneous lower continental crust by mafic magmas, followed by mixing or mingling with silicic magmas. These may be formed by the partial melting of initially high $\delta^{18}\text{O}$ continental crust ($\sim +9.0$) with hybridized lower $\delta^{18}\text{O}$ ($\sim +6.0$) mantle-derived basaltic magmas toward the center of the system. Normal mantle values ($+5.5 \pm 0.4\%$, (Eiler et al., 2000)) are found in all volcanic centres with the exception of Shasta and Lassen (Borg et al., 2002).

Previous studies demonstrate that the ranges in the Sr, Nd and Pb isotopic compositions of intermediate and silicic volcanic rocks in the LVC overlap those of regional mafic volcanic rocks, though they consistently fall near the more crust-like extremes of the ranges (i.e. high $^{87}\text{Sr}/^{86}\text{Sr}$, very high $\delta^{18}\text{O}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$) (Borg and Clynne, 1998). These features have been interpreted to indicate that the silicic magmas are well-homogenized partial melts (20–25%) of lower continental crust that is compositionally similar to regional calc-alkaline mafic lavas erupted during the late Cenozoic in the southernmost Cascades ((Guffanti et al., 1990); (Borg and Clynne, 1998)).

Extension in the Back Arc

The tectonic regime (Figure 10) behind the volcanic front may be extensional or compressional, dependent on the age of the seafloor being subducted. Chilean-type subduction zones are characterized by strong coupling between the two plates. The upper

plate moves outward over the lower plate and the upper plate experiences compression. Mariana-type subduction zones typically have steep dipping slabs, generating a force on the above plate that induces extension (Davies, 1999).

The extensional Basin and Range province is expanding westward and impinging upon the active Cascade arc in northern California (Guffanti and Weaver, 1988), resulting in prominent normal faults recorded in the Lassen area and northwards (Borg et al., 2002). This may explain why more mantle derived basaltic material (HAOT, alkalic basalts) are found around the LVC and north. The HAOT lavas may actually originate from intra-arc vents that stretch out along Basin and Range type extensional faults (Guffanti et al., 1990).

(Hammond, 1989) explains that extensional volcanism may now be the dominating mechanism behind most of the volcanism in the Cascades from the LVC to southern Washington. Extensional volcanism in the High Cascades is likely a result of two processes, an increase in Basin and Range extension into the Cascade Arc, and a decrease in the convergence rate between the Juan de Fuca and North American plates as a result of the decreased size of the subducting plate.

Generalizations on the Geology of the Western Cascades

Volcanism related with the Western Cascades arc may have initiated as early as 45 Ma and may extend to ages as young as 5 Ma. This extensive age range may allow unfortunate addition of some pre- and post-Western Cascades samples (du Bray et al., 2006). In many parts of the Cascade Range, no obvious discontinuity between

Figure 10: Map of tectonics located in the Modern Cascades. The tectonics shown above are associated with Basin and Range faulting and are not associated with Cascade volcanism. Figure from (Observatory, 2009). Black lines represent known faults; red dashed lines represent progression of B&R faulting into the Cascades; red arrows show overall plate motion between N.American plate and the Pacific, Explorer and Juan de Fuca plates; orange curved lines show depth of Juan de Fuca plate beneath the N.American plate; triangles represent major Cascade volcanoes; circles represent crater complexes.

magmatism associated with the Western Cascades and High Cascades arcs has been identified.

Abundant fieldwork has shown that there is no visual unconformity along the length of the Cascade Range between the two sub provinces. Middle Miocene volcanic rocks of the voluminous Columbia River basalt are not related to Western Cascades arc magmatism (Dickinson, 1997) (du Bray et al., 2006).

McBirney, (1978) noticed that the proportion of basaltic rocks has increased steadily with time. The Oligocene-Early Miocene episode produced the largest volume of siliceous rocks, mainly rhyolite and dacite; andesite was the dominant rock type produced in the Cascades during the Mid-Miocene episode, and since that time, basalt has outweighed all other rock types combined.

There are no known reports that document any existence of LKTs in the Western Cascades. Alkaline rocks, which lack the typical subduction signature, are not found in the Western Cascades (Bacon, 1990; Madsen et al., 2006), but are found in the Modern Cascade rocks in the Mt. Lassen area (Strong and Wolff, 2003). This does not suggest that LKTs or WIPS did not exist during this time, as heavy soil and vegetation cover, extensive glaciation and burial by Modern Cascades volcanism may have obscured any existence of these unique basaltic end members.

Generalizations on the Geology of the Southern Ancestral Cascades

In western Nevada, Ancestral Cascades arc volcanism began as early as ~30 Ma, much earlier than previously recognized (Dickinson and Hatherton, 1967). In the Lake Tahoe area of northern California, ages vary from as old as 28 Ma to approximately 3 Ma

and arc magmatism generated numerous stratovolcanoes throughout the region (Cousens et al., 2008). Lavas are primarily andesitic, consisting of moderately to coarsely plagioclase porphyritic rocks with accessory hornblende, pyroxene and rare biotite while the more mafic (basalt to basaltic andesite) volcanism began at least as early as 15 Ma in broadly the same region as the intermediate activity (Henry and Perkins, 2001) (du Bray et al., 2006). Ancestral Cascades arc lavas are sub alkaline and follow a calc-alkaline fractionation trend, very similar to that of the Modern South Cascades. While Henry and Perkins) 2001) and Cousens et al., (2008) noted that the Southern Ancestral Cascades differ from the Modern Southern Cascades in that there are no high-alumina olivine tholeiites (HAOT) lavas, all Southern Ancestral Cascade lavas are light REE-enriched while their REE patterns are concave upwards. Finally, all Southern Ancestral Cascade lavas have $^{87}\text{Sr}/^{86}\text{Sr}$ greater than 0.7040 and $^{143}\text{Nd}/^{144}\text{Nd}$ less than 0.5128.

Summary of South Ancestral Cascades and Southern Modern Cascades

There are similarities and differences between lavas from the Modern South Cascades compared to lavas from the Southern Ancestral Cascades. Petrologically, lavas from both arcs are similar as they both exhibit varying phenocryst abundances and compositions, while also exhibiting numerous volcanic complexes comprised of debris flows and/or lahar flows (Cousens et al., 2008). Lavas from the Southern Modern Cascades are dominantly more mafic (Hildreth, 2007), while lavas from the Southern Ancestral Cascades in the Lake Tahoe-Reno area are more andesitic (Cousens et al., 2008; Dickinson, 1997). There are no known LKT or WIP occurrences within the Ancestral Cascades, compared to the numerous occurrences of LKTs and WIPs, not only

within the Modern South Cascades, but also throughout the entire Modern Cascades arc (Bacon et al., 1997; Hildreth, 2007; Schmidt et al., 2008). Overall, lavas from the Southern Modern Cascades have less radiogenic in both $^{87}\text{Sr}/^{86}\text{Sr}$ and more radiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ compared to lavas from the Southern Ancestral Cascades, although there is some overlap among the high $^{87}\text{Sr}/^{86}\text{Sr}$ lavas from the LVC and the low $^{87}\text{Sr}/^{86}\text{Sr}$ samples of the South Ancestral Cascades.

CHAPTER 4: FIELD WORK AND PETROLOGY

Field Methods

From August 5th – August 18th, 2008, samples were collected from the field area, referred to as the “North Sierra”, using two geological maps, the Chico Quadrangle and the Westwood Sheet, and three California State Forestry maps: Lassen, Plumas and Tahoe National Forests. The mode of transportation was a 2008 Jeep Patriot 4X4, while navigation to outcrops thought to be Mio-Pliocene in age based on information from the geological map was completed by using the forestry map for that specific area. Samples were collected from outcrops that represented a lava flow only; no samples were selected from the numerous volcanoclastic units north of the Lake Tahoe area due to the uncertainty in their origin. GPS locations were determined using a handheld unit. A map of the area with sample locations is presented in Figure 11.

Physical Observations

Closer to the LVC, the quality and paleogeographic location of each outcrop was a key factor in trying to roughly determine whether an outcrop was Quaternary, or Miocene-Pliocene in age. Quaternary flows are expected to be less weathered than the Miocene-Pliocene flows and should also fill in paleocanyons, while Miocene-Pliocene flows were generally located at higher elevations and are slightly more weathered.

Petrology

Photographs of representative rock samples are included from Figure 12 to Figure 14. Table 1 summarizes the basic geological information from the field for the 58 samples used in this study. The thin-section petrography of the volcanic rocks is summarized in Appendix I for all samples.

Overall, the lavas range in composition from basalts through to rhyolites and are all calc-alkaline. Lavas from the northern Sierra can be divided into three categories based on the percentage of phenocrysts visible in hand specimens: weakly porphyritic (<10%), moderately porphyritic (10-20%) and porphyritic (>20%). Volcanic lavas for the northern Sierra are commonly weakly porphyritic (~50% of all samples), with rocks from all groups, including all of the more evolved rocks (i.e. dacites and rhyolites), exhibiting limited phenocrysts. Both the moderately porphyritic and the porphyritic groupings include roughly the same percentage of rocks ($\sim 25 \pm 3\%$). Phenocrysts range in size from millimeters to no larger than two centimeters, and size is independent of the total percentage of phenocrysts in the rock (i.e., a weakly or moderately porphyritic rock was not limited to the smallest size in phenocryst and vice versa). The common texture amongst all the rock types is microporphyritic, whereas most of the weakly porphyritic samples exhibited a trachytic texture. A few samples (two basalts, one basaltic andesite and three andesites) were glomeroporphyritic, with clusters comprised of olivine $\pm$ plagioclase + pyroxene. The plagioclase phenocrysts commonly exhibit a strong sieve texture (top picture from Figure 15). Another disequilibrium texture found was that of “ghost” amphiboles, which exhibit an opaque rim that sometimes continues to the core (bottom picture from Figure 15). Cousens et al (2008) interpreted this unique feature as

Figure 11: Location and rock type for samples collected during the 2008 field season plotted on the Chico (southern map) and the Westwood sheet (northern map). Orange and dark red represent Miocene - Pliocene volcanic rocks, beige represent Quaternary volcanic rocks, yellow represents Quaternary sediments and pink represent Sierra Nevada granites.

Figure 12: TOP: Aphyric basalt from the N.Sierra. BOTTOM: Moderately porphyritic basaltic andesite from the N.Sierra. Phenocrysts are plagioclase and slightly weathered olivines.

**Figure 13: TOP: Basaltic andesite with a rare larger amphibole phenocryst (~1cm).
BOTTOM: Phenocryst poor andesite. Blocky pyroxenes are the dominant phenocryst.**

Figure 14: TOP: Porphyritic plagioclase + amphibole dacite. The only dacite collected during the 2008 season. BOTTOM: Sample 08-LT-32, a phenocryst poor (glassy) rhyolite.

Figure 15: TOP: Phenocryst of plagioclase exhibiting oscillatory zoning, and at the centre of the crystal, a sieve texture. BOTTOM: Amphibole ghost (see text) in a basaltic andesite. Diameter for both pictures is 2.5mm.

amphibole being broken down to an anhydrous assemblage of plagioclase, pyroxene and magnetite due to the degassing of the magma, which is common in calc-alkaline andesites.

All of the rock groups include lavas with $\leq 40\%$ phenocrysts, although most samples have $<25\%$. Plagioclase is the most abundant phenocryst phase in all rock types (as shown in the top picture of Figure 14). In mafic rocks (basalts and basaltic andesites), the abundance of plagioclase ranges from 2% to 15%, with no samples from any rock suite having higher plagioclase phenocryst abundances than 15%. Oscillatory zoning occurs in numerous plagioclase phenocrysts and is not restricted to a specific rock suite. The other phenocryst phases include clino- and orthopyroxene ($\leq 8\%$), olivine ($\leq 10\%$), magnetite ($\leq 2\%$), and amphibole ($\leq 2\%$), with most phenocrysts being subhedral. The groundmass contains the same phases as are present in phenocrysts, as well as glass.

The andesite samples contain phenocrysts of plagioclase (2%-15%), pyroxene (1-10%), olivine ($\leq 5\%$), hornblende (2-5%) and magnetite ($\leq 2\%$). With the exception of some anhedral pyroxene and olivine, all phenocrysts are subhedral. The groundmass consists primarily of plagioclase, with minor clinopyroxene, olivine and glass. Sample 08-LT-17 includes rounded xenocrysts of biotite (3%), quartz (3%) and potassium feldspar (2%), suggesting significant interaction with a granitic source during magma ascent.

Phenocryst abundances in the dacites and rhyolites were noticeably less in comparison to the other rock types. Plagioclase (1-5%), quartz ($\leq 5\%$), pyroxene ($\leq 5\%$),

amphibole ($\leq 5\%$), biotite ($\leq 2\%$), and potassium feldspar ($\leq 1\%$) were all present as phenocrysts. Plagioclase phenocrysts range from euhedral to subhedral, whereas other phenocryst phases range from subhedral to anhedral. Rounded grains of quartz in some of these samples are interpreted as xenocrysts, while more angular grains of quartz are believed to be phenocrysts. There was evidence of a minor secondary phase present in the rhyolitic sample 08-LT-32, where it appeared that potassium feldspars were being replaced by calcite.

Table 1: Rock descriptions and TAS name with sample number and location.

