Pliocene to late Pleistocene magmatism in the Aurora
Volcanic Field, Nevada and California, USA: A
petrographic, geochemical, and isotopic study

by

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Abstract

The Pliocene-Pleistocene Aurora Volcanic Field (AVF) covers 325 km², stretches across the California-Nevada border north of Mono Lake, and overlaps the Miocene Bodie Hills Volcanic Field (BHVF). It lies within the northwest-striking Walker Lane shear zone at the western margin of the Basin and Range province. Although erupted in a post-subduction setting, normalized incompatible element patterns show a continental arc-like signature, similar to those of other late Cenozoic volcanic fields within the Walker Lane, interpreted to reflect a metasomatized lithospheric mantle source.

Petrographic features suggest syn-magmatic open system processes. While trace element compositions show few signs of these processes, the range in isotopic ratios confirm them. The BHVF and the AVF rocks have similar Pb isotope ratios, however, the AVF rocks have higher ⁸⁷Sr/⁸⁶Sr, lower ¹⁴³Nd/¹⁴⁴Nd, and stronger correlations between ⁸⁷Sr/⁸⁶Sr, silica, and δ¹⁸O, indicating that crustal assimilation was a more influential process in the AVF than in the BHVF.

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List of Abbreviations

ACA	Ancestral Cascades Arc
AFC	Assimilation and Fractional Crystallization
AVF	Aurora Volcanic Field
BHVF	Bodie Hills Volcanic Field
BSE	Bulk Silica Earth
DMM	Depleted MORB Mantle, reservoir
EMI	Enriched Mantle I, reservoir
EMII	Enriched Mantle II, reservoir
HFSE	High Field Strength Elements: elements with small, highly charged cations (i.e. REE, Sc, Y, Th, Hf, Zr, Ti, Nb, Ta, P; Pearce, 1983)
HIMU	High μ (μ = present-day $^{238}\text{U}/^{204}\text{Pb}$ value), reservoir
HREE	Heavy Rare Earth Elements: atomic numbers 68 to 71 (i.e. Er to Lu)
ICP-AES	Inductively Coupled Plasma-Atomic Emission Spectrometry
ICP-MS	Inductively Coupled Plasma-Mass Spectrometry
LILE	Large Ion Lithophile Elements: elements with large cations of low charge (i.e. Cs, Rb, K, Ba, Sr, Eu^{2+} , Pb^{2+} ; Rollinson, 1993)
LOI	Loss On Ignition
LREE	Light Rare Earth Elements: REE with atomic number 57 to 61 (i.e. La to Pm)
MORB	Mid-Ocean Ridge Basalt, reservoir
MREE	Middle Rare Earth Elements: REE with atomic numbers 62 to 67 (i.e. Sm to Ho)
NHRL	Northern Hemisphere Reference Line

OGS	Ontario Geological Survey
REE	Rare Earth Elements: a series of metals with atomic numbers 57 to 71 (i.e. La to Lu). Also known as the lanthanide series.
SNB	Sierra Nevada Batholith
TIMS	Thermal Ionization Mass Spectrometer
TTVF	Tahoe-Truckee Volcanic Field
USGS	U.S. Geological Survey
VSMOW	Vienna Standard Mean Ocean Water
WDXRF	Wavelength Dispersive X-Ray Fluorescence spectrometry
WGB	Western Great Basin
XRF	X-Ray Fluorescence spectrometry

Aurora Volcanic Field map unit abbreviations

Qaca	Trachyandesite of Aurora Crater
Qacc	Trachyandesite of Cow Camp Creek
Qmh	Trachyandesite of Mt. Hicks
Qmsa	Trachyandesite of Mud Spring
Qta1	Trachyandesite flow 1
Qta2	Trachyandesite flow 2
QTaap	Trachyandesite South of Aurora Peak
QThch	Hornblende trachydacite of Cedar Hill
QTopb	Olivine-pyroxene basaltic trachyandesite
Tach	Trachyandesite of Cedar Hill

Tba	Basaltic trachyandesite of Beauty Peak
Tbch	Trachybasalt of Cedar Hill
Tdap	Trachydacite of Aurora Peak
Thd	Hornblende trachydacite
Tmlp	Basaltic andesite of Locomotive Point
Tmr	Martinez Hill rhyolite
Tpch	Pyroxene trachyandesite Cedar Hill
Trsp	Spring Peak rhyolite

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1 Chapter: Introduction

The Aurora Volcanic Field (AVF) is Pliocene to late Pleistocene (3.9–0.1 Ma), covers ~325 km², and extends across the California-Nevada border north of Mono Lake (Figure 1.1; Gilbert et al., 1968; Al-Rawi, 1969; Lange et al., 1993; Lange and Carmichael, 1996). The AVF consists primarily of monogenetic lava and cinder cones (e.g., Aurora Crater), valley-filling lava flows (e.g., the trachyandesite of Mud Spring), shield volcanoes (e.g., Mount Hicks and Beauty Peak), and trachydacitic to rhyolitic domes (e.g., Cedar Hill, Aurora Peak, and Spring Peak), yielding ~25 km³ of volcanic material with broadly bimodal compositions (Lange and Carmichael, 1996; Reheis et al., 2002; John et al., 2015). Several well-preserved volcanic features include Aurora Crater, Beauty Peak, and the Mud Spring lava flow, which can be easily delineated from aerial photography and satellite imagery, while other volcanic products are overlapping and faulted, thereby obscuring original stratigraphic relations (e.g., the four Cedar Hill lava domes/flows; Figure 1.2).

The AVF partially overlaps with the east and southeast margins of the Miocene Bodie Hills Volcanic Field (BHVF; light grey field in Figure 1.2). Both volcanic fields overlie pre-Tertiary basement composed of (1) Lower Paleozoic hornfelsed argillite, sandstone, chert, and pebble conglomerate; (2) Triassic (?) metamorphosed sandstone, siltstone, chert, tuff, and pillow basalt; (3) Mesozoic (Jurassic?) meta-andesite, metatuff, and metavolcaniclastic rocks; and (4) Late Cretaceous (98–83 Ma; John et al., 2015) granitic rocks that are part of the Sierra Nevada batholith (Chesterman and Gray, 1975; Stewart et al., 1982; Robinson and Kistler, 1986; John et al., 2015). The granitic rocks are exposed

at the margins of the Bodie Hills and Aurora volcanic fields, the metavolcanic rocks are found in the northern and eastern parts, and the metasedimentary rocks crop out in the southwestern and northern parts (dark grey fields in Figure 1.2; John et al., 2015).

The AVF lies within a complex tectonic setting located at the western margin of the Basin and Range physiographic province known as the Great Basin and also lies within the Walker Lane shear zone at the northwest limit of the Mina deflection (Figure 1.3). The Basin and Range province is a vast arid territory renowned for its regularly spaced, north-south trending mountain ranges, separated by flat desert basins (Eaton, 1982). The Great Basin is the most tectonically active subsection of the Basin and Range Province; its margins are especially active having high seismicity, abundant fault scarps with young displacements, and active thermal springs (Eaton, 1982). The Walker Lane is ~80 km wide, northwest-striking shear zone at the western margin of the Great Basin, and marks the eastern boundary of the Sierra Nevada microplate. It contains a complex fault system that accommodates ~20-25% of the dextral transtensional shear between the Pacific and North American plates (Bennett et al., 1999; Thatcher et al., 1999; Dixon et al., 2000; Oldow, 2000; 2003; Faulds and Henry, 2008; John et al., 2012; Busby, 2013). The Walker Lane is dominated by array of discontinuous northwest-striking right-lateral faults that are interrupted by zones of northeast-striking normal faults and east- to east-northeast-striking left-lateral faults (Stewart, 1980; Oldow, 1992; Wesnousky, 2005). A belt of east-striking left-lateral faults transition to northeast-striking normal faults, and accommodate a 60-km right step in the Walker Lane; this array of faults is known as the Mina deflection (also known as the Excelsior-Coaldale block; Stewart, 1988; Oldow,

1992). The Mina deflection coincides with a major bend in pre-Cenozoic structures and may reflect the east-northeast-trending Proterozoic continental margin marked by the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ isopleth (see section 1.2.1; Figure 1.3; Kistler and Peterman, 1973, 1978; Farmer and dePaolo, 1983; Stewart, 1988; Oldow et al., 1989; Tosdal et al., 2000; Faulds and Henry, 2008). The Walker Lane shear zone continues to the south of the Garlock fault, where it is more commonly known as the eastern California shear zone (Figure 1.3).

1.1 Overview of work completed on the Aurora Volcanic Field

1.1.1 Previous work

Seventeen of the twenty-two AVF units (Qmsa, Qaca, Qta2, Qa, Qmh, Qacc, Tmh, Tdap, Tbch, QTaap, Tpch, Tba, QThch, Thd, Tach, Trsp, Tmlp; see the Aurora Volcanic Field map unit abbreviations within the List of Abbreviations in the front matter p. xiii) have been dated using ^{40}Ar - ^{39}Ar and K-Ar methods (^{40}Ar - ^{39}Ar : groundmass, whole rock, and plagioclase, K-Ar: groundmass, whole rock, biotite, hornblende, plagioclase, and sanadine) (Gilbert et al. 1968; Silberman and McKee, 1972; Silberman and Chesterman, 1972; Lange et al., 1993; Reheis and others, 2002; John et al., 2015). Fleck et al. (2015) compiled all geochronological analyses, and calculated a preferred age for each of the AVF units in this study based on a calculated an average age and relative age relations to define a single age for each unit.

Whole rock major and trace element data, and microprobe work (analyses on cores and rims of olivine, plagioclase, orthopyroxene, clinopyroxene, hornblende, biotite, and Cr-

spinel) were completed in order to calculate the temperature, water content, and oxygen fugacity of selected basalt to dacite composition AVF units (Lange et al., 1993; Lange and Carmichael, 1996). Lange and Carmichael (1996) used the microprobe data from andesites containing both orthopyroxene and augite phenocrysts (i.e., the trachyandesite of Aurora Crater, Qcac, the trachyandesite of Mud Spring, Qmsa, and the pyroxene trachyandesite Cedar Hill, Tpch) with the two-pyroxene thermometer of Lindsley and Frost (QUILF; 1992). Calculated eruption temperatures of these units were ~1040–1080°C. Pre-eruptive water contents were both high and variable, ranging from 1.5 wt. % in the trachyandesite of Cedar Hill (Tach) to ~6 wt. % in the trachydacite of Aurora Peak (Tdap). Lange and Carmichael (1996) determined this water content using the exchange of Ca and Na between plagioclase and liquid (K_D^{Ca-Na}). Calculated oxygen fugacities fall within –0.4 and 12.4 log units of the Ni-NiO buffer (Lange and Carmichael 1996).

Gravity and airborne magnetic surveys were conducted over the Bodie Hills including the BHVF and the AVF (John et al., 2012). A new geologic map of the BHVF and the AVF was compiled from existing data, previous maps, and new fieldwork (John et al., 2015).

1.1.2 Geological problems

The AVF lavas have subduction-like geochemical signature similar to the BHVF rocks that they overlie; both volcanic fields have an enrichment of large ion lithophile elements (LILE) and depletion of high field strength elements (HFSE) relative to the primitive mantle (Lange and Carmichael, 1996; John et al., 2012). However, this geochemical signature is normally characteristic of subduction-related continental margin arcs, and the

AVF lavas erupted well after the northward migration of the Mendocino triple junction point, which marked the termination of arc-related volcanic activity to the east of the San Andreas transform fault (Atwater 1970, 1973; Atwater and Stock, 1998). Additionally, John et al. (2012) showed that the BHVF arc magmas had much higher eruption rates and larger inferred magmatic reservoirs than the AVF magmas, indicating very different processes formed each volcanic field. The AVF may also be compared to other Pliocene-Pleistocene volcanic fields within the western Great Basin; however, the AVF tends to be more evolved compositionally (i.e., rocks almost exclusively with compositions >50% silica), masking its initial mantle source signature and subsequent differentiation processes. While there has been extensive geological and geophysical work done on the Miocene BHVF, particularly on the mineral deposits in the Bodie, Aurora, and Masonic mining districts, and the Pliocene-Pleistocene Long Valley and Big Pine volcanic fields, only limited geological work has been published on the AVF.

1.1.3 Current study goals

The goals of this study are to (1) compare the geochemistry of the post-subduction AVF with the subduction-related BHVF and with the extension-related Western Great Basin volcanic fields (e.g., Long Valley, Big Pine, Kings River), (2) constrain the composition magma source(s) and magma evolution for the AVF, and (3) determine the magmatic and tectonic processes responsible for development of the AVF.

1.2 Regional tectonic processes

1.2.1 Formation of the Sierra Nevada Batholith and the $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ isopleth

The Farallon plate and its remnants (e.g., Juan de Fuca and Cocos plates) have been obliquely subducting under the west margin of the continental North American plate since at least the early Mesozoic era. The longstanding subduction of the west margin of North America created a series of superimposed plutonic belts in the Triassic, Jurassic, and Cretaceous periods (ca. 240 Ma to ca. 80 Ma), each having slight differences in geographic trends (Hietanen, 1973; Lanphere and Reed, 1973; Hay, 1976; Kistler, 1990; Armstrong and Ward, 1991). These plutonic belts form a large composite body known as the Sierra Nevada Batholith (SNB). The SNB is a far from homogenous regional basement rock, consisting predominately of Mesozoic granitic to granodioritic plutons, but also including diorite and tonolite mafic enclaves (Barbarin et al., 1989) and metamorphic roof pendants.

The SNB retains the characteristics of a major tectonic boundary. The $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ isopleth of 0.706 (i.e., the “0.706 line”; Figure 1.3) is interpreted to represent the boundary between young island arc terrains (upper Paleozoic oceanic crust and mantle) accreted during the Jurassic period and Precambrian-aged in the western Sierra Nevada with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} < 0.706$, and crystalline continental basement in the eastern Sierra with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} > 0.706$ (Kistler and Peterman, 1973). The 0.706 line is based on the initial isotopic-Sr values of the Mesozoic plutonic rocks from the mid- to upper-crust (Kistler and Peterman, 1978). The rarely-preserved Mesozoic mafic volcanic rocks have similar isotopic ratios and distributions. Having both volcanic and plutonic reflect this boundary

implies that it separates lithospheric provinces as well as crustal provinces from one another (Leeman, 1982; Kistler, 1990; Kempton et al., 1991). This boundary is also reflected in recent mafic alkaline volcanic rocks (e.g., Ormerod et al., 1988). Aside from geochemical evidence, seismic evidence also places the edge of the Proterozoic North America craton near the 0.706 line (Saleeby et al., 1984).

1.2.2 Creation of the slab window and San Andreas transform fault system

The East Pacific Rise is the site of a divergent tectonic plate boundary (i.e., a spreading center) and an axial zone of plate growth. Once the subduction rate of the Farallon plate exceeded the half-spreading rate of the East Pacific Rise during the Mesozoic, the ridge began to gradually approach the North American continent. Around 28-30 Ma, the East Pacific Rise intersected with the trench along the continental margin, bringing the Pacific and the North American plates into direct contact (Figure 1.4; Atwater, 1970). This marks the first contact between the two plates and initiation of 1) the ridge subduction and 2) the right-lateral transform faulting along the Pacific-American plate boundary.

At mid-ocean ridges like the East Pacific Rise, convection in the mantle allows hot mantle to rise beneath the spreading center, which causes decompression partial melting and generates new basaltic magma. The basaltic magma rises to the earth's surface, where it cools and crystallizes at trailing plate edges of diverging plates, creating new oceanic crust (Parfitt and Wilson, 2008). At ridge-trench intersections, one or both the diverging plates will subduct into asthenosphere. Basaltic magma may still be produced between the diverging plates; however, as the trailing plate edges sink into the

asthenosphere, they no longer act as a cold barrier where the basaltic magma can cool and crystallize and no new crust will be created (Thorkelson, 1994). Without new plate growth, a gap between the diverging plates widens, creating a slab-free zone or a slab window (Figure 1.4; Figure 1.5; Dickinson and Snyder, 1979).

The Pacific-American transform fault (i.e., the San Andreas transform fault) gradually lengthened with the simultaneous northward and southward migration of two plate margin triple junctions, known as the Mendocino and Rivera, respectively (points labeled “x” and “y” in Figure 1.5; Dickinson and Snyder, 1979). As each triple junction point migrated in opposite directions, the slab window grew, progressively terminating arc volcanic activity to the east of the transform fault (Figure 1.5). The southern edge of the subducting Farallon/Juan de Fuca plate was interpolated by Atwater and Stock (1998; Figure 1.6). Based on this interpolation, subduction at the latitude of the BHVF and AVF is inferred to have ended around 10 Ma when the Pacific–Juan de Fuca–North American triple junction (i.e., the Mendocino triple junction) migrated north of this area (Figure 1.6). This marks a change the tectonic setting in the region; plate convergence is replaced by transform faulting at the west margin of the continental North American plate ending and subduction to the south of the Juan de Fuca plate margin (Atwater and Stock, 1998; Busby et al., 2007).

1.2.3 The Walker Lane fault system

Tectonic activity in the Walker Lane shear zone was initiated ~13-12 Ma (Faulds and Henry, 2008; Busby, 2013). It coincides with an increase in relative motion between the

Pacific and North American plates (33 mm/yr to 52 mm/yr) and the ~1000 km southward jump in the Rivera triple junction, causing both an eastward shift of the southern section of the San Andreas fault system inland and east to east-northeast extension, which ultimately led to opening of the Gulf of California (Figure 1.3; Atwater and Stock, 1998; Stock and Hodges, 1989; Oskin and Stock, 2003; Fletcher et al., 2007). This eastward shift in the southern San Andreas fault system generated a large restraining bend in the fault system, termed the Big Bend, which induced a ~60 km right slip and associated deformation, currently known as the eastern Walker Lane (i.e., the Las Vegas shear zone; Deibert, 1989; Faulds and Henry, 2008).

Between 11 and 6 Ma the Mendocino triple junction point changed direction from an azimuth of N60°W to N37°W (Atwater and Stock, 1998), reflecting a change in relative motion of the Pacific and the North American plates roughly coincided with a change from east-west to a northwest motion and a ~10° clockwise rotation of the Sierra Nevada block relative the Colorado Plateau (Wernicke and Snow, 1998; Atwater and Stock, 1998; McQuarrie and Wernicke, 2005). The right-lateral shear in the Walker Lane shifted westward to the western Great Basin to align with the southern section of the San Andreas transform (i.e., the Gulf of California), and to parallel and accommodate the plate motion of the Sierra Nevada block (Faulds and Henry, 2008). Transtension in the Walker Lane propagated northwestward into the northern Walker Lane (NW Nevada – NE California) in unison with northward migration of the Mendocino triple junction point (Faulds and Henry, 2008). Currently, the Walker Lane terminates at the southern end of

the modern South Cascade arc near Lassen Peak, roughly at the same latitude the Mendocino triple junction point (Figure 1.3).

1.2.4 Extension within the western Great Basin

The Basin and Range province experienced a major and widespread period of extension that began abruptly at ca. 17–16 Ma, creating westward-propagating extensional faulting and NNE trending mountain ranges and intervening basins that currently typify the north and center of the physiographic province (Stewart, 1978; Eaton, 1982; Surpless et al., 2002; Camp et al., 2015). This extensional event was catalyzed by the Yellowstone mantle plume, which ascended through a propagating fracture in the Juan de Fuca subducted plate (Camp et al., 2015). After 12–10 Ma, the extension continued at lower magnitude, focused on the western, eastern, and northernmost margins of the Basin and Range (Camp et al., 2015).

Other contributing factors generating extension in the western Great Basin region are 1) the mid-Cenozoic extension is due to rollback of the subducting Farallon plate (Dickinson and Snyder, 1979; Dickinson, 2006), 2) the low-magnitude extension resulting from the jump in the Riviera triple junction and oblique Pacific-North American plate motion, which produced a series of large, shallow basins in the Walker Lane (Stewart, 1988, 1992; Dilles and Gans, 1995; Stockli et al., 2003; Henry and Perkins, 2001), and 3) the extension in response to the slab window development, which created faults that may have served as conduits for later magmatic activity (section 1.2.2; Dickinson and Snyder 1979; John et al., 2012).

1.3 Local geology

1.3.1 Bodie Hills Volcanic Field

The Miocene Bodie Hills Volcanic Field (BHVF) is part of the southern segment of the Ancestral Cascades Arc (ACA). The BHVF covers more than 1200 km², straddles the California-Nevada border north of Mono Lake, and is unconformably overlain by the Pliocene-Pleistocene AVF (John et al., 2015). The BHVF formed by episodic magmatism between about 15.0 Ma and 5.5 Ma, producing 21 distinct volcanic eruptive centers and several smaller centers, underlain by middle- to shallow-crustal-level magma reservoirs (John et al., 2012; John et al., 2015). During this magmatic period, several hydrothermal systems were also active and formed extensive alteration zones and associated gold-silver mineral deposits such as the Bodie, Aurora, and Masonic mining districts (Vikre et al., 2015).

BHVF rocks have compositions that vary continuously from 50 to 78 wt. % SiO₂, although few contain <55 wt. % SiO₂. The erupted rocks are high-K calc-alkaline and oscillate in silica composition between mafic and intermediate to felsic compositions through time. They are porphyritic and commonly contain 15–35 volume percent phenocrysts of plagioclase, pyroxene, and hornblende ± biotite. The trace element geochemical signature is consistent with subduction-related continental margin volcanic arcs: high large ion lithophile elements (LILE; e.g. Ba, Pb, K, Sr) contents relative to the high field strength elements (HFSE; e.g. Ta, Nb, Zr, Hf, Ti). Radiogenic isotope ratios range from a mantle signature (low Sr and Pb isotopic ratios, high Nd isotopic ratios) to a

crustal signature (high Sr and Pb isotopic ratios, low Nd isotopic ratios). Neither age nor whole rock compositions of BHVF rocks are well correlated with isotopic variations (du Bray et al., in press).

1.3.2 Tectonic trends within the Aurora Volcanic Field

Two main structural trends are evident in the BHVF and AVF: NNW-striking faults parallel to the East Sierran frontal fault zone, and northeast-striking faults bounding the Mono Basin (Parkiser, 1970). The stress field within the Bodie Hills changed at about the same time as the Mendocino triple junction passed through the area at 9-10 Ma. This change in stress is reflected in the BHVF rocks in which units older than about 9 Ma are cut by conjugate northeast-striking left-lateral and north-striking right-lateral oblique-slip faults, whereas late Miocene BHVF rocks are cut by north- and northwest-striking dip-slip faults (John et al., 2012, 2015). The BHVF magmas may have used the normal faults as conduits to the surface (John et al., 2012).

The AVF lavas are more strongly controlled by local faulting than the BHVF (John et al., 2012). The AVF is characterized by north-northeast-elongated monogenetic volcanoes that are locally offset by a series of north-northeast-elongate stepover grabens, which are linked to the faults on the northern boundary of Mono Basin at the northwest edge of the Mina deflection (John et al., 2015). The youngest faults are north-striking normal and northeast-striking left-lateral oblique-slip faults that cut the AVF units (John et al., 2015).

Figure 1.1 Google Earth satellite image of central California and Nevada.

Indicated for reference are: the state boundary between California and Nevada (in yellow), local cities (white stars), lakes, and the study location (lime green box shows the boundaries of Figure 1.2).

The red field marks the limits of the Ancestral Cascades Arc (after, Dickinson, 1997).

Figure 1.2 Geologic map of the study field area.

This map shows the distribution of various Aurora Volcanic Field components (coloured), in relation to rocks of the Bodie Hills Volcanic Field (grey), basement rocks (dark grey) and Late Cenozoic sedimentary rocks (pale yellow). Geologic units are based on the work of Lange and Carmichael, 1996; John et al., 2015; and subsequent fieldwork and geochemical sampling. Red dots show the locations of analyzed samples. Volcanic vents are shown as red asterisks. The border between California and Nevada is shown as a black dashed line. Inset map shows the location of the map relative to California and Nevada. After John et al., 2015.

Figure 1.3 Map of the eastern margin of the Pacific plate and the western margin of the North American plate.

The Sierra Nevada block as a dark green field, the Walker Lane as an orange field (Mina Deflection is labeled as “M”), the 706 line as a white line, faults as black lines, the San Andreas fault as a purple line, and the modern Cascades Arc as red fields. After Faults and Henry, 2008.

Figure 1.4 A simplified block diagram showing a mid-oceanic ridge-trench intersection and creation of a slab-free zone or “slab window” beneath the overriding plate.

This figure shows diverging plates are separated by a asthenosphere-filled gap as they subduct. The subducted plates inhibit flow of asthenosphere; however, the asthenosphere may freely flow up from beneath the descending oceanic plates to fill the gap created in the slab-free zone. After Thorkelson, 1994.

Figure 1.5 A model of slab window formation (stippled) under the North American plate at (a) 20 Ma, (b) 10Ma, and (c) 5Ma.

Notice that slab-free region expands as the triple junctions, labeled x and y, migrate simultaneously to the north and south. The southern edge of the Cascades arc corresponds to the northern edge of the slab window. After Dickinson and Snyder, 1979.

Figure 1.6 Google Earth satellite image of central California and Nevada showing the location of the southern edge of the Juan de Fuca plate from 20 Ma to the present (subhorizontal grey lines; after Atwater and Stock, 1998).

The grey lines mark the incremental extinction of the magmatic arc to the south of the Juan de Fuca plate margin. Note that the 10 Ma chron goes through the BHVF and AVF corresponding to the cessation of subduction in the study area. For orientation purposes only, in their present-day locations are: the state boundary between California and Nevada (yellow line), local cities (white stars), and the study area (green rectangle is the boundaries of Figure 1.2).

2 Chapter: Methods

2.1 Summary of methods

Thirty-three samples of AVF lavas were collected in the field, supplemented by twenty-five additional AVF samples (covered thin sections and powdered samples) provided by David John and Ed du Bray of the United States Geological Survey (USGS). Seven basement rock samples were also provided by the USGS to represent local, potential crustal contaminants. All samples were collected based on freshness (e.g., the least weathered or altered). These samples (except for two that were collected for petrography only) were analyzed by X-ray fluorescence spectrometry (XRF) and inductively coupled plasma-mass spectrometry (ICP-MS) for major and trace element composition. All of the AVF rocks were sent to the same laboratory for whole rock major and trace element geochemistry; therefore, interlaboratory bias is not a significant issue for the AVF data suite. The 63 samples were analyzed by thermal ionization mass spectrometry (TIMS) for their strontium and neodymium isotopic compositions. Thirty-one samples were analyzed for lead isotopic compositions using TIMS, and seventeen samples for oxygen isotopic compositions using a dual inlet gas mass spectrometer (see Table 1).

Table 1 Summary of analyses completed in this study

Sample Number	Unit Name Complete	Shorthand	Age ¹ (Ma)	Analyses Completed for this Project				
				Thin Section	Whole Rock Geochemistry	Pb	Sr	Nd
13-AU-01	(Rancheria plug, Basaltic andesite diatreme) ²	(Trwai) ²	11.70	✓	✓		✓	✓
13-AU-02	(Mafic clast from Rancheria plug) ²	(Trwai) ²	11.70	✓	✓		✓	✓
13-AU-03	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓		✓	✓
13-AU-04	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓		✓	✓
13-AU-05	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓		✓	✓
13-AU-06	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-07	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-08	Trachyandesite flow 2	Qta2	1.32	✓	✓		✓	✓
13-AU-09	Trachyandesite of Aurora Crater	Qaca	0.36	✓	✓		✓	✓
13-AU-10	Trachyandesite of Aurora Crater	Qaca	0.36	✓	✓		✓	✓
13-AU-11	Trachyandesite South of Aurora Peak	QTaap	2.45	✓	✓		✓	✓
13-AU-12	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-13	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-14	Trachyandesite of Cedar Hill	Tach	3.70	✓				
13-AU-15	(Mount Biedeman, Hbld-bt-plag dacite) ²	(Tamb) ²	9.27	✓	✓		✓	✓
13-AU-16	(Bodie Hills Rhyolite) ²	(Trbh) ²	9.79					
13-AU-17	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-18	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓		✓	✓
13-AU-19	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓		✓	✓
13-AU-20	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓
13-AU-21	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓		✓	✓

¹ Ages are based on "preferred age" from duBray et al., 2015.

² Units in round brackets, (), are from BHVF.

Sample Number	Unit Name Complete	Shorthand	Age ¹ (Ma)	Analyses Completed for this Project				
				Thin Section	Whole Rock Geochemistry	Pb	Sr	Nd
13-AU-22	Pyroxene trachyandesite Cedar Hill	Tpch	2.80	✓	✓	✓	✓	✓
13-AU-23	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓	✓	✓	✓
13-AU-24	Trachyandesite of Cedar Hill	Tach	3.70	✓	✓	✓	✓	✓
13-AU-25	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓	✓	✓	✓
13-AU-26	Trachyandesite of Mt. Hicks	Qmh	1.60	✓	✓	✓	✓	✓
13-AU-27	Trachydacite of Aurora Peak	Tdap	2.70	✓	✓	✓	✓	✓
13-AU-28	Trachydacite of Aurora Peak	Tdap	2.70	✓	✓	✓	✓	✓
13-AU-29	Trachyandesite of Mt. Hicks	Qmh	1.60	✓	✓	✓	✓	✓
13-AU-30	Trachyandesite of Aurora Crater	Qaca	0.36	✓	✓	✓	✓	✓
13-AU-31	Trachyandesite of Aurora Crater	Qaca	0.36	✓	✓	✓	✓	✓
13-AU-32	Olivine-pyroxene basaltic trachyandesite	QTopb	4.00	✓	✓	✓	✓	✓
13-AU-33	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	✓	✓	✓	✓
Thin section and powdered samples courtesy of the USGS								
203176	Hornblende trachydacite of Cedar Hill	QThch	2.95	(✓) ³	(✓) ³	✓	✓	✓
203181	Pyroxene trachyandesite Cedar Hill	Tpch	2.80	(✓) ³	(✓) ³	✓	✓	✓
203184	Basaltic andesite of Locomotive Point	Tmlp	3.90	(✓) ³	(✓) ³	✓	✓	✓
203195	Basaltic andesite of Locomotive Point	Tmlp	3.90	(✓) ³	(✓) ³	✓	✓	✓
203204	(Trachyandesite of Masonic) ²	(Tma) ²	14.70	(✓) ³	(✓) ³	✓	✓	✓
203230	Trachyandesite of Cow Camp Creek	Qacc	2.00	(✓) ³	(✓) ³	✓	✓	✓
203232	Trachyandesite of Cow Camp Creek	Qacc	2.00	(✓) ³	(✓) ³	✓	✓	✓
203246	Pyroxene trachyandesite Cedar Hill	Tpch	2.80	(✓) ³	(✓) ³	✓	✓	✓
203247	Pyroxene trachyandesite Cedar Hill	Tpch	2.80	(✓) ³	(✓) ³	✓	✓	✓
203326	Basaltic trachyandesite of Beauty Peak	Tba	2.90	(✓) ³	(✓) ³	✓	✓	✓
203354	Pyroxene trachyandesite Cedar Hill	Tpch	2.80	(✓) ³	(✓) ³	✓	✓	✓

³ Check marks with rounded brackets, (✓), indicate samples or analyses were provided by the USGS.

Sample Number	Unit Name Complete	Shorthand	Age ¹ (Ma)	Analyses Completed for this Project				
				Thin Section	Whole Rock Geochemistry	Isotope Geochemistry		
						Pb	Sr	Nd
203382	Spring Peak rhyolite	Trsp	3.70	(✓) ³	(✓) ³	✓	✓	✓
203393	Trachyandesite of Aurora Crater	Qaca	0.36	(✓) ³	(✓) ³		✓	✓
203394	Trachyandesite of Aurora Crater	Qaca	0.36	(✓) ³	(✓) ³	✓	✓	✓
203395	Trachyandesite of Mud Spring	Qmsa	0.28	(✓) ³	(✓) ³	✓	✓	✓
203398	Trachyandesite flow 1	Qta1	1.45	(✓) ³	(✓) ³	✓	✓	✓
203402	Trachyandesite of Cow Camp Creek	Qacc	2.00	(✓) ³	(✓) ³	✓	✓	✓
203443	Trachyandesite flow 1	Qat1	1.45	(✓) ³	(✓) ³	✓	✓	✓
01-BA-1	Hornblende trachydacite	Thd	3.50	(✓) ³	(✓) ³	✓	✓	✓
08-BA-3	Trachyandesite of Aurora Crater	Qaca	0.36	(✓) ³	(✓) ³	✓	✓	✓
08-BA-6	Martinez Hill rhyolite	Tmr	2.60	(✓) ³	(✓) ³	✓	✓	✓
08-BA-7	Trachyandesite flow 2	Qta2	1.32	(✓) ³	(✓) ³	✓	✓	✓
11-BA-46	Spring Peak rhyolite	Trsp	3.70	(✓) ³	(✓) ³	✓	✓	✓
12-BA-32	Trachybasalt of Cedar Hill	Tbch	2.60	(✓) ³	(✓) ³	✓	✓	✓
12-BA-33	Hornblende trachydacite of Cedar Hill	QThch	2.95	(✓) ³	(✓) ³	✓	✓	✓
203449	Trachydacite of Aurora Peak	Tdap	2.70	✓	(✓) ³			
203450	Trachyandesite of Mt. Hicks	Qmh	1.60	✓	(✓) ³			
13-BA-04	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓	(✓) ³			
13-BA-10	Hornblende trachydacite of Cedar Hill	QThch	2.95	✓				
Basement rock samples courtesy of the USGS								
203350	Granite of Murphy Creek		83	✓	✓	✓	✓	✓
203377	Granodiorite of Masonic Mountain		95	✓	✓	✓	✓	✓
203387	Monzogranite of Rattlesnake Gulch		95	✓	✓	✓	✓	✓
BCP-13-1	Siliceous shale		Paleozoic	✓	✓	✓	✓	✓
CC-13-1	Black argillite		Paleozoic	✓	✓	✓	✓	✓
HOM-13-1	Meta-andesite, amphibole grade		Jurassic	✓	✓	✓	✓	✓
MAS13-2	Metadacite tuff (Masonic Mountain)		Jurassic	✓	✓	✓	✓	✓

2.2 Fieldwork

Fieldwork was conducted to familiarize the author with the local geology and to collect samples to supplement those provided by the USGS. Samples were collected from AVF outcrops (1) to obtain geochemical data from all of the major map units (eruptive centers), (2) to better define distinctions between BHVF and AVF volcanic centers and (3) to refine the Bodie Hills geologic map (e.g., Cedar Hill lava flows were poorly characterized in previous research; John and du Bray per. comm. 2013).

Fieldwork was completed from July 11 to July 16, 2013 with Prof. Brian Cousens of Carleton University, and David John of the USGS, Menlo Park, California, and Edward du Bray of the USGS, Denver, Colorado. Thirty-three samples were collected, including samples from two previously unmapped units: olivine-pyroxene basaltic (QTopb) trachyandesite and, trachyandesite south of Aurora Peak (QTaap). Two samples (13-AU-01 and 13-AU-02) were collected from the Rancheria diatreme, part of the BHVF, to compare with the most mafic units of AVF. All samples were selected based on their freshness. Due to the dry climate, weathering and alteration of the rock samples is minimal; however, all weathered surfaces of samples were removed in the field.

The field area was accessed by vehicle via the National Forest Development Road 028 from the north and via Route 167 CA (Pole Line Road) from the south. Sample locations were recorded with a handheld Magellan Explorer 610 GPS instrument, using geodetic coordinates in the North American Datum 1927 (NAD 27) and UTM grid coordinates

(Zone 11 S). The sample locations are indicated on Figure 1.2 and their coordinates are provided in Appendix A.

