GEOLOGIC MAP OF THE DAVIS MOUNTAIN 7.5-MINUTE QUADRANGLE, CAMAS AND GOODING COUNTIES, IDAHO: STRATIGRAPHIC RELATIONS OF MIOCENE RHYOLITES

By:

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To the Graduate Faculty:

The members of this committee appointed to examine the thesis of William L. Oakley III find it satisfactory and recommend that it be accepted.

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A special thanks goes to Dan Weisz. He passed away in fall 2003. I spent a lot of time with Dan drilling basalt and panning gold. He was always willing to discussing girls, gold, and geology (usually in that order). I had my first discussion of ignimbrites with him. He was a good friend and is very much missed. This is for you.

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ABSTRACT

This study focuses on detailed mapping of the Davis Mountain 7.5’ Quadrangle in the central Mount Bennett Hills north of the Snake River Plain, Idaho. The rock units in the study area include Eocene Challis Volcanic Group, late Miocene Idavada Volcanic Group, and post-Idavada age units.

The plagioclase and pyroxene bearing, crystal-vitric rhyolitic ignimbrites of the central Mount Bennett Hills are correlative with the Idavada Group of central and southwest Idaho as suggested by petrographic, geochemical, and geochronological parameters.

The rhyolite of Deer Springs (11.21±0.06 Ma), tuff of Fir Grove (11.17±0.08 Ma), and tuff of Knob (10.6± 0.2 Ma), are correlated with the Bruneau-Jarbidge volcanic field. These are separated by an angular unconformity from four overlying rhyolite units which are part of the Twin Falls volcanic field.

Normal faulting related to the formation of the western Snake River Plain began before 10.1 Ma with movement on the northeast-striking, down-to-the-south Pole Creek fault which cuts the Bruneau-Jarbidge units.

Two new stratigraphic units were recognized: basalt of Davis Mountain and Pole Creek vitrophyre. The olivine tholeiite basalt of Davis Mountain (10.1-10.6 Ma) caps the Bruneau-Jarbidge volcanic field age sequence and was faulted and tilted before the eruption of the Pole Creek vitrophyre, a 10.1-10.6 Ma vitric tuff that is the lowest and oldest unit of the Twin Falls age sequence in the Davis Mountain area. The overlying Twin Falls volcanic field units include the vitrophyre of Pole Creek, tuff of Thorn Creek,

tuff of Gwin Springs, McHan basalt (9.4 Ma), diatomite of Clover Creek and tuff of City of Rocks (9.2 Ma).

Two sets of northeast and northwest striking down-to-the-south faults deform the units of the Twin Falls volcanic field. At 9.4 Ma, during the eruptions of the Twin Falls volcanic field a half graben basin opened or a canyon was dammed. In this basin over 100 meters of the Clover Creek tuffaceous diatomite was deposited, in as few as 100 k.y. The diatomite contains a large resource which if mined would flood the world diatomite market, drive the price down, and be unprofitable under current economic conditions.

By 9.2 Ma the diatomite of Clover Creek was dissected by south-flowing streams, cutting down to a lowering base level. This base-level drop at 9.2 Ma created north-south trending transverse canyons in the south tilting volcanic construct of the Mount Bennett Hills. These canyons originally had headwaters in the highlands of the Soldier Mountains 15km to the north. They were beheaded in Late Miocene time (post-9.1 Ma), by east-striking, down-to-the-north faulting on the northern margin of the Mount Bennett Hills related to the rifting of the Camas Prairie.

Tilting progressed from about 10Ma to at least 3 Ma, as all units dip southward(at 13 to 2 degrees), though with decreasing angle in younger units. The sequence of deformation after 10 Ma, and formation of the Camas Prairie, is consistent with progressive detumescence of a topographically elevated Twin Falls volcanic field and north-south extension related shorting of the down warping central Snake River Plain to the south.

The south-tilted horst that now forms the MBH is a potential region of aquifer recharge for the central Snake River Plain. Major springs in the MBH are sourced from the intercalated sediment at the contact zones of the volcanic units suggesting that deep aquifers are present in the Idavada Group.

CHAPTER 1:
INTRODUCTION

Location

The Davis Mountain 7.5’ Quadrangle (DMQ) is located approximately 66km east of Mountain Home, Idaho, and nearly 80km west-northwest of Twin Falls, Idaho (43°7.5’ to 43°15’ north latitude, 114°52.5’ to 115° west longitude). The DMQ straddles an east-west trending 100km-long mountain range called the Mount Bennett Hills (MBH). The MBH is a horst-like tilted fault block that is bounded by the down-warped Snake River Plain to the south and the down-faulted Camas Prairie to the north. Figure 1.1 shows the relationship of the DMQ to the regional physiographic and geologic provinces.

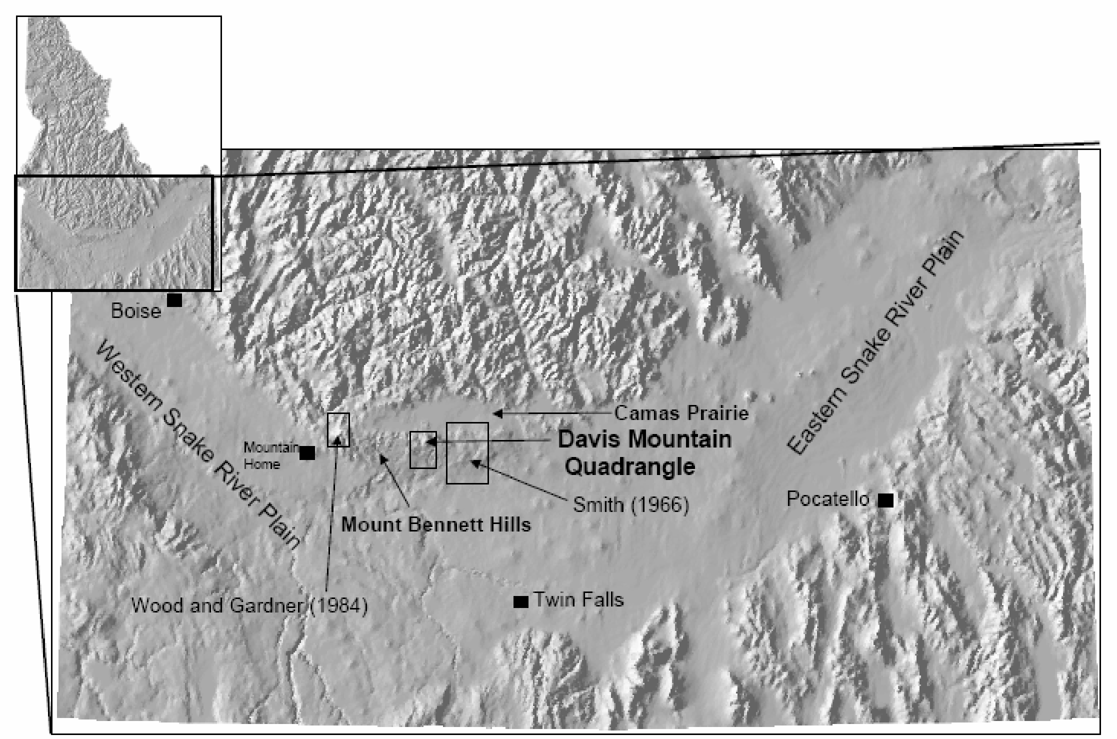

Figure 1.1. Regional shaded relief map denoting the geologic provinces and illustrating the pertinent physiographic features of southern Idaho.

General Geologic Setting

The Mount Bennett Hills form an east-west-trending, south-dipping horst consisting of Miocene Idavada Group rhyolite with intercalated basalt and sediments that are underlain by the Eocene Challis Volcanic Group and Tertiary-Cretaceous granitic rocks of the Idaho batholith (Malde and others, 1963; Worl and others, 1995), and capped by Miocene and Pliocene Snake River Plain olivine tholeiite basalt. All of these lithologic groups crop out in the DMQ except for the granitic rocks. The granitic rocks crop out in both quadrangles to the east and west, in the Fir Grove Mountain and Dempsey Meadows Quadrangles respectively (Malde and others, 1963; Smith, 1966).

Statement of the Problem

The Idavada Group is a term used to identify Miocene rocks on the north and south sides of the central and western Snake River Plain. Units of the central Snake River Plain have been separated based on the age and location and assigned to various volcanic fields. Internal stratigraphic relations of the Idavada record the development and deformational history of the SRP, consistent with a stationary hotspot over which the North American tectonic plate has moved southwestward. The initiation of rhyolitic volcanism generally decreases in age northeastward. Figure 2.1 shows the spatial relation and approximate ages of the volcanic fields on the SRP. In general, these include rhyolitic and basaltic volcanic products of the Bruneau-Jarbidge and Twin Falls volcanic fields between about 13 to 7 Ma (Pierce and Morgan, 1992).

The Bruneau-Jarbidge (12.5-11.3 Ma), and Twin Falls (10-8 Ma) volcanic fields are potential source regions for the Idavada Group rocks of the Mount Bennett Hills

(Bonnichsen, 1982; Honjo 1990; McCurry et al, 1996). I use these terms only in an age and stratigraphic sense, and make no judgment about the number and type of volcanic vents.

The Bruneau-Jarbidge volcanic center has been delineated by Bonnichsen (1982) as an arcuate structure in the Owyhee Plateau. Volcanic textures suggesting flow direction in ignimbrites point away from the center which allows its definition.

The nature and location of the Twin Falls volcanic field is poorly understood because of burial by post-Idavada basalt; no such structure is visible at the surface for the Twin Falls volcanic field. Structural and textural data from the Cassia Mountains south of the city of Twin Falls suggest a northerly source area (McCurry et al, 1996), thus suggesting a source near the city of Twin Falls.

Similar lithologic characteristics of adjacent units, origin, mode of emplacement (i.e. lava versus pyroclastic flow), and complex inter-fingering of units have complicated stratigraphic and spatial correlations of Snake River Plain rhyolites (Smith, 1966; Bonnichsen and Citron, 1982; Wood and Gardener, 1984, Honjo, 1990; Williams et al, 1991: Clemens, 1996; etc.).

Wood and Gardner (1984) delineated 10 rhyolite units at Bennett Mountain, northwest of Mountain Home, approximately 40km to the west of the DMQ. Smith (1966) observed 5 separate Idavada Group units 4.5km east of the DMQ. Figure 1.2 compares the stratigraphy presented by Wood and Gardner (1984), Smith (1966), Honjo (1990), and this study.

Also several of the Idavada Group units have been observed on the north side of the Camas Prairie, which could provide a maximum age of the down-faulted Camas

Prairie Rift (Cluer and Cluer, 1986). Detailed mapping in the MBH allows better definition of which units are present on the north side of the Camas Prairie.

Purpose and Significance

The purpose of this study is to: 1) create a geologic map of the Davis Mountain 7.5’ Quadrangle, 2) build upon the existing stratigraphic framework set forth by previous workers by adding only a minimal amount of new unit names and extending the known units; the goal is a simple and regionally useful set of map units for compilation at 1:100,000 by the Idaho Geological Survey, 3) evaluate the stratigraphic conclusions presented by Smith (1966); early reconnaissance work indicated problems associated with Smith’s stratigraphic order of the rhyolites, 4) provide details about the structural evolution of the central MBH, southern Camas Prairie and the central SRP, and 5) suggest potential economic opportunities in the MBH; the area has abundant springs and diatomaceous earth deposits.

A unified stratigraphic nomenclature for the MBH would be useful when attempting to correlate units that crop out to the east in the area of Picabo and Lake Hills as well as across to the south side of the SRP in the Cassia Mountains, Browns Bench or the Owyhee Plateau area.

The Snake River Plain is divided into two geologic provinces: the western and eastern SRP. The Mount Bennett Hills are located where these two provinces come together. The western Snake River Plain is a northwest-trending basin defined by high angle normal faults and filled by a nearly 2km-thick sequence of sediments and basalts.

The eastern SRP in a northeast-trending basin filled with bi-modal volcanic products: rhyolite domes covered by olivine tholeiite basaltic lava flows (Malde, 1991).

Scope and Procedures

This study includes: 1) a 1:24,000-scale map of the DMQ and associated cross-sections, 2) an interpretation of the structural development of the MBH, 3) correlation with the stratigraphy of Smith (1966), 4) discussion of the petrography and geochemistry of the Idavada Group, 5) discussion of the volcanological aspects of the rhyolitic units, and 6) location and distribution of ground water sources and economic deposits. The geologic map and cross-sections display the extent and orientation of the units measured and observed in the field and on color aerial photographs.

Geochemical data was provided by ALS Chemex in Reno, Nevada, and Washington State University's GeoAnalytical Laboratory. Thin sections were prepared by Spectrum Petrographics and University of Utah College of Mines and Earth Sciences Sample Preparation and Thin Section Laboratory. Geochronological analyses were provided by

Funding for this project was provided by U.S. Geological Survey EdMap grant 05-HQAG0022 to Paul Link, the Idaho Geological Survey, and Round Mountain Gold Corporation.

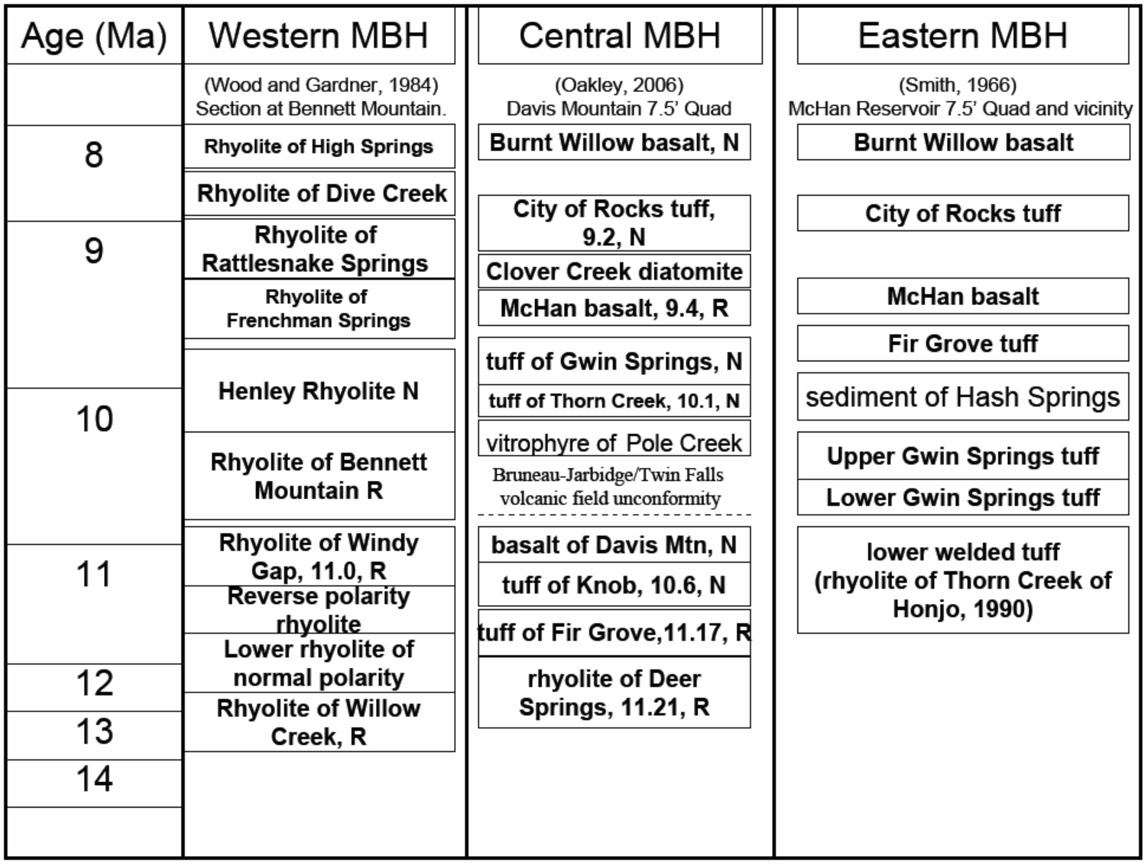

Figure 1.2 Three stratigraphic columns showing the nomenclature for the Mount Bennett Hills: Smith, 1966; Wood and Gardner, 1984; and this study. K-Ar ages from Armstrong et al., 1980; and Honjo et al., 1986. Magnetic polarity data is from Wood and Gardner, 1984, and this study.

The geology of the DMQ elucidates the deformational history of the MBH in the context of structures and volcanic deposits. The results of this study build on the work performed by others trying to understand the SRP rhyolites. The MBH are in a unique location where the down-warped eastern SRP meets the down-faulted western SRP (Mabey, 1982). Therefore, the DMQ is in a suitable location to study the relationship between the two types of SRP deformation. My intention is to develop a succinct, unified stratigraphic nomenclature for use in future research.

CHAPTER 2: REGIONAL GEOLOGY

The Davis Mountain quadrangle is located in the Mount Bennett Hills on the north side of the central Snake River Plain. The Mount Bennett Hills are an east-west trending, 100km-long mountain range with elevations around 1,800-2,100 meters above sea level. They are bordered by several geologic provinces. To the north of the MBH is the Camas Prairie Rift (Cluer and Cluer, 1986) and farther north are the Soldier Mountains. The Soldier Mountains consist of Tertiary and Cretaceous intrusive rocks of the Idaho batholith Atlanta lobe overlain and intruded by Eocene volcanic and hypabyssal rocks of the Challis Volcanic Group (Worl et al, 1991; Lewis, 2001).

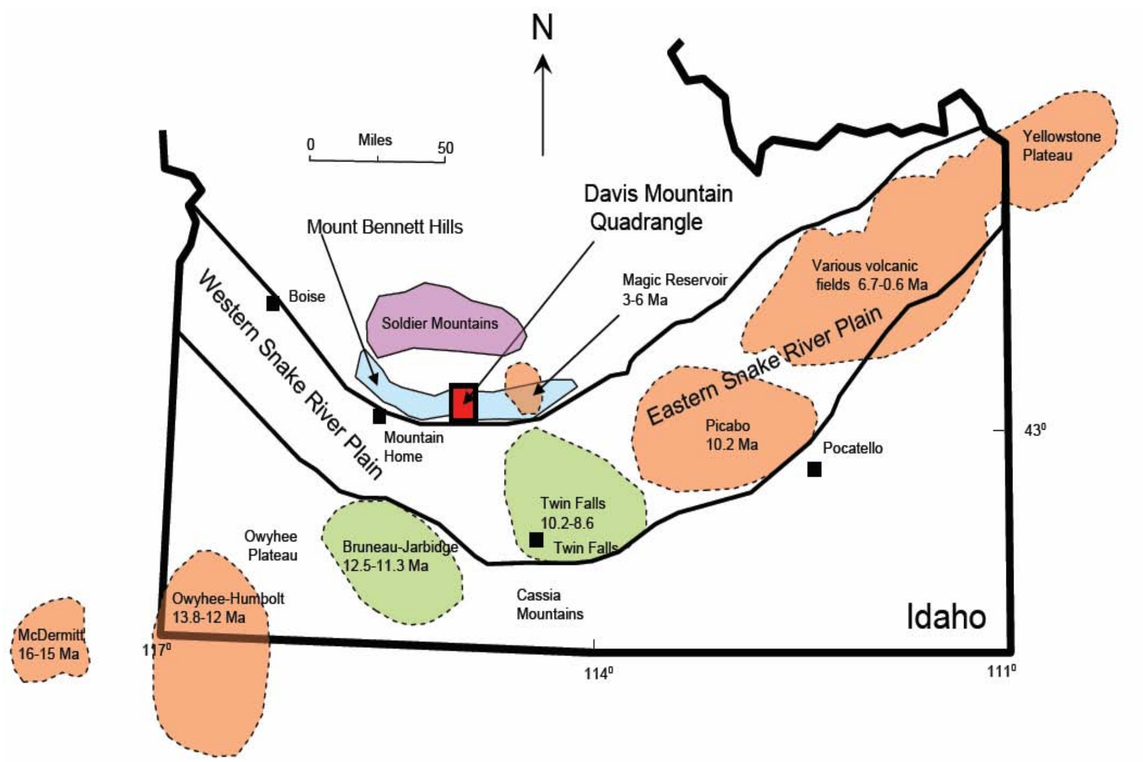

Figure 2.1. An outline of southern Idaho showing the locations of Yellowstone–Snake River Plain hotspot volcanic centers. Dashed outlines of volcanic fields and dates denoting time of activity generally decrease in age from southwest to northeast as suggested by Pierce and Morgan, 1992. The red square is the location of the Davis Mountain Quadrangle. Modified from Hughes et al., 1996 (Figure 1).

The Idavada Group rhyolites are eruptive products of the Yellowstone-Snake River Plain hotspot track as defined by Pierce and Morgan (1992) more specifically the Twin Falls and Bruneau-Jarbidge volcanic fields. Figure 2.1 shows the linear distribution and ages associated with the various volcanic fields of the Y-SRP hotspot track.

The Camas Prairie Rift separates the Soldier Mountains from the Mount Bennett Hills. The Camas Prairie is a Miocene to Quaternary rift basin bounded on the north and south by normal faults (Cluer and Cluer, 1986). Cluer and Cluer present an interpretation of the rift suggesting that shortening in the Snake River Plain due to subsidence is facilitated to the north by extension. The result is a rift with south-dipping normal faults and antithetic north-dipping faults. The antithetic faults form the south margin of the Camas Prairie Rift and the Mount Bennett Hills.

The rift basin is a graben filled with Miocene to Quaternary age sediments derived from the granitic intrusions and volcanic rocks of the Soldier Mountains. Intercalated in the sediment are basalt flows of various ages as young as Quaternary (Honjo et al., 1986). Several erosional remnants or faulted blocks of Challis Volcanic Group rocks and rocks of the Idaho Batholith protrude above the nearly flat-lying sediments and basalts that define the Camas Prairie.

The Mount Bennett Hills form the southern margin of the Camas Prairie. The MBH are composed of Challis Volcanic Group rocks that are underlain by the Idaho batholith and overlain by rhyolites and basalts of the Idavada Group (Malde et al. 1963). The Idavada units are Late Miocene in age, ranging from ~11.2-9.2 Ma (K-Ar ages from Armstrong et al, 1980 and Honjo et al, 1986; and Ar-Ar ages from this study).

Down-to-the-north normal faults of the Camas Prairie define the north side of the MBH. From the crest of the mountain range to the south the Idavada basalts and rhyolites form a south-dipping ramp. Attitudes are commonly 6° and range from 2-13° (Smith, 1966; Wood and Gardner, 1984, and this study).

At the eastern-most end of the Mount Bennett Hills in the vicinity of Magic Reservoir is a small volcanic field off the axis of the SRP. This volcanic field is younger than the Idavada Group of the MBH. The field includes quartz latite and hybrid ferrolatite lava flows 6-3 Ma (Honjo et al., 1986; Honjo and Leeman, 1987). These rocks define the Magic Reservoir volcanic field (Leeman, 1982). None of the Magic Reservoir volcanic products are found in the Davis Mountain Quadrangle. Smith (1966) mapped several of the Magic Reservoir units. The Square Mountain basalt of Smith (1966) is found 20km east of the DMQ. This is the closest of the Magic Reservoir volcanic units.

The modern Snake River Plain is a low arcuate depression covered by Neogene and Quaternary sediments and volcanic products. The SRP is 600km long and up to 70km wide and spans from Yellowstone National Park in the east to the eastern portion of Oregon in the west.

Recently concluded work (Beranek, 2005) regarding the drainage history of the SRP using detrital zircons has shown that the Y-SRP hotspot "existed as a continental divide as it migrated across southern Idaho." The study sampled modern and ancient stream deposits.

To the south of the Snake River Plain are several mountain belts and highlands that are composed of similar rocks to the MBH. These areas include the Cassia Mountain (or South Hills), Brown's Bench, and the Owyhee Plateau. Work is being done (Bill

Bonnichsen, Martha Godchaux of the Idaho Geological Survey, and Lisa Morgan of the U.S. Geological Survey; etc) to elucidate the geology of these regions and potentially correlate the rocks on the south side of the SRP to the rocks in the MBH.

Detrital Zircons of Clover Creek

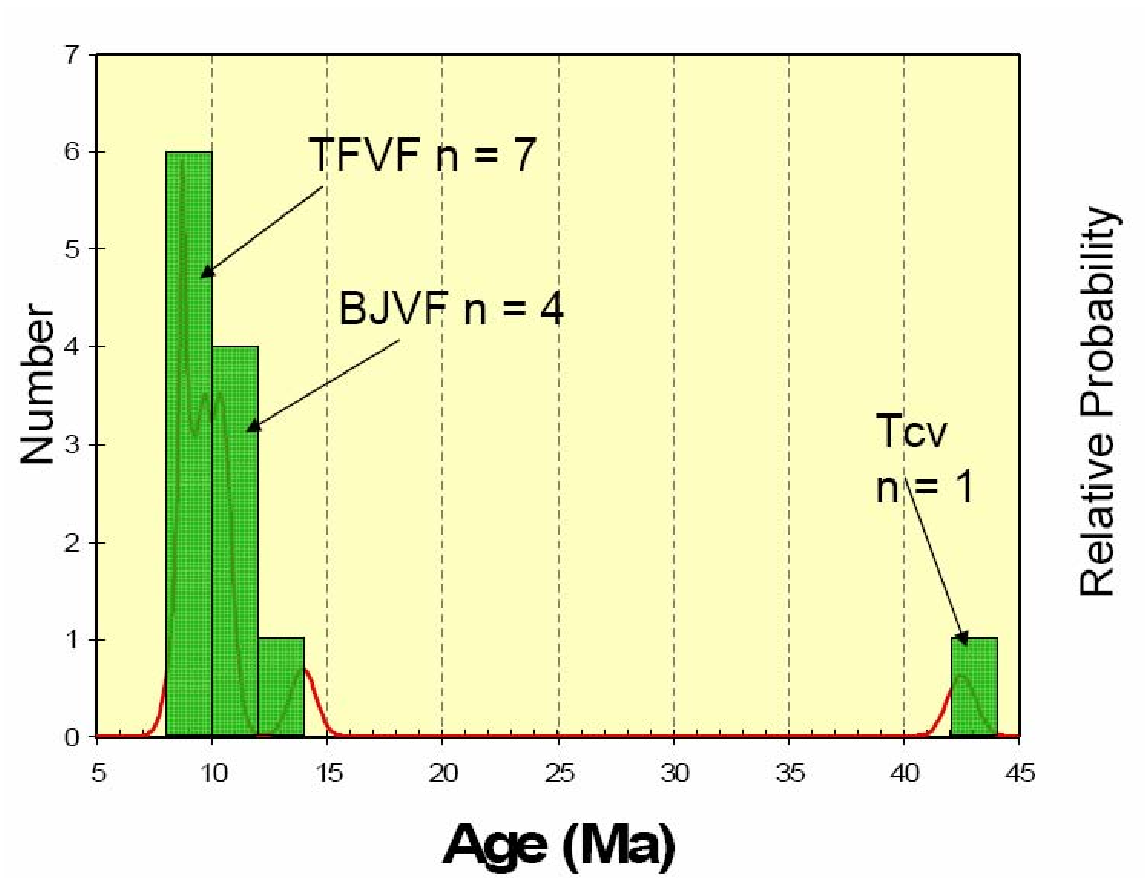

Figure 2.2. Histogram plot of the detrital zircon age spectrum for Clover Creek. 12 detrital zircon grains were analyzed. Each bar spans 2 Ma. TFVF is the Twin Falls volcanic field; BJVF is the Bruneau-Jarbidge volcanic field. Tcv represents age correlative with the Challis Volcanic Group. The detrital zircons suggest Idavada and Challis-aged (8-14 Ma and 40-45 Ma) rocks are drained by Clover Creek. Data can be found in Appendix IV.

A detrital zircon sample was taken by H.G. McDonald and analyzed by Link in 2002. The sample is from just south of the Davis Mountain Quadrangle. 12 grains were analyzed yielding ages that overlap with the Bruneau-Jarbidge and Twin Falls volcanic fields. Figure 2.2 shows the age spectrum of detrital zircons for the modern Clover Creek.

Ages of detrital zircons were determined by P.K. Link. The sample was taken from the modern sediment of Clover Creek which drains the majority of the DMQ. The spectrum of ages of the zircon grains gives an indication of the age of the units in the DMQ. These data are presented here for the first time. Twelve zircon grains were probed for U-Pb geochronology using the SHRIMP (Sensitive High Resolution Ion Microprobe) at Australian National University, using standard techniques for detrital zircon samples (Williams and Claesson, 1987; Compston et al., 1992; Williams, 1998; Link et al., 2005). Each analysis consists of 4 scans through the mass range. The data were reduced in a manner similar to that described by Williams (1998, and references therein), using the SQUID Excel Macro of Ludwig (2003).

Age (Ma)	±	Potential host unit of the DMQ
8.6	0.6	tuff of City of Rocks
8.7	0.1	tuff of City of Rocks
8.8	0.4	tuff of City of Rocks
9.1	0.5	tuff of City of Rocks
9.2	0.4	tuff of City of Rocks
9.7	0.2	tuff of Gwin Springs
10.2	0.3	tuff of Thorn Creek
10.3	0.7	tuff of Knob
10.5	0.3	tuff of Knob
10.6	0.4	tuff of Knob
13.9	0.6	unknown unit
42.5	0.6	Challis Volcanics

Table 2.1 This table lists the ages obtained from detrital zircons of Clover Creek just south of the DMQ. The grains are listed chronologically with associated uncertainties. These grains are of identical age to K-Ar and Ar-Ar estimates of Idavada ages (Armstrong et al., 1980, and Honjo et al., 1986).

This technique provides a fast reconnaissance geochronological sample of ages of zircons in the Clover Creek drainage, and as I will show subsequently in this thesis, the ages determined are totally compatible with the correlations of map units that I made based on stratigraphic and volcanological techniques.

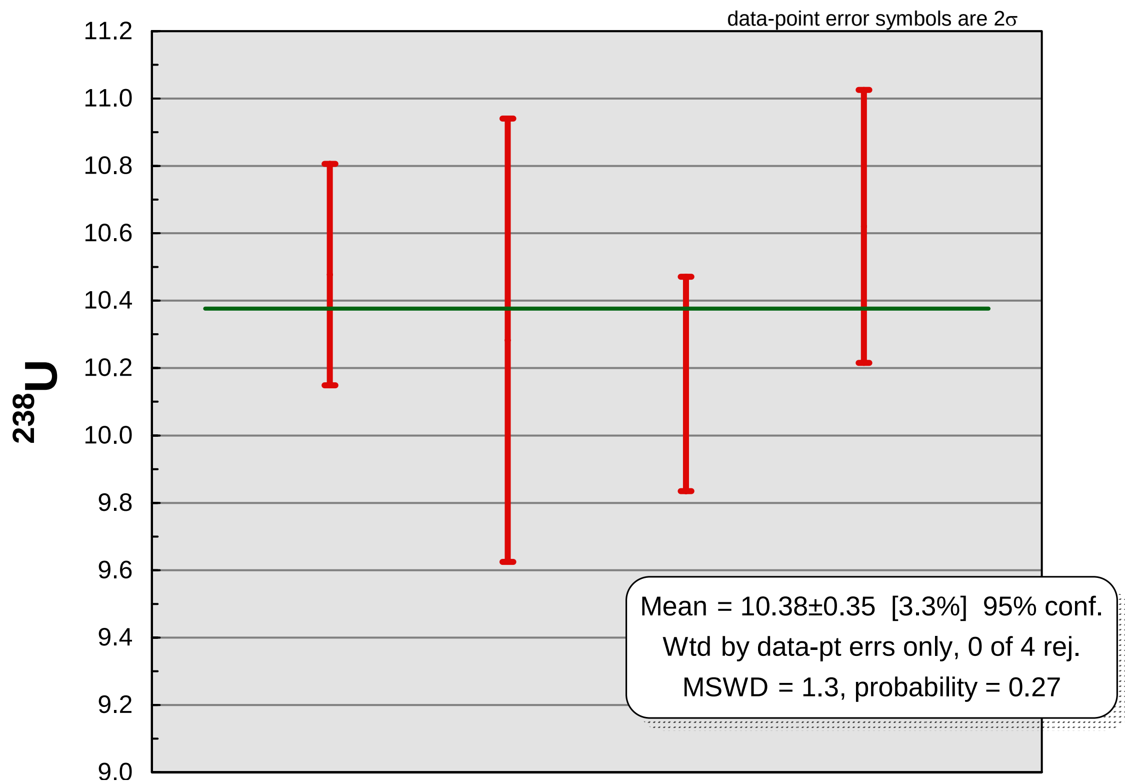

Figure 2.3 Graph showing ages of detrital zircon grains about 10 Ma. Lines through error bars represents the mean average age of these 4 grains. This age is statistically equivalent to the K-Ar age (10.1±0.3 Ma) by Armstrong et al., 1980 for the tuff of Thorn Creek.