Sample Number	Hand Specimen Description	Easting	Northing
08-LT-01	Basalt. Light-medium gray aphanitic matrix. 2-5% olivine, slightly to medium weathered. Rock has a sheared appearance (flow banding).	670358	4510025
08-LT-02	Andesite. Platy medium gray matrix. 2-5% olivine, less than 1mm, fresh. 10% plagioclase; forms in clusters (snowflake plag). Possible Quaternary lava flow. Close to Whaleback Mtn.	678713	4504299
08-LT-03	Andesite. Platy med – gray matrix. 30% plag phenocrysts, less than 1mm and fresh. Rare phenocrysts of pyroxene and olivine. Non-vesicular flow foliation. Sampled close to Borg Sample LB91-142	680557	4498571
08-LT-04	Basalt. Med- dark grey matrix. 2-5% olivine, slightly weathered, less than 1mm. 1-2% plagioclase, small, fresh laths (less than 1mm). Possible correlation to Borg sample LB91-149 and 150.	687982	4506647
08-LT-05	Andesite. Black aphanitic groundmass. Trace of both olivine and plagioclase phenocrysts (Less than 0.5mm).	687982	4506647
08-LT-06	Andesite. 10-15% fresh olivine, rarely larger than 1mm. 20% plagioclase, lath shaped, fresh and less than 1mm. Medium gray matrix. Outcrop on top of Antelope Mtn, near fire station. Borg sample LB-90-80 is close by.	676925	4495401
08-LT-07	Andesite. Medium gray matrix. Minor pyroxenes, elongate, less than 1mm. 2-5% plagioclase, blocky, less than 1mm. 5-10% olivine, less than 1mm and fresh. Glomeroporphyritic clusters of olivine and plagioclase.	677940	4495615
08-LT-08	Basalt. Medium – light gray matix. 1-2 mm clasts of plagioclase and olivine are very common. 5% olivine phenocrysts, less than 1mm. Merrill Mountain location.	683280	4491107
08-LT-09	Basalt. Dark grey matrix. Olivines are small (~0.5mm), fresh to slightly weathered. 2% plagioclase, less than 1mm, fresh. Numerous small vesicles.	665674	4448873
08-LT-10	Basaltic Andesite. Medium gray aphanitic matrix. 1% olivine, fresh and ~1mm. Plagioclase was around 1%, fresh and euhedral. North-east of Butt Lake Reservoir.	651891	4448416
08-LT-11	Basaltic Andesite. Medium gray aphanitic matrix. 20-30% plagioclase, less than 1mm and dominantly fresh. 5-10% blocky pyroxenes, fresh, less than 1mm	642192	4449359

<i>Sample Number</i>	<i>Outcrop Description</i>	<i>Easting</i>	<i>Northing</i>
08-LT-12	Basalt. Aphyric, medium gray matrix. Microvesicular, microcrystalline, no noticeable phenocrysts. Possible Quaternary flow.	648704	4452385
08-LT-13	Andesite. Aphyric matrix. 1-2% olivine, less than 1mm and very fresh. Near the town of Westwood.	672229	4472233
08-LT-14	Dacite. Dense, medium gray matrix. Minor olivine, fresh and less than 1mm. Platy rock cleavage. Possibly same unit as 08-LT-13. Close to Borg sample 08-LT-14.	675578	4473329
08-LT-15	Andesite. Medium to dark grey matrix. Possible glass in matrix. Slight conchoidal fracture of rock. 5% of large (~1-2cm) of olivine and/or olivine plagioclase glomerocrysts. Outcrop at fire lookout on Pegleg Mtn.	673840	4474732
08-LT-16	Andesite. Similar to 08-LT-13. Aphyric platy volcanic rock. Chris Henry sampled here for geochronology.	674582	4473985
08-LT-17	Andesite. Finely crystalline, light grey matrix. 5-10% biotite, 1-3mm, euhedral and fresh. 5-10% amphibole, 2-4mm, euhedral and slightly weathered. 30% plagioclase, 3mm – 1cm, fresh, blocky and euhedral. Spooner Summit.	250039 (Zone 11)	4332512 (Zone 11)
08-LT-18	Basalt. Medium gray matrix with microvesicles. 2% olivine, less than 1mm and slightly weathered. 2-5% plagioclase, less than 0.5mm, fresh and euhedral. South of Susanville.	699025	4473772
08-LT-19	Basaltic Andesite. Sampled from a monolithic debris flow. Dark grey aphanitic groundmass with plagioclase microlites. 15-20% plagioclase, blocky and fresh. 2-5% pyroxenes, highly weathered, ~0.5mm.	680211	4467878
08-LT-20	Andesite. Light to medium gray matrix, microcrystalline, possible plagioclase and amphibole microlites. 5% amphibole phenocrysts, euhedral to subhedral, ~1mm. 25% plagioclase, blocky, euhedral and fresh. ~1mm, some unknown inclusions in the plagioclase. Minor sieve texture noticed in some plagioclase.	679556	4467851

<i>Sample Number</i>	<i>Outcrop Description</i>	<i>Easting</i>	<i>Northing</i>
08-LT-21	Basaltic Andesite. Aphanitic medium grey groundmass. 10-15% plagioclase, 1-2mm, blocky and fresh to slightly weathered. 2-5% pyroxene, euhedral and ~0.5mm. 5% olivine, fresh to moderately weathered, 0.5-3mm and commonly glomerophytic. Hamilton Mtn.	681782	4464055
08-LT-22	Basaltic Andesite. Finer grained, less plagioclase rich flow than in 08-LT-21. Matrix is dark grey and microcrystalline. Very fissile. 5% olivine, weathered, ~1mm. 1% plagioclase, blocky, fresh. South of Hamilton Mtn.	682861	4463205
08-LT-23	Basalt. Medium gray aphanitic matrix. 5-10% plagioclase, 0.5-1mm plagioclase, euhedral and fresh. 2-5% pyroxene, blocky, 0.5-3mm, euhedral and dominantly fresh. SW side of Coyote Mtn.	684569	4460062
08-LT-24	Basaltic Andesite. Light to medium gray, almost aphanitic with plagioclase microlites. 1-2% pyroxene, fresh to slightly weathered, subhedral, 0.5-2mm. 5-10% plagioclase, 0.5-1mm, fresh and blocky. On the road to Red Rock lookout.	695586	4460493
08-LT-25	Basalt. Dark grey matrix, microcrystalline. 2% pyroxene, slightly weathered, 0.5-1mm. 2-5% olivine, less than 0.5mm and slightly weathered. 10-15% plagioclase, 0.5-1.5mm, fresh to slightly weathered. South edge of Wildcat Ridge, near a gravel pit.	702642	4456582
08-LT-26	Basaltic Andesite. Black aphanitic to glassy matrix. Minor olivine ~1%, 0.5mm, blebby looking and slightly weathered. 1% plagioclase, less than 0.5mm, euhedral and fresh. Common granitic xenoliths, cm's in size. Angular to sub-rounded with reaction rims. Rare gabbroic xenoliths (cm's). This sample contained a granitic xenolith.	709265	4423775
08-LT-27	Basaltic Andesite. Same as 08-LT-26, but with a rare gabbroic xenolith.	709265	4423775
08-LT-28	Basalt. Light grey matrix, fine grained to microcrystalline. Variably porphyritic. 5% plagioclase, 0.5mm, subhedral and fresh. 5-10% olivine, 0.5-2mm, weathered. 5-10% pyroxenes, 0.5-3mm, weathered.	705704	4427968

Sample Number	Outcrop Description	Easting	Northing
08-LT-29	Basalt. Close to previous location, outcrop is more crystalline than previous sample. Holocrystalline, fine-grained matrix, medium gray with plagioclase microlites. 2-5% pyroxene, 0.5-2mm, euhedral and dominantly fresh. 10-15% olivine, subhedral, fresh to slightly weathered, 0.5-3mm.	704590	4427935
08-LT-30	Basaltic Andesite. Aphanitic light grey matrix. 10-15% pyroxene, 0.5-1.5mm, euhedral, fresh to slightly weathered. Trace amount of olivine, fresh, ~0.5mm. Top of Mt. Ingalls.	702650	4429570
08-LT-31	Basaltic Andesite. Medium gray matrix, very fine grained. 2-5% olivine, euhedral, slightly weathered, ~0.5mm. 5-10% pyroxene, 0.5mm, fresh and euhedral.	702650	4429570
08-LT-32	Rhyolite. Fresh, glassy matrix. Massive lava flow with no apparent layering. Lightly sheared. Phenocrysts are weathered and hard to ID.	697980	4419420
08-LT-33	Rhyolite. Very fine, grayish white matrix with microlites of plagioclase and biotite. 5% biotite, euhedral, fresh to slightly weathered, 0.5-1mm. Minor amounts of plagioclase, which were subhedral, fresh to slightly weathered, less than 0.5mm.	719653	4419563
08-LT-34	Andesite. Aphanitic medium gray matrix. 30% plagioclase, 0.5-3mm, euhedral, fresh, elongate and blocky.	718523	4419621
08-LT-35	Andesite. Light grey matrix, plenty of plagioclase microlites, rock was moderately to strongly weathered. 2-5% amphiboles, 0.5-2mm, euhedral. 15% plagioclase, 0.5-2mm, slightly weathered, blocky to elongate. Coyote Hill.	719679	4428007
08-LT-36	Basaltic Andesite. Light grey, aphanitic matrix. Minor pyroxene, moderately weathered, less than 0.5mm. Minor plagioclase, less than 0.5mm and fresh.	721826	4428876
08-LT-37	Basaltic Andesite. Light to medium gray, aphanitic matrix. 2-5% hornblende, euhedral, dull in colour and 1-3mm. 10-15% small pyroxenes, less than 0.5mm. 20-25% plagioclase, ~0.5mm.	728610	4428707
08-LT-38	Basaltic Andesite. Light grey, fine-grained matrix with plagioclase microlites. 10% amphibole, 0.5-4mm, euhedral, fresh to slightly weathered. 10-15% plagioclase, 0.5-1mm euhedral and fresh. Microvesicles, less than 0.5mm.	732155	4423978

<i>Sample Number</i>	<i>Outcrop Description</i>	<i>Easting</i>	<i>Northing</i>
08-LT-39	Basaltic Andesite. Aphanitic, medium gray matrix. Trace amounts of plagioclase, 1-2mm, euhedral. 5-10% olivine, subhedral, 0.5-1mm, fresh to slightly weathered. Near the Sagehen area.	737979	4370073
08-LT-40	Basalt. Medium to dark grey matrix. Very fine grained, some plagioclase microlites and no vesicles. 5% olivine, less than 0.5mm, fresh.	737847	4369603
08-LT-41	Basaltic Andesite. Dark to medium gray matrix, fine grained. 15-20% pyroxene, blocky, 0.5-3mm, euhedral and fresh. Trace hornblende, minor olivine and minor plagioclase present, all less than 0.5mm in size. Rare xenoliths found in the outcrop.	737222	4369352
08-LT-42	Andesite. Medium gray matrix. 5-10% plagioclase, 1-2mm, slightly weathered, blocky, appear to have reaction rims.	736320	4365558
08-LT-43	Basaltic Andesite. Medium gray matrix with a glassy look. Rock is jointed, platy and aphyric.	732792	4385582
08-LT-44	Basaltic Andesite. Aphanitic, medium gray matrix. 5% pyroxene, less than 1mm and fresh. 10% amphibole, euhedral, 0.5-2mm and fresh. 10-15% plagioclase, euhedral, fresh, 0.5-2mm, elongate laths.	731969	4384105
08-LT-45	Basaltic Andesite. Medium gray aphanitic matrix. 15% plagioclase, slightly weathered, 0.5-1.5mm, blocky to elongate. Some crystals may be xenocrysts. 2% pyroxene, square cross section, less than 0.5mm. 5-10% amphibole, lath shaped, euhedral, 0.5-3mm. Close to Frenchman Lake Dam.	740297	4419332
08-LT-46	Basaltic Andesite. Glassy, fresh matrix. Fractures in rock are altered. Rare plagioclase crystals, 0.5-1mm. Rock overall is aphyric, dark to medium grey and outcrop was roughly jointed.	740276	4420073
08-LT-47	Basaltic Andesite. Medium gray, trachytic matrix, with orientated plagioclase microlites. Minor pyroxene, 0.5-1mm, subhedral and slightly weathered. 10-15% olivine, euhedral to subhedral, 0.5-1.5mm, fresh to slightly weathered.	738477	4420845
08-LT-48	Andesite. Medium gray, aphanitic matrix. 2% pyroxene, fresh, euhedral, less than 1mm. 2% hornblende, fresh, euhedral, less than 1mm. 10% plagioclase, blocky and elongate, 1-3mm and fresh.	742741	4387426

Sample Number	Outcrop Description	Easting	Northing
08-LT-49	Andesite. Fine grained, medium gray matrix with microlites of plagioclase and unknown mafic mineral. 10% hornblende, subhedral, slightly weathered and 0.5-1mm. 15-20% plagioclase, 0.5-2mm, fresh to slightly weathered, euhedral and blocky.	743595	4385619
08-LT-50	Basaltic Andesite. Fine grained aphyric rock. Matrix is medium gray and looks to have a conchoidal fracture to it. Trace of plagioclase phenocrysts, less than 0.5mm, blocky and fresh.	741528	4376563
08-LT-51	Andesite. Light to medium gray, aphanitic matrix. 20% vesicles, elongate, 0.5-3mm, seem to be aligned. 5-10% plagioclase, 0.5-2mm, slightly weathered, subhedral, platy and blocky. Near Byers Pass Rd.	714632	4468317
08-LT-52	Andesite. ~ 150m up slope from 08-LT-51. Light to medium gray matrix, aphanitic. Mafic minerals are highly weathered making ID not possible, ~5-10%. ~15% plagioclase, fresh to slightly weathered, 0.5-3mm, euhedral to subhedral.	714856	4468532
08-LT-53	Basaltic Andesite. Medium gray matrix, very fine grained to aphanitic. Possible flow banding in plagioclase microlites. 2-5% olivine, euhedral, slightly weathered. Trace amounts of plagioclase phenocrysts, ~5%, less than 1mm.	699129	4450585
08-LT-54	Basaltic Andesite. Aphanitic to glassy matrix. Dark grey. Outcrop was very knobbly and jointed. ~20% plagioclase phenocrysts ~3-4mm, and subhedral.	697750	4450433
08-LT-55	Basalt. Black, aphanitic to glassy matrix. 1-2% olivine, slightly weathered, less than 0.5mm. Trace of plagioclase, almost microlite in size. Possible Lovejoy?	715897	4446313
08-LT-56	Basaltic Andesite. Dark grey aphanitic matrix. 1-2% olivine, slightly to moderately weathered, 0.5-2mm.	724328	4434840
08-LT-57	Basaltic Andesite. Medium gray aphanitic matrix. 10-15% hornblende, 1-6mm, elongate, dull black colour, possible plagioclase cores. Hard to determine if pyroxenes are present. Rare larger xenocrysts of plagioclase, ~6mm. ~20m thick flow.	729416	4439400
08-LT-58	Basaltic Andesite. Medium to dark grey, trachytic matrix. Aphyric.	732185	4439042

CHAPTER 5 – MAJOR, TRACE ELEMENT AND ISOTOPIC GEOCHEMISTRY

General Observations

Whole rock, trace element and isotopic data from the study area are presented in Appendix II, while the analytical methods are available in Appendix III. Geochemical results from the 58 rock samples are plotted on the conventional total alkalis silica (TAS) plot (Le Maitre et al., 2002). Of the 58 samples analyzed, 13 are classified basalt, 25 are basaltic andesite, 17 are andesite, one is a dacite and two are rhyolite (Figure 16). When the samples were plotted on a $K_2O\%$ vs. $SiO_2\%$ graph from (Le Maitre et al., 2002) (Figure 17) the majority of the samples plotted in the med-K field. Two basalts, one basaltic andesite, two andesites, one dacite and one rhyolite plot in the high-K field. There are also three basalt samples that plot in the low-K field. With the exception of four samples that are slightly alkaline, the samples are sub-alkaline. Magnesium (Mg) numbers ($Mg \# = 100 \times (Mg/Mg+Fe^{2+})$ where $Fe^{2+} = 0.9 \text{ Fe total}$) range from 23 to 66 with an average of 53 (N=58). While the basalts and the basaltic andesites have the highest Mg#'s, the andesites are more restrictive and show less scatter than the more primitive basaltic rocks.

Whole Rock and Trace Element Geochemistry

Variations in the major element chemistry, taken into account with the observed phenocryst phases, can help constrain the differentiation history of these volcanic rocks. The decrease in CaO, increase in SiO_2 , and negative Eu anomaly present in most samples (especially the most evolved: dacites and rhyolites) indicate fractionation of plagioclase, the major phenocrystic phase found in all rock types.

Figure 16: Diagram showing chemical classification of samples collected during 2008 field season (LeBas et al., 1986).

Figure 17: K₂O vs SiO₂ plot. Note the three samples grouped together in the low K field of the basalts. These samples are the low K tholeiites (LKT). Description in text. Felsics are comprised of the one dacite and two rhyolites from the N.Sierra.