2.3 Petrography

Thirty-two samples collected in the field were prepared for covered thin sections (sample 13-AU-16 is a mingled-magma sample from BHVF and does not have a corresponding thin section). Covered thin sections of 7 basement rock samples provided from the USGS were also prepared. The samples were cut to 3 cm x 4 cm sized pucks using a Highland Park trim saw with a continuous rim embedded-diamond blade. The pucks were smoothed and leveled down using a Highland Park Flat Lapping Machine with 320-grit sand, and then finished with 600-grit sand. The thin sections were examined under plane- and cross-polarized light to facilitate mineral identification and textural relations (Appendix B).

2.4 Sample processing for geochemical analysis

Thirty-three AVF samples were collected as fist sized blocks. Two samples were not analyzed geochemically: sample 13-AU-14 is extensively weathered and was collected for thin section only (to designate the unit for mapping purposes), and sample 13-AU-16 is a BHVF unit. All other samples were trimmed of their weathered surfaces and broken into chips in the field, and sent to the USGS for whole rock major and trace element geochemical analysis. Each of the AVF samples were crushed down to <2 mm with an ASC Scientific Jaw Crusher (Model JC-300ST), then pulverized down to <200 mesh (<74 microns) in an Angstrom Ring Pulverizer (Model TE100) using an agate container.

To limit contamination, the jaw crusher and the ring pulverizer were cleaned thoroughly before use, and between each sample, and at the end of use. The jaw crusher machine and its plates were cleaned of dust and debris using an air hose before and after ~90 g of >2 mm of pure Arkansas quartz was run thru the crusher. The plates were removed, wiped down with a scrubber sponge, and then cleaned with ultra-pure grade methanol (>99.8%). The ring pulverizer's agate container, puck, ring, and O-ring were cleaned using a brush and air hose before and after ~90g of pure Arkansas Quartz (< 2 mm) was added to the agate container and run for 5 minutes in the ring pulverizer. Then agate container, the puck, the ring, and the O-ring were cleaned with ultra-pure grade methanol (>99.8%). Split powders of chemically analyzed samples were then processed for Pb, Sr, and Nd isotope analysis at Carleton University. Selected samples (Table 1) were processed for oxygen isotope work at University of Western Ontario.

Seven basement rock samples were crushed and powdered at Carleton University. Each sample was crushed in a Bico Braun Chipmunk jaw crusher and was ground to a fine-powder in a Rocklabs Standard Ring Mill batch pulveriser with an agate head. The jaw crusher and the ring mill were cleaned thoroughly before use, and between each sample, and at the end of use to limit cross-contamination of the samples. The jaw crusher was scrubbed with a steel brush, dusted off with compressed air, and wiped down with ethanol. A small amount of the following sample was run through the jaw crusher and discarded to pre-contaminate the instrument. To eliminate the caking of finer grained samples to the side of the bowl, the ring mill was cleaned by a run with quartz sand before use, after 3 or 4 samples, and at the end of use. Between samples, the ring mill was

washed with soap and water, wiped with ethanol, dried with compressed air, and pre-contaminated with the next sample. An aliquot of each sample was sent to Geoscience Laboratories in Sudbury, Ontario for whole rock major and trace element concentrations. Another aliquot was processed for Pb, Sr, and Nd isotope analysis at Carleton University.

2.4.1 Whole rock and trace element analysis at SGS Minerals (Toronto, ON)

Thirty-one samples (all 13-AU-series samples, except for 13-AU-14 and 13-AU-16) were sent to SGS Minerals (under contract for the U.S. Geological Survey, USGS) and analyzed for whole-rock geochemistry using wavelength dispersive X-ray fluorescence spectrometry (WD-XRF), inductively coupled plasma-atomic emission spectrometry (ICP-AES), and inductively coupled plasma-mass spectrometry (ICP-MS) to determine major and trace element composition.

Ten major-element oxides (SiO_2 , Al_2O_3 , CaO , MgO , Na_2O , K_2O , total iron as $\text{Fe}_2\text{O}_{3\text{T}}$, MnO , P_2O_5 , and TiO_2) and gravimetric loss of ignition, LOI, (at 925°C) were analyzed by WDXRF. Each rock sample is fused with a 50/50 lithium metaborate, lithium tetraborate flux. The resulting glass disk is placed within a wavelength dispersive X-ray spectrometer. A rhodium X-ray tube irradiates the disk, exciting the atoms in the sample to release additional X-ray photons, which have wavelengths characteristic of the elements in the sample. The intensity of the X-ray photons emitted by the sample is counted and used to determine the concentration. Concentrations of each element are established from pre-existing calibration curves, which are derived from international reference materials and synthetic standards, and converted into a percent oxide.

Eight major (Al, Ca, Fe, K, Mg, Mn, P, Ti) and forty-seven rare earth and trace elements (Ag, As, Ba, Be, Bi, Cd, Ce, Co, Cr, Cs, Cu, Dy, Er, Eu, Ga, Gd, Ge, Hf, Ho, In, La, Li, Lu, Mo, Nb, Nd, Ni, Pb, Pr, Rb, Sb, Sc, Sm, Sn, Sr, Ta, Tb, Th, Tl, Tm, U, V, W, Y, Yb, Zn, Zr) were determined by inductively coupled plasma-atomic emission spectrometry (ICP-AES) and inductively coupled plasma-mass spectrometry (ICP-MS). Each sample is decomposed at 450°C using a sodium peroxide sinter. The resulting cake is leached with water, and acidified with nitric acid. Tartaric acid is added, and the sample is digested. Aliquots of the sample are aspirated into the ICP-AES and the ICP-MS, and concentrations of the elements are determined.

The ICP-MS is calibrated with aqueous standards; internal standards are used to counterbalance matrix effects and instrument drift. The ICP-AES calibrated with standards of digested rock reference materials, and a series of multi-element solution standards. The precision of the data is based on duplicates (a porphyritic hornblende trachyandesite 13-AU-13 from QThch, and in house reference material DGPM-1, and GSP-QC). The international reference material SY-4 was used to monitor laboratory bias. The analyses on SY-4 were within the 95% confidence interval for all elements except for Cr and Cu (<10 ppm compared to 12 ± 1 ppm, and <5 ppm compared to 7 ± 1 ppm, respectively). The values measured for the blank were less or equal to the detection limits, except for Ni and Sb (10 ppm compared to 5 ppm, and 0.2 ppm compared to 0.1 ppm, respectively). See Appendix C.1 for analyses and associated statistics on the

duplicate samples, the in house reference material, the international reference material, and the blank sample.

2.4.2 Whole rock and trace element analysis at Geoscience Laboratories (Ontario Geological Survey, OGS, Sudbury, ON)

Seven basement rock samples were sent to the Geoscience Laboratories and analyzed for whole-rock geochemistry using X-ray fluorescence spectrometry (XRF), inductively coupled plasma mass spectrometry (ICP-MS), to determine major and trace element composition.

Loss on ignition (LOI) was determined for each sample from a two-step procedure: at 100°C under nitrogen atmosphere, then at 1000°C under oxygen atmosphere until a constant weight percent is determined. The majority of the LOI was derived from the latter process. Then each sample was fused with a borate flux to produce a glass bead. Ten major-element oxides (SiO_2 , Al_2O_3 , CaO , MgO , Na_2O , K_2O , total iron as $\text{Fe}_2\text{O}_3^{\text{T}}$, MnO , P_2O_5 , and TiO_2) were analyzed using XRF. An X-ray tube irradiates the glass bead. X-ray photons are emitted by the elements in the sample, and are counted. Concentrations of each element are established from pre-existing calibration curves and converted into a percent oxide.

Forty four trace and rare earth elements (Ba, Be, Bi, Cd, Ce, Co, Cr, Cs, Cu, Dy, Er, Eu, Ga, Gd, Hf, Ho, La, Li, Lu, Mo, Nb, Nd, Ni, Pb, Pr, Rb, Sb, Sc, Sm, Sn, Sr, Ta, Tb, Th, Ti, Tl, Tm, U, V, W, Y, Yb, Zn, Zr) were analyzed by ICP-MS. Samples prepared using

an open beaker multi-acid digestion (hydrofluoric acid, hydrochloric acid, nitric acid, and perchloric acid). The samples were diluted and run on the ICP-MS using conventional solution nebulization, and aspirated into argon plasma.

Sample calibration was carried out using a combination of synthetic multi-element solutions and certified reference materials. The precision of the data is based on duplicate samples (a fine grained siliceous shale, BPC 13-1, and a basalt that is routinely used at Carleton University as an internal standard, 10-LT-05) and OGS internal reference materials (MRB-29 and NPD-1). International reference material BHVO-2 and SY-4 were used to monitor laboratory bias. The analyses on BHVO-2 were within the 95% confidence interval for all elements except for Ho and Ti (0.95 ppm compared to 1.04 ± 0.08 ppm, and 17335.42 ppm compared to 16300 ± 400 ppm, respectively). Titanium oxide was within the 95% confidence interval for SY-4 (0.29 ppm compared to 0.287 ± 0.003 ppm), and was used for plots. The values measured for the blank were less or equal to the detection limits, except for Mo (0.05 ppm compared to 0.03 ppm). See Appendix C.2 for analyses and associated statistics on duplicate samples, the international reference materials, and the blank sample. Only one run was completed for each internal reference material, MRB-29 and NPD-1, therefore no statistical analysis was calculated.

2.4.3 Radiogenic isotope analysis at Carleton University (Ottawa, ON)

The methods for rock dissolution and element extraction are modified after Cousens (1996). Lead, strontium, and neodymium were separated by chromatography. Sr and Nd isotopic analyses were performed on all 63 samples (56 AVF and 7 basement samples),

and thirty samples were selected for lead isotope analysis (23 AVF and 7 basement samples, see Table 1). All isotopic measurements were completed on a ThermoFinnigan Triton thermal ionization mass spectrometer (TIMS). Parent-daughter ratios for initial Pb, Sr, and Nd isotope ratios were calculated using U, Th, Pb, Rb, Sr, Sm, and Nd concentrations from the ICP-AES and ICP-MS data (provided by the SGS Minerals and OGS laboratories) and the radiometric ages of each vent (see Appendix E for analyses). ϵ_{Nd}^t is calculated using a CHUR value of 0.512630 (Bouvier et al., 2008).

2.4.3.1 Whole rock dissolution procedures

Each powdered sample was placed in a 15 mL Savillex Teflon screw cap vial and acid-washed to remove weathering products with 1.5 N of hydrochloric acid (HCl), then rinsed once with ultra-pure water. The samples were then dissolved in a closed beaker system at 120°C in a three step process: first in a mixture of hydrofluoric acid (HF) and nitric acid (HNO₃), then in 7N HNO₃, and finally in 6N HCl. Select samples (see Table 1) had Pb extracted were then taken up in 1N hydrobromic acid (HBr). All other samples were taken up in 2.5N HCl for Sr and Nd extraction. Refractory mineral phases would have a minimal effect on the isotopes in the AVF field. No garnet was observed optically in previous studies nor in the geochemical signature (flat MREE to HREE slope in chondrite-normalized diagrams; see Figure 4.2). Zircon is present in the two AVF rhyolite units in trace amounts (small negative Zr anomaly in the rhyolites in the primitive mantle-normalized diagrams; Figure 4.3 b) and would have a minimal effect on their isotopes.

2.4.3.2 Lead isotope procedures

Lead was extracted by chromatography in Bio-Rad 10mL polyethylene Econo-columns with Dowex AG1-X8 anion resin with 1N HBr, 1.5 N HCl, and 6N HCl. Once Pb was extracted, 7N HNO₃ was added to each sample to break down bromides (if present) and dried down. The extracted Pb for each sample was loaded onto an outgassed Re filament using silica gel and phosphoric acid (H₃PO₄). The Pb samples are run manually at filament temperatures of 1280-1320°C (1-16 V). All mass spectrometer runs are corrected for fractionation using NIST SRM981, based on the values of Todt et al. (1984). The SRM981 Pb standard was analyzed every 12 or fewer samples. The average ratios measured (uncorrected) for the Pb standard, SRM981 are: $^{206}\text{Pb}/^{204}\text{Pb} = 16.891 \pm 0.011$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.428 \pm 0.016$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.502 \pm 0.047$ (2 σ), based on 65 runs between October 2010 and August 2014. The agreed-upon values for NBS 981 (fractionation-corrected) from Todt et al. (1996) are $^{206}\text{Pb}/^{204}\text{Pb} = 16.9356 \pm 0.0007$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4891 \pm 0.0009$, $^{208}\text{Pb}/^{204}\text{Pb} = 36.7006 \pm 0.0034$. Total procedural blanks for Pb are < 100 picograms.

2.4.3.3 Strontium isotope procedures

Strontium was extracted by chromatography in Bio-Rad borosilicate glass columns with Dowex AG50-X8 cation resin with 2.5N HCl. REE were then collected using 6N HCl, from which Nd was extracted (see section 2.3.3.4). The extracted Sr for each sample was loaded onto an outgassed Ta filament using H₃PO₄. The Sr samples are run manually at filament temperatures of 1400-1520°C (higher temperatures were due to low signal and climbing temperature during collection; more sample was dissolved for subsequent runs).

Strontium isotope ratios are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.11940$ to correct for instrument fractionation. A Sr standard was analyzed every 12 or fewer samples. The average ratio measured for the Sr standard, Eimer and Amend (E&A) SrCO_3 , is $^{86}\text{Sr}/^{88}\text{Sr} = 0.708023 \pm 35$ (2σ), based on 19 runs between August 2010 and August January 2014. The agreed-upon value for E&A is $^{86}\text{Sr}/^{88}\text{Sr} = 0.708000$. Total procedural blanks for Sr are < 250 picograms.

2.4.3.4 Neodymium isotope procedures

Neodymium was extracted by chromatography in Eichrom Ln Resin columns containing Teflon powder coated with HDEHP [di(2-ethylhexyl) orthophosphoric acid, Richard et al., 1976] with 0.26N HCl. The extracted Nd for each sample was loaded onto one side of a Re double filament assembly using H_3PO_4 . The Nd samples are run automatically at filament temperatures of 1650-1820°C (0.2-2 V). Neodymium isotope ratio is normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$. A Nd standard was analyzed once every 20 or fewer samples. The average $^{143}\text{Nd}/^{144}\text{Nd}$ measured for the internal lab Nd standard is 0.511830 ± 16 (2σ) (equivalent to La Jolla value of 0.511849) based on 48 comparative runs between January 2013 and August 2014. Total procedural blanks for Nd are < 50 picograms.

2.4.4 Oxygen isotope analysis at University of Western Ontario (London, ON)

Seventeen samples were sent to University of Western Ontario for whole-rock oxygen isotope analysis. Whole-rock powders were dried overnight, under vacuum, at 200°C. The dried samples were loaded into Ni-reaction vessels under dry nitrogen gas, and heated at 300°C for 2 hours under vacuum. Overnight, the samples were reacted with

chlorine trifluoride (ClF_3) in sealed vessels at 550°C to release oxygen gas. Oxygen was extracted from the silicates using the method of Clayton and Mayeda (1963), as modified by Borthwick and Harmon (1982) for use with ClF_3 . The oxygen was converted quantitatively to CO_2 over red-hot graphite. Carbon dioxide yields were checked to ensure complete conversion and/or absence of contamination. The oxygen-isotope composition of the CO_2 gas was analyzed using an Optima dual inlet-mass spectrometer. All isotopic results are reported in the usual δ -notation relative to Vienna Standard Mean Ocean Water (VSMOW; Potter et al., 2008; Appendix E).

Duplicates of both rhyolites (08-BA-6 and 11-BA-46) and in addition to three laboratory standards were run during the course of the analysis (see Appendix C.3 for their analysis and precision calculations). The within analytical session data shows that the laboratory standards are within $\pm 0.2\text{‰}$ of the accepted $\delta^{18}\text{O}$ value, and within 96% confidence. The reproducibility is much better for the CO_2 laboratory standard (within $\pm 0.1\text{‰}$) than the basalt and quartz silicate laboratory standards, suggesting that half of the total error arises from sample heterogeneity and the methods used to liberate the oxygen.

2.4.5 Assessment of uncertainty: Precision replicate analysis, standards, and blanks.

Precision replicate analysis was based on the standard deviation, percent relative standard deviation, percent error and percent difference on all duplicate sample runs, and standards. The standard deviation is equal to:

$$s = \sqrt{\frac{\sum_{i=1}^n (x_i - \bar{x})^2}{n-1}}, \quad (1)$$

where x_i is the i^{th} value, $\bar{x}$ is the mean of all the values, and n is the number of values.

The percent relative standard deviation is equal to:

$$\% \text{RSD} = \frac{s}{\bar{x}} \times 100\% . \quad (2)$$

The percent error is equal to:

$$\% \text{ error} = \frac{\tilde{x} - x}{x} \times 100\% , \quad (3)$$

where $\tilde{x}$ is the measured value, and x is the accepted or certified value. Finally, the percent difference is equal to:

$$\% \text{ diff} = \frac{|x_1 - x_2|}{\bar{x}} \times 100\% , \quad (4)$$

where x_1 and x_2 are measured values. Measured blanks were compared with the detection limits of the instrument, and measured international reference material compared with accepted values and their 95% confidence interval (2σ). See Appendix C for analyses and their assessment of uncertainty.

3 Chapter: Field relations and petrography

The Aurora Volcanic Field (AVF) rocks are lava flows related to small shield volcanoes, monogenetic cones, and small domes (John et al., 2012; 2015). The volcanoes are north-northeast-elongated (e.g., Aurora Crater) and volcanic units are locally offset by north-northeast elongate step-over grabens linked to east-northeast-striking left-lateral faults (Gilbert and others, 1968; John et al., 2012; 2015).

Several lava flow units are clearly delineated in aerial and satellite photographs, such as the younger units (e.g., trachyandesite of Aurora Crater, Qaca, and trachyandesite of Mud Spring, Qmsa). Lava flows are commonly blocky or platy, and primary features, such as brecciated (a'a) flow tops, oxidation textures, and/or the flow front jointing, may be preserved (Figure 3.1, Figure 3.2).

On a macroscopic scale, most units are porphyritic with plagioclase phenocrysts. Other phenocrysts may be present thus aiding rock-type identification in the field, although normalized major element analyses and the total alkali-silica (TAS) diagram boundaries of LeBas et al. (1986) were used to assign rock names. Some units span several compositions. For example, compositions of the trachyandesite of Cow Camp Creek (Qacc) range from basaltic trachyandesite to trachydacite, with the phenocryst assemblages varying accordingly. Thus, many units are more easily identified using both petrographic and geochemical characteristics. A summary of field and microscopic descriptions are presented below in alphabetical order based on the map unit abbreviation. Additional petrographic detail can be found in Appendix B. Ages in the

following sub-section are based on preferred ages of du Bray et al. (2016), where they calculated an average age, potentially modifying it to account for known relative age relations, to define a single age for each unit.

Figure 3.1 Platy flow with blocky, a' flow top of the trachyandesite of Aurora Crater (Qaca).
Hammer handle (bottom left) for scale.

Figure 3.2 Flow front jointing in the trachyandesite of Aurora Crater (Qaca).

Intersection of horizontal and vertical joints in the lava flow front. Scale card for scale.

3.1 Map units descriptions

Qaca, Trachyandesite of Aurora Crater (0.36 Ma):

Dark-grey, sparsely porphyritic trachyandesite lava flows and cinder deposits that emanated from an elliptical composite cone with central crater. Lava flows are very platy and have a'a flow tops. Flow fronts show columnar jointing with widely spaced fractures parallel to flow front and more narrowly spaced fractures perpendicular to flow front (Figure 3.2). This unit contains ~2 % (volume percent of thin section samples) phenocrysts of clinopyroxene, minor orthopyroxene, and trace plagioclase in a moderately devitrified (25-55% glass) groundmass of plagioclase, pyroxenes, opaque

oxide crystals, and is minimally vesicular (1-5%) except at the flow front (Fig. 3.2; 25-35% in samples 13-AU-30 and 13-AU-31; orthopyroxene noted in microprobe analyses of Lange and Carmichael, 1996).

Qacc, Trachyandesite of Cow Camp Creek (2.00 Ma):

Medium-grey, sparsely to moderately porphyritic trachyandesite lava flows also include basaltic trachyandesite and trachydacite compositions. A series of lava flows form the ≤ 170 m-thick unit that associated with an inferred shield volcano (John et al., 2015). The groundmass is weakly to strongly devitrified, and contains plagioclase laths. The phenocryst content ranges from 6 to 25%, and is dominated by plagioclase crystals that are commonly sieved (except in the trachydacite). Minor phenocrysts include olivine, pyroxene, and hornblende, varying according to rock composition. This unit is sparsely vesicular (0-5%).

Qmh, Trachyandesite of Mt. Hicks (1.60 Ma):

Dark-grey, nearly aphanitic, trachyandesite lava flows; also include basaltic trachyandesite compositions. Flow tops are very vesicular (10-50%). This unit contains 1-2% olivine and trace plagioclase phenocrysts in a strongly devitrified groundmass of plagioclase and opaque oxide crystals, and is moderately vesicular (15%), with minor infill of carbonate. The lava flows emanated from a shield volcano centered on Mount Hicks (John et al., 2015).

Qmsa, Trachyandesite of Mud Spring (0.28 Ma):

Medium-dark-grey, sparsely to moderately porphyritic, rubbly trachyandesite lava flow (John et al., 2015). This unit contains ~7% phenocrysts of hornblende, pyroxene, and plagioclase in a weakly devitrified groundmass of plagioclase, pyroxene, hornblende and opaque oxide crystals. This is the youngest unit of the AVF.

Qta1, Trachyandesite flow 1 (1.45 Ma):

Several medium-light-grey, sparsely porphyritic, vesicular (4-10%) trachyandesite lava flows with minor basaltic trachyandesite composition that form a shield volcano (John et al., 2015). This unit contains 1-8% phenocrysts of plagioclase, hornblende, and olivine in a completely devitrified groundmass of plagioclase, pyroxene, opaque oxides, and olivine.

Qta2, Trachyandesite flow 2 (1.32 Ma):

Medium-grey, nearly aphyric, very vesicular (minor carbonate infill) trachyandesite lava flow. John et al. (2015) also observed flow lamination, and that forms a small volcanic cone and a plug. This unit contains ~1% olivine and rare plagioclase phenocrysts in strongly devitrified groundmass of plagioclase and opaque oxides.

QTaap, Trachyandesite South of Aurora Peak (2.45 Ma):

Medium-dark-grey, sparse to moderately porphyritic, moderately vesicular (10%) trachyandesite lava flows. This unit contains phenocrysts of hornblende and plagioclase (~6%; John et al., 2015 report 15% phenocrysts) in a weakly devitrified, dusty glass

groundmass. Remnants of two lobes of lava flows are shown on the Bodie Hills map (John et al., 2015).

QThch, Hornblende trachydacite of Cedar Hill (2.95 Ma):

Medium- to dark-grey, moderately porphyritic, trachyandesite to trachydacite lava flows and lava cones. The flows are sparsely vesicular (<2%), massive and blocky, with irregular a'a flow tops, flow banding, and oxidation textures preserved. This unit contains ~22% plagioclase and hornblende phenocrysts with trace biotite and pyroxene phenocrysts in a weak to moderately devitrified, dusty glass groundmass containing microphenocrysts of plagioclase, hornblende $\pm$ pyroxene.

QTopb, Olivine-pyroxene basaltic trachyandesite (age unknown, a maximum age of 4.00 Ma was used for plots and initial isotope calculations):

Medium-grey, moderately porphyritic, basaltic trachyandesite in a small, isolated outcrop. This unit contains ~25% plagioclase, olivine, and pyroxene phenocrysts in a completely devitrified groundmass containing microphenocrysts of plagioclase, pyroxene, and opaque oxide crystals.

Tach, Trachyandesite of Cedar Hill (3.70 Ma):

Light- to dark-grey, moderate to strongly porphyritic, trachyandesite lava flows and lava cones, with minor basaltic trachyandesite composition. This unit contains ~27% phenocrysts of plagioclase, pyroxene, and olivine in a moderately devitrified groundmass. The lava flows are sparsely vesicular (1-5%), but can be very frothy (30-

40% vesicles) in places. Pyroxene phenocrysts can display subophitic texture (plagioclase intergrowths).

Tba, Basaltic trachyandesite of Beauty Peak (2.90 Ma):

Light-grey, sparsely porphyritic basaltic trachyandesite lava flows that form a ≤ 250 -m thick sequence emanating from a shield volcano (John et al., 2015). This unit contains ~3% phenocrysts of olivine surrounded by intergrowths of pyroxene and flow-orientated plagioclase crystals (trachyte texture), in a groundmass of plagioclase, olivine, and opaque oxides.

Tbch, Trachybasalt of Cedar Hill (2.60 Ma):

Medium- to dark- grey, moderately porphyritic trachybasalt lava flow. This unit contains ~10% (Lange and Carmichael report 5%; John et al., 2015 report 5-20%) phenocrysts of plagioclase, olivine, and pyroxene in a strongly devitrified groundmass of plagioclase, pyroxene, olivine, and opaque oxide crystals. Rare carbonate may infill the vesicles. The lava flows are interbedded with cinder to form a composite cone (John et al., 2015).

Tdap, Trachydacite of Aurora Peak (2.70 Ma):

Light- to pinkish-medium-grey, moderately porphyritic trachydacite lava flows forming a dome. This unit contains ~17% phenocrysts of plagioclase, hornblende, and minor biotite in a weak to moderately devitrified groundmass of plagioclase, hornblende, and opaque oxide crystals. This unit also contains glomerocrysts (crystal clusters) of hornblende $\pm$ plagioclase 2-6 mm in diameter.

T_{hd}, Hornblende trachydacite (3.50 Ma):

Dark-grey, sparsely porphyritic trachydacite lava flows, which form a small outcrop. This unit contains ~8% phenocrysts of plagioclase, hornblende, and pyroxene in a moderately devitrified groundmass of dusty glass, plagioclase, and opaque oxide crystals.

T_{mlp}, Basaltic andesite of Locomotive Point (3.90 Ma):

Medium-light-grey, sparsely porphyritic basaltic andesite lava flows emanating from a plug-filled vent along the same trend as the BHVF Big Alkali domes (John et al., 2015).

This is the oldest AVF unit and is isolated from the rest of the Pliocene-Pleistocene exposures within the Bodie Hills by about 10 km (John et al., 2015). The lava contains ~4% phenocrysts of pyroxene and olivine in a moderate to strongly devitrified groundmass of plagioclase, pyroxene, opaque oxides, and olivine.

T_{mr}, Martinez Hill rhyolite (2.60 Ma):

Light-grey, sparsely porphyritic, rhyolite lava flows forming 3 coalescing domes (John et al., 2015). This unit contains 4% phenocrysts of quartz, alkali-feldspar (sanidine), plagioclase, and biotite in a fresh glassy groundmass. Accessory minerals are titanite and opaque oxides. Spherulites are also present. John et al. (2015) also noted accessory zircon and allanite.

Tpch, Pyroxene trachyandesite Cedar Hill (2.80 Ma):

Medium- to dark-grey, moderately porphyritic, trachyandesite lava flows from composite lava cones (John et al., 2015). This unit contains 15-24% phenocrysts of plagioclase, pyroxene, olivine, and trace hornblende in moderately devitrified groundmass of plagioclase, olivine (iddingsite), and pyroxene with 5 to 15% vesicles.

Trsp, Spring Peak rhyolite (3.70 Ma):

Very light-grey to very pale-orange, sparsely porphyritic, and voluminous series of rhyolite flow domes and plugs, and scarce dikes. This unit contains 3-4% phenocrysts of quartz, plagioclase, and biotite, and trace amounts of hornblende and alkali-feldspar (sanidine) in a glassy groundmass.

Four additional AVF map units were not part of this study: Qa, Andesite of Reheis et al. (2002), QTai, Trachyandesite, QTob, Olivine basaltic trachyandesite, and QTbbc, Basaltic debris-flow deposits of Cedar Hill. The author did not encounter these units during fieldwork, and there are no known analyses of these units from previous publications.

3.2 Mineralogy

3.2.1 Mafic to intermediate units

3.2.1.1 Opaque oxides (Titanomagnetite)

Cubic, opaque oxide crystals do not occur as a phenocryst phase in any lava, but they are a ubiquitous groundmass phase. Based on the microprobe analyses of Lange and

Carmichael (1996), this phase is titanomagnetite (Ti-magnetite); no ilmenite was identified in any of the units they analyzed (i.e., Qaca, Qmh, Qmsa, Qta1, QThch, Tach, Tba, Tbch, Thd, Tpch).

3.2.1.2 Olivine

Olivine occurs as phenocrysts and microphenocrysts in the trachybasalt (Tbch), the basaltic andesite (Tmlp), both basaltic trachyandesite units (Tba and QTopb), and several trachyandesite units (Qmh, Qta1, Qacc, Qta2, Tpch, Tach). Phenocrysts are often skeletal, and are no larger than 2 mm. Although they can be broken and/or rounded, they form euhedral to subhedral crystals, except in trachyandesites where they are enclosed by pyroxenes and are heavily altered. Olivine is commonly altered to iddingsite. Iddingsite alteration can be weak or strong, even between similar-sized crystals within the same sample (e.g., Qta1 203443, Figure 3.3). Microphenocrysts are generally completely altered to iddingsite. Cubic, 0.01-0.05 mm sized, magnetite inclusions may occur along fractures in olivine, especially in basaltic trachyandesites and trachyandesites (e.g., Qmh, 13-AU-29 and Qta 1, 203443). In samples where more than 1% volume of hornblende is present, olivine phenocrysts are rare and olivine may only occur within the groundmass.

Figure 3.3 Weak and pervasive iddingsite alteration of olivine phenocrysts.

The phenocrysts are in devitrified glass containing plagioclase laths. Transmitted light photomicrographs in plane polarized light (top) and cross polarized light (bottom). Field of view is 2.0 mm. (Unit Qta1, sample 203443).

3.2.1.3 Pyroxene

Pyroxene commonly occurs as phenocrysts and microphenocrysts in the mafic and intermediate compositions except nearly aphyric rocks (e.g., Qmh and Qta2), and where hornblende becomes a dominant phase (e.g., Qta1, QTaap, Tdap, and QThch). Pyroxene phenocrysts are often zoned: examples of step zoning are noted with prominent changes in color (green to colorless in plane polarized light), birefringence, and/or inclusions (of

olivine, plagioclase, or melt) restricted to a particular zone, while gradual or progressive zoning is usually only noted through changes in birefringence. The cores of phenocrysts can have vermicular or cellular sieve texture (Qaca, Qacc, Qmsa, Tbch, Tmlp; e.g., Figure 3.4a). Pyroxenes in trachyandesites are strongly associated with anhedral, vermicular titanomagnetite and can form a symplectic texture (Tpch and Qacc-203232; Figure 3.4b). Although pyroxene phenocrysts are rarely larger than 2 mm, they can reach 4 mm.

The microprobe analyses of Lange and Carmichael (1996) indicate that orthopyroxene occurs in two trachyandesite units as phenocrysts and microphenocrysts (in units Qaca and Tpch). They also found a single orthopyroxene crystal in Qmsa, but interpreted it as a xenocryst. John et al. (2015) also reported orthopyroxene in Tpch. In this study, occasional reversely zoned pyroxene crystals with green-core and colorless rims in plain polarized light were observed in Tpch (Figure 3.5c) and as an isolated crystal in Qaca (sample 08-BA-03). The remainder of the pyroxenes were determined to be clinopyroxene. Based on the microprobe analyses of Lange and Carmichael (1996), these clinopyroxene crystals are augite.

Figure 3.4 Notable textures in clinopyroxene: reverse zonation, symplectic, and sieve textures.

(a) Sieve texture in core of clinopyroxene. Cross polarized light. Field of view is 2.0 mm. (Unit Tmlp, sample 203195). (b) Symplectic texture. Vermicular titanomagnetite in pyroxene, emanating from crystal boundary between pyroxene and iddingsite-altered olivine. Groundmass is devitrified glass (shaded region is from sample number engraving in glass below sample). Plane polarized light. Field of view is 2.0 mm. (Unit Tpach, sample 203181). (c) Green core, reverse zoned pyroxene phenocryst in weakly devitrified groundmass. A cluster of pyroxene (to the right), heavily iddingsite-altered olivine crystals, occasional plagioclase laths, and opaque crystals are also present. Plane polarized light. Field of view is 2.0 mm. (Unit Tpach, sample 203247).

3.2.1.4 Hornblende

Nearly all hornblende phenocrysts are mantled by a fine-grained opaque (in transmitted light) material, commonly referred to as opacite (one sample contains opacite-free hornblende phenocrysts, 13-AU-07 of unit QThch). Likewise, all hornblende crystals in the groundmass have been entirely converted to opacite. This opacite texture is also found in rock of the modern and ancestral Cascades arcs and is characterized by microcrystalline anhydrous assemblage of plagioclase, pyroxene, and magnetite as a result of a breakdown reaction caused by H₂O loss from the coexisting melt during magma ascent (e.g., Rutherford and Hill, 1993; Cousens et al., 2008).

Hornblende phenocrysts are euhedral to subhedral, with occasional embayed faces or absent cores. Rounded relict cores of phenocrysts are common where opacite rims are thick. Opacite rims are commonly 0.05 mm thick but range from 0.01-0.08 mm and extend to the crystal core, entirely converting hornblende to opacite in some samples (Figure 3.5). Hornblende usually occurs as isolated crystals but may form clusters, commonly with clinopyroxene crystals. Only one sample contains opacite-free hornblende (QThch, 13-AU-07), although Lange and Carmichael (1996) noted opacite-free rims in two additional units (Thd and Tdap).

Figure 3.5 Cluster of hornblende and pyroxene crystals in weakly devitrified groundmass.

Hornblende is rimmed by opacite. Plane polarized light. Field of view is 2.0 mm. (Unit Thdch, sample 12-BA-33).

3.2.1.5 Biotite

Biotite phenocrysts occur in minor amounts (trace to 1%) in two intermediate flows (units QThch and Tdap). They are euhedral to subhedral, tabular or columnar. They rarely have opacite rims (QThch: 13-AU-13, 13-AU-20, and 13-AU-21).

3.2.1.6 Plagioclase

Plagioclase is the most varied and most common mineral within the mafic to intermediate composition rocks. It has a wide range in phenocryst modal abundance (1-20%), exhibits a continuous range of sizes, and displays a range of textures (e.g., sieve, subophitic, “swallow-tail”) and zonation (commonly concentric and oscillatory, rarely reverse or convoluted). Phenocrysts are usually euhedral to subhedral, rarely rounded or embayed. They may display subophitic texture with pyroxene, have inclusions of pyroxene or apatite, or occur as inclusions in pyroxene. Plagioclase is ubiquitous as a groundmass phase, even when it does not occur as phenocrysts (e.g., Tmlp, Tba), and may form a trachytic or felty fabric, or delineate flow banding.

Plagioclase with a mesh-like, cellular, or dusty sieve-texture typically comprises a minor portion of the plagioclase; however, this texture can be widespread in Qacc (particularly in the basaltic trachyandesite and trachyandesite compositions) and Tpch. Sieve texture is typically identified within the cores of crystal, where the sieved zones are generally well rounded, but it may also be found within a specific zone of a crystal or may be pervasive throughout the crystal core and have a clear, sieve-free rim (Figure 3.6). The sieve textures are likely due to resorption or internal corrosion resulting from thermal or compositional disequilibrium between the crystal and the melt (Cox et al., 1979; Streck, 2008; see section 5.3.2). The sieve texture is pronounced in megacrysts, but may be present in any size of crystal, though rarely within the groundmass feldspars.

Only two units have reversely zoned plagioclase phenocrysts (Qta 1, Qmsa). These phenocrysts display pervasive sieve texture throughout the crystal with a thin, clear rim. Based on the microprobe analyses of Lange and Carmichael (1996), these reversely zoned crystals have sodic cores composed of An_{28} and An_{40} respectively, with calcic rims of An_{60} and An_{54} .

Figure 3.6 Sieve texture in plagioclase (Pl).

This texture ranges from pervasive to absent and may occur within the core or rim of grains. Plane polarized light (top) and cross polarized light (bottom). Other phenocrysts are pale green clinopyroxene (Cpx), and opaque minerals. Field of view is 4.5 mm. (Unit Tpch, sample 203246).