Figure 2.3 and 2.4 show statistically indistinguishable U-Pb ages of detrital zircon grains. Figure 2.3 shows analyses of four grains that averaged 10.38±0.35 Ma. This age is roughly equivalent to the K-Ar age of the tuff of Thorn Creek (10.1±0.3 Ma; Armstrong et al., 1980).

Figure 2.4 shows three grains aged 8.72±0.11 Ma. This age is a younger than the K-Ar age for the youngest unit of the DMQ. Honjo et al., (1986), report 9.15±0.15 Ma for the tuff of City of Rocks which is the youngest rhyolitic unit of the DMQ. More data would constrain this age.

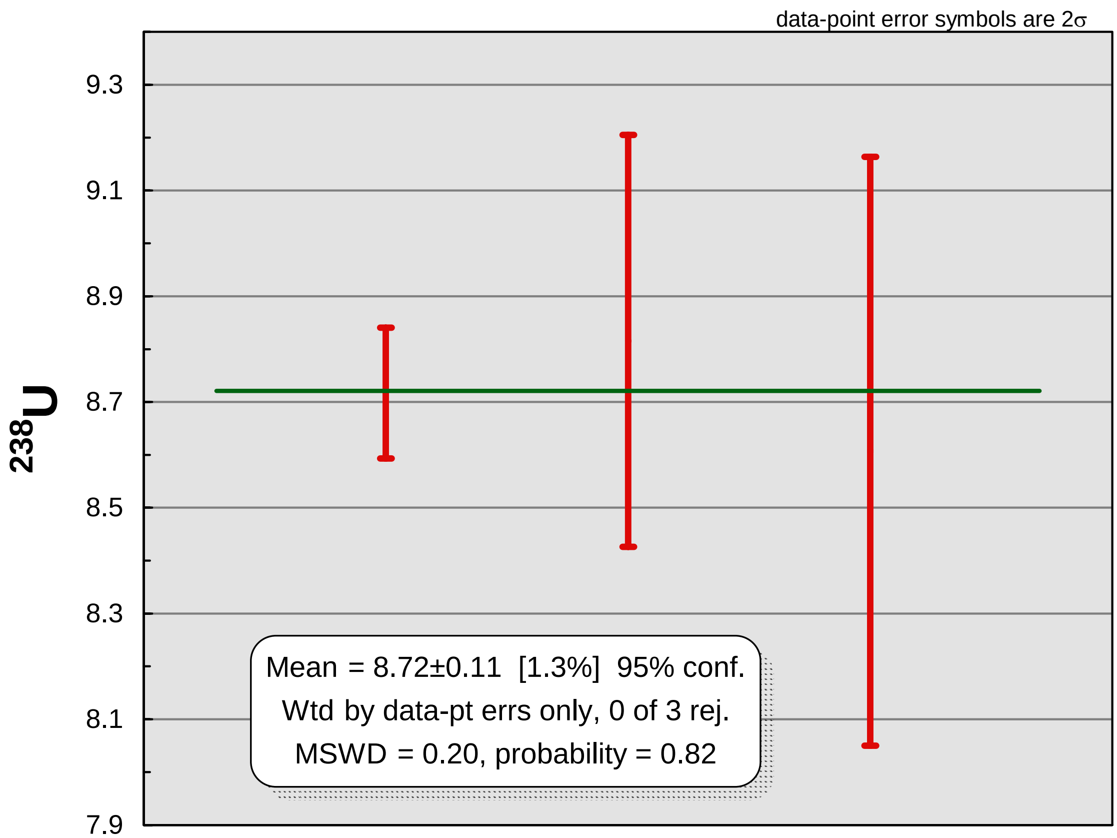

Figure 2.4. Graph showing ages of three detrital zircon grains. (8.72±0.11) is slightly younger than the K-Ar age presented by Honjo (1990) 9.15±0.13 Ma for the tuff of City of Rocks which is the youngest tuff in the DMQ .

CHAPTER 3:
METHODS

Mapping Procedures and Criteria

The idea to create a map of the DMQ was suggested by Dr. Reed Lewis, Associate State Geologist, Idaho Geological Survey. He recommended the DMQ because of its good outcrop, proportion of public land, and value for the Idaho Geological Survey project to map the Fairfield 1:100,000 Quadrangle.

The rocks were observed in outcrop and in hand sample with a 10X hand lens. Measurements of bedding, foliations, and lineations were done with a Brunton compass with inclinometer. Foliations within the body of the rhyolitic units were measured at the site of observation, whereas most unit orientations were measured from a distance and are only estimates of the actual orientation. This was done to reduce the variations that could be obtained from measuring foliation in ignimbrites (created by emplacement turbulence and non-tectonic folding, i.e. pyroclastic or lava flow textures), and which may mask the true orientation of the bedding. The foliation of an outcrop may in fact be dipping counter to the overall dip direction of the unit (Branney and Kokelaar, 1992).

Nearly 50 samples were taken for geochemical and petrographic analysis. The units sampled include the Idavada Volcanic Group, Challis Volcanic Group and overlying basaltic units. The Challis and Idavada groups were distinguished by mineralogy, outcrop pattern, and color in the field and in aerial photographs.

The Idavada Group units can be traced laterally based on differences in crystal content between the vitrophyres. The crystal-poor tuff of Fir Grove consists of 2-5% phenocrysts where as tuff of Knob contains over 20%. These differences are best viewed in hand sample. The conspicuous difference in total crystal content made distinguishing units quite simple given the generally homogenous crystal content and rather sparse large-scale faulting.

Geochemical Sampling

Whole-rock geochemical sampling techniques and analyses used in this study are focused on correlating and distinguishing the units. Major elements were analyzed using X-ray Fluorescence (XRF) at the Washington State University GeoAnalytical Laboratory, Pullman, Washington, and at ALS Chemex in Reno, Nevada. This method has been used by previous workers including Honjo (1990), Leeman (unpublished data), and Bonnichsen (unpublished data). Also, the Idaho Geological Survey uses this method.

Most trace element analyses were determined by Inductively Coupled Plasma-Mass Spectrometry (ICP-MS) at ALS Chemex in Reno. I chose this method for its ability to analyze many elements using a small sample size, and the improved detection limits over ICP-AES (Rollinson, 1993).

Some of the analytical work was done by the Washington State University GeoAnalytical Laboratory in Pullman, Washington. This lab processed six samples using the XRF method. All major elements and 18 trace elements were reported. These analyses were funded by the IGS.

ALS Chemex in Reno, Nevada, analyzed 40 samples with XRF and ICP-MS: XRF for major elements and ICP-MS for 38 trace elements. One sample was assayed for gold and silver using atomic absorption and ICP-MS for 26 other pathfinder elements

(elements used for locating and analyzing ore bodies) and this was funded by Round Mountain Gold Corporation.

Some controversy arose regarding which type of rock (glassy of crystalline) to sample for the rhyolites. The glass of the vitrophyre will return results most like the original magma chemistry. However the glass has to be unaltered. Hydration of glass alters the original glass composition, particularly alkali contents (Lipman, 1965; Scott, 1971). Hydration of rhyolitic glasses is dominated by the reaction $Na^+ \leftrightarrow H_2O^+$. This type of alteration increases over time and is identified in volcanic glasses by low major element totals, low Na_2O, and high LOI values (>0.5 % to 6 %).

Devitrification and crystallization processes can also alter the values of elements (Lipman, 1965; Scott, 1971). Therefore analyses from the devitrified zones may not return as accurate results compared to non-hydrated glass.

The glassy rocks of the DMQ are altered to varying degrees by hydration. As well, the chemical signatures of the devitrified rocks are altered by crystallization. Unaltered glass chemistries, which are difficult to find in the DMQ, are used as a proxy for original magma chemistry at time of eruption; thus the problem of what to sample. As a result, I sampled both the glassy and the devitrified zones of many of the units.

Thin Section Sampling and Analysis

Nineteen thin sections were cut from ten rhyolites, seven basalts, and two Challis Volcanic Group units. I made at least one thin section to represent each of the major volcanic units. Standard size and thickness sections were cut by Spectrum Petrographics of Vancouver, Washington, and the University of Utah College of Mines and Earth

Sciences Sample Preparation and Thin Section Laboratory at Salt Lake City, Utah. Felsic units were etched with hydrofluoric acid and stained with sodium cobaltinitrate to identify sanidine.

The slides were analyzed with an Olympus BH-2 binocular polarizing-light microscope. The rhyolites were examined primarily to document the mineral assemblages, textures, abundances of phenocrysts and degree of sample alteration. The Michel-Levy method was used for determining the anorthite content of plagioclase phenocrysts. A description of this method can be found in Nesse's Introduction to Mineralogy (2000). Most of the plagioclase phenocrysts of the rhyolites are unzoned, which is necessary to use this method.

Paleomagnetic Distinctions and Procedures

Paleomagnetic polarities were determined with a MEDA µMAG field fluxgate magnetometer. The instrument consists of a console with a digital readout screen, several knobs and selectors for zeroing the unit and adjusting for exceptionally strong or weak magnetic samples. From the console extends a probe on a six-foot-long cable that is to be positioned in an east-west direction to nullify the current magnetic field as much as possible. The probe measures intensities of magnetic fields in milligauss, so all field magnets and metal objects must remain at rest during the measurement of samples. Movement of such items will create undue 'noise' that the instrument will detect. These unnecessarily induced magnetic fields can alter the performance of the instrument.

A minimum of three samples were taken at each site. The results of the paleomagnetic survey are located in Appendix III and they all matched lithostratigraphic correlations.

Potential errors can occur when sampling in an area that has been struck by lightning or re-magnetized by Earth's current field. Spurious results will be obtained if surveys are conducted in the vicinity of power lines. Electrical power lines were not an issue with the sampling in the Davis Mountain Quadrangle.

Because of potential re-magnetization issues, measurements resulting in reverse polarity have greater credibility than those resulting in normal polarity. Normal magnetic polarity result may have been altered (personal communications from Idaho Geological Survey personnel, Dan Weisz, 1998; Dean Garwood, 2002; Kurt Othberg, 2005).

I have also found that more reliable and 'good' samples come from the base of the outcrop where the samples are insulated from the current magnetic field. Samples taken from the tops of spires or pinnacle-like features seem to be more often re-magnetized or return poor results (i.e., the opposite reading is not produced when the sample is turned around). So I tried to sample in areas that were not highly exposed or pillar-like without material closely surrounding it in the sampled area.

CHAPTER 4:
UNIT DESCRIPTIONS

This chapter will include a description of the all of the map units of the Davis Mountain Quadrangle.

Eocene Units

Challis Volcanic Group (Tcv)/(Tcvs)

The Challis Volcanic Group (Johnson et al., 1995) includes Eocene intermediate-composition lavas and associated pyroclastic and sedimentary rocks. The Challis Volcanic Group in the DMQ is distinguished from the Idavada Volcanic Group based on the presence of hydrous mineral phases including biotite and hornblende. The outcrops are deeply weathered, characterized by crumbly and rubbly outcrop with steep rock faces with associated light colored, sandy soil. In the DMQ, I divided the Challis into two units based on rock type: tuffaceous sedimentary unit (Tcvs) and a felsic volcanic rock unit (Tcv). The flow rock unit includes, andesitic lava flows, dacitic rock, and rhyolitic ignimbrites. The sedimentary unit is exposed in the northwest corner of the map and appears to be younger than the flow unit, resting unconformably on top of dacitic and rhyolitic rocks.

Challis Volcanic Flow Rocks (Tcv)

This unit includes the lithologies introduced above: rhyolitic ignimbrite deposits, andesitic flows, and dacitic rock. The andesitic flows are found in the northeastern corner of the map to the east of Davis Mountain, and are unconformably overlain by the

Miocene Fir Grove Tuff of the Idavada Group. This unit is recognized by fist-sized cinder-like rocks lacking appreciable phenocryst content in a rubbly outcrop. The cinder-like and spatter-like clasts are reddish-black and brown with stretched vesicles and flow textures. This portion of the member is probably a crumble-breccia mantling the flow. The interior of the flow is variably colored due to alteration. Deep weathering and abundant lichen-cover distort or obliterate any textural features visible at hand sample scale. From a distance the unit can be seen as a grayish colored massive unit extending from Perkins Gulch, at the base of Davis Mountain, east to the north face of the east peak of Twin Peaks.

At the base of the Twin Peaks, green and white rhyolitic rocks are exposed. Welding varies and facies of this unit are local and discontinuous. Contorted foliation can be found in the lower rocky slopes north of the pothole. In the loose colluvial material associated with this unit are 3-5cm diameter clasts of chalcedonic quartz veining, evidence of hydrothermal activity.

The steptoe knoll that is surrounded by the basalt of Pothole is of a similar lithology to the interior of the flow at the base of Davis Mountain; perhaps more felsic? More detailed observation may produce a rudimentary stratigraphic framework for the Challis Volcanic Group of the Twin Peaks area.

In the northeastern portion of the DMQ Challis dacitic rocks and the tuffaceous sedimentary unit are exposed. The dacitic units are also deeply weathered and altered obscuring many primary textures. The dacitic rocks contain abundant phenocrysts and lithics of a fine-grained nature. The principal crystals include plagioclase, sanidine, hornblende, biotite, quartz and Fe-Ti oxides. The most abundant phase is plagioclase ranging in shape and size from sub- to euhedral 2-5mm discrete crystals. Biotite and hornblende are the distinguishing phases that set this unit apart from the Idavada Group in the DMQ. The hornblende is generally euhedral black-green grains, 1-3 mm long. The biotite grains appear mostly as anhedral blebs probably due to the orientation in the matrix, biased by the direction of the thin section cut. Some of the biotite forms 1-3mm diameter euhedral books. Subhedral sanidine grains are minor in proportion to plagioclase. Anhedral quartz grains are also a minor phase. Lithics in this unit include fine-grained volcanics with a potassium-rich matrix and few small phenocrysts. The lithic grains and phenocrysts are supported by an altered, variably welded, gray matrix.

These units have a meter-scale layering visible from Stokes Road along the Camas Prairie Centennial Marsh just off the northwest corner of the map.

Challis Volcanic Group Sediment (Tcvs)

Tcvs represents a tuffaceous sediment (once quarried for gravel) resting unconformably on the dacitic units of Tcv. This unit is comprised of a fine white reworked ash with well-rounded clasts of Challis Volcanics and granitic rocks. The unit strikes northeast and dips 25° to the southeast. Small offsets in the bedding are common, and generally filled with chalcedonic material.

The unit ranges from well-lithified to quite friable with clasts easily removed from the bedding. Friable material is much more poorly sorted than the lithified zones. Highly fractured and veined portion of the strongly lithified material was assayed. The 29-element assay can be found in Appendix I (Sample # 121978).

Clasts in the lithified sedimentary rocks vary in size, texture, and lithology. Varieties include: polycrystalline quartz grains, micaceous granitoid fragments, quartzofeldspathic schist/gneiss clasts, tuffaceous volcanics, and intergranular volcanics. Lithic fragments exhibit a rounded to sub-rounded shape.

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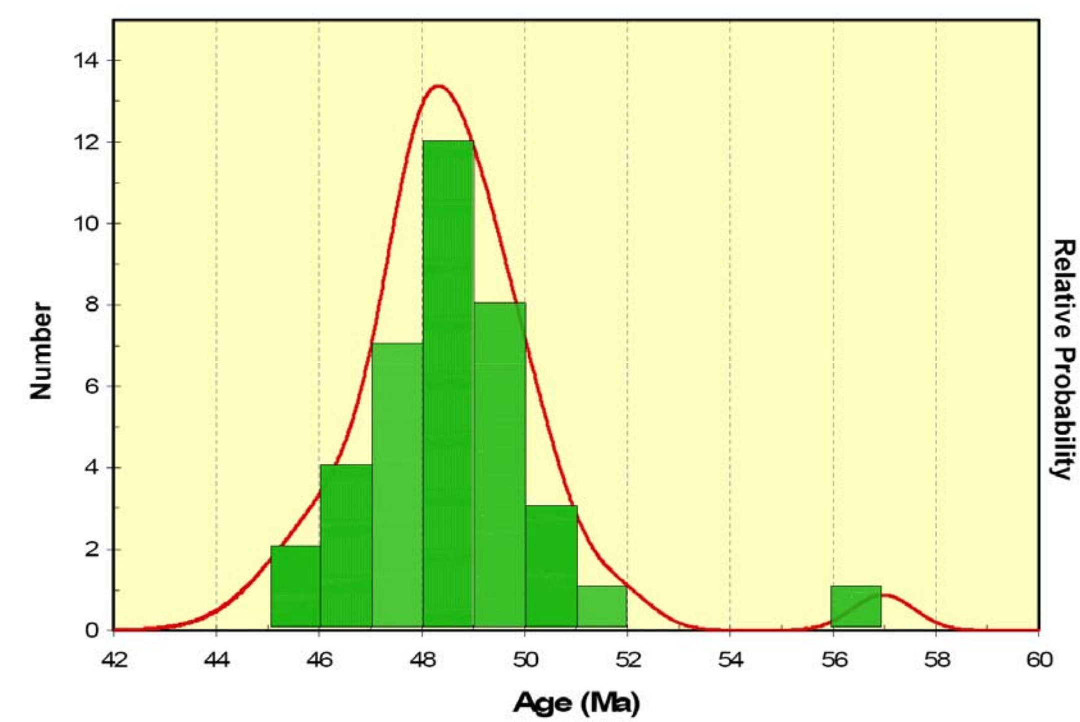

Figure 4.1 Histogram showing the distribution of detrital zircon ages from sample 01PL05. This plot includes 39 of the 40 samples analyzed.

Sample 01PL05 is a detrital zircon sample from this unit. Analysis reveals an average age of 48.1 Ma from 38 grains. The grains are no younger than Eocene which suggests there was no Oligocene volcanic activity in the area from which this sediment was derived or that the sediment is Eocene in age. The sediment contains clasts of granitic material, presumed to be from the Cretaceous Idaho Batholith, yet there are no Cretaceous age zircons in the sediment. This suggests that the granitic clasts are from

Tertiary intrusions of the Challis Volcanic Group and not the Idaho Batholith. Figure 4.1 is a histogram displaying the results of the detrital zircon analysis of the unit.

Miocene Units

These units are constrained by radiometric dates that suggest late Miocene-age volcanism and sedimentation. The name Idavada was used by Malde and Powers (1962) to describe non-mineralized silicic rocks near Salmon Falls Creek at what is called Brown's Bench. Axelrod (1964) gave Idavada group status, including Beaverdam, Jenny Creek, and Cougar Point Formations. Malde et al., 1963 included the silicic rocks of the MBH in the Idavada Volcanic Group. Smith (1966) assigned the name Idavada Volcanic Group in which he includes 10 mappable formations. This study attempts to stay consistent with the terminology set forth by Smith.

The Idavada and Challis Groups are distinguished by the minerals present. Biotite and hornblende are abundant mineral phases in the Challis Group and are extremely rare or non-existent in the Idavada Group. Separation of the groups can be seen in air photos; the Challis Group is deeply eroded and has very irregular outcrops lacking the distinct, prominent columns of the Idavada Group. Also the Challis Group appears lighter in color, buff to white, on the air photos and has distinct orthogonal joints.

Rhyolite of Deer Springs (Tds)

This is a reverse magnetic polarity, crystal-rich (20%), sanidine bearing, vitric rhyolitic ignimbrite with plagioclase, quartz, and pyroxene. Trace amounts of fayalitic olivine, zircon, apatite, and ilmenite are present. The unit is exposed in Deer Creek in the vicinity of Deer Springs and in the east fork of Deer Creek. The base of the unit is not exposed but the exposed thickness reaches 25 meters. Typical exposures in Deer Creek are 15 meters thick.

The informal unit rhyolite of Deer Springs unconformably overlies the Challis Volcanic Group and pinches out at the head of Deer Creek canyon against the Challis Volcanic Group.

This unit was first recognized by Honjo (1990). His study suggests a tentative correlation of the rhyolite of Deer Springs with the 11.0 Ma rhyolite of Windy Gap, dated using the K-Ar method by Armstrong et al., (1980). This is the oldest Idavada unit in the DMQ and is the oldest of three inferred Bruneau-Jarbidge age units. Ar-Ar analysis of this unit reveals an age of 11.21±0.08 Ma.

Most of the exposed unit is a black and red upper vitrophyre with abundant 3-5cm spherulites. The vitrophyres weather to black sand. Just below the upper vitrophyre is a layered sequence that is highly contorted and occasionally lineated. This unit does stand in prominent cliffs but they are limited due to sparse exposures. Where devitrified, the unit stands in pinnacle or hoodoo-like columns, typical of many MHB rhyolite units.

The unit consists of 20% phenocrysts. The phenocryst assemblage is dominantly sub-euhedral, 2mm long, embayed sanidine, 30% of total phenocrysts. Plagioclase is 1.5mm, euhedral and comprises 20% of the phenocrysts with anorthite content ~An_{30}. Broken and embayed quartz phenocrysts are <1mm in diameter accounting for 10% of the crystals. The proportion of pyroxene is 10% (as equal amounts of augite and pigeonite). The pyroxene phenocrysts are less than 1 mm in diameter. Magnetite grains are generally subhedral grains and represent 5% of the phenocrysts. Trace mineral phases (<1%) include: fayalitic olivine, zircon, apatite, and ilmenite.

The rhyolite of Deer Springs is distinguished geochemically as a peralkaline, high Rb (210 ppm), high Zn (95 ppm), high silica (75 wt.%) rhyolite.

Tuff of Fir Grove (Tfg)

This is a reverse magnetic polarity, crystal-poor (2%), plagioclase-bearing, vitric rhyolitic ignimbrite with quartz, augite, sanidine, pigeonite, magnetite and lithics. The tuff of Fir Grove is exposed north of 43°11.5’ N, along Deer Creek and East Clover Creek, and along the north side of Davis Mountain. The ignimbrite reaches 130 meters thick on the westside of Deer Creek at the head. Typical thickness for the tuff of Fir Grove in the DMQ is 100 meters.

This ignimbrite unconformably overlies the rhyolite of Deer Springs and the units of the Challis Volcanic Group. In the northern part of the map area the tuff of Fir Grove is in unconformable contact with the Challis Volcanic Group.

The unit was first described by Smith (1966) who suggested two cooling units. No evidence for two cooling units has been recognized in the DMQ. Smith’s type section is located at Fir Grove Mountain about 10 km to east in the Fir Grove Mountain Quadrangle. This unit is the second of the inferred Bruneau-Jarbidge units.

The typical outcrops of this unit have well-defined upper and lower vitrophyres and lower platy zone grading up into a contorted devitrified zone between the vitrophyres. Immediately beneath the upper vitrophyre is a zone of large rough-walled vesicles 3-4cm in diameter (lithophysal cavities). These cavities exhibit vapor-phase crystallization and alteration halos up to 5mm. The vapor-phase crystals are light-tan with dendritic manganese oxides on exposed surfaces in a rough-walled open space with a radial texture. On fresh surfaces the cavities are off-white and slightly pink against a matrix of shardy, devitrified pink material with rims or halos of brownish matrix.

The lithophysal cavity zone of this unit stands in prominent cliffs. These cliffs are distinct, usually 5-10 meters tall with conspicuous holes. The platy and contorted zones below lithophysal cavities occasionally stand in columns. The upper and lower vitrophyres are slope forming zones.

This unit is low in phenocrysts content (2%). The most abundant phase is broken subhedral plagioclase 1.5mm. The plagioclase has an anorthite content of An_{35}. Granophyric textures can be seen with plagioclase and sanidine. The unit also contains euhedral, embayed quartz grains <1mm. Augite and pigeonite are in near equal proportions. Opaque minerals include magnetite and lesser ilmenite. The unit also contains accessory zircon and apatite. Lithic grains in this unit are minor, 5% of the rock. They include fine-grained felsic volcanic lithologies. The source of these grains is unknown.

The tuff of Fir Grove is distinguished geochemically as low Ba (590-830 ppm), low Sr (<50 ppm), high silica (75-76 wt %), high Rb (205-240 ppm) rhyolite.

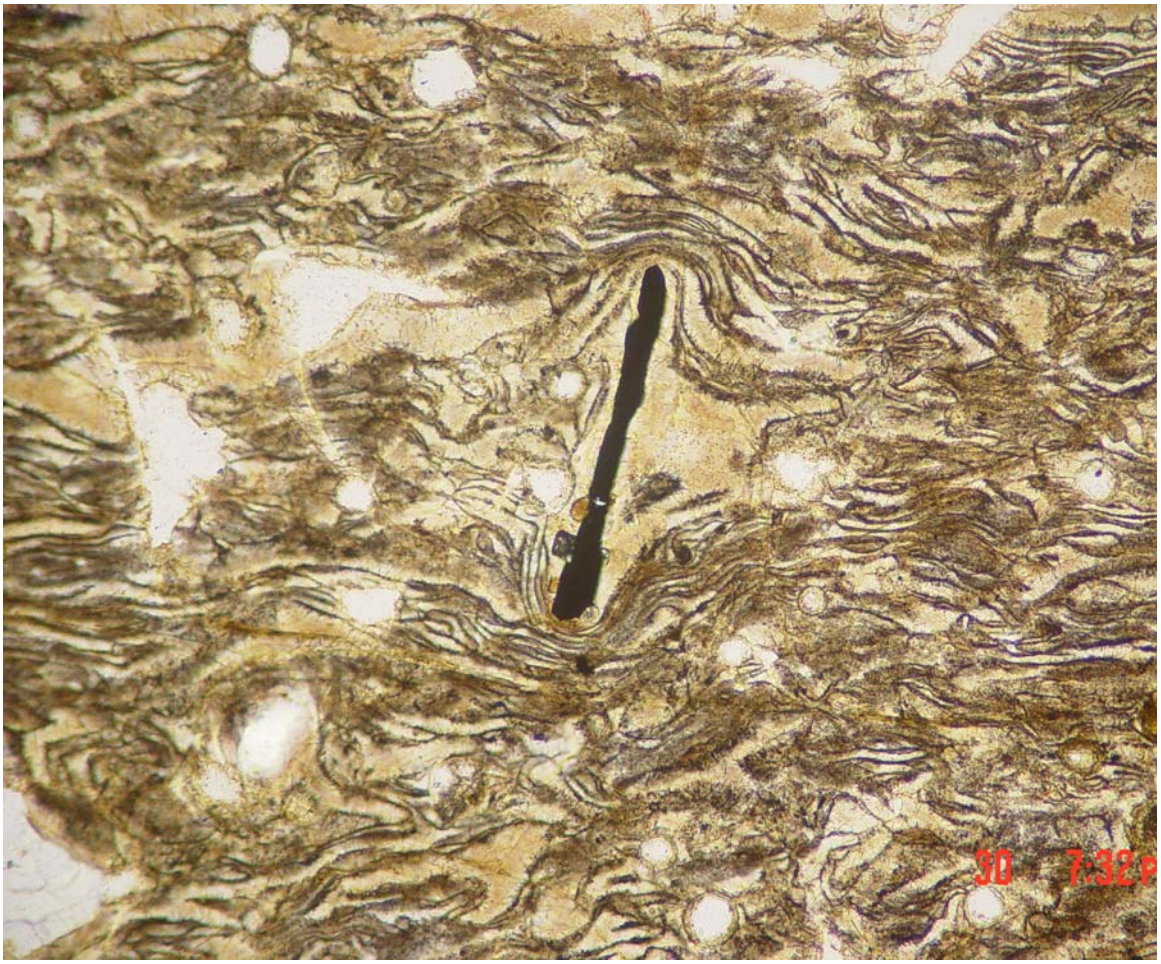

Figure 4.2 Photomicrograph of the stoney interior of tuff of Fir Grove. The long slender opaque mineral is ilmenite. Note the aspect ratio of the shards and the way they wrap around the ilmenite grain suggesting a high degree of welding.

Tuff of Knob (Tk)

This is a normal polarity, crystal-rich (20%), plagioclase-pyroxene-bearing, vitric rhyolitic ignimbrite with minor amounts of quartz, magnetite, and sanidine. It contains accessory zircon and apatite. Knob is dominantly lithic-poor, except in section 20, northwest of Davis Mountain, where it contains 10-20% lithics of Challis Volcanic Group affinity, up to 5cm in diameter.

The tuff of Knob is typically 120 meters thick, up to 140 meters. It is exposed in Deer Creek and East Clover Creek canyons north of the Pole Creek fault zone. The tuff of Knob has been traced 6 km to the east to Fir Grove Mountain, and by Honjo (1990) 16 km to the west to Deer Heaven Mountain.

The tuff of Knob unconformably overlies the tuff of Fir Grove. The unit thickens to the south and is inferred in the subsurface south of the Pole Creek fault. In the north the tuff of Knob is in unconformable contact with the Challis Volcanic Group.

The tuff of Knob is an informal unit first described by Honjo (1990). This is the youngest of the inferred Bruneau-Jarbidge units at 10.6±0.2 Ma, determined using Ar-Ar methods.

The tuff of Knob has black upper and lower vitrophyres with a rough and grainy texture. The unit has a planar foliation that becomes highly contorted at certain intervals near the top of the ignimbrite. Moderate amounts of weathering of the vitrophyres bring out the foliation.

The transition between glassy and devitrified rock is usually very conspicuous, having an abrupt change in color, texture, and weathering profile. The devitrified portions have a bluish, light-gray color when fresh, and become reddish as the amount of weathering intensifies.

Fresh vitrophyre is generally black (light-colored where weathered and hydrated). As with the younger tuff of Thorn Creek, spherulites become abundant at the transition of vitrophyre to devitrified material. Also the contact is characterized by alternating layers or pods of glassy and devitrified material.

Columns are common in this unit and resemble the outcrops of tuff of City of Rocks and the tuff of Thorn Creek. Where the tuff of Thorn Creek and the tuff of Knob are in direct contact with each other they are difficult to distinguish. The columns are generally 5-10 meters tall and 3 meters in diameter with a vertical orientation.

The upper portions of the tuff of Knob are generally quite vesicular. In section 34, just to the east-southeast of Davis Mountain, the upper portion of the devitrified zone is very vesicular, approaching 50% of the rock by volume. At that location, the vesicles range from irregular disk-shapes to cigar-like irregular tubes, giving evidence for differential stretching, and compaction or welding.