Both Al_2O_3 and Sr vary widely amongst rock types (with the exception of the dacite and rhyolites). The presence of pyroxene in the basalts through andesites, and the observed decrease of $\text{CaO}/\text{Al}_2\text{O}_3$, CaO, and Sc with increasing SiO_2 are consistent with clinopyroxene fractionation. The decrease in MgO and Co, most likely reflect olivine fractionation in the more mafic rocks. The decrease in FeO^* , TiO_2 , and V (not shown) with decreasing SiO_2 and the presence of opaque phenocrysts in most rock samples illustrate the importance of titanomagnetite as a fractionating phase. The increase in P_2O_5 (Figure 18) reflects the lack of fractionation of apatite, until very high SiO_2 . The presence of xenocrysts and the large range in Pb, Sr, Nd concentration in the most of the rock types, regardless of silica content, suggest that crustal contamination may also be a process affecting the chemistry.

With the expected fractionation trends in the oxides discussed above and represented in Figure 18 excluded, trends in among individual rock types are hard to find when looking at the trace elements. Samples from the study area have a wide range of ratios of "fluid-mobile" (also often referred to as the large-ion-lithophile elements [LILE]) to "fluid-immobile" (high-field-strength elements [HFSE], and REE) elements. Ratios such as Ba/Nb [2.4-35], Ba/La [2.0-16], Ba/Zr [2.0-17.4]), and light (L)REE to HFSE La/Nb [1.2-11] and La/Zr [0.3-4.3], and finally ratios of immobile to mobile elements Nb/U [1.4-15.1] and Ce/Pb [1.3-26] also exhibit a wide range of values.

Figure 18: Harker diagram for N.Sierra lavas. Data for Modern South Cascades from (Borg et al., 1997) and data for South Ancestral Cascades is from (Cousens et al., 2008).

The incompatible-element diagrams (Figure 19), normalized to primitive mantle after Sun and McDonough (1988), are strongly enriched. Pronounced troughs occur at Nb and Ta in all samples. Peaks occur in most, but not all samples, at Pb and Sr. In general, fluid-mobile elements, such as Rb, Ba, U, K, and Sr, are enriched relative to less fluid-mobile elements, such as Nb, Ta, Zr, Hf and the REE. The basalts exhibit 3 distinct groups; one group is defined by overall low concentration of all of the incompatible elements, especially the LILE; another group is defined by the noticeable lack of the negative Nb – Ta trough; and the third group is comprised of samples that have large (but variable) Nb-Ta troughs. A few basalts are an exception to this with their ratios between LILE and the HFSE being low. Incompatible-element diagrams for the andesites and basaltic andesites are remarkably homogeneous, with the exception of three samples. 08-LT-44 and 45 show a HREE depletion while 08-LT-54 shows a higher overall abundance of all incompatible elements, while still reflecting the same pattern as seen in the other basaltic andesites.

Trace element ratio plots such as $\text{La}/\text{Sm}_{\text{pmn}}$ vs Ba/Nb (Figure 20) and Zr/Nb vs Ce/Pb (Figure 21) show a pronounced overlap among basalts, basaltic andesites and andesites. Although trends are evident in some ratio plots when reviewing the data as a whole, grouping among individual rock types is less common and Zr/Nb vs Ba/Nb and Ce/Pb vs Cs/Rb are plotted in Figures 22 and 23 respectively, these plots distinguish fluid characteristics from mantle characteristics. The common feature for these plots is that the basalts are found at both extremes of the plots, as well as between them. The basaltic andesites also exhibit a similar range in

Figure 19: Incompatible element pattern for the basalts, basaltic andesites and andesites. All values are normalized to Primitive Mantle ((Sun and McDonough, 1989)). In the basalt pattern, the green = CAB, black = LKT and red = WIP.

Figure 20: Ratio plot of La/Sm pmn vs Ba/Nb.

Figure 21: Ce/Pb vs. Zr/Nb. Symbols are the same as in Figure 20.

Figure 22: Zr/Nb vs. Ba/Nb. Symbols are the same as in Figure 20.

Figure 23: Ce/Pb vs. Cs/Rb. Symbols are the same as in Figure 20.

values, and may vary just as much as the basalts (Figures 20, 21 and 22). The andesites plot in a more restricted range than do the basalts and basaltic andesites. An obvious trend is illustrated when plotting Zr/Nb vs Ba/Nb (Figure 22), where an increase on the y-axis indicates increasing % of partial melting or a change in the mantle source, and an increase on the x-axis indicates an increase in the fluid component. Again, the basalts plot in both the high and low ranges for both the partial melting and the fluid component axes. Other rock types widely vary, and again plot in between the extreme ranges of the basalts. Overall the samples plot on a slight positive correlation, suggesting that the higher the degree of partial melting, the higher the degree of the fluid component in the source.

Isotope Geochemistry

The $^{87}\text{Sr}/^{86}\text{Sr}$ of the basalts range from 0.70387 to 0.70419. The $^{143}\text{Nd}/^{144}\text{Nd}$ of the basalts ranged from 0.51267 to 0.51307 (Figure 24). The basaltic andesites range in $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.70306 - 0.70480, while their $^{143}\text{Nd}/^{144}\text{Nd}$ range from 0.51255-0.51293. The andesites have an $^{87}\text{Sr}/^{86}\text{Sr}$ range of 0.7040-0.7052 and have a $^{143}\text{Nd}/^{144}\text{Nd}$ range of 0.51252-0.51281. Overall, an increase in $^{87}\text{Sr}/^{86}\text{Sr}$ does correlate with an increase in silica content, while a very slight negative correlation can be found when looking at the $^{143}\text{Nd}/^{144}\text{Nd}$ vs. SiO_2 plot. Overall, the Sr and Nd isotopic values overlap between all rock groups, however the andesites do show a more restrictive range compared

Figure 24: Nd, Pb and Sr isotopic ratios in the N.Sierra.

to the basalts and basaltic andesites. Note that two samples of basalt plot above the main array of mafic lavas. Sample 08-LT-23 has a higher $^{87}\text{Sr}/^{86}\text{Sr}$ value (0.70679) than any other lava that may be indicative of either crustal contamination, high sediment input into the mantle wedge, presence of a fluid enriched in $^{87}\text{Sr}/^{86}\text{Sr}$ or a more enriched mantle end member. Sample 08-LT-08 has the highest radiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ value (0.51307) that may be in relation to being derived from the least fluid-modified part of the mantle wedge.

Pb isotopic ratios also plot similarly to the Sr and Nd isotopic ratios. Individually, the rock groups are somewhat heterogeneous, yet when compared as a group, the lavas plot in a restrictive range. The $^{206}\text{Pb}/^{204}\text{Pb}$ of the rock groups from the study area all overlap. Basalts range from 18.88 to 19.06, while the basaltic andesites range from 18.79 to 19.07 and the andesites range from 18.88 to 19.03 (Figure 24). Also shown in Figure 24 are similar trends amongst the rock types of $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$. Plots of $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ show a strong positive correlation, with an observable overlap amongst all of the rock groups.

Unfortunately, only limited oxygen data were collected for rocks from the study area, which makes it difficult to discern any noticeable trends (Figure 25). In a somewhat similar trend to that of $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\delta^{18}\text{O}$, the $\delta^{18}\text{O}$ vs. SiO_2 plot (Figure 25) also expresses a vague positive correlation, while rock types show a range in values. The andesites showed the most restraint in their range, with $\delta^{18}\text{O}$ isotopic values from +6.7 to +7.6.

Figure 25: $\delta^{18}\text{O}$ values plotted against $^{87}\text{Sr}/^{86}\text{Sr}$ (TOP) and SiO_2 (BOTTOM). Both show a slight positive correlation. Symbols in the top figure represent all rock samples while symbols used in the bottom figure is the same as in Figure 23. Figure modified from (Cousens et al., 2008).

CHAPTER 6: DISCUSSION

Petrological Comparisons

The geographic location of the North Sierra study area makes it a useful region for the study of possible variations in magma processes from the Lassen Volcanic Centre to the Lake Tahoe-Reno area. As Atwater and Stock (1998) indicated, reconstruction of the Mendocino triple junction and the southern edge of the Juan de Fuca (JDF) plate suggest that a continental arc should have existed in the northern Sierra Nevada during the Miocene-Pliocene. Currently, the southern edge of the subducting JDF plate is situated north of Lake Tahoe at $\sim 39.5^{\circ}\text{N}$ (Atwater and Stock, 1998), which is roughly 300 km south of the southern end of the active modern Cascades at Lassen (Borg et al., 2002) (Clynne, 1990). Cousens et al. (2008) stated that a possible reason why the LVC lavas and the southern Ancestral lavas exhibit geological and petrological similarities may be the northward progression of the southern edge of the JDF slab.

The lavas from the North Sierra exhibit petrographic features similar to those of Quaternary Lassen lavas (Borg et al., 2002) (Clynne, 1990) and the Tahoe-Reno lavas (Cousens, 2008). Features in common include: sieve texture in phenocrystic plagioclase, oscillatory zoned plagioclase, non-equilibrated amphibole phenocrysts, plagioclase microlites and the presence of glass in the groundmass of several of the more weakly porphyritic rocks. These textures suggest that shallow magmatic processes are similar in the three regions, although they do not allow for greater insights into source characteristics.

The porphyritic lavas have large phenocrysts of plagioclase, pyroxene and in smaller amounts, amphibole. Zonation (Figure 15; top) may be found in many of the large plagioclase phenocrysts indicating possible recharge by magmas from below as the recharged magmas may have a different An composition. In the Ancestral Cascade lavas, the An contents in a single rock may vary by as much as 35 An units (Cousens et al. 2008). The amphiboles show evidence of being in disequilibrium as they have opaque reaction rims (Figure 15, bottom) which sometimes extend further into the core, suggesting that either the melt degassed beneath a volcano, or the magma ascended slowly across the amphibole stability field limit (Cousens et al., 2008).

Hildreth (2007) established that the Southern Modern Cascades produce higher amounts of felsic material, volumetrically, than the rest of the Modern Cascades. He also noted that the amount of felsic rocks produced is small in proportion to the more primitive rocks types. Cousens et al., (2008) noted that the volcanic rocks around in the Lake Tahoe-Reno area contain volumetrically less felsic rocks than the Southern Modern Cascades.

While the petrography may be insightful regarding the crustal residency time of the magmas, it cannot be fully relied upon for a means of confidently determining whether a sample is “Ancestral” or “Quaternary”. There is no petrographic transition between the LVC lavas to lavas from the Lake Tahoe – Reno area.

Geochemical Comparisons

As noted in Chapter 5, lavas from the Northern Sierra are all calc-alkaline, yet are slightly less alkalic than those from the Reno-Tahoe area, but slightly more alkalic than

samples from the Lassen Volcanic Centre. Due to the alkalis being more incompatible and enter the early melts, successive melting will dilute the melt and resulting in a more silica-saturated melts. This suggests that lavas from the N.Sierra may have melted at deeper depths and/or may be lower degrees partial melts of the mantle than compared to the LVC lavas, while compared to South Ancestral Cascade lavas N.Sierra lavas may be generated at shallower depths in the mantle and/or may be higher degree partial melts of the mantle.

The following breakdown of rock type is based on SiO_2 the trace element data, normalized to Primitive Mantle from (Sun and McDonough, 1989), which were presented in Figure 19.

Basalts

As reported in previous studies (Bacon et al., 1997; Leeman, 1983; Borg et al., 1997; Hildreth, 2007), there are different basaltic rock types that occur along the entire volcanic axis of the Modern Cascades Arc. These different rock types are believed to reflect melting of different parts of a heterogeneous mantle (Bacon et al., 1997). It is proposed that a LKT (N-MORB like) may represent anatexis of a depleted sub-arc mantle, an intraplate or within-plate basalt (OIB –like, or E-MORB) represents melting of an enriched mantle (relative to Primordial Mantle of Sun and McDonough (1989), and the dominant calc-alkaline basalts and primitive basaltic andesites represents melting of mantle with a classic subduction component added to it.

In the study area, examples of all three groups of basalts are found, distinguished from the trace element plots: a within-plate group, an LKT-like group and a calc-alkaline group.

LKT Group

Samples 08-LT-09, 08-LT-12 and 08-LT-18, are three samples that may be derived from a depleted mantle source. Referred to as LKTs (or HAOTs) throughout the Cascades Arc, these samples from the N. Sierra show the characteristic LKT trend which includes low LILE abundances, a small Nb-Ta trough and a flat REE pattern (Figure 26), as well as a low K₂O% (<0.3). These three samples also have a low La/Sm_{pmn} (<1.3), and an Nb/Zr ratio of less than 0.45. These rocks were collected well south of the Lassen Volcanic field (refer to Figure 11), and have a restricted spatial distribution between latitudes 40.17 and 40.4 N and longitudes 121.3 and 120.65 W. Petrologically the samples from the study area also fit the criteria for LKT as they are only weakly porphyritic to aphyric.

N.Sierra LKT Genesis

It has been well documented by many authors (Hildreth, 2007; Leeman, 1983; Bacon, 1990; Schmidt et al., 2008) that LKTs are produced by 6-10% decompression partial melting of a nearly anhydrous sub-arc mantle which had previously been slightly modified by a subduction signature. Elkins-Tanton et al., (2001) showed that LKTs from the LVC area were derived from ~40 km depth. LKTs from the N.Sierra closely resemble LKTs from the LVC and Mt. Shasta in terms of their incompatible elements (Figure 26) while the N.Sierra LKTs have slightly higher

Figure 26: Comparative Incompatible Element patterns. Data for Lassen rocks is from (Borg et al., 1997) and (Clynne, 1990), data for Shasta and Crater Lake rocks is from (Bacon et al., 1997), and data for Ancestral Cascades is from (Cousens et al., 2008). All values are normalized to Primitive Mantle ((Sun and McDonough, 1989))

$^{87}\text{Sr}/^{86}\text{Sr}$ values. With the LKTs thought to be derived from a depleted sub arc mantle, the increase in $^{87}\text{Sr}/^{86}\text{Sr}$ just south of the LVC is a strong case for a slightly more enriched overall mantle south of the LVC. Over time, the $^{87}\text{Sr}/^{86}\text{Sr}$ in the mantle will increase due to the radioactive decay of ^{87}Rb , which may suggest that a reason behind the N.Sierra LKTs exhibiting a slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$ compared to the Southern Modern Cascades is that the source for the N.Sierra LKT may be slightly older.

LKT Geographical Comparisons

Like the intraplate basalts, LKTs are not found in the Reno-Tahoe area (Cousens et al., 2008) but are common in the Southern Modern Cascades (Bacon et al., 1997; Borg et al., 1997). In the N. Sierra, two LKTs (samples 08-LT-09 and 08-LT-12) were found in an area well south of the Lassen Volcanic Centre around Lake Almanor, California, where no Quaternary rocks have been previously documented. It is considered that these unique rock types found in the Ancestral Cascades are LKTs of Miocene-Pliocene age, and are the only exposure of this basalt type to be found this far south. The other LKT, 08-LT-18, is located close to the town of Susanville CA, where rocks of the LVC older than Quaternary in age have been identified. The trace element patterns and isotopic ratios for all three LKTs are nearly identical and may have all been derived from the same mantle source. These three rocks have the lowest SiO_2 content of all the basalts sampled from the N.Sierra (47.81, 47.89 and 48.13 respectively), which are very similar to the SiO_2 contents of the LKTs from the LVC and Mt.Shasta (Bacon et al., 1997) (Reiners et al., 2000).

The LKTs from the N.Sierra are on the Basin and Range side of the Cascades, but still ~500 km from the Basin and Range itself, and the LKT's from the N.Sierra are not in any known back-arc setting. If the LKTs located in the N.Sierra are in fact Miocene-Pliocene in age, and if these LKTs were erupted in a similar mechanism as elsewhere in the Modern Cascades (i.e. close to Basin and Range faults), then these LKTs may be considered as the first evidence of Basin and Range related volcanism in the N. Sierra and Reno-Tahoe areas.