3.2.1.7 Xenocrysts and xenoliths

Xenocrysts and xenoliths are crystals and rock fragments, respectively, that have been incorporated into magma from a foreign source; they have no genetic relationship to the host rock. These are easier to identify if they are inappropriate for the magma composition in which they are found, for example, a quartz crystal found in basalt will be surrounded by breakdown mantle or reaction rim. Their identification becomes much more difficult as they approach the host magma composition. The addition of xenolithic material causes both thermal and compositional disequilibrium. In order to dissolve the foreign material, heat is taken from the surrounding magma causing local undercooling, causing fine-grained crystallization of the magma surrounding the xenolithic material (Vernon, 2004).

In several AFV intermediate lava units, quartz crystals are mantled by clinopyroxene. These xenocrysts are found in a range of compositions from basaltic andesite to trachydacite (Tmlp, sample 203184; Qta 1, sample 203398; Qacc, sample 203402; Figure 3.7). There are also two occurrences of a fine-grained xenolith mantled by clinopyroxene (Tbch, sample 12-BA-32; Tmlp, sample 203195).

Figure 3.7 Quartz (Qtz) xenocryst mantled by pyroxene

The xenocryst is in a devitrified glassy groundmass with plagioclase microlites (Pl), opaque minerals (altered hornblende?) and clinopyroxene (Cpx) (to right of photomicrograph). Plane polarized light (top) and cross polarized light (bottom). Field of view is 4.5 mm. (Unit Tacc, sample 203402).

3.2.2 High-silica rhyolites

The AVF high-silica rhyolites are sparsely porphyritic. They contain ~ 5% phenocrysts of quartz, sanidine, plagioclase, biotite, and trace hornblende (Trsp, 11-BA-46), in a glassy groundmass. Accessory minerals include opaque oxides and titanite (Tmr). Dense masses of fine intergrown quartz and feldspar needles radiating out from a common nucleus, or spherulites, occur in both rhyolites. Spherulitic aggregates may also fringe phenocrysts or

glomerocrysts (growth mantles) (Figure 3.8). Spherulites crystallize under strong supersaturation conditions (Vernon, 2004).

Figure 3.8 Spherulitic aggregates fringing a glomerocryst of plagioclase crystals.

Plane polarized light (top) and cross polarized light (bottom). Field of view is 2.0 mm. (Unit Tmr, sample 08-BA-06).

3.2.3 Basement rocks

Seven basement rock samples from the Bodie Hills region were collected for evaluation of potential basement contamination of the AVF magmas: two Paleozoic sedimentary rocks (a siliceous shale and a black argillite), three Cretaceous granitic rocks (samples of the granite from Murphy Creek, the granodiorite of Masonic Mountain, and the

monzogranite of Rattlesnake Gulch), and two Jurassic(?) metavolcanic rocks (a meta-andesite and a metadacite from Masonic Mountain). The sedimentary rocks are both very fine-grained and have late-stage fracturing and veining. The minerals in the granitic rocks are all anhedral, and dominated by quartz, altered (dusty) alkali-feldspar, and plagioclase. Alkali-feldspar crystals can have exsolution lamellae (perthite), while plagioclase can have myrmikite texture at crystal margins (sample 203377), and quartz can have undulatory extinction. The mafic mineral component (opaque oxides, biotite, and secondary/alteration phase, chlorite) of the granitic rocks is small, ranging from trace amounts to 5%. Accessory minerals include titanite, apatite, and secondary epidote. The metavolcanic rocks have highly altered feldspars, with quartz, chlorite ($\pm$ biotite), and opaque crystals in the groundmass. The metadacite from Masonic Mountain has strong foliation defined by biotite.

4 Chapter: Geochemistry

The total iron was recalculated to obtain iron expressed as FeO^T [wt. %] using 1.1111 as the weight ratio between Fe_2O_3 and FeO :

$$\text{FeO}^T [\text{wt. \%}] = \frac{\text{Fe}_2\text{O}_3^T [\text{wt. \%}]}{1.1111}. \quad (5)$$

The major and trace element data were recalculated as volatile free prior to plotting in figures. Major element oxides (SiO_2 , Al_2O_3 , CaO , MgO , Na_2O , K_2O , Fe_2O_3^T , MnO , P_2O_5 , and TiO_2) were thus recalculated to total 100 wt. %, and the dry factor ratio is used to recalculate the trace element concentrations. These data were combined with data from previously published work (primarily Lange and Carmichael, 1996).

4.1 Lithological classification of the Aurora Volcanic Field

The Aurora Volcanic Field (AVF) rocks have bimodal compositions, dominated by a near continuous range from mafic to intermediate compositions and high-silica rhyolite; the intermediate and rhyolitic compositions are separated by a prominent gap between about 66 and 75 wt. % SiO_2 (Figure 4.1a). The mafic to intermediate rocks range from trachybasalt to trachydacite (Le Bas et al., 1986). The trachybasalt, basaltic trachyandesite, trachyandesite, and trachydacite rock types can be further subdivided based on their potassium content, whereby they are predominantly potassic trachybasalts, shoshonites, latites, and trachydacites, respectively (Le Maitre, 2002). The AVF rocks are transitional between alkaline and subalkaline with increasing silica (Figure 4.1a, boundary from Irvine and Baragar, 1971). They are also classified as magnesian to weakly ferroan (not shown), calc-alkalic to alkali-calcic (Figure 4.1b; Frost et al., 2001),

and metaluminous to weakly peraluminous (Figure 4.1c; Shand, 1943). The high potassic nature of the AVF suite is also shown by the K_2O versus SiO_2 diagram (Figure 4.1d) on which most samples plot within or just above the high-K field within the shoshonitic field (boundaries based on Ewart, 1982 and Le Maitre, 1989).

Figure 4.1 Classification diagrams and Harker diagrams.

These diagrams show compositions of Aurora Volcanic Field, AVF (colored symbols each represent a unique rock unit, see legend for details; data from Lange and Carmichael, 1996, and this study) compared to compositions of samples from Western Great Basin, WGB (blue field; data from Omerod et al., 1988; 1991; Cousens, 1996; Feldstein and Lange, 1999), Post-Arc volcanic rocks (white field; data from Cousens et al., 2011; 2012; 2013; Anderson, 2014), Bodie Hills Volcanic Field, BHVF (grey field; data from John et al., 2012), and the Ancestral Cascades Arc, ACA (red field; data from Cousens et al., 2008; B. Cousens, unpublished data, 2015).

(a) Total Alkali vs. Silica (TAS) variation diagram (field boundaries from LeBas et al., 1986; alkaline-subalkaline boundary from Irvine and Baragar, 1971). Inset shows the AVF units (black crosses) compared with other volcanic fields (WGB, Post-Arc, BHVF, ACA);

(b) $\text{Na} + \text{K}_2\text{O} - \text{CaO}$ versus SiO_2 , from Frost et al., 2001. The AVF lavas are calc-alkalic to alkali-calcic; (c) aluminum saturation index, ASI, from Shand, 1943. The AVF lavas are metaluminous to weakly peraluminous (only rhyolites and sample 203433 are peraluminous); (d) K_2O vs. silica (boundaries between low-K, medium-K, and high-K are from Le Maitre, 1989; high-K–shoshonitic boundary is from Ewart, 1982); (e) CaO vs. silica;

(f) TiO₂ vs. silica; (g) Ni vs. silica; (h) Sc vs. silica; (i) V vs. silica;

(j) $\text{CaO}/\text{Al}_2\text{O}_3$ vs. silica; (k) Al_2O_3 vs. silica; (l) Sr vs. silica; (m) Eu/Eu^* vs. silica. (Values of Eu_N , Sm_N , and Gd_N were normalized to a chondritic (N) reservoir using values from Anders and Ebiharra, 1982. See text for equation used to calculate the Eu anomaly);

(n) Ba vs. silica; (o) Th vs. silica, (the yellow field shows the high Th-U-Cs-Rb andesites); (p) P_2O_5 vs. silica; (q) Nb vs. silica.

4.2 Significance of covariant diagrams using major and trace elements

Major elements, and compatible and incompatible trace elements are plotted graphically to identify any underlying geologic relationship(s), which may exist between the different AVF units. The abscissa in bivariate plots is chosen based on maximum variability within the sample suite and/or based on distinguishing particular geochemical processes.

Compositional variations that correlate to the observed phenocryst assemblage may indicate the mineral phases that have crystallized and been removed from the melt (fractional crystallization). Since the AVF rocks have a large range in silica (49-78 wt. % SiO_2) and crystal fractionation processes are suspected, a series of variation diagrams with SiO_2 along the x-axis, also known as Harker diagrams, were plotted (Figure 4.1 d-q).

The Harker diagrams are interpreted to reflect the bulk system changes, for example, as elements are incorporated into crystallizing minerals or remain the melt phase during crystal fractionation, or the addition of elemental abundance to the system through magma mixing or assimilation. Trace elements (elements that are not stoichiometric constituents of phases, often expressed in parts per million) especially give insight into the petrological processes the rock experienced (e.g., partial melting, crystal fractionation; Rollinson, 1993; White, 2013). During these processes, trace elements can substitute for major ions in many rock-forming and accessory minerals. Ions with the same ionic charge and similar size tend substitute more readily. Partition coefficients (D) are used to predict whether a trace element will be accepted into a specific solidifying mineral crystal structure ($D > 1$) or preferentially remain in the melt ($D < 1$). They are

sensitive to the temperature, pressure, water content, oxygen activity, composition of the melt, and chemistry and crystal structure of the mineral.

4.3 Harker diagrams and magmatic evolution of Aurora Volcanic Field

Linearly decreasing concentrations of FeO^{T} (not shown), MgO (not shown), TiO_2 , CaO , MnO (not shown), $\text{CaO}/\text{Al}_2\text{O}_3$, Co (not shown), Ni , V , and Sc with increasing silica corresponds to fractionation of olivine, clinopyroxene, titanomagnetite, and hornblende (Figure 4.1 e-j). V and TiO_2 have a strong, negative, linear correlation with silica (Pearson correlation coefficient, r^2 , of 0.77 and 0.91, respectively). Although V will substitute for Ti , Ti is incompatible in all basalt and basaltic andesite phases (e.g., D_{Ti} of 0.02 in olivine, 0.1-0.4 in pyroxenes, 0.9 in plagioclase; Arth, 1976; Fujimaki et al., 1984) except for magnetite (D_{Ti} of 7.5; Schock, 1979). This suggests fractionation of titanomagnetite played an important role in the AVF, even if this phase does not occur in the phenocryst assembly.

Ni concentrations display two decreasing trends, one from 49 to 55 weight percent silica and another from 57 to 62 weight percent silica (Figure 4.1 g). These trends correspond to lavas containing both olivine and clinopyroxene phenocryst phases (Tbch, Tba, QTopb, Qacc, and Tpch) or nearly aphyric ($\leq 2\%$ volume phenocrysts, Qmh).

Sc and $\text{CaO}/\text{Al}_2\text{O}_3$ both show a negative linear trend with increasing silica content providing support for fractionation of clinopyroxene (Figure 4.1 h, j). Olivine or plagioclase fractionation will not produce a decrease in $\text{CaO}/\text{Al}_2\text{O}_3$ with increasing silica;

this ratio will remain constant with fractionation of olivine because it contains neither element, whereas the plagioclase fractionation may cause a slight increase. Sc may also partition into hornblende especially in andesite and dacite rock compositions, where this phase more commonly crystallizes.

The compositional variations are consistent with fractionation of olivine, clinopyroxene, and hornblende, which are observed phenocryst phases in the AVF. Titanomagnetite only occurs in the groundmass (Lange and Carmichael, 1996); however, it is a ubiquitous phase in the mafic to intermediate composition rock units, and may also play an important role in fractionating AVF magmas.

Plagioclase is the only phase in AVF lavas that is inconsistent with the inferred fractionating mineral assemblage. As previously mentioned, plagioclase fractionation will have a minimal effect on $\text{CaO}/\text{Al}_2\text{O}_3$ versus silica content, since plagioclase removes both Ca and Al from the magmas. Although the AVF magmas exhibit a strong negative correlation of CaO with silica, there is only a relatively flat Al_2O_3 correlation, with a distinct drop in Al_2O_3 corresponding to the high silica rhyolites (Figure 4.1 e, k). The same drop occurs in Sr concentration (Figure 4.11). If the rhyolites are excluded, the correlation coefficient of the Harker diagrams is much lower for Al_2O_3 and Sr, and there is a notable decrease for CaO (i.e., the $r^2_{\text{Al}_2\text{O}_3}$ for all AVF rocks is 0.66 and drops to 0.17 when excluding the rhyolites; similarly r^2_{Sr} of 0.58 drops to 0.22, and r^2_{CaO} of 0.90 drops to 0.82). The distinct drops in Al_2O_3 , CaO, and Sr contents at high silica values likely correspond to abundant fractionation of plagioclase. There is also a continuous decline in

Eu with increasing silica (not shown); however, the Eu anomaly is a better measure of Eu fractionation. The Eu anomaly compares the measured concentration of Eu with the expected value of Eu. The Eu anomaly can be quantified by the equation suggested by Taylor and McLennan (1985):

$$\frac{\text{Eu}}{\text{Eu}^*} = \frac{\text{Eu}_N}{\sqrt{(\text{Sm}_N) \times (\text{Gd}_N)}}, \quad (6)$$

where $\frac{\text{Eu}}{\text{Eu}^*}$ is the Eu anomaly, Eu^* is the “expected” value of Eu extrapolated between Sm_N and Gd_N , and Eu_N , Sm_N , and Gd_N are the measured concentrations of Eu, Sm, and Gd, normalized to chondrite or primordial mantle values. Figure 4.1m was plotted using chondritic reservoir values from Anders and Ebiharra, 1982 (more on use of REE and normalizing to chondrite values in the following section, 4.4). The Eu anomaly versus silica plot shows that only minor amounts of Eu are partitioned into plagioclase until high silica contents are reached (Figure 4.1m).

Many incompatible elements show weak or no correlation with silica (e.g., Ba, Th, Nb in Figure 4.1 n, o, q). The LREE contents are scattered, and although the rest of the REE have a tighter trend, with a slight negative slope. Nb and Zr (not shown) each show very weak negative correlations with silica. Th has a near horizontal slope with silica, aside from two positive spikes in abundance: one in the andesites and another in the rhyolites. Ba is scattered, but shows a weak, positive correlation with silica before a drop in abundance corresponding to the rhyolites; this may reflect fractionation of biotite and/or potassium feldspar. Of the incompatible elements, Rb has the strongest correlation with

silica, and shows a moderate, positive correlation with silica, with a spike in abundance in some andesites (not shown).

4.4 Rare earth element diagrams

4.4.1 Geological significance of rare earth elements

The rare earth elements (REE) are a unique set of trace elements; they have very similar physical and chemical properties, largely resulting from having similar sized atomic radii, and they form stable 3+ ions (except for Pm, whose isotopes are all unstable and now extinct; Rollinson, 1993). Low-atomic-number members (atomic number 57 to 61, i.e., lanthanum, La, to promethium, Pm) are called light rare earth elements (LREE), and those with high atomic numbers (68 to 71, i.e., erbium, Er, to lutetium, Lu) are called heavy rare earth elements (HREE). Occasionally the mid-range members (atomic numbers 62-67, samarium, Sm, to holmium, Ho) are referred to as the middle rare earth elements (MREE). Like other trace elements, the REE occur in many rock-forming silicates by replacing major ions. However, due to slight differences in their size (lanthanide contraction: atomic radii decreases with increasing atomic number from 1.15 Å for La to 0.93 Å for Lu), minerals will selectively accept REE into their crystal structure (Faure and Mensing, 2005). For example, garnet will preferentially accept HREE with smaller atomic radii into its crystal structure (e.g., Lu). Therefore, partial melting of garnet-bearing rocks (e.g., garnet peridotite or eclogite) where garnet is left in the residue results in a melt that is depleted in HREEs, especially Lu, creating a steep negative slope in the HREE part of normalized REE diagrams (Rollinson, 1993).

Some REE can have stable ions in other oxidation states; however, only Ce^{4+} and Eu^{2+} are of geologic importance. These ions are slightly smaller and larger than their respective $3+$ ions. In specific environments, these ions will be preferentially incorporated into certain minerals and fractionate out. For example, in oxidizing sedimentary environments, Ce^{3+} oxidizes to Ce^{4+} , which precipitates out of solution as CeO_2 and creates a negative Ce anomaly; Eu^{2+} will be preferentially incorporated minerals (e.g., plagioclase) by replacing Ca^{2+} in igneous melts under low O_2 fugacity conditions (Rollinson, 1993).

Rare earth elements are incompatible elements, and therefore have an affinity for the melt phase. They are often compared to their concentrations in a chondritic meteorite, because these meteorites are believed to be an unfractionated sample of the solar system (Rollinson, 1993). Deviations from the chondritic values (e.g., change in slope, abundance, and/or anomalies) indicate an evolution of the magma. Trends on normalized REE diagrams (i.e., REE patterns) are influenced by the source rock chemistry and geochemical processes affecting the crystal-melt equilibrium, such as variable amounts of partial melting of the source rock, crystal fractionation, and/or assimilation of wallrock. Due to their low solubility in aqueous fluids, REE are relatively immobile and original geochemical signatures are often preserved despite subsequent weathering, hydrothermal alteration, and low-grade metamorphism (Michard, 1989; Rollinson, 1993).

4.4.2 Rare earth element patterns in Aurora Volcanic Field

Rare earth element concentrations for each AVF map unit were averaged and plotted normalized to a chondritic (N) reservoir using values from Anders and Ebihara (1982;

Figure 4.2 ab). All AVF rocks show greater enrichment for the LREE and MREE than for the HREE. This creates a prominent slope for the LREE and MREE (La/Ho_N between 6.1 to 25.0), and a relatively flat slope for the HREE (Er/Lu_N between 0.6 and 1.3). In geological systems dominated by crystal fractionation, the REE tend to remain in the melt phase and increase in abundance as silica increases (Cox et al., 1979). However, instead of increasing REE abundance with increasing silica in the AVF rock suite, the most mafic unit, the trachybasalt (Tbch), has the highest REE contents. The basalt and andesite REE patterns overlap (Figure 4.2a), and the dacites and rhyolites (Figure 4.2b) tend have a lower abundance of REE than the more mafic units. Several units show a slight negative Eu anomaly (e.g., Tpch, Qta2); however, only the rhyolite units show a pronounced negative Eu anomaly, suggesting late fractionation of plagioclase or the rhyolites were derived from or incorporated melt from a gabbroic crustal rock with residual plagioclase. The slightly negative Eu anomalies also may indicate minor fractionation of plagioclase.

The high LREE slope and high abundances in REE relative to a chondritic or primitive mantle reservoirs in the more mafic AVF rocks (e.g., Tbch, Figure 4.2a) suggest either a low degree of partial melting of the mantle source and/or a LREE-enriched mantle source. The flat HREE slope and high abundance in HREE suggests a garnet-free source.

Figure 4.2 Average (by unit) chondrite-normalized (Anders and Ebiharra, 1982) REE diagrams for AVF samples.

REE abundances of the BHVF rocks are bounded by the two grey patterns (John et al., 2012). (a) Trachybasalt to trachyandesite samples. (b) Trachydacite to rhyolite samples. Both rhyolites (Spring Peak and Martinez Hill) have distinct negative Eu anomalies.

4.5 Primitive mantle normalized multi-element diagrams

Multi-element diagrams compare the abundance of elements that are incompatible in both mantle minerals and in minerals crystallizing from the magma. Certain large ion lithophile elements, LILE (elements with large cations of low charge, e.g., Rb, Ba, K, Sr), and high field strength elements, HFSE (elements with small, highly charged cations, e.g., Th, Nb, Ta, P, Zr, Hf, Ti, Y) are added to the REE in order of increasing compatibility with respect to a small fraction of mantle melt (Rollinson, 1993).

Peaks and troughs are common in multi-element diagrams due to the behavior difference between the various groups of trace elements. The LILE are mobile during hydrothermal alteration or metamorphism; they can serve as an indication of interaction with a fluid phase. They are also concentrated in the earth's crust and may indicate crustal contamination of magmas. The HFSE are fluid immobile and therefore may reflect the chemistry of the source rock or discriminate crystal-melt processes during magma evolution (e.g., fractionation of P into apatite in rhyolites; Pearce 1983; Rollinson, 1993).

Trace element concentrations for each AVF vent were averaged and plotted on a multi-element diagram, normalized to a primitive mantle reservoir (reservoir composition before formation of the continental crust) using values from Sun and McDonough (1989; Figure 4.3ab). The AVF units show distinct volcanic arc- or subduction-related signatures, with enrichment in Ba and Pb, and depletion in Nb and Ta. The rhyolites differ from the rest of the units with a relative depletion in Ba, and a pronounced depletion of P and Ti. This could be due to fractionation of potassium feldspar, apatite, and titanite, respectively, all of which are observed in thin section.

Figure 4.3 Average (by unit) primitive mantle-normalized (Sun and McDonough, 1989) trace element variation diagrams for AVF samples.

BHVF rocks are bounded by the two grey patterns (John et al., 2012). Both volcanic fields have arc- or subduction-related signatures, with enrichment in Ba and Pb, and depletion in Nb and Ta. (a) Trachybasalt to trachyandesite samples. (b) Trachydacite to rhyolite samples.

4.6 Radiogenic isotopes

4.6.1 Geological significance of radiogenic isotopes

Radiogenic isotopes are commonly used as tracers for petrogenetic studies to identify geological processes and sources of magmas, because they are not fractionated by crystal-melt processes (e.g., partial melting, crystal fractionation; Rollinson, 1993). In particular, isotope analysis can be used to determine whether magma mixing, crustal assimilation, or both have occurred in the exchange of matter within underlying magma reservoirs (Rollinson, 1993). Using more than one set of isotopes allow geological processes to be more easily identified; due to their different chemistry, each parent-

daughter pair of isotopes behaves uniquely, and varies in sensitivity to specific processes. Therefore, trends that are apparent in multiple isotope pairs have significant geological implications (Rollinson, 1993).

Sm and Nd are part of the same chemical group (lanthanides or REE group); they are weakly susceptible to fractionation due to petrological processes controlled by crystal-melt equilibrium. They typically occur in small concentrations, ranging from <1 ppm in ultramafic rocks to 8 ppm Sm and 45 ppm Nd in granitic rocks (Faure and Mensing, 2005). Their limited mobility during rock alteration allows them to retain their primary concentrations during weathering, hydrothermal alteration, and low-grade metamorphism (Rollinson, 1993). Therefore, the Sm-Nd isotope pair will maintain their source isotopic signature, but they will not be sensitive to small proportions of crustal mixing.

In contrast, Rb and Sr belong to different chemical groups (alkali metal and alkaline earth groups, respectively), causing differences in their crystal-melt affinities, and a high degree of mobility in crustal processes. Sr has an affinity to plagioclase, while Rb prefers to remain in the melt phase, and Sr will remain immobile during hydrothermal alteration while Rb is fluid mobile. The chemical differences between Rb and Sr allow them to be easily separated, causing extreme fractionation between mantle and crust (0.2 ppm Rb and 1 ppm Sr in ultramafic rocks compared to 110 ppm Rb and 440 ppm Sr in high-Ca granitic rocks; Faure and Mensing, 2005). Thus, older continental crust has very high Sr isotope ratios compared to the mantle or to young crust. This isotope pair is very sensitive to crustal mixing and hydrothermal alteration.

The Pb isotopes have three parent-daughter isotope pairs, creating a more complicated system, but useful since each isotope pair has a unique decay series and half-life. The short half-life of ^{235}U compared to ^{238}U means that old crustal rocks commonly have higher $^{207}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$ compared to younger crustal rocks and the mantle. Uranium and Th belong to the same chemical group (actinide element group), and have a stronger affinity for the melt phase than Pb. Uranium and Th share similar chemical behavior, except in oxidizing conditions where U forms a highly soluble 6+ ion in making it fluid mobile. Therefore, U, like Pb, will be mobile during crustal alteration processes while Th will remain immobile.

4.6.2 Radiogenic isotopes in the Aurora Volcanic Field

Covariate isotope plots show a strong, negative, near linear correlation between Nd and initial Sr isotope ratios, plotting in the enriched source field. In other words, the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ is higher (0.705109 to 0.706933) and ϵ_{Nd} is lower (-0.74 to -5.00) than the Bulk Silicate Earth value (BSE) (Figure 4.4). Some units have considerable spread of Nd and Sr isotope ratios (e.g., Qacc, Tach, and QThch) and several units overlap isotopically (e.g., Qaca and Qmh). Lead isotopes vary slightly, much less than those of Nd or Sr ($^{206}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 18.974\text{-}19.246$, $^{207}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 15.628\text{-}15.691$, $^{208}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 38.671\text{-}38.954$), and lie above the Northern Hemisphere Reference Line (NHRL) for oceanic basalts, and between the upper mantle and upper continental crust reservoirs (Zartman and Haines, 1988). Covariate Pb-Pb isotope plots show a strong, positive correlation,

with linear trend in the data towards and into the Enriched Mantle II (EMII) reservoir (continental sediment reservoir; Zindler and Hart, 1986; Figure 4.5).

There is no correlation between silica content and either Pb isotope ratios, or the Nd isotope ratio (r^2 between 0.006 and 0.03, Figure 4.6 ac, 4.7ace). There is, however, a weak positive correlation between silica and the Sr isotope ratio ($r^2 = 0.24$). None of the isotope ratios vary systematically with time (Figure 4.6bd, 4.7bdf).

Figure 4.4 ϵ_{Nd} versus $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ plot.

This figure shows isotopic compositions of AVF samples relative to the BHVF rocks, the ACA rocks, and the Post-arc rocks. It is noteworthy that the AVF rocks have significantly higher $^{86}\text{Sr}/^{87}\text{Sr}_{\text{initial}}$ and lower ϵ_{Nd} values than the BHVF rocks. Mantle reservoirs from Zindler and Hart (1986): DMM (depleted MORB mantle), BSE (Bulk Silica Earth, also known as the primary uniform reservoir, is the earth without a core), HIMU mantle (subducted oceanic crust), EMII (Enriched Mantle 2, subducted continental sediment), EMI (Enriched Mantle 1, subducted ocean sediment). For comparison: MORB (Mid-Ocean Ridge Basalts).

Figure 4.5 $^{207}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ versus $^{206}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ plot.

This figure shows a strong, positive correlation, linear trend towards and into the EMII field. AVF and BHVF overlap almost entirely, however, AVF is slightly more radiogenic than BHVF. Mantle reservoirs from Zindler and Hart (1986): DMM (depleted MORB mantle), BSE (Bulk Silica Earth, also known as the primary uniform reservoir, is the earth without a core), HIMU mantle (subducted oceanic crust), EMII (Enriched Mantle 2, subducted continental sediment), EMI (Enriched Mantle 1, subducted ocean sediment). For comparison: Northern Hemisphere Reference Line (NHRL) for oceanic basalts from Hart (1984).

Figure 4.6 $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ (top; a & b) and $^{143}\text{Nd}/^{144}\text{Nd}_{\text{initial}}$ (bottom; c & d) compared with silica content (left, a & c) and age (right, b & d).

Generally, the mafic units show greater range in isotope values than the felsic units; likewise, there is greater isotopic range in older lavas, than younger lavas. Radiogenic Sr is moderately correlated with silica. All other plots show no or weak correlation.

Figure 4.7 Initial Pb isotopic ratios compared with silica content (a, c, & e) and age (b, d & f).

Generally, the mafic units show greater range in isotope values than the felsic units; likewise, there is greater isotopic range in older lavas, than younger lavas. The rhyolites tend to be more radiogenic (c & e).

4.7 Oxygen isotope geochemistry

Like radiogenic isotopes, oxygen isotopes in volcanic rocks can reflect the source rock isotope signature, but are easily modified due to crustal interaction or hydrothermal alteration (Taylor, 1968). Globally, the $\delta^{18}\text{O}$ value of ultramafic to mafic rocks is +5.5‰ to +6.5‰. The $\delta^{18}\text{O}$ value will increase with silica content, typically 0.5‰ for each 10 wt. % SiO_2 increase during closed-system differentiation, such that granitic rocks tend to have values of +8‰ to +10‰ (Sheppard and Harris, 1985; Taylor and Sheppard, 1986). However, unlike radiogenic isotopes, stable isotopes have much smaller masses and are susceptible to temperature sensitive fractionation. The $\delta^{18}\text{O}$ values can be lowered by pervasive post-emplacement, high-temperature (>300°C) alteration with a low $\delta^{18}\text{O}$ fluid (e.g., hydrothermal alteration with meteoric waters) to values much less than +6‰. Alternatively, $\delta^{18}\text{O}$ values can be raised by low temperature (<200°C) post-emplacement water-rock interaction with sea water or meteoric water, increasing the $\delta^{18}\text{O}$ values by 6-8‰.

The AVF rocks tend to have $\delta^{18}\text{O}$ consistent with global values for mafic and granitic rocks. The trachybasalt (Tbch) and basaltic andesite (Tmlp) have $\delta^{18}\text{O}$ values of 6.5‰ and 6.4‰, respectively, and $\delta^{18}\text{O}$ values tend to increase with increasing silica content (Figure 4.8a). However, one rhyolite (Trsp) has the lowest $\delta^{18}\text{O}$ value of the AVF suite (4.0‰). In a covariate plot of $\delta^{18}\text{O}$ values against $^{86}\text{Sr}/^{88}\text{Sr}_{\text{initial}}$, two samples do not follow the moderate, positive correlation trend (Trsp and Tbch); these samples have highly

radiogenic Sr but low $\delta^{18}\text{O}$ (Figure 4.8b). The remainder of the AVF rocks show a moderate, positive correlation between $\delta^{18}\text{O}$ and silica contents, and $^{86}\text{Sr}/^{88}\text{Sr}_{\text{initial}}$.

Figure 4.8 $\delta^{18}\text{O}$ compared with silica (a), and with $^{86}\text{Sr}/^{88}\text{Sr}_{\text{initial}}$ (b) for AVF, ACA, and post-arc rocks.

Each rock group shows a positive correlation with silica and $^{86}\text{Sr}/^{88}\text{Sr}$. A couple rocks within the AVF rock suite fall outside the trend (Trsp for $\delta^{18}\text{O}$ vs. silica, and Trsp and Tbch for $\delta^{18}\text{O}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$).

4.8 Summary of Aurora Volcanic Field geochemistry

The AVF lavas have bimodal compositions, ranging from 49 to 78 wt. % SiO_2 , with a prominent gap between 66 and 75 wt. % SiO_2 . They are high-K calc-alkaline to shoshonitic in composition, and are metaluminous to weakly peraluminous. They are enriched in REE, particularly in the LREE. Primitive mantle-normalized incompatible

element patterns show a volcanic arc- or subduction-related signature, with enrichment in Ba and Pb, and depletion in Nb and Ta.

The compositional variations for mafic to intermediate composition rock units are consistent with fractionation of the observed phenocryst phases, olivine, clinopyroxene, and hornblende, and groundmass phase, titanomagnetite. They also suggest only minor fractionation of plagioclase until high silica contents are reached.

Neither lead nor neodymium isotope ratios covary systematically with silica content; however, strontium isotope ratios show a weak positive correlation. Pb, Nd, and Sr isotope compositions do not vary systematically with time. The AVF rock suite plots in the enriched source field, with higher $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ (0.705109 to 0.706933) and lower ϵ_{Nd} (-0.74 to -5.00) than the BSE value. Covariate lead-lead isotope plots show a strong, positive correlation, with linear trend in the data towards and into the EMII reservoir; the AVF rocks lie above the NHRL, in between the upper mantle and upper continental crust reservoirs ($^{206}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 18.974\text{-}19.246$, $^{207}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 15.628\text{-}15.691$, $^{208}\text{Pb}/^{204}\text{Pb}_{\text{initial}} = 38.671\text{-}38.954$).

The AVF rocks tend to have $\delta^{18}\text{O}$ values consistent with global values for mafic and granitic rocks, and tend to have a moderate, positive correlation between silica and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$. Two samples lie outside these trends: the rhyolite of Spring Peak (Trsp) with a low $\delta^{18}\text{O}$ value (+4.01‰), and basalt of Cedar Hill (Tbch) with high initial Sr and low Nd isotope ratios, as well as a low $\delta^{18}\text{O}$ value.

5 Chapter: Discussion:

5.1 Comparison to local volcanic fields

5.1.1 Comparison to Bodie Hills Volcanic Field (BHVF)

The AVF lavas are interpreted to have erupted in a post-subduction, extensional tectonic setting but have geochemical signatures similar to those of the older, subduction-related BHVF rocks. Unlike the AVF rocks, the BHVF show a continuous range in silica contents. The AVF rocks have higher concentrations of K_2O and P_2O_5 , and lower concentrations of Th, U (not shown), Cs (not shown), Rb (not shown), Cr (not shown), and Ni (Figure 4.1d, g, o, p), but otherwise the major and trace elements follow similar trends and largely overlap in abundance. Notably, the REE patterns of the AVF rocks are indistinguishable from those of the BHVF (Figure 4.2 a, b). Like the AVF, the BHVF mafic to intermediate rocks show no increase in REE abundance with increasing silica; the mafic to intermediate rocks have overlapping REE patterns. Likewise, the AVF primordial mantle-normalized incompatible element patterns are indistinguishable from those of older, subduction-related BHVF rocks: both show volcanic arc- or subduction-related signatures, with enrichment in Ba and Pb, and depletion in Nb and Ta (Figure 4.3 a, b). Enrichment in K and Sr and depletion in Ti are less pronounced in the AVF rocks than in the BHVF rocks.

Despite the compositional similarities between volcanic fields, covariate isotope plots show that the AVF rocks can have significantly higher $^{86}Sr/^{87}Sr_{initial}$ and lower ϵ_{Nd} values than the BHVF and other volcanic rocks of the ancestral Cascades arc (ACA) (Figure 4.4). Also, there is a stronger correlation between $^{86}Sr/^{87}Sr_{initial}$ and SiO_2 within the AVF

suite than in the BHVF suite ($r^2_{\text{AVF}} = 0.24$ versus $r^2_{\text{BHVF}} = 0.20$), and stronger correlation between initial $^{87}\text{Sr}/^{86}\text{Sr}$ and the Pb isotope ratios (moderate to strong, $r^2_{\text{AVF}} = 0.56\text{-}0.65$; weak to moderate, $r^2_{\text{BHVF}} = 0.15\text{-}0.43$). Lead isotopic plots show that the AVF and BHVF rocks overlap almost entirely in the EMII field; the AVF rocks tend to have slightly higher Pb isotope values than the BHVF rocks (Zindler and Hart, 1986; Figure 4.5). Although there are no isotopic oxygen isotope data for the BHVF, there are data for other rocks within the ACA. The AVF rocks show a much stronger correlation of $\delta^{18}\text{O}$ with either silica or $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ than the ACA rocks: for $\delta^{18}\text{O}$ and SiO_2 $r^2_{\text{ACA}} = 0.08$ and $r^2_{\text{AVF}} = 0.50$ (excluding Trsp), and for $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ $r^2_{\text{ACA}} = 0.06$ and $r^2_{\text{AVF}} = 0.19$ (excluding Trsp).

Both the AVF and the BHVF units oscillate between mafic to intermediate compositions and felsic compositions through time (AVF, Figure 5.1; BHVF, du Bray et al., 2016), and show no correlation of isotopic ratios with either silica content or time (AVF, Figures. 4.6, 4.7; BHVF, du Bray et al., 2016).