The top of this unit, in rare exposures, has a crystal-poor (<1% phenocrysts), orange, non-welded ash facies. This can be seen in the northern portions of section 27 to the north-northeast of Davis Mountain. These ash beds above the upper vitrophyres are considered post-ashflow fall-out deposits from the same eruption or an eruption shortly after the main ignimbrite-forming eruption.

Phenocrysts of the tuff of Knob include 50% 3mm euhedral, sieve-textured plagioclase, An_{30}. Pyroxenes account for 40% of the phenocrysts: 30% augite, 60% pigeonite, and 10% orthopyroxene. The pyroxenes are subhedral 0.5mm grains. The plagioclase and pyroxene grains occasionally can be observed as 2-3mm glomerocrysts. Quartz grains are minor (<2%), 0.3mm anhedral grains. Only a few grains of sanidine can be seen in thin section. Sanidine grains are subhedral 1mm long grains. The remaining phenocrysts (7%) are anhedral opaque grains.

The tuff of Knob is distinguished geochemically as a metaluminous, low silica (71-72 wt. %), low Ta (2.5 ppm), high Zn (85-95 ppm) rhyolite.

Basalt of Davis Mountain (Tbd)

This is a normal polarity, porphyritic, intergranular, diktytaxitic, olivine tholeiite basalt with plagioclase and olivine phenocrysts. This basalt is typically 20 meters thick, but reaches 70 meters in the western portion of the DMQ.

The basalt is exposed mostly to the north of the Pole Creek fault zone. It is the upper-most unit north of the fault zone. Minor exposures of the basalt of Davis Mountain can be seen south of the fault zone. This is the youngest unit deformed by the Pole Creek fault. It is displaced down to the south. This relationship is best observed in the Squaw Creek canyon near the Davis Mountain road where the top of the basalt of Davis Mountain is exposed.

The basalt of Davis Mountain unconformably overlies the tuff of Knob. South of the fault zone the basalt has an unconformable angular relationship with the overlying Twin Falls-aged sequence. This basalt separates the inferred Bruneau-Jarbidge sequence from the Twin Falls-aged sequence. To the south of Davis Mountain peak the unit is tilted 4° to the south. North of the peak, the unit is mostly flat-lying.

This unit was sampled for geochemical analysis by Honjo (1990) and was included in that study as Tertiary basalt undivided. In this study this unit is given the informal name basalt of Davis Mountain for the outcrops at the peak of Davis Mountain.

The unit generally forms low-amplitude, hilly topography. It is not a cliff-forming unit. Outcrops are blocky and rubbly with large-diameter (up to 5 meters) blocks and boulders. The boulders frequently spall spheroidally.

The upper portion of the unit is vesicular. The vesicles are large, up to 3cm, and usually partially filled with white calcium carbonate or reticulated zeolite (?) minerals. Vesicle pipes are common and are 10cm wide.

In hand sample the unit appears reddish-brown on weathered surfaces with a dark-gray matrix and clear phenocrysts of plagioclase and minor red-altered olivine grains. Also in hand sample and at outcrop-scale, the unit appears medium-brown to dark-brown in association with a rusty-brown soil.

The exposure at the Hill City-Bliss road in the southeast corner of the map is a much more intensely weathered. Pillow structures and associated palagonitic material can also be seen with vesicular zones. The glomerocrysts are much larger, up to 3cm, and contain larger and more crystals. Olivine grains are visibly (seen without a hand lens) altered to a reddish iddingsite(?). Vesicles are coated with a blue-gray clay-like material and have dendritic manganese oxides (pyrolusite?).

The basalt of Davis Mountain is porphyritic with ~10% phenocrysts of plagioclase and olivine. The estimated phenocryst content of the unit varies ranging from 7% to 20%. Phenocryst abundance is heterogeneous at outcrop-scale. Plagioclase grains are euhedral, 3mm-long and account for ~60% of the phenocrysts. The olivine grains are subhedral, 1.5mm, and account for the ~40% of the total phenocrysts. Magnetite and ilmenite are present as a minor phase.

Many of the samples show a weakly diktytaxitic texture with stellate plagioclase and olivine glomerocrysts. The matrix of the rock dominantly exhibits an intergranular texture. However, one sample (04WL019) exhibits a subophitic texture.

The basalt of Davis Mountain is chemically distinguished by high silica (48-49 wt %), low P_2O_5 (0.16-0.21wt %), high Cu (55-82), low Pb (4-9 ppm), low Dy (4.3-5.3 ppm), low Er (2.5-3.1 ppm), low Tm (0.3-0.4 ppm), low Yb (2.2-2.8ppm).

Vitrophyre of Pole Creek (Tpv)

The vitrophyre of Pole Creek is a crystal-poor (3%), plagioclase lithic-vitric rhyolitic ignimbrite with pyroxene, quartz, and magnetite. The unit is exposed along Pole Creek between Deer Creek and Squaw Creek. Where exposed, the unit is typically 6 meters thick, with a maximum thickness of 9 meters in Deer Creek.

The unit overlies the basalt of Davis Mountain south of the Pole Creek fault. To the west of Pole Creek in Deer Creek the unit dips up to 13° to the south. In Squaw Creek the unit dips ~5° to the south.

The vitrophyre of Pole Creek is a new unit first recognized in this study. The unit is directly overlain by the 10.1 Ma tuff of Thorn Creek. This thin unit is the oldest of the Twin Falls-aged units exposed in the DMQ.

The exposures of this unit represent the distal facies of the ignimbrite. The unit terminates against the Pole Creek fault. It has very fine-grained lithic fragments. The direction of elongation of stretched vesicles is southeast/ northwest measured from a sub-horizontal foliation.

Much of the unit is comprised of black, glassy material. The upper 1 meter of the unit is orange-brown non-welded lithic-rich ash. Outcrops are generally weathered and poorly exposed. The best exposures are along the west side of Deer Creek just north of the confluence of Pole Creek on the east side of Deer Creek. The vitrophyre of Pole

Creek is a slope-forming unit. Weathered surfaces of the vitrophyre are gray to brown color with spots of rusty color (hematite?) surrounding the oxides.

The vitrophyre of Pole Creek has ~3% phenocrysts of subhedral 1mm plagioclase grains, subhedral 0.5mm pyroxenes (pigeonite 45%, augite 50%, and orthopyroxene 5%) with minor amounts of quartz and magnetite. Trace amounts of zircon and apatite can be seen in thin section. Similar to the tuff of Fir Grove the vitrophyre of Pole Creek has a shardy texture. Phenocryst aggregates of plagioclase, pigeonite, and augite are found in this unit. The plagioclase and pigeonite phases exhibit a sieve texture.

The vitrophyre of Pole Creek is metaluminous and can be chemically distinguished from the other crystal-poor ignimbrites using these elements: Rb (190 ppm), Sr (90 ppm), TiO_2 (0.55 wt %), Ni (10ppm), Cr (50 ppm), and V (11 ppm).

Tuff of Thorn Creek (Tt)

The tuff of Thorn Creek is a normal-polarity, crystal-rich (20%), plagioclase-bearing, vitric rhyolitic ignimbrite with pyroxenes and magnetite. Trace elements include zircon and apatite.

The unit is exposed to the south of the Pole Creek fault zone and extends into the subsurface at Clover Creek. The best exposures are in Deer, Squaw, East Clover and Clover creeks. The ignimbrite thickens dramatically to the south in the Deer and Squaw Creeks where the base of the unit is exposed. Thickness ranges from 12 meters in vicinity of Pole Creek to over 100 meters at Clover Creek where the basal vitrophyre is not exposed.

The tuff of Thorn Creek unconformably overlies the vitrophyre of Pole Creek. The lower contact of the unit dips more than the upper contact. The unit pinches out rapidly against the Pole Creek fault scarp.

This ignimbrite was originally described by Smith (1966), which he called the "lower welded tuff." Honjo (1990) suggested the name rhyolite of Thorn Creek for the exposure at Thorn Creek in the McHan Reservoir Quadrangle to the east (not to be confused with the Thorn Creek to the west, south of Deer Heaven Mountain). Smith's observations indicated good exposures of the unit at Thorn Creek; however the base of the unit is not exposed in that region. Armstrong et al., (1980) present a K-Ar age of 10.1±0.3 Ma for this unit.

The unit is characterized by large-diameter (up to 3 meters) columns that are similar to the columns of the younger tuff of City of Rocks, but with less regularity. Along the east margin of the study area, the tuff of Thorn Creek is in unconformable contact with the tuff of Knob. In hand sample these two units are indistinguishable. At outcrop-scale the tuff of Thorn Creek has short, wide columns and tuff of Knob has taller, narrower columns.

A section of the tuff of Thorn Creek was measured in the Squaw Creek canyon where the exposure is nearly continuous. The ignimbrite has a vesicular zone at the upper contact but lacks the large lithophysal cavities found in the crystal-poor units. Spherulitic textures are common at the upper and lower transitions from glassy to crystallized material.

Phenocryst content is generally homogenous in terms of variety and abundance of crystals. The assemblage is dominated by 3mm euhedral plagioclase phenocrysts contributing 20% of the total rock by volume. Less abundant phases include 1mm subhedral pyroxenes (40% augite, 60% pigeonite). Plagioclase and pigeonite can be found together in crystal aggregates. Sub-anhedral opaque minerals contribute 2% of the rock.

The tuff of Thorn Creek can be geochemically distinguished from tuff of Knob with higher Ta values (3.0-3.3 ppm for Thorn Creek), lower Zn (75 ppm) and higher Nb (46-52 ppm).

Tuff of Gwin Springs (Tgs)

The tuff of Gwin Springs is a normal-polarity, crystal-poor (5% crystals), plagioclase-bearing, vitric rhyolitic ignimbrite with pyroxene and lithic fragments. Trace amounts of sanidine and ilmenite can be seen in the unit.

The tuff of Gwin Springs is found south of the Pole Creek fault zone. Near the confluence of Deer and Clover creeks in section 19 the unit is a very thin vitrophyre (3 meters) capped by an orange, non-welded ash layer with fragments of the vitrophyre incorporated into the ash.

The tuff of Gwin Springs thickens to the east and south. From lower Squaw Creek to the east, the unit is typically 60 meters thick. Exposures of this unit extend into the subsurface to the south along Clover Creek at the lower boundary of the DMQ.

Like the tuff of Thorn Creek, the tuff of Gwin Springs has differing upper and lower contact orientations, suggesting that the unit was emplaced onto a non-horizontal surface. In Squaw Creek canyon the upper contact dips 5° to the south where as the lower contact dips 6.5° to the south. At Squaw Creek the tuff of Gwin Springs thickens at a rate of 26 meters per kilometer south of the Pole Creek fault where the deposit terminates.

Smith (1966) first identified this unit. He used the term “Gwin Springs Tuff” attempting to formalize the unit. Here, I change the name to the informal “tuff of Gwin Springs” which is consistent with the North American Stratigraphic Code (1983).

The type section of tuff of Gwin Springs is at Gwin Springs in Thorn Creek canyon, east of Highway 46 in McHan Reservoir Quadrangle approximately 20 km east of the DMQ. Smith suggests multiple cooling units for the tuff of Gwin Springs. Only one cooling unit was recognized during this study.

The tuff of Gwin Springs is a Twin Falls-age ignimbrite that is < 10.1 Ma constrained by a K-Ar age by Armstrong et al., (1980).

The unit has well-defined upper and lower vitrophyres and a platy section grading up into a contorted devitrified zone. Immediately beneath the upper vitrophyre is a zone of large rough-wall vesicles 3-4cm in diameter, with minor vapor-phase crystallization and alteration halos around the lithophysal cavities.

The vitrophyres are black to light gray (perlite), generally 3-5 meters thick with spheroidal weathering, and a waxy smooth texture.

The formation varies in color from gray-tan to burgundy-reds and pinks in the devitrified zone. Stretched vesicles are common in near horizontal, non-contorted facies and suggest a final direction of movement in a northeast/ southeast direction.

The unit contains 5% phenocrysts dominantly 1mm euhedral plagioclase (An_{26}) with minor <1mm subhedral pyroxenes (30% augite, 40% pigeonite, 30% orthopyroxene), quartz, and magnetite. Trace amounts of subhedral sanidine and ilmenite can be found with accessory zircon and apatite. Some of the plagioclase and pigeonite grains have a sieve texture. Crystal aggregates contain plagioclase, pigeonite, and augite.

The tuff of Gwin Springs is a metaluminous, moderately high silica (73-76 wt %), high K (~5%) rhyolite. It can be distinguished from the tuff of Fir Grove with Ba (1140-1240 ppm), Rb (172-185 ppm), Sr (64-79 ppm), and Ta (3.2-3.7 ppm).

Basalt of McHan (Tmb)

This is a reverse-polarity, aphyric, subophitic, diktytaxitic, olivine tholeiite basaltic lava flow with abundant shapeless opaque grains.

Exposures of the basalt are limited to the south and southeastern portions of the Davis Mountain Quadrangle. Typically this basaltic lava flow is 12 meters thick. In lower Squaw Creek, the basalt attains a thickness of over 60 meters, probably filling in a depression. To the west in section 19 along Clover Creek the unit thins to 5 meters where it terminates.

The basalt of McHan unconformably overlies the tuff of Gwin Springs. The basalt cannot be traced laterally to the Pole Creek fault zone. The unit pinches out to the north and the west.

The basalt of McHan was originally defined by Smith (1966). The unit can be trace to the McHan Reservoir along Highway 46, ~20 km to the east. Honjo et al., (1986) dated this unit with K-Ar dating methods and determined the age to be 9.4±0.2 Ma.

Smith (1966) suggests a thickening to the north for Tbm. The exposures in the DMQ are insufficient to make any assertion to in that respect due to float and cover from the overlying diatomite. However, the unit pinches out to the north and west, suggesting a southerly or easterly source direction or there was a lower base level in those directions. Also, Smith observed several flows comprising the basalt of McHan. In the DMQ only one lava flow is recognized.

The basalt frequently appears deeply weathered and weathers to dark brown soil. The best exposures are in sections 20 and 28. Where exposed, the top of the basalt is vesicular and the base is massive to blocky with columnar jointing. Vesicles are dominantly spherical or composite where several coalesced.

In hand sample the basalt has alternating zones of dense and diktytaxitic material giving it a layered appearance. It ranges in color, depending on the degree of weathering, from a mottled red and medium-gray (diktytaxitic zones are red) to dark-gray and tan which from a distance gives the rock a dark green appearance.

In unaltered zones, vesicle fillings and coating are rare. Where the rock has been altered the vesicles are coated with tan, white, and yellow clayey calcium carbonate material and dendritic manganese oxides. Weathered surfaces are medium-brown to rusty red.

South of the DMQ the basalt has a pillow-like texture, suggesting it flowed into water. The unit is poorly exposed in most locations.

The basalt of McHan is a low TiO_2 (1.4 wt %), high Al_2O_3 (17.6 wt %), olivine tholeiite basaltic lava flow. The basalt flow can be distinguished from other DMQ basalt flow with Rb and Sr (both low; 4 ppm and 200 ppm respectively). Zr is also lower than the other basalts (85 ppm).

Diatomite of Clover Creek (Tcd)

The diatomite of Clover Creek is a thick deposit (100 meters) of basin-filling sediments: coarse- and fine-grained fluvial siliciclastic layers and lenses, fine-grained lacustrine deposits with diatoms, limestone pods and lenses, and beds of volcanic shards.

This unit is exposed in the southeast corner of the map area. The deposit thins to the north and west, and is thickest along the lower parts of Squaw and East Clover Creek: up to 130 meters.

The diatomite of Clover Creek has an irregular upper contact suggesting the unit was dissected before the overlying units (tuff of City of Rocks and basalt of Burnt Willow) were emplaced. This relationship can be seen on the east side of Catchall Creek canyon where the tuff of City of Rocks is in contact with the basalt of McHan (the diatomaceous sediments were completely eroded away).

The diatomite is in unconformable contact with the basalt of McHan beneath it in the south. At the northern extent of the unit near Pole Creek and Squaw Creek, the diatomite of Clover Creek was deposited directly on the tuff of Gwin Springs to the north of the termination of the McHan basalt.

Moyle (1985) and Toth et al., (1987) evaluated the economic potential of the diatomite deposit. They informally refer to the lacustrine sediments as the “Clover Creek diatomite deposit”. Here, I refer to all of the sedimentary deposits at this stratigraphic horizon as the diatomite of Clover Creek. This includes lacustrine as well as fluvial sediments deposited at this interval between the basalt of McHan and the tuff of City of Rocks.

The main portion of the diatomite body is white to pale tan (depending on the water content; darker varieties have more pore water), massive, ash-rich, fine-grained siliceous sedimentary rock. The unit is approximately 90 meters thick. Lenses of gravel and sand, and pod-like bodies of bedded limestone can be found in the main body of the deposit. Also, lenses and layers of coarse-grained (~1mm), nearly pure ash shards are included in the unit.

The upper and lowermost portions of the diatomite are composed of stream-like sediments consisting of coarse-grained, well-rounded, cobbles and pebbles in a sandy matrix. The clasts are Idavada Volcanics > Challis Volcanics > basalt and opaline material (from hydrothermal alteration and veining of the Challis Volcanic Group). The average clast size is 2cm, with a maximum size of 25cm. Clasts are weakly imbricated and poorly sorted. Proportions of clasts are based on a 100-grain pebble count. This observation was made below sample 05WL043 in section 23 on the west side of the north fork of Catchall Creek in a paleo-channel cut into the soft diatomite.

Diatoms in the white massive layers can be seen with a polarizing light microscope by making grain-mounts of the dust from a fresh surface of a sample. The diatoms vary from whole, unbroken specimens to fine filaments of broken-off spines. Most of the individual diatoms are <0.1 mm. Several types of diatoms (~10) can be found in the deposit. See Bradbury and Krebs (1995) for more information on Snake River Plain diatoms.

Glass shards from this unit were dated with the K-Ar method by Evernden et al., (1964) and an age of 10.3 ±0.3 Ma is presented. This age is also presented in Bradbury and Krebs (1995) and they suggest a corrected age of 10.0±0.3 Ma. This age is inconsistent with K-Ar ages on the underlying basalt of McHan (9.4 Ma) and overlying tuff of City of Rocks (9.2 Ma; Honjo et al., 1986) and is regarded as erroneous.

There are 25 active placer claims in the diatomite of Tcd within the DMQ. The area was first claimed by the Conaway Family in 1910. Currently 42 claims covering 6700 acres cover the deposit (Toth et al., 1987). Diatoms are used as filtration aids, fillers, and insulation (Dolley, 2001).

Figure 4.3 Photograph of a small quarry in the diatomite of Clover Creek. The gray layer in the upper right of the photograph is a 0.3-0.6 meter thick baked zone in the diatomite at the contact with the basalt of Burnt Willow (top right in brown).

Tuff of City of Rocks (Tcort)

This unit is a normal-polarity crystal-rich (25%), plagioclase and pyroxene-bearing, crystal-vitric rhyolitic ignimbrite with minor magnetite.

The tuff of City of Rocks crops out in the extreme southeast corner of the map. Rocks mapped by Smith (1966) extend to the east side of Idaho Highway 46 in Gooding County ~25 km east of the DMQ. The unit pinches out beneath the basalt of Burnt Willow in Section 14. The unit was emplaced on top of the tilted and eroded diatomite of Clover Creek. This relationship can be seen at the Hole-in-the-wall quarry southeast of the DMQ along the Bray Lake-East Clover road, where the irregular surface of the diatomite is mimicked by the basal vitrophyre of the ignimbrite. The tuff pinches out against the diatomite north of Catchall Creek.

This unit was named by Smith and was dated by Honjo et al., (1986) who present an age of 9.15±0.13 Ma. The unit is known for the Gooding County recreational area: Gooding City of Rocks, off Idaho Highway 46 north of Gooding. It is the youngest of the Twin Falls-age rhyolites.

Smith describes the unit's outcrop pattern as "…large, rounded, deeply weathered monoliths bounded by wavy joints and surrounded by reddish-brown granular sand." This is also an accurate description of the outcrops in the DMQ where the stoney, devitrified portion of the ignimbrite is exposed.

This tuff appears to end abruptly in the southeast corner of the DMQ and in some places only a vitrophyre is exposed. This unit is interpreted to pinch-out to the west against a deeply eroded surface of the Clover Creek diatomite.

Figure 4.4 These are pillars weathered out of the City of Rocks tuff. Pillars are 10-25 meters tall. Small red dot in the center of the photograph is P.K. Link for scale. Foliation in the pillars is subhorizontal as see by the platy partings less than one meter thick with isoclinal folds at centimeter-decimeter scales.

The tuff of City of Rocks is a welded crystal-vitric tuff with 25% phenocrysts dominated by 2mm euhedral plagioclase, 1mm subhedral augite (45% of total pyroxene), 1mm subhedral pigeonite (20%), and 1mm subhedral orthopyroxene (35%). Smith reported plagioclase as andesine An_{33-42}, I measured an average of An_{31} for the vitrophyres, with maximum and minimum values of An_{20-46}. The tuff of City of Rocks is one of the few units with an appreciable amount of orthopyroxene which makes up 35% of the total pyroxenes. This is significant considering that the other similar looking units (tuff of Thorn Creek, tuff of Knob, and rhyolite of Deer Springs) have only a trace amount of orthopyroxene. Also, no grains of quartz or sanidine were observed. Thorn Creek and Knob tuffs both have minor amounts of quartz and sanidine.

The tuff of City of Rocks is low silica rhyolite (~70 wt %), nearly dacitic based on chemical distinctions drawn by LeBas et al. (1986). The unit is high in TiO_2 (0.85 wt%), MgO (0.75 wt%), and CaO (2.4 wt%). The tuff is low in Ba (1050ppm), La (75 ppm), Pr (15.5 ppm), Tb (1.6 ppm), and Yb (5.2 ppm).

Basalt of Burnt Willow (Tbw)

This is a normal-polarity, porphyritic, plagioclase and olivine-bearing, weakly diktytaxitic olivine tholeiite basaltic lava flow with a subophitic matrix. The unit ranges from 3 meters thick the along the south rim of Clover Creek at the convergence of Deer Creek, to over 80 meters thick adjacent to the landslide on Clover Creek in section 29, T.3S R.13E. Typical outcrop are 8 meters thick.

This unit is correlative with the Burnt Willow Basalt of Smith (1966). Smith was the first to describe this lava flow. The flows can be trace along the rim of Clover Creek south for 10 km and into Smith's map area 15 km to the east. The unit terminates in the vicinity of the Pole Creek fault zone.

The unit lies in angular unconformable contact with the tuff of City of Rocks, diatomite of Clover Creek and tuff of Gwin Springs. The basalt is dipping 2° to the southeast. The upper surface of the tuff of Gwin Springs dips 5° southeast.

The unit was emplaced onto a surface with discrete gullies. The thickness is uniform except in the Clover Creek canyon on the west side of the map area (where it appears to have filled in a paleo-Clover Creek channel) and adjacent to the section 29 landslide previously mentioned.

It forms a prominent rim-rock along much of the southern half of the DMQ with 1-meter diameter columns up to 25-meters tall. Occasional small slumps and topples can be seen along the rims of the canyons.

In hand sample, the basalt of Burnt Willow has phenocrysts of plagioclase and olivine, and is in varying stages of weathering with thin to moderately thick weathering rinds (1-7mm), dark to medium-gray fresh surfaces, and pods or zones of diktytaxitic textured material in an otherwise massive groundmass.

The unit consists of ~10% 3-5mm-long clear tabular euhedral plagioclase phenocrysts, 2-3% 1-2mm diameter equant olivine phenocrysts that are altered to a dark brown/red color, and rare euhedral 1mm magnetite grains.

The basalt of Burnt Willow is an olivine tholeiite that (compared to the older basalts of the DMQ) is high in TiO_2, FeO*, P_2O_5, V, Rb, Nb, Co, Ce, Dy, Eu, Gd, Nd, Pr, Sm, Ta, Tb, Tm, and Yb. The basalt has slightly lower silica content (45 wt%) than the other basalts of the DMQ.

Quaternary/Tertiary Units

These units include landslides and sediments intercalated in the basalts and rhyolites of questionable ages.

Landslide Deposits (QTls)

Hummocky surfaces with rocky or soil-covered material largely out of place characterize the landslide deposits. I mapped 10 landslides in the DMQ. The southernmost landslides are a manifestation of stream erosion of the diatomite of Clover Creek beneath the basalt of Burnt Willow. The basalt stands in vertical cliffs and the diatomite, a much less resistant rock, is eroded out by the meanders of Clover Creek. The toe of the hillside is removed and allows the weight of the overlying basalt to propagate a toe failure, i.e. the bottom of the hill slides out. Another observed scenario has the diatomite removed to the point of undercutting the basalt resulting in a toppling failure where the top of the basalt tips out and falls away from the hillside.

The landslides in the north (rooted in rhyolitic rocks) generally occur where there is an abrupt change in dip or where several faults come together. These features in the rock are likely places of ground water collection and flow. Abundant groundwater will lubricate fractures and enhance mass wasting processes. The large slide in section 31 has basalt of Davis Mountain at the bottom of the canyon, 200 meters below the nearest outcrop along the canyon rim. This particular slide is a good place to observe hummocky topography and multiple slumps and slide scarps. These landslides are likely recurrently active, with initiation in Pliocene time.

Sediments Undivided (QTs)/(Ts)

This unit includes unconsolidated sandy and gravelly material in between and on top of the Miocene Idavada Group, marked by the symbol Ts. The lithologies of the clasts consist of friable to competent granitic rocks, well-rounded competent rocks of the Challis Volcanic Group, and less dominantly vesicular sub-rounded basalt pebbles. Very rarely yellow-orange-red-mottled silicified wood eroded from the Challis Volcanic Group can be found in the gravel.

The largest exposure of QTs can be found along the east side in the cut of Deer Creek in the east half of section 31 T.2S., R.13E. I suggest this deposit to be a result of the damming of the creek after movement of the landslide farther downstream. These deposits are thus perched gravel terraces. They contain reworked material from the Challis Volcanic Group sediment (Tcvs) at the head of the canyon. QTs deposits are generally less the 1 meter thick.

Quaternary Units

These units consist of materials know to be Holocene (post-12,000 years). They are either currently being deposited (Qal) or have delicate features not yet weathered away (Qbp).

Basalt of Pothole (Qbp)

This is a normal-polarity porphoritic, plagioclase and olivine-bearing, strongly diktytaxitic, olivine tholeiite basaltic lava flow of the Camas Prairie.

The vent is located just off the northeast side of the study area at the Pothole. Cluer and Cluer (1986) have an aerial photograph of this feature. It appears to have produced a small-volume lava flow. It ranges in thickness from 3-7 meters (on the DMQ) and covers an area from the southeast shore of Mormon Reservoir (along the northern border of the Fir Grove Mountain Quadrangle) to the northeast corner of the DMQ (approximately 7 sq. km). The lava flowed to the north, east, and west out of the Pothole away from the Mount Bennett Hills front.

Malde et al., (1963) mapped the unit as Tertiary basalt of the Camas Prairie. Here, I suggest that the unit is Quaternary age based on the preservation of the vent and delicate textures of the flow. Smith's (1966) map area did not include this unit. This is a newly recognized unit.

The basalt of Pothole unconformably overlies the weathered surface of the Challis Volcanic Group in the extreme northeast corner of the map area.

The unit has a vesicular flow top with shelly pahoehoe surfaces still observable. Near the top of the flow some of the vesicles are filled with a light tan to pink calcium carbonate material. Lower in the flow, the vesicles are open, spherical, and visibly diktytaxitic.

The basalt of Pothole has a brownish/medium to dark gray weathered surface with thin (3mm) to no weathering rinds. The surfaces are partially covered with lichen. The fresh surfaces are dark to medium gray with plagioclase phenocrysts creating a felty texture.

The basalt of Pothole has a strikingly similar geochemical signature to the basalt of Burnt Willow. The basalt of Pothole has elevated concentrations of many of the elements that the Burnt Willow basalt is enriched with compared to the older basalts (McHan and Davis Mountain). The major difference in the two basalts is the Al_2O_3 concentration. The basalt of Pothole is depleted in Al_2O_3 with respect to the other DMQ basaltic lava flows.

Alluvium (Qal)

This unit consists of modern stream deposits located in and adjacent to streams. Occasionally deposits of mappable volume exist on top of the plateaus where seasonal runoff is augmented by springs. The units are probably no more than 15 meters thick at most. However the base of the deposit can not be seen.

The alluvial materials contain sediments derived from the Challis Volcanic Group, but are dominated by clasts of Idavada Group rhyolites and basalts and rarely contain granitoid rocks of Idaho batholith affinity. The sediments range in grain size from silt to boulders.

A meandering stream system has developed in places along Clover Creek in the southern portion of the DMQ. Much of the alluvial material in the study area is located along this main drainage. Where the banks are eroded, coarse and fine-grained sequences can be observed. At these locations, the grain-size ranges from silt to pebbles, with cobbles and boulders rarely exposed. The material is generally loosely consolidated and easily disturbed. The banks will stand in vertical exposures. Overhanging material is generally rare and ephemeral.

In the valleys containing flowing streams such as Deer and Squaw Creek, semi-controlled channels (as defined by Summerfield, 1991) produce occasional deposits of alluvial material. These deposits are generally too small to map at this scale. The smaller ephemeral streams have only minor alluvial development.

CHAPTER 5:
PRESENTATION OF DATA AND INTERPRETATIONS SUGGESTED BY MAP RELATIONS

This chapter will address several topics including: petrography, geochemistry, volcanology, economic geology, stratigraphy, structural geology, and a section explaining the geologic development of the Davis Mountain Quadrangle.

Petrography

Nineteen thin sections were examined, one from each of the rock units. Groundmass textures, shape and proportion of opaque grains, and phenocryst assemblages were compared for the basalt samples. Rhyolite samples were analyzed for phenocryst assemblages, and any notable fragmental textures. From these observations several distinctions can be made. Appendix II has tables displaying the results of the analyses organized by rock type and sample number.

Rhyolites

The rhyolite units are very similar in hand sample and at outcrop scale, only distinguishable by phenocryst content, which defines two groups: crystal-rich (15-25% crystals) and crystal-poor (0-10% crystals). There are 4 crystal-rich units and 3 crystal-poor units. Further division of the crystal-rich and crystal-poor units is difficult at the hand sample-scale. They, however, have some distinctions in thin section.