Within-plate Group (WIP)

Although uncommon in both the Ancestral Cascade Arc and the Modern Cascades arc, intraplate basalts were found in the northern part of the study area, within 50 km of the modern lavas from the LVC (Figure 11) where Bullen and Clynne (1990), and Bacon et al., (1997) also described rocks with an intraplate component. Hildreth (2007) described the same type of rock but referred to it as a within-plate basalt (WIP), which is what the rocks from this study, which exhibited a similar intraplate component, will be referred to throughout the rest of this study. Samples 08-LT-01, 08-LT-04 and 08-LT-08 follow the WIP pattern described in Chapter 3 very closely, with the only exception being that they exhibit an Nb-Ta anomaly which is quite small compared to the other basalts of this study area. Although these are not true alkali basalts, as are found in the Simcoe Mountains volcanic field (Bacon et al., 1997), there is little doubt that these three samples do exhibit a largely intraplate signature, exemplified by a steep REE pattern.

North Sierra WIP Basalts Genesis

Much like the proposal that Strong and Wolff (1993) and Hildreth (2007) suggest, there may be thick sheets or discrete mantle domains that are only slightly affected by slab derived fluids. The REE abundances of the WIP suggest that beneath the N.Sierra there are irregular lenses of enriched mantle that undergo low degree partial melting that produce a steeper REE pattern. The presence of a small but noticeable Nb-Ta trough suggests that a small flux off the slab may be initiating melting of the enriched mantle lenses.

WIP Geographical Comparison

Basalts with an intraplate signature are not found in the Reno-Tahoe area (Cousens et al., 2008), but as previously mentioned they are found in the LVC and sporadically in other areas of the Modern Cascades. WIP basalts from the LVC, Mt. Shasta and the N.Sierra have almost identical incompatible element patterns (Figure 26). However, due to the close (~10 km) proximity of the WIP collected for this study to Quaternary rocks studied by others in the area (eg. (Clynne, 1990) (Borg et al., 1997; Guffanti et al., 1990)), and without any known ages for the N.Sierra WIP, it is possible that the WIP from the N.Sierra are Quaternary in age.

It is clear that age dating of the WIP lavas from the N.Sierra would substantially help in determining which geographical domain these lavas should be grouped with.

Lovejoy Basalt

(Wagner et al., 2000) identified basalt flows in the Northern Sierran Nevada that supposedly are parts of the 15 Ma Lovejoy basalt, which may be associated with the

Miocene Columbia River flood basalts. The Lovejoy flood basalt is characterized by their higher REE abundances than any other basalts (CAB, WIP, LKT) found in the area.

Sample 08-LT-55 shows this REE trend and also exhibited petrographic features, unique to the Lovejoy basalt, such as an aphyric to glassy groundmass, and, on the macro scale its small and weak columnar joints. The Lovejoy basalt sampled as part of this study was collected within roughly 7 km of the type locale for this type of basalt (Garrison et al., 2003).

Lovejoy Geographical Comparison

According to (Garrison et al., 2008) the Lovejoy Basalt consists of up to 13 exposed flows of aphyric olivine basalt. It erupted from the Thompson Peak area near Honey Lake, CA, and flowed down a paleovalley which cut into Sierran basement that trended south 30 km to Red Clover Creek before bending southwestward and flowing 65 km to the Sacramento Valley. Dating work by (Garrison et al., 2008) put its age in the middle Miocene at 15.4 Ma.

There are no samples of the Lovejoy basalt documented around the Reno-Lake Tahoe area (Cousens et al., 2008) or in the LVC.

Source of CAB rocks

The source of the CAB, basaltic andesites and andesites are thought to originate from a depleted mantle peridotite as the ratios of primitive mantle normalized Nb/Zr are all less than 1. In contrast the WIPs have a ratio of 1.2-1.8, which suggests an enriched source while the LKTs have a ratio of 0.2-0.5.

Not all of the basaltic andesites from the N.Sierra can be produced by crystal fractionation of a CAB from this same area, due to the much greater range in their Sr and Nd isotopic values compared to the CAB, thus ruling out a direct parental relationship (Figure 27).

Primitive Basaltic andesites ($Mg\# > 60$) from this region, and from other parts of the Cascades arc (Bacon et al., 1997; Borg et al., 1997) can be derived in a similar way as CABs are formed, by partial melting of a sub-arc mantle source. The geochemistry of the CAB and the basaltic andesites show similarities as both rock types show similar isotopic ranges for Pb, Sr, Nd and O, while they also show similar incompatible elements patterns. With the range of isotopic compositions observed in the CAB, basaltic andesites and andesites, it is clear that these rock types tap a mantle source with a variable isotopic composition.

CABs are the most abundant rock type from this study area and one of the more abundant rock types in the entire modern Cascade Range. The CABs of the N.Sierra show significant similarities in both the petrographic and geochemical signatures to CABs described by Bacon (1990) from the LVC, Mt. Shasta, Mt. Adams, Medicine Lake and Crater Lake. CABs from the study area have much higher percentages of olivine and plagioclase phenocrysts compared to the WIP or LKTs, as well as possibly having a clinopyroxene crystallization stage, but the percentages of clinopyroxene are less than olivine or plagioclase. Geochemically, rocks from the study area perfectly fit the calc alkaline basalt trends presented by Bacon (1990) and described and noted elsewhere in the modern Cascades. They have LILE enrichment, a significant negative Nb-Ta anomaly, and decreasing primitive-mantle normalized REE abundances from light to

Figure 27: Nd and Sr isotopic ratio plot of lavas, with the basalts split into the four end-members (WIP, LKT, Lovejoy and CAB). Fields for MORB and lithosphere modified from (Yogodzinski et al., 1996).

heavy.

N.Sierra CAB Genesis

Although few of the samples collected turned out to be CAB, the CAB in the N.Sierra clearly follow the typical subduction signature that most CAB of the Modern Cascades show enrichment in LILE and LREE and a noticeable depletion in the HFSE. Variance in the incompatible element pattern for the CAB may be the result of: 1) the degree of partial melting, 2) the varying degree of depletion (or) enrichment of the mantle wedge from which the above partial melts are generated from and 3) the geochemical variability of the fluids released from the slab.

Figures 20 – 23 illustrate that the samples from the N.Sierra show that fluid from the subducting slab is affecting their geochemistry, as noted by high values of Ba/Nb (100-300), low Ce/Pb (2-4) and high Cs/Rb (0.05-0.09).

The enrichment or peaks of the mobile elements Rb, U, K, Pb, and Sr compared to the depletion or troughs of the immobile elements (Th and HFSE) could reflect melts of mantle sources that are metasomatized by hydrous fluids from the subducting slab as hydrous fluids released off the slab would be enriched in the mobile elements relative to the immobile elements.

Basaltic Andesites

Basaltic andesites (52-57% SiO₂) are more abundant than basalts along the Cascades Arc (Hildreth, 2007), which is also true for the N.Sierra. Comparing the trace element plots for the basaltic andesites and andesites from the N. Sierra, the basaltic andesites show a trace element pattern that is very similar to the pattern shown by the

CAB. Most of the basaltic andesites have the same abundances for most trace elements and REE as the CAB. There are a few exceptions where the overall abundances of LIL and HFSE are higher than the others. Samples 08-LT-44 and 45, which are close together geographically, exhibit the same pattern as the majority of the other basaltic andesites except for their HREE (Ho, Tm, Yb and Lu) patterns that show an appreciable depletion. This depletion suggests that the source for these rocks came from the garnet lherzolite field (Borg, 1997), as garnet is known to retain the HREE in its chemical structure. The presence of garnet suggests that the depth of melting for these two samples was greater than any other samples as the stability field for garnet is $\sim 1400^{\circ}\text{C}$ and > 85 km depth (Robinson and Wood, 1998). There are no other samples from the N.Sierra, LVC or the Lake Tahoe area to have this noticeable HREE depletion that suggests that the other samples were generated at depths within the spinel lherzolite field.

N. Sierra Basaltic Andesite and Andesite Genesis

The basaltic andesites cannot be related by crystal fractionation to either the LKTs or the WIP due to the greater Sr isotopic variance of the basaltic andesite and the difference in incompatible element patterns among the three groups. Crystal fractionation of a melt will not alter the isotopic signature of the residual liquid, yet it can increase the overall incompatible element abundances of the liquid as the silica content increases (Figure 28). The basaltic andesites are thought to be produced by the partial melting of a hydrous mantle peridotite (Baker, 1988) (Hildreth, 2007). Similarities in geochemistry between the CAB, basaltic andesites, and andesites indicate that basaltic andesites are derived from the same mantle wedge material as the CAB. Compared to the CAB the

basaltic andesites are higher degree partial melts as they exhibit lower abundances of the MREE and HREE, which behave incompatibly during melting but are diluted as the percentage of melting increases.

There is some evidence that the basaltic andesites and andesites show effects of crustal contamination, such as zoned plagioclase, minor crustal xenoliths and $\delta^{18}\text{O}$ values of around (+6.5), all of which suggest at least some interaction with crustal rocks. Lower Sr and Nd isotopic ratios compared to Sierra Nevada granitoids suggest that the crustal interactions are not significant enough to noticeably shift the geochemistry of these rocks away from that of their mantle sources.

Andesites may be produced by crystal fractionation from a basaltic andesites melt. If this were the case for andesites of the N.Sierra, then the REE patterns (Figure 29) would exhibit a negative Eu anomaly due to plagioclase fractionation, which they do not.

Alternatively, the andesites may be formed due to magma mixing. On a Nd and Sr isotopic ratio plot (Figure 27), the Sr isotopic values of the andesites from the N.Sierra exhibit overall, more radiogenic Sr values than those of the basaltic andesites. This leads to the suggestion that the andesites may have formed by mixing a more primitive melt that has a more MORB-like isotopic signature (higher $^{143}\text{Nd}/^{144}\text{Nd}$, lower $^{87}\text{Sr}/^{86}\text{Sr}$) with a source that has a higher Sr and Nd isotopic value (eg. Lithospheric mantle ((Cousens et al., 2008))). A mixing line between these two end members (MORB and lithospheric mantle) would pass through most of the samples. This process would produce a trace element figure that exhibits no negative Eu anomaly and would account for the range in Nd, Pb, and Sr isotopic values.

Figure 28: REE plot of basalts and basaltic andesites, normalized to Primitive Mantle values (Sun and McDonough, 1989).

Whatever the process may be that is responsible for the formation of andesites in the N.Sierra, it must be the dominant process because geochemically, the andesites from the N. Sierra are very similar to one another when looking at their trace element plots.

Andesite Geographic Comparison

Although common in the N. Sierra, andesites are less abundant than the basalts and the basaltic andesites. Petrologically, the andesites fit into two categories, a porphyritic and non-porphyritic suite, similar to what (Cousens et al., 2008) reported for lavas in the Lake Tahoe-Reno area. There are fewer andesites in the N.Sierra compared to the Lake Tahoe-Reno area, which may be related to the fact that there are fewer major volcanic edifices in the N.Sierra, however glaciation and mechanical erosion makes this proposal hard to confirm.

N.Sierra Dacite and Rhyolite Genesis

Dacites and rhyolites may be formed by the crystal fractionation of an andesitic melt, which is the dominant process for dacite and rhyolite production in the Modern Cascades (Hildreth, 2007). Another possible way to create a dacite or rhyolite is to partially melt a basaltic rock. Looking at the dacites and rhyolites REE plots and incompatible element patterns are a good way to distinguish between the two above processes. Negative Eu, P, Sr and Ba anomalies (Figure 29) would represent substantial crystal fractionation, as plagioclase fractionation will lead to the Eu and to some of the Sr anomaly, while fractionation of apatite would produce all four anomalies. Partial melting of a basaltic parent would produce a higher, yet similar incompatible element

Figure 29: Incompatible Element pattern of the dacite (08-LT-14) and two rhyolites. All values are normalized to Primitive Mantle ((Sun and McDonough, 1989).

pattern and would also produce a melt with the same isotopic value as the parental rock. Evidence which supports that the only dacite found in the N.Sierra is likely a partial melt of a basaltic parent rather than a crystal fractionated andesite is the lack of negative Eu anomaly (Figure 29) and more importantly, the similar incompatible element pattern the dacite has compared to the CABs.

As is the case with the lone dacite from the area, the rhyolites were also phenocryst poor, which according to Hildreth (2007) favours melt extraction from pluton sized crystal mush reservoir. Sample 08-LT-32 is a rhyolite, which unlike the dacite has undergone substantial crystal fractionation as shown by the large negative Eu, P, Sr and Ba anomalies in its incompatible element pattern. The second rhyolite, 08-LT-33, shows hints of being both a partial melt from a basaltic parent and undergoing at least moderate fractionation of apatite and plagioclase prior to eruption.

By studying the three most felsic rocks found in the N.Sierra study area, it was discovered that there are two ways in which these rocks are being formed: 1) by means of crystal fractionation of a primitive melt, and/or 2) by means of partially melting a basaltic rock.

Evidence of Crustal Contamination?

Although contamination or mixing of mantle derived melts with continental crustal material can generate incompatible-element characteristics similar to those observed in some of the samples, crustal assimilation alone (during ascent) cannot explain the incompatible-element chemistry of the evolved volcanic rocks. For example, the enriched basaltic samples extend to higher concentrations of many elements (e.g., Ba,

Figure 30: REE patterns of N. Sierra andesites and felsic rocks. Values normalized to Primitive Mantle ((Sun and McDonough, 1989)).

Sr, Pb, Th, U, LREEs) than generally found in MORB or the continental crust (Hofmann, 1988).

$\delta^{18}\text{O}$ (and in conjunction with SiO_2 and $^{87}\text{Sr}/^{86}\text{Sr}$) is a useful geochemical tool in helping to identify continental crust as a possible contaminant of ascending magmas. The asthenospheric mantle is believed to have $\delta^{18}\text{O}$ values of $\sim +5.5 \pm 1$ (Eiler et al., 2000), while the continental crust is thought to have $\delta^{18}\text{O}$ values around $+9 \pm 1$ (Eiler et al., 2000). Crustal contamination should increase the $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$, and this effect will increase with increasing SiO_2 in a contaminated liquid.

Borg (1997) used $\delta^{18}\text{O}$ as an indicator of possible crustal contamination for rocks from the LVC. Borg (1997) reported that the more felsic rocks from the LVC exhibited $\delta^{18}\text{O}$ values of around $+8 - +10$, which lead the author to believe that the dacites and rhyolites were produced by partial melting of the lower continental crust.

Figure 25 (Chapter 5) shows the $\delta^{18}\text{O}$ values for the three study areas. $\delta^{18}\text{O}$ values for basalts and basaltic andesites from the LVC, Tahoe-Reno area and the N. Sierra all plot in a restricted range from $+6 - +7.5$. The number of basaltic andesites and andesites sampled for $\delta^{18}\text{O}$ values from the LVC and the Tahoe Reno area is limited, making correlations between these areas and the N.Sierra difficult. The limited intermediate rock data available from the LVC show that overall these rocks maintain a low $\delta^{18}\text{O}$ value. While the N.Sierra andesites exhibit slightly higher $\delta^{18}\text{O}$ values than LVC, the values still suggest that the andesites from both locations have very little contact time with the continental crust. $\delta^{18}\text{O}$ values increase only slightly from the basaltic andesites to

andesites, which would suggest that this slight increase might be attributed to fractional crystallization of the basaltic andesites (Eiler et al., 2000).