Some samples within the AVF andesite lavas show unusual enrichment in the incompatible trace elements Th (> 9 ppm, 13-24 ppm), U (>3ppm, 4.5-8.5 ppm), Cs (>3ppm, 3.5-8.5 ppm), and Rb (>95 ppm, 109-156 ppm) (samples 203181, 203246, 203247, and 203354 from Tpch; 203232 from Qacc; and 13-AU-05 from Tach; e.g., yellow field in Figures. 4.1v and 4.1w). These high Th-U-Cs-Rb AVF andesites also overlap with the most radiogenic strontium and least radiogenic neodymium isotope ratios of the BHVF rock suite (Figure 4.4). The high Th-U-Cs-Rb AVF andesites do not

show the extreme enrichment of Rb and Cs of the Bodie Hills rhyolites (Rb up to 435 ppm, Cs up to 186 ppm); instead, they show the greatest similarities to some samples of the trachyandesite of Del Monte and the trachydacite of Aurora Canyon (samples 00-BA-7, 00-BA-11, 10-BA-18, 00-BA-30, 00-BA-34, 08-BA-57), which also have higher Th, U, Cs, and Rb contents than the other BHVF andesites. Most of these high Th-U-Cs-Rb BHVF andesite samples even show a slight depletion in Ba (1025-1214 ppm) that is also apparent in the high Th-U-Cs-Rb Tpch samples (1050-1300 ppm, whereas other Tpch samples, 13-AU-22 and SN-326, have 1450-1550 ppm).

Figure 5.1 Age vs. silica plot for AVF units.

This plot shows an oscillating pattern for the AVF units from low to high silica with time; there is no correlation. Older rock units (≥ 2.5 Ma) show a greater range in silica (49-78 wt. % silica) composition than the younger rocks.

5.1.2 Comparison to post-ancestral Cascade arc volcanic fields

The AVF, the Tahoe-Truckee Volcanic Field (TTVF) of the Lake Tahoe region, and several volcanic fields south of the AVF within the Western Great Basin (WGB; e.g., Long Valley, Big Pine, and Kings River volcanic fields) are interpreted to have erupted after the southern edge of the subducting Juan de Fuca plate (Farallon plate) had migrated north of each respective volcanic field (Cousens, 1996; Cousens et al., 2011; Ormerod et al., 1988, 1991; Fieldstein and Lange 1999). They erupted after subduction had ceased in the area, hence they are grouped them as “post-ancestral Cascade arc”, or post-ACA, volcanic fields.

The post-ACA volcanic fields display similar element trends; their compatible elements show negative correlation with silica (e.g., Sc, V, Ni; Figures 4.1 g-i), while many of the incompatible elements show no correlation with silica (e.g., Ba, Th, Nb, shown in Figures 4.1 n, o, Q; and U, Pb, Rb, Zr, Y, REE, which are not shown). Their REE patterns overlap, and they have high LREE contents, a steep LREE to MREE slope, and a flat HREE slope. The incompatible trace element patterns of the TTVF and the WGB fields follow the arc-like patterns of the AVF, with some elements showing a slight increase or decrease in abundance (e.g., the Kings River region rocks show greater enrichment in Ba and K). Covariate isotope plots show considerable overlap between the volcanic fields. They have strong, negative, near-linear correlation between Nd and Sr isotope ratios, and show considerable spread of isotope values, transitioning from the depleted source field to the enriched source field (Figure 4.4). The AVF and WGB tend to include a greater proportion of an enriched source than the TTVF (and the BHVF and the ACA in the

previous subsection); the AVF and WGB have higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ than the TTVF. Lead isotopes vary significantly less between fields, and all overlap with the Enriched Mantle II reservoir and lie above the Northern Hemisphere Reference Line for oceanic basalts (Figure 4.5). There are no oxygen isotopic data for the WGB fields. The AVF rocks show a much stronger positive correlation between oxygen isotopes and silica or strontium isotopes than the TTVF rocks (for $\delta^{18}\text{O}$ and SiO_2 , $r^2_{\text{TTVF}} = 0.01$, and for $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$, $r^2_{\text{TTVF}} = 0.09$). However, if all the post-ancestral Cascades arc rocks (i.e., TTVF, and volcanic fields in Buffalo Valley, Upsal Hogsback, and Reno area) are grouped together, the correlations become much stronger (for $\delta^{18}\text{O}$ and SiO_2 , $r^2_{\text{post-ACA}} = 0.60$, and for $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$, $r^2_{\text{post-ACA}} = 0.65$).

5.2 Mantle sources

5.2.1 Asthenospheric versus lithospheric mantle source for the AVF magmas

Relative to the lithospheric mantle, the asthenospheric mantle has lower Ba/Nb, Ba/Zr, $^{87}\text{Sr}/^{86}\text{Sr}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and higher $^{143}\text{Nd}/^{144}\text{Nd}$, TiO_2 , and Nb/La (Fitton et al., 1991). The intraplate “Ocean-Island Basalt-OIB”-like source of the Basin and Range province produces lavas that are mildly alkaline, no HFSE depletion, and have lower Sr- and higher Nd-isotopic ratios (Fitton et al., 1991; Rogers et al., 1995; DePaolo and Daley, 2000). Key geochemical features in the AVF lavas are compared with two nearby late Cenozoic volcanic fields from the Basin and Range province that have intraplate-like, alkaline asthenospheric mantle sources: the Cima volcanic field (southeastern California; Farmer et al., 1995), and the Buffalo Valley volcanic field (northern Nevada; Cousens et al., 2013) (Figure 5.2).

Primitive mantle normalized multi-element diagrams show that the AVF rocks have distinct multi-element patterns distinct from MORB and OIB-like basalts typical of asthenospheric mantle sources. Compared with the primitive mantle, the AVF rocks are enriched in incompatible elements, like the OIB-like source rocks, but are depleted in Nb and Ta, and have a slight negative Ti anomaly (Figure 5.2). The AVF rocks are also enriched in Ba and Pb, with a slight positive Sr anomaly. The AVF rocks also have distinctly different isotopic signatures from asthenospheric mantle sources; the AVF rocks plot at higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}\text{Pb}/^{204}\text{Pb}$, and lower $^{143}\text{Nd}/^{144}\text{Nd}$ than MORB or OIB-like basalts (Figures 4.4 a, b; Figures 4.5 a, b). These incompatible element and isotope ratio trends are evident on asthenosphere-lithosphere source discrimination plots on which the AVF rocks tightly cluster in the lithospheric mantle source field with higher Ba/Nb, Ba/Zr, $^{87}\text{Sr}/^{86}\text{Sr}$, and lower $^{143}\text{Nd}/^{144}\text{Nd}$, TiO_2 , and Nb/La than the Cima or Buffalo Valley fields (e.g., Figures 5.3 a, b).

The basaltic andesite of Locomotive Point, Tmlp, tends to plot closer to an asthenosphere mantle source in isotopic ratio plots (i.e., lower $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$, $^{207}\text{Pb}/^{204}\text{Pb}$, higher $^{143}\text{Nd}/^{144}\text{Nd}$), and has less abundant incompatible trace elements, possibly indicating a greater asthenospheric mantle component. However, this unit plots near the center of the cluster of AVF samples in any incompatible trace element asthenosphere-lithosphere source discrimination plots (e.g., Figure 5.3b). The Tmlp unit is geographically separated from the rest of the AVF units (Figure 1.2), and so this slight difference in geochemical signature may indicate some heterogeneity in the lithospheric mantle source.

Heterogeneity in the lithospheric mantle is also observed in xenoliths from the region (e.g., Beard and Glazner, 1995). Therefore, the asthenosphere has made little, if any, contribution to the AVF geochemistry.

Figure 5.2 Primitive mantle-normalized (Sun and McDonough, 1989) trace element variation diagrams for AVF samples (< 55wt. % SiO₂ only basalt and andesite rock compositions) compared to mantle sources.

The minimum and maximum values for Aurora, Cima, and Buffalo Valley volcanic fields are included in this plot. The Cima and the Buffalo Valley volcanic fields are similar to ocean island basalts. The reference fields are from Sun and McDonough (1989): ocean island basalts, OIB, enriched and normal mid-ocean ridge basalts, E-type MORB and N-type MORB respectively, and primitive mantle.

Figure 5.3 Asthenosphere-lithosphere discrimination plots

These plots show rocks characteristic of lithospheric mantle sources (e.g., AVF and BHVF; Big Pine volcanic field and Long Valley volcanic field, in inset) and asthenospheric mantle sources (e.g., Buffalo Valley region rocks, in inset); a dashed grey line separates the mantle sources. Only (trachy)basalts and basaltic (trachy)andesites <55 wt. % SiO_2 are plotted. Nb, La, Ba, and Zr are primitive mantle normalized according to Sun and McDonough, 1989. (a) Plot of $(\text{Nb}/\text{La})_N$ vs. $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ showing the AVF rocks overlapping with the BHVF, the TTVE, and the Big Pine and Long Valley volcanic fields. Tmlp has lower $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$, however, also has low $(\text{Nb}/\text{La})_N$ and overlaps with other lithospheric mantle source rocks instead of overlapping with asthenospheric source rocks like the Buffalo Valley volcanic field rocks. (b) The AVF clusters in the lithospheric mantle source field with higher $(\text{Ba}/\text{Zr})_N$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ than the asthenosphere mantle source field.

5.2.2 Source of arc-like geochemical signature within the Walker Lane

The AVF, the BHVF, the TTVF, and the WGB fields all show arc-like trace element signatures, with enrichment of LILE (e.g., Ba, Pb, K, Sr) relative to HFSE (e.g., Ta, Nb, Zr, Hf, Ti). An arc-like signature is expected for the BHVF and the rest of the ACA, because it formed due to the subduction of the Juan de Fuca plate (remnant of the Farallon plate) under the North American plate. However, subduction ended at the latitude of BHVF and AVF at ~10 Ma as the trailing southern edge of the Juan de Fuca plate and the Mendocino triple junction migrated northwards, replacing the destructive plate margin with the San Andreas transform system (Atwater, 1970, 1973; Atwater and Stock, 1998). The BHVF continued to produce magma until 8 Ma, and then renewed activity briefly at ~6 Ma. After another ~1.5 Ma hiatus, volcanism recommenced at 3.9 Ma, resulting in the formation of the AVF (John et al., 2012). Without a subducting slab to add volatiles to a mantle wedge for at least 5 Ma prior to the eruption of the AVF, why would the AVF have an arc signature in its incompatible trace element patterns?

Ormerod et al. (1991) showed that the negative Nb, Ta, and Ti anomalies (subduction signature) and the low Rb/Sr ratios in lavas from the WGB were independent of the degree of melting, and concluded that these were characteristics of the magma source region. This chemical arc signature is not restricted to volcanic rocks along the eastern margin of the Sierra Nevada; this trend is characteristic of Cenozoic basalts throughout the western USA (Fitton et al., 1988, 1991; Farmer et al., 1989; Kempton et al., 1991; Ormerod et al., 1991; Bradshaw et al., 1993; Cousens et al., 2008). This pervasive, regional-scale pervasive signature is interpreted to originate from lithospheric mantle that

was metasomatized by fluids associated with either early rifting (younger than 820 Ma; Beard and Glazner, 1995) or longstanding subduction (possibly Precambrian, but at least Mesozoic through until late Cenozoic) beneath the western margin of the U.S.A. (e.g., Ormerod et al., 1988). Partial melts of this metasomatized lithospheric mantle source have incompatible element patterns that resemble subduction-derived magmas (e.g., volcanic fields of the WGB). The lithospheric mantle under the AVF would have been metasomatically enriched in the same manner, and could produce similar incompatible element patterns from its partial melts.

Although the Pliocene and younger WGB volcanism (e.g., Long Valley and Big Pine volcanic fields) is interpreted to have formed in an extensional environment (i.e., Owens Valley, California, is a rift valley), rocks in this region do not show a depleted, upper mantle isotope signature, and instead have isotopic trends indicative of an enriched mantle source (i.e., high Sr and low Nd isotope ratios compared to depleted mantle, DMM). Ormerod (1988, 1991) and Cousens (1996) suggested that the WGB source was a metasomatized lithospheric mantle with contributions from the asthenosphere (i.e., intraplate or within-plate ocean island basalt). The existence of this enriched, metasomatized lithospheric mantle source was confirmed analysis of mantle xenoliths from the Big Pine field (Beard and Glazner, 1995).

5.2.3 Melting mantle xenoliths: partial melt model and degree of melting

A mantle partial melting model was developed using REE concentrations to evaluate if the metasomatized lithospheric mantle xenoliths from Big Pine volcanic field (eastern

California, USA) could be the source of the most mafic lava in the AVF. The model utilizes equation 11 in Shaw (1970) based on batch melting, where the liquid remains in equilibrium with the residual source rock until the melt separates and rises, and minerals are melted in the same proportions as their abundance in the source rock:

$$\frac{C_L}{C_o} = \frac{1}{[D_o + F(1 - D_o)]}, \quad (7)$$

where C_L is the concentration of a trace element in the melt, C_o is the concentration of the trace element in the original source rock, D_o is the bulk distribution coefficient of that element in the original mineral assemblage in the source rock, and F is the weight fraction of the melt produced.

Excel Solver was used to find the optimal mineral assemblage to form the most mafic unit within AVF (highest MgO and Mg-number, lowest SiO₂, i.e., Tbch, trachybasalt of Cedar Hill). The sum of the weighted squared differences between calculated partial melt for a particular mineral assemblage and average observed value (from Tbch rock samples) was minimized, using variances of the samples as weights. In order to ensure that the mantle source was either a spinel-bearing peridotite or garnet-bearing peridotite, additional constraints were imposed on Excel Solver (olivine $\geq 40\%$, and either spinel or garnet $\geq 3\%$). Mineral-melt partition coefficients for basaltic and basaltic andesite liquids were taken from the compilation in Rollinson (1993) and GERM (2014). The mineral phases used were olivine, clinopyroxene, hornblende, magnetite, plagioclase, apatite, spinel, and garnet.

The mantle xenolith that fit best this model is an enriched wehrlite from Big Pine volcanic field metasomatized lithospheric mantle (sample Pi-2-5; Beard and Glazner, 1995). This peridotite has a steep negative LREE slope (i.e., La/Sm) that matches the REE slope of the AVF rocks. Also, invoking a moderate degree of partial melting (e.g., 10%, Figure 5.2) allowed a better fit with the LREE pattern of AVF rocks. The flat HREE slope of the AVF samples made it difficult to involve a peridotite with significant amount of residual garnet (e.g., >7% garnet; see section 4.4 for more details on HREE fractionation in garnet) and required a higher F value. However, the melting curves matched best with minimal to no residual garnet (i.e., 3%, Figure 5.2a). Excluding hornblende from the melted mineral assemblage resulted in rather high F values (over 15%) compared to the Putirka and Busby (2007) model (<6% melting for the WGB, 6-10% for TTVF and ACA; Putirka and Busby, 2007; Cousens et al., 2008), indicating the need for a metasomatized mantle source for the AVF rocks. Increasing the percent of spinel of a spinel-bearing mantle source changed the curves slightly but requires decreasing the F value; at least a 10% spinel content is needed if the degree of partial melting is below 6% (Figure 5.2b).

Thus, the AVF model falls within the low degree of partial melting range calculated for the Tahoe-Reno region and the ACA rocks (6-10%). The lower degree of partial melting would produce a smaller volume of magma (Harry and Leeman, 1995). Lange and Carmichael (1996) estimated that the AVF produced a magma volume of only $\sim 25 \text{ km}^3$ of erupted lava.

Figure 5.4 Melting model of a mantle xenolith compared with the REE pattern of the AVF.

An enriched wehrlite from Big Pine region (sample Pi-2-5; Beard and Glazner, 1995) was used in the equilibrium melting model of Shaw (1970) and compared with the trachybasalt, Tbch, of the AVF. The mineral-melt partition coefficients for basaltic and basaltic andesite liquids were taken from the compilation in Rollinson (1993) and GERM (2014). The melting model shows that a spinel- or a garnet- bearing peridotite can fit the data for the Tbch. A) 6.1% partial melt of a garnet-bearing peridotite (61.2% olivine, 21.1% clinopyroxene, 14.8% hornblende, and 3% garnet). B) 8.7% partial melt of a spinel-bearing peridotite (40% olivine, 48.4% clinopyroxene, 8.6% hornblende, and 3% spinel).

The multi-element incompatible trace element pattern, isotopic ratios, and trace element ratio asthenosphere-lithosphere discrimination plots coupled with the partial melt model of the lithospheric mantle xenolith indicate that the AVF magmas were derived from the lithospheric mantle.

5.2.4 Depth of source melting

The depth of the mantle source of AVF basalts can be estimated using the Al-mineral phase stability (i.e., presence of spinel or garnet within mantle depths; Figure 5.4), and the boundary between lithospheric and asthenospheric mantle. The garnet-spinel transition zone is between 70 and 85 km, with garnet stable at greater depths and spinel at shallower depths (Longhi, 1995; Kinzler, 1997; Robinson and Wood, 1998). The asthenosphere-lithosphere boundary beneath the western United States is approximately 70-75 km depth (DePaolo and Daley, 2000; Li et al., 2007; Putirka et al., 2012). Within the western U.S., the garnet-spinel and asthenosphere-lithosphere boundaries are roughly coincident. Because the AVF source rock has little to no garnet (i.e., no HREE depletion in AVF magmas, Figure 4.2a; modeling with Tbch in section 5.2.2), the 70-75 km boundary is a maximum depth of mantle source melting.

Figure 5.5 Stability fields for plagioclase-, spinel-, and garnet-peridotites using the peridotite KLB-1 of Takahashi (1986).

5.2.5 Mantle source of enriched isotopic signature

The $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth was originally defined based on the Mesozoic plutonic rocks and on the Proterozoic basement rocks with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} > 0.706$ and Phanerozoic basement rocks with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} < 0.706$ (Kistler and Peterman 1973; 1978). This line was also reflected in the Mesozoic mafic volcanic rocks, as well as recent mafic alkaline volcanic rocks implying that it separates lithospheric provinces as well as crustal

provinces from one another (see section 1.2.1; Leeman, 1982; Ormerod et al., 1988; Kempton et al., 1991). Rogers et al. (1995) noted that there were no discernable differences in major and trace contents of the basalts on either side of the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth except that those erupted through the Proterozoic basement tend to have lower silica contents, and isotopic values with higher $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$. If the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth extends into the lithospheric mantle, the highly enriched radiogenic isotope signature ($^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} > 0.706$) in the AVF cannot be derived purely from a Phanerozoic lithospheric mantle source as defined (i.e., $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} < 0.706$). The rate of decay for the parent radiogenic isotope is based on the atomic concentration of the element and the amount of time to decay. The concentrations of Rb to Sr is converted to the ($^{87}\text{Rb}/^{87}\text{Sr}$) ratio by the equation

$$\left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)_{\text{atom}} = \left(\frac{\text{Rb}}{\text{Sr}}\right)_{\text{conc.}} \times \frac{\text{Ab} \times ^{87}\text{Rb} \times \text{WSr}}{\text{Ab} \times ^{86}\text{Sr} \times \text{WRb}} \quad (8)$$

where ^{87}Rb and ^{87}Sr is the number of atoms in each respective isotope in a unit weight of a mineral at the present time, and (Rb/Sr) is the ratio of the measured concentrations of these elements. The time for the parent radiogenic isotope, ^{87}Rb , to decay is taken into account in following equation

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{\text{int.}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{\text{pres.}} - \frac{^{87}\text{Rb}}{^{86}\text{Sr}}(e^{\lambda t} - 1) \quad (9)$$

where $^{87}\text{Sr}/^{86}\text{Sr}_{\text{int.}}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{pres.}}$ are the initial and present $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios respectively, $^{87}\text{Rb}/^{87}\text{Sr}$ is calculated from equation 8, λ is a decay constant of ^{87}Rb (1.42×10^{-11} in years⁻¹), and t is the time in years since time of formation (Faure and Mensing, 2005). Using the above equations 8 and 9, and the concentrations of a metasomatized

lithospheric mantle (i.e., Rb of 0.002-0.32 ppm and Sr of 8.88-288.00 ppm found in the xenoliths from Big Pine volcanic field; Beard and Glazner, 1995) the maximum change in $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ to $^{87}\text{Sr}/^{86}\text{Sr}_{\text{present}}$ is only 0.000368 over 248 Ma (i.e., from the beginning of the Mesozoic to present). This maximum value added to $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.706 creates a maximum $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.706368, while the AVF $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ ranges from 0.705109 to 0.706933. Therefore, the AVF enriched radiogenic isotope signature must be derived from or include components from an ancient source that has a slightly higher Rb/Sr and lower Sm/Nd.

5.3 Differentiation

5.3.1 Fractional crystallization

Elemental trends for the intermediate composition AVF rocks suggest fractional crystallization of olivine, clinopyroxene, hornblende, and titanomagnetite. Concentrations of compatible major element oxides and trace elements (i.e., MgO, FeO^{T} , TiO_2 , MnO, $\text{CaO}/\text{Al}_2\text{O}_3$, Co, Ni, V, Sc) decrease linearly with increasing silica. Only the Sr abundances and $\frac{\text{Eu}}{\text{Eu}^*}$ decrease at high silica contents, indicating that plagioclase is a major fractionating phase later than anticipated from the petrography (see section 4.3). However, if fractional crystallization is the only source of differentiation for the AVF rocks, there would be no change in the isotopic ratio for any of the isotopes studied; isotopes are not fractionated by crystal-melt processes (see section 4.6.1). Instead, we see a wide range in radiogenic isotope ratios: $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.705109 to 0.706933, ϵ_{Nd} of -0.74 to -5.00, $^{206}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 18.974 to 19.246, $^{207}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 15.628 to 15.691, $^{208}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 38.671 to 38.954 (Figures. 4.4, 4.5). This implies that either different

AVF lava units are unrelated to each other but share the same process(es) of major and trace element differentiation, or they have incorporated variable types of material through open system processes such as magma mixing or assimilation.

5.3.2 Petrographic evidence indicating syn-magmatic open system processes

Open system processes give rise to an exchange of mass and energy between a magma body and its surrounding environment. This can include processes such as magma mixing and contamination/assimilation, or degassing of volatiles during magma ascent (e.g., CO₂, H₂O) where volatiles may be transferred from one magma to another magma at a shallower depth. There are several petrographic indications that the AVF lavas differentiated in a syn-magmatic open system manner: the presence of reversely zoned pyroxene, spongy cellulose sieve texture in plagioclase and pyroxene, xenocrystic quartz, and opacite rims on hornblende.

5.3.2.1 Reverse zoning

Rare reverse zoning in clinopyroxene crystals is prominently displayed as step zoning between green cores and colorless rims in at least two AVF units: crystals with discordant (anhedral) zoning with rounded core (sample 203247, Tpch; Figure 3.5) and a single megacryst with jaggedly, subhedral, sieved core (sample 08-BA-3, Qaca). Different color intensities noted through optical microscopy denote a strong compositional contrast, with the greater color saturation attributed to iron-rich pyroxene (e.g., Streck, 2008). Examples with the largest contrast are the “green core” clinopyroxenes from alkaline magmas (e.g., Brooks and Printzlau 1978; Barton and van Bergen 1981; Duda and Schmincke 1985;

Dobosi 1989; Barton et al. 1982). Similar compositional/color saturation contrasts are also apparent in sub-alkaline magmas (e.g., Anderson, 1974; Pe-Piper, 1984; Streck et al., 2007). In each of these examples, reverse zoning textures associated with step zoning are interpreted as signs of open system processes such as magma mixing (magma recharge) and/or contamination. These open system processes may have also influenced the AVF magmas with green core clinopyroxenes. Microprobe analyses confirmed a single reversely zoned clinopyroxene in Qmsa (Lange and Carmichael, 1996).

Reverse zoning was identified in plagioclase phenocrysts in at least two of the AVF units: Qta 1 and Qmsa (based on microprobe work by Lange and Carmichael, 1996). These reversely zoned plagioclase crystals also have pervasive sieve-textured cores and thin, clear rims (Lange and Carmichael, 1996). However, unlike reverse zonation in pyroxene, caution is needed when interpreting reverse zoning in plagioclase by open system behavior. One example of zoned plagioclase in a closed system is through convection within magma body, or “self-mixing” (Cox et al., 1979; Streck, 2008). If a sodic plagioclase crystallizes at the top of a magma chamber, and subsequently sinks to a lower level within the magma chamber, it experiences an increased temperature gradient. This causes resorption of the Na-rich crystal rim, which is soluble at higher temperatures, followed by renewed crystallization of a rim that is more calcic than its core (Singer et al., 1995). Self-mixing due to temperature differences may occur also by recharge (i.e., new magma input) of a magma reservoir with more mafic magma. Magma recharge would boost the temperature gradient, and provide a more calcic matrix for new plagioclase growth. Alternatively, the reversely zoned plagioclases could be xenocrysts.

5.3.2.2 Sieve texture

Spongy cellulose (i.e., sieve) textures in plagioclase and pyroxene are present in several units in the AVF and can indicate dissolution or resorption as a result of open system behavior in a magma body (Streck, 2008). Resorbed pyroxenes with a spongy cellular texture are a clear indication of strong textural and compositional disequilibrium resulting from internal melting of a crystal (Streck et al., 2007; Streck, 2008). On the other hand, plagioclase often exhibits a multi-stage growth and resorption history, leading to a much more complex array of textures (Pearce and Kolisnik, 1990; Johannes et al., 1994). Both coarse-grained (vermicular) or fine-grained (dusty) sieved textures are found in AVF plagioclase, whereas only the coarse-grained size is found for AVF pyroxenes.

Sieve texture can indicate the internal melting of a crystal and its xenocrystic nature, particularly if the crystal's sieved zone is well rounded. Heating above the mineral's solidus temperature (Figure 5.5) may cause dissolution of the phenocryst and rounding of its crystal edges (Pearce et al., 1987; Shelley, 1993). If the anorthite (i.e., Ca-rich plagioclase) content of the phenocryst is less than that of the plagioclase in equilibrium with the melt, the plagioclase crystal surface corrodes, becoming rough and finely indented, preferentially melting the albite (i.e., Na-rich plagioclase) component of the plagioclase first (Pearce et al., 1987; Shelley, 1993). Melt nucleates and collects at the crystal surface and at any cracks that connect to the crystal surface, creating a coarse-grained (vermicular) or fine-grained (dusty) texture. Anorthite-rich plagioclase is then precipitated from the melt at the plagioclase crystal edges (Johannes et al., 1994).

Plagioclase is more readily dissolved when it is in presence of hydrous fluid or exposed to pressures outside of its stability field (Figure 5.4).

5.3.2.3 Xenocrysts

Xenoliths and xenocrysts are foreign materials that have been incorporated into the magma. They may have been ripped off the walls of conduits or reservoirs during magma ascent, incorporated from the roof of a magma chamber, or rarely picked up as the lava flowed. However, their absence does not indicate the magma remained uncontaminated; xenoliths may react with the surrounding magma and completely dissolve, changing the chemistry of the magma.

In the AVF, there are two occurrences of a fine-grained aggregate mantled by pyroxene; these may be a heavily altered and/or resorbed xenocryst or xenolithic shale from the local basement rock (Tbch, 12-BA-32; Tmlp, 203195). Quartz xenocrysts are also conspicuously mantled by pyroxene and found in a range of rock types from basaltic andesite to trachydacite (Tmlp, 203184; Qta 1, 203398; Qacc, 203402; Figure 3.8).

5.3.2.4 Opacite rims

The crystal-melt reaction breakdown rim of opaque material on hornblende, commonly referred to as opacite, is an anhydrous assemblage of plagioclase, pyroxene, and magnetite (Rutherford and Hill, 1993). This breakdown reaction is caused by a relative loss in H₂O from hornblende and the coexisting melt due to pre-eruption degassing and

loss of volatiles in a shallow magma reservoir, or decompression-related disequilibrium resulting from slow ascent through the crust (Rutherford and Hill, 1993; Rutherford, 2008). Opacite texture, commonly found within the modern and ancient Cascades arc rocks, is attributed to degassing and/or decompression (e.g., Rutherford and Hill, 1993; Cousens et al., 2008); these processes tend to cause near-rim breakdown reactions in hornblende crystals (Rutherford, 2008). Additional open system processes (e.g., magma recharge), which may change the magma temperature, composition, and/or volatile content, can lead to a pervasive alteration of amphibole and may contain altered amphibole adjacent to unaltered amphiboles (Rutherford, 2008).

Nearly all hornblende crystals in AVF rocks have a thin opacite rim (~0.05 mm), suggesting decompression during magma ascent. Although smaller crystals such as microphenocrysts or groundmass crystals show pervasive replacement to opacite, this is likely a function of size, as opposed to an indication of multiple populations of hornblende.

Opacite rims also locally mantle biotite crystals (samples 13-AU-13, 13-AU-20, and 13-AU-21 of unit QThch). Biotite is also sensitive to the amount of water in the melt and will become unstable as the magma devolatilizes (e.g., Wones and Eugster, 1965).

5.3.3 Magma recharge

The AVF lava compositions oscillate between lower and higher silica compositions with time (Figure 5.1). As magmas evolve to higher silica compositions through the

fractionation of mafic components as it cools and crystallizes, fresh mafic magma must be introduced in order to lower the silica content. However, it is unlikely that the AVF units were all derived from a single magma reservoir or chamber.

Although the petrographic textures indicate a change in magmatic conditions, there are no mafic enclaves or mingled magma in the AVF lavas. The glomerocrysts in the trachydacite of Aurora Peak (unit Tdap) are the only petrographic evidence that may indicate magma mixing within the AVF. Additionally, a large temporal (3.9 – 0.1 Ma) and spatial (~38 km x ~24 km; Figure 1.2) range between the AVF units, coupled with a low volume of volcanic material produced (~25km³; Lange and Carmichael, 1996) suggest that small amounts of magma likely cooled and crystallized at depth before the next magmatic pulse. Each magmatic pulse will have a specific starting composition based on the source rock (e.g., 49% the lowest silica composition for the AVF), and may differentiate to a different degree before it cools and crystallizes. This creates an array of compositions within the AVF. The silica composition lowers again with a new magmatic pulse from the source rock. The repetitive pulses of new magma batches coupled with the variable differentiation of each magma batch create the oscillating silica compositions over time.

Lange and Carmichael (1996) proposed a model of differentiation from basalt to dacite by repeated partial melting of basaltic magma, which had accumulated and crystallized at the lower crust-mantle boundary. In light of the new isotopic data for the AVF, this model cannot explain the process by which the magmas were differentiated; repeated

partial melts from one source would show no change in isotopic signature for the AVF. Instead, a range in isotopic signature indicates mixing between multiple sources.

Each batch of magma may not only have differentiated to a differing degree and may also incorporate variable amounts of surrounding crustal rock to produce the range in isotopic signature of the AVF.

5.3.4 Crustal sources

As lithospherically-derived magma passes upward through the crust to reach the surface, there is a possibility of interaction between the magma and the crust. Pieces of crustal rock may be picked up, incorporated into, and dissolved by the magma (bulk assimilation), or the magma may absorb part of the crustal rock as it passes through it (selective assimilation). Another possibility is that the magma incorporated a partial melt of the surrounding country rock (contamination). Any one of these processes, or a combination thereof, will change the composition of the magma, unless the added crustal material has the same composition. If the magma has a different composition, phenocrysts may become unstable, creating reverse zonation and inclusions enveloped by mantles/coronas (noted above in sections 5.2.1.1 and 5.2.1.3). However, there may only be chemical evidence that such processes took place.

5.3.4.1 Chemical evidence of crustal signature: local contamination

Th, U, Cs, and Rb are incompatible elements that are abundant in rocks that have undergone extreme crystal fractionation, such as rhyolite or granite. Although Rb and U

can be very mobile in fluids, Th is immobile. Enrichment of Th, U, Cs, and Rb is uncommon in andesites, suggesting that AVF andesites with high Th-U-Cs-Rb contents have been contaminated by a granitoid.

The high Th-U-Cs-Rb AVF andesites (samples 203181, 203246, 203247, and 203354 from Tpch; 203232 from Qacc; and 13-AU-05 from Tach; section 5.1.1; indicated by the yellow field in Figures 4.1o) are not limited to a single AVF unit, nor do they have similarities in abundances of Th, U, Cs, and/or Rb to other samples in these units (Tpch, Tach, and Qacc). Within Qacc, for example, the trachydacite sample (203402) has geochemical characteristics similar to other AVF rocks, whereas the basaltic trachyandesite sample (203230) has a transitional composition between the remainder of the AVF rocks and another source. This suggests heterogeneous mixing between the AVF rocks and a local contaminant.

The high Th-U-Cs-Rb AVF andesites show similarities to some samples of the BHVF trachyandesite of Del Monte and the trachydacite of Aurora Canyon; they overlap in the abundances of incompatible elements Th, U, Cs, Rb, and show the same narrow range in radiogenic isotope ratios ($^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.70550 to 0.70587, ϵ_{Nd} of -1.89 to -3.01, $^{206}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 19.067 to 19.104, $^{207}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 15.633 to 15.658, $^{208}\text{Pb}/^{204}\text{Pb}_{\text{initial}}$ of 38.783 to 38.878; Figures 4.4, 4.5; John et al, 2012; du Bray et al., in press). The Sierra Nevada Batholith (SNB) underlies both the AVF and the BHVF, and contains high abundances of Th, U, Cs, and Rb (e.g., Barbarin et al., 1989). Granitic pluton(s) from the

SNB could easily span the distance between the vents for Tpch, Tach, and Qacc, and the two BHVF units, and acted as a contaminant to these units.

5.3.4.2 Contamination of the volcanic field

5.3.4.2.1 Cryptic contamination signature

Other AVF rocks show no obvious crustal contamination signature in their major and trace element compositions, although the presence of chemical disequilibrium is shown petrographically by reverse zoning and sieve textures in phenocrysts and mantled xenocrysts. As lavas evolve, they are more likely to show evidence of contamination, because they are more likely to have a longer residence time within the continental crust (Rollinson, 1993). Progressive enrichment of incompatible trace elements is expected for lavas that have evolved by fractional crystallization and incurred assimilation of crust due to longer crustal residency. However, incompatible trace elements (e.g., REE, Ba, Sr, Rb, Th, Nb) within the AVF rock suite show a large scatter and/or very shallow positive slopes when plotted against silica indicating little or no correlation with silica. Trace element ratios (e.g., Th/La, Rb/Sr) only discriminate the AVF rhyolites and high Th-U-Cs-Rb AVF andesites (discussed in section 5.1.1 and 5.3.4.1). The Ce/Pb ratio shows a positive correlation with silica and has been used as an index of crustal contamination, but is inconclusive on its own; the Ce/Pb ratio is also used as a slab-derived sediment and/or fluid index, and may indicate a remnant mantle metasomatic signature in the mantle source. If the Pliocene-Pleistocene mafic magmas already have a subduction-like trace element signature, as the mafic rocks in the Sierra Nevada region imply (e.g.,

Omerod et al., 1988; 1991; Cousens, 1996; 2011; Feldstein and Lange, 1999), and the crustal rocks they erupt through also have subduction-like trace element signature (e.g., Barbarin et al., 1989; this study), then assimilation of crustal material may only be discernable in the isotopic compositions. This also necessitates that the trace element and/or isotopic ratios of crustal rocks are distinct from the starting mafic magmas to be distinguished through crustal contamination models.

Mantle-crustal mixing is supported by strong correlations in the AVF isotope covariate plots; crystal fractionation alone will not change the radiogenic isotopic ratios. A mixing relationship between two or more isotopic sources is further suggested by large variations in their isotopic signature within a single lava unit; for example unit QThc ranges from $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.70629 to 0.70687 and ϵ_{Nd} of -2.07 to -4.52 (Figure 4.4). Examples of heterogeneous crustal assimilation and mixing have been observed within other volcanic fields (e.g., Laughlin et al., 1972). A variety of rocks with different compositions may have been assimilated into the AVF magma, obscuring a simple mantle-crustal trace element mixing signature while displaying a conspicuous range in isotopic signature.