The crystal-rich rhyolites include the rhyolite of Deer Springs, tuff of Knob, tuff of Thorn Creek, and tuff of City of Rocks. These units are all welded, so that much of the fragmental textures have been obliterated. There is no one criterion, property, or feature that will distinguish all these units. However, certain features can

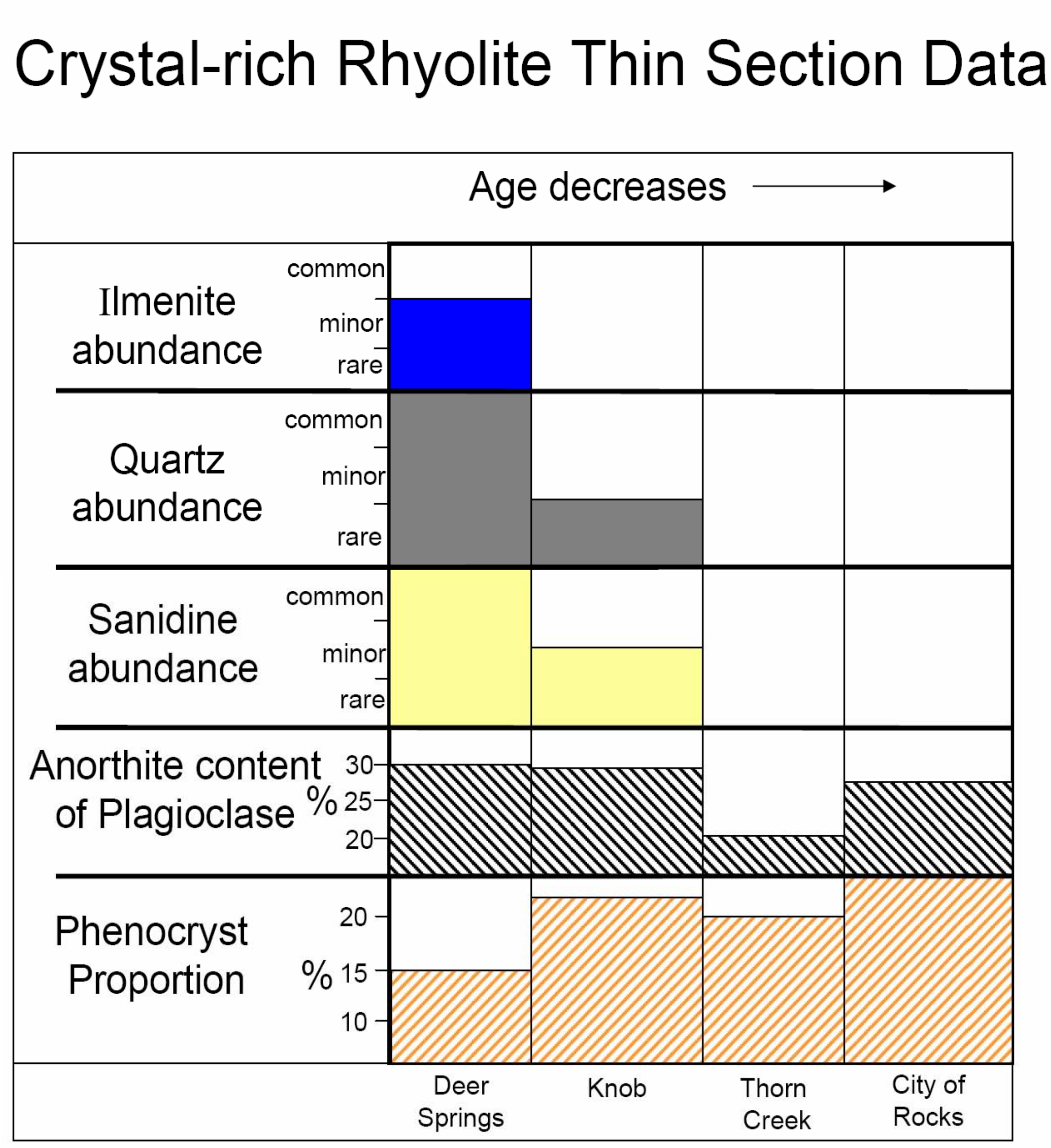

Figure 5.1. Stacked bar graphs showing relative abundances of select phenocryst assemblages of the crystal-rich rhyolites. The data is complied from visual estimates of standard thin-sections using a petrograhic microscope. “Phenocryst Proportion” is a corollary to plagioclase content; plagioclase is always the dominant phenocryst phase in all rhyolite units. Note the lack of sanidine, quartz and ilmenite in the tuffs of Thorn Creek and City of Rocks versus the abundance of the same phenocrysts in Deer Creek. Also note the low anorthite content of the plagioclase in the tuff of Thorn Creek. Number of thin sections analyzed per unit: Deer Springs- 1; Knob-2; Thorn Creek-1; City of Rocks -2.

distinguish certain units, such as the proportion of sanidine. The rhyolite of Deer Springs has appreciably greater sanidine content than the rest of the crystal-rich units, but the other units cannot be distinguished from each other on the basis of their sanidine content.

Figure 5.1 graphically illustrates the distinguishing petrographic features of the crystal-rich rhyolite units. The main distinguishing petrographic features of the crystal-rich (15-25%) units are:

Deer Springs- abundant quartz, sanidine, ilmenite; lower proportion of phenocrysts (15%)
Knob- minor amounts of sanidine and quartz
Thorn Creek- low An-content of plagioclase
City of Rocks- highest proportion of total phenocrysts (25%)

The three crystal-poor units include the tuff of Fir Grove, vitrophyre of Pole Creek, and tuff of Gwin Springs. These units all exhibit a fragmental/shardy texture in thin section. Figure 4.1 demonstrates this texture quite well. Figure 5.2 compares the same features for the crystal-poor units as Figure 5.1 did for the crystal-rich units. As seen in Figure 5.2 the crystal-poor units are less distinct in thin section than the crystal-rich units. Abundance of sanidine and An-content of plagioclase distinguish tuff of Fir Grove from tuff of Gwin Springs but no data could be collected for the An-content of the vitrophyre of Pole Creek. Ilmenite and quartz abundance and total proportions of phenocrysts are all nearly the same for the crystal-poor units and thus not a useful tool when trying to distinguish the crystal-poor rhyolites.

I also looked at the proportions of pyroxenes in thin section to distinguish the rhyolites. Orthopyroxene, pigeonite and augite are the three types of pyroxene present. Most of the rhyolites have nearly equal proportions of pigoenite and augite with little or no orthopyroxene. However, the City of Rocks tuff and Gwin Springs tuff both contain appreciable amounts of orthopyroxene. Figure 5.3 is a ternary diagram displaying the proportions of pyroxenes seen in thin section.

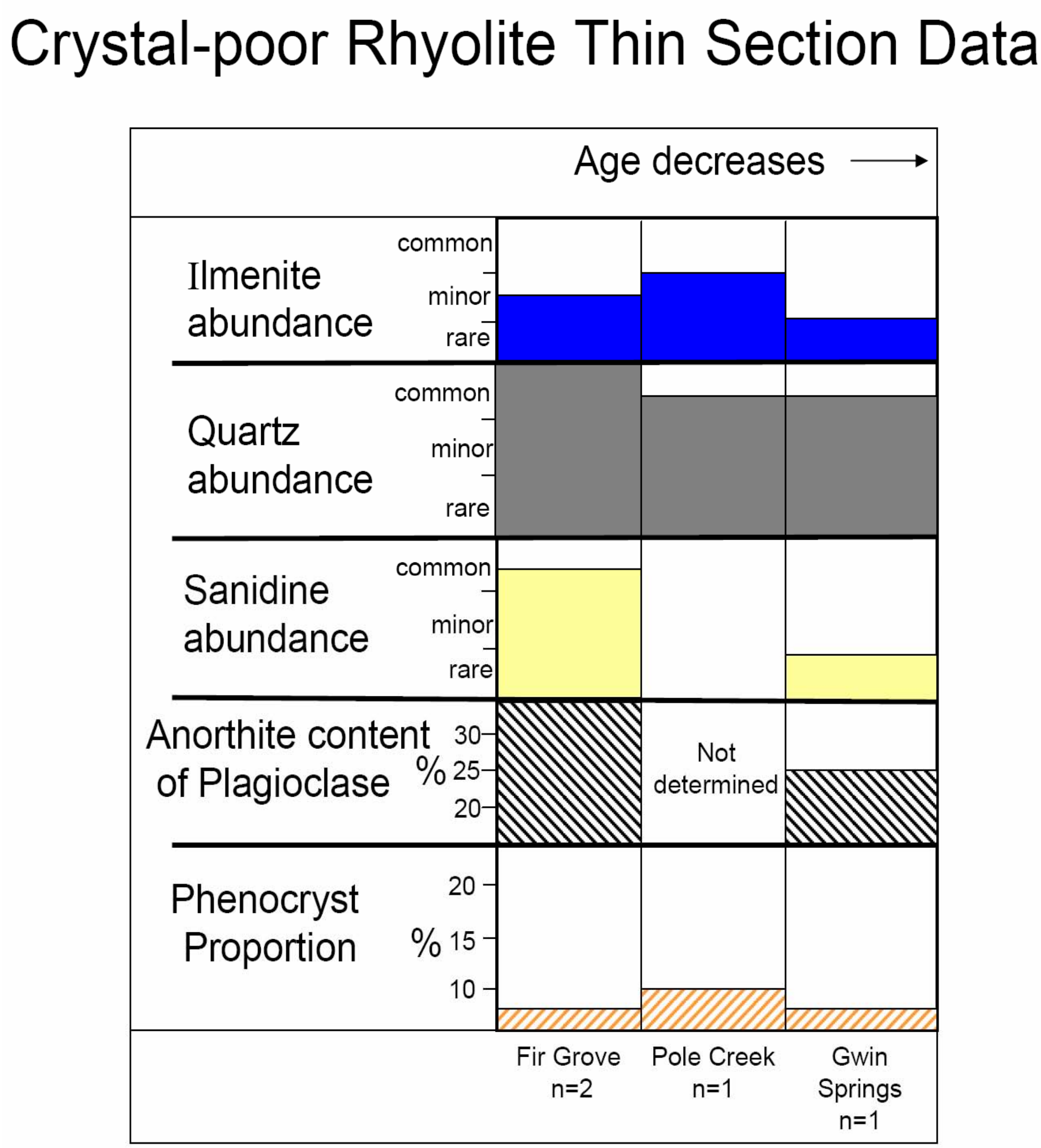

Figure 5.2. Stacked bar graphs comparing selected phenocryst assembledges and abundances for the crystal-poor rhyolites. The data is complied from visual estimates of standard thin-sections using a petrograhic microscope. “Phenocryst Proportion” is a corollary to plagioclase content; plagioclase is always the dominant phenocryst phase in all rhyolite units. Note the difference in the An-content of plagioclase grains. Pole Creek vitrophyre did not have enough grains oriented appropriately such that the An-content could be determined. The notation beneath the unit name is the number thin sections analyzed for each unit.

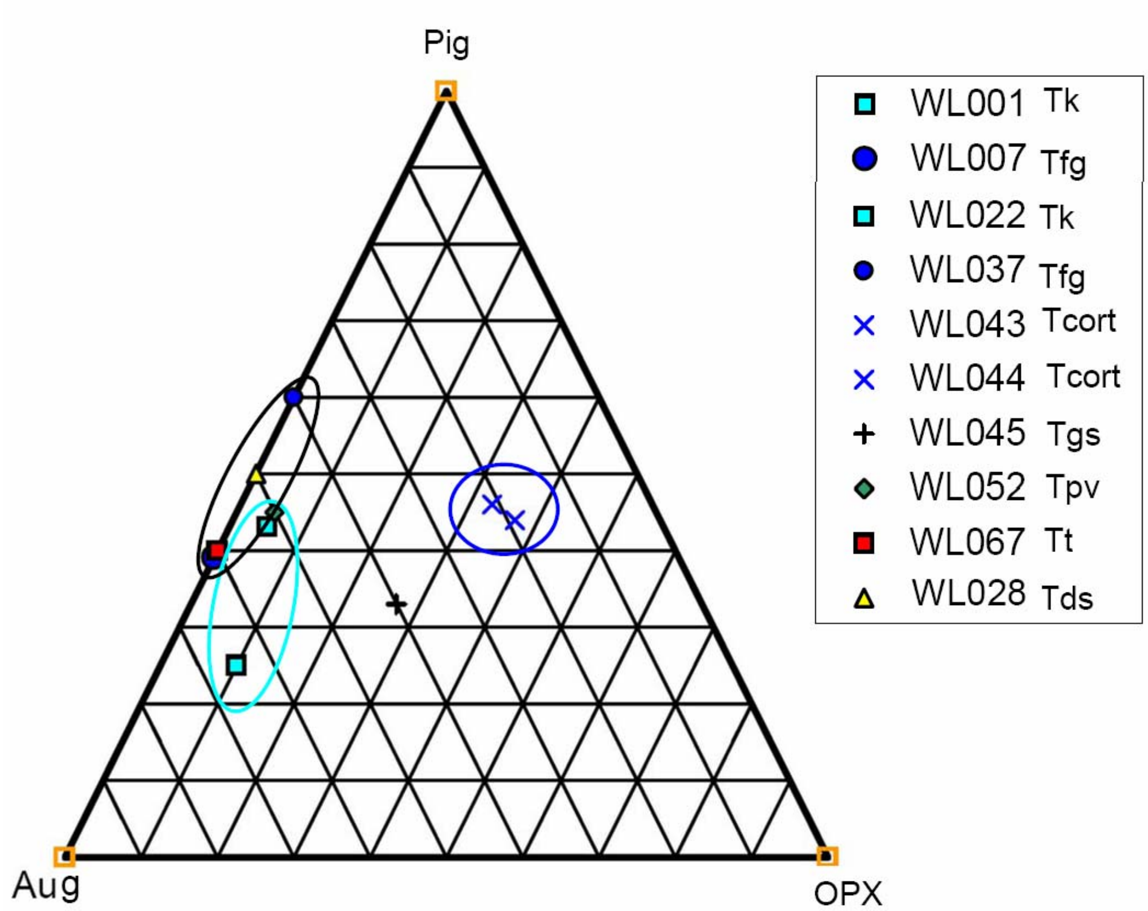

Figure 5.3. Ternary diagram comparing the proportions of pyroxenes of the rhyolites. Circles encompass units with multiple samples. Tk = tuff of Knob; Tfg = tuff of Fir Grove; Tcort = tuff of City of Rocks; Tgs = tuff of Gwin Springs; Tpv = vitrophyre of Pole Creek; Tt = tuff of Thorn Creek; Tds = rhyolite of Deer Springs; OPX = orthopyroxene; Pig = Pigeonite; Aug = augite. Note that Tgs and Tcort have the most OPX and that Tk and Tpv have minor amounts of OPX; Tds and Tfg have no OPX. Also note the near equal proportions of pigeonite and augite in the samples with and without orthopyroxene.

Basalts

Thin section analyses of the basaltic rocks focused on several parameters in an effort to distinguish separate flow units. These parameters include: the proportions of opaque grains (broken down into several sub-groups), plagioclase and olivine, the length of the longest plagioclase, and the total phenocryst abundance.

I have three parameters for comparing the basaltic units using opaque grains. These three parameters are based on the abundance of shapeless, ilmenite-like (long, slender blades), and magnetite-like (equant or square) opaque grains. The "amorphous opaques" parameter includes all opaque grains that are not shaped like ilmenite or magnetite.

The most notable unit for the "amorphous opaques" section is the McHan basalt. This unit is very rich in opaques without a defining shape. I saw no distinct shapes that would suggest either ilmenite or magnetite in this unit.

The other three basaltic rock units contain varying amounts of ilmenite-shaped opaque grains. The basalt of Davis Mountain and basalt of Burnt Willow contained grains that were shaped like magnetite. Tbw has more opaque grains than Tbd as well as greater number of "shaped" grains. The basalt of Pothole has ilmenite-shaped grains and shapeless grains but no magnetite-like grains.

The length of the plagioclase phenocrysts was taken into consideration. As Figure 5.4 shows, the basalt of Davis Mountain has the longest plagioclase grains at 8mm. The other basalts have plagioclase laths less than 4mm. The thin section from the McHan basalt has only one plagioclase grain. The unit is aphyric, whereas the other basalts all exhibit porphryritic textures.

Plagioclase and olivine are common phases present in most of the Snake River Plain basalts (Leeman, 1982; Bonnichsen and Godchaux, 2002; McCurry, M., Hughes, S., personal communication, 2004-2005). Considering this, I compared the proportions of these phases. The basalt of Davis Mountain has the highest olivine-to-

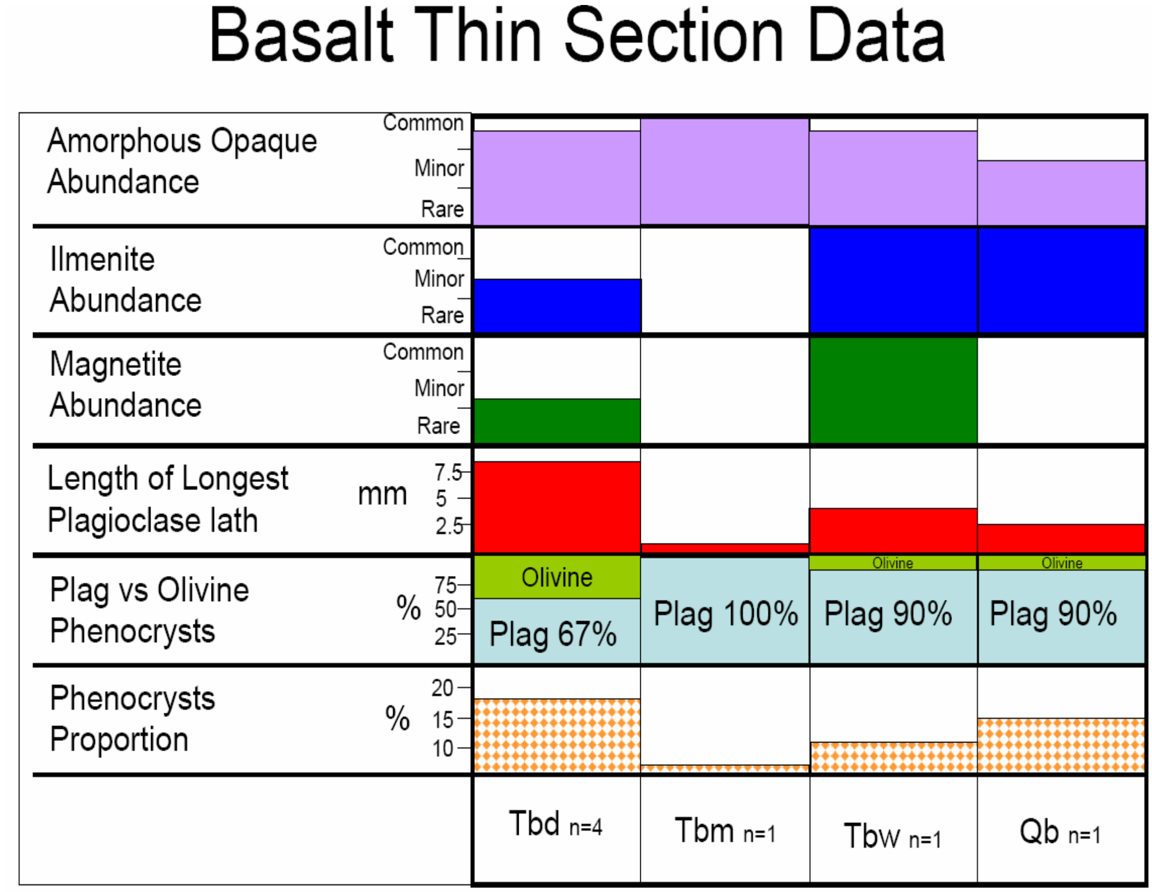

Figure 5.4. Stacked bar graphs comparing various parameters for distinguishing the basalt flows. The upper three parameters are relative abundances where "common" means the grains were easily identified in the section. The notation next to the unit symbol along the bottom represents the number of thin sections examined for that unit. Unit symbols at the bottom: Tbd = basalt of Davis Mountain; Tbm = basalt of McHan; Tbw = basalt of Burnt Willow; Qb = basalt of Pothole.

plagioclase ratio. No olivine grains were observed in the McHan basalt. The basalt of Pothole and the Burnt Willow basalt both have a 1:9 olivine-plagioclase ratio.

The next parameter is total phenocryst content. The younger units stratigraphically above the Pole Creek unconformity become more porphyritic. This does not include the basalt of Davis Mountain which is below the unconformity.

Geochemistry

48 samples were analyzed for major and trace element concentrations. The results of these analyses can be found in Appendix I. The purpose of these data was to find chemical discriminators that can be used to identify particular units, and define the major element chemistry of the rhyolitic units. The rhyolites of the DMQ range from low-silica rhyolites (nearly dacitic) to high-silica < 75% rhyolites.

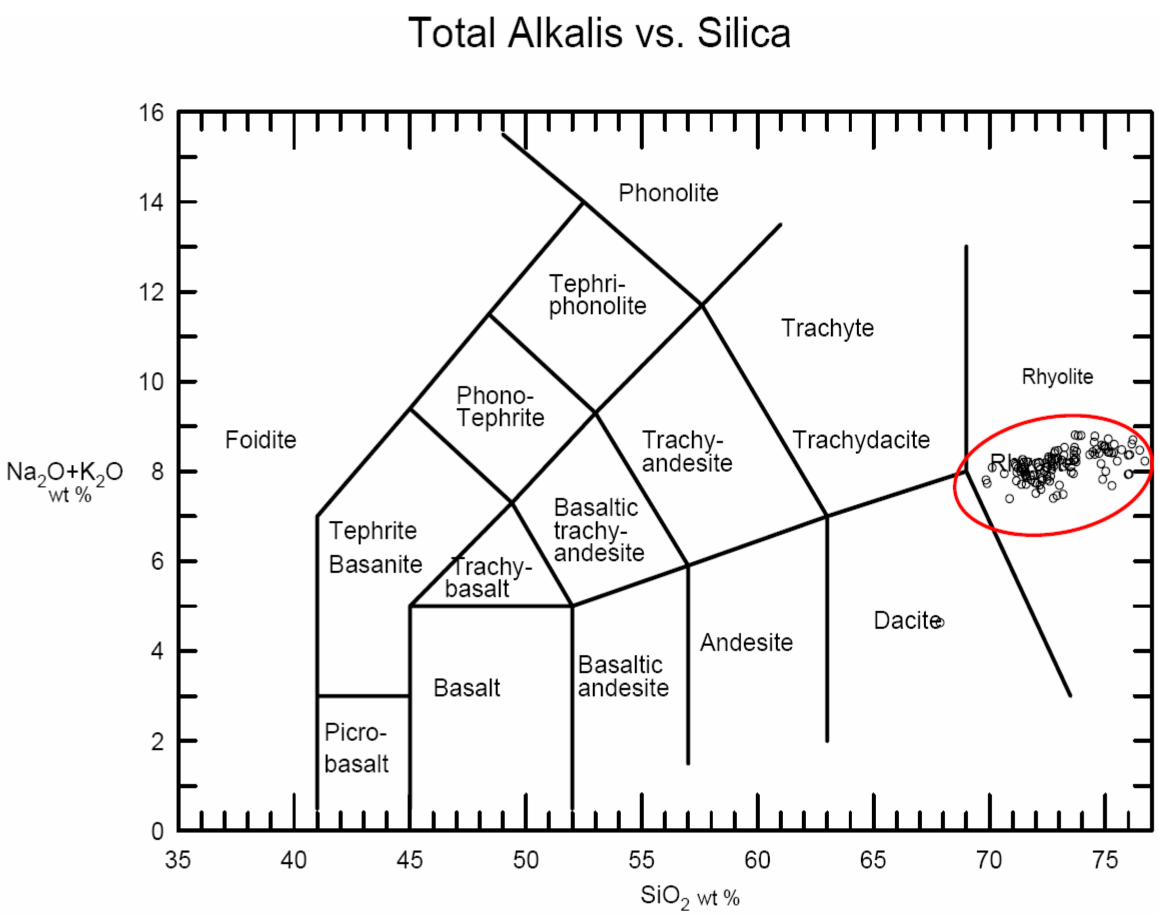

Figure 5.5. Total Alkali-Silica diagram showing the distribution of rhyolites from the MBH (open black circles). The area within the red circle is the field of distribution for the Idavada Group. Field lines on the diagram from LeBas et al., 1986. The data, 113 points, are from: Honjo, 1990; Bill Leeman, unpublished data; Bill Bonnichsen, unpublished data; and this study.

The rhyolites of the Mount Bennett Hills are included in the Idavada Volcanic Group (Malde et al., 1963). This correlation has been supported on an age-basis by Armstrong et al., (1980), Wood and Gardner (1984) and Honjo (1990). It can also be demonstrated on a geochemical basis (see igures 5.5 to 5.10).

Various studies have presented chemical data on rocks of Idavada affinity (Bonnichsen and Citron, 1982; Watkins, 1996; Wright, 1996; Parker, 1998; McCurry and

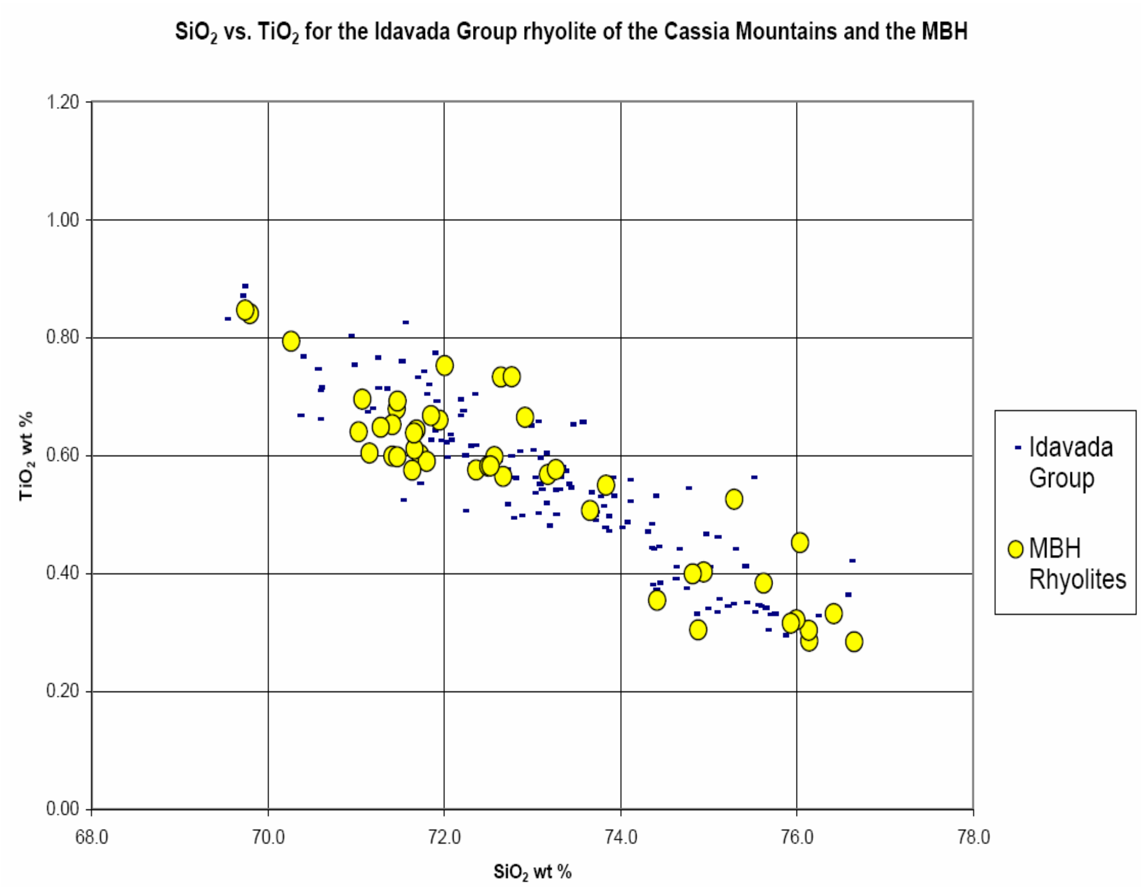

Figure 5.6. The graph illustrates the relationship of the Idavada group (dash marks) in the Cassia Mountains and the silicic volcanic rocks of the Mount Bennett Hills (yellow dots). This graph illustrates the geochemical similarity of the Cassia Mountains Idavada Group rocks and those in the Mount Bennett Hills. Data is recalculated on an anhydrous ferrous basis. Data from Honjo, 1990; Watkins, 1996; Wright, 1996; Parker, 1998; McCurry and Hughes, 2002; and this study.

Hughes, 2002; etc.) and the rocks of the DMQ have similar elemental concentrations. Figure 5.6 shows the relationship of the DMQ rhyolites with the rhyolites of the Cassia Mountains to the south of the Snake River Plain.

Similar to the utility of the petrographic analyses, there is no one parameter that defines each individual unit. There are however, several major and trace elements that differ systematically between particular units.

The utility of a specific element for correlation depends on that element's stability in the environment (i.e. the surrounding minerals and glass will not absorb or expel this element during devitrification and diagenesis in an unpredictable way), invariant or consistent concentration within the unit, ease of measurement, and statistically significant difference in concentration between units. Appendices IB and IC have major and trace element data arranged according to sample.

The following graphs illustrate the function of geochemical data for the purposes of this study: to differentiate units that are otherwise indistinguishable with field methods (paleomagnetically or in hand sample).

The first two graphs, Figures 5.7 and 5.8, illustrate the utility of four elements (Zn, Nb, Ta, and Rb) when trying to distinguish the tuff of Knob from tuff of Thorn Creek. These two units have the same magnetic polarity (see Appendix III) and very similar petrographic features (see Appendix II). Their major elements, when displayed with bivariate diagrams, show every similar major element chemistry, as well.

Plotted on these following two graphs are all of the crystal-rich rhyolites. Some of the units are statically indistinguishable from each other, but remember the purpose is only to distinguish the tuff of Knob from the tuff of Thorn Creek.

In the first graph, Ta versus Rb, note that the Knob and Thorn Creek units have similar Rb concentration of ~ 170 ppm but distinct Ta values (~2.6 ppm and 2.9-3.3 ppm respectively). The samples from City of Rocks plot between Knob and Thorn Creek and are not well distinguished using these elements. City of Rocks can be distinguished from these units using SiO_2.

Ta vs Rb for the crystal-rich rhyolites of the DMQ

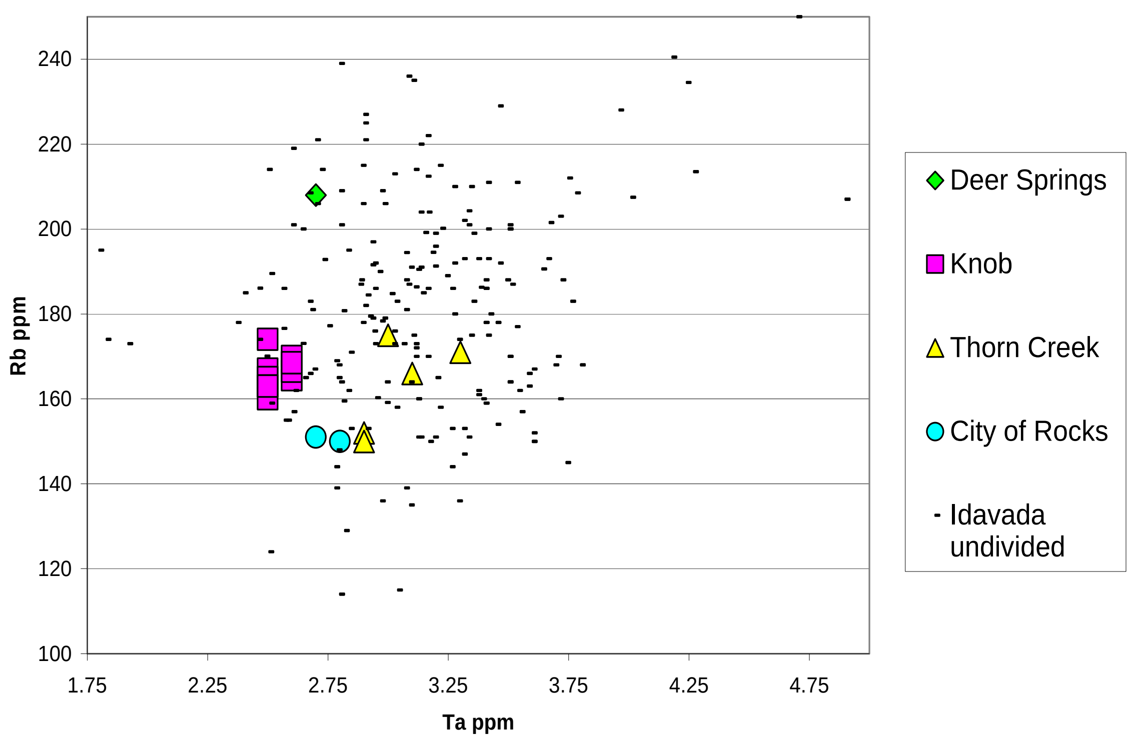

Figure 5.7. The bivariate plot of Ta versus Rb suggests that the tuff of Thorn Creek is enriched in Ta compared to tuff of Knob. Note that the values for Knob are clustered more tightly than those for Thorn Creek. Notice the enrichment of Rb in the Deer Springs unit. The black dashes represent analyses of other Idavada Group units, supporting the premise that the MBH rhyolites belong to the Idavada Group. Data for Idavada Undivided are from the ESPR Database (Mike McCurry, personal communication, 2005) all other analyses are from this study.