The dacite and rhyolites from the N.Sierra exhibit $\delta^{18}\text{O}$ values, which are very similar to the $\delta^{18}\text{O}$ values of N.Sierran andesites, show that crustal contamination does not play a significant role in magma generation and/or magma modification for N. Sierra lavas. An increase in $\delta^{18}\text{O}$ values does not correlate well with either SiO_2 or $^{87}\text{Sr}/^{86}\text{Sr}$ (Figure 25) and N.Sierra dacites and rhyolites have values that are considerably less than the contaminated dacites and rhyolites from the LVC.

Sample 08-LT-33 is a rhyolite from the N.Sierra that was analyzed for $\delta^{18}\text{O}$ values. This rhyolite is very similar to the rhyolites of Borg (1997) in that they are light in colour and exhibit a strong porphyritic (~25% phenocrysts) texture. The $\delta^{18}\text{O}$ value for this rhyolite was +7.4, which is noticeably less than the +8 - +9.5 values of the crustally derived rhyolites that Borg (1997) sampled. This suggests that even the most evolved rocks from the N.Sierra are not a product of partial melting of continental crust, as is the case around the LVC (Borg 1997). Combined with the fact that there are very few dacitic to rhyolitic lava flows in the N.Sierra (either previously mapped or located in this study), also suggests that significant melting of the continental crust has not occurred in this region.

A possible explanation for the slight increase in the $\delta^{18}\text{O}$ values while exhibiting no correlation with SiO_2 and $^{87}\text{Sr}/^{86}\text{Sr}$, may be that the fluid coming off the dehydrating slab may have variably high $\delta^{18}\text{O}$ (Eiler et al., 2000), or that the mantle wedge beneath

the N.Sierra is variable in $\delta^{18}\text{O}$. Either of the above scenarios would account for the variable $\delta^{18}\text{O}$ values noticed in all rocks from the N.Sierra.

Crustal Contamination of Basalts?

As mentioned previously, CAB and WIP basalts both share very similar $\delta^{18}\text{O}$ values ($\sim +6.5$) and they all also fail to exhibit further evidence of significant crustal contamination such as an increase in radiogenic Sr with increasing SiO_2 and/or granitic xenocrysts. The LKT from this area exhibited an $\delta^{18}\text{O}$ value of +5.2, which suggests either that the mantle from which the LKT is derived is a less modified mantle or that the LKT underwent less interaction with continental crust during ascent than did the CABs and WIPs. However, because of the lack of a positive correlation between SiO_2 and Sr isotopic ratios or $\delta^{18}\text{O}$ noted earlier, the latter option for the lower $\delta^{18}\text{O}$ value is less likely. It has been shown that low temperature alteration of oceanic crust may raise its $^{18}\text{O}/^{16}\text{O}$ value (Eiler et al., 1998). Thus when the altered slab dehydrates beneath the mantle wedge, the fluid coming off the slab should exhibit a higher $\delta^{18}\text{O}$ value than if the slab were not altered. CABs from the N.Sierra are known to exhibit geochemical characteristics that are known to come from the fluid flux of a subducting slab (increase in LILE, decrease in HFSE) and the CABs $\delta^{18}\text{O}$ value is on average +6.5. If the fluid flux from the dehydrating slab is enriched in $^{18}\text{O}/^{16}\text{O}$, then the LKTs have very little to no fluid flux component in them.

Figure 31: Comparison plot of Sr, Nd and Pb isotopic values. Outside of the more radiogenic values of the South Ancestral Cascades in the Sr and Nd plot, there is noticeable overlap among the three regions. South Ancestral Cascades data from (Cousens et al., 2008); Modern South Cascades data from Borg et al., 1997 and 1998.

Sediment Component

Sediments add radiogenic Sr and Pb to the overlying mantle wedge, but in different concentrations that ultimately affect the isotopic ratio of these elements (Elliott et al., 1997). The $^{87}\text{Sr}/^{86}\text{Sr}$ from the N.Sierra is less radiogenic than those of the Lake Tahoe-Reno area and similar to that of the Modern South Cascades (Figure 31). Lower $^{87}\text{Sr}/^{86}\text{Sr}$ compared to lavas from the Lake Tahoe-Reno area may represent a younger mantle due to lower amounts of Rb/Sr, or it may represent a melt less modified by a lithospheric mantle component (Cousens et al., 2008). Elevated Ba/La and Ba/Zr compared to Cascadia sediments compounded with a $^{207}\text{Pb}/^{204}\text{Pb}$ array that also extends to values higher than Cascadia sediments (Cousens et al., 2008, and sources within), suggests that sediments added to the mantle wedge (if any) are not the source of the high $^{87}\text{Sr}/^{86}\text{Sr}$ and high $^{207}\text{Pb}/^{204}\text{Pb}$ component in N.Sierra lavas. Subducted sediments are also expected to be a component of the flux added to the mantle from the subducting slab, and this holds true for the N.Sierra, however the geochemical signature of the sediments are being masked by the more radiogenic Pb, high Ba/La and high Ba/Zr component.

Source of High $^{87}\text{Sr}/^{86}\text{Sr}$ and Low $^{144}\text{Nd}/^{143}\text{Nd}$

If sediments and crustal contaminants are not responsible for the higher Sr and Pb and lower Nd isotopic values found in samples from the N.Sierra, then there must be some source within the mantle wedge that is responsible for the shift in isotopic values.

Cousens et al., (2008) noted that the high $^{87}\text{Sr}/^{86}\text{Sr}$, low $^{143}\text{Nd}/^{144}\text{Nd}$ and high $^{207}\text{Pb}/^{204}\text{Pb}$ compared to Southern Modern Cascade lavas is consistent with an enriched ancient lithospheric component beneath the Lake Tahoe-Reno area. With samples from

the N.Sierra exhibiting values of $^{87}\text{Sr}/^{86}\text{Sr}$, which are slightly more radiogenic compared to Southern Modern Cascades but less radiogenic than Southern Ancestral Cascades, it could be plausible that the lithospheric component, which may be responsible for the higher Sr isotopic values, thins northwards from the Lake Tahoe – Reno area.

A thinner lithosphere beneath the N.Sierra and northwards, would account for the less radiogenic Sr and more elevated Nd isotopic values compared to the Southern Ancestral Cascades, as it may allow for shorter periods of interaction between the rising magma and the more radiogenic lithospheric component. A thicker lithosphere for the Southern Ancestral Cascades may also explain why andesitic lavas are the more dominant lava type, as a thicker lithosphere may force basaltic magmas to pond at deeper crustal levels, allowing the magmas to crystal fractionate to more evolved compositions.

Cascadia Mantle beneath the N. Sierra

From studying the basalts of the study area, it is clear that there are distinctive geochemical domains in the mantle. These distinctive signatures are quickly identifiable when looking at the Incompatible Element plots. There are three different geochemical components present in the sub-arc mantle beneath the N.Sierra, which are consistent with other author's findings for the Modern Cascades ((Bacon et al., 1997; Hildreth, 2007), (Hammond, 1989), (Reiners et al., 2000). These are a depleted source (LKT), an enriched source (WIP), and of course, subduction component(s) (CAB).

Strong and Wolff (2003) concluded that the mantle just north of LVC should be considered to be a heterogeneous mantle. These authors, and later Hildreth (2007), also propose that crustal contamination processes do not greatly affect (if at all) basaltic melts.

This lead the authors to suggest that as opposed to basaltic melts being stored in crustal depth batch melts awaiting ascent, the basaltic melts ascend as dike-like structures through the crust (Figure 32) which rarely have a chance to undergo any assimilation-fractional crystallization (AFC) processes which would lead to the generation of more evolved composition. This mantle model nicely describes the possible origins of melts from the N.Sierra.

The one magma type that Stong and Wolff (2003) do not have in their model that needs to be addressed is LKT magma genesis. LKTs from the Mt. Shasta area have origins related to decompression-dominated melting of a MORB like mantle (Hart et al., 1984), while LKT trace element compositions suggest a melting model (Schmidt, 2008) where the degree of melting is high and the fluid input is low. As Schmidt (2008) further explains, a warm mantle, due to the subduction of a young and warm oceanic plate may allow for high degrees of melting of the mantle wedge without needing the aid of fluids derived from the slab to initiate melting, as is required for the CAB.

Due to the above constraints on LKT genesis noted elsewhere in the Modern South Cascades, LKT genesis for the N.Sierra is proposed to be close to the base continental crust, and much like the WIP (Figure 32, “OIB-like magma”), are contained in lenses within the mantle wedge.

Figure 32: Model modified from (Strong and Wolff, 2003) for magma generation to the northeast (~150km) of the LVC. This model is accepted as a best-fit model for magma generation for the N.Sierra. Curved black arrows indicate fluid flux from subducting slab. OIB-like magma is equivalent to WIP from the N.Sierra. Lines through crust represent simplified indication of Basin and Range faulting.

CONCLUSIONS

Tertiary volcanic rocks collected from the northern Sierra Nevada area, between Lake Tahoe and Reno to the south and the Lassen Volcanic Centre to the north, are calc-alkaline basalts through to rhyolites. Common phenocrysts in the samples include plagioclase, olivine and clinopyroxene, with less common phenocrysts of hornblende, biotite, and an unidentified opaque mineral, probably a Fe-Ti oxide.

Fieldwork focused on collecting samples from an area that lies between two geochemical different segments of continental arc volcanism, the Southern Modern Cascades, which terminates at the Lassen Volcanic Centre (LVC) and the Tertiary Southern Ancestral Cascades, which is located around the Lake Tahoe-Reno area and to the south. The goal was to determine if any transition, either geochemically, petrological or both occurs between the two different volcanic sequences.

The twelve basalts collected fall into four end member compositions. Three are low-potassium tholeiites (LKT), three are within plate basalts (WIP), one is a Lovejoy basalt related to the Columbia River flood basalts, and five are calc-alkaline basalts (CAB). Also, 25 samples are basaltic andesites, 17 are andesite, one is a dacite and two are rhyolites.

WIP basalts are derived from an enriched mantle source, while LKTs are thought to be derived from an anhydrous, depleted mantle. CABs are generated by fluid flux melting of a heterogeneous peridotite mantle. The basaltic andesites and andesites also

are considered to be direct partial melts from the mantle wedge, however, they may have undergone some crystal fractionation and possibly temporary stagnation and contamination at crustal levels as indicated by multi-zoned plagioclases and slightly elevated $\delta^{18}\text{O}$ values. The dacite is formed by partially melting a basaltic parent, which exhibits a MORB-like isotopic signature, while one rhyolite is formed by the partial melting of a basaltic andesite parent. The other rhyolite, with its large negative Eu anomaly is suggested to have formed due to the crystal fractionation of an andesitic parent.

Sr isotopic values of the LKTs and the WIPs for the N.Sierra are very restricted in their range, while the CAB, basaltic andesites, andesites and felsic rocks exhibit a wide variance of $^{87}\text{Sr}/^{86}\text{Sr}$.

Comparisons of the N. Sierra lavas to those of the Ancestral Cascades arc and of the Modern South Cascades show both similarities and differences. The trace element abundances are very similar among all three regions, which suggest that the fluids that are coming off the slab were similar in elemental composition. The noticeable difference among the three regions is their Sr isotopic values, which become more radiogenic from the LVC through the N. Sierra to the Lake Tahoe-Reno area segment of the Ancestral Cascades. It is therefore concluded that petrologically N.Sierra lavas more closely resemble lavas from the LVC. N.Sierran lavas are dominantly mafic which is similar to LVC lavas, while Ancestral Cascades are dominantly andesitic. WIPs and LKTs are found in the N.Sierra and the LVC but not in the Ancestral Cascades. Geochemically however, N.Sierran lavas show affiliation to neither the Modern Southern Cascade lavas nor to Ancestral Cascades in the Lake Tahoe-Reno region. N.Sierran lavas

geochemically show that the area between the two Cascades arc may be considered a transition zone where an ancient enriched lithosphere thins northwards from the Lake Tahoe-Reno area, through the N.Sierra, towards the LVC and possibly beyond.

FURTHER WORK

Although the geochemical database for that of the entire Cascades arc is growing with every study, there clearly needs to be an emphasis put on the dating of more samples. A very helpful tool in this study would have been the ability to have some of the samples dated, especially the LKTs and WIP as this would help in piecing together the complicated tectonic and volcanic setting of the area.

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Appendix I: Thin Section Summary

Sample #	Rock type	Ol (%)	Cpx (%)	Plg (%)	Amph (%)	Bio (%)	Qtz (%)	Potassic Feldspar (%)	Mt (%)	Groundmass	Xenocrysts	Weathering
08-LT-01	Basalt	2	—	—	2	—	—	—	—	1 Ol, plq, mt	—	Moderate
08-LT-02	Andesite	5	5	5	—	—	—	—	—	Pvx, plq, mt	—	—
08-LT-03	Andesite	2	10	10	—	—	—	—	—	1 Pvx, plq, mt, ol	Glomerophrys of ol and plq.	—
08-LT-04	Basalt	2	2	10	—	—	—	—	—	1 Ol, plq, mt	Glomerophrys of ol and plq.	Weak (Ol only)
08-LT-05	Andesite	2	1	2	—	—	—	—	—	plq, pvx, ol, mt	—	—
08-LT-06	Andesite	5	10	15	—	—	—	—	—	plq, pvx, ol, mt	—	—
08-LT-07	Andesite	—	—	1	10	—	—	—	—	plq, pvx, ol, mt	Glomerophrys of pyr and plq.	—
08-LT-08	Basalt	10	—	—	10	—	—	—	—	Ol, plq	Glomerophrys of ol and plq.	Slight
08-LT-09	Basalt	5	—	—	5	—	—	—	—	Ol, plq	—	Moderately
08-LT-10	Basaltic Andesite	2	2	2	—	—	—	—	—	plq, pvx, ol, mt	Glomerophrys of ol and pvx.	—
08-LT-11	Basaltic Andesite	10	5	15	—	—	—	—	—	plq, pvx, ol, mt	—	Slight
08-LT-12	Basalt	10	—	—	5	—	—	—	—	Ol, plq	—	Slight
08-LT-13	Andesite	—	—	2	2	—	—	—	—	plq, pvx, mt	—	—
08-LT-14	Dacite	—	—	5	5	—	—	—	—	plq, pvx, mt	—	—
08-LT-15	Andesite	—	—	2	5	—	—	—	—	plq, pvx, qls, mt	Glomerophrys of pvx and plq.	Slight
08-LT-16	Andesite	<1	—	2	2	—	—	—	—	plq, pvx, mt	—	—
08-LT-17	Andesite	—	—	15	15	—	3	2	2	plq, pvx, mt	—	—
08-LT-18	Basalt	5	1	5	—	—	—	—	—	Ol, plq, mt	—	Moderately
08-LT-19	Basaltic Andesite	2	5	15	—	—	—	—	—	Plq, ol	—	—
08-LT-20	Andesite	—	—	10	15	3	—	—	—	pvx, plq, mt	—	—
08-LT-21	Basaltic Andesite	2	10	15	—	—	—	—	—	plq, ol	—	—
08-LT-22	Basaltic Andesite	2	2	5	—	—	—	—	—	plq, pvx, ol, mt	—	—
08-LT-23	Basalt	5	—	—	5	—	—	—	—	plq, qls, mt	—	—
08-LT-24	Basaltic Andesite	2	2	10	<1	—	—	—	—	plq, pvx, ol, mt	—	—
08-LT-25	Basalt	5	2	10	—	—	—	—	—	plq, pvx, ol	—	—
08-LT-26	Basaltic Andesite	2	—	—	2	—	—	1	—	plq, ol, qls	Otz	—
08-LT-27	Basaltic Andesite	2	—	—	2	—	—	1	—	plq, ol, qls	Qtz and possible Gabbro fragments	—