In addition to a cryptic geochemical signature, petrographic textures may also mask evidence for crustal incorporation. Beard et al. (2005) demonstrated with a conceptual model that bulk crustal assimilation may not greatly change the crystallizing mineral assemblage when partially melted xenoliths yield crystal phases that are already present in the host magma (e.g., plagioclase, pyroxene, Fe-Ti oxides, and hydrous melt). The cryptic nature of contamination may be further enhanced in hydrous magmas, like those

of the AVF that had 1.5 to ~6 wt. % H₂O (Lange and Carmichael, 1996), whereby new hydrous minerals may be crystallized in response to the reaction between hydrous melt and anhydrous minerals causing the reverse reaction of dehydration melting (Lange and Carmichael, 1996; Beard et al., 2005). This reaction, known as “hydration crystallization,” may affect any anhydrous phase in the system: phenocrysts, xenocrysts, and/or crystals with a hybrid signature and would obscure or erase evidence the mixing record (e.g., zoned pyroxenes; Beard et al., 2004; Beard et al., 2005). Late-stage formation of hornblende at the expense of clinopyroxene is evident in some of the AVF magmas where pyroxene is mantled by hornblende or hornblende contains pyroxene inclusions (thin section petrography of units Qmsa, QThch, Thd; Appendix B).

5.3.4.2.2 Enriched source of AVF isotopic signature

The AVF, the BHVF, and the TTVF overlie the same Precambrian-Phanerozoic lithospheric mantle, delineated by the ⁸⁷Sr/⁸⁶Sr isopleth (i.e., ⁸⁷Sr/⁸⁶Sr_{initial} = 0.706 line; see Figure 1.3). Although the BHVF, the AVF, and the TTVF all show strong, negative correlations between radiogenic Sr and Nd isotope ratios, most of the AVF samples have higher initial ⁸⁷Sr/⁸⁶Sr and lower ¹⁴³Nd/¹⁴⁴Nd ratios than those of the BHVF or the TTVF. Therefore, this correlation suggests an older and/or more enriched source for the AVF than that of the BHVF or the TTVF. The AVF also has a stronger positive correlation between ⁸⁷Sr/⁸⁶Sr and SiO₂ ($r^2_{\text{AVF}} = 0.24$ versus $r^2_{\text{BHVF}} = 0.20$; Figure 4.4a), and between ⁸⁷Sr/⁸⁶Sr and the Pb isotope ratios (moderate to strong, $r^2_{\text{AVF}} = 0.56\text{--}0.65$; weak to moderate, $r^2_{\text{BHVF}} = 0.15\text{--}0.43$); these correlations indicate that the AVF rocks contain a larger crustal component than the BHVF rocks. Many of the basement rocks analyzed in

this study and in Barbarin et al. (1989) have high initial $^{87}\text{Sr}/^{86}\text{Sr}$ and low ϵ_{Nd} (Figure 4.4) and could modify the isotopic signature of the AVF magmas (increase the initial $^{87}\text{Sr}/^{86}\text{Sr}$ and lower the ϵ_{Nd}) particularly if the concentrations of REE and Sr in the crust are greater than in the lithospheric mantle by several factors.

The AVF and the BHVF not only share the same lithospheric mantle source, but they also overlie the same basement rocks. The BHVF and the AVF fall within the narrow Pb isotopic ratio range of the SNB ($^{206}\text{Pb}/^{204}\text{Pb} = 18.20$ to 20.06 , $^{207}\text{Pb}/^{204}\text{Pb} = 15.58$ to 15.76 , $^{208}\text{Pb}/^{204}\text{Pb} = 38.44$ to 39.61 ; Doe and Delevaux, 1973; Chen and Tilton, 1991). Likewise, the Pb isotope values of the WGB (e.g., Long Valley and Big Pine volcanic fields), and the TTVF also fall within this range. High U and Th concentrations are common in highly fractionated magmas, such as rhyolites and granites. The SNB has high concentrations of U and Th (up to 11.7 ppm and 27.4 respectively; Chen and Tilton, 1991), such that assimilation of these rocks and respective Pb isotope signatures can drown out the original mantle signature for any of these volcanic fields.

Both radiogenic and stable isotope ratios are sensitive to mantle versus crustal sources (Rollinson, 1993). Initial $^{87}\text{Sr}/^{86}\text{Sr}$ correlates positively with $^{18}\text{O}/^{16}\text{O}$ (excluding Trsp and Tbch samples; these samples will be addressed section 5.4) and both isotope ratios correlate positively with silica contents, suggesting that the AVF has incorporated crustal material. The AVF shows a larger increase in $\delta^{18}\text{O}$ than expected for closed-system crystal fractionation (0.5‰ per 10 wt. % SiO_2 ; Taylor and Sheppard, 1986); aside from values below the primitive mantle value of $+5.7 \pm 0.3\text{‰}$ (Taylor and Sheppard, 1986;

i.e., Trsp, +4.01‰), the AVF has a range of ~2.5‰ (+6.41 to +8.96‰) over a range of ~30wt. % SiO₂ (49.00 to 77.61 wt. % SiO₂), also suggesting interaction with crustal material.

5.3.4.3 Modelling crustal contributions

Metasomatism of a lithospheric mantle source, as a result of subduction, may produce isotopic signatures that resemble those of the crust (e.g., the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} > 0.706$ signature of enriched lithosphere underlying the Precambrian continental basement in the Big Pine region; Beard and Glazner, 1995). Therefore, isotopic discrimination diagrams alone are insufficient to show the source of the isotopic signature. Crustal and mantle contributions can be modelled through processes of end-member mixing curves and crustal assimilation combined with fractional crystallization (AFC) to test if the isotopic signature may be derived by crustal contribution, and if mantle-crustal end member mixing or AFC are possible models for differentiation of the AVF. If either model is applicable, the amount of crustal contribution may be estimated.

5.3.4.3.1 Mantle-crust mixing curves

Several Sierra Nevada intrusive rocks have been interpreted as the result of mantle-crustal end member mixing (Reid et al., 1983; Kistler et al., 1986; Frost and Mahood, 1987; Sisson et al., 1996). Linear correlations on isotope ratio plots (e.g., Figures 4.4, 4.5, 4.8b) suggest that mixing between mantle and crustal sources may be a viable differentiation process to produce the AVF rocks. Therefore, mantle-crust end-member

mixing curves were used to test crustal incorporation in the AVF magmas. Covariant isotope plots were used to order to produce end member mixing curves that exclude crystal-melt processes (e.g., fractional crystallization; Figure 5.5). This necessitates a mantle end member that is isotopically depleted relative to the AVF rocks, and a crustal end member that is isotopically enriched relative to the AVF rocks. Although the xenoliths of the Big Pine volcanic field were used in the partial melt model, all the lithospheric mantle xenoliths have isotopic values similar to the AVF rocks (see Figure 4.4). Therefore, the mafic end members used in this model were the most mafic unit of the BHVF, Trwai (Rancheria basaltic andesite diatreme) and the most isotopically depleted mafic unit of the AVF, Tmlp (basaltic andesite of Locomotive Point). The crustal end members used in this model were the SNB granitoid rocks from Barbarin et al. (1989) and local basement rocks from the Bodie Hills region (2 Paleozoic? sedimentary rocks, 2 Jurassic? metavolcanic units and 3 83-95 Ma granitic units from local SNB). The bulk mixing curve was calculated by incrementally adding a crustal assimilant to the parental mafic melt using the formulas

$$\left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} \right)_M = \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} \right)_A \times f_A \left(\frac{\text{Sr}_A}{\text{Sr}_M} \right) + \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} \right)_B \times (1 - f_A) \left(\frac{\text{Sr}_B}{\text{Sr}_M} \right) \quad (10)$$

and

$$\left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_M = \left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_A \times f_A \left(\frac{\text{Nd}_A}{\text{Nd}_M} \right) + \left(\frac{{}^{143}\text{Nd}}{{}^{144}\text{Nd}} \right)_B \times (1 - f_A) \left(\frac{\text{Nd}_B}{\text{Nd}_M} \right), \quad (11)$$

where $({}^{87}\text{Sr}/{}^{86}\text{Sr})_M$ and $({}^{143}\text{Nd}/{}^{144}\text{Nd})_M$ are the isotopic ratios of a mixture, $({}^{87}\text{Sr}/{}^{86}\text{Sr})_A$ and $({}^{143}\text{Nd}/{}^{144}\text{Nd})_A$ are the isotopic ratios of component A (e.g., mantle end member), $({}^{87}\text{Sr}/{}^{86}\text{Sr})_B$ and $({}^{143}\text{Nd}/{}^{144}\text{Nd})_B$ are the isotopic ratios of component B (e.g., crustal end

member), f_A is the weighing factor expressed as the abundance of component A of a given mixture (where the weighing factor of component A and of B together sum to 1.0), Sr_A and Nd_A are the elemental concentrations of component A (in ppm), Sr_B and Nd_B are elemental concentrations of component B (in ppm), and Sr_M and Nd_M are the weighted elemental concentrations of the mixture for example, $Sr_M = Sr_A f_A + Sr_B (1 - f_A)$ likewise $Nd_M = Nd_A f_A + Nd_B (1 - f_A)$. Isotopic ratios of the crustal rocks were corrected to 4 Ma, approximately when they could have been incorporated into the AVF magma.

The basement granitoid rock that best followed a linear $\epsilon_{Nd} - ^{87}Sr/^{86}Sr$ mixing trend of the AVF is the Mount Givens granodiorite (sample MHb3-H; Barbarin et al., 1989).

Although a mixing curve can be derived from a melt of either Tmlp (sample 203184) or Trwai (sample 203173) and an assimilated of the granodiorite (sample MHb3-H), both mantle end members create a mixing curve where Tbch is over 60% contaminated (~90% contamination in Figure 5.6a). The Pb-Pb isotopic plots also show Tbch as the most enriched AVF rock (Figure 4.5). Even with other mixing pairs, Tbch is at best shown at ~59% contaminated, but they necessitate multiple crustal contaminants to fit the whole AVF data suite (e.g., a plot of ϵ_{Nd} versus Ce/Sm with end members Big Pine volcanic field xenolith sample IN9220H-4 and a SNB granitoid sample MHb3-H, Figure 5.6b).

This high proportion of contamination is unreasonable for a trachybasalt. Models of heat and mass transfer in magma-country rock system at shallow crustal depths (low pressure of 1 kbar) showed a maximum 5-18% assimilation in basalts while restricted magma differentiation (change in SiO_2 or MgO of <0.3 wt. %) and limiting olivine crystallization (5-7%; Reiners et al., 1995). In addition to its low silica content, Tbch has the lowest

Pb/Ce and K_2O/P_2O_5 , and low Rb/Sr, further limiting the amount of possible contamination of this unit.

Figure 5.6 Mantle-crust end member mixing plots.

(a) Plot of ϵ_{Nd} vs. $^{87}Sr/^{86}Sr_{initial}$ between an AVF basaltic andesite (representing mantle proxy end member) and a granodiorite from the SNB (crustal end member). The black mixing curve is between the AVF basaltic andesite, Tmpl (sample 203184), and the Mount Givens Granodiorite (sample MHb3-H; Barbarin et al., 1989). (b) Plot of $^{143}Nd/^{144}Nd$ vs. Ce/Sm showing multiple crustal end members are required to fit the AVF data. Tick marks on the mixing curves every 20% crustal component added.

5.3.4.3.2 Assimilation Fractional Crystallization (AFC)

The crustal assimilation and fractional crystallization (AFC) model can yield significantly different results than end member mixing models; however, it is a more accurate and more complex contamination model. The AFC model involves heat balance and incorporates the combined influence of concurrent assimilation and fractional crystallization; the latent heat of crystallization during fractional crystallization drives assimilation of country rock, which further drives crystallization of the host magma (Taylor, 1980; DePaolo, 1981). This model also accounts for situations where country rock surrounding slow or stalled magma may be significantly heated, especially rocks that lie above a magma chamber (DePaolo, 1981).

Modeling AFC was based on equations developed by DePaolo (1981):

$$\frac{C_L}{C_0} = f' + \left(\frac{r}{r-1+D} \right) \left(\frac{C_A}{C_0} \right) (1-f') \quad (12)$$

and

$$f' = F^{-\left(\frac{r-1+D}{r-1} \right)} \quad (13)$$

where the change in concentration of a trace element is dependent on the fractionation of an element between the crystallizing minerals and the magma (bulk distribution coefficient, D), the ratio of assimilation rate to the fractional crystallization rate (r), the concentration of the trace element in the magma (C_0) and in the assimilated country rock (C_A), and the mass fractionation of the remaining magma (F).

The axes of bivariate plots were chosen such that one axis would be sensitive to assimilation but not to fractional crystallization and the opposite for the other axis (e.g., ϵ_{Nd} versus Ce/Sm, Figure 5.6b). To broaden the range of potential assimilants, mafic inclusions, aggregates, dikes, and their host granitoid rocks from the central SNB (data from Barbarin et al., 1989), in addition to basement rocks from the Bodie Hills region (data from this study, John et al., 2012, and du Bray et al., 2016) were used.

A two-step AFC model was used. The starting mafic end member, representing a starting magma composition in the first step, is the unit Tmlp (sample 203184), which is the most isotopically depleted AVF unit and one of the most mafic units in this volcanic field. During the first step, olivine, clinopyroxene, $\pm$ plagioclase were fractionally crystallized out, using mineral percentages found in the AVF units and normalized to 100% (see Table 2 for mineral percentages used in each AFC model and corresponding bulk partition coefficients). The K_D values for the fractionating minerals were taken from alkali basalts found in the Geochemical Earth Reference Model, (GERM) (Schnetzler and Philpotts, 1970; Fujimaki et al., 1984; GERM, 2016). Multiple crustal end members were necessary to produce curves to fit the range in AVF data. The model with excluding plagioclase as a fractionating phase used lower F values.

In the second step of the AFC model, the starting magma composition was the trachyandesite from Aurora Crater (sample 13-AU-31), with mid-range isotopic and trace element values. During the second step, clinopyroxene, hornblende $\pm$ plagioclase were fractionally crystallized out. The mineral percentages used were based off the AVF units

or the literature, and then normalized to 100%. The K_D values for the fractionating minerals were taken from the trachyandesite found in GERM (Luhr et al., 1984; GERM, 2016). Again, the model with less plagioclase fractionation used lower F values.

As the mantle-crust end member model, multiple crustal contaminants must be used to model the range in the AVF data; different crustal assimilants were needed for each element and/or isotope ratio used for the AFC model. The ideal ratio of assimilation rate to the fractional crystallization rate (r ; equation 12) varied from 0.6 to 0.9, depending on the assimilant used and on the elements plotted in the model.

Table 2 Mineral percentages used in each AFC model and corresponding bulk partition coefficients.

	Step 1: Alkali basalt Kd		Step 2: Trachyandesite Kd	
	Ol, Cpx [%]	Ol, Cpx, Plag [%]	Cpx, Hbld [%]	Cpx, Hbld, Plag [%]
Ol	25	28	-	-
Cpx	75	12	29	4
Plag	-	60	14	68
Hbld	-	-	57	28

Above mineral percentages are based on the following AVF units or references:

Tmlp QTopb Qmsa Luhr et al., 1984

Abbreviations:

Ol = olivine, Cpx = clinopyroxene, Plag = Plagioclase, Hbld = hornblende

Element	Bulk partition coefficients (based on each model above)			
Ce	0.044	0.075	0.738	0.378
Nd	0.092	0.057	1.526	0.680
Sm	0.179	0.050	2.223	0.949

Note: Grey fields were used for AFC models in Figure 5.7.

Kd references:

Step 1: Ol and cpx Kd values are from Fujimaki et al. (1984), plag Kd values are from Schnetzler and Philpotts (1970).

Step 2: Cpx, hbld, and plag Kd values are from Luhr et al. (1984).

Figure 5.7 $^{143}\text{Nd}/^{144}\text{Nd}$ vs. Ce/Sm with AFC model curves.

A two-step AFC model. In the first step, the Tmpl (sample 203184) represents the starting magma composition (most isotopically depleted and one of the most mafic units of AVF), 25% olivine and 75% clinopyroxene were crystallized, and multiple assimilants were used. The K_D values for the fractionating minerals were taken from alkali basalts (see Table 2). In the second step, the Qaca (sample 13-AU-31) represents the starting magma composition (mid-range isotopic and trace element values), 29% clinopyroxene, 57% hornblende, and 14% plagioclase were crystallized. The K_D values for the fractionating minerals were taken from the trachyandesite (see Table 2). Tick marks on the AFC curves indicate F from 1 to 0.5, in increments of 0.1.

5.4 Special cases

5.4.1 High $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ and low ϵ_{Nd} signature in the trachybasalt, Tbch

The trachybasalt of Cedar Hill, Tbch, is the most mafic unit in the AVF and shows no obvious geochemical signature of crustal contamination in major element oxides or trace elements. This unit has the low silica, Rb/Sr , Pb/Ce , K/La , and $\delta^{18}\text{O}$ and high MgO , $\text{P}_2\text{O}_5/\text{K}_2\text{O}$, and Ni compared to all the other AVF units, as well as relative to the BHVF units. It lies west of the 0.706 line based on upper- and mid-crustal Mesozoic plutons

with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} < 0.706$ (Kistler and Peterman, 1973, 1978). The $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ boundary is interpreted to carry through to the lithospheric mantle (Ormerod, 1988; Kempton et al., 1991; Beard and Glazner, 1995). If this is true, the high $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.70693 in Tbch cannot be derived from the young, accreted lithosphere: an $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.706 cannot decay to 0.707 given the low concentrations of Rb that characterize the mantle (e.g., 0.002-0.32 ppm, Big Pine xenolith; Beard and Glazner, 1995) and the youthful age of the AVF (i.e., <4 Ma) (See also section 1.2.1 on $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth definition and section 5.2.5 showing that the AVF rocks greater than $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ 0.706368 were not derived from a lithospheric mantle source with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} < 0.706$ of Mesozoic age.) Thus Tbch must have been derived from or exchanged with source with higher concentrations of Rb and/or an older source that had enough time to decay to produce an $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ of 0.70693.

Partial melting of labile phases (e.g., micas, amphiboles) can radically shift the magmatic $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ without markedly changing the magmatic $\delta^{18}\text{O}$ values (Lackey et al., 2008; Rushmer and Knesel, 2011). Hornblende has a higher partition coefficient for Nd than biotite (e.g., r^2 is 4.260 versus 0.290; rhyolites from Arch, 1976), and can not only produce a melt with a high $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$, but also a lower ϵ_{Nd} signature. Melting and assimilation can be rapid processes that do not necessitate long residence times; as a result, contaminant compositions may deviate from those expected from equilibrium melting models (e.g., Kaczor et al., 1988). At low-pressure conditions within the upper crust, the dehydration breakdown reaction of high-Rb/Sr phases, such as micas and amphiboles, leads to partial melt with high $^{87}\text{Sr}/^{86}\text{Sr}$ (Kaczor et al., 1988). As mica and

amphibole are the first phases to melt and they have high concentrations of Rb (e.g., Knesel and Davidson, 1996), early small degrees of partial melts will have higher $^{87}\text{Sr}/^{86}\text{Sr}$ values than the whole rock value of the assimilated (Rushmer and Knesel, 2011). However, at higher pressures (e.g., 600 MPa at mid-crustal depths, ~20 km), the $^{87}\text{Sr}/^{86}\text{Sr}$ value of small degrees of partial melting has been observed to be lower than the whole rock value of the assimilated in melting experiments conducted by Rushmer and Knesel (2011). Therefore, the isotopic value of a partial melt not only depends on its source, but also on the pressure-temperature conditions under which it formed. The SNB contains both biotite and hornblende phases in mafic inclusions, dikes, aggregates and host granitoid rocks; however, the pressure conditions during crystal evolution for the AVF units are currently unknown.

Alternatively, the high $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ and low ϵ_{Nd} of Tbch may have been derived from the same mantle source as the mafic volcanic rocks in the Big Pine and/or Long Valley regions. The $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth is not a discrete boundary, and some interleaving of low- and high- $^{87}\text{Sr}/^{86}\text{Sr}$ sources is not unreasonable. Some researchers have shifted the “706 line” (i.e., the inferred edge of Precambrian crust) based on mafic volcanic rocks in the region (e.g., Rogers et al., 1995). The $^{208}\text{Pb}/^{204}\text{Pb} = 38.8$ isopleth (Tosdal et al., 2000) is also used to indicate the edge of the Proterozoic-Phanerozoic boundary, although it is often offset from the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.706$ isopleth.

5.4.2 Low $\delta^{18}\text{O}$ signature in the rhyolite, Trsp

The low $\delta^{18}\text{O}$ value (+4.01 ‰) in the Spring Peak rhyolite, Trsp, suggests it was derived from or exchanged with a precursor material that at one time interacted with meteoric water. Direct addition or mixing of meteoric water with magma at depth is improbable (Taylor and Sheppard, 1986). The $\delta^{18}\text{O}$ values can be lowered by pervasive post-magmatic, high-temperature (>300°C) alteration with a low $\delta^{18}\text{O}$ fluid. Although the samples collected were fresh and showed no evidence of hydrothermal alteration (personal observation; du Bray and John, personal communication, 2015), evidence of meteoric water infiltration has been observed within the BHVF (e.g., John et al., 2012; 2015; Vikre et al., 2015), including a region proximal to the vent for this rhyolite (Vikre et al., 2015).

The large alteration zones in the BHVF were dominantly formed by meteoric water and contain low $\delta^{18}\text{O}$ whole rock and quartz vein values (Vikre et al., 2015). The $\delta^{18}\text{O}_{250^\circ\text{C}}$ value of water in fluid inclusions in one vein sample is -12 ‰ (O'Neil and Silberman, 1974), which is comparable to the whole rock $\delta^{18}\text{O}$ values in hydrothermally altered Oligocene rhyolites in the Stillwater Range, Nevada (-9 ‰; John and Pickthorn, 1996). The amount mixture of hydrothermally altered rock with the Trsp can be estimated using the formula

$$\delta^{18}O_M = \delta^{18}O_A f + \delta^{18}O_B (1 - f) \quad (14)$$

where $\delta^{18}O_M$ is the $\delta^{18}\text{O}$ value of a mixture, $\delta^{18}O_A$ and $\delta^{18}O_B$ are the $\delta^{18}\text{O}$ value of component A and component B, and f is the abundance of component A of a given mixture. If the present day Trsp ($\delta^{18}O_M$, using equation 14) originally had a $\delta^{18}\text{O}$ similar

to the Martinez Hill rhyolite (whole rock $\delta^{18}O_A$ value of +8.70‰), only ~27% contamination is required to reduce the $\delta^{18}O$ to +4.01 ‰ using these meteoric water $\delta^{18}O$ values of the fluid inclusion in the BHVF or the Oligocene rhyolites in the Stillwater Range, Nevada ($\delta^{18}O_B$ values of -9 ‰). An exchange (e.g., assimilation of hydrothermally altered rock) between the Trsp magma and hydrothermally altered BHVF plutons at depth is possible.

A smaller alteration zone, Spring Peak sinter (Hg), is within 1.5 km of the Trsp volcanic vent and contains altered rhyolite fragments, which may be Trsp (Vikre et al., 2015). However, detailed mapping, paragenetic investigations, fluid inclusion microthermometric, oxygen isotope analyses of the Spring Peak alteration zone have not been completed, and additional information is needed to know whether this is a viable option to lower the $\delta^{18}O$ of the Trsp.

5.5 Evolution models for the AVF rhyolites and bimodal volcanic rock suite

There are two common methods of producing a rhyolitic lava: 1) by mafic underplating and crustal anatexis and 2) by fractional crystallization. Both methods can also produce a bimodal (basalt-rhyolite) volcanic rock suite (Winter, 2001).

5.5.1 Mafic underplating and crustal anatexis

This method of rhyolite genesis is common in extensional environments, and will produce bimodal compositions. Mantle-derived mafic magmas commonly underplate or pond where there are large density contrasts, for example, at the mantle-crust boundary.

The magma may stall at this boundary, particularly if the crust is thickened as it is in the Mono basin region (~44 km; Gilbert, 2012), heating and partially melting the surrounding rock (Parfitt and Wilson, 2008; Winter, 2001). Partially melting mafic lower crust of the SNB produces felsic melts, which may be enriched in the LILE and light REE, and depleted in Nb and Ta (e.g., Beard and Lofgren, 1991; Rapp and Watson, 1995), such as is observed in the AVF volcanic rock suite.

Generating enough heat to melt the crust is an issue for a single pulse of mantle-derived magma, where an equal amount of heat is lost through the base and the roof of the intrusion. Multiple injections of hot, mantle-derived magma successively emplaced above the earlier ones, as expected due to greater buoyancy of hotter magmas, enables the maximum average temperature of the granitic melt remain higher for longer time periods than that of a single intrusion (Pederson et al., 1998; Petford and Gallagher, 2001). Likewise, the maximum melt thickness will be greater (Petford and Gallagher, 2001). Changing the number and thickness of intrusions can generate a range of melt compositions from a single, homogeneous source, accounting for the range of compositions for the AVF (Petford and Gallagher, 2001). Decreasing the time between each successive basalt intrusion also reduces the heat loss to the crust and increases the rate and amount of partial melting (Petford and Gallagher, 2001). Evidence for multiple-injected mafic sills and stocks underlie east-central Sierra Nevada in Onion Valley (Sisson et al., 1996).

Although it is possible that the AVF rhyolites were produced by partial melting of the mid- or upper crust, their flat HREE patterns indicate that it is unlikely they were derived by partial melting a garnet-bearing mafic lower crust. The AVF rhyolites do not plot in fields commonly used to identify melts of mafic crustal rocks on trace element diagrams (e.g., $\text{La/Yb}_{\text{pmn}}$ versus Yb_{pmn} and Sr/Y_{pmn} versus Y_{pmn} ; Drummond and Defant; 1990).

A subsequent open source process (e.g., crustal contamination) is required to produce the range in isotopic ratios observed in the mafic and intermediate AVF rocks.

5.5.2 Fractional crystallization

Differentiation models of (assimilation) fractional crystallization fit the AVF volcanic suite (section 5.3.4.3.2). Although there are a variety of mechanisms for producing a gap in silica content (i.e., the “Daly Gap”; Daly, 1927), removal of specific mineral phase(s) (i.e., Fe-Ti oxides and hornblende) through crystal fractionation is consistent with the major element oxides and trace element Harker diagrams for the AVF rocks (Figure 4.1, section 4.3).

The Daly Gap can be produced through fractional crystallization of Fe–Ti oxides.

Because there is no silica in this phase, its removal increases SiO_2 in the residual liquid.

Although, ilmenite is not present in the AVF lavas, titanomagnetite is present as a ubiquitous phase in the groundmass (Lange and Carmichael, 1996). Fractionation of hornblende has also been used to explain a silica gap in a calc-alkaline differentiation trend (e.g., Mt. St. Helens and Medicine Lake; Rutherford et al., 1985; Grove and

Donnelly-Nolan, 1986) and is present in the several AVF trachyandesite and trachydacite compositions. However, Lange and Carmichael (1996) refuted using titanomagnetite and/or hornblende to explain the Daly Gap in the AVF rocks. They used the oxygen fugacity of the trachybasalt (Tbch), the trachyandesite of Mt. Hicks (Qmh) and the trachyandesite of Aurora Crater (Qaca), in which the ferric iron compositions of these fractionating minerals (Fe-Ti oxides and hornblende) would cause too great of a drop of the ferric iron composition of the residual liquid with no feasible method to oxidize the magma. This may be the case, but it is also unlikely that the Tbch is genetically related to either Qmh or Qaca (see section 5.3.3.).

5.6 Summary of source and differentiation modeling of the AVF

The AVF, the TTVF, and the WGB all show arc-like trace element signatures, with enrichment of LILE (e.g., Ba, Pb, K, Sr) relative to HFSE (e.g., Ta, Nb, Zr, Hf, Ti) interpreted to be derived from the metasomatized lithospheric mantle, which is supported through the use of asthenosphere-lithosphere discrimination diagrams (e.g., Figure 5.4a, b), and partial melting modeling (section 5.2.2). The high abundance of LREE and steep La/Sm slope is difficult to reproduce through modeling except by low degree of partial melting of metasomatized lithospheric mantle xenoliths with high LREE contents and minimal, if any, residual garnet. The melting model derived using the LREE-enriched Big Pine xenoliths of Beard and Glazner (1995) best fit the AVF trachybasalt data. This partial melting model easily falls within the low degree of partial melting range calculated for the WGB fields, Tahoe-Reno region, and the ACA rocks, which produced <6% and 6-10% partial melts (Putirka and Busby, 2007, Cousens et al., 2008,

respectively), and agrees with the small estimated volume of volcanic products within the AVF ($\sim 25 \text{ km}^3$, Lange and Carmichael, 1996). The REE patterns and multi-element patterns of the intermediate AVF rocks parallel those of the mafic unit, indicating a similar source signature and subsequent differentiation processes.

The compositional variations are consistent with the inferred crystal fractionation of the observed crystal phases, olivine, clinopyroxene, hornblende, and titanomagnetite in the AVF rocks. Plagioclase is not a major phenocryst phase until the high silica rhyolite compositions (Figure 4.1, section 4.3). However, pure fractional crystallization will not account for a range in isotopic ratios, and petrographic evidence suggests differentiation in an open system. Differentiation by magmatic recharge is an unlikely model for the AVF; the large time span ($\sim 4 \text{ Ma}$) and small amount of magmatic products ($\sim 25 \text{ km}^3$ total for the AVF) suggest each magmatic pulse would cool and solidify before the next injection of magma. Each batch of magma may have differentiated to a differing degrees and incorporate variable amounts of crustal rock to produce the oscillating silica contents over time and also the range in isotopic signature for the AVF.

Crustal contamination is only evident geochemically in the high Th-U-Cs-Rb andesites in the AVF (such as units Qacc, Tach, Tpch). While Th, U, Ca, and Rb are all incompatible elements, Th is immobile and abundant in rocks that have undergone extreme crystal fractionation, like the SNB granitoid basement rocks. The role of crustal contamination is more ambiguous for the rest of the volcanic field, with no evidence of the incompatible element enrichment typically associated with crustal rocks (e.g., Th and Rb). This may be

accounted for through cryptic contamination (e.g., Beard et al., 2005). The interaction between a lithospheric mantle derived magma and a crustal source by end-member mixing or by assimilation fractional crystallization processes may produce reasonable methods for differentiation of the AVF rocks and can also account for the range in isotopic ratios. One major problem with these models is that they require multiple crustal contaminants to satisfy a variety of elements modeled. However, the most likely candidate for crustal contamination is the SNB, which is very heterogeneous (e.g., Barbarin et al., 1989); heterogeneous mixing of a crustal assimilant has been known to produce a range in isotopic signature within a single lava flow (Laughlin et al., 1972), as is shown within certain AVF units (e.g., Tach, QThch).

The isotopic ratios from the most mafic AVF unit (Tbch) may indicate extensive isotopic exchange with an enriched source (high initial $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$). This isotopic signature could be derived by partial melting of hornblende in the SNB at low-pressures conditions within the upper crust. Alternatively, the Tbch may have been derived from some of the older, enriched lithospheric mantle source. Instead of a discrete Proterozoic-Phanerozoic boundary normally shown on maps (e.g., Fig. 1.3), the accreted Phanerozoic lithosphere may be interleaved with the low- $^{143}\text{Nd}/^{144}\text{Nd}$ and high- $^{87}\text{Sr}/^{86}\text{Sr}$ Proterozoic enriched lithospheric mantle.

The low $\delta^{18}\text{O}$ (+4.01 ‰) of the high silica rhyolite, Trsp, could be derived through high-temperature (>300°C) alteration with a low $\delta^{18}\text{O}$ fluid such as meteoric water. Large alteration zones in the BHVF, with low $\delta^{18}\text{O}$ values (e.g., -12 ‰; O'Neil and Silberman,

1974), have been observed and could produce the low observed $\delta^{18}\text{O}$ value by ~27% contamination.

There are at least two potential methods of producing rhyolitic lavas and bimodal (basalt-rhyolite) volcanic suite such as found in the AVF: mafic underplating with crustal anatexis and fractional crystallization. In either case, an additional open system process is necessary to explain the range in isotopic data for the AVF.

5.7 Tectonic environment influencing generation and eruption of the AVF lavas

The WGB, the TTVF, and the AVF rocks erupted in the post-subduction (Atwater 1970, 1973; Atwater and Stock, 1998), mildly extensional, transtensional environment of the Walker Lane shear zone (e.g., Oldow, 2003), and their mafic magmas were derived from a metasomatized lithospheric mantle source (e.g., Fitton et al., 1988, 1991; Farmer et al., 1989; Kempton et al., 1991; Ormerod et al., 1991; Bradshaw et al., 1993; Cousens et al., 2008). The northward migration of the subducting Juan de Fuca plate combined with regional extension in the western Great Basin and Walker Lane lead to partially melting of the lithospheric mantle and eruption of the AVF.

After the removal of the subducting slab (by northward migration; see section 1.2.2), the asthenosphere was able to flow up from beneath the descending oceanic plates to fill the gap created in the slab-free zone (Figure 1.4; Thorkelson, 1994). This in turn caused the overlying mantle to heat. If the overlying mantle is given sufficient time, it will approach liquidus temperatures and begin to partially melt. Harry and Leeman (1995) show that the

metasomatized subcontinental lithospheric mantle will melt much earlier than the asthenospheric mantle. Similarly, Putirka and Busby (2007) suggested that dry magmas could not be erupted through thick Sierra Nevada crust (~44 km under the AVF; Gilbert, 2012) due to a lack of sufficient magma buoyancy. Volatiles (e.g., H₂O) increase buoyancy, increasing the chances that the magma will reach the surface. Putirka and Busby (2007) suggested a minimum of 0.4%–1.3% H₂O is required in order to reach the surface. Several periods of extension within the western Great Basin and Walker Lane created a series of north-striking faults, stepover grabens, and northeast-striking left-lateral faults, which strongly the AVF magmatism and volcanism by providing conduits through which the magmas could ascend to the surface (John et al., 2012; Cousens et al., 2008).

5.8 Recommendations for future work

There is great difficulty in constraining the source for the range in isotopic signature considering that both the mantle lithosphere at Big Pine, represented by xenoliths from the region (Beard and Glazner, 1995), and the SNB granitoids (e.g., Barbarin et al., 1989) show a large range in composition and isotopic signature. Syn-magmatic open system processes are evident through petrographic textures in the AVF lavas. To further investigate the role and/or extent of crustal contamination, single crystal microdrilling and isotope analyses are recommended to show how the isotope signature changes as the AVF magmas are differentiated (Ramos and Tepley, 2008). These analyses may be used to distinguish whether or not contamination is a major factor in chemical differentiation, provided that evidence of mixing record has not been obliterated (e.g., by hydration

crystallization; Beard et al., 2005). Plagioclase and other minerals have shown that individual crystals may preserve evidence of magma mixing, contamination, and other magmatic hybridization processes (e.g., Davidson and Tepley, 1997; Knesel et al., 1999; Gagnevin et al., 2004). Analyzes of both populations (i.e., sieved and non-sieved) of plagioclase and clinopyroxene would determine whether one population was incorporated from an alternate source (e.g., a contaminant) or experienced a process creating instability on one population (e.g., temperature/pressure disequilibrium). Detailed high-resolution quantitative imaging (e.g., backscattered electron images) of each mineral is advised prior to micro-drilling and isotope analysis.

Microprobe analyses can be used to investigate the pressure conditions (i.e., barometry) of the magmas at depth, where phenocrysts were formed before final ascent to the surface. Information on pressure and depth of crystallizing magmas will disclose whether partial melting and assimilation of country rock are viable processes; these processes will not occur where magmatic temperatures are below liquidus temperatures of minerals within the country rock, e.g., at shallow depths or while the magma erupts or (e.g., 10-15 km). If crystallization occurs in response to assimilation, the pressure can predict if partially melting of likely assimilated phases would increase or decrease the $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratio value of the host magma. There are several mineral techniques to measure barometry (e.g., Al in hornblende, Anderson and Smith, 1995; Zr in titanite, Hayden et al., 2008; plagioclase, Putirka, 2008; clinopyroxene, Putirka, 2008, Eq. A3; calcic amphiboles, Ridolfi and Renzulli, 2012). Knowing the crystallization depth may help

answer additional questions, e.g., whether the depth is constant for each AVF unit, or whether the ascent paths overlap with other magmas.