In Figure 5.7, note that Rb in the rhyolite of Deer Springs is high relative to other MBH units. Other elements that distinguish Thorn Creek and Knob are Zn and Nb. In

Figure 5.8, Deer Springs is not statistically distinguishable from tuff of Knob using Zn and Nb. City of Rocks shows a smaller distinction from Knob.

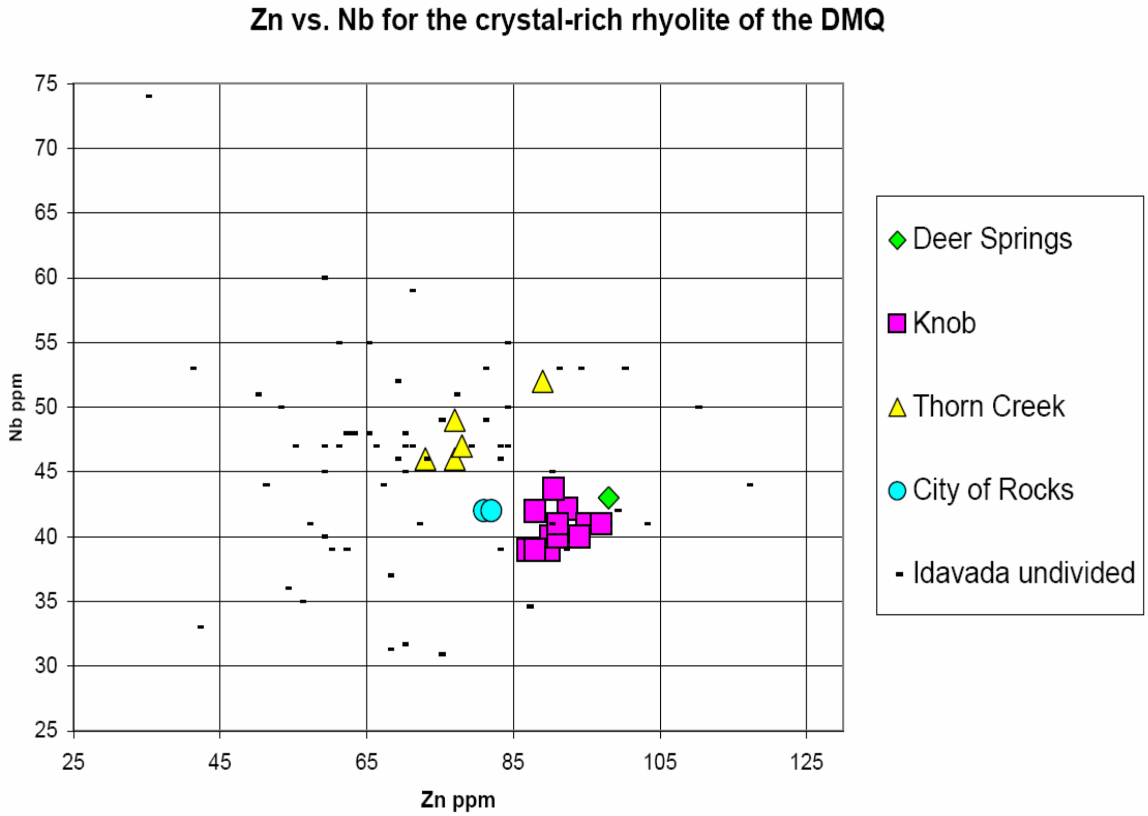

Figure 5.8. The bivariate plot of Zn vs Nb is another way to differentiate the tuff of Knob from tuff of Thorn Creek. Notice the tight cluster of Knob data points. The lone point of tuff of Thorn Creek is a sample (05WL067) from the southern margin of the DMQ. Data is from same sources as figure 5.7.

The data for tuff of Knob clusters much tighter than the data for tuff of Thorn Creek on both graphs. This may suggest that the tuff of Thorn Creek is chemically more heterogeneous than tuff of Knob at the scale it was sampled (hand sample).

The previous two graphs demonstrated the utility of specific elements to distinguish Knob from Thorn Creek, two otherwise similar crystal-rich rhyolites. The following two graphs, figures 5.9 and 5.10 will similarly show distinction of two crystal-poor units.

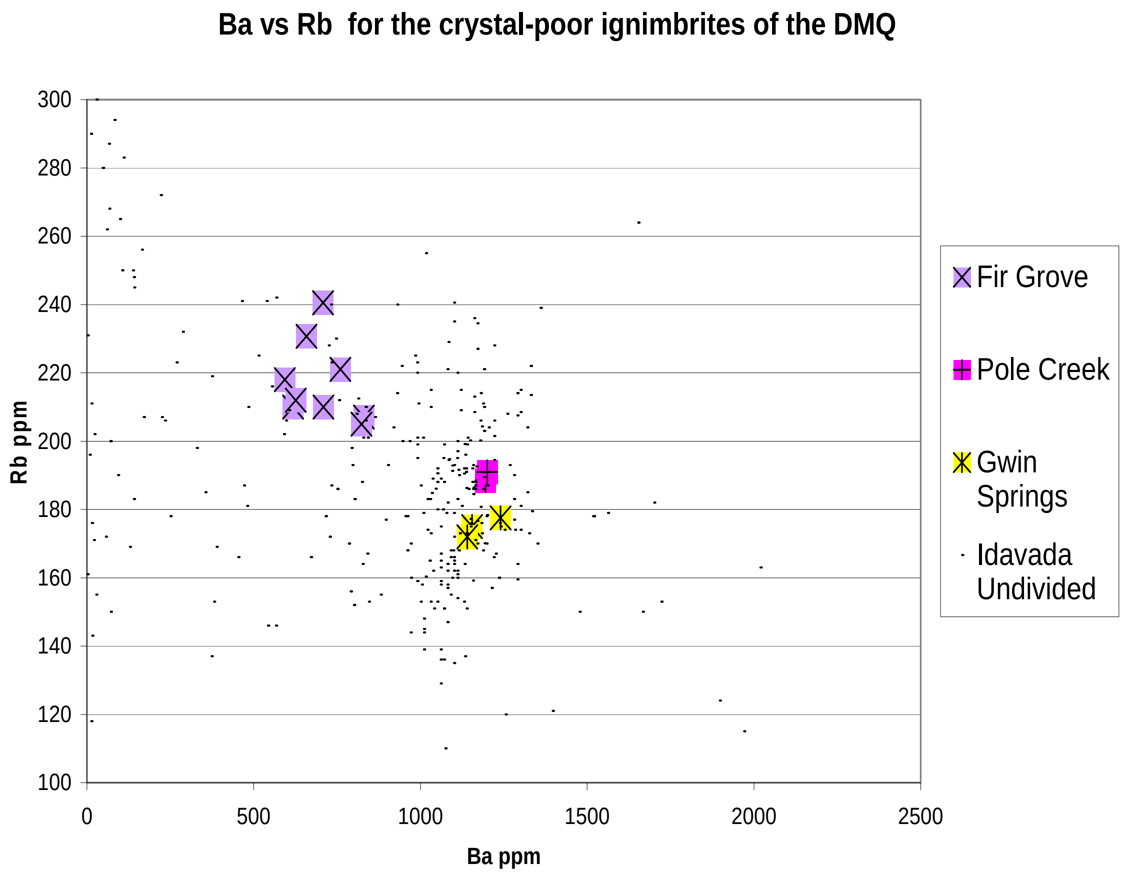

Figure 5.9. This bivariate plot shows the enrichment of Ba in Gwin Springs and the enrichment of Rb in Fir Grove. These elements can be used to differentiate the units if magnetic field measurements cannot be made. Data for Idavada Undivided are from the ESPR Database (Mike McCurry, personal communication, 2005) all other analyses are from this study.

The two units to be differentiated are Gwin Springs and Fir Grove. These two units are very similar in outcrop and hand sample. Smith (1966) said "if it were not for the sediment between them, they would be indistinguishable." Honjo (1990) suggested Fir Grove tuff was probably part of (the same unit as) the Gwin Springs tuff.

Wood and Gardner (1984) went to Smith's type localities to conduct a paleomagnetic survey and found that Gwin Springs and Fir Grove were, in fact, of opposite polarity, suggesting that they are different units. There are also chemical distinctions involving concentrations of Rb, Sr, and Ba.

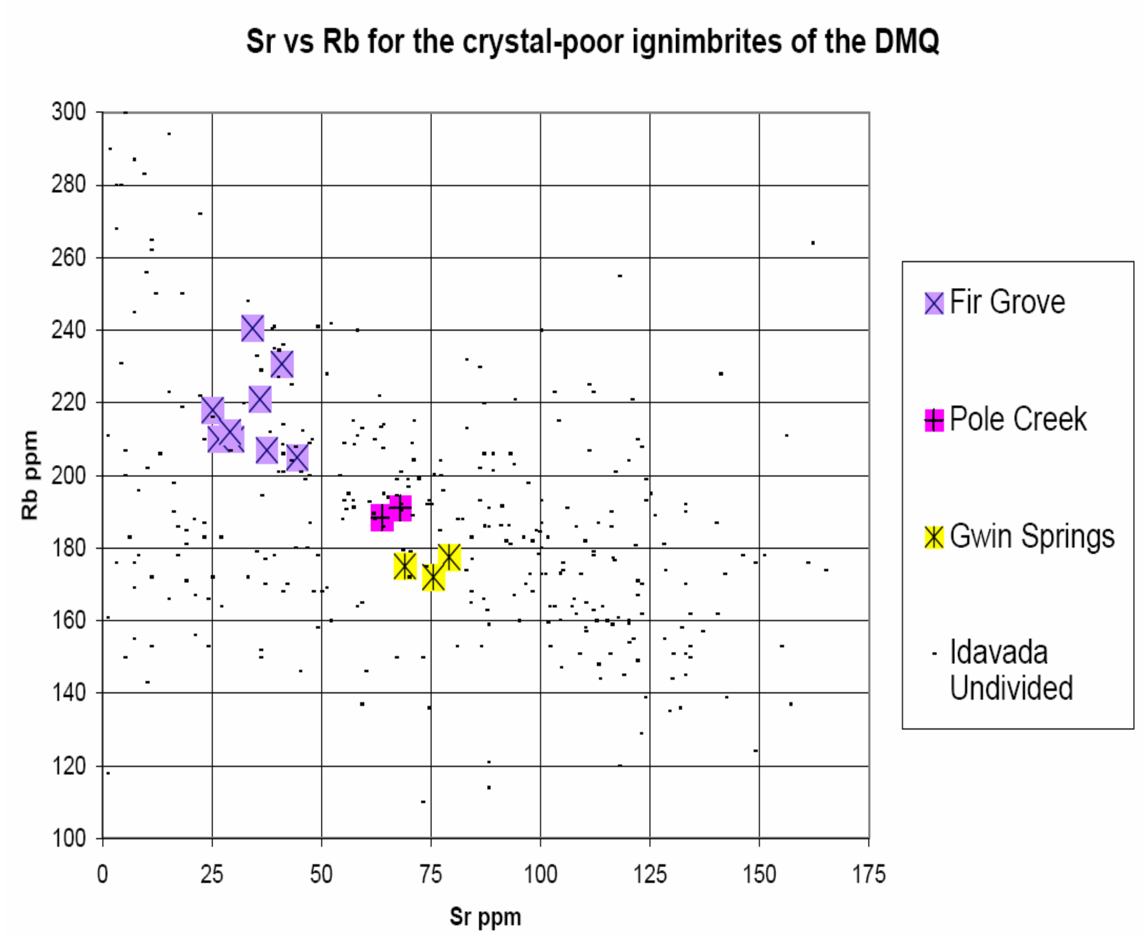

Figure 5.10. Sr and Rb are used here to segregate the tuff of Fir Grove from tuff of Gwin Springs. Note the near-linear trend of the units as they get younger. Fir Grove is the oldest unit identified and Gwin Springs is the youngest. Data for Idavada Undivided are from the ESPR Database (Mike McCurry, personal communication, 2005) all other analyses are from this study.

As seen in Figures 5.9 and 5.10, the geochemistry validates the paleomagnetic work. The Gwin Springs tuff is enriched in both barium and strontium compared to the Fir Grove tuff. Conversely, Gwin Springs is depleted in rubidium with respect to Fir Grove.

The Pole Creek vitrophyre, stratigraphically between Gwin Springs and Fir Grove, plots between them also (i.e. can also be chemically distinguished).

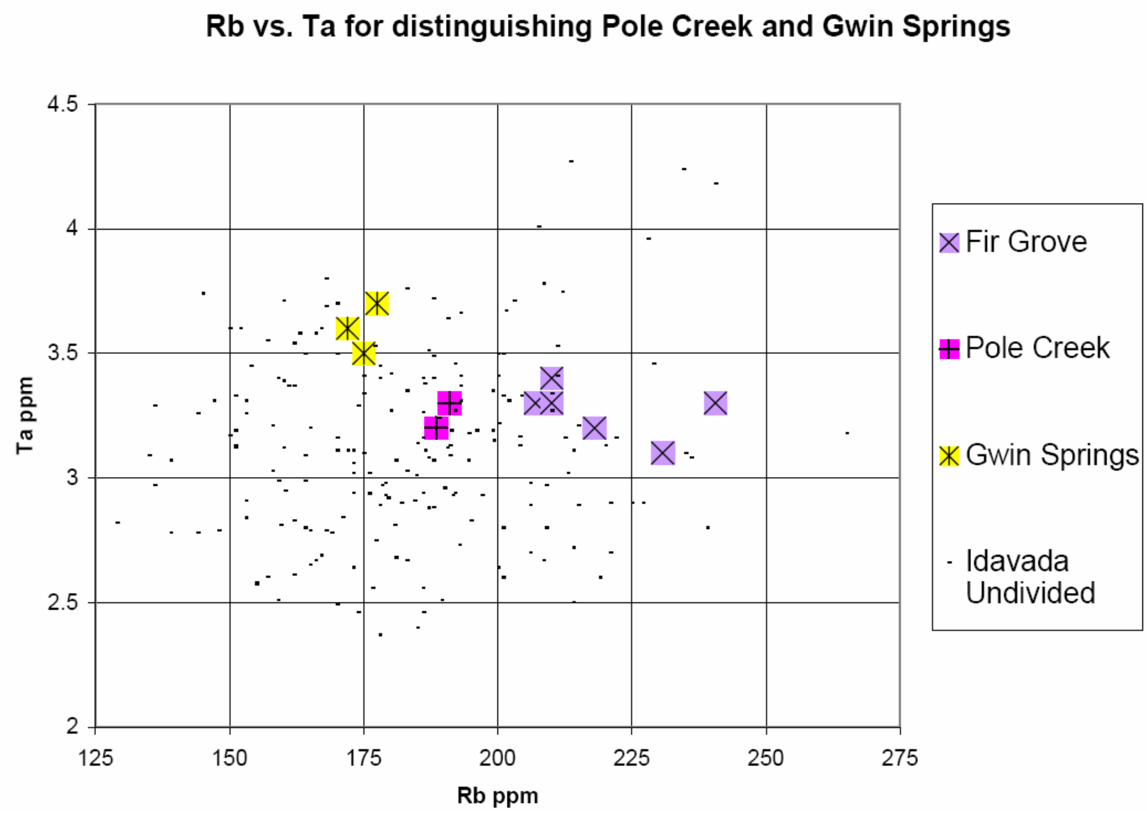

Figure 5.11 This bivariate plot shows a distinction between all three crystal-poor ignimbrites of the DMQ most notably Pole Creek and Gwin Springs. Data for Idavada Undivided are from the ESPR Database (Mike McCurry, personal communication, 2005) all other analyses are from this study.

Figure 5.11 shows a chemical distinction between the Pole Creek vitrophyre, the Gwin Springs tuff, and the Fir Grove tuff. These latter two units however can be distinguished using paleomagnetic signatures. Pole Creek and Gwin Springs cannot be split based on paloemagnetic signatures. As shown in Figure 5.9, Ba values are not distinct for Pole Creek and Gwin Springs (~1200ppm). Rb vs Ta biavariate plot supports the field relations that the units are indeed separate.

Volcanology

Ignimbrite versus Lava Flow Deposits

The rhyolitic rocks of the MBH exhibit features that suggest both pyroclastic flow and lava flow emplacement styles. In general I concluded that all the units were pyroclastic flows. Features suggested by Manley (1996) that indicate pyroclastic flow include wide distribution, broken phenocrysts, heterogeneous phenocryst distribution, and welding. These features are explained in more detail by Manley (1996). Most of these features and textures are not present at each outcrop and thus many outcrops of an individual unit need to be observed to be able to label the deposit as an ignimbrite or effusive lava flow.

Many features of these rhyolites are common to both the lava flow rocks and the fragmented pyroclastic flow rocks. Table 5.1 compares the features generally used to discriminate lava flow deposits and ignimbrites.

The best place to look for pyroclastic flow features is at the top and base of the unit where the shards may have cooled sufficiently to resist being strongly welded or completely coalesced such that the original fragmental texture is destroyed (Manley, 1992; Mike Branney, personal communication, 2005). The basal zones of silicic lava flows will generally consist of blocky breccias (Bonnichsen, 1982). In the MBH, the bases of the units are rarely exposed so using this criterion for distinguishing an ignimbrite from a lava flow is not of much use.

Also, the tops of the deposits are generally either eroded or poorly exposed. Co-ignimbrite ash or highly vesiculated glassy surfaces of lava flows are features expected of the upper zones of silicic units. These features are delicate and easily eroded (Fink and Manley 1987; Manley, 1992).

Thin sections of the DMQ units have some utility in regard to distinguishing the nature of emplacement. The crystal-poor units all exhibit a shardy texture in the interior of the unit with various degrees of welding, and common lithic fragments. Thus, they are fragmental and must be ignimbrites.

The problem of ignimbrite-or-lava flow arises with respect to the crystal-rich units. The units are all very similar physically. They all have large prominent columns with planar and contorted foliations and basal and upper vitrophyres with autobrecciation (Figure 5.12 is a photograph of this feature). The tops and bottoms of these units (the vitrophyres) are generally slope-forming and not well exposed creating uncertainty as to the exact location of the contact. The criteria used to determine the type of deposit are broken crystals and heterogeneous phenocrysts distribution, thickness at termination, and basal breccias.

Heterogeneous phenocryst distribution suggests that the crystals were winnowed out of the ash in the pyroclastic cloud and deposited in a manner similar to sand and gravel bars in fluvial systems. Creating lenses or pods of crystals by plug flow of a viscous medium would invoke unknown processes in the flow. Therefore, zones of crystal enrichment suggest particulate flow, indicating that the deposit is an ignimbrite (Branney and Kokelaar, 1992).

Common features of Lava and Ignimbrites

From Manley (1996)

Discriminator	**Ignimbrite**	**Lava Flow**
Bubble wall shards	ubiquitous	locally in carapace and basal breccias
Pumice	ubiquitous	locally in carapace and basal breccias
Fiamme	ubiquitous	not found
Broken Crystals	common throughout	only locally in welded microbreccia
Lithic Fragments	common	very rare
Gas elutriation pipes	common, but destroyed or distorted during rheomorphism except in static zones such as the base	not found
Welding Zonation	ubiquitous	none
Discrete internal breccias	none: rheomorphism can brecciate chilled margins	found on many scales
True basal crumble breccias	none: however rheomorphism can create one if it follows beyond the original deopsitional boundary	ubiquitous
oxide shells	none	found with mircobreccias
Extreme textural heterogeneity from outcrop to hand sample scale	no, generally gradational	common
Coherent, crystallite-defined flow layering in transparent glass of vitrophyre	none	most common texture in vitrophyres

Table 5.1. This table compares the features commonly used to distinguish lavas from ignimbrites. The table is modified from Manley (1996); see his study for further references.

Figure 5.12. Photograph of breccia in the upper vitrophyre of tuff of Knob. Black clasts are unaltered portions of the glass surrounded by devitrified red matrix. These features are attributed to flow after emplacement, (See Branney and Kokelaar, 1992). The pyroclasts were hot enough to coalesce and remobilize after emplacement. This breccia is a result of the breakup of an outer shell that cooled before the molten interior ceased movement. This feature can also be found in the upper vitrophyres of tuff of City of Rocks and the tuff of Thorn Creek. Photo by author.

Also the variation in crystal content may be a function of gradational welding; allowing phenocrysts to be pushed closer together giving the appearance of varied distribution. Evidence for welding may be used as a proxy for particulate flow. Welding is the compaction of particles.

The rationale behind the broken-crystals-criterion for identifying ignimbrites is this:

1) crystals grow in a magma chamber that is higher than surface pressures,
2) crystals grow to include some of the surrounding magma as inclusions at the elevated pressure,
3) magma ascends to region of lower pressure,
4) the pressure on the surfaces of the crystal is greatly reduced,

5) the magma inclusion is still at the elevated pressure, held by the rigid structure of the crystal.
6) pressure differential between included and excluded magma causes the crystal to break.

Assumptions are that: 1) the magma was under sufficient pressure such that the loss of that pressure will facilitate breaking the crystal as well as fragmenting the magma, and 2) that the magma feeding a lava flow can not depressurize fast enough to facilitate breaking the crystal without fragmenting the magma, (Best and Christiansen, 1997). The phenocrysts do not break due to grain-to-grain contact in the pyroclastic flow.

The rhyolitic units of the DMQ can be divided into two groups: 1) those that are indisputably fragmental and 2) those that are disputable. The indisputably fragmental units are generally crystal-poor (0-10%), and have an interior matrix defined by glass shards exhibiting various degrees of welding and common lithic fragments (Table 5.2). These units include: the tuff of Gwin Springs, the tuff of Fir Grove, and the vitrophyre of Pole Creek. Figure 5.13 is generalized column of the crystal-poor units.

These units, where they are of appreciable thickness, have zones just below (and sometimes including) the upper vitrophyre that include large conspicuous rough-walled vesicles. These vesicles are generally not folded and therefore formed after or at the end of emplacement. These features are absent in the crystal-rich tuffs.

The units that are disputably fragmental include the tuff of Knob, rhyolite of Deer Springs, tuff of Thorn Creek, and tuff of City of Rocks. These silicic units are crystal-rich (10-25%), have a foliation that is contorted at microscopic-scale up to outcrop-scale. They have breccias in the upper vitrophyre, suggesting lava flow, or lava-like flow after emplacement.

Features of the Silicic Volcanics of the Davis Mountain Area

Red = features of ignimbtires **Pink = lava features**

Unit	Bubble wall shards	Pumice	Fiame	Broken Crystals	Lithic Fragments	Gas elutriation pipes	Welding Zonation	Discrete internal breccias	True basal crumble breccias	Extreme textural heterogeneity from outcrop to hand sample scale
Deer Springs	**no**	no	**no**	*minor*	**none found**	none found	*common*	none	no	*no*
Fir Grove	*common*	no	**no**	*rare*	*minor*	none found	*common*	none	no	*no*
Knob	**no**	*?*	**no**	*minor*	*minor, common on NE side of DM*	none found	*common*	none, upper vitrophyre is brecciated	no	*no*
Pole Creek	*common*	no	**no**	none found	*minor*	none found	*unknown*	none	no	*no*
Thorn Creek	**no**	no	**no**	*rare*	**none found**	none found	*common*	none, upper vitrophyre is brecciated	no	*no*
Gwin Springs	*common*	no	**no**	none found	*common*	none found	*common*	none	no	*no*
City of Rocks	**no**	no	?	*rare*	**none found**	none found	*common*	none, upper vitrophyre is brecciated	**Smith (1966) notes an occurrence of basal breccia**	*no*

Table 5.2. Features of the rhyolites of the Davis Mountain area that suggest lava or ash flow emplacement style. Red-colored words suggest ignimbrite and pink words suggest lava flow features.

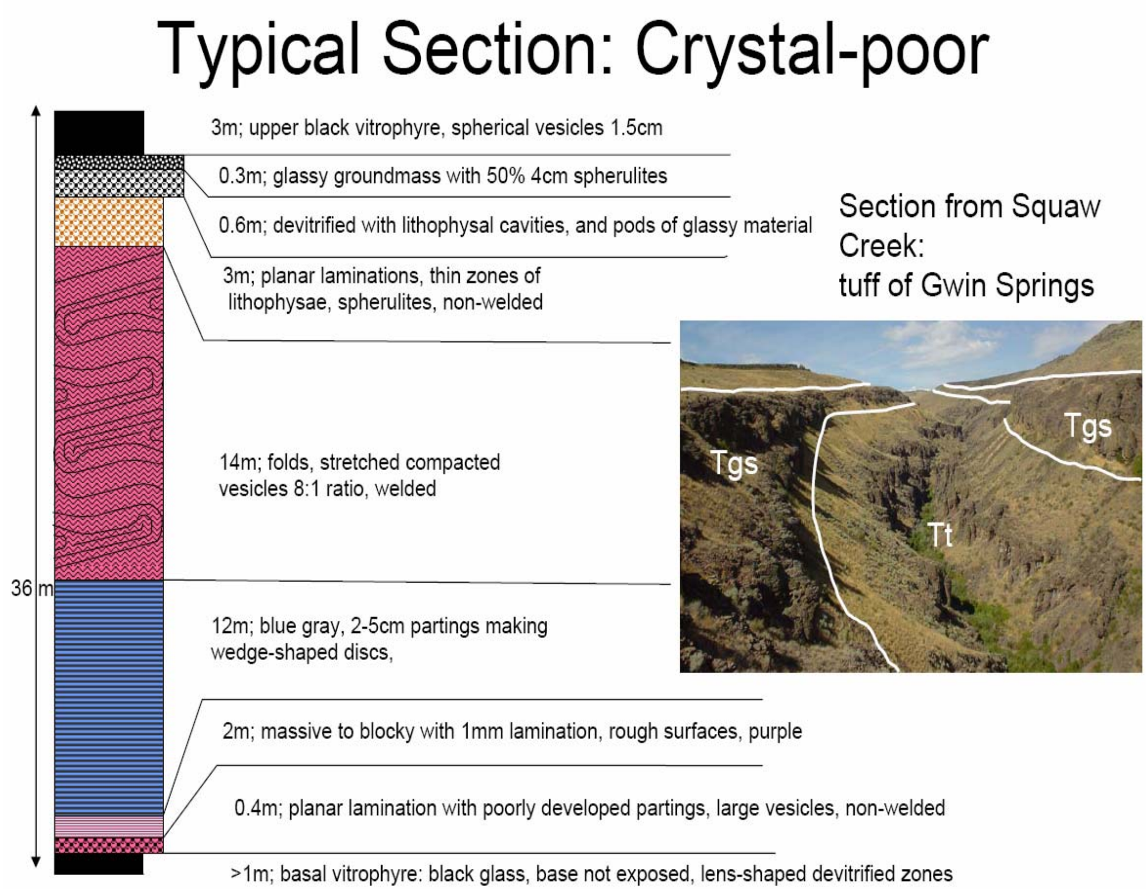

Figure 5.13. Typical crystal-poor ignimbrite of DMQ. This is a generalized stratigraphic column of the crystal-poor rhyolites. Note that there are two major zones of planar lamination divided by a zone in which the foliation is strongly folded. Also notice that the upper vitrophyre is thicker than the lower vitrophyre. This is probably partly due to the welding of the lower vitrophyre. The upper zone of lithophysal cavities is also unique to these units. The large vesicles are not found in the unit when the total thickness does not exceed 10m.

The vesicles are variably compacted suggesting welding, and in turn, a fragmental nature. In places, an oxidized non-welded co-ignimbrite ash is found at the top of the upper vitrophyre (most notable for tuff of Knob). These features are evidence for particulate flow and suggest these units are ignimbrites. A generalized stratigraphic column is presented in Figure 5.14.

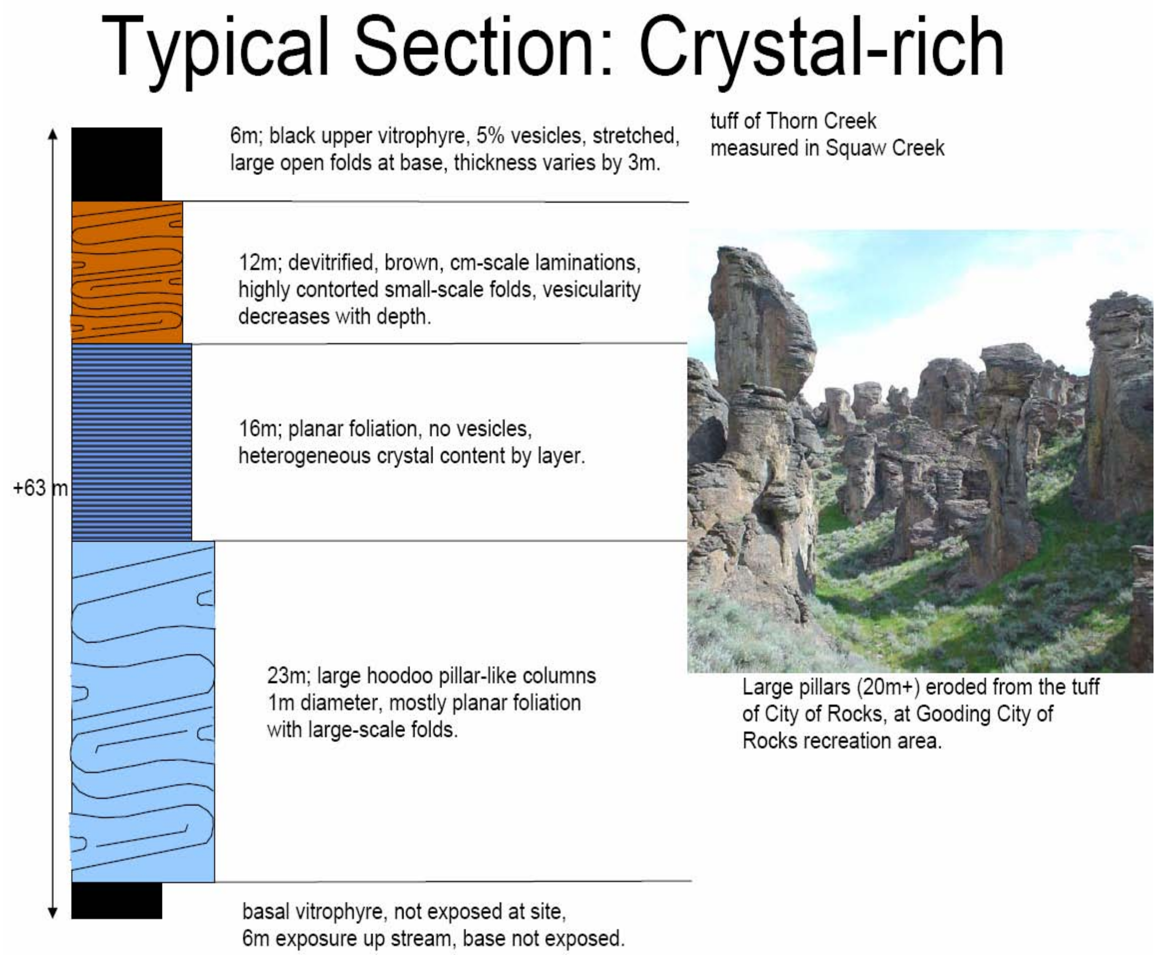

Figure 5.14. This is a generalized stratigraphic column of the crystal-rich rhyolite tuff of Thorn Creek measured in Squaw Creek canyon. The crystal-rich tuffs exhibit two zones with a contorted foliation divided by a zone of mostly planar fabrics. This is in contrast to the crystal-poor units with a two zones of planar fabric divided by a contorted zone. These units also lack the conspicuous large vesicles in the upper zones.