08-LT-28	Basalt	5	1	10	—	—	—	—	plq, ol, mt	Moderately
08-LT-29	Basalt	5	—	10	—	—	—	—	plq, ol, mt	
08-LT-30	Basaltic Andesite	2	5	5	—	—	—	—	plq, pyx, ol, mt	
08-LT-31	Basaltic Andesite	2	5	5	—	—	—	—	plq, pyx, ol, mt	
08-LT-32	Rhyolite	—	—	5	5	—	—	—	plq, qls	
08-LT-33	Rhyolite	—	—	1	5	2	5	—	plq, qls, mt	
08-LT-34	Andesite	—	5	15	5	—	—	—	plq, pyx, ol, qls, mt	
08-LT-35	Andesite	2	—	15	5	1	—	—	plq, pyx	
08-LT-36	Basaltic Andesite	1	—	5	—	—	—	—	plg, ol, mt	
08-LT-37	Basaltic Andesite	—	5	15	—	—	—	—	2 plg, pyx, ol	Qtz crystals
08-LT-38	Basaltic Andesite	—	5	10	—	—	1	—	plg, pyx, mt	
08-LT-39	Basaltic Andesite	5	1	5	—	—	—	—	plg, ol, mt	
08-LT-40	Basalt	2	1	2	—	—	—	—	2 plq, ol, qls, mt	
08-LT-41	Basaltic Andesite	5	5	10	—	—	—	—	plg, ol	
08-LT-42	Andesite	—	2	10	5	—	—	—	plq, pyx	Qtz and granitic xenoliths
08-LT-43	Basaltic Andesite	<1	—	5	—	—	—	—	plq, qls	Qtz and granitic xenoliths
08-LT-44	Basaltic Andesite	—	10	15	5	—	—	—	plq, pyx, mt	Qtz
08-LT-45	Basaltic Andesite	10	—	15	—	—	—	—	plq, pyx	
08-LT-46	Basaltic Andesite	1	—	5	—	—	—	—	plq, qls, mt	
08-LT-47	Basaltic Andesite	5	—	5	—	—	—	—	plq, pyx, ol, mt	
08-LT-48	Andesite	—	2	10	1	—	—	—	plq, pyx, mt	Qtz and granitic xenoliths
08-LT-49	Andesite	—	2	15	5	—	—	—	plq, pyx, qls	Qtz and granitic xenoliths
08-LT-50	Basaltic Andesite	—	1	5	—	—	—	—	plq, pyx, qls, mt	
08-LT-51	Andesite	2	5	10	1	—	—	—	plq, pyx, qls, mt	
08-LT-52	Andesite	—	2	10	2	—	—	—	plq, pyx, qls	Stronaly
08-LT-53	Basaltic Andesite	2	1	5	—	—	—	—	plq, ol, qls	Qtz and granitic xenoliths
08-LT-54	Basaltic Andesite	1	—	10	—	—	—	1	plq, qls, ol	
08-LT-55	Basalt	2	—	2	—	—	—	—	qls, plq	

08-LT-56	Basaltic Andesite	2	2	2	—	—	—	—	—	plg, ol, pyx, gls, mt	Moderately
08-LT-57	Basaltic Andesite	2	—	5	2	—	—	—	—	plg, ol, pyr	
08-LT-58	Basaltic Andesite	2	1	2	—	—	—	—	—	plg pyr, ol, gls	Moderately

Notes: Ol = olivine, px = pyroxene, cpx = clinopyroxene, opx = orthopyroxene, plg = plagioclase, amph = amphibole, bio = biotite,
alk = alkali, fsp = feldspar, qtz = quartz, mt = magnetite, gls = glass, aug = augite. — = no phenocrysts were present in thin section.

Appendix II: Major, Trace and Isotopic Data

A) Major Element Data

Label	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ t	MnO	MgO
08-LT-01	51.96	1.301	17.54	9.142	0.150	5.61
08-LT-02	59.72	0.651	16.91	6.149	0.092	3.51
08-LT-03	58.60	0.716	17.01	6.580	0.112	3.93
08-LT-04	51.35	1.345	17.43	9.144	0.150	6.24
08-LT-05	57.93	0.711	16.84	6.943	0.116	4.21
08-LT-06	59.24	0.672	16.95	6.380	0.103	3.74
08-LT-07	58.11	0.774	17.29	7.018	0.119	4.13
08-LT-08	50.79	1.554	17.69	9.353	0.149	5.99
08-LT-09	48.13	0.912	17.52	10.294	0.172	8.73
08-LT-10	53.90	0.650	15.48	8.308	0.144	7.20
08-LT-11	55.80	0.785	17.09	7.135	0.107	5.56
08-LT-12	47.81	0.929	17.56	10.087	0.171	8.89
08-LT-13	61.23	0.638	16.05	5.640	0.098	3.16
08-LT-14	63.34	0.601	16.45	5.175	0.074	2.47
08-LT-15	62.50	0.583	16.21	5.068	0.086	2.48
08-LT-16	61.21	0.668	16.52	5.886	0.079	2.99
08-LT-17	59.46	0.717	17.38	5.734	0.097	2.75
08-LT-18	47.89	0.830	17.90	10.482	0.173	8.22
08-LT-19	52.40	0.795	19.30	9.052	0.169	3.50
08-LT-20	59.73	0.517	17.52	6.630	0.130	2.69
08-LT-21	52.09	0.813	17.71	9.261	0.149	5.94
08-LT-22	52.54	0.655	15.69	8.843	0.155	7.45
08-LT-23	50.56	0.887	16.66	10.240	0.415	5.61
08-LT-24	53.65	0.787	18.39	8.934	0.146	4.53
08-LT-25	50.07	0.741	18.93	9.633	0.156	6.02
08-LT-26	55.71	0.714	18.12	6.851	0.119	3.40
08-LT-27	55.27	0.720	18.06	7.167	0.121	3.85
08-LT-28	51.81	0.822	16.72	9.841	0.172	6.94
08-LT-29	51.26	0.812	16.50	9.738	0.174	7.01
08-LT-30	52.41	0.893	16.85	9.109	0.154	6.25
08-LT-31	52.96	0.775	16.41	8.910	0.151	6.55
08-LT-32	72.21	0.168	13.19	2.596	0.043	0.35
08-LT-33	71.62	0.190	15.47	1.453	0.067	0.29
08-LT-34	58.25	0.657	17.93	6.602	0.102	2.95
08-LT-35	59.41	0.556	18.17	6.035	0.116	2.72
08-LT-36	56.91	0.544	19.75	7.182	0.117	1.68
08-LT-37	54.77	0.736	19.22	8.556	0.137	3.32
08-LT-38	55.95	0.580	19.83	7.376	0.124	3.03
08-LT-39	53.43	0.818	16.76	9.437	0.170	7.13
08-LT-40	52.00	0.803	16.45	9.282	0.160	6.92
08-LT-41	52.37	0.737	16.26	9.089	0.151	6.93
08-LT-42	59.00	0.592	18.18	6.646	0.123	2.70
08-LT-43	56.46	0.586	19.36	7.567	0.129	2.92
08-LT-44	56.73	0.794	17.74	7.740	0.112	3.37
08-LT-45	55.42	0.909	19.26	7.882	0.108	2.21
08-LT-46	56.55	0.790	19.33	7.358	0.088	2.04
08-LT-47	52.74	0.742	17.28	8.442	0.138	7.34
08-LT-48	58.48	0.644	18.15	6.941	0.130	2.30
08-LT-49	58.50	0.618	18.00	7.212	0.124	3.63
08-LT-50	56.96	0.543	18.96	7.589	0.134	2.31
08-LT-51	58.30	0.771	17.35	7.936	0.120	2.42
08-LT-52	60.46	0.691	17.71	7.241	0.134	1.11
08-LT-53	53.36	0.812	17.75	8.545	0.140	5.80
08-LT-54	55.13	1.422	16.09	8.681	0.136	3.70
08-LT-55	50.02	2.317	14.17	14.192	0.235	4.61
08-LT-56	55.27	0.635	16.97	7.532	0.130	5.80
08-LT-57	53.34	0.804	18.29	8.821	0.154	5.22
08-LT-58	54.93	0.687	18.36	7.931	0.129	4.27
Precision*	0.5	1.9	1.2	1.6	3.4	1.7

* Note: Precision is based on percent difference on duplicate samples.

Major Element data con't.

Label	CaO	Na2O	K2O	P2O5	FeO	Mg#	LOI
08-LT-01	7.97	4.08	1.249	0.468	8.23	57.4	0.27
08-LT-02	6.37	3.49	1.945	0.201	5.53	55.7	0.40
08-LT-03	6.58	3.44	1.891	0.204	5.92	56.8	0.74
08-LT-04	8.12	3.73	1.256	0.442	8.23	60.0	0.56
08-LT-05	6.60	3.32	1.962	0.187	6.25	57.2	1.77
08-LT-06	6.69	3.51	1.771	0.207	5.74	56.3	0.36
08-LT-07	6.81	3.60	1.565	0.278	6.32	56.4	0.14
08-LT-08	8.23	3.98	1.375	0.495	8.42	58.5	0.31
08-LT-09	11.07	2.46	0.149	0.083	9.26	65.1	0.61
08-LT-10	9.73	4.27	0.938	0.138	7.48	65.6	0.44
08-LT-11	8.27	3.19	1.133	0.172	6.42	63.2	0.39
08-LT-12	10.53	2.64	0.176	0.095	9.08	66.0	0.26
08-LT-13	5.51	3.32	2.811	0.252	5.08	55.2	0.56
08-LT-14	4.97	3.59	2.576	0.218	4.66	51.2	0.70
08-LT-15	5.11	3.37	2.694	0.207	4.56	51.8	1.25
08-LT-16	5.55	3.64	2.298	0.266	5.30	52.8	0.10
08-LT-17	5.92	4.05	1.563	0.262	5.16	51.3	2.15
08-LT-18	10.65	2.37	0.256	0.117	9.43	63.3	1.42
08-LT-19	9.56	2.84	0.786	0.134	8.15	46.0	2.03
08-LT-20	6.18	3.64	1.279	0.189	5.97	47.2	2.06
08-LT-21	9.57	2.92	0.828	0.172	8.33	58.5	0.58
08-LT-22	9.58	3.51	1.076	0.150	7.96	65.0	1.18
08-LT-23	7.54	3.13	2.318	0.142	9.22	54.7	2.80
08-LT-24	8.42	3.01	1.315	0.143	8.04	52.7	0.33
08-LT-25	9.77	2.36	1.015	0.146	8.67	57.9	1.81
08-LT-26	8.19	3.60	1.282	0.198	6.17	52.2	1.72
08-LT-27	8.04	3.58	1.250	0.197	6.45	54.2	1.69
08-LT-28	10.55	2.43	0.967	0.179	8.86	60.8	0.54
08-LT-29	10.48	2.43	0.949	0.176	8.76	61.3	0.09
08-LT-30	9.60	2.69	1.009	0.223	8.20	60.2	0.36
08-LT-31	10.02	2.53	1.025	0.168	8.02	61.8	0.48
08-LT-32	1.41	3.34	4.957	0.015	2.34	22.9	1.71
08-LT-33	1.67	4.97	2.472	0.083	1.31	30.5	1.69
08-LT-34	6.59	3.93	1.338	0.218	5.94	49.6	3.04
08-LT-35	6.23	4.03	1.269	0.220	5.43	49.8	1.10
08-LT-36	7.01	4.06	0.913	0.317	6.46	34.0	2.26
08-LT-37	7.49	3.55	0.999	0.214	7.70	46.1	1.53
08-LT-38	7.71	3.77	1.054	0.211	6.64	47.5	0.33
08-LT-39	10.32	2.58	1.188	0.217	8.49	62.4	0.13
08-LT-40	10.17	2.55	1.156	0.214	8.35	62.1	0.47
08-LT-41	9.98	2.68	0.949	0.170	8.18	62.7	0.70
08-LT-42	6.42	3.71	1.684	0.249	5.98	47.2	1.23
08-LT-43	7.20	3.92	0.995	0.253	6.81	45.9	0.36
08-LT-44	6.73	3.42	1.897	0.231	6.97	48.9	1.55
08-LT-45	7.15	4.25	1.117	0.263	7.09	38.2	0.94
08-LT-46	6.87	4.29	1.083	0.248	6.62	37.9	0.78
08-LT-47	8.88	3.17	0.699	0.172	7.60	65.7	0.19
08-LT-48	6.45	4.11	1.563	0.304	6.25	42.2	1.59
08-LT-49	7.16	3.52	1.399	0.192	6.49	52.5	0.38
08-LT-50	7.12	3.53	1.278	0.208	6.83	40.1	1.53
08-LT-51	5.50	4.02	1.741	0.213	7.14	40.2	1.43
08-LT-52	4.79	4.14	1.907	0.216	6.52	25.2	1.53
08-LT-53	8.58	3.23	1.038	0.217	7.69	59.9	0.79
08-LT-54	6.47	3.48	2.861	0.635	7.81	48.4	1.22
08-LT-55	8.04	3.10	1.825	0.923	12.77	41.7	0.36
08-LT-56	7.89	3.33	0.999	0.156	6.78	62.9	1.09
08-LT-57	8.25	3.47	0.928	0.208	7.94	56.6	0.26
08-LT-58	7.95	3.72	0.971	0.207	7.14	54.2	0.66
Precision*	2	2.7	0.7	5			3

* Note: Precision is based on percent difference on duplicate samples.