Although less significant to understanding the AVF as a whole, additional work on the hydrothermal alteration of the Spring Peak sinter and oxygen isotopes of the Trsp may highlight if the two are related, and the pervasiveness of low $\delta^{18}\text{O}$ values in the Trsp. Another sample of Trsp could be analyzed for whole rock oxygen isotopes to see if the low $\delta^{18}\text{O}$ value found in this study is localized or pervasive across the unit. An oxygen isotope analysis on refractory phenocrysts of the Trsp, such as quartz or zircon, when compared with the whole rock oxygen isotope analysis, will confirm if the low $\delta^{18}\text{O}$ is magmatic (e.g., derived from a low $\delta^{18}\text{O}$ protolith) or post magmatic (e.g., derived from hydrothermal alteration).

Appendices

Appendix A Sample locations and coordinates

Sample Number	Unit Name Shorthand	Latitude (N)⁴	Longitude (E)⁴	Elevation (m)	Easting	Northing	US State	Date Collected
13-AU-01	Trwai	38.08647	-119.14363	2289	312001	4217374	CA	July 11, 2013
13-AU-02	Trwai	38.08647	-119.14363	2289	312001	4217374	CA	July 11, 2013
13-AU-03	Tach	38.17448	-118.92614	2149	331280	4226721	CA	July 11, 2013
13-AU-04	Tach	38.19273	-118.91980	2200	331877	4228736	CA	July 11, 2013
13-AU-05	Tach	38.19804	-118.92188	2352	331627	4229526	CA	July 11, 2013
13-AU-06	QThch	38.20102	-118.91386	2281	332416	4229644	CA	July 11, 2013
13-AU-07	QThch	38.21463	-118.90026	2297	333638	4231130	CA	July 11, 2013
13-AU-08	Qta2	38.36186	-118.83071	2024	340051	4247345	NV	July 12, 2013
13-AU-09	Qaca	38.35775	-118.89468	1866	334452	4247002	NV	July 12, 2013
13-AU-10	Qaca	38.31948	-118.90932	1990	333084	4242781	NV	July 12, 2013
13-AU-11	QTaap	38.27256	-118.87419	2385	336051	4237513	NV	July 12, 2013
13-AU-12	QThch	38.14443	-118.96147	2088	328115	4223451	CA	July 13, 2013
13-AU-13	QThch	38.14278	-118.97952	2249	326528	4223302	CA	July 13, 2013
13-AU-14	Tach	38.13260	-119.01921	2186	323026	4222246	CA	July 13, 2013
13-AU-15	Tamb	38.13365	-119.01835	2212	323103	4222363	CA	July 13, 2013
13-AU-16	Trbh	38.13365	-119.01835	2224	323156	4222394	CA	July 13, 2013
13-AU-17	QThch	38.16043	-118.90283	2132	333290	4225120	CA	July 14, 2013
13-AU-18	Tach	38.15770	-118.92379	2089	331446	4224855	CA	July 14, 2013
13-AU-19	Tach	38.17276	-118.88896	2110	334533	4226463	CA	July 14, 2013
13-AU-20	QThch	38.17153	-118.88057	2119	335266	4226312	CA	July 14, 2013
13-AU-21	QThch	38.18961	-118.87365	2079	335913	4228306	CA	July 14, 2013
13-AU-22	Tpch	38.19090	-118.87123	2087	336128	4228445	CA	July 14, 2013

⁴ All Easting and Northing were collected in NAD 27, Zone 11 S.

Sample Number	Unit Name Shorthand	Latitude (N) ⁴	Longitude (E) ³	Elevation (m)	Easting	Northing	US State	Date Collected
13-AU-23	QThch	38.16506	-118.79920	2176	342379	4225452	NV	July 15, 2013
13-AU-24	Tach	38.18782	-118.79659	2129	342657	4227973	NV	July 15, 2013
13-AU-25	QThch	38.23149	-118.82353	2129	340394	4232865	NV	July 15, 2013
13-AU-26	Qmh	38.23682	-118.82573	2184	340212	4233461	NV	July 15, 2013
13-AU-27	Tdap	38.28243	-118.85816	2601	337476	4238579	NV	July 15, 2013
13-AU-28	Tdap	38.27756	-118.85991	2658	337312	4238041	NV	July 15, 2013
13-AU-29	Qmh	38.29236	-118.84699	2435	338477	4239660	NV	July 15, 2013
13-AU-30	Qaca	38.32452	-118.85579	2125	337776	4243245	NV	July 15, 2013
13-AU-31	Qaca	38.33475	-118.85823	2015	337586	4244385	NV	July 15, 2013
13-AU-32	QTopb	38.15556	-118.95087	2097	329071	4224667	CA	July 16, 2013
13-AU-33	QThch	38.16336	-118.98196	2404	326364	4225590	CA	July 16, 2013

Appendix B Petrography: Field and thin section (supplement to Chapter 3)

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Qaca	203393 203394 08-BA-3 13-AU-09 13-AU-10 13-AU-30 13-AU-31	Trachy-andesite	Usually strongly devitrified. Very platy lava flow but has a'a flow top. (13-AU-09 & -10). Columnar jointing at flow front (13-AU-30 & -31): wide fractures parallel to flow front, small fractures, perpendicular to flow front. Sample is almost aphanitic. Dark grey, glassy groundmass. Moderately devitrified (25-55% glass, plag, pyx, op). Minimal (1-5%) vesicles except for 13-AU-30 and 13-AU-31(25-35%).	pyx	2%	0.2-1.0, 4.0 Eu- to subhedral, forms clusters (esp. smaller xtals), associated with op. Core can be fragmented/ sieved/missing +/- embayments especially xtals bigger than 1 mm. Rare, reversely zoned xtal (green core, colorless rim).
				plag	tr	0.2-0.4, 1.0 Rare in (08-BA-3, and 13-AU-10).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Qacc	203230	Basaltic trachy-andesite	No vesicles, weakly devitrified (minor plag xtals), dusty.	plag	15%	0.5-1.5, 2.5 Eu- to subhedral, more than half have some amount of sieve texture, usually in the core or less commonly within rim. Rare concentric zoning.
				pyx	7%	1-1.7, 2.2 Subhedral, sieve texture (0.1mm plag Inclusions or ol inclusions), can have idd along rim, inclusions in zoning near rim.
				ol	3%	0.4-0.8, 2 Sub- to anhedral, skeletal (rounding and embayed). Nearly all are covered in idd altn.
	203232	Trachy-andesite	Symplectutue (intergrowth in a vermicular fashion) of op and pyx, particularly in clusters of pyx xtals (3mm). 5% vesicles. Cluster of plag and pyx 4mm.	plag	15%	1.5-2, 4.4 Eu- to subhedral. Concentric zoned. Sieve texture (dusty to spongy cellular) is common, rarely completely sieved. Rare pyx inclusions.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
	203402	Trachy-dacite	Strongly devitrified, (plag and hbld/op xtals). Clot 0.5 x 2mm of hbld/op and plag, with glass (not devitrified like the rest of the slide).	pyx	7%	0.3-1.0, 1.9 Eu- to subhedral, Isolated xtals, clusters +/- plag, may enveloping ol., Inclusions (0.01-0.05 mm), Zoning can be seen from inclusions, or birefringence.
				ol	tr	0.3-0.5 Rimmed by pyx or as inclusions (heavy idd altn) isolated xtals (minor idd altn).
				plag	3%	0.2-0.5 Euhedral with swallow tail texture. Rare sieve texture (a couple xtals heavily sieved and anhedral).
				pyx	tr	0.2-0.7, 1.5 Commonly in clusters (0.5-1.3 mm), also isolated. Rare large xtal (1.5 mm).
				hbld	3%	0.2-0.6, 1 Dominated by opacite. Only larges xtals have a hbld core.
				qtz	tr	1.5-2 A couple occurrences, anhedral xtals, each with reaction rims (pyx? xtals).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Qmh	203450 13-AU-26	Trachy-andesite	Flow top; lots of vesiculated blocks (10-50% vesicles, rounded, 2-3 cm diameter). Almost aphanitic. Dark grey groundmass. Strongly devitrified, (plag xtals 0.06-0.3 mm, op 0.02-0.05 mm). Vesicles (1-15%).	ol	2%	0.5-0.9, 1.4 Eu- to subhedral, +/-skeletal. Associated with op (ave 0.04 mm xtals in ol fractures), rare idd.
				plag	tr	0.3-0.8, 0.8 Subhedral, with or without inclusions (only in 203450).
				ol	1%	0.6-1.1, 2.0 Eu- to subhedral +/-skeletal. Associated with opaques (in fractures), rare idd. Microphenos (ave 0.2mm).
Qmsa	203395	Trachy-andesite	Weakly devitrified (plag laths, pyx, hbld, op), mostly isotropic with dusty texture. 15% Vesicles. Minor carbonate infilling.	hbld	4%	0.2-0.5, 2.0 Eu- to subhedral (rounding on prism ends where opacite is thicker), thin op rim commonly 0.05 mm regardless of xtal size (can be as thin as 0.01 mm), cores are missing (sometimes lined with opacite), zoned.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Qta 1				plag	1%	0.6-0.8, 1.9 Larger xtals have sieve texture in rim or throughout. Rare anhedral xtal (sived, and not sieved).
				pyx	2%	0.4-0.8, 1.5 Sub- to anhedral, most missing cores, occasional sieve texture, some form clusters of pyx +/- hbl, a few have sieve texture, some are surrounded by hbl xtals (replacement).
				plag	5%	0.4-1.0, 3.5 Euhedral or anhedral and embayed. 1% have sieve texture: sieved in cores (microphenos and phenos), others in rims, or not at all. One has convolution zonation (irregular).
				hbl	2%	0.4-0.8, 2.6 Subhedral and rounded, most are entirely opaque (only 2 largest xtals have some hbl cores preserved).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
			1.0 mm width.	pyx	tr.	0.3-0.7, 0.7 Most pyx is <0.4mm and in the groundmass.
				qtz	tr	1, 1.9 A couple occurrences, each surrounded by reaction rim corona.
				ol	tr	0.6 Rare pheno (usually as micropheno in groundmass), subhedral, minor idd staining, 0.01 mm inclusions.
				ol	1%	0.4-0.9, 0.9 Eu- to subhedral, fractured, op in fractures surrounded by strong idd. rim (0.01- 0.05 mm).
	203443	Basaltic trachy-andesite	Devitrified (plag, pyx, op, ol) and dusty. Vesicles (7-10%) are smaller than 203398. 1% vesicles 0.05-0.2mm width.	plag	tr	0.3 One zoned and euhedral, another is subhedral and strained.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Qta 2	08-BA-7 13-AU-08	Trachy-andesite	Strongly devitrified (plag, op). Nearly aphanitic, very vesicular (some are infilled) but chose sample with few vesicles. 0.5-2% vesicles, rounded and equant (ave 0.3-0.5 mm), rare dusty mineral (13-AU-08).	ol	1%	0.2-0.5, 0.5 Subhedral, skeletal (fast growth), most have pervasive idd (few with minor/no idd), can have missing core, or embayments.
				plag	tr	3.5 Subhedral, no sieve texture.
				hbld	4%	0.1-0.6, 1.8 Almost entirely replaced by opaques, only in largest crystal that shows original crystal. Elongate, tabular or acicular occasionally pinacoidal.
QTaap	13-AU-11	Trachy-andesite	Blocky weathering, but see thick plates from flowage. Med. grey groundmass. Weakly devitrified, dusty. 10% rounded vesicles (none infilled) ave. 0.1-0.3 mm diameter.	plag	2%	0.5-1.0, 1.8 Strong concentric zoning, a few larger crystals have sieve texture in a zone, with clear rim (trace).
				pyx	tr.	0.25 Sub- to anhedral.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
QThch	13-BA-4 13-AU-12 13-AU-13 13-AU-23 13-AU-25	Trachy-andesite	Massive, blocky flow. Sheared blocks toward road, south of outcrop (due to topography?). No vesicles. Dark grey groundmass. Moderately devitrified (few xtals of plag, hbld, +/- pyx; glass is dusty and not longer isotropic). Few vesicles (<2%).	plag	10-15%	Eu- to subhedral, some dusty inclusions along concentric zoning, ap (0.01-0.1 mm) in larger plag. Rare anhedral xtals with pervasive inclusions (dusty).
				hbld	3-7%	Eu- to subhedral, rounding and few embayments or missing cores, opacite rims up to 0.05mm thick. Smaller xtals are entirely opacite.
				bio	tr	Subhedral, rounding. Rare xtals have opacite rim (13-AU-13).
				pyx	tr	Eu- to anhedral, more edges when xtals are isolated, anhedral xtals are in hbld cores. Few clusters surrounded by hbld and opacite.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts			
				Mineral	Abundance (vol. %)	Size: ave., largest (mm)	Notes on Phenocrysts
	203176 12-BA-33 13-AU-06 13-AU-07 13-AU-17 13-AU-20 13-AU-21 13-AU-33	Trachy-dacite	Porphyritic. Flow banding is common. Rare plag megacrysts up to 1 cm in size. Flow top with irregular a'a and original oxidation texture (13-AU-33). Med. grey groundmass, (13-AU-33 has red streaks).	plag	10-15%	0.3-1.3, 4	Euhedral, swallow tails, Concentric zoning common, rare dusty plag (dusty texture follows zoning), can have ap incl. (13-AU-20, -21 & -33 where plag is larger).
			Weak to moderately devitrified. Microphenocrysts of plag, +/- pyx, +/- hbld (13-AU-20, -33). Dusty. Usually very few vesicles (<1%) except for 13-AU-07 (15%).	hbld	3-7%	0.2-0.6, 4	Eu- to subhedral. Usually isolated, but can form clusters +/- pyx up to 2 mm diameter. Can have pyx inclusions. Commonly has an opacite rim 0.01-0.08 mm (none in 13-AU-07).
				bio	tr	0.4, 2	Eu- to subhedral (few rounded edges), rare plag incl. Rare xtals have an opacite rim (13-AU-20 & -21). Largest xtal in 13-AU-07.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
				pyx	tr	0.4, 1.2 Forms clusters with either plag or hbl, or may be inclusions in hbl.
QTopb	13-AU-32	Basaltic trachy-andesite	Small outcrop. Med. grey groundmass. Strongly devitrified (plag, pyx, op xtals).	plag	15%	0.3-1, 4.5 Most form clusters, few are zoned, +/- ap inclusions, rare megacryst is heavily sieved.
				ol	7%	0.4-2, 2 Subhedral, strong idd altn on edges (0.15 mm), associated with op, some embayments/holes in core.
				pyx	3%	0.5 -1.5, 2 Strong association with ol, can form clusters.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm)
Tach	13-AU-03 13-AU-04 13-AU-18 13-AU-19 13-AU-24	Trachy-andesite	Some parts of the flow are very frothy (30-40% 1 mm vesicles), but most of the flow is 1-5% 1 cm elongate vesicles. Vesicles are rarely infilled. Flow banding common in some outcrops. Lt to dk grey groundmass.	plag	15-20%	0.5-0.9, 4.0
						Eu- to subhedral. Sieve texture in 1-3% xtals in a wide xtal size range (as small as 0.5 mm but usually in larger xtals). Rare sieve texture exclusive to rim or less commonly pervasive throughout xtal. Sieve texture can occur along a rim of plag cluster (13-AU-19). Rarely anhedral (when in clusters with pyx). Rare pyx inclusions.
			Devitrified. Crystal clusters (0.8 - 3.0 mm diameter) of subhedral to anhedral pyx and plag phenocrysts. Can also have subophitic texture in pyx (with plag).	pyx	3-5%	0.5-0.8, 1.4
						Sub- to anhedral. Some are concentrically zoned with 0.05 mm ol xtals in a zone. Some embayments on larger xtals. May be associated with anhedral (0.05-0.1 mm) opaque xtals (13-AU-19).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
	13-AU-05	Bas. trachyand.		ol	1-2%	0.1-0.4, 0.5 Eu- to subhedral, isolated xtals or in clusters with anhedral pyx xtals, weak to strong idd. altn (depending on the sample).
Tba	203326	Basaltic trachy-andesite	Strongly devitrified (no glass, plag, ol/idd, op). Plag is strongly orientated by flow direction (trachyte texture).	ol	3%	0.3-0.6, 1.0 Subhedral, skeletal. Some idd. rims (up to 0.08 mm thick) and can be as much as 50% of the ol crystal. Other ol have minor staining.
				pyx	2%	0.8-1.0, 1.1 Most pyx have intergrowths of plag (subophitic texture; not common in other AVF lavas). Some pyx growing around ol (replacement).
Tbch	12-BA-32	Trachy-basalt	Cinder cone buried by later lavas at quarry, near where the sample was taken. Dk. grey groundmass.	ol	4%	0.6, 1.8 Subhedral. Larger xtals are broken and/or rounded/ embayed with 0.02 mm idd. rim along fractures/ edges.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Tdap			<p>Strongly devitrified (very little glass. Plag, pyx, ol, and op xtals).</p> <p>Rare carb. infilling vesicles. A tr. anhedral mineral is entirely altered to a fine grained aggregate (xenocryst).</p>	pyx	2%	0.5, 3.2 Eu- to subhedral, some strong zonation, some sieve texture in the core, often associated with ol., can occur in clusters.
				plag	4%	0.6-1.3, 3.4 Eu- to subhedral. Continuous range of sizes. All megacrysts have a dusty, sieved cores (trace microphenos and phenocrysts). Minor inclusions of pyx up to 0.02 mm. Rarely in clusters.
				plag	8%	0.5-3.0, 4 Eu- to subhedral, strong zoning. Rare dusty/sieve texture. Inclusions of ap. (usually 0.04 mm, up to 0.5 mm long).
				hbld	8%	0.4-2.5 Eu- to subhedral, opacite rims (ave 0.05 mm thick).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
				bio	1%	0.6-1.2 Eu- to subhedral, tabular, or columnar.
Thd	01-BA-1	Trachydacite	Moderately devitrified (plag op), very dusty texture. Micro-phenocrysts of hbl.	plag	5%	0.2-1.4, 4.5 Eu- to anhedral (rounded). Most are zoned. Rare sieve texture (pervasive in a megacryst). Inclusions of pyx, ap. One large xtal 4.5 mm long.
				hbl	2%	0.2-0.7, 1.4 Well rounded with op. rim (0.01-1.5 mm). Some cores are gone. Can have pyx inclusions.
				pyx	1%	0.3-0.5, 0.6 Subhedral., Isolated or in clusters +/- plag. Inclusions in plag and hbl.
Tmlp	203184 203195	Basaltic andesite	Moderately to strongly devitrified. 203184 is mostly crystalline (plag, pyx, op, ol), 203195 is no longer isotropic.	pyx	3%	0.6, 1.4 Missing/fragmented core, sieved, occasionally forms clusters, strong association with op, can be zoned.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Tmr	08-BA-6	Rhyolite	Plag is <0.5 mm but larger xtals have ap inclusions, rare crystals have dusty texture, (203195).	ol	1%	0.5, 1.2 Subhedral, occasionally rounded or missing core, skeletal. Minor idd. Can be enveloped by pyx.
				qtz	tr	0.9 A single xtal with reaction rim corona enveloped by pyx(?) (203184).
				qtz	1%	0.4-1, 1.6 Subhedral, can have embayments. One xtal is fringed by fibrous, spherulite.
				alk-fspr	1%	0.5-1.8 Subhedral, large fractures cut across xtal, small fractures perpendicular to rim.
				plag	1%	0.8-1.0, 1.6 Mostly euhedral, but some xtals are completely rounded, one cluster is fringed by fibrous, spherulite.
				bio	1%	0.2-0.8 Eu- to subhedral, elongate, tabular.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
				ti	tr	0.05-0.1, 0.6 Eu- to subhedral. Can be associated with op or isolated.
Tpch	203181 203246 203247 203354 13-AU-22	Trachy-andesite	Symplectic texture: vermicular intergrowth of op and pyx at ol-pyx contacts emanating from ol into pyx. Also found to a lesser extent in pyx isolated from ol (203181, 203246). Moderately devitrified (plag, op, ol/idd? pyx). Range in abundance of vesicles (5-15%). Med- dk grey groundmass.	plag	8-15%	Euhedral (rare anhedral with embayments and sieve texture at rim). Concentric zoned. Can have pyx inclusions. Sieve texture is common; sometime pervasive with clean rim, rarely pervasive throughout. Rare ap inclusions. Rare megacryst (4mm).
				pyx	4-7%	Subhedral, often forms clusters. Associated with anhedral op (along rim or as inclusions). Core can be fragmented and/or have plag inclusions. Zoning can be seen from a change in color (green core to colorless rim in ppl, reverse zoning), inclusions, or birefringence.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Trsp	203382 11-BA-46	Rhyolite	Visited another outcrop (southwest of Aurora Crater, i.e. not where these samples were collected). Aphyric and sheared. Lt. grey groundmass. Small (0.015 mm) feldspar laths with swallow tails +/- skeletal texture. Fewer xtals and more glass shards in 302282 than 11-BA-46. Larger feldspars have fracturing in their rims.	ol	1-2%	0.4-1.2 Sub- to anhedral, skeletal. Some ol have heavy idd altn (ass with py as inclusions/rim), others have almost no idd. (isolated xtals), can have concentric zones.
				hb	tr	0.5-1 Subhedral, enveloped by 0.05 mm rim of opacite.
				qtz	1%	0.6-1.4, 2.1 Eu- to anhedral with embayments, can fringed by fibrous, spherulite (thinner spherulitic rim than Tmr).
				plag	1%	0.6-1.7, 2.8 Euhedral when isolated, anhedral in clusters. Concentric zoning.
				alk-fspr	tr	1-1.7 Subhedral to anhedral (rounded, and embayed). Can be strained.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Trwai BHVF				bio	1%	0.1-1.1 (not in 203382).
				hbld	tr	0.6 Subhedral, one xtal in 11-BA-46 with bio inclusion in the center, highly fractured, no op rim.
				ol	7%	0.25-2.0 Subhedral, pinacoidal, fractured/fragmented, opaque inclusions (0.01-0.05 mm).
	13-AU-01	Basalt	Dk. grey groundmass. Devitrified (plag, pyx, op up to 0.1 mm).	pyx	1%	0.3-0.6 Subhedral. Larger xtals have missing and/or fragmented cores.
				plag	tr.	1.0-2.0 Rare. Rounded/irregular edges.
				ol	3-10%	0.5-1 Isolated, no idd rim.
				pyx	5%	0.5-2.1 Subhedral, fractured, and plucked.
	13-AU-02	Basalt	Devitrified (plag, op up to 0.1 mm). 5% vesicles.	plag	tr.	1.3 Dusty, not sieved.
				qtz	tr	1.8 Mostly plucked out, mantled by pyx.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Base-ment	203350	Granite (Murphy Creek)		alk-fspr	35%	0.7-2, 4 Anhedral. Altered (dusty). Exsolution lamellae. Tartan twinning.
				plag	35%	0.5-2 Anhedral. Altered (dusty).
				qtz	30%	1.7-5.5 Anhedral (elongate, space filling). Undulose extinction.
				bio/ chl	tr	0.3-1.5 Anhedral. Greenish brown. (Chl is a secondary/altn phase)
Base-ment	203377	Grano-diorite (Masonic Mt.)	Accessory mineral ap (0.05-0.2 mm) associated with bio.	alk-fspr	25-30%	1.0-2.1, 5 Altered (dusty). Exsolution lamellae.
				plag	40-45%	0.8-2.5, 4 Subhedral to anhedral. Concentric zoned. Myrmikite texture at xtal margins.
				qtz	25-30%	0.5-2, 5 Anhedral. Undulose extinction.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Base- ment	203387	Monzo- granite (Rattle- snake Gulch)	Trace amount of subhedral opaques (0.05-0.3 mm). Accessory mineral ap (up to 0.05 mm).	bio/ chl	3%	0.5-1.2 Anhedral. Tabular with jagged edges to fibrous (especially with secondary/altn phase, chl). Bio can be glassy-looking and subhedral.
				ti	1%	0.1-0.3, 1.0 Euhedral. Associated with bio and op.
				alk-fspr	25-30%	1.5-3.0 Altered (dusty). Exsolution lamellae.
				plag	40-45%	1.3-2, 3 Altered (dusty). Saussuritization alteration (mostly dusty, epidote xtals up to 0.1 mm).
				qtz	20-25%	1.5-2, 3 Anhedral.
				bio	5%	0.5-2 Anhedral. Bio can be glassy-looking and subhedral.

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
				ti	1%	0.1-0.5 Subhedral. Associated with bio and op.
Base-ment	BPC-13-1	Siliceous shale	Fine-grained blotchy red-brown groundmass. Late-stage fracture infilling (qtz? +chlorite).	N/A	N/A	N/A
Base-ment	CC-13-1	Black argillite	Fine-grained blotchy black groundmass. Late-stage fracture infilling (anhedral, ave. 0.02-0.05 mm sized crystals: qtz?)	N/A	N/A	N/A
Base-ment	HOM-13-1	Meta-andesite	Groundmass is made up of qtz (only occurs as small, anhedral clusters ave 0.1 mm, largest 1.1 mm), and chl+/- bio (ave 0.05-0.1 mm, largest 0.3 mm).	plag	55%	1.2-2, 4.5 Altered (light brown to patchy medium brown, and dusty; not the cellular/sieve texture of the AVF samples).

Unit	Sample #	Rock Type	Field & thin section notes	Phenocrysts		
				Mineral	Abundance (vol. %)	Size: ave., largest (mm) Notes on Phenocrysts
Base-ment	MAS-13-2	Meta-dacite? (Masonic Mt.)	Strongly foliated (bio). Groundmass is made of chlorite (ave 0.05 mm), qtz + feldspar? (ave 0.01-0.05, largest 0.2 mm), and op (largest 0.2 mm, overprints foliation).	alk-fspr?	5%	0.5-1, 2.5 Subhedral to anhedral (rounded). Simple twins. Foliation bends around xtals.
				bio	2%	0.1-0.3 Euohedral to subhedral, platy xtals create dominant foliation.

Appendix C Assessment of uncertainty

C.1 SGS Minerals

ICP-AES Units [wt%]	Sample 13-AU-13	Duplicate sample	Mean	Std. dev. (1 σ)	RSD [%]	Percent diff. [%]
Al	9.37	9.68	9.53	0.22	2.30	3.25
Ca	3.80	3.80	3.80	0	0	0
Fe	4.16	3.67	3.92	0.35	8.85	12.52
K	2.90	2.90	2.90	0	0	0
Mg	1.68	1.37	1.53	0.22	14.37	20.33
Mn	0.08	0.06	0.07	0.01	20.20	28.57
P	0.15	0.14	0.15	0.01	4.88	6.90
Ti	0.49	0.49	0.49	0	0	0
ICP-AES Units [ppm]						
Ag	<1	<1				
As	<30	<30				
Ba	1320	1420	1370	70.71	5.16	7.30
Be	<5	<5				
Bi	<0.1	<0.1				
Cd	<0.2	<0.2				
Ce	53.4	53.3	53.35	0.07	0.13	0.19
Co	16.8	16.8	16.8	0	0	0
Cr	60	10	35	35.36	101.02	142.86
Cs	1.1	1.1	1.1	0	0	0
Cu	27	24	25.5	2.12	8.32	11.76
Dy	2.86	2.78	2.82	0.06	2.01	2.84
Er	1.54	1.57	1.56	0.02	1.36	1.93
Eu	1.16	1.08	1.12	0.06	5.05	7.14
Ga	21	21	21	0	0	0
Gd	3.77	3.69	3.73	0.06	1.52	2.14
Ge	1	1	1	0	0	0
Hf	4	5	4.5	0.71	15.71	22.22
Ho	0.52	0.54	0.53	0.01	2.67	3.77
In	<0.2	<0.2				
La	26.4	25.9	26.2	0.35	1.35	1.91
Li	20	10	15.00	7.07	47.14	66.67
Lu	0.20	0.18	0.19	0.01	7.44	10.53
Mo	2	2	2	0	0	0
Nb	9	9	9	0	0	0
Nd	23.8	23.7	23.75	0.07	0.30	0.42
Ni	21	12	16.5	6.36	38.57	54.55
Pb	21	27	24	4.24	17.68	25.00
Pr	6.08	5.99	6.04	0.06	1.05	1.49
Rb	65.7	65.2	65.45	0.35	0.54	0.76
Sb	0.5	0.5	0.5	0	0	0

ICP-AES Units [ppm]	Sample 13-AU-13	Duplicate sample	Mean	Std. dev. (1σ)	RSD [%]	Percent diff. [%]
Sc	12	10	11	1.41	12.86	18.18
Sm	4.3	4.2	4.25	0.07	1.66	2.35
Sn	1	<1				
Sr	709.0	880.0	794.5	120.92	15.22	21.52
Ta	<0.5	<0.5				
Tb	0.61	0.49	0.55	0.08	15.43	21.82
Th	4.4	4.9	4.65	0.35	7.60	10.75
Tl	<0.5	<0.5				
Tm	0.22	0.22	0.22	0	0	0
U	1.56	1.59	1.58	0.02	1.35	1.90
V	113	115	114	1.41	1.24	1.75
W	<1	<1				
Y	15.5	15.4	15.45	0.07	0.46	0.65
Yb	1.5	1.5	1.5	0	0	0
Zn	84	72	78	8.49	10.88	15.38
Zr	170	174	172.00	2.83	1.64	2.33

Note: Duplicate sample run with original sample set.

ICP-AES Units [wt %]	Sample H13-41	Duplicate sample	Mean	Std. dev. (1σ)	RSD [%]	Percent diff. [%]
Al	6.44	6.27	6.36	0.12	1.89	2.68
Ca	0.2	0.2	0.20	0.00	0.00	0.00
Fe	1.06	1.04	1.05	0.01	1.35	1.90
K	3.9	3.9	3.90	0.00	0.00	0.00
Mg	0.06	0.06	0.06	0.00	0.00	0.00
Mn	0.02	0.02	0.02	0.00	0.00	0.00
P	<0.01	<0.01				
Ti	0.07	0.06	0.07	0.01	10.88	15.38
ICP-AES Units [ppm]						
Ag	<1	<1				
As	<30	<30				
Ba	67.1	60.1	63.60	4.95	7.78	11.01
Be	<5	<5				
Bi	<0.1	<0.1				
Cd	<0.2	<0.2				
Ce	85.1	84.5	84.80	0.42	0.50	0.71
Co	<0.5	<0.5				
Cr	<10	<10				
Cs	1.7	1.7	1.70	0.00	0.00	0.00
Cu	<5	<5				
Dy	5.57	5.5	5.54	0.05	0.89	1.26
Er	3.43	3.3	3.37	0.09	2.73	3.86
Eu	0.11	0.11	0.11	0.00	0.00	0.00
Ga	23	23	23.00	0.00	0.00	0.00
Gd	5.64	5.7	5.67	0.04	0.75	1.06
Ge	2	2	2.00	0.00	0.00	0.00
Hf	7	7	7.00	0.00	0.00	0.00
Ho	1.11	1.07	1.09	0.03	2.59	3.67
In	<0.2	<0.2				
La	32.6	32.2	32.40	0.28	0.87	1.23
Li	40	40	40.00	0.00	0.00	0.00
Lu	0.54	0.45	0.50	0.06	12.86	18.18
Mo	4	3	3.50	0.71	20.20	28.57
Nb	28	28	28.00	0.00	0.00	0.00
Nd	29.5	29.5	29.50	0.00	0.00	0.00
Ni	8	<5				
Pb	35	35	35.00	0.00	0.00	0.00
Pr	8.24	8.09	8.17	0.11	1.30	1.84
Rb	197	192	194.50	3.54	1.82	2.57
Sb	0.9	1.1	1.00	0.14	14.14	20.00
Sc	<5	<5				
Sm	6	5.9	5.95	0.07	1.19	1.68
Sn	2	2	2.00	0.00	0.00	0.00
Sr	10.7	13	11.85	1.63	13.72	19.41

ICP-AES Units [ppm]	Sample H13-41	Duplicate sample	Mean	Std. dev. (1σ)	RSD [%]	Percent diff. [%]
Ta	1.6	1.6	1.60	0.00	0.00	0.00
Tb	0.85	0.85	0.85	0.00	0.00	0.00
Th	24	23.9	23.95	0.07	0.30	0.42
Tl	0.9	0.9	0.90	0.00	0.00	0.00
Tm	0.51	0.5	0.51	0.01	1.40	1.98
U	6.5	6.6	6.55	0.07	1.08	1.53
V	<5	<5				
W	2	2	2.00	0.00	0.00	0.00
Y	33.6	33.2	33.40	0.28	0.85	1.20
Yb	3.3	3.4	3.35	0.07	2.11	2.99
Zn	83	69	76.00	9.90	13.03	18.42
Zr	200	213	206.50	9.19	4.45	6.30

Note: Duplicate sample run with original sample set.