These units lack a basal crumble breccia (at least where exposures permit observation at or near the base if the unit). However Smith noted that one location the tuff of Thorn Creek ("lower welded tuff" of Smith, 1966) did have a basal breccia. The base of the Thorn Creek unit in the DMQ at the distal reach of the unit in Squaw Creek is not brecciated. If the unit were a true lava flow, one would expect to see a "true basal breccia" (as defined by Manley, 1996) at the termination of the flow. Similarly, the tuuf of City of Rocks does not exhibit a basal breccia near its termination.

Ignimbrite	Gwin Springs	Fir Grove	Pole Creek	Knob	Thorn Creek	City of Rocks
Unknown	Deer Springs					

Table 5.3 This table categorizes the silicic units of the DMQ based on textural evidence and features presented in Table 5.2 as: ignimbrites or unknown. The emplacement nature of the rhyolite of Deer Springs is unknown due to the lack of exposure. Thorn Creek is thought to be an ignimbrite but Smith presents evidence that suggests lava flow or ignimbrite.

Ignimbrites may exhibit specific textures and/or features that imply a direction of flow. These direction fabrics include elongated vesicles, sheath folds, and imbricated or linear alignment of particles such as crystal, lithic, or pumice fragments (Branney and Kokelaar, 1992). Vesicle elongation was the easiest of these features to measure. The direction of elongation of the vesicles tends to point toward the southeast, which is toward the city of Twin Falls and the Twin Falls volcanic field. Although this one data point may support the idea that this unit is sourced from the Twins Falls area, it only tells us the last direction of flow for the unit. Local variations in flow direction can be quite substantial, especially if the unit is emplaced on varied, irregular topography (Branney et al., 2002), such as over a fault scarp. Faults in the DMQ were active during the emplacement of several of the units and are discussed later in greater detail in the structural geology and geologic development sections.

Economic Geology

The diatomite of Clover Creek

There are 25 active mining claims within the Davis Mountain Quadrangle. These claims are located in the southeast corner of the map area. They hold the mineral rights of the Clover Creek diatomite. The area was first claimed by the Conaway Family in 1910. Currently 42 claims covering 6700 acres cover the entire deposit.

The first production of diatomite from the Clover Creek deposit began in 1930 (Toth et al., 1987), where it was shipped to Gooding by horse-drawn wagons and then sent by railroad to a Utah sugar factory for use as a filter aid. By the mid 1930's 50-150 tons/year were sold to the Sterling Lumber Company in Twin Falls for use as insulation. In the '50s and '60s the claim owners sold minor amounts of crude diatomite for use as marker lines on ball fields.

Diatomite (a.k.a. diatomaceous earth or DE) is an industrial mineral commodity composed of the skeletal remains of diatoms, used in filters and paints, as a food additive, light-weight building blocks, and insulation (Toth et al., 1987). Historically it has been used in building blocks when the weight of the block was an issue, such as in the domes of St. Sophia in Istanbul Turkey (built in 535 A.D.). Also in the 1860's, it was used as a stabilizer of nitroglycerin for dynamite (Dolley, 2001).

Diatoms are microscopic planktonic plants that secrete an internal skeleton made of silica. They can be found in lacustrine and marine environments. They thrive in waters that are rich in dissolved silica and other nutrients used in their biological functions (Toth et al., 1987). The Clover Creek deposit reaches a thickness of about 90 meters and includes diatomaceous earth, sand and gravel lenses, and primary and

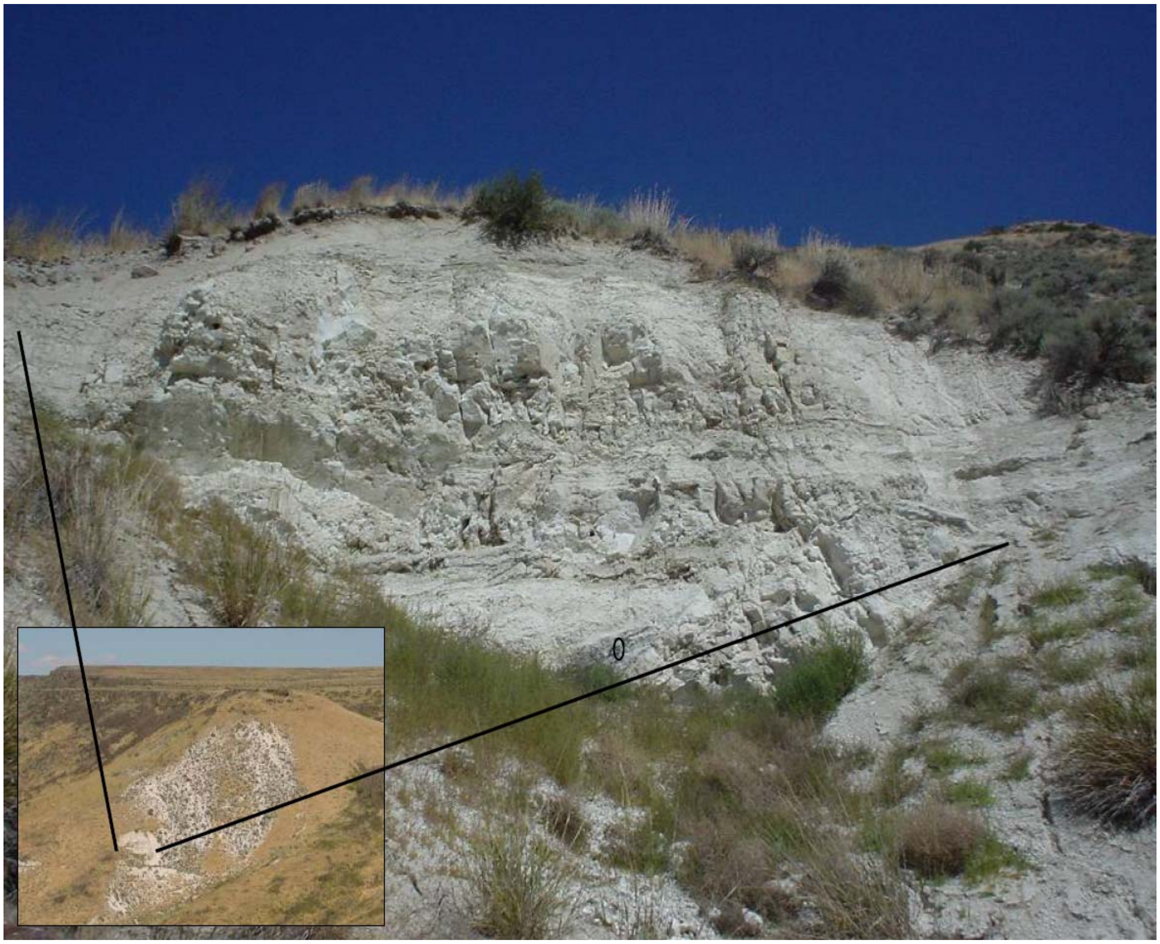

Figure 5.15. Photograph of the Clover Creek diatomite at prospect just south of the DMQ along the east side of Clover Creek. The cut face is about 18 ft tall. Note the horizontally layered strata at the top and the dipping strata at the bottom of the face. This may represent a growth fault that was later onlapped and overlapped by more diatomite; soft sediments deformation may be another explanation. The gray lens at the bottom of the face is a bed of coarse-grained glass shards. The black oval encircles a 500mL water bottle for scale. The inset photograph was taken from across Clover Creek canyon and shows the total thickness of the deposit. Photographs by author.

reworked ash beds suggesting either a basin 90 meters deep existed or a deepening lacustrine basin that accumulated 90 meters of ash, detritus, and diatom skeletons opened and filled with lake sediments. The unit, Tcd, pinches out to the west and north in the DMQ (fault displacement may control the thinning of the unit). The top and bottom of the unit contain many clasts suggesting a fluvial environment at the opening and after filling the basin with diatomaceous earth.

The Clover Creek diatomite deposit contains approximately 416 million tons of diatomite. If modern surface mining techniques were employed to produce 500 tons/day for 300 days/year the mine could produce diatomite for 120 years. Production at this level would account for 50% of the total U.S. non-marine DE output. This would flood the market with DE and unless the demand for DE increases, then the value of DE would drop. The profitability of a Clover Creek operation would be greatly reduced, probably to the point of infeasibility (Dick Conroy, personal communication, 2005).

Ground Water Resources

The Davis Mountain quadrangle has a plethora of springs. Most of the springs are seasonal, waning or completely drying up in the late summer. There are several springs that continue to flow year-round and sustain enough water flow as to support the healthy trout and crayfish populations of Clover Creek and Deer Creek. The area is used by local fisherman and hunters whose game depends on the waters in these creeks.

The springs in the Davis Mountain quadrangle are mostly structurally controlled. Ground water flows along the faults, down the axes of synclinally deformed strata, and along unconformities. Many of the large springs are located at or very near large structures, or where several structures (faults, folds, and unconformities) intersect.

Many of the springs fed by ground water in fault zones are ephemeral. These are the low discharge, seasonal springs. This may be due to a much higher hydraulic conductivity in fault zones compared to sedimentary interbeds between the volcanic units. The fracture spaces associated with the faults may be less restricting than the unconformities allowing the water to drain out faster than it can be recharged and thus the springs dry up.

One of the most notable springs is located near the head of Deer Creek's east fork. The spring is fed by subsurface waters flowing along contacts of units that have been deformed tectonically into a west-dipping syncline. Along the axis of the syncline to the east is a tributary drainage. This drainage area is the head waters area for Deer Creek. The springs to the west in Deer Creek canyon proper generally dry up in the summer, but the east fork always has water flowing.

At the confluence of the east and main forks of Deer Creek is a landslide. This landslide is located where several faults come together, with movement enhanced by copious amounts of ground water in the area. Just to the south of the landslide deposit, are a multitude of springs. They seem to be fed from the contact zone of tuffs of Knob and Fir Grove.

Between many of the rock units is a very thin (< 1m thick) discontinuous layer of sediment composed mostly of sand-sized unconsolidated fluvial material and occasionally of pebbles and cobbles. These thin interbeds of coarse sediment are likely flow paths for ground water. They are potentially discordant bands or lenses of sediment deposited by juvenile streams draining the subdued topography of the ignimbrites. The thin sand and gravel lenses now have a regional south dip; ground water flows preferentially along them to the south.

Considering the thin and potentially discordant nature of these sediment interbeds, the flow path of the ground water could be quite complicated. For instance, if these sediment interbeds were formed by a meandering stream, then the path of the ground water could be quite tortuous.

Regardless of the nature of the fluvial environment that deposited the sediments, the modern streams cut through the sequence of rhyolite, basalt, and sediment, cutting these flow paths. The result is springs at contacts.

However, the units of the DMQ are generally tilted to the south toward the Snake River Plain. If the interbedded sediments extend to the south there is great potential for a deep ground water resource. The flow direction of these paloe-streams is unknown. If the streams were flowing east-west, the intercalated sediments may have east-west trends. As such they would not likely be well-connected with the aquifers of the SRP to the south.

Evidence for southward flow of ground water from Davis Mountain is the multitude of springs along the north side of Clover Creek and lack of springs on the south side of the canyon. This suggests some connectivity of the paleo-streambeds and a potential for ground water resources to the south fed from the MBH.

The Hill City-Bliss road crosses Clover Creek twice, once on the west side of the DMQ and again in the vicinity of White Arrow Hot Springs about six miles to the south. Throughout the late summer and fall the creek crossings are both dry, but the stretch of creek between these two crossings is always flowing. Clover Creek supports abundant aquatic wildlife which would not exist if the creek dried up in the summer. However the water does not flow on the surface to the Snake River at King Hill. It infiltrates the stream bed well before reaching the Snake River.

There may be two water flow paths, 1) water flows southward toward the town of Bliss in the subsurface at some undefined break between the crossings or 2) follow the stream bed in the subsurface toward the west to the Snake River. If the water follows the path of option 1) then the area around Bliss may have an untapped water supply beneath the basalt that is sourced at the top of the Mount Bennett Hills.

Stratigraphy

Detailed mapping of the Davis Mountain Quadrangle has revealed Eocene Challis Volcanic Group, 10 units of Miocene age previously mapped by Malde et al. (1963) as Idavada Group undivided and two post-Miocene basalt flows. I divided the Miocene units into 7 rhyolites, 2 basalts, and one lacustrine/fluvial sedimentary unit.

The oldest rhyolites are from the Bruneau-Jarbidge eruptive center (~11 Ma),. These units include the 11.21 Ma rhyolite of Deer Springs, the 11.17 Ma tuff of Fir Grove, and the 10.6 Ma tuff of Knob. They unconformably overlie Eocene Challis Volcanic Group. These relations can be seen in the northern part of the DMQ.

Unconformably overlying the three rhyolites is the basalt of Davis Mountain. This package of rocks (three rhyolites and a basalt) is cut by the Pole Creek fault trending northeast across the center of the map with down-to-the-southeast displacement.

In the northern map area these units are deformed by young normal, down-to-the-north, faults (deformation extending into or wholly within Quaternary time) creating step-like geomorphic features descending northward to the surface of the Camas Prairie.

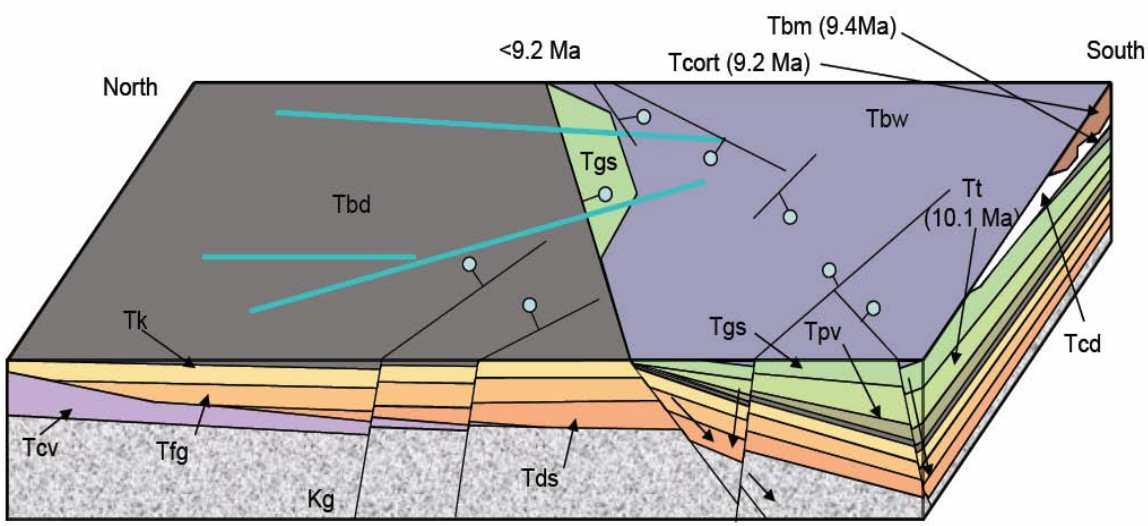

Figure 5.16 Generalized block diagram illustrating the relationship of the Bruneau-Jarbidge age rhyolites (orange and yellow) and Twin Falls age rhyolites (greens). B-J units include rhyolite of Deer Springs (Tds), Fir Grove Tuff (Tfg), and Tuff of Knob (Tk). Twin Falls units include the Pole Creek vitrophyre (Tpv), Tuff of Thorn Creek (Tt), Gwin Springs Tuff (Tgs), McHan Basalt (Tbm), Clover Creek Diatomite (Tcd), and tuff of City of Rocks. The basalt of Burnt Willow (Tbw) may be considerably younger than the Twin Falls volcanic field (11-9.2 ma) units.

Another younger series of rhyolite and basalt was emplaced after the phase of deformation associated with the Pole Creek fault. Unconformably overlying the basalt of Davis Mountain is the younger Twin Falls sequence, the age of which corresponds to the Twin Falls volcanic field (Armstrong et al., 1980; Honjo, 1990). (Note the term volcanic field is used in the most general lithostratigraphic sense, without inference about the geometry of the eruptive vents). The units (in green in Figure 5.16) terminate against the Pole Creek fault scarp and thicken dramatically to the southeast as seen in the Squaw Creek canyon. These units (vitrophyre of Pole Creek (Tpv), tuff of Thorn Creek (Tt), and tuff of Gwin Springs (Tgs)) can be seen onlapping the older Bruneau-Jarbidge sequence in the Deer Creek canyon. See Figure 5.16 for a diagrammatic illustration of this relationship. Deer Creek is represented by the western-most blue line.

Border Contact Issues with Smith (1966)

The onlapping relationship of the younger Twin Falls-aged rhyolites onto the older Bruneau-Jarbidge aged rhyolites is not obvious to the east in the Fir Grove Mountain and McHan Reservoir 7.5-minute quadrangles mapped by Smith (1966). The canyons that are transverse to the physiographic trend of the MBH have not eroded deeply enough to expose all units, so Smith could not see the onlapping relations of the tuff of Thorn Creek and Gwin Springs tuff over tuff of Knob and tuff of Fir Grove. The thinning and termination of the tuff of Thorn Creek to the north and discovery of Pole Creek vitrophyre are features of the MBH that Smith could not recognize due to lack of exposure. Here, I propose a new stratigraphic succession corrected from the work of Smith (1966). Figure 5.17 shows the stratigraphic relations presented by Smith (1966) compared to those proposed in this study.

The Pole Creek vitrophyre and basalt of Davis Mountain are important in understanding the relationship of the Bruneau-Jarbidge aged and Twin Falls aged units. Tpv thins to the east and north. The unit is not exposed in the East Clover Creek Canyon but can be found to the west in Squaw Creek canyon and Pole Creek with the basalt of Davis Mountain.

The basalt of Davis Mountain has been eroded away on the eastern side of Davis Mountain. This unit separates the Pole Creek vitrophyre from the tuff of Knob. Several erosional remnants of the basalt of Davis Mountain remain but the unit cannot be found east of the DMQ. The basalt of Davis Mountain was not recognized by Smith because it does not extend into his map area. Smith did not recognize the tuff of Knob as a separate unit because he did not have exposures of the basalt of Davis Mountain or vitrophyre of

Pole Creek. He instead interpreted the tuff of Knob as the northern-most extent of the very similar looking tuff of City of Rock. From this he erroneously mapped tuff of Fir Grove as tuff of Gwin Springs.

I found problems with correlating the tuff of Fir Grove into the DMQ. In the DMQ, the tuff of Fir Grove is stratigraphically several units below the tuff of Gwin Springs. Smith has the Fir Grove above the Gwin Springs. Elevations of the units would suggest this relationship only if the units were completely flat lying and not faulted or dipping down to the south.

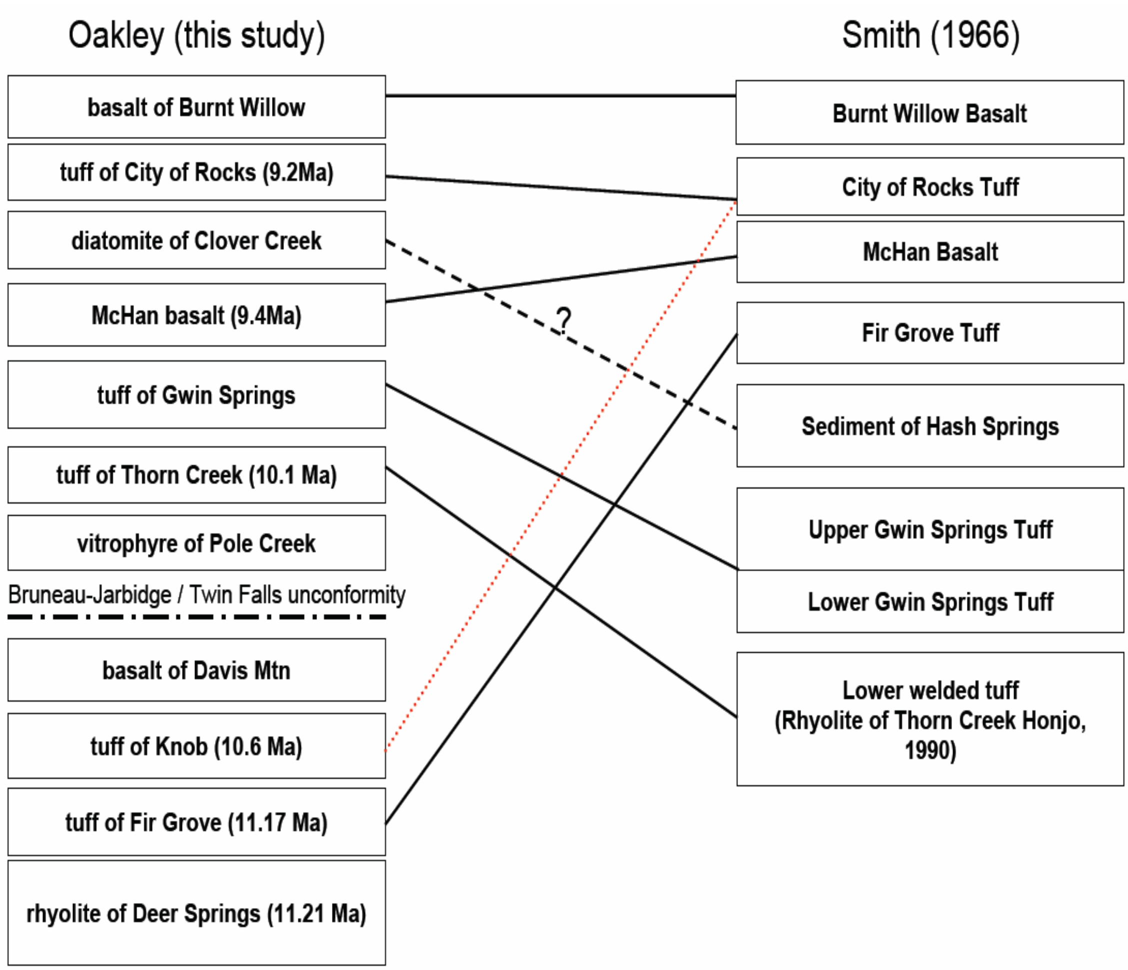

Figure 5.17. Stratigraphic correlation chart showing the order of units as presented by Smith (1966) and the revised order as seen in this study. Note that the units without tie-lines are units that were not emplaced, or are not exposed in the area mapped by Smith to the east of the Davis Mountain quadrangle. Dates are from Armstrong et al., (1980) and Honjo (1990) and this study.

I traced the tuff of Fir Grove to the type section at Fir Grove Mountain in the western portion of Smith's map area. The tuff of Fir Grove is a crystal-poor, reverse polarity rhyolite, (same paleomagnetic polarity reported by Wood and Gardner, 1984). P.K. Link and I measured the paleomagnetic pole at Fir Grove Mountain. We found it to be consistent with Wood and Gardner's analysis, which verifies Smith's identification of the tuff of Fir Grove at that location.

Link and I found a crystal-rich unit at the peak of Fir Grove Mountain stratigraphically above the tuff of Fir Grove. This is consistent with the relations in the DMQ as well. However in the DMQ the crystal-rich unit is the tuff of Knob (this can be demonstrated geochemically as well) not tuff of City of Rocks as mapped by Smith. Based on the relations observed in the DMQ, I suggest the crystal-rich unit above the tuff of Fir Grove at Fir Grove Mountain is actually tuff of Knob and was erroneously labeled tuff of City of Rocks by Smith.

In the western portion of Smith's field area, in the vicinity south of Fir Grove Mountain, he shows the tuff of Gwin Springs cut by a down-to-the-north, northeast-trending normal fault. This fault is shown juxtaposing tuff of City of Rocks on the north (hanging wall of the fault) with tuff of Gwin Springs on the south (footwall of the fault). Based on the relations of these units in the DMQ and erroneous labeling of the tuff of Knob as tuff of City of Rocks, I suspect this fault is of the same age as the down-to-the-south Pole Creek fault in the DMQ, if it exists at all. This would juxtapose tuff of Gwin Springs with tuff of Knob. It also obviates the need for Smith's inferred down-to-the-north normal fault.

Gwin Springs and Thorn Creek tuffs, stratigraphically above the vitrophyre of Pole Creek, both thin to the north and terminate against the Pole Creek fault scarp in the DMQ. The base of tuff of Thorn Creek is only seen in the vicinity of the Pole Creek where the unit thins enough to facilitate exhumation and exposure of the basal contact. In Deer Creek and Squaw Creek canyons south of Pole Creek, the tuff of Thorn Creek thickens and the base of the unit is not exposed; the lower contact dips below the surface.

The fault scarp of the Pole Creek fault is manifested in the outcrop pattern of tuff of Knob. Sufficient dissection at a high angle to the trend of the fault in the vicinity of Pole Creek (i.e. the cross-sectional view created by the Deer Creek canyon) allows for observation of the separation of tuff of Knob and tuff of Thorn Creek, two physically similar units. The two units are separated by the thin but distinct vitrophyre of Pole Creek.

As seen in Figure 5.17, there is a tentative correlation between Smith's Hash Spring formation and the Clover Creek diatomite. The stratigraphic relations are similar for both units. Smith's Hash Spring formation and the Clover Creek diatomite are both stratigraphically above the Gwin Springs tuff. The Hash Spring unit and the Clover Creek diatomite both pinch out northward against a paleo-highland of older units. However, Smith has the McHan basalt stratigraphically above the Hash Springs unit whereas I mapped the McHan basalt below the Clover Creek diatomite in the Catchall Creek canyon. This suggests that although the units are similar in lithologic nature they are two different units.

The Hash Springs formation and the diatomite of Clover Creek both consist of water-laden ash with coarse, well-rounded pebbles and cobbles of the Challis Volcanic

Group, Idavada Volcanic Group, and granitic lithologies. Smith suggests a fluvial dominated depositional environment for the hash Springs formation, whereas the Clover Creek diatomite is dominated by lacustrine deposits. The Hash Springs unit may include the deltaic or fluvial system which flowed westward into the paleo-Clover Creek lake. This hypothesis is supported by the occasional lenses and layers of siliciclastic material in the Clover Creek diatomite reported by Toth et al. 1987. The nature of the siliciclastic lenses, the directions that they pinch out or thicken, is however uncertain. If the lenses thinned to the west this would support a basin filled from the west, potentially by systems depositing the Hash Springs formation.

If the Hash Springs formation is not correlative with the Clover Creek diatomite then two similar lakes existed in the eastern MBH, one before 9.4 Ma and one after. The earlier lake would not have extended into the DMQ for there is no evidence of the lake beneath the 9.4 Ma McHan basalt. The latter lake (9.3 Ma) would have extended as far as the Clover Creek diatomite which covers an area of about 60 square kilometers. Much of the deposit is covered by basalt and so an accurate reconstruction is difficult.

Structural Geology

I divided the structural development of the DMQ into four phases based on trend of structures and the ages of the units affected. Most of the structures are vertical, down-to-the-south normal faults with 10-15 meters of displacement. I did, however, map three tectonic folds in the DMQ, one monocline dipping north into the Pole Creek fault, a syncline at the west end of Davis Mountain in the east fork of Deer Creek dipping away from the Davis Mountain fault, and a drape fold in the Burnt Willow basalt. There are other folds in the DMQ but they are small and difficult to recognize and trace. Several of the normal faults have over 30 meters of dip-slip motion. No oblique-slip was noted for any fault.

Figure 5.18. Exposures of discrete, planar fault surfaces are rare in the DMQ. This photograph is the only example found of such a feature. The magnitude of offset on this structure is unknown but it cuts through a lithic-rich Challis Volcanic Group rock. The structure strikes to the south dipping 60° to the west. Slickenlines trend toward the northwest plunging steeply. This may suggests the last movement of the fault was obliquely to the northwest.

Fault zones are recognized by abrupt changes in the orientation and elevation of the unit contacts. The internal foliation was not used to determine strike and dip of ignimbrites due to the lack of consistency of these measurements. Discrete planes with slicknesides and slickenlines are rare (shown in Figure 5.18). Fault zones are characterized by broad, rubbly bands with loose material covering the structure. Because most faults have less than 15 meters of offset, covered by a zone of rubbly material, they initially look like monoclinal folds.

Post-Eocene, pre-Late Miocene highland

At the head of Deer Creek, the oldest Idavada Group units are exposed. This includes the rhyolite of Deer Springs and Fir Grove tuff. The exposure of the older rhyolite of Deer Springs ends (the unit onlaps an irregular surface on the Challis Volcanic Group) however the exposures of younger Fir Grove tuff overlap Challis Volcanic Group rocks to the north. The rhyolite of Deer Springs terminates before it reaches the crest of the Challis Volcanic Group rocks.

The termination of the rhyolite of Deer Springs is interpreted to be a function of the volcanic flow onlapping a pre-existing highland composed of the Challis Volcanics and unexposed Idaho batholith. The Challis Volcanic Group can be seen to the north and the Idaho Batholith is inferred in the subsurface based on mapping in the adjacent regions by Malde et al. (1963) and Smith (1966).

The Challis Volcanic Group existed as a highland, either as a volcanic construct or as an uplifted block, with the rhyolite of Deer Springs wedging-out before the emplacement of the tuff of Fir Grove. The first phase deforms Fir Grove as well as tuff of Knob and the basalt of Davis Mountain; all the units with ages that overlap in time with activity of the Bruneau-Jarbidge volcanic field.

First Phase:
Pole Creek Fault System (Late Miocene, post-11.0 Ma, pre- 10.1 Ma)

The first phase of deformation is a series of en echelon, left-stepping, down-to-the-southeast normal faults. This is the most prominent structure in the DMQ. I named these faults the Pole Creek fault zone. Pole Creek and several other drainages follow the structure. This creates a trellis stream pattern (of Easterbrook, 1993) with the trunk streams following the trend of the Pole Creek fault (flowing over the less resistant rhyolites) and the tributary streams entering the trunk streams at high angles from the north (draining the surface of the more resistant basalt of Davis Mountain). This fault system is illustrated in Figure 5.19 as red lines.

Displacement along the fault zone varies but the most accurate place to measure the offset is in the vicinity of Squaw Creek. The upper contact of the basalt of Davis Mountain can be used as a marker for measuring the displacement. Dip-slip movement is about 120 meters.

Units emplaced after the basalt of Davis Mountain lap against this structure and are restricted to the area south of the fault zone

Second Phase:
Faulting prior to the basalt of Burnt Willow

The next phase of faulting deforms the younger onlapping Twin Falls volcanic field sequence (post 10.1 Ma). This second phase is dominated by northwest-trending faults with less-numerous conjugate northeast-trending faults. Figure 5.19 shows the location of these faults in orange.

The faults are distinguished from the other sets by cutting through the Pole Creek vitrophyre, tuff of Thorn Creek, tuff of Gwin Springs, McHan basalt, and Clover Creek diatomite. However, they do not cut the Burnt Willow basalt. The faults generally are near vertical planes trending 325° with down-to-the-north and down-to-the-south shear sense.

Displacement magnitudes range from 3 to 15 meters. Many of the faults are not mappable at 1:24,000-scale. Faults with offset less than 5 meters do not show mappable unit displacement and have been omitted from the map. Most of the "second phase" faults shown on the map have 10 meters of offset or more.