B) Trace Element Data

Label	Ba	Ce	Co	Cr	Cs	Dy	Er
08-LT-01	585	46.13	33.90	108	0.30	4.88	2.85
08-LT-02	766	41.77	21.11	67	1.68	3.79	2.17
08-LT-03	781	43.45	23.10	86	1.27	3.61	2.19
08-LT-04	536	45.91	36.49	148	0.22	4.74	2.75
08-LT-05	636	35.07	26.01	77	2.54	3.25	1.96
08-LT-06	755	41.83	21.63	72	1.28	3.43	2.00
08-LT-07	704	44.31	23.61	92	0.59	3.69	2.16
08-LT-08	514	51.44	34.87	126	0.16	4.86	2.82
08-LT-09	121	10.13	49.06	271	0.05	4.32	2.94
08-LT-10	246	20.38	32.31	243	0.48	3.44	2.13
08-LT-11	286	28.15	26.07	113	0.38	3.16	1.85
08-LT-12	84	10.17	50.93	219	0.02	4.37	2.91
08-LT-13	935	51.59	18.04	75	2.44	3.52	2.03
08-LT-14	963	47.44	14.68	51	2.42	4.95	2.73
08-LT-15	952	49.34	14.67	40	3.07	3.27	1.94
08-LT-16	934	48.97	15.97	80	2.16	4.71	2.79
08-LT-17	889	43.63	13.15	23	1.60	2.39	1.20
08-LT-18	210	12.28	51.43	240	0.07	4.20	2.91
08-LT-19	426	19.39	25.34	24	1.38	3.41	2.21
08-LT-20	599	28.90	17.60	19	1.21	3.26	2.03
08-LT-21	327	24.05	35.26	103	0.36	3.26	1.97
08-LT-22	391	32.87	35.61	278	0.64	3.67	2.26
08-LT-23	784	12.15	37.50	80	0.75	3.81	2.37
08-LT-24	524	32.63	29.53	32	0.40	2.72	1.55
08-LT-25	437	36.95	35.68	48	0.92	2.94	1.64
08-LT-26	537	24.80	22.25	46	1.70	2.42	1.35
08-LT-27	508	23.85	24.84	52	1.56	2.37	1.30
08-LT-28	288	28.49	31.90	169	0.22	3.49	2.09
08-LT-29	276	29.35	32.78	171	0.21	3.62	2.19
08-LT-30	353	31.89	29.86	138	0.29	3.54	2.01
08-LT-31	295	27.54	31.11	160	0.53	3.19	1.82
08-LT-32	119	151.40	0.45	<10	1.45	7.71	4.27
08-LT-33	1243	55.35	0.59	<10	2.57	1.47	0.92
08-LT-34	746	32.32	17.64	20	1.08	2.51	1.38
08-LT-35	738	26.06	14.67	16	0.65	2.22	1.26
08-LT-36	519	24.20	13.88	<10	0.94	2.33	1.29
08-LT-37	510	23.42	22.12	13	0.55	2.89	1.67
08-LT-38	568	22.90	20.74	20	0.60	2.53	1.48
08-LT-39	370	29.99	32.09	188	0.20	3.55	2.11
08-LT-40	355	29.86	32.83	177	0.37	3.47	2.08
08-LT-41	345	22.24	35.74	197	0.46	2.51	1.50
08-LT-42	973	40.42	16.08	<10	0.88	3.02	1.85
08-LT-43	476	30.11	18.53	11	0.49	2.80	1.62
08-LT-44	841	40.81	22.80	46	2.10	3.88	2.35
08-LT-45	477	26.97	18.08	<10	0.45	2.46	0.88
08-LT-46	557	22.13	17.30	15	0.64	2.39	0.89
08-LT-47	444	17.80	34.54	358	0.24	2.73	1.59
08-LT-48	873	37.99	14.34	32	0.76	2.86	1.66
08-LT-49	698	30.85	20.16	55	1.08	2.93	1.68
08-LT-50	653	29.78	15.82	<10	0.73	3.04	1.86
08-LT-51	1216	38.11	20.54	21	1.38	3.90	2.30
08-LT-52	849	35.50	17.12	14	0.99	3.13	1.84
08-LT-53	548	29.66	33.51	145	0.52	3.85	2.42
08-LT-54	878	94.53	24.60	87	9.63	6.24	3.42
08-LT-55	1500	50.92	36.08	38	1.30	8.25	4.70
08-LT-56	683	18.22	28.94	260	1.57	2.30	1.33
08-LT-57	501	21.53	30.60	104	0.69	2.76	1.55
08-LT-58	811	22.85	25.88	71	0.68	2.63	1.56
Precision*	7	2.00	15.00	6	15.00	6.00	7.00

* Note: Precision is based on percent difference on duplicate samples.

Trace Element data con't.

Label	Eu	Gd	Hf	Ho	La	Lu	Nb
08-LT-01	1.76	5.39	3.87	1.03	22.12	0.38	12.05
08-LT-02	1.26	4.41	3.75	0.75	30.30	0.31	6.39
08-LT-03	1.21	3.98	3.62	0.74	23.54	0.31	6.87
08-LT-04	1.69	5.11	3.88	0.95	21.87	0.39	15.87
08-LT-05	1.09	3.46	2.96	0.66	17.57	0.29	5.82
08-LT-06	1.16	3.85	3.53	0.69	21.84	0.29	6.21
08-LT-07	1.31	4.17	3.59	0.75	21.75	0.31	6.35
08-LT-08	1.80	5.29	4.40	0.97	24.71	0.39	21.83
08-LT-09	1.00	3.49	1.73	0.95	4.04	0.44	1.48
08-LT-10	1.12	3.57	1.90	0.72	10.89	0.30	1.52
08-LT-11	1.23	3.63	2.56	0.65	14.54	0.25	2.31
08-LT-12	1.06	3.55	1.76	0.94	3.84	0.44	1.09
08-LT-13	1.21	3.96	4.45	0.70	26.10	0.31	7.91
08-LT-14	1.89	6.17	4.43	0.95	40.60	0.38	7.58
08-LT-15	1.09	3.68	4.33	0.65	25.42	0.28	7.42
08-LT-16	1.56	5.59	4.37	0.96	33.83	0.41	8.06
08-LT-17	1.29	3.34	2.31	0.44	20.97	0.16	5.07
08-LT-18	1.02	3.43	1.58	0.94	5.39	0.44	1.73
08-LT-19	0.99	3.24	1.74	0.73	9.06	0.33	2.07
08-LT-20	1.07	3.32	2.70	0.68	14.25	0.32	3.48
08-LT-21	1.09	3.32	2.04	0.68	10.61	0.28	3.05
08-LT-22	1.20	3.88	2.37	0.76	18.95	0.32	2.18
08-LT-23	0.93	3.36	2.44	0.80	4.86	0.35	2.30
08-LT-24	1.05	3.01	2.19	0.55	17.24	0.23	2.38
08-LT-25	1.14	3.36	1.82	0.59	19.38	0.23	1.83
08-LT-26	0.94	2.74	1.89	0.49	11.63	0.19	3.01
08-LT-27	0.94	2.60	1.95	0.45	11.00	0.18	2.93
08-LT-28	1.12	3.65	2.22	0.71	12.70	0.30	4.04
08-LT-29	1.16	3.85	2.22	0.76	13.40	0.30	4.02
08-LT-30	1.27	3.93	2.32	0.72	14.46	0.28	4.28
08-LT-31	1.07	3.43	2.09	0.64	12.87	0.25	3.57
08-LT-32	0.42	8.74	7.50	1.49	77.41	0.65	17.08
08-LT-33	0.66	1.70	3.48	0.29	30.40	0.18	5.37
08-LT-34	1.03	2.98	2.61	0.49	16.09	0.20	3.66
08-LT-35	0.91	2.52	2.38	0.45	13.00	0.20	3.43
08-LT-36	1.06	2.77	2.27	0.47	12.37	0.19	3.34
08-LT-37	1.04	3.07	2.15	0.59	11.14	0.24	3.16
08-LT-38	0.92	2.76	2.28	0.51	11.22	0.23	2.70
08-LT-39	1.24	3.75	2.59	0.73	13.88	0.30	2.72
08-LT-40	1.23	3.80	2.55	0.71	13.38	0.30	2.76
08-LT-41	0.98	2.86	1.82	0.51	10.15	0.21	2.27
08-LT-42	1.09	3.49	3.77	0.63	21.64	0.29	5.75
08-LT-43	1.03	3.15	2.47	0.57	14.63	0.26	4.57
08-LT-44	1.24	4.22	4.04	0.80	21.83	0.34	4.93
08-LT-45	1.16	3.62	2.97	0.41	12.39	0.08	3.75
08-LT-46	1.24	3.71	2.59	0.38	13.25	0.09	3.44
08-LT-47	0.89	2.81	1.70	0.55	8.37	0.23	2.13
08-LT-48	1.17	3.48	3.26	0.57	19.05	0.23	5.14
08-LT-49	1.06	3.27	2.74	0.59	16.54	0.26	3.90
08-LT-50	1.04	3.25	2.58	0.63	14.71	0.29	3.82
08-LT-51	1.31	4.22	3.47	0.78	21.93	0.34	4.26
08-LT-52	1.10	3.51	3.38	0.65	18.27	0.28	4.74
08-LT-53	1.30	4.34	2.22	0.83	18.88	0.30	3.39
08-LT-54	1.76	7.71	8.87	1.24	45.15	0.46	10.66
08-LT-55	3.17	9.00	3.96	1.67	22.71	0.61	6.41
08-LT-56	0.85	2.52	2.00	0.47	8.56	0.19	2.29
08-LT-57	1.02	3.08	2.00	0.55	10.34	0.22	2.68
08-LT-58	0.99	2.90	2.19	0.53	11.14	0.22	2.71
Precision*	5.00	2.00	5.00	1.00	2.00	3.00	6.00

* Note: Precision is based on percent difference on duplicate samples.

Trace Element data con't.

Label	Nd	Ni	Pb	Pr	Rb	Sc	Sm
08-LT-01	26.07	95.40	6.60	6.09	23.22	23.60	5.58
08-LT-02	25.57	59.20	10.50	6.53	54.25	19.40	5.00
08-LT-03	21.26	53.10	9.30	5.46	51.69	20.50	4.27
08-LT-04	24.86	113.70	4.60	5.89	23.90	25.00	5.28
08-LT-05	17.23	61.10	9.70	4.24	53.28	19.50	3.68
08-LT-06	20.52	53.80	10.40	5.16	46.50	19.10	4.13
08-LT-07	22.06	57.60	8.50	5.54	34.57	19.20	4.63
08-LT-08	27.12	99.30	4.60	6.51	24.86	24.30	5.70
08-LT-09	8.25	131.40	1.30	1.63	1.19	46.50	2.63
08-LT-10	14.11	38.60	4.00	3.10	10.02	35.40	3.42
08-LT-11	18.64	44.10	6.60	4.27	12.01	29.70	3.91
08-LT-12	8.35	178.00	1.10	1.65	2.62	42.60	2.73
08-LT-13	23.48	45.90	12.70	6.06	72.04	15.50	4.66
08-LT-14	39.43	33.90	13.10	10.11	76.22	14.10	7.70
08-LT-15	22.02	30.20	14.30	5.81	79.26	13.20	4.27
08-LT-16	31.72	44.10	11.60	7.97	64.43	15.90	6.22
08-LT-17	22.37	9.60	9.50	5.53	33.68	11.40	4.32
08-LT-18	9.28	216.50	1.30	1.89	3.09	43.30	2.70
08-LT-19	11.88	10.40	5.30	2.63	20.58	31.60	3.07
08-LT-20	15.76	10.90	6.90	3.71	33.97	15.60	3.53
08-LT-21	14.52	36.10	2.80	3.30	12.90	32.70	3.37
08-LT-22	19.70	80.60	4.30	4.76	22.10	36.70	4.16
08-LT-23	8.69	39.70	8.80	1.82	122.89	38.50	2.81
08-LT-24	16.72	19.40	8.10	4.09	22.85	23.10	3.45
08-LT-25	19.09	38.20	9.80	4.62	14.20	30.30	3.94
08-LT-26	13.99	32.60	9.60	3.27	25.27	19.60	3.07
08-LT-27	13.20	38.80	9.30	3.16	24.11	18.40	2.97
08-LT-28	16.91	22.50	3.20	3.88	11.97	37.40	3.91
08-LT-29	17.57	23.70	8.40	4.05	12.22	38.80	4.00
08-LT-30	18.82	19.00	5.60	4.40	13.06	33.30	4.22
08-LT-31	16.35	22.00	7.90	3.83	14.25	36.10	3.63
08-LT-32	58.48	<1.6	5.70	16.80	155.77	2.70	10.47
08-LT-33	19.80	<1.6	18.30	6.02	51.31	2.00	2.76
08-LT-34	17.30	16.30	25.00	4.26	27.67	14.90	3.46
08-LT-35	13.77	8.50	9.80	3.39	23.14	10.60	2.84
08-LT-36	15.35	1.80	7.40	3.59	16.49	7.70	3.24
08-LT-37	13.67	12.80	5.90	3.25	19.43	15.20	3.25
08-LT-38	13.32	12.80	7.90	3.17	18.95	12.60	3.03
08-LT-39	18.15	27.00	3.30	4.19	12.97	36.60	4.03
08-LT-40	18.30	26.10	4.10	4.18	12.99	37.60	4.09
08-LT-41	13.83	41.70	5.80	3.15	12.57	36.30	3.14
08-LT-42	21.03	5.90	9.90	5.35	43.12	13.00	4.07
08-LT-43	16.72	4.50	7.40	4.06	18.08	13.10	3.47
08-LT-44	22.82	21.90	16.40	5.71	56.83	22.90	4.75
08-LT-45	16.32	3.30	8.20	3.79	18.77	6.60	3.89
08-LT-46	17.26	6.50	8.50	3.87	14.45	6.40	3.92
08-LT-47	10.91	124.70	6.60	2.51	9.98	27.80	2.70
08-LT-48	20.56	2.20	13.00	5.00	36.22	10.40	4.02
08-LT-49	17.03	23.80	8.40	4.27	31.70	18.40	3.55
08-LT-50	16.55	1.70	8.30	3.98	33.02	12.30	3.50
08-LT-51	21.18	13.00	13.40	5.31	29.04	17.90	4.62
08-LT-52	18.56	7.70	10.50	4.55	41.95	16.30	3.86
08-LT-53	18.87	82.80	7.00	4.50	19.71	27.30	4.15
08-LT-54	48.61	32.90	26.80	12.17	122.44	22.60	9.65
08-LT-55	34.20	21.00	5.90	7.35	36.35	39.60	8.36
08-LT-56	11.05	97.90	8.40	2.51	18.47	22.40	2.61
08-LT-57	13.53	59.70	8.70	3.00	12.66	21.10	3.19
08-LT-58	13.29	46.60	8.40	3.12	18.10	19.70	3.07
Precision*	2.00	15.00	8.00	4.00	3.00	8.00	2.00

* Note: Precision is based on percent difference on duplicate samples.

Trace Element data con't.