WDXRF	Measured	DGPM-1 Duplicate	SGS defined	Mean	Std. dev. (1σ)	RSD [%]	Percent diff. [%]
Units [wt%]	value	value	values				
Al ₂ O ₃	9.59	9.57	9.4	9.58	0.01	0.1	0.2
CaO	0.2	0.19	0.19	0.20	0.01	3.6	5.1
Cr ₂ O ₃	0.02	0.02	0.014	0.02	0	0	0
Fe ₂ O ₃	1.98	1.97	1.9	1.98	0.01	0.4	0.5
K ₂ O	2.79	2.76	2.72	2.78	0.02	0.8	1.1
LOI	4.1	4.27	3.68	4.19	0.12	2.9	4.1
MgO	0.57	0.57	0.58	0.57	0	0	0
MnO	<0.01	<0.01	<0.01				
Na ₂ O	0.05	0.06	0.07	0.06	0.01	12.9	18.2
P ₂ O ₅	0.1	0.09	0.087	0.10	0.01	7.4	10.5
SiO ₂	80.7	80.2	80.2	80.45	0.35	0.4	0.6
TiO ₂	0.56	0.56	0.55	0.56	0	0	0

LOI = gravimetric loss of ignition

ICP-AES

Units [wt%]

Al	5.12	5.14	4.82	5.13	0.01	0.3	0.4
Ca	0.2	0.2	0.17	0.20	0	0	0
Fe	1.29	1.32	1.36	1.31	0.02	1.6	2.3
K	2.2	2.2	2.23	2.20	0	0	0
Mg	0.28	0.29	0.322	0.29	0.01	2.5	3.5
Mn	<0.01	<0.01	0.0025				
P	0.04	0.03	0.041	0.04	0.01	20.2	28.6
Ti	0.32	0.32	0.325	0.32	0	0	0

ICP-AES

Units [ppm]

Ag	<1	<1	<1				
As	180	180	169	180.00	0	0	0
Ba	1200	1210	1272	1205.00	7.07	0.6	0.8
Be	<5	<5	<5				
Bi	<0.1	<0.1	<0.1				
Cd	0.3	0.3	0.34	0.30	0	0	0
Ce	89.8	90.6	90.5	90.20	0.57	0.6	0.9
Co	1.8	1.7	1.36	1.75	0.07	4.0	5.7
Cr	130	130	120	130.00	0	0	0
Cs	8.5	8.8	8.9	8.65	0.21	2.5	3.5
Cu	13	14	13.7	13.50	0.71	5.2	7.4
Dy	3.15	3.31	3.22	3.23	0.11	3.5	5.0
Er	2.04	2.07	2.08	2.06	0.02	1.0	1.5
Eu	0.77	0.8	0.78	0.79	0.02	2.7	3.8
Ga	12	12	11.3	12.00	0	0	0
Gd	3.64	3.6	3.9	3.62	0.03	0.8	1.1
Ge	3	3	2.34	3.00	0	0	0
Hf	12	12	9.8	12.00	0	0	0

ICPAES Units [ppm]	DGPM-1			Mean	Std. dev. (1 σ)	RSD [%]	Percent diff. [%]
	Measured value	Duplicate value	SGS defined values				
Ho	0.64	0.65	0.67	0.65	0.01	1.1	1.6
In	<0.2	<0.2	<0.2				
La	48.3	48.2	51.7	48.25	0.07	0.1	0.2
Li	40	40	41.9	40.00	0	0	0
Lu	0.31	0.33	0.315	0.32	0.01	4.4	6.3
Mo	15	15	13.6	15.00	0	0	0
Nb	11	11	9.8	11.00	0	0	0
Nd	29.6	29.9	30.5	29.75	0.21	0.7	1.0
Ni	16	19	14.3	17.50	2.12	12.1	17.1
Pb	12	13	9.8	12.50	0.71	5.7	8.0
Pr	9.03	9.22	9.4	9.13	0.13	1.5	2.1
Rb	100	99.6	94	99.80	0.28	0.3	0.4
Sb	12.8	13.3	13	13.05	0.35	2.7	3.8
Sc	9	9	8.8	9.00	0	0	0
Sm	4.6	4.6	4.7	4.60	0	0	0
Sn	2	2	1.8	2.00	0	0	0
Sr	85.1	82.9	91	84.00	1.56	1.9	2.6
Ta	0.8	0.7	0.67	0.75	0.07	9.4	13.3
Tb	0.53	0.54	0.59	0.54	0.01	1.3	1.9
Th	11.2	11.4	9.9	11.30	0.14	1.3	1.8
Tl	8.4	8.6	8.2	8.50	0.14	1.7	2.4
Tm	0.32	0.3	0.29	0.31	0.01	4.6	6.5
U	3.67	3.71	3.3	3.69	0.03	0.8	1.1
V	101	101	104	101.00	0	0	0
W	77	76	73.6	76.50	0.71	0.9	1.3
Y	20.6	20.6	20	20.60	0	0	0
Yb	2.1	2.1	2.1	2.10	0	0	0
Zn	25	23	22	24.00	1.41	5.9	8.3
Zr	447	447	410	447.00	0	0	0

	GSP-QC			Mean	Std. dev.	RSD	Percent
WDXRF	Measured	Duplicate	SGS defined		(1σ)	[%]	diff. [%]
Units [wt%]	value	value	values				
Al ₂ O ₃	14.6	14.7	14.4	14.65	0.07	0.5	0.7
CaO	2.09	2.1	2.24	2.10	0.01	0.3	0.5
Cr ₂ O ₃	<0.01	0.01	<0.01				
Fe ₂ O ₃	4.16	4.18	4.03	4.17	0.01	0.3	0.5
K ₂ O	5.19	5.21	5.12	5.20	0.01	0.3	0.4
LOI	1.56	1.67	1.05	1.62	0.08	4.8	6.8
MgO	1.05	1.05	1.06	1.05	0	0	0
MnO	0.03	0.03	0.032	0.03	0	0	0
Na ₂ O	2.52	2.56	2.55	2.54	0.03	1.1	1.6
P ₂ O ₅	0.27	0.28	0.27	0.28	0.01	2.6	3.6
SiO ₂	67.8	67.8	67.4	67.80	0	0	0
TiO ₂	0.67	0.66	0.66	0.67	0.01	1.1	1.5

LOI = gravimetric loss of ignition

ICP-AES

Units [wt%]

Al	7.85	7.89	7.4	7.87	0.03	0.4	0.5
Ca	1.5	1.4	1.55	1.45	0.07	4.9	6.9
Fe	2.8	2.77	2.85	2.79	0.02	0.8	1.1
K	4.3	4.2	4.12	4.25	0.07	1.7	2.4
Mg	0.58	0.56	0.06	0.57	0.01	2.5	3.5
Mn	0.03	0.03	0.027	0.03	0	0	0
P	0.1	0.11	0.118	0.11	0.01	6.7	9.5
Ti	0.39	0.38	0.376	0.39	0.01	1.8	2.6

ICP-AES

Units [ppm]

Ag	3	4	2.8	3.50	0.71	20.2	28.6
As	40	<30	<30				
Ba	1270	1240	1320	1255.00	21.21	1.7	2.4
Be	<5	<5	<5				
Bi	4.4	4.5	4.3	4.45	0.07	1.6	2.2
Cd	0.2	0.3	0.26	0.25	0.07	28.3	40.0
Ce	380	379	369	379.50	0.71	0.2	0.3
Co	8.3	8.1	6.4	8.20	0.14	1.7	2.4
Cr	20	20	22.1	20.00	0	0	0
Cs	1.9	1.9	1.99	1.90	0	0	0
Cu	31	32	31.1	31.50	0.71	2.2	3.2
Dy	5.48	5.6	5.56	5.54	0.08	1.5	2.2
Er	2.34	2.17	2.4	2.26	0.12	5.3	7.5
Eu	2.05	2.05	2.02	2.05	0	0	0
Ga	24	24	22	24.00	0	0	0
Gd	11.7	11.4	14.8	11.55	0.21	1.8	2.6
Ge	1	2	1.6	1.50	0.71	47.1	66.7
Hf	15	16	9.7	15.50	0.71	4.6	6.5

ICPAES Units [ppm]	GSP-QC			Mean	Std. dev. (1 σ)	RSD [%]	Percent diff. [%]
	Measured value	Duplicate value	SGS defined values				
Ho	0.86	0.88	0.92	0.87	0.01	1.6	2.3
In	<0.2	<0.2	<0.2				
La	156	154	166	155.00	1.41	0.9	1.3
Li	30	30	33.6	30.00	0.00	0.0	0.0
Lu	0.22	0.25	0.248	0.24	0.02	9.0	12.8
Mo	2	<2	<2				
Nb	27	27	22.6	27.00	0	0	0
Nd	180	180	186	180.00	0	0	0
Ni	13	14	10.6	13.50	0.71	5.2	7.4
Pb	41	39	43.3	40.00	1.41	3.5	5.0
Pr	47.7	47.5	47.7	47.60	0.14	0.3	0.4
Rb	253	254	240	253.50	0.71	0.3	0.4
Sb	1	0.8	1	0.90	0.14	15.7	22.2
Sc	6	6	5.75	6.00	0	0	0
Sm	22.6	22.6	23.8	22.60	0	0	0
Sn	5	5	1.9	5.00	0	0	0
Sr	231	226	219	228.50	3.54	1.5	2.2
Ta	7.5	7.7	7.9	7.60	0.14	1.9	2.6
Tb	1.18	1.17	8	1.18	0.01	0.6	0.9
Th	106	107	100	106.50	0.71	0.7	0.9
Tl	2.1	2.3	2.1	2.20	0.14	6.4	9.1
Tm	0.29	0.3	0.27	0.30	0.01	2.4	3.4
U	2.66	2.79	2.6	2.73	0.09	3.4	4.8
V	77	76	73	76.50	0.71	0.9	1.3
W	7	7	6.3	7.00	0	0	0
Y	25.7	25.9	25	25.80	0.14	0.5	0.8
Yb	1.6	1.8	1.7	1.70	0.14	8.3	11.8
Zn	117	113	115	115.00	2.83	2.5	3.5
Zr	568	585	396	576.50	12.02	2.1	2.9

WDXRF	SY-4			BLANK-13-11443	
	Measured value	Certificate value	2-σ (95% confidence)	Measured value	Detection limit
Units [wt%]					
LOI	4.65	4.56	0.07	100	0.01
SiO ₂	50.00	49.9	0.1	<0.01	0.01
Al ₂ O ₃	20.80	20.69	0.08	<0.01	0.01
Fe ₂ O ₃	6.29	6.21	0.03	<0.01	0.01
MgO	0.57	0.54	0.01	<0.01	0.01
CaO	8.09	8.05	0.04	<0.01	0.01
K ₂ O	1.68	1.66	0.02	<0.01	0.01
Na ₂ O	7.12	7.1	0.05	<0.01	0.01
TiO ₂	0.29	0.287	0.003	<0.01	0.01
MnO	0.11	0.108	0.001	<0.01	0.01
P ₂ O ₅	0.13	0.131	0.004	<0.01	0.01
Cr ₂ O ₃	<0.01			<0.01	0.01
ICPAES					
Units [wt%]					
Al	11.00			<0.01	0.01
Ca	5.30			<0.1	0.1
Fe	4.19			<0.01	0.01
K	1.40			<0.1	0.1
Mg	0.29			<0.01	0.01
Mn	0.08			<0.01	0.01
P	0.05			<0.01	0.01
Ti	0.17			<0.01	0.01
Units [ppm]					
Ag	<1			<1	1
As	<30			<30	30
Ba	307	340	5	<0.5	0.5
Be	<5	2.6	0.1	<5	5
Bi	<0.1			<0.1	0.1
Cd	<0.2			<0.2	0.2
Ce	121	122	2	<0.1	0.1
Co	2.9	2.8	0.2	<0.5	0.5
Cr	<10	12	1	<10	10
Cs	1.6	1.5	0.1	<0.1	0.1
Cu	<5	7	1	<5	5
Dy	19.4	18.2	0.6	<0.05	0.05
Er	14.6	14.2	0.5	<0.05	0.05
Eu	1.87	2.00	0.04	<0.05	0.05

ICPAES	SY-4			BLANK-13-11443	
	Measured value	Certificate value	2- σ (95 % confidence)	Measured value	Detection limit
Units [wt %]					
Ga	38	35	1	<1	1
Gd	14.9	14.0	0.5	<0.05	0.05
Ge	1			<1	1
Hf	11	10.6	0.4	<1	1
Ho	4.13	4.3	0.1	<0.05	0.05
In	<0.2			<0.2	0.2
La	54.2	58	1	<0.1	0.1
Li	40	37	2	<10	10
Lu	1.98	2.1	0.1	<0.05	0.05
Mo	<2			<2	2
Nb	15	13	1	<1	1
Nd	56.9	57	1	<0.1	0.1
Ni	7	9	1	10	5
Pb	12	10	1	<5	5
Pr	14.30	15.0	0.3	<0.05	0.05
Rb	58.9	55	1.5	<0.2	0.2
Sb	<0.1			0.2	0.1
Sc	<5	1.1	0.1	<5	5
Sm	12.2	12.7	0.4	<0.1	0.1
Sn	7			<1	1
Sr	1120.0	1191	12	<0.1	0.1
Ta	0.7	0.9	0.1	<0.5	0.5
Tb	2.55	2.6	0.1	<0.05	0.05
Th	1.5	1.4	0.2	<0.1	0.1
Tl	<0.5			<0.5	0.5
Tm	2.35	2.3	0.1	<0.05	0.05
U	0.84	0.8	0.1	<0.05	0.05
V	13	8	0.1	<5	5
W	<1			<1	1
Y	121.0	119	2	<0.5	0.5
Yb	15.2	14.8	0.4	<0.1	0.1
Zn	95	93	2	<5	5
Zr	527	517	16	<0.5	0.5

Appendix C Assessment of uncertainty

C.2 Geoscience Laboratories

XRF Units [wt %]	Measured value	BHVO-2	2-σ (95% confidence)
		Certificate value	
SiO ₂	50.37	49.9	1.2
TiO ₂	2.73	2.73	0.08
Al ₂ O ₃	13.74	13.5	0.4
Fe ₂ O ₃	12.33	12.3	0.4
MnO	0.170		
MgO	7.38	7.23	0.24
CaO	11.443	11.4	0.4
Na ₂ O	2.22	2.22	0.16
K ₂ O	0.52	0.52	0.02
P ₂ O ₅	0.284	0.27	0.04
LOI	-0.64		
ICP-MS			
Units [ppm]			
Ba	132.4	130	26
Be	1.03		
Bi	0.039		
Cd	0.12		
Ce	36.73	38	4
Co	47.0	45	6
Cr	304	280	38
Cs	0.100		
Cu	136	127	14
Dy	5.28		
Er	2.54		
Eu	2.06		
Ga	21.33	21.7	1.8
Gd	6.14	6.3	0.4
Hf	4.42	4.1	0.6
Ho	0.95	1.04	0.08
La	15.19	15	2
Li	4.0	5	
Lu	0.27	0.28	0.02
Mo	3.84		
Nb	17.87	18	4
Nd	24.82	25.0	3.6
Ni	126	119	14
Pb	1.5		
Pr	5.33		

= Outside 95% confidence of certificate value.

ICP-MS Units [ppm]	BHVO-2		
	Measured value	Certificate value	2- σ (95% confidence)
Rb	9.49	9.8	2.0
Sb	0.11		
Sc	33.6	32	2
Sm	6.08	6.2	0.8
Sn	1.89	1.9	
Sr	393	389	46
Ta	1.2	1.4	
Tb	0.894	0.9	
Th	1.21	1.2	0.6
Ti	17335.42	16300	400
Tl	0.02		
Tm	0.328		
U	0.42		
V	333.62	317	22
W	0.22		
Y	25.73	26	4
Yb	1.930	2	0.4
Zn	97.95	103	12
Zr	175	172	22

XRF Units [wt%]	SY-4		
	Measured value	Certificate value	2-sigma (95% confidence)
SiO ₂	49.94	49.9	0.1
TiO ₂	0.29	0.287	0.003
Al ₂ O ₃	20.84	20.69	0.08
Fe ₂ O ₃	6.20	6.21	0.03
MnO	0.106	0.108	0.001
MgO	0.52	0.54	0.01
CaO	7.987	8.05	0.04
Na ₂ O	7.13	7.1	0.05
K ₂ O	1.63	1.66	0.02
P ₂ O ₅	0.134	0.131	0.004
LOI	4.87	4.56	0.07

= Outside 95% confidence of certificate value.

BLANK-13-11443

ICP-MS Units [ppm]	Measured value	Detect limit
Ba	0.3	0.9
Be	0	0.06
Bi	0.002	0.009
Cd	0.01	0.01
Ce	0.03	0.2
Co	0	0.1
Cr	1	24
Cs	0.000	0.006
Cu	0	2
Dy	0	0.02
Er	0	0.01
Eu	0	0.005
Ga	0	0.05
Gd	0	0.02
Hf	0.02	0.09
Ho	0	0.003
La	0.01	0.09
Li	0	0.2
Lu	0	0.002
Mo	0.05	0.03
Nb	0	0.04
Nd	0.01	0.08
Ni	0	3
Pb	0	0.4
Pr	0	0.02
Rb	0.04	0.2
Sb	0.01	0.04
Sc	0.3	0.9
Sm	0	0.02
Sn	0.01	0.08
Sr	0	2
Ta	0	0.2
Tb	0	0.003
Th	0	0.09
Ti	0.87	26
Tl	0	0.005
Tm	0	0.002
U	0	0.02
V	0.08	10
W	0.01	0.5
Y	0.01	0.08
Yb	0.002	0.009
Zn	1.02	8
Zr	1	3
= Outside 95% confidence of certificate value.		

	Sample BPC 13-1	Duplicate sample	Mean	Std. dev. (1 σ)	RSD [%]	Percent diff. [%]
Units [ppm]						
Ba	>2400	3392.5				
Be	1.39	1.29	1.34	0.07	5.28	7.46
Bi	0.173	0.194	0.18	0.01	8.09	11.44
Cd	0.08	0.07	0.08	0.01	9.43	13.33
Ce	34.07	33.6	33.84	0.33	0.98	1.39
Co	1.9	1.9	1.90	0.00	0.00	0.00
Cr	80	81	80.50	0.71	0.88	1.24
Cs	2.93	2.92	2.93	0.01	0.24	0.34
Cu	66	65	65.50	0.71	1.08	1.53
Dy	2.74	2.7	2.72	0.03	1.04	1.47
Er	1.51	1.52	1.52	0.01	0.47	0.66
Eu	0.68	0.71	0.70	0.02	3.05	4.32
Ga	13.48	13.73	13.61	0.18	1.30	1.84
Gd	2.91	2.88	2.90	0.02	0.73	1.04
Hf	1.18	1.18	1.18	0.00	0.00	0.00
Ho	0.52	0.52	0.52	0.00	0.00	0.00
La	15.71	15.33	15.52	0.27	1.73	2.45
Li	18.5	18.2	18.35	0.21	1.16	1.63
Lu	0.24	0.23	0.24	0.01	3.01	4.26
Mo	0.6	0.57	0.59	0.02	3.63	5.13
Nb	6.38	6.49	6.44	0.08	1.21	1.71
Nd	15.01	15.07	15.04	0.04	0.28	0.40
Ni	42	41	41.50	0.71	1.70	2.41
Pb	6	5.9	5.95	0.07	1.19	1.68
Pr	3.93	3.89	3.91	0.03	0.72	1.02
Rb	67.59	69.16	68.38	1.11	1.62	2.30
Sb	2.06	2.04	2.05	0.01	0.69	0.98
Sc	10.6	11	10.80	0.28	2.62	3.70
Sm	3.3	3.29	3.30	0.01	0.21	0.30
Sn	1.37	1.36	1.37	0.01	0.52	0.73
Sr	31	31	31.00	0.00	0.00	0.00
Ta	0.4	0.4	0.40	0.00	0.00	0.00
Tb	0.441	0.44	0.44	0.00	0.16	0.23
Th	5.48	5.37	5.43	0.08	1.43	2.03
Ti	2123.97	2159.2	2141.59	24.91	1.16	1.65
Tl	0.42	0.42	0.42	0.00	0.00	0.00
Tm	0.229	0.224	0.23	0.00	1.56	2.21
U	1.47	1.47	1.47	0.00	0.00	0.00
V	115.81	118.46	117.14	1.87	1.60	2.26
W	0.87	0.87	0.87	0.00	0.00	0.00
Y	13.11	13.2	13.16	0.06	0.48	0.68
Yb	1.532	1.554	1.54	0.02	1.01	1.43
Zn	120.86	122.74	121.80	1.33	1.09	1.54
Zr	48	50	49.00	1.41	2.89	4.08

Note: Duplicate sample run with original sample set.

	Sample 10-LT-05	Duplicate sample	Mean	Std. dev. (1σ)	RSD [%]	Percent diff. [%]
Units [wt %]						
SiO₂	49.42	49.24	49.33	0.36	0.13	0.26
TiO₂	1.42	1.42	1.42	0.00	0.00	0.00
Al₂O₃	18.49	18.45	18.47	0.22	0.03	0.15
Fe₂O₃	9.43	9.37	9.4	0.64	0.04	0.45
MnO	0.14	0.142	0.141	1.42	0.00	1.00
MgO	6.69	6.69	6.69	0.00	0.00	0.00
CaO	9.579	9.578	9.579	0.01	0.00	0.01
Na₂O	3.58	3.55	3.565	0.84	0.02	0.60
K₂O	1.03	1.03	1.03	0.00	0.00	0.00
P₂O₅	0.389	0.387	0.388	0.52	0.00	0.36
LOI	0.62	0.62	0.62	0.00	0.00	0.00
Total	100.79	100.48	100.64	0.31	0.22	0.22

Note: Duplicate sample run with original sample set.

Appendix C Assessment of uncertainty

C.3 University of Western Ontario

Sample	$\delta^{18}\text{O}$ [‰] VSMOW	Yield [$\mu\text{mol/mg}$]	Accepted $\delta^{18}\text{O}$ value	Std. dev. (1 σ) [‰]	RSD [%]	Percent error [%]	Percent diff. [%]
Laboratory standard quartz	11.4499404	16.14	11.5			0.44	
Laboratory standard quartz	11.5010880	16.22	11.5	0.06	0.53	0.01	
Laboratory standard quartz	11.3804512	16.4	11.5			1.04	
Laboratory standard basalt	7.5364940	12.77	7.5	0.06	0.78	0.49	
Laboratory standard basalt	7.6203920	13.79	7.5			1.61	
Laboratory standard CO ₂ (II)	10.2271028		10.3			0.71	
Laboratory standard CO ₂ (II)	10.3354287		10.3	0.06	0.59	0.34	
Laboratory standard CO ₂ (II)	10.3301910		10.3			0.29	
08-BA-6	8.8682680	14.9		0.23	2.67	N/A	3.8
08-BA-6-dup	8.5390568	15.53					
11-BA-46	4.1951404	14.76		0.26	6.54	N/A	9.3
11-BA-46-dup	3.8240588	15.02					

Std. dev. = Standard deviation

Percent diff = percent difference

All isotopic results are reported in the usual δ - notation relative to VSMOW (Vienna Standard Mean Ocean Water).

Appendix D

D.1 Aurora Volcanic Field rocks

Sample	DL	13-AU-01	13-AU-02	13-AU-03	13-AU-04	13-AU-05
Latitude		38.08647	38.08647	38.17448	38.19273	38.19804
Longitude		-119.14363	-119.14363	-118.92614	-118.91980	-118.92188
Units [wt %]						
SiO ₂	0.01	49.70	50.90	58.10	57.40	54.50
TiO ₂	0.01	0.96	0.94	1.02	0.95	1.17
Al ₂ O ₃	0.01	15.20	14.90	17.10	16.90	18.40
Fe ₂ O ₃	0.01	9.21	9.01	6.51	6.23	8.25
MnO	0.01	0.14	0.13	0.10	0.09	0.11
MgO	0.01	10.40	8.06	3.44	3.52	2.99
CaO	0.01	8.42	8.71	5.56	5.81	6.18
Na ₂ O	0.01	2.82	2.87	3.98	3.57	3.55
K ₂ O	0.01	1.61	1.78	3.86	3.62	3.54
P ₂ O ₅	0.01	0.41	0.41	0.63	0.62	0.56
LOI	0.01	1.87	3.24	0.32	1.27	1.26
Cr ₂ O ₃	0.01	0.08	0.08	<0.01	0.01	<0.01
Total (wet)		100.74	100.95	100.619	99.98	100.51

Method: wavelength dispersive X-ray fluorescence spectrometry (WDXRF).

DL = Detection Limits, LOI = gravimetric loss of ignition

Units [ppm]						
Ag	1	<1	<1	<1	<1	<1
As	30	<30	<30	<30	<30	<30
Ba	0.5	787	919	1570	1520	1530
Be	5	<5	<5	<5	<5	<5
Bi	0.1	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	0.2	0.2	0.2	0.2	<0.2	0.2
Ce	0.1	70.5	69.3	116.0	102.0	85.1
Co	0.5	48.6	45.0	22.0	21.7	27.4
Cr	10	530	500	60	60	10
Cs	0.1	0.4	2.0	1.5	1.2	3.5
Cu	5	62	56	25	28	39
Dy	0.05	3.46	3.45	4.45	4.03	5.49
Er	0.05	1.81	1.75	2.46	2.12	2.77
Eu	0.05	1.57	1.59	1.79	1.77	2.20
Ga	1	19	19	21	20	22
Gd	0.05	4.78	4.75	6.11	5.46	7.98
Ge	1	2	1	1	1	1
Hf	1	3	3	8	8	6
Ho	0.05	0.61	0.61	0.79	0.69	0.94
In	0.2	<0.2	<0.2	<0.2	<0.2	<0.2
La	0.1	33.4	34.3	59.1	50.0	44.4
Li	10	10	20	20	20	20
Lu	0.05	0.23	0.23	0.33	0.28	0.33
Mo	2	<2	<2	<2	3	2

Sample	DL	13-AU-01	13-AU-02	13-AU-03	13-AU-04	13-AU-05
Units [ppm]						
Nb	1	7	6	25	22	11
Nd	0.1	34.1	33.5	48.4	42.8	48.1
Ni	5	192	178	32	38	16
Pb	5	16	9	21	21	16
Pr	0.05	8.36	8.28	13.00	11.30	11.50
Rb	0.2	35.6	41.1	91.2	74.6	124.0
Sb	0.1	0.4	<0.1	0.1	0.3	0.6
Sc	5	24	24	13	12	16
Sm	0.1	6.2	5.8	7.6	6.9	9.0
Sn	1	1	<1	1	1	1
Sr	0.1	723.0	741.0	790.0	788.0	780.0
Ta	0.5	<0.5	<0.5	1.2	1	0.6
Tb	0.05	0.62	0.61	0.80	0.73	0.92
Th	0.1	5.6	5.6	6.8	6.3	13.3
Tl	0.5	<0.5	<0.5	<0.5	<0.5	0.8
Tm	0.05	0.25	0.26	0.35	0.33	0.40
U	0.05	1.29	1.51	2.00	2.05	4.66
V	5	220	208	128	122	181
W	1	<1	<1	<1	<1	1
Y	0.5	17.4	17.3	24.6	20.9	28.6
Yb	0.1	1.6	1.6	2.3	2.1	2.4
Zn	5	103	94	90	85	101
Zr	0.5	128	124	359	326	240

Method: inductively coupled plasma-atomic emission spectrometry (ICP-AES).

Sample	DL	13-AU-01	13-AU-02	13-AU-03	13-AU-04	13-AU-05
Units [wt%]						
Al		8.12	7.91	9.15	8.82	9.72
Ca		5.90	6.00	3.90	4.00	4.20
Fe		6.31	6.11	4.47	4.21	5.53
K		1.30	1.40	3.10	2.90	2.80
Mg		5.78	4.55	1.93	1.95	1.69
Mn		0.11	0.10	0.08	0.08	0.09
P		0.16	0.16	0.22	0.22	0.21
Ti		0.57	0.54	0.61	0.56	0.68

Method: ICP-AES.

Sample	13-AU-06	13-AU-07	13-AU-08	13-AU-09	13-AU-10
Latitude	38.20102	38.21463	38.36186	38.35775	38.31948
Longitude	-118.91386	-118.90026	-118.83071	-118.89468	-118.90932
Units [wt%]					
SiO2	63.00	64.70	56.70	56.70	59.50
TiO2	0.72	0.58	1.14	0.97	0.88
Al2O3	16.60	16.30	17.70	17.00	16.80
Fe2O3	4.47	3.59	7.06	6.84	6.17
MnO	0.05	0.06	0.11	0.11	0.10
MgO	2.30	1.60	3.32	3.90	2.93
CaO	4.46	3.53	5.80	6.49	5.17
Na2O	3.80	4.08	4.16	4.04	4.09
K2O	3.97	3.88	3.32	2.81	3.34
P2O5	0.44	0.31	0.70	0.54	0.53
LOI	1.33	1.66	0.31	1.12	0.67
Cr2O3	<0.01	<0.01	0.01	<0.01	<0.01
Total (wet)	101.14	100.29	100.318	100.52	100.184
Method: WDXRF.					

Units [ppm]					
Ag	<1	<1	<1	<1	<1
As	<30	<30	<30	<30	<30
Ba	1720	1800	1500	1350	1430
Be	<5	<5	<5	<5	<5
Bi	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	<0.2	<0.2	0.2	0.3	0.3
Ce	86.6	78.9	113.0	92.4	104.0
Co	14.2	10.0	21.5	24.4	18.8
Cr	30	20	30	70	50
Cs	1.4	1.7	1.1	0.7	0.7
Cu	11	10	22	29	24
Dy	2.95	2.79	4.77	4.00	4.16
Er	1.67	1.40	2.52	2.24	2.26
Eu	1.40	1.17	1.90	1.74	1.73
Ga	21	20	22	22	22
Gd	4.35	3.80	6.18	5.42	5.61
Ge	1	1	1	1	1
Hf	6	6	7	7	8
Ho	0.53	0.45	0.84	0.73	0.77
In	<0.2	<0.2	<0.2	<0.2	<0.2
La	43.4	40.2	55.3	45.5	52.0
Li	20	20	20	20	20
Lu	0.22	0.20	0.33	0.28	0.31
Mo	3	3	3	3	5
Nb	14	13	28	18	20
Nd	36.0	31.8	47.3	40.1	43.6
Ni	24	17	24	34	23
Pb	23	25	17	19	22

Sample	13-AU-06	13-AU-07	13-AU-08	13-AU-09	13-AU-10
Units [ppm]					
Pr	9.47	8.54	12.50	10.30	11.40
Rb	85.3	90.7	67.5	48.8	57.8
Sb	0.5	0.4	0.2	0.3	0.2
Sc	8	6	12	14	11
Sm	6.0	5.1	7.6	6.9	7.2
Sn	1	1	1	1	2
Sr	883.0	898.0	937.0	951.0	705.0
Ta	0.7	0.7	1.3	0.9	1.0
Tb	0.55	0.47	0.78	0.69	0.75
Th	6.2	6.4	5.3	4.0	4.6
Tl	<0.5	<0.5	<0.5	<0.5	<0.5
Tm	0.24	0.21	0.34	0.31	0.34
U	1.90	2.42	1.58	1.31	1.52
V	93	71	137	144	115
W	<1	1	<1	<1	1
Y	16.8	14.5	24.4	21.5	22.9
Yb	1.5	1.3	2.4	2.1	2.3
Zn	65	151	101	98	91
Zr	251	238	318	296	345

Method: ICP-AES.

Sample	13-AU-06	13-AU-07	13-AU-08	13-AU-09	13-AU-10
Units [wt%]					
Al	8.66	8.72	9.43	8.95	8.93
Ca	3.00	2.50	3.90	4.50	3.50
Fe	3.00	2.41	4.76	4.65	4.16
K	3.20	3.30	2.80	2.30	2.60
Mg	1.28	0.86	1.86	2.25	1.63
Mn	0.06	0.05	0.09	0.09	0.08
P	0.14	0.10	0.25	0.21	0.19
Ti	0.42	0.33	0.67	0.58	0.51

Method: ICP-AES.

Sample	13-AU-11	13-AU-12	13-AU-13	13-AU-15	13-AU-17
Latitude	38.27256	38.14443	38.14278	38.13365	38.16043
Longitude	-118.87419	-118.96147	-118.97952	-119.01835	-118.90283
Units [wt%]					
SiO₂	59.00	60.60	61.60	58.20	64.30
TiO₂	0.98	0.79	0.77	0.87	0.60
Al₂O₃	17.70	17.50	17.50	17.50	16.60
Fe₂O₃	5.78	5.31	5.19	6.28	3.93
MnO	0.09	0.07	0.07	0.10	0.06
MgO	2.20	2.11	2.20	3.00	1.74
CaO	4.96	5.18	5.06	5.73	4.29
Na₂O	4.16	4.00	3.87	3.47	4.18
K₂O	3.82	3.02	3.15	3.60	3.61
P₂O₅	0.67	0.26	0.28	0.43	0.32
LOI	0.78	1.07	1.13	1.83	1.01
Cr₂O₃	<0.01	<0.01	<0.01	<0.01	<0.01
Total (wet)	100.139	99.91	100.82	101.01	100.64
Method: WDXRF.					

Units [ppm]					
Ag	<1	<1	<1	<1	<1
As	<30	<30	<30	<30	<30
Ba	1820	1310	1320	1280	1560
Be	<5	<5	<5	<5	<5
Bi	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	<0.2	<0.2	<0.2	0.2	<0.2
Ce	112.0	53.9	53.4	90.4	72.4
Co	18.4	17.7	16.8	20.9	12.0
Cr	<10	<10	60	60	20
Cs	1.5	0.9	1.1	4.5	0.9
Cu	19	23	27	27	17
Dy	3.89	2.86	2.86	3.74	2.43
Er	2.05	1.63	1.54	1.92	1.26
Eu	1.79	1.19	1.16	1.67	1.23
Ga	22	21	21	22	20
Gd	5.60	4.03	3.77	5.36	3.53
Ge	1	1	1	2	1
Hf	7	4	4	7	5
Ho	0.71	0.51	0.52	0.66	0.43
In	<0.2	<0.2	<0.2	<0.2	<0.2
La	55.7	26.4	26.4	45.8	36.2
Li	20	10	20	20	20
Lu	0.28	0.20	0.20	0.26	0.17
Mo	7	2	2	2	<2
Nb	20	9	9	17	11
Nd	47.2	24.0	23.8	41.0	28.9
Ni	9	16	21	23	16
Pb	22	21	21	16	27

Sample	13-AU-11	13-AU-12	13-AU-13	13-AU-15	13-AU-17
Units [ppm]					
Pr	12.40	6.22	6.08	10.60	7.75
Rb	83.1	67.1	65.7	124.0	76.4
Sb	0.3	0.5	0.5	0.3	0.4
Sc	9	10	12	12	7
Sm	7.2	4.4	4.3	6.7	4.6
Sn	1	<1	1	1	1
Sr	1000.0	783.0	709.0	713.0	839.0
Ta	0.9	<0.5	<0.5	0.9	0.6
Tb	0.68	0.49	0.61	0.65	0.46
Th	6.7	4.5	4.4	13.0	6.7
Tl	<0.5	<0.5	<0.5	0.9	<0.5
Tm	0.30	0.22	0.22	0.29	0.19
U	2.02	1.44	1.56	3.38	1.82
V	121	119	113	115	64
W	<1	<1	<1	1	<1
Y	21.7	15.3	15.5	19.9	13.0
Yb	2.0	1.4	1.5	1.9	1.2
Zn	95	70	84	82	63
Zr	287	166	170	258	180

Method: ICP-AES.

Sample	13-AU-11	13-AU-12	13-AU-13	13-AU-15	13-AU-17
Units [wt%]					
Al	9.22	9.40	9.37	9.46	8.82
Ca	3.40	3.40	3.80	3.90	2.90
Fe	3.73	3.54	4.16	4.16	2.58
K	3.00	2.40	2.90	2.90	2.90
Mg	1.21	1.14	1.68	1.66	0.95
Mn	0.07	0.06	0.08	0.08	0.05
P	0.25	0.09	0.15	0.17	0.12
Ti	0.56	0.45	0.49	0.50	0.34

Method: ICP-AES.

Sample	13-AU-18	13-AU-19	13-AU-20	13-AU-21	13-AU-22
Latitude	38.15770	38.17276	38.17153	38.18961	38.19090
Longitude	-118.92379	-118.88896	-118.88057	-118.87365	-118.87123
Units [wt%]					
SiO₂	56.60	57.50	64.00	60.90	59.10
TiO₂	1.08	0.97	0.61	0.78	0.99
Al₂O₃	17.30	17.50	16.80	16.80	17.80
Fe₂O₃	6.94	5.89	4.19	4.96	6.09
MnO	0.11	0.09	0.07	0.07	0.06
MgO	3.67	3.15	1.69	2.07	3.23
CaO	5.85	6.04	4.17	4.73	5.99
Na₂O	3.89	3.79	4.33	3.90	3.92
K₂O	3.82	3.40	3.84	3.93	3.27
P₂O₅	0.73	0.48	0.35	0.45	0.52
LOI	0.67	0.91	0.49	0.42	0.42
Cr₂O₃	0.01	<0.01	<0.01	<0.01	0.01
Total (wet)	100.658	99.718	100.538	99.011	101.392
Method: WDXRF.					

Units [ppm]					
Ag	<1	<1	<1	<1	<1
As	<30	<30	<30	<30	<30
Ba	1540	1450	1420	1640	1470
Be	<5	<5	<5	<5	<5
Bi	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	<0.2	0.2	<0.2	<0.2	<0.2
Ce	119.0	84.1	79.2	97.2	84.8
Co	23.5	22.5	11.5	23.4	23.5
Cr	60	30	<10	<10	30
Cs	1.8	1.2	1.9	1.4	1.1
Cu	27	25	13	15	26
Dy	4.50	3.26	2.61	3.13	3.46
Er	2.43	1.80	1.43	1.72	1.80
Eu	1.93	1.56	1.27	1.55	1.63
Ga	20	20	20	20	22
Gd	6.10	4.82	3.66	4.26	5.04
Ge	1	1	1	1	1
Hf	8	5	5	6	5
Ho	0.79	0.60	0.48	0.57	0.61
In	<0.2	<0.2	<0.2	<0.2	<0.2
La	58.5	40.3	40.2	47.8	40.3
Li	20	10	20	20	20
Lu	0.30	0.26	0.18	0.22	0.21
Mo	3	2	2	2	2
Nb	28	13	12	13	13
Nd	48.8	37.2	32.2	40.4	39.0
Ni	36	25	11	12	28
Pb	19	21	26	25	21

Sample	13-AU-18	13-AU-19	13-AU-20	13-AU-21	13-AU-22
Units [ppm]					
Pr	13.10	9.62	8.68	10.70	9.76
Rb	95.3	66.5	89.0	87.3	62.2
Sb	0.4	0.5	0.7	0.3	0.5
Sc	13	11	6	8	12
Sm	7.6	6.2	4.7	6.1	6.4
Sn	2	1	1	<1	2
Sr	838.0	1180.0	910.0	992.0	1240.0
Ta	1.3	0.6	0.7	0.7	0.6
Tb	0.80	0.60	0.46	0.60	0.63
Th	6.2	5.4	8.4	8.5	5.3
Tl	<0.5	<0.5	<0.5	<0.5	<0.5
Tm	0.37	0.25	0.21	0.25	0.25
U	2.02	1.90	2.76	2.61	1.65
V	129	137	68	110	146
W	1	<1	1	1	<1
Y	23.8	17.7	14.6	16.9	18.3
Yb	2.3	1.7	1.4	1.6	1.7
Zn	86	71	65	97	71
Zr	363	219	196	220	221

Method: ICP-AES.