Third Phase:
Post-Burnt Willow basalt faults (Pre-10.1 Ma to Pliocene)

The third phase of faulting involves structures that offset the basalt of Burnt Willow. Some of these structures were active prior the emplacement of the Burnt Willow basalt; such as the one in section 15 that crosses East Clover Creek with down-to-the-north displacement. The faults displace the older units more than the younger units.

For example, the fault in section 15 (informally called the East Clover fault) offsets the Gwin Springs tuff nearly 75 meters, whereas the Burnt Willow basalt is only offset 35 meters. This discrepancy is evidence for recurring displacement along the fault, with deformation prior to, and after emplacement of the basalt of Burnt Willow.

Deformation of the Burnt Willow basalt is generally minor in magnitude. Low-magnitude displacements of the older units result in flexure of the basalt instead of offset of the basalt. This style of deformation, called drape folding, can be seen at faults with minor movement after the emplacement of the basalt flows, such as the west end of the East Clover fault.

The basalt of Burnt Willow does not change thickness across the fault. It was emplaced onto an even surface. The basalt does however change orientation in the vicinity of the fault. The Clover Creek diatomite is stratigraphically beneath the Burnt Willow basalt at the East Clover fault and any fault scarp that existed in the Diatomite could easily have been eroded flat.

The fault in East Clover Creek existed prior to the basalt of Burnt Willow as seen by the large displacement of the units in East Clover Creek that lie stratigraphically beneath the basalt. Offset of the basalt cannot be seen in Squaw Creek (to the west of East Clover Creek) along the trend of the fault in East Clover Creek. The displacement of the fault attenuated to the west.

Deformation of the Burnt Willow basalt along the East Clover fault can be seen up to the east edge of the Squaw Creek canyon. Deformation at the eastern-most extent of the structure is manifested as a dip-slope tilted down to the north, instead of fault displacement. There is no break or offset of the basalt at Squaw Creek but deformation in the form of tilting increases to the east. The trend of the dip-slope created by the basalt is parallel to and aligned with the East Clover fault, which actually offsets the basalt in East Clover Creek.

Evidence for recurrent faulting is quite common in the DMQ. Many of the units change thickness across faults. The fault in section 15 along East Clover Creek is the best exposure of such a relationship. This relationship suggests that the structures in this phase were part of an earlier phase, but continued to deform after the basalt of Burnt Willow was emplaced. Figure 5.19 displays these structures in green.

Fourth Phase:
Camas Prairie faults, post-9 Ma, continuing into Pleistocene

The final phase of deformation resulted in the formation of the Camas Prairie Rift. These structures, the Camas Prairie faults, are down-to-the-north normal faults, located near the crest of the MBH and northward. Displacement on these structures measures up to 150 meters (at Davis Mountain) and combine to displace the basalt of Davis Mountain nearly 450 meters within the DMQ. These faults displace Eocene- to Quaternary-aged units.

The Camas Prairie faults are dominantly east-west-trending, splintered (as defined by Easterbrook, 1993) normal faults. The fault segments generally break into two (or more) segments accommodating deformation over a broad area. This can be seen just to the north of Davis Mountain along the Knob fault.

These faults have steeply dipping or vertical orientations. The Davis Mountain fault dips at ~65° to the north and ties in with the vertical faults to the north as a detachment-type or master fault system. Evidence for the dip of the Davis Mountain fault is based on the trace of the fault over topography.

The age of this phase of faulting is constrained by the units it deforms. These faults are post-basalt of Davis Mountain (post-10.3Ma) but cut the Quaternary-aged basalt of Pothole (<100,000 k.y.).

The offset of the basalt of Pothole is minor compared with the displacement on the Knob and Davis Mountain faults. The basalt of Pothole is offset 4-5 meters in the

DMQ whereas the Knob and Davis Mountain faults have 10's to 100's of meters of offset. The offset of the basalt of Pothole may not be the maximum displacement of the fault. Only the western extent of the fault was examined in detail. Mapping of the Fir Grove Mountain Quadrangle to the east would give insight into the nature of the faults cutting the Pothole basalt.

The rate of deformation can be constrained using the magnitude of the offset (5 meters) of the basalt of Pothole and the age of the unit (<100,000 k.y.). The rate of deformation thus is approximately $1/20^{th}$ of millimeter per year. At this rate, deformation of the Davis Mountain fault (magnitude of 450 meters) began around 9 Ma. However, many of the faults show ~150 meter displacement suggesting inception around 3 Ma.

The fourth phase of deformation, the Camas Prairie faults, displaced some of the modern drainages. The canyons of East Clover Creek, Deer Creek, and the east fork of Deer Creek are beheaded canyons. Cluer and Cluer (1986) have noted that these (and other canyons of the MBH) have been cut-off on the north side. They suggest that the deformation that created the Camas Prairie Rift is manifested as the beheaded canyons. The canyons were initially cut in Late Miocene time by streams draining the Soldier Mountains to the north of the Camas Prairie, before the prairie existed. During deformation the prairie was down dropped, leaving the MBH as an asymmetric south-tilted highland with gentle slopes dipping to the south and steeper slopes (fault scarps) dipping to the north. The existence of these formerly through-going canyons is supported by the presence of Challis Volcanic Group and Idaho batholith clasts in sediments (Ts) of the MBH, and in the basal portion of the ~9.3 Ma Clover Creek diatomite at its northern extent.

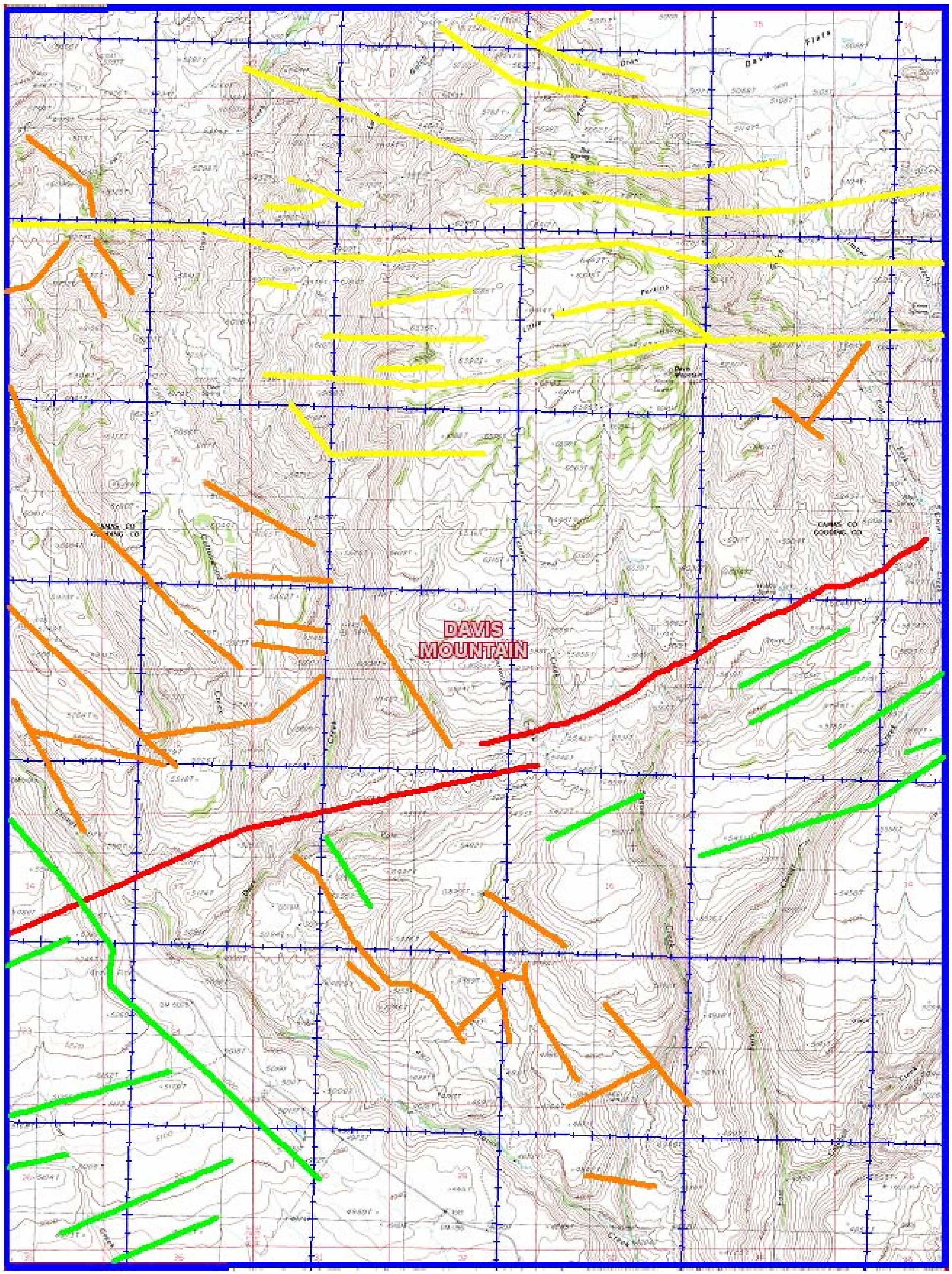

Figure 5.19. Topographic map showing the major faults of Davis Mountain. In red is the first phase of tectonic deformation, the Pole Creek fault. In orange are the faults of the second phase. In green is the third phase of faulting. The yellow faults in the north are the Camas Prairie faults, the fourth phase. These faults were active up into Quaternary time, but do not cut recent deposits of alluvial material. These structures are near vertical down-to-the-north normal faults.

Geologic Development of the Davis Mountain Quadrangle

The following discussion is a synopsis of the emplacement and deformation of the rocks in the vicinity of the Davis Mountain Quadrangle. Starting with the oldest known rocks of the area, I present my interpretation of the evolution of the DMQ and evidence that supports my assertions. The interpretations are accompanied by block diagrams illustrating pertinent features.

Cretaceous Idaho Batholith- Atlanta lobe

The oldest known rocks of the DMQ, granitic rocks of the Idaho Batholith Atlanta lobe (Lewis, 2001), do not crop out in the map area but can be found in adjoining quadrangles to the east and west and are thus inferred in the subsurface of the DMQ. U-Pb ages of detrital zircons from sediment deposits of streams draining the regions underlain by these granitic rocks presented by Link et al. (2005) are in accord with previous K-Ar dating (Lewis et al., 1993) and indicate a Late Cretaceous age of emplacement (70-110 Ma).

The granitic rocks were uplifted and eroded quite rapidly in Paleocene time, in the Hailey area (Jordan, 1994). The batholith locally intruded Devonian siltites and the overlying Pennsylvanian and Permian Dollarhide Formation of the Sun Valley Group (Worl et al., 1991).

Leeman et al., (1985) document Archean felsic gneiss xenoliths in the units of the Magic Reservoir Volcanic Center ~20 miles to the east of the DMQ. Leeman suggests felsic xenoliths may have been part of the country rock hosting the batholith and Tertiary volcanic rocks of the SRP.

Eocene-age intrusive rocks have been reported (Bennett and Knowles, 1985, Johnson et al., 1988; Lewis and Kiilsgaard, 1991) in the Solider Mountains north of the Camas Prairie (to the north of the DMQ). The rocks of the Eocene Challis Volcanic Group are in nonconformable contact with the granitic rocks in the MBH. Smith (1966) notes a sloping surface (up to 13°) of a "…deeply eroded granite."

Eocene Challis Volcanics

Unconformably overlying the granitic rocks are a series of Eocene 42-56 Ma (Fisher and Johnson, 1994; Worl et al., 1991; Link et al., 2005) intermediate volcanic lavas and sedimentary units, the Challis Volcanic Group. Vent areas for these rocks is unknown, but it is likely that they came from the north due to extensive exposures of CVG vents.

After the emplacement of the Challis Volcanic Group but before the emplacement of the Late Miocene rhyolite of Deer Springs (11.21 Ma) the Challis Volcanic Group was dissected. This erosional event is manifested by the irregular nature of the contact between the Challis Volcanic Group rocks and the overlying Idavada Group rhyolites.

The rhyolite of Deer Springs, the tuff of Fir Grove, and tuff of Knob onlap against the Challis Volcanic Group suggesting the Challis Volcanic Group exhibited topographic relief. The Fir Grove tuff has an irregular lower contact only when in contact with the Challis Volcanic Group. The tuff of Fir Grove/rhyolite of Deer Springs contact is flat. The Deer Springs/Challis contact is poorly exposed and no observations were made concerning the nature of the contact.

A highland existed in the north-central part of the DMQ where neither the rhyolite of Deer Springs nor Fir Grove tuff was deposited. At this location the tuff of Knob is in contact with the Challis Volcanic Group.

The Challis Volcanic Group does not have an irregular surface at every location in the DMQ. The upper surface of the Challis Volcanic Group is mostly flat with occasional irregularities. These irregularities may be either paleo-canyons or fault scarps. If they are stream valleys, they suggest that the dissection of the Challis Volcanic Group was fairly short-lived before the emplacement of the Idavada units, leaving large flat tracts bound by small steep-walled gullies. More detailed investigations of the Challis Volcanic Group in the MBH would provide answers to these questions.

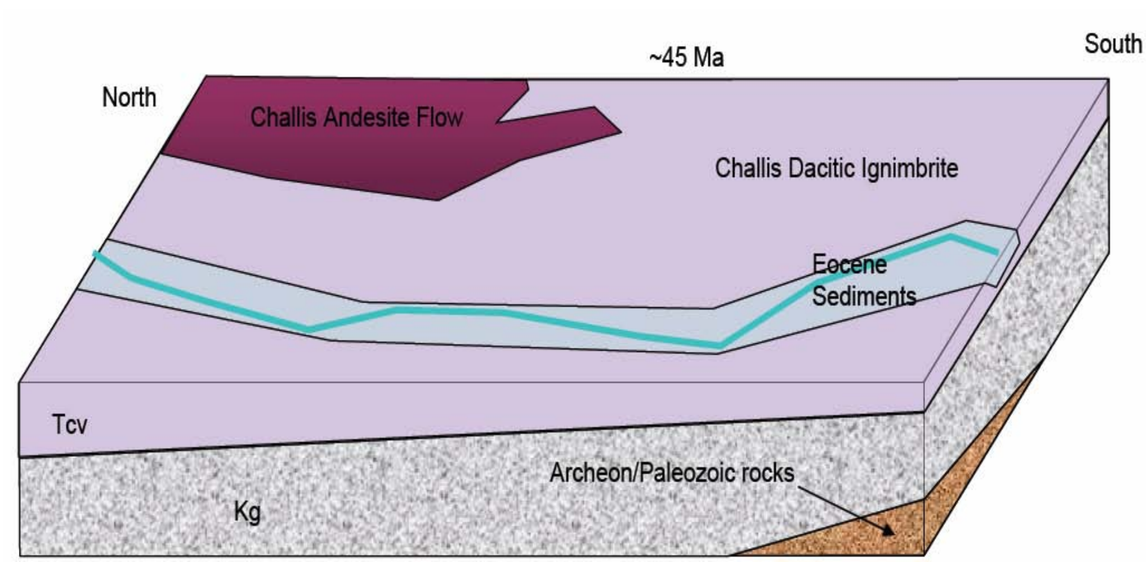

Figure 5.20. Block diagram 1 shows the Challis Volcanic Group (Tcv) overlying the granitic rocks of the Idaho Batholith. Archean/Paleozoic rocks represent the pre-batholith basement rocks.

Late Miocene Idavada Group:
Bruneau-Jarbidge Volcanic Field

Overlying the Challis Volcanic Group is a bimodal sequence of Miocene rhyolitic and basaltic volcanic units. These units mantle the irregular paleo-topography and fill in the lower topography of the Challis Volcanic Group resulting in more planar topography with each unit upward.

The first of the silicic units is the 11.21 Ma rhyolite of Deer Springs. Honjo (1990) suggested a tentative correlation with a unit (rhyolite of Windy Gap) recognized at Bennett Mountain by Wood and Gardner (1984). The rhyolite of Windy Gap was determined to be 11.0 Ma using the K-Ar method (Armstrong et al., 1980).

Several lines of evidence suggest the rhyolite of Deer Springs was erupted from the Bruneau-Jarbidge eruptive center of Bonnichsen and Kauffmann (1987): 1) its age of 11.21 Ma, and 2) its location stratigraphically below the “Pole Creek unconformity.”

The Bruneau-Jarbidge volcanic field is located to the southwest and kinematic indicators would be expected to suggest such a direction. However, the kinematic indicators suggest east-west flow directions. The indicators record the last movement of the flow and may not indicate the overall direction of flow, especially at the flow margins and over irregular topography.

The unit is only exposed in the Deer Creek canyons and thus little is known about its relationship with other units. It seems to terminate against the Challis Volcanic Group units and is overlain by the tuff of Fir Grove. Exposures in east Deer Creek suggest the unit was not deeply eroded. The upper surface of the unit is planar and uniform. The upper vitrophyre maintains constant thickness.

Following the emplacement of the rhyolite of Deer Creek a crystal-poor ignimbrite was emplaced, the 11.17 Ma tuff of Fir Grove. This unit can be found well to the north of the exposures of the rhyolite of Deer Springs. The Fir Grove tuff did not mantle all of the Challis Volcanic Group north of the termination of the rhyolite of Deer Springs. The stratigraphically higher tuff of Knob is in contact with the Challis volcanic rocks near the physiographic peak called Knob. The tuff of Fir Grove corresponds in age to Cougar Point tuff unit XI in the Bruneau-Jarbidge volcanic field (B. Bonnichsen, personal communication, 2006).

After the Fir Grove tuff was emplaced, streams deposited coarse sediment on the surface. These beds are discontinuous and highly transmissive in regard to ground water. Large springs flow year round were these sediments exist. The sediment beds are too thin to map at 1:24,000-scale but imply a period of time after the emplacement of the tuff of Fir Grove without silicic volcanic deposition.

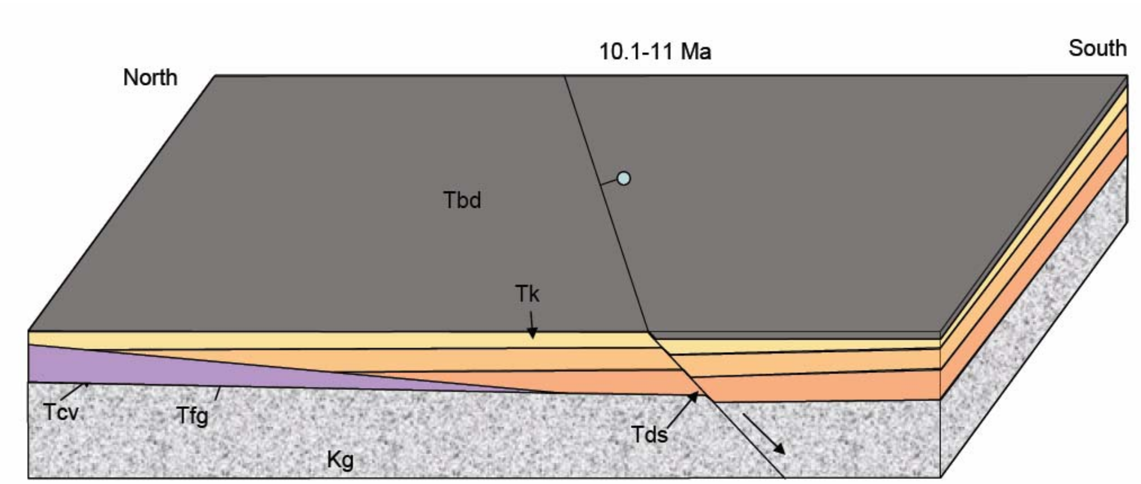

Figure 5.21 Block diagram 2 shows the relationship of the Bruneau-Jarbidge sequence of the Idavada Group: Tk, Tfg, and Tds; with the older Eocene Challis Group (Tcv) and Cretaceous granitic rocks (Kg). Note the way the rhyolite of Deer Springs (Tds) pinches out.

The next silicic unit is the 10.6 Ma tuff of Knob. It is the last of the Bruneau-Jarbidge age rhyolitic units to be deposited in the DMQ. This is a thick unit (over 100 meters in most places) that extends across the Camas Prairie to the north to the foothills of the Solider Mountains west of Hailey. It also exists in unconformable contact with Challis Volcanic Group rocks.

Following the deposition of the tuff of Knob, a thin layer of sandy sediment and a 5-25 meter thick basaltic lava flow (basalt of Davis Mountain) were deposited and emplaced. The basalt seems to thicken to south implying a southerly source or that a basin was forming to the south. The down-to-the-south faulting began on the Pole Creek fault system following the emplacement of the basalt of Davis Mountain. Figure 5.23 shows the general orientation of the fault and the units that are offset. The age of this faulting event is prior to 10.1 Ma. The youngest unit displaced by the structure is the basalt of Davis Mountain and it is offset by 120 meters.

Late Miocene Idavada Group:
Twin Falls Volcanic Field

The Twins Falls volcanic field is defined by 8-10 Ma Idavada Group units located near the central Snake River Plain and believed to have been derived from vents located in the vicinity of the city of Twin Falls (Pierce and Morgan, 1992; Perkins et al., 1995; McCurry et al., 1996). The next sequence of rocks includes: (from oldest to youngest) the vitrophyre of Pole Creek, tuff of Thorn Creek, tuff of Gwin Springs, McHan basalt, diatomite of Clover Creek, and tuff of City of Rocks.

The first unit to be emplaced after the initiation of the Pole Creek fault is the vitrophyre of Pole Creek. This unit is thin, less than 5 meters thick in exposures where the total thickness can be seen. In Deer Creek canyon this unit can be seen lapping up against the eroded Pole Creek fault scarp of tuff of Knob. In Squaw Creek, the vitrophyre of Pole Creek is very thin (~2 meters) and does not appear to reach the fault scarp.

This unit has stretched vesicles that indicate the last flow direction was northwest-southeast. These lineations project towards the proposed Twin Falls Volcanic Center of Pierce and Morgan (1992). Lineations in similar age rhyolites south of the central Snake River Plain project northward into the same region supporting the existence of the volcanic field (Hughes et al., 1996, McCurry et al. 1996). However this is only one data point and more detailed analysis of the directional fabrics is required before indicating a vent location.

The tuff of Thorn Creek (10.1 Ma; Armstrong et al., 1980) lies stratigraphically above the Pole Creek vitrophyre. It is a crystal-rich tuff very similar to tuff of Knob. The tuff of Thorn Creek thins dramatically to the west and to the north, terminating against the Pole Creek fault scarp. The dramatic thinning of the tuff of Thorn Creek indicates that the unit ends and did not accumulate an appreciable thickness to the north of the structure. There are no erosional remnants of the units on top of the basalt of Davis Mountain to the north of the Pole Creek Fault. In Squaw Creek, the upper and lower vitrophyres come together pinching out the internal devitrified zone.

Above the tuff of Thorn Creek is the tuff of Gwin Springs. This rhyolite also thins dramatically to the north and west, again suggesting that the eruption did not emplace an appreciable thickness north of the Pole Creek fault.

In Squaw Creek, the tuff of Gwin Springs is well exposed such that one can observe the general locations of the upper and lower contacts. Using an inclinometer, I measured the attitude of the upper and lower contacts. They differ by 1.5°, with the lower contact dipping 6.5° to the south. The upper contact dips 5° to the south. The disparity in the orientation of top and bottom of the unit suggests that the surface on which this unit was deposited dipped gently southward, and the emplacement of the unit filled in the lower portion. This relationship can be seen in Figure 5.24.

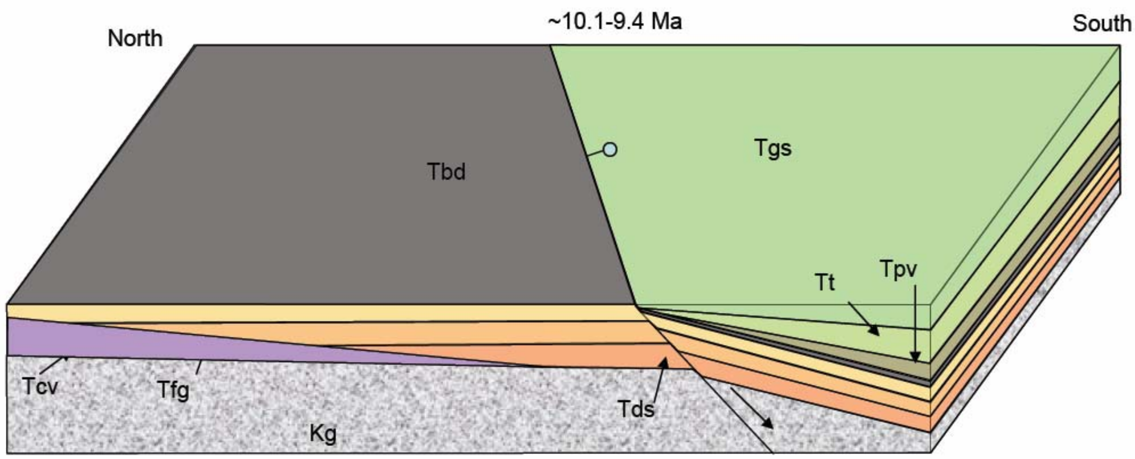

Figure 5.22 Block diagram 3 shows the onlap and thickening relationship of the green Twin Falls volcanic field units. Dips on the units are exaggerated for illustrative purposes.

After tilting of the Gwin Springs tuff, the McHan basalt (9.4 Ma, Honjo et al., 1986) was emplaced. Smith (1966) suggested a northerly source area due to thickening to the north. I interpret the McHan basalt in the DMQ to have flowed in from the east. The basalt wedges out to the west and to the north, therefore the unit is restricted to the south and east portions of the map. The spatial extent of the basalt implies a basin or lowland pathway to the east at 9.4 Ma. Figure 5.25 shows the relative spatial extent of the McHan basalt.

After the emplacement of the McHan basalt, down-to-the-north and down-to-the-south, northwest-trending faults propagated. These faults displace the McHan basalt but may have initiated before the basalt was emplaced and controlled the distribution of the basalt. Figure 5.25 displays the larger of these faults.

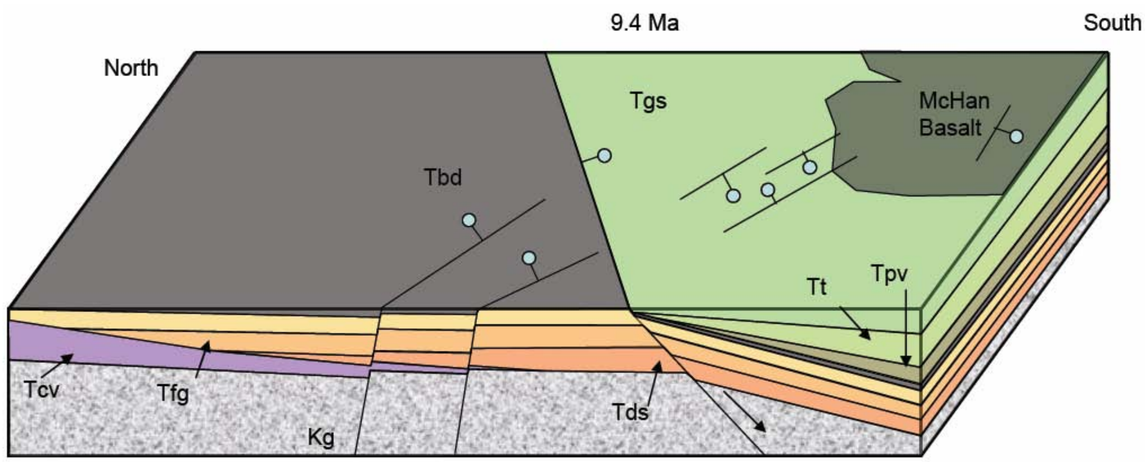

Figure 5.23 Block Diagram 4 showing the increased tilting of the Twin Falls-aged units, and the relative distribution of the McHan basalt. Normal faults (with ball and stick on the down-dropped side) initiated movement around the time of the emplacement of the McHan basalt and continued after basalt emplacement.

During or shortly after the emplacement of the McHan basalt, less then 200,000 years (based on K-Ar age constraints of Armstrong et al., 1980, and Honjo et al., 1986), a basin opened up facilitating the deposition of the diatomite of Clover Creek. The lower portion of the deposit is coarse clastic material, the basal fluvial overfilled basin phase of a half-graben (Lambiase, 1990). Deposited on top of the clastic material is a thick sequence of lacustrine sediments consisting of tuffaceous diatomite with limestone lenses. This represents Lambiase's under-filled phase of basin development that forms during rapid subsidence. The top of the unit is interpreted as the over-filled phase of Lambiase's model: defined by coarse clastic material deposited by fluvial processes.

The lake that occupied the basin was generally clear, clean water as evidenced by the purity of the diatomite (Dick Conroy, personal communication). The more clastic particles raining out of the water column the less pure a diatomite deposit will be.

The diatomite thins to the north and west in the DMQ. As it thins, the unit becomes increasingly clast-rich which supports the idea that the basin was being fed locally by streams from the north. South of the DMQ the basin appears to shallow, but more gently than it does to the north.

Smith (1966) has a unit, the Hash Springs formation, that is similar to (although out of stratigraphic sequence with) the Clover Creek diatomite. The Hash Springs is found in an east-west-trending belt across the center of Smith's map area. If the Hash Springs is the same unit as the Clover Creek diatomite then this implies an east-west-trending basin bordering the MBH at ~9.3 Ma (age constrained by K-Ar dating of the overlying tuff City of Rock and underlying McHan basalt by Honjo et al., 1986).

Following the deposition of the Clover Creek diatomite, there was a period of erosion. The Clover Creek diatomite was dissected and in places completely removed. This relationship can be seen on the east side of Catchall Creek. There, the tuff of City of Rocks is sitting directly on the McHan basalt. This suggests that the Clover Creek diatomite was completely removed at that location

Just to the south of that exposure in section 26 the City of Rocks tuff has an irregular lower contact with the Clover Creek diatomite implying that the diatomite had not been completely eroded at the time the City of Rocks tuff was emplaced. The lake basin may have been deeper, the unit was thicker to the south of the DMQ, or there is a meander of the paleo-canyon that was cut into the diatomite.

South of the DMQ boundary along the East Clover-Bray Lake road, near the old Chalk Mine and Hole-in-the-Wall quarries are great exposures of the contact between the City of Rocks tuff and Clover Creek diatomite. At these exposures the contact is quite tortuous and rather conspicuous, having juxtaposed the black lower vitrophyre of the City of Rocks and the white diatomite. The Clover Creek diatomite was partially eroded before the City of Rocks was emplaced as suggested by the uneven contact between the diatomite and the overlying tuff of City of Rocks.