Label	Sr	Ta	Tb	Th	Tm	U	V
08-LT-01	576.70	0.69	0.86	2.40	0.46	0.83	179
08-LT-02	464.10	0.43	0.63	5.22	0.31	1.82	132
08-LT-03	513.20	0.46	0.59	5.14	0.31	1.63	139
08-LT-04	532.00	0.92	0.78	2.27	0.39	0.75	183
08-LT-05	462.30	0.46	0.53	5.93	0.29	2.41	146
08-LT-06	484.90	0.42	0.56	4.84	0.29	1.70	143
08-LT-07	530.40	0.38	0.62	3.55	0.31	1.20	157
08-LT-08	580.40	1.30	0.79	2.53	0.40	0.81	191
08-LT-09	200.90	0.09	0.62	0.64	0.43	0.14	214
08-LT-10	580.80	0.09	0.54	2.21	0.30	0.92	210
08-LT-11	1086.60	0.13	0.52	2.13	0.26	0.71	203
08-LT-12	237.90	0.07	0.64	0.31	0.43	0.08	211
08-LT-13	442.60	0.54	0.58	6.36	0.30	2.26	110
08-LT-14	416.50	0.55	0.85	6.98	0.39	2.60	106
08-LT-15	411.00	0.55	0.54	7.61	0.28	2.76	105
08-LT-16	467.40	0.53	0.79	5.88	0.40	2.09	105
08-LT-17	1031.50	0.33	0.43	3.99	0.17	1.79	123
08-LT-18	245.10	0.10	0.60	0.51	0.43	0.13	220
08-LT-19	485.90	0.13	0.51	2.32	0.32	0.87	248
08-LT-20	460.90	0.24	0.51	3.34	0.30	1.48	104
08-LT-21	765.00	0.17	0.52	1.45	0.28	0.52	232
08-LT-22	654.90	0.13	0.58	3.74	0.33	1.19	222
08-LT-23	297.10	0.16	0.57	3.52	0.35	1.49	257
08-LT-24	578.20	0.17	0.45	5.88	0.23	2.10	227
08-LT-25	683.30	0.11	0.49	7.07	0.24	1.97	247
08-LT-26	746.90	0.19	0.40	1.96	0.20	0.91	176
08-LT-27	756.30	0.18	0.38	1.89	0.19	0.94	173
08-LT-28	613.00	0.23	0.57	2.67	0.31	0.90	265
08-LT-29	635.60	0.23	0.59	2.77	0.31	0.92	260
08-LT-30	780.80	0.24	0.59	2.73	0.30	0.90	240
08-LT-31	700.40	0.20	0.52	2.85	0.27	0.99	263
08-LT-32	71.20	1.11	1.29	29.69	0.64	6.88	<10
08-LT-33	446.00	0.35	0.25	8.37	0.15	2.97	12
08-LT-34	758.90	0.21	0.41	2.90	0.20	1.26	141
08-LT-35	724.40	0.21	0.36	1.93	0.19	0.96	103
08-LT-36	895.10	0.17	0.40	1.08	0.19	0.61	90
08-LT-37	715.40	0.18	0.47	1.38	0.24	0.59	145
08-LT-38	760.30	0.16	0.41	1.28	0.22	0.60	135
08-LT-39	831.70	0.15	0.58	2.22	0.31	0.93	248
08-LT-40	853.60	0.15	0.56	2.17	0.31	0.91	253
08-LT-41	859.90	0.13	0.42	1.92	0.21	0.81	247
08-LT-42	639.90	0.32	0.51	3.53	0.28	1.34	119
08-LT-43	716.10	0.26	0.45	1.64	0.24	0.82	114
08-LT-44	635.50	0.31	0.63	7.63	0.35	3.26	192
08-LT-45	808.40	0.21	0.49	1.02	0.11	0.48	137
08-LT-46	743.30	0.20	0.48	0.69	0.11	0.40	136
08-LT-47	599.00	0.12	0.43	1.15	0.23	0.46	199
08-LT-48	740.70	0.30	0.49	3.31	0.24	1.38	125
08-LT-49	629.70	0.24	0.48	3.34	0.26	1.43	139
08-LT-50	645.60	0.22	0.49	2.01	0.28	0.86	117
08-LT-51	502.10	0.28	0.63	5.92	0.34	2.41	151
08-LT-52	432.20	0.32	0.52	4.49	0.28	1.85	140
08-LT-53	577.60	0.19	0.62	3.01	0.32	0.99	195
08-LT-54	554.40	0.62	1.06	17.93	0.50	6.70	215
08-LT-55	393.20	0.39	1.32	3.58	0.65	1.42	375
08-LT-56	643.20	0.12	0.38	0.88	0.19	0.45	162
08-LT-57	799.30	0.15	0.45	1.83	0.22	0.80	193
08-LT-58	752.40	0.15	0.43	1.73	0.22	0.80	158
Precision*	2.00	20.00	2.00	5.00	7.00	5.00	4

* Note: Precision is based on percent difference on duplicate samples.

Trace Element data con't.

Label	Y	Yb	Zn	Zr
08-LT-01	27.28	2.51	83	166
08-LT-02	21.98	2.00	67	155
08-LT-03	21.75	2.04	68	147
08-LT-04	26.68	2.56	76	176
08-LT-05	18.74	1.88	69	113
08-LT-06	19.78	1.91	67	143
08-LT-07	20.86	2.02	76	147
08-LT-08	28.26	2.59	79	201
08-LT-09	25.67	2.87	67	67
08-LT-10	20.95	1.98	69	66
08-LT-11	18.62	1.67	67	90
08-LT-12	25.22	2.84	72	70
08-LT-13	20.16	1.98	66	184
08-LT-14	26.01	2.53	65	181
08-LT-15	18.44	1.86	61	177
08-LT-16	29.08	2.64	69	183
08-LT-17	12.25	1.06	91	77
08-LT-18	26.39	2.86	68	62
08-LT-19	19.76	2.15	77	61
08-LT-20	18.81	2.03	75	101
08-LT-21	18.17	1.87	80	74
08-LT-22	20.67	2.12	70	84
08-LT-23	21.74	2.27	272	87
08-LT-24	14.54	1.51	71	77
08-LT-25	15.12	1.57	78	64
08-LT-26	12.21	1.28	82	66
08-LT-27	11.98	1.22	82	67
08-LT-28	18.68	1.97	83	80
08-LT-29	19.22	2.06	94	78
08-LT-30	18.34	1.90	85	84
08-LT-31	16.79	1.73	82	75
08-LT-32	40.18	4.28	30	315
08-LT-33	8.63	1.06	54	133
08-LT-34	13.11	1.35	87	95
08-LT-35	11.62	1.24	82	89
08-LT-36	11.93	1.24	107	83
08-LT-37	15.60	1.62	87	78
08-LT-38	14.15	1.47	92	86
08-LT-39	21.49	2.05	78	90
08-LT-40	19.78	2.00	77	94
08-LT-41	14.16	1.36	90	64
08-LT-42	18.28	1.84	86	155
08-LT-43	16.24	1.63	93	95
08-LT-44	23.27	2.26	82	156
08-LT-45	10.82	0.59	94	112
08-LT-46	11.12	0.62	97	94
08-LT-47	15.17	1.51	79	59
08-LT-48	16.30	1.57	101	130
08-LT-49	15.93	1.69	84	102
08-LT-50	18.30	1.87	93	98
08-LT-51	22.13	2.27	84	130
08-LT-52	17.94	1.80	70	123
08-LT-53	27.80	1.92	78	84
08-LT-54	33.29	3.11	104	353
08-LT-55	42.97	4.08	130	146
08-LT-56	13.29	1.25	78	73
08-LT-57	15.46	1.42	92	72
08-LT-58	15.21	1.46	90	81
Precision*	6.00	3.00	3	3

* Note: Precision is based on percent difference on duplicate samples.

C) Isotopic Data

Label	87Sr/86Sr	143Nd/144Nd	208Pb/204Pb	207Pb/204Pb	206Pb/204Pb	Oxygen O16
08-LT-01	0.704024	0.512806	38.716	15.642	18.972	+ 5.1
08-LT-02	0.704478	0.512722	38.733	15.644	19.005	
08-LT-03	0.704176	0.512728	38.723	15.637	18.998	
08-LT-04	0.703883	0.512820	38.732	15.648	19.968	+ 6.5
08-LT-05	0.704340	0.512727	38.808	15.659	19.034	
08-LT-06	0.704387	0.512720	38.813	15.665	19.018	
08-LT-07	0.704270	0.512719	38.714	15.639	18.985	+ 7.3
08-LT-08	0.704058	0.513075	38.727	15.646	18.964	
08-LT-09	0.039320	0.512864	38.853	15.669	19.034	
08-LT-10	0.703414	0.512913	38.574	15.618	18.926	
08-LT-11	0.702964	0.512938	38.407	15.593	18.792	+ 6.1
08-LT-12	0.703877	0.512863	38.778	15.645	19.019	
08-LT-13	0.704361	0.512753	38.707	15.639	18.971	
08-LT-14	0.704418	0.512747	38.715	15.640	18.976	+ 7.6
08-LT-15	0.704482	0.512726	38.788	15.661	18.992	
08-LT-16	0.704301	0.512751	38.775	15.661	18.986	
08-LT-17	0.704816	0.512642	38.735	15.648	18.995	+ 6.9
08-LT-18	0.704142	0.512816	38.839	15.662	19.059	+ 6.3
08-LT-19	0.704283	0.512798	38.784	15.659	19.035	
08-LT-20	0.704256	0.512804	38.73	15.643	18.992	
08-LT-21	0.703589	0.512817	38.697	15.640	19.002	
08-LT-22	0.703791	0.512729	38.738	15.636	19.076	+ 6.8
08-LT-23	0.706792	0.512862	38.611	15.629	18.877	+ 6.7
08-LT-24	0.703846	0.512856	38.523	15.598	18.894	
08-LT-25	0.703934	0.512837	38.684	15.640	18.950	
08-LT-26	0.704644	0.512679	38.671	15.632	18.949	+ 7.7
08-LT-27	0.704623	0.512685	38.681	15.629	18.962	
08-LT-28	0.704014	0.512675	38.594	15.595	18.939	
08-LT-29	0.704011	0.512679	38.702	15.625	18.962	
08-LT-30	0.704034	0.512679	38.699	15.626	18.961	
08-LT-31	0.704083	0.512648	38.707	15.627	18.964	
08-LT-32	0.703399	0.512109				
08-LT-33	0.704288	0.512752	38.574	15.605	18.922	+ 7.4
08-LT-34	0.704546	0.512625	38.643	15.623	18.944	+ 7.6
08-LT-35	0.704469	0.512728	38.594	15.613	18.935	
08-LT-36	0.704442	0.512731	38.772	15.660	18.975	+ 7.5
08-LT-37	0.704486	0.512558	38.733	15.653	18.940	
08-LT-38	0.704422	0.512734	38.739	15.657	18.941	
08-LT-39	0.703441	0.512811	38.547	15.620	18.878	+ 6.8
08-LT-40	0.703445	0.512777	38.526	15.615	18.864	
08-LT-41	0.703891	0.512732	38.625	15.627	18.928	
08-LT-42	0.705265	0.512529	38.839	15.666	19.023	
08-LT-43	0.704781	0.512641	38.737	15.649	18.973	
08-LT-44	0.704510	0.512598	38.802	15.667	19.000	
08-LT-45	0.704352	0.512765	38.66	15.637	18.908	
08-LT-46	0.704273	0.512774	38.671	15.648	18.888	
08-LT-47	0.704361	0.512762	38.679	15.640	18.921	
08-LT-48	0.704693	0.512693	38.637	15.629	18.907	
08-LT-49	0.704671	0.512676	38.695	15.632	18.980	+ 7.4
08-LT-50	0.704739	0.512691	38.699	15.640	18.973	
08-LT-51	0.704017	0.512817	38.579	15.617	18.888	
08-LT-52	0.704250	0.512801	38.689	15.649	18.921	+ 6.7
08-LT-53	0.704162	0.512802	38.604	15.615	18.928	
08-LT-54	0.704507	0.512796	38.759	15.639	19.048	
08-LT-55	0.704197	0.512833	38.605	15.605	19.038	+ 7.1
08-LT-56	0.704482	0.512733	38.632	15.632	18.899	
08-LT-57	0.704370	0.512742				
08-LT-58	0.704368	0.512756				
Precision*	0.708070	0.511820	36.4943	15.425	16.888	
2 sigma	0.000011	0.000011	0.0567	0.017	0.014	0.05

* Pb stds = NBS 981 measured over 1.5 yrs, Sr std = E & A std measured over 2 years and Nd std measured over 4 years.

Appendix III: Analytical Techniques

A) Thin Section Preparation

Rock samples were cut and thin sections were cut, mounted and select sections polished at Carleton University. Petrography was done using a binocular transmitted light microscope at Carleton University.

B) X-ray Fluorescence (XRF) and Inductively Coupled Plasma Mass Spectrometry (ICP-MS)

Samples were clean crushed using a chipmunk jaw crusher and then ground in an agate ring mill to a fine powder. Before crushing and powdering both machines were pre contaminated with small amounts of the sample being crushed. Larger volumes of powder than necessary were produced and then mixed in order to obtain a representative sample. The powders were then split into three parts, one for X-Ray Fluorescence analysis at the University of Ottawa, one for Rare Earth Element analysis (ICP-MS) at the Ontario Geological Survey Geochemical Laboratory in Sudbury and one for isotopic analysis at Carleton University.

XRF was carried out at the University of Ottawa by a fused disc technique in order to determine the major element concentrations of SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , P_2O_5 , MgO , NaO , K_2O and CaO . A 3:1 flux to sample ratio was heated to approximately 1200°C in a porcelain crucible and the liquid produced was then poured into a platinum disc-shaped mould and then cooled. The product after cooling is an approximately 35mm, homogenous glass disc that is ready for analysis. The data was collected using an automated computer system and the data was reduced using an in-house written program in a spreadsheet.

Rare earth element concentrations as well as Th, U, Hf and Ta concentrations were determined using ICP-MS in Sudbury at the Ontario Geological Survey, by an acid dissolution technique using HF. The resulting residue is then dissolved in nitric acid, dried down and then diluted to the concentration level it will run at prior to taking it to the mass spectrometer.

C) Isotope Geochemistry and Mass Spectrometry Methods

Procedural steps for Sr, Nd, and Pb isotope analyses included crushing, acid washing, dissolution, column chromatography, filament loading, and mass spectrometry. All chemistry and mass spectrometry was performed at the Isotope Geochemistry and Geochronology Research Center (IGGRC) at Carleton University, Ottawa, Canada. The mass spectrometer used was a ThermoFinnigan Triton TI, operated in static mode. The chemical procedures, originally published by Cousens (1996), are outlined below.

Samples were clean crushed using a chipmunk jaw crusher and then ground in an agate ring mill to a fine powder. Before crushing and powdering both machines were pre contaminated with small amounts of the sample being crushed. Larger volumes of powder than necessary were produced and then mixed in order to obtain a representative sample. The powders were then split into three parts, one for X-Ray Diffraction analysis at the University of Ottawa, one for Rare Earth Element analysis (ICP-MS) at the Ontario Geological Survey Geochemical Laboratory in Sudbury and one for Sm-Nd isotopic analysis at Carleton University.

200 mg were weighed out for Sr, Pb and Nd chemistry and were ready for acid washing.

Acid Washing:

The washing procedure used 200 mg of sample that was then washed in 1.5N HCl in a room-temperature ultrasonic bath (with the water level below the cap) for agitation

(rather than on a hot plate). The samples were then rinsed with ultrapure H₂O and ready for dissolution.

Dissolution: Rock powders were dissolved in three steps:

(1) 50% HF + 16N HNO₃, (2) 7N HNO₃, and (3) 6N HCl. Following complete dissolution, the dried residue from the dissolved powder was taken up in 1 N HBr.

Pb was separated using Bio-Rad 10-mL polyethylene columns and Dowex AGI-8X anion resin, using 1N HBr to elute other elements and 6N HCl to elute Pb. Samples were centrifuged prior to loading. The captured HBr solution (containing the rare earth elements) was dried and the residue taken up in 2.5N HCl.

Separation: Ion-exchange column chromatography was used to isolate the various elements of interest. The dissolved sample was centrifuged and loaded into 14-mL Bio-Rad borosilicate glass chromatography columns with 3 mL of Dowex 50-X8 cation resin. Sr was eluted with 2.5N HCl.

The rare earth elements (REE) were separated with the Sr columns (as above), but using 6N HCl. The REE fraction, dried and dissolved in 0.26N HCl, was loaded into 14 mL Bio-Rad borosilicate glass chromatographic columns containing a 2 cm-high bed of Teflon powder coated with HDEHP [di(2-ethylhexyl) orthophosphoric acid, Richard et al. 1976. Nd was then eluted using 0.26N HCl.

D) Oxygen isotopic analysis was completed at Queen's University and was extracted from samples by the fluorination method of (Clayton and Mayeda, 1963). ¹⁸O/¹⁶O ratios were measured using a Finnigan Mat 252 Isotope Ratio Mass Spectrometer.