Sample	13-AU-18	13-AU-19	13-AU-20	13-AU-21	13-AU-22
Units [wt%]					
Al	9.05	8.93	8.86	9.00	9.41
Ca	4.00	4.00	2.70	3.10	4.10
Fe	4.66	3.95	2.71	3.32	4.08
K	3.10	2.70	3.10	3.10	2.90
Mg	2.02	1.78	0.88	1.14	1.81
Mn	0.09	0.07	0.06	0.07	0.07
P	0.25	0.18	0.13	0.14	0.18
Ti	0.63	0.55	0.34	0.45	0.57

Method: ICP-AES.

Sample	13-AU-23	13-AU-24	13-AU-25	13-AU-26	13-AU-27
Latitude	38.16506	38.18782	38.23149	38.23682	38.28243
Longitude	-118.79920	-118.79659	-118.82353	-118.82573	-118.85816
Units [wt%]					
SiO₂	57.70	56.30	60.70	55.40	63.40
TiO₂	0.80	1.09	0.84	1.34	0.62
Al₂O₃	17.70	17.20	16.40	18.00	16.50
Fe₂O₃	5.33	7.01	4.98	7.45	3.93
MnO	0.09	0.11	0.07	0.10	0.06
MgO	2.25	4.13	2.91	3.34	2.03
CaO	5.33	6.23	4.89	6.06	3.90
Na₂O	4.09	3.80	3.52	4.14	4.07
K₂O	3.40	3.60	3.92	3.54	3.81
P₂O₅	0.45	0.69	0.50	0.91	0.34
LOI	0.98	0.64	1.37	0.77	0.97
Cr₂O₃	<0.01	0.02	<0.01	<0.01	<0.01
Total (wet)	98.119	100.798	100.1	101.046	99.634
Method: WDXRF.					

Units [ppm]					
Ag	<1	<1	<1	<1	<1
As	<30	<30	<30	<30	<30
Ba	1440	1580	1660	1660	1660
Be	<5	<5	<5	<5	<5
Bi	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	<0.2	<0.2	<0.2	<0.2	<0.2
Ce	87.0	115.0	95.0	130.0	80.9
Co	16.4	25.8	18.8	21.9	12.6
Cr	<10	70	50	40	30
Cs	1.5	1.4	2.0	1.7	1.3
Cu	20	29	15	23	13
Dy	3.03	4.50	3.36	4.85	2.80
Er	1.64	2.27	1.71	2.51	1.45
Eu	1.5	1.95	1.54	2.27	1.27
Ga	21	20	21	21	21
Gd	4.42	6.02	4.88	6.69	4.10
Ge	1	1	1	2	1
Hf	5	7	6	8	6
Ho	0.55	0.77	0.60	0.88	0.49
In	<0.2	<0.2	<0.2	<0.2	<0.2
La	41.9	56.1	46.5	64.2	40.8
Li	10	20	20	10	20
Lu	0.22	0.30	0.22	0.34	0.18
Mo	3	3	3	3	3
Nb	13	25	14	33	13
Nd	37.5	48.1	40.6	56.0	33.5
Ni	12	41	34	23	23
Pb	22	18	23	31	27

Sample	13-AU-23	13-AU-24	13-AU-25	13-AU-26	13-AU-27
Units [ppm]					
Pr	9.87	12.80	10.60	14.90	8.97
Rb	70.5	84.4	89.3	77.1	85.4
Sb	0.3	0.2	0.4	0.8	0.3
Sc	8	14	10	13	7
Sm	5.9	7.6	6.4	8.4	5.3
Sn	1	1	<1	2	<1
Sr	1160.0	887.0	932.0	1080.0	957.0
Ta	0.6	1.2	0.8	1.5	0.7
Tb	0.58	0.79	0.62	0.83	0.51
Th	6.7	6.1	6.5	5.8	6.5
Tl	<0.5	<0.5	<0.5	<0.5	1.1
Tm	0.24	0.36	0.27	0.38	0.21
U	2.40	1.88	2.01	1.72	2.08
V	112	138	111	140	78
W	<1	<1	1	<1	<1
Y	16.3	23.6	18.0	27.0	15.4
Yb	1.6	2.2	1.5	2.4	1.4
Zn	68	87	66	92	111
Zr	209	339	245	339	241

Method: ICP-AES.

Sample	13-AU-23	13-AU-24	13-AU-25	13-AU-26	13-AU-27
Units [wt%]					
Al	9.40	9.02	8.74	9.63	8.85
Ca	3.60	4.20	3.30	4.10	2.70
Fe	3.55	4.67	3.34	4.96	2.70
K	2.80	2.90	3.20	2.80	3.10
Mg	1.23	2.32	1.63	1.91	1.14
Mn	0.07	0.09	0.06	0.09	0.05
P	0.15	0.25	0.19	0.31	0.12
Ti	0.46	0.63	0.49	0.78	0.37

Method: ICP-AES.

Sample	13-AU-28	13-AU-29	13-AU-30	13-AU-31	13-AU-32
Latitude	38.27756	38.29236	38.32452	38.33475	38.15556
Longitude	-118.85991	-118.84699	-118.85579	-118.85823	-118.95087
Units [wt%]					
SiO₂	65.10	54.40	59.70	58.00	52.20
TiO₂	0.61	1.34	0.86	0.98	1.06
Al₂O₃	16.60	17.40	16.80	17.00	17.30
Fe₂O₃	3.84	8.00	6.13	6.82	9.34
MnO	0.04	0.12	0.10	0.10	0.14
MgO	1.85	4.55	2.91	3.42	6.31
CaO	3.74	6.54	5.18	5.75	8.93
Na₂O	4.11	4.00	4.01	4.03	3.66
K₂O	3.89	2.94	3.32	3.04	1.51
P₂O₅	0.33	0.73	0.53	0.53	0.34
LOI	0.92	0.76	0.88	0.80	0.43
Cr₂O₃	0.03	<0.01	<0.01	<0.01	0.03
Total (wet)	101.026	100.777	100.417	100.468	101.219
Method: WDXRF.					

Units [ppm]					
Ag	<1	<1	<1	<1	<1
As	<30	<30	<30	<30	<30
Ba	1640	1490	1450	1300	1000
Be	<5	<5	<5	<5	<5
Bi	<0.1	<0.1	<0.1	<0.1	<0.1
Cd	<0.2	<0.2	<0.2	<0.2	0.2
Ce	79.5	114.0	101.0	95.8	59.1
Co	11.6	28.6	18.6	21.6	43.3
Cr	20	90	50	50	150
Cs	1.4	0.7	0.7	0.8	0.3
Cu	10	26	25	30	54
Dy	2.59	4.75	4.27	4.30	3.86
Er	1.43	2.55	2.27	2.43	2.22
Eu	1.2	2.04	1.67	1.72	1.53
Ga	20	22	22	22	20
Gd	3.75	6.36	5.53	5.68	4.72
Ge	1	1	1	2	2
Hf	6	7	8	7	3
Ho	0.46	0.85	0.75	0.76	0.71
In	<0.2	<0.2	<0.2	<0.2	<0.2
La	39.7	55.2	50.4	46.3	27.0
Li	20	20	20	20	10
Lu	0.19	0.34	0.30	0.32	0.26
Mo	3	3	4	4	<2
Nb	13	27	20	19	7
Nd	32.0	49.5	40.8	40.9	28.6
Ni	23	49	22	26	81
Pb	26	16	21	20	18

Sample	13-AU-28	13-AU-29	13-AU-30	13-AU-31	13-AU-32
Units [ppm]					
Pr	8.74	13.30	11.00	10.70	7.05
Rb	86.9	52.8	58.4	52.8	22.8
Sb	0.5	0.3	0.1	<0.1	0.5
Sc	6	14	11	13	23
Sm	4.9	8.0	7.0	7.0	5.1
Sn	<1	2	2	1	<1
Sr	914.0	1030.0	740.0	775.0	982.0
Ta	0.7	1.2	0.9	0.9	<0.5
Tb	0.45	0.84	0.74	0.75	0.61
Th	6.4	5.1	4.5	4.2	3.7
Tl	<0.5	<0.5	<0.5	<0.5	<0.5
Tm	0.20	0.35	0.33	0.35	0.30
U	2.15	1.43	1.55	1.37	1.09
V	75	159	116	135	184
W	<1	<1	<1	<1	<1
Y	14.3	25.9	23.0	23.4	20.5
Yb	1.4	2.3	2.2	2.3	2.1
Zn	54	214	144	101	88
Zr	237	310	349	315	133

Method: ICP-AES.

Sample	13-AU-28	13-AU-29	13-AU-30	13-AU-31	13-AU-32
Units [wt%]					
Al	8.67	9.24	8.91	8.92	9.06
Ca	2.40	4.40	3.50	3.80	5.90
Fe	2.53	5.34	4.11	4.52	6.22
K	3.00	2.40	2.70	2.50	1.30
Mg	1.01	2.53	1.63	1.88	3.51
Mn	0.05	0.10	0.08	0.08	0.11
P	0.11	0.25	0.18	0.18	0.11
Ti	0.35	0.79	0.50	0.55	0.62

Method: ICP-AES.

Sample	13-AU-33	13-BA-4
Latitude	38.16336	38.24296
Longitude	-118.98196	-118.82885
Units [wt%]		
SiO2	64.10	60.00
TiO2	0.60	0.87
Al2O3	17.00	17.90
Fe2O3	4.03	5.33
MnO	0.07	0.07
MgO	1.25	1.76
CaO	3.59	4.19
Na2O	4.17	4.37
K2O	3.84	4.13
P2O5	0.26	0.51
LOI	1.14	0.68
Cr2O3	<0.01	<0.01
Total (wet)	100.05	99.812
Method: WDXRF.		

Units [ppm]		
Ag	<1	<1
As	<30	<30
Ba	1550	1870
Be	<5	<5
Bi	<0.1	<0.1
Cd	<0.2	<0.2
Ce	65.6	115.0
Co	8.5	14.9
Cr	<10	<10
Cs	1.3	1.2
Cu	7	13
Dy	3.00	3.85
Er	1.53	1.95
Eu	1.14	1.75
Ga	21	22
Gd	3.56	5.12
Ge	1	1
Hf	6	8
Ho	0.51	0.66
In	<0.2	<0.2
La	33.4	57.9
Li	20	20
Lu	0.23	0.28
Mo	3	3
Nb	11	21
Nd	27.3	45.8
Ni	7	9
Pb	26	25

Sample	13-AU-33	13-BA-4
Units [ppm]		
Pr	7.28	12.70
Rb	83.4	89.9
Sb	1.4	0.3
Sc	5	7
Sm	4.6	7.0
Sn	1	<1
Sr	684.0	902.0
Ta	0.6	1.0
Tb	0.47	0.82
Th	6.1	8.3
Tl	<0.5	<0.5
Tm	0.23	0.29
U	1.88	2.44
V	62	104
W	<1	1
Y	15.8	20.1
Yb	1.6	2.0
Zn	68	78
Zr	221	325

Method: ICP-AES.

Sample	13-AU-33	13-BA-4
Units [wt%]		
Al	8.99	9.50
Ca	2.40	2.80
Fe	2.63	3.55
K	3.10	3.30
Mg	0.67	0.95
Mn	0.05	0.06
P	0.09	0.17
Ti	0.34	0.50

Method: ICP-AES.

Appendix D

D.2 Basement rocks

Sample Units [wt%]	DL	203350	203377	203387	BPC 13-1	CC 13-1	HOM 13-1	MAS 13-2
SiO ₂	0.04	75.85	71.74	71.72	83.22	91.17	63.08	71.33
TiO ₂	0.01	0.18	0.24	0.39	0.37	0.08	0.61	0.18
Al ₂ O ₃	0.02	13.78	15.25	14.35	7.61	1.63	16.15	14.09
Fe ₂ O ₃	0.01	0.82	1.81	2.80	3.55	0.68	5.20	3.51
MnO	0.002	0.010	0.047	0.071	0.014	0.036	0.092	0.010
MgO	0.01	0.14	0.52	0.66	1.76	1.38	2.25	0.48
CaO	0.006	0.437	2.128	1.684	0.043	2.381	3.625	0.126
Na ₂ O	0.02	4.35	3.92	3.80	0.04	0.06	3.63	1.28
K ₂ O	0.01	4.79	4.07	3.97	1.71	0.60	4.49	8.85
P ₂ O ₅	0.002	0.031	0.093	0.203	0.051	0.164	0.236	0.060
LOI		0.26	0.65	0.79	2.01	1.00	0.95	0.60
Total (wet)		100.66	100.46	100.45	100.37	99.18	100.32	100.52

Method: X-ray fluorescence spectrometry (XRF). LOI = loss of ignition

Units [ppm]								
Ba	0.9	724.4	846.0	1489.7	>2400	853.1	1481.6	>2400
Be	0.06	2.10	2.64	2.16	1.39	0.46	1.69	1.18
Bi	0.009	0.129	0.032	0.046	0.173	0.141	0.185	0.871
Cd	0.01	0.03	0.10	0.05	0.08	2.55	0.05	0.02
Ce	0.2	48.19	46.81	51.33	34.07	12.75	48.65	25.91
Co	0.1	0.3	2.2	2.4	1.9	2.8	13.4	3.6
Cr	24	<24	<24	<24	80	34	34	<24
Cs	0.006	1.98	3.72	3.80	2.93	1.05	4.16	1.56
Cu	2	2	<2	4	66	42	12	147
Dy	0.02	2.47	1.00	3.03	2.74	1.54	3.22	1.04
Er	0.01	1.20	0.49	1.66	1.51	1.02	1.89	0.56
Eu	0.005	0.59	0.59	1.00	0.68	0.30	1.05	0.39
Ga	0.05	15.98	19.38	18.12	13.48	2.51	17.69	13.20
Gd	0.02	2.94	1.44	3.39	2.91	1.62	3.56	1.50
Hf	0.09	0.61	0.76	1.20	1.18	0.41	1.28	2.08
Ho	0.003	0.44	0.17	0.59	0.52	0.34	0.64	0.19
La	0.09	24.05	31.85	26.78	15.71	9.19	25.58	13.94
Li	0.2	7.1	22.9	22.9	18.5	4.0	13.1	21.4
Lu	0.002	0.16	0.08	0.24	0.24	0.15	0.27	0.11
Mo	0.03	0.31	0.31	0.33	0.60	3.94	1.24	1.42
Nb	0.04	8.27	6.61	10.19	6.38	2.01	8.78	7.08
Nd	0.08	20.18	15.15	20.75	15.01	8.10	21.25	10.81
Ni	3	<3	<3	<3	42	39	12	<3
Pb	0.4	12.3	16.5	15.2	6.0	3.0	11.6	7.1
Pr	0.02	5.67	4.64	5.65	3.93	2.07	5.76	3.00
Rb	0.2	127.33	156.72	155.56	67.59	22.52	112.14	219.44
Sb	0.04	0.33	0.12	0.36	2.06	3.56	1.18	1.16
Sc	0.9	3.7	2.4	6.4	10.6	2.5	14.2	3.8

Sample	DL	203350	203377	203387	BPC	CC	HOM	MAS
Units [wt %]					13-1	13-1	13-1	13-2
Sm	0.02	3.97	2.30	4.00	3.30	1.54	4.22	2.05
Sn	0.08	1.22	0.80	1.69	1.37	0.36	1.33	0.95
Sr	2	40	508	247	31	58	447	306
Ta	0.2	0.8	0.5	0.8	0.4	<0.2	0.7	0.7
Tb	0.003	0.426	0.180	0.503	0.441	0.230	0.522	0.205
Th	0.09	14.21	16.93	17.88	5.48	1.70	15.52	22.17
Ti	26	877.21	1484.96	2474.87	2123.97	425.12	3685.01	1003.63
Tl	0.005	0.48	0.66	0.73	0.42	0.17	0.42	0.68
Tm	0.002	0.172	0.073	0.241	0.229	0.147	0.281	0.088
U	0.02	3.20	2.52	3.17	1.47	1.90	4.53	7.20
V	10	<10	22.25	25.40	115.81	103.27	108.63	26.82
W	0.5	0.89	<0.5	<0.5	0.87	<0.5	1.86	3.77
Y	0.08	12.11	5.63	17.16	13.11	12.72	18.84	5.43
Yb	0.009	1.101	0.488	1.598	1.532	0.969	1.793	0.659
Zn	8	18.31	41.88	64.52	120.86	146.56	63.63	12.70
Zr	3	13	15	26	48	19	45	72

Method: inductively coupled plasma mass spectrometry (ICP-MS).

Appendix E Whole rock isotopic data: Lead, strontium, neodymium, and oxygen isotope analyses

Sample	U [ppm]	Th [ppm]	Pb [ppm]	Age [Ma]	206Pb/204Pb initial	2-sigma	207Pb/204Pb initial	2-sigma	208Pb/204Pb initial	2-sigma
01-BA-1	1.26	4.6	12	3.50	19.091	0.001	15.681	0.001	38.921	0.004
08-BA-3	1.61	4.8	23	0.36	19.030	0.001	15.653	0.001	38.784	0.005
08-BA-6	6.66	17.5	39	2.60	19.154	0.001	15.691	0.001	38.947	0.004
08-BA-7	1.87	6.0	17	1.32	19.091	0.001	15.653	0.001	38.802	0.003
11-BA-46	5.55	13.1	37	3.70	19.151	0.001	15.681	0.001	38.914	0.002
12-BA-32	1.61	6.4	11	2.60	19.238	0.002	15.690	0.003	38.954	0.009
12-BA-33	2.17	6.7	25	2.95	19.148	0.001	15.677	0.001	38.893	0.004
203176	1.59	4.9	25	2.95	19.079	0.000	15.669	0.000	38.874	0.009
203181	5.54	15.5	22	2.80	19.097	0.002	15.658	0.002	38.878	0.004
203184	0.97	3.4	14	3.90	18.988	0.001	15.640	0.001	38.715	0.005
203195	1.05	3.5	15	3.90	18.974	0.001	15.628	0.001	38.671	0.002
203230	2.02	6.2	13	2.00	19.052	0.001	15.644	0.001	38.801	0.003
203232	7.77	23.2	21	2.00	19.076	0.001	15.633	0.001	38.789	0.004
203246	8.42	23.8	21	2.80	19.070	0.001	15.635	0.001	38.783	0.004
203247	5.16	15.0	18	2.80	19.069	0.002	15.646	0.002	38.831	0.005
203326	1.18	3.9	14	2.90	19.147	0.001	15.679	0.001	38.905	0.003
203354	6.88	20.0	18	2.80						
203382	8.43	25.2	34	3.70	19.026	0.002	15.651	0.002	38.825	0.007
203393	1.31	4.0	18	0.36						
203394	1.30	4.2	19	0.36	19.034	0.000	15.646	0.000	38.763	0.002
203395	1.21	4.2	18	0.28	19.060	0.003	15.652	0.003	38.790	0.011
203398	1.69	5.8	15	1.45	19.103	0.001	15.653	0.002	38.802	0.005
203402	1.44	4.2	24	2.00	19.246	0.001	15.688	0.001	38.945	0.003
203443	1.51	5.3	15	1.45	19.096	0.001	15.662	0.001	38.837	0.002

Sample	U [ppm]	Th [ppm]	Pb [ppm]	Age [Ma] ⁴	206Pb/204Pb initial	2-sigma	207Pb/204Pb initial	2-sigma	208Pb/204Pb initial	2-sigma
Basement										
203350	3.2	14.21	12.3	4	19.077	0.001	15.658	0.002	38.813	0.005
203377	2.52	16.93	16.5	4	19.107	0.001	15.674	0.001	38.924	0.004
203387	3.17	17.88	15.2	4	19.158	0.002	15.667	0.002	38.796	0.006
BPC-13-1	1.47	5.48	6	4	19.611	0.001	15.708	0.000	39.307	0.001
CC-13-1	1.9	1.7	3	4	23.355	0.004	15.874	0.003	39.512	0.007
HOM-13-1	4.53	15.52	11.6	4	19.562	0.004	15.690	0.003	39.114	0.009
MAS-13-2	7.2	22.17	7.1	4	20.453	0.001	15.695	0.000	39.788	0.001

⁴ Age of 4 Ma is used for basement rocks as it is the oldest age for the AVF.

Sample	Rb [ppm]	Sr [ppm]	⁸⁷ Rb/ ⁸⁶ Sr present ⁵	⁸⁷ Sr/ ⁸⁶ Sr present	2-sigma	Age [Ma]	⁸⁷ Sr/ ⁸⁶ Sr initial
01-BA-1	62.5	701	0.2579	0.706357	0.000010	3.5	0.706344
08-BA-3	50.7	818	0.1793	0.706038	0.000007	0.36	0.706037
08-BA-6	132	175	2.1823	0.706890	0.000010	2.6	0.706809
08-BA-7	68.2	1080	0.1827	0.706307	0.000011	1.32	0.706303
11-BA-46	150	254	1.7086	0.706998	0.000007	3.7	0.706908
12-BA-32	38.9	1610	0.0699	0.706935	0.000007	2.6	0.706933
12-BA-33	81.2	940	0.2499	0.706781	0.000013	2.95	0.706771
203176	73.6	676	0.3150	0.706306	0.000045	2.95	0.706293
203181	142	643	0.6389	0.705656	0.000007	2.8	0.705631
203184	29.4	907	0.0938	0.705133	0.000019	3.9	0.705128
203195	27.8	917	0.0877	0.705113	0.000012	3.9	0.705109
203230	53.9	803	0.1942	0.705518	0.000011	2	0.705513
203232	143	679	0.6092	0.705694	0.000014	2	0.705677
203246	156	676	0.6676	0.705789	0.000004	2.8	0.705763
203247	109	781	0.4037	0.705881	0.000014	2.8	0.705865
203326	38.6	1000	0.1117	0.706166	0.000012	2.9	0.706162
203354	145	636	0.6595	0.705667	0.000016	2.8	0.705641
203382	194	29.5	19.0280	0.707714	0.000013	3.7	0.706715
203393	48.5	872	0.1609	0.705919	0.000011	0.36	0.705918
203394	51.8	817	0.1834	0.705930	0.000012	0.36	0.705929
203395	45.6	1060	0.1244	0.705851	0.000005	0.275	0.705851
203398	71.4	1040	0.1986	0.706291	0.000004	1.45	0.706286
203402	52.6	1020	0.1492	0.706668	0.000011	2	0.706664
203443	65.5	1130	0.1677	0.706254	0.000008	1.45	0.706251
13-AU-01	35.6	723	0.1424	0.705645	0.000014	11.7	0.705621
13-AU-02	41.1	741	0.1605	0.705637	0.000046	11.7	0.705611
13-AU-03	90.9	788	0.3340	0.706427	0.000020	3.7	0.706410
13-AU-04	74.6	788	0.2739	0.706442	0.000008	3.7	0.706427
13-AU-05	124	780	0.4599	0.705775	0.000008	3.7	0.705750
13-AU-06	85.3	883	0.2795	0.706777	0.000007	2.95	0.706765

⁵The atomic isotope ratio, ⁸⁷Rb/⁸⁶Sr is calculated (from equation 8 in text) using the ICP-MS data.

Sample	Rb [ppm]	Sr [ppm]	⁸⁷ Rb/ ⁸⁶ Sr present ^s	⁸⁷ Sr/ ⁸⁶ Sr present	2-sigma	Age [Ma]	⁸⁷ Sr/ ⁸⁶ Sr initial
13-AU-07	90.7	898	0.2922	0.706884	0.000015	2.95	0.706872
13-AU-08	67.5	937	0.2084	0.706256	0.000011	1.32	0.706252
13-AU-09	48.8	951	0.1484	0.705895	0.000013	0.36	0.705895
13-AU-10	57.8	705	0.2372	0.706014	0.000017	0.36	0.706013
13-AU-11	83.1	1000	0.2404	0.706624	0.000015	2.5	0.706615
13-AU-12	67.1	783	0.2479	0.706461	0.000012	2.95	0.706450
13-AU-13	65.7	709	0.2681	0.706433	0.000012	2.95	0.706422
13-AU-15	124	713	0.5031	0.705856	0.000036	9.4	0.705788
13-AU-17	76.4	839	0.2635	0.706868	0.000017	2.95	0.706857
13-AU-18	95.3	838	0.3290	0.706409	0.000016	3.7	0.706392
13-AU-19	66.5	1180	0.1630	0.706653	0.000020	3.7	0.706645
13-AU-20	89	910	0.2830	0.706690	0.000013	2.95	0.706679
13-AU-21	87.3	992	0.2546	0.706814	0.000011	2.95	0.706804
13-AU-22	62.2	1240	0.1451	0.706685	0.000020	2.8	0.706679
13-AU-23	70.5	1160	0.1758	0.706667	0.000019	3.0	0.706660
13-AU-24	84.4	887	0.2753	0.706504	0.000014	3.7	0.706489
13-AU-25	89.3	932	0.2772	0.706775	0.000008	2.95	0.706764
13-AU-26	77.1	1080	0.2065	0.706260	0.000011	1.6	0.706255
13-AU-27	85.4	957	0.2582	0.706864	0.000010	2.7	0.706854
13-AU-28	86.9	914	0.2751	0.706874	0.000009	2.7	0.706864
13-AU-29	52.8	1030	0.1483	0.706134	0.000012	1.6	0.706130
13-AU-30	58.4	740	0.2283	0.706036	0.000011	0.36	0.706035
13-AU-31	52.8	775	0.1971	0.705944	0.000014	0.36	0.705943
13-AU-32	22.8	982	0.0672	0.706131	0.000015	4 ⁶	0.706127
13-AU-33	83.4	684	0.3527	0.706378	0.000010	2.95	0.706363
13-BA-04	89.9	902	0.2883	0.706560	0.000016	2.95	0.706548

⁶ Age for sample 13-AU-32 is unknown, so the oldest age for the Aurora Volcanic Field is used (4Ma).

Sample	Rb [ppm]	Sr [ppm]	⁸⁷ Rb/ ⁸⁶ Sr present ⁵	⁸⁷ Sr/ ⁸⁶ Sr present	2-sigma	Age [Ma] ⁴	⁸⁷ Sr/ ⁸⁶ Sr initial
Basement rocks							
203350	127.33	40	9.2187	0.716755	0.000012	4	0.716231
203377	156.72	508	0.8926	0.706985	0.000021	4	0.706934
203387	155.56	247	1.8222	0.707020	0.000012	4	0.706917
BPC-13-1	67.59	31	6.3200	0.726158	0.000010	4	0.725799
CC-13-1	22.52	58	1.1248	0.720188	0.000010	4	0.720124
HOM-13-1	112.14	447	0.7258	0.706161	0.000007	4	0.706120
MAS-13-2	219.44	306	2.0755	0.710186	0.000011	4	0.710068

Sample	Sm [ppm]	Nd [ppm]	$^{147}\text{Sm}/^{144}\text{Nd}$ present ⁷	$^{143}\text{Nd}/^{144}\text{Nd}$ present	2-sigma	Age [Ma]	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	Epsilon Nd, T ⁸
01-BA-1	4.6	27.8	0.0996	0.512464	0.000006	3.5	0.512462	-3.19
08-BA-3	7.9	46.8	0.1017	0.512488	0.000008	0.36	0.512488	-2.77
08-BA-6	3.6	24.0	0.0903	0.512434	0.000008	2.6	0.512433	-3.78
08-BA-7	8.7	52.7	0.0994	0.512460	0.000006	1.32	0.512459	-3.30
11-BA-46	2.4	16.7	0.0865	0.512426	0.000005	3.7	0.512424	-3.92
12-BA-32	10.4	65.4	0.0958	0.512384	0.000005	2.6	0.512382	-4.77
12-BA-33	5.7	34.1	0.1007	0.512438	0.000006	2.95	0.512436	-3.72
203176	4.8	26.8	0.1079	0.512530	0.000053	2.95	0.512528	-1.91
203181	6.3	35.0	0.1084	0.512511	0.000011	2.8	0.512509	-2.29
203184	4.7	23.1	0.1225	0.512585	0.000009	3.9	0.512582	-0.85
203195	5.2	25.6	0.1223	0.512598	0.000009	3.9	0.512595	-0.58
203230	5.5	29.6	0.1119	0.512489	0.000008	2	0.512488	-2.73
203232	6.5	37.2	0.1052	0.512515	0.000007	2	0.512514	-2.22
203246	7.7	39.9	0.1162	0.512506	0.000003	2.8	0.512504	-2.39
203247	6.5	35.2	0.1112	0.512494	0.000011	2.8	0.512492	-2.62
203326	6.2	36.3	0.1029	0.512497	0.000007	2.9	0.512495	-2.57
203354	7.0	37.7	0.1118	0.512535	0.000006	2.8	0.512533	-1.83
203382	2.6	14.6	0.1072	0.512510	0.000009	3.7	0.512508	-2.30
203393	7.5	43.4	0.1041	0.512494	0.000010	0.36	0.512494	-2.64
203394	7.6	42.7	0.1072	0.512502	0.000006	0.36	0.512502	-2.49
203395	6.9	39.1	0.1063	0.512508	0.000005	0.275	0.512508	-2.38
203398	8.9	55.2	0.0971	0.512472	0.000004	1.45	0.512471	-3.06
203402	4.6	24.7	0.1121	0.512458	0.000005	2	0.512457	-3.33
203443	8.0	54.1	0.0890	0.512469	0.000010	1.45	0.512468	-3.12
13-AU-01	6.2	34.1	0.1095	0.512440	0.000006	11.7	0.512432	-3.57
13-AU-02	5.8	33.5	0.1043	0.512455	0.000005	11.7	0.512447	-3.28
13-AU-03	7.6	48.3	0.0946	0.512421	0.000006	3.7	0.512418	-4.04
13-AU-04	6.9	42.8	0.0971	0.512445	0.000003	3.7	0.512443	-3.56

⁷ The atomic isotope ratio, $^{147}\text{Sm}/^{144}\text{Nd}$ is calculated using the ICP-MS data.

⁸ Epsilon Nd (T) is calculated using a CHUR value of 0.512630 (Bouvier et al., 2008).

Sample	Sm [ppm]	Nd [ppm]	¹⁴⁷ Sm/ ¹⁴⁴ Nd present ⁷	¹⁴³ Nd/ ¹⁴⁴ Nd present	2-sigma	Age [Ma]	¹⁴³ Nd/ ¹⁴⁴ Nd initial	Epsilon Nd, T ⁸
13-AU-05	9.0	48.1	0.1127	0.512539	0.000008	3.7	0.512536	-1.73
13-AU-06	6.0	36.0	0.1004	0.512441	0.000009	2.95	0.512439	-3.64
13-AU-07	5.1	31.8	0.0966	0.512449	0.000006	2.95	0.512448	-3.48
13-AU-08	7.6	47.3	0.0968	0.512475	0.000005	1.32	0.512474	-3.01
13-AU-09	6.9	40.1	0.1036	0.512496	0.000006	0.36	0.512496	-2.61
13-AU-10	7.2	43.6	0.0994	0.512485	0.000007	0.36	0.512485	-2.83
13-AU-11	7.2	47.2	0.0919	0.512433	0.000005	2.45	0.512431	-3.81
13-AU-12	4.4	24.0	0.1104	0.512474	0.000006	2.95	0.512472	-3.01
13-AU-13	4.3	23.8	0.1088	0.512465	0.000004	2.95	0.512463	-3.19
13-AU-15	6.7	41.0	0.0984	0.512438	0.000006	9.4	0.512432	-3.63
13-AU-17	4.6	28.9	0.0959	0.512404	0.000010	2.95	0.512403	-4.36
13-AU-18	7.6	48.8	0.0938	0.512430	0.000004	3.7	0.512428	-3.86
13-AU-19	6.2	37.2	0.1004	0.512416	0.000003	3.7	0.512414	-4.13
13-AU-20	4.7	32.2	0.0879	0.512417	0.000008	2.95	0.512415	-4.12
13-AU-21	6.1	40.4	0.0909	0.512408	0.000007	2.95	0.512406	-4.30
13-AU-22	6.4	39.0	0.0988	0.512418	0.000004	2.80	0.512416	-4.11
13-AU-23	5.9	37.5	0.0947	0.512413	0.000006	2.95	0.512411	-4.20
13-AU-24	7.6	48.1	0.0951	0.512399	0.000007	3.7	0.512397	-4.46
13-AU-25	6.4	40.6	0.0949	0.512419	0.000006	2.95	0.512417	-4.08
13-AU-26	8.4	56.0	0.0903	0.512468	0.000006	1.6	0.512467	-3.14
13-AU-27	5.3	33.5	0.0953	0.512437	0.000009	2.7	0.512436	-3.72
13-AU-28	4.9	32.0	0.0922	0.512450	0.000006	2.7	0.512448	-3.48
13-AU-29	8.0	49.5	0.0973	0.512496	0.000009	1.6	0.512495	-2.60
13-AU-30	7.0	40.8	0.1033	0.512485	0.000010	0.36	0.512485	-2.82
13-AU-31	7.0	40.9	0.1031	0.512505	0.000013	0.36	0.512505	-2.43
13-AU-32	5.1	28.6	0.1074	0.512477	0.000009	4 ^b	0.512474	-2.94
13-AU-33	4.6	27.3	0.1015	0.512486	0.000007	2.95	0.512484	-2.77
13-BA-04	7.0	45.8	0.0920	0.512446	0.000007	2.95	0.512444	-3.56

Sample	Sm [ppm]	Nd [ppm]	$^{147}\text{Sm}/^{144}\text{Nd}$ present ⁷	$^{143}\text{Nd}/^{144}\text{Nd}$ present	2-sigma	Age [Ma] ⁴	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	Epsilon Nd, T ⁸
Basement								
203350	3.97	20.2	0.1185	0.512640	0.000009	4	0.512637	0.23
203377	2.30	15.2	0.0914	0.512449	0.000012	4	0.512446	-3.48
203387	4.00	20.8	0.1161	0.512521	0.000006	4	0.512518	-2.09
BPC 13-1	3.30	15.0	0.1324	0.511997	0.000006	4	0.511994	-12.31
CC 13-1	1.54	8.1	0.1145	0.512013	0.000010	4	0.512010	-12.00
HOM 13-1	4.22	21.3	0.1196	0.512651	0.000007	4	0.512648	0.46
MAS 13-2	2.05	10.8	0.1142	0.512651	0.000007	4	0.512648	0.45

Sample	$\delta^{18}\text{O}$ VSMOW	Yield $\mu\text{mol/mg}$
203181	6.98	13.69
203184	6.41	13.24
203230	7.52	13.48
203232	8.60	13.72
203326	7.38	12.83
203395	7.89	13.81
203402	8.25	14.22
08-BA-6	8.87	14.90
08-BA-6-dup	8.54	15.53
11-BA-46	4.20	14.76
11-BA-46-dup	3.82	15.02
12-BA-32	6.47	12.90
12-BA-33	8.96	14.04
13-Au-04	8.45	13.56
13-Au-08	7.47	13.27
13-Au-09	8.26	13.30
13-Au-23	8.65	14.07
13-Au-28	8.35	14.40
13-AU-29	7.32	13.41

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