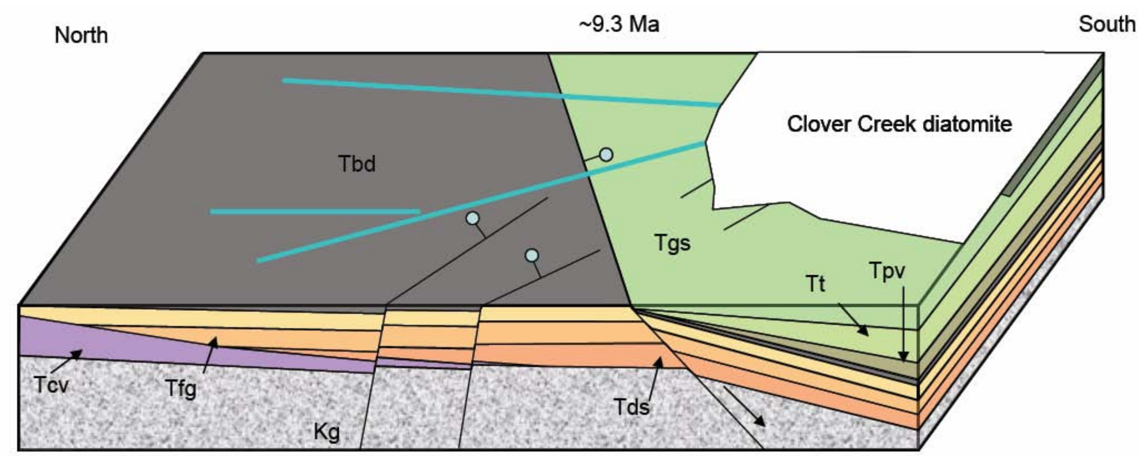

Figure 5.24 Block diagram 5 showing the distribution of the Clover Creek diatomite. The diatomite probably filled the north-tilted east-southeast trending basin that was beginning to form during the emplacement of McHan basalt. Following the deposition of the diatomaceous unit, the basin was uplifted (or the base level dropped) potentially due to the formation of the western Snake River Plain or the detumescence of the hotspot bulge, and the diatomite was eroded.

At 9.2 Ma the tuff of City of Rocks was deposited on the southern flank of the MBH highland created by the tilting and faulting of the Twin Falls age units. The rapid thinning and termination of the tuff of City of Rocks to the north and west suggests that the pyroclastic flow that deposited the tuff met a barrier that inhibited any further transport. This barrier is a manifestation of the uplift/base level drop that allowed for dissection of the diatomite. The barrier was a highland composed of the previously deposited and subsequently faulted and/or eroded diatomite of Clover Creek and underlying units. So the local basin filled with sediment, then was incised, and filled again but with rhyolite instead of sediments.

Exposures of the tuff of City of Rocks in the DMQ are limited to several square kilometers in the extreme southeast corner. At this location Catchall Creek flows along the westward termination of the tuff of City of Rocks. The tuff is located only on the east side of Catchall Creek and the west fork that splits off at the section line boundary for sections 26 and 23. This modern drainage is interpreted to mark a paleo-drainage cut into the diatomite of Clover Creek and filled in by the tuff of City of Rocks. The paleo-drainages that cut into the diatomite of Clover Creek must have extended north, because the sediments associated with the diatomite and adjacent to the tuff of City of Rocks contain clasts of Challis Volcanic Group rocks; the nearest outcrops are to the north. Also the City of Rocks terminated against a topographic barrier suggesting that the pyroclastic flow was beginning to move up hill(?). Therefore the drainages were flowing to the south.

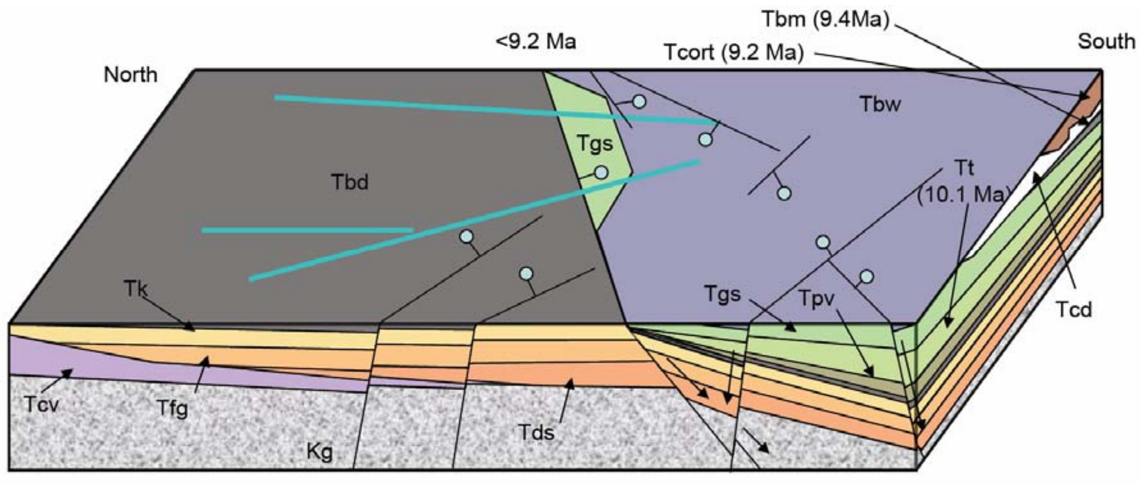

Figure 5.25 Block diagram 6 shows the erosion of the Clover Creek diatomite, emplacement of the City of Rocks tuff and the emplacement of the Burnt Willow basalt. Also shown are faults that have movement that post-dates the Burnt Willow basalt. The age of tuff of Thorn Creek is from Armstrong et al., 1980; McHan basalt and tuff of City of Rocks ages are from Honjo et al., 1986)

Basalt of Burnt Willow

Following the emplacement of the tuff of City of Rocks, the Burnt Willow basalt was erupted and emplaced. This channel-filling basalt flow varies greatly in thickness. Near the confluence of Deer and Clover creeks it is only several meters thick. At the landslides in sections 28 and 29 along Clover Creek, the basalt stands in vertical cliffs 30 meters tall. Along the Hill City-Bliss road just west of the DMQ the basalt filled in a paleo-Clover Creek channel. The paleo-channel trends to the northwest. The Burnt Willow basalt ends abruptly just past the west boundary of DMQ along the road. The paleo-channel was bounded by faults offsetting the basalt of Davis Mountain. This may be a similar channel to the modern Clover Creek. The basalt at that location has been dissected by the modern Clover Creek.

This unit has been affected by some of the late stages of deformation of the Snake River Plain. Faults that cut the Burnt Willow basalt only offset it by several meters. However the underlying units may be offset by several 10's of meters such as the fault in East Clover Creek mentioned in the discussion on structures. Movement along those faults continued after emplacement of the basalt.

The Burnt Willow basalt may be as young as Pliocene-age. It has primary volcanic features that are not preserved in the older basalts such as shelly and glassy pahoehoe ropes. These features are rare but the still exist in this unit. The surface of the basalt makes flat table-like features compared to the basalt of Davis Mountain. The basalt of Davis Mountain has a dissected hilly surface.

Faulted topography existed before the Burnt Willow basalt was emplaced. The paleo-channel has basalt of Davis Mountain at a higher elevation to the south. The peak on the south side of Clover Creek is basalt of Davis Mountain surrounded by the younger Burnt Willow basalt.

Following the emplacement of the basalt of Burnt Willow was a period of erosion. South-flowing canyons were incised prior to the formation of the Camas Prairie. These canyons now have been beheaded by the Camas Prairie structures. The canyons are dry 100+ meter-deep features that do not fit in the current environment. The drainage divide at the crest of the MBH should not be incised with large canyons unless these canyons existed prior to the rifting of the Camas Prairie. Modern runoff could not account for the erosion of these canyons. Conversely, the younger canyons such as Squaw Creek shallow and terminate at the crest of the MBH.

Cluer and Cluer (1986) suggest the canyons were cut by streams flowing south off the batholith north of the prairie and across the Idavada “volcanic construct,” to the axis of the proto-Snake River Plain.

In the beheaded Deer Creek canyon is a large landslide deposit, potentially caused by groundwater saturation where several structures come together. To the north of this landslide is a cobble-rich stream deposit composed mainly of Challis volcanic rocks and batholith clasts up to boulder-size. These clasts are very well-rounded in coarse, sandy matrix. This deposit is thought to be a manifestation of the rise in base level caused by the damming of the canyon from landslide. The stream loses energy as it is impeded by the newly fallen debris and drops any sediment load leaving the canyon mantled with sediment. The sediments are only found in and around the north side of the landslide in the main fork and the east fork of Deer Creek therefore both channels (the main and east forks) were fed from a northerly source prior to the Camas Prairie rifting events.

Age of the Camas Prairie Rift

The southern margin of the Camas Prairie Rift is controlled by down-to-the-north normal faults. These faults cut the Bruneau-Jarbidge age Idavada volcanic units and the young basalt of Pothole. Displacement on these faults is much greater in the Idavada Group than in the basalt of Pothole (100’s of meters versus 10’s of meters).

The Camas Prairie Rift deformation began sometime after 9.2 Ma (after 5.8 Ma based on Cluer and Cluer’s (1986) study) and ceased after the emplacement of the basalt of Pothole. Seismological studies of the western United States suggest that the Camas Prairie is aseismic (Pennington et al., 1974) No earthquakes have been recorded.

Cluer and Cluer suggest the 5.8 Ma inception of the rift based on dated rocks cut by the faults near Magic Reservoir at the eastern extent of the Camas Prairie. The Moonstone rhyolite was dated by Struhsacker et al., (1982) to be 5.8 Ma. Cluer and Cluer suggest this rhyolite is associated with rifting.

Camas Prairie:
Basalt of Pothole

The final rock unit to be deposited is the basalt of Pothole, shown in Figure 5.28 as Qb. This unit is thought to be Middle to Late Pleistocene in age. There are several reasons for this, most notably the exceptional preservation of the vent. The vent, called the Pothole (photo in Cluer and Cluer, 1986; page 96) is located just east of the DMQ's northeast corner. It is a crater-like depression with steep sides and is over 25 meters deep. The lava flow has delicate features such as shelly pahoeohoe preserved. A northwest-trending fault offsets the basalt and cuts directly through the vent.

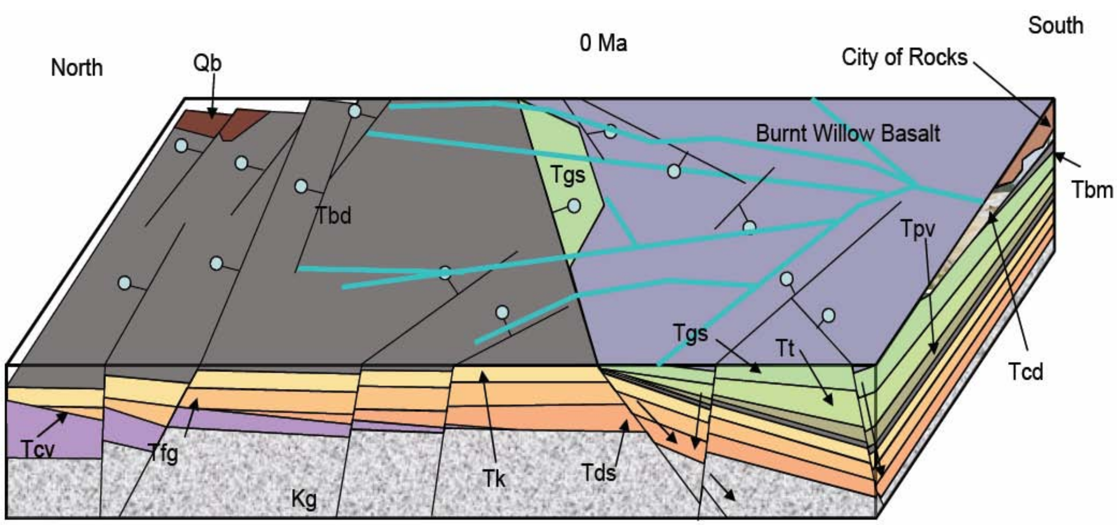

Figure 5.26 Block diagram 7 shows the current state of development in the DMQ. Note the streams have been cut short from Figure 5.25 . Also note the down-to-the-north east-west trending normal faults at the north end of the block. These are the faults controlling the formation of the Camas Prairie Rift (Cluer and Cluer, 1986).

The following is a sequential step-like description of the events that suggest the maximum age of the Camas Prairie rift.

1. Emplacement of Bruneau-Jarbidge age Idavada Group (including the basalt of Davis Mountain) ~11-10.5 Ma onto the Challis Volcanics.

2. Uplift or southward tilt of these units (Pole Creek fault) and inception of south flowing streams in minor transverse drainages (Deer and East Clover Creeks). Pre-10.1 Ma.

3. Emplacement of the Twin Falls-aged Idavada Group, 10.1 -9.2 Ma.

3.1. Emplacement of the Pole Creek vitrophyre, tuff of Thorn Creek, and McHan basalt.

3.2. Opening and filling of a basin dominated by a lacustrine facies (diatomite of Clover Creek) with minor fluvial facies from the transverse canyons post-9.4 Ma, pre-9.2 Ma. Basin is created by damming of paleo-canyon or controlled by local tectonics.

3.3. Drop in base level caused by tilting or uplift of the MBH, or erosion of the canyon dam. This causes the diatomite to erode out. Basin deposits are reworked and now dominated by fluvial facies. Fluvial deposits include Challis Volcanic Group rocks and batholith clasts from exposures to the north.

3.4. Emplacement of the tuff of City of Rocks at 9.2 Ma, preserving the paleo-canyon profile in the diatomite of Clover Creek. Paleo-canyons filled in by tuff and streams flow around the margins of the tuff of City of Rocks.

4. Emplacement of the Burnt Willow basalt. Pliocene

4.1. Burnt Willow basalt eroded by transverse flowing streams.

5. Transverse canyons cut-off and beheaded by Camas Prairie rifting, Pliocene and later.

The Burnt Willow basalt is cut by the transverse canyons. This is the youngest unit cut by the canyons and therefore the basalt existed before the canyons were cut. After the basalt was emplaced, the canyons were cut, and then beheaded by the inception of rifting. So the timing of inception of the Camas Prairie Rift can be placed after the emplacement of the Burnt Willow basalt during Pliocene time.

CHAPTER 6:
CONCLUSIONS AND SUMMARY

Conclusions

Eruptive Style of Rhyolitic Units

One of the questions this study set out to answer is: How were the rhyolites of the DMQ emplaced? The rhyolites were broken in to two textural groups: crystal-rich and crystal-poor. All of the crystal-poor units exhibit primary fragmental textures in thin section, indicating emplacement as ignimbrites rather than lava flows. Bubble-wall glass shards can be seen in various degrees of welding. These units include the tuff of Gwin Springs, vitrophyre of Pole Creek, and tuff of Fir Grove.

Determining the nature of emplacement for the crystal-rich units was much more difficult. Each of the crystal-rich units exhibit features common to both lava flows and ignimbrites. The rhyolite of Deer Springs is poorly exposed. The few outcrops that do exist are generally weathered and eroded such that the nature of emplacement cannot be determined. The tuffs of Knob, Thorn Creek, and City of Rocks exhibit features suggestive of both a lava flows and ignimbrites, however the features suggesting ignimbrites are more conclusive.

Considering that the evidence for a silicic lava flow has not been observed in a convincing nature, I suggest that all the rhyolitic units are ignimbrites. These appear to be distal facies of Twin Falls and Bruneau-Jarbidge volcanic fields.

Timing of Local Deformation

The Bruneau-Jarbidge age units (11.21-10.6 Ma) were emplaced against a highland of Challis Volcanic Group rocks as seen by their stratigraphic relationship in Plate 1. Between 11.0 and 10.1 Ma these units were uplifted and faulted (Pole Creek fault) with displacement down-to-the-south on a moderately dipping, northeast-trending fault. Following the inception of the Pole Creek fault, Twin Falls age rhyolites were emplaced and faulted synchronously. The offset on these faults is greater for older units (as seen east Clover Creek canyon). During the eruptions of the Twin Falls volcanic field, the emplaced units were being increasingly tilted to the south (albeit only 1.5°) after 10.1 Ma.

Further tilting is suggested by the spatial distribution of the McHan basalt at 9.4 Ma. The unit does not extend as far to the north as the previous rhyolite units. The lava flow was inhibited from flowing to the north by a tilted tuff of Gwin Springs surface.

Following the deposition of the McHan basalt a basin opens (or is created by the damming of a canyon), and is filled with lacustrine and fluvial sediments. The fluvial systems were draining highlands to north as evidenced by gravels that contain lithologies with source areas to the north.

Near 9.2 Ma the base level of the streams changed. The lake was uplift/tilted or the base level dropped (i.e. the SRP was deflecting or faulting downward). This is suggested by the erosion of the lacustrine sediments. After depositing the diatomite the base level dropped (or the canyon dam was breached) and the basin-filling sediment was partially removed. The gullies that were eroded into the diatomite were preserved by inundation of the tuff of City of Rocks at 9.2 Ma.

Following the emplacement of the tuff of City of Rocks, north-south rifting initiated, forming the Camas Prairie. Since 9.2 Ma, the Camas Prairie has been displaced downward at least 425 meters, based on the offset of the basalt of Davis Mountain. Displacement along Camas Prairie structures continued into Middle to Late Quaternary time, deforming the Quaternary basalt of Pothole in the northeast corner of the Davis Mountain quadrangle.

Future Research

Samples 05WL046-05WL051 were collected from the Bruneau-Jarbidge aged units for geochronologic analysis. Two samples were collected from three units (rhyolite of Deer Springs, tuff of Fir Grove, and tuff of Knob) in an attempt to constrain the age of the earliest stage of Idavada volcanism in the DMQ using Ar-Ar methods.

Dr. Dave Rodgers of Idaho State University has proposed to map Idavada aged rocks in the Lake Hills area in an effort to develop the structural history of the region between Picabo and Craters of the Moon National Monument.

Washington State University Master's student Will Starkell has begun geochemical work in the MBH in the vicinity of Bennett Mountain, approximately 50 km west of the DMQ. He is being assisted by Bill Bonnichsen, emeritus of the Idaho Geological Survey.

The Idaho Geological Survey is preparing to map the Fairfield 1:100,000 quadrangle which the DMQ is on the western edge.

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APPENDIX I: GEOCHEMISTRY

I-A SAMPLE LOCATION

Sample	Location	Northing	Easting	Unit	Rock	Analysis
04WL001	Davis Mtn	662530	4785750	Knob	rhyolite: upper vitrophyre	XRF/TS
04WL005	Davis Mtn	662980	4786070	Fir Grove	rhyolite: below lithophysae	XRF/ICPMS
04WL006	Davis Mtn	663490	4786260	Fir Grove	rhyolite: nonwelded top	XRF
04WL007	Davis Mtn	663660	4786290	Fir Grove	rhyolite: upper vitrophyre	XRF/TS
04WL012	Davis Mtn	665810	4787510	Knob	rhyolite: upper vitrophyre	XRF
04WL014	Davis Mtn	666460	4787640	Fir Grove	rhyolite: above lithophysae	XRF/ICPMS
04WL015	Davis Mtn	667060	4787380	basalt of Davis Mtn	basalt	XRF
04WL019	Davis Mtn	668830	4785580	basalt of Davis Mtn	basalt	XRF/TS
04WL021	Davis Mtn	670950	4786260	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL022	Davis Mtn	670820	4786250	Knob	rhyolite: lower devitr	XRF/ICPMS
04WL024	Davis Mtn	669720	4786200	basalt of Davis Mtn	basalt	XRF/ICPMS/TS
04WL025	Davis Mtn	670220	4786000	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL027	Davis Mtn	664700	4786150	Fir Grove	rhyolite: lower vitrophyre	XRF/ICPMS
04WL028	Davis Mtn	665280	4785880	Deer Springs	rhyolite: vitrophyre	XRF/ICPMS/TS
04WL030	Davis Mtn	666460	4787000	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL031	Davis Mtn	666460	4787000	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL032	Davis Mtn	666460	4787000	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL033	Davis Mtn	666460	4787000	Knob	rhyolite: lower vitrophyre	XRF/ICPMS
04WL034	Davis Mtn	666460	4787000	Knob	rhyolite: lower devitr	XRF/ICPMS
04WL035	Davis Mtn	666460	4787000	Knob	rhyolite: lower devitr	XRF/ICPMS
04WL036	Davis Mtn	666460	4787000	Knob	rhyolite: lower devitr	XRF/ICPMS
04WL037	Davis Mtn	666350	4786730	Fir Grove	rhyolite: vitrophyre	XRF/ICPMS/TS
04WL038	Davis Mtn	666350	4786730	Fir Grove	rhyolite: devitr	XRF/ICPMS
04WL039	Davis Mtn	663830	4787440	Fir Grove	rhyolite: devitr	XRF/ICPMS
04WL040	Davis Mtn	663830	4787440	Fir Grove	rhyolite: vitrophyre	XRF/ICPMS
05WL041	Davis Mtn	667855	4778286	Burnt Willow	basalt	XRF/ICPMS
05WL043	Davis Mtn	672100	4778830	City of Rocks	rhyolite: lower vitrophyre	XRF/ICPMS/TS
05WL044	Davis Mtn	672364	4779055	City of Rocks	rhyolite: upper vitrophyre	XRF/ICPMS/TS
05WL045	Davis Mtn	663798	4781427	Pole Creek	rhyolite: vitrophyre	XRF/ICPMS/TS
05WL052	Davis Mtn	667730	478760	Pole Creek	rhyolite: vitrophyre	XRF/ICPMS/TS

Sample	**Location**	**Northing**	**Easting**	**Unit**	**Rock**	**Analysis**
05WL053	Davis Mtn	666258	4779257	Gwin Springs	rhyolite: upper vitrophyre	XRF/ICPMS
05WL055	Davis Mtn	663024	4780807	Burnt Willow	basalt	XRF/ICPMS
05WL056	Davis Mtn	663342	4781069	Thorn Creek	rhyolite: devitr	XRF/ICPMS
05WL056A	Davis Mtn	663342	4781069	Thorn Creek	rhyolite: lower vitrophyre	XRF/ICPMS
05WL057	Davis Mtn	666767	4778600	Mchan	basalt	XRF/ICPMS/TS
05WL058	Davis Mtn	667947	4778700	Gwin Springs	rhyolite: lower devitr	XRF/ICPMS
05WL059	Davis Mtn	667501	4779059	Thorn Creek	rhyolite: upper vitrophyre	XRF/ICPMS
05WL060	Davis Mtn	669898	4782870	Knob	rhyolite: upper vitrophyre	XRF/ICPMS
05WL061	Davis Mtn	669648	4782534	basalt of Davis Mtn	basalt	XRF/ICPMS/TS
05WL062	Davis Mtn	668761	4783060	basalt of Davis Mtn	basalt	XRF/ICPMS
05WL063	Davis Mtn	672034	4789551	Pothole	basalt	XRF/ICPMS/TS
05WL064	Davis Mtn	663317	4781150	basalt of Davis Mtn	basalt	XRF/ICPMS/TS
05WL065	Davis Mtn	667610	4777032	Burnt Willow	basalt	XRF/ICPMS/TS
05WL066	Davis Mtn	669620	4776644	Gwin Springs	rhyolite: upper devitr	XRF/ICPMS
05WL067	Davis Mtn	669478	4776889	Thorn Creek	rhyolite: upper vitrophyre	XRF/ICPMS/TS
05WL068	Davis Mtn	669846	4777111	Mchan	basalt	XRF/ICPMS
121978 (04WL010)	Davis Mtn	664619	4786911	Challis Sediment	sediment	AA/ICPMS/TS

I-B MAJOR ELEMENT ANAYSIS BY SAMPLE

Sample	04WL001	04WL005	04WL006	04WL007	04WL012	04WL014	04WL015	04WL019	04WL021	04WL022	04WL024	04WL025	04WL027
Unit	Tk	Tfg	Tfg	Tfg	Tk	Tfg	Tbd	Tbd	Tk	Tk	Tbd	Tk	Tfg
SiO_2	71.15	76.71	74.88	76.14	71.66	76.49	49.99	49.96	71.80	71.77	49.32	71.58	76.19
TiO_2	0.60	0.28	0.30	0.28	0.61	0.33	1.48	1.32	0.64	0.64	1.47	0.68	0.30
Al_2O_3	13.20	12.66	13.35	12.05	12.88	12.70	15.99	16.22	13.57	13.62	16.20	13.44	12.03
FeO*	4.16	1.71	2.49	2.21	4.18	1.40	10.92	10.39	3.80	3.71	11.45	4.09	2.06
MnO	0.07	0.00	0.02	0.02	0.06	0.01	0.18	0.17	0.05	0.05	0.18	0.06	0.02
MgO	0.46	0.00	0.25	0.04	0.43	0.00	7.51	8.20	0.38	0.33	7.60	0.38	0.02
CaO	2.13	0.39	0.86	0.62	1.86	0.55	10.52	10.39	1.82	1.91	10.53	1.96	0.66
Na_2O	2.88	2.69	1.58	2.16	2.63	3.00	2.68	2.60	2.73	2.68	2.53	2.63	2.09
K_2O	5.23	5.54	6.24	6.45	5.57	5.47	0.55	0.57	5.08	5.22	0.53	5.08	6.61
P_2O_5	0.12	0.02	0.02	0.02	0.11	0.04	0.18	0.17	0.11	0.08	0.17	0.10	0.01
Original Total	98.37	99.78	96.43	97.79	98.72	100.15	99.79	99.99	99.94	99.87	99.90	99.88	98.98
LOI		0.93				0.53			1.63	2.13	0.28	2.22	3.15

I-B MAJOR ELEMENT ANAYSIS BY SAMPLE

Sample	04WL028	04WL030	04WL031	04WL032	04WL033	04WL034	04WL035	04WL036	04WL037	04WL038	04WL039	04WL040	05WL041
Unit	Tds	Tk	Tk	Tk	Tk	Tk	Tk	Tk	Tfg	Tfg	Tfg	Tfg	Tbw
SiO_2	74.55	71.55	71.16	71.38	71.18	72.05	71.60	71.96	76.06	75.99	75.02	75.71	45.60
TiO_2	0.35	0.65	0.64	0.65	0.70	0.66	0.69	0.67	0.32	0.32	0.40	0.38	2.94
Al_2O_3	13.06	13.45	13.63	13.46	13.63	13.57	13.66	13.39	12.43	12.53	13.12	12.57	16.35
FeO*	2.29	4.06	4.16	4.20	4.08	4.09	4.17	4.21	2.15	2.30	2.63	2.10	14.70
MnO	0.03	0.07	0.08	0.06	0.06	0.04	0.05	0.05	0.02	0.00	0.03	0.02	0.21
MgO	0.08	0.36	0.39	0.43	0.38	0.25	0.29	0.25	0.00	0.01	0.00	0.01	7.05
CaO	0.83	1.84	2.01	1.98	1.86	1.53	1.55	1.60	0.61	0.48	0.80	0.71	9.29
Na_2O	2.55	2.44	2.83	2.72	2.93	2.98	3.06	3.09	2.24	2.89	2.10	1.90	2.45
K_2O	6.23	5.47	4.96	4.99	5.05	4.75	4.80	4.66	6.13	5.47	5.90	6.58	0.84
P_2O_5	0.03	0.11	0.13	0.12	0.12	0.08	0.12	0.12	0.03	0.02	0.01	0.03	0.56
Original Total	99.56	99.45	99.81	100.00	100.10	99.74	99.52	99.98	99.85	99.51	99.99	99.89	99.89
LOI	3.28	2.41	2.48	2.26	1.87	0.79	0.83	0.64	3.04	1	2.79	3.14	0.31

I-B MAJOR ELEMENT ANAYSIS BY SAMPLE

Sample	05WL043	05WL044	05WL045	05WL052	05WL053	05WL055	05WL056	05WL056A	05WL057	05WL058	05WL059	05WL060	05WL061
Unit	Tcort	Tcort	Tgs	Tpv	Tgs	Tbw	Tt	Tt	Tbm	Tgs	Tt	Tk	Tbd
SiO_2	69.90	69.85	73.76	75.29	74.82	47.86	72.76	72.00	47.81	76.03	72.92	72.64	48.20
TiO_2	0.84	0.85	0.51	0.53	0.40	1.86	0.73	0.75	1.36	0.45	0.66	0.73	1.56
Al_2O_3	14.09	13.89	12.78	12.05	12.33	16.48	13.28	13.07	17.82	12.16	13.00	12.93	16.99
FeO*	4.03	4.34	3.03	2.97	2.96	11.93	3.75	4.13	12.05	2.48	3.93	3.76	12.22
MnO	0.06	0.06	0.05	0.03	0.05	0.18	0.05	0.05	0.17	0.01	0.02	0.05	0.18
MgO	0.79	0.77	1.18	0.29	0.16	8.42	0.66	0.47	7.74	0.09	0.20	0.39	7.66
CaO	2.39	2.26	0.17	1.10	1.06	10.79	1.23	1.88	10.46	0.76	1.67	1.49	10.56
Na_2O	3.15	3.03	2.72	2.18	2.65	1.94	3.05	2.46	2.03	3.01	3.12	2.19	2.04
K_2O	4.58	4.78	5.72	5.50	5.52	0.32	4.35	5.05	0.31	4.93	4.34	5.70	0.39
P_2O_5	0.16	0.18	0.08	0.06	0.05	0.24	0.13	0.14	0.26	0.06	0.14	0.13	0.20
Original Total	99.96	99.60	99.95	98.67	98.55	99.40	99.00	98.99	99.81	98.80	98.43	98.94	99.16
LOI	1.98	2.26	2.97	3.1	2.87	0.98	1.61	2.7	1.79	1.04	1.58	2.89	1.16

I-B MAJOR ELEMENT ANAYSIS BY SAMPLE

Sample	05WL062	05WL063	05WL064	05WL065	05WL066	05WL067	05WL068	121978 (04WL010)
Unit	Tbd	Qb	Tbd	Tbw	Tgs	Tt	Tbm	Tcvs
SiO_2	49.90	47.12	49.81	45.91	75.98	73.18	48.55	
TiO_2	1.37	3.25	1.41	2.80	0.44	0.57	1.46	0.1
Al_2O_3	16.37	14.52	16.86	16.12	12.23	12.75	17.42	5.51
FeO*	10.93	14.91	11.74	14.39	2.48	3.75	10.17	
MnO	0.16	0.21	0.16	0.20	0.00	0.06	0.15	0.57
MgO	7.49	6.76	6.62	7.37	0.12	0.33	8.20	
CaO	10.56	9.78	9.92	9.44	0.74	1.76	11.23	0.17
Na_2O	2.46	2.29	2.54	2.56	3.01	2.51	2.23	1.19
K_2O	0.58	0.65	0.73	0.72	4.92	4.98	0.35	1.73
P_2O_5	0.16	0.49	0.21	0.48	0.06	0.11	0.23	1.94
Original Total	99.47	99.81	99.61	100.00	98.51	98.32	99.45	
LOI	0.56	-0.3	1.11	-0.71	1.43	2.65	0.96	

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	04WL001	04WL005	04WL006	04WL007	04WL012	04WL014	04WL015	04WL019	04WL021
Unit	Tk	Tfg	Tfg	Tfg	Tk	Tfg	Tbd	Tbd	Tk
Ag		<1				<1			<1
As									
Au_{ppb}									
Ba	1280	709	658	708	1328	830	462	372	1205
Be									
Bi									
Cd									
Ce	146	137	181	189	149	165.5	34	38	150
Co		<0.5				0.5			4
Cr	5	20	2	2	5	10	315	380	50
Cs		3.1				3.6			3.2
Cu	7	10	7	9	8	<5	75	63	6
Dy		10.8				12.2			9.7
Er		6.4				7.9			6
Eu		1.4				1.5			2
Ga	25	22	25	20	22	23	20	19	21
Gd		13				13.6			11.4
Hf		14				15			14
Hg									
Ho		2.1				2.5			2
La	87	92.9	123	109	84	96.6	26	20	80.7
Lu		0.9				1.2			0.9
Mn									
Mo		4				4			4
Nb	42.1	50	62.8	55.1	43.7	51	8.6	7.8	39
Nd		73.7				71			59.6
Ni	5	<5	6	6	6	<5	119	140	<5
P									
Pb	26	24	37	36	27	29	5	4	21
Pr		21				20.3			16.3
Rb	172	210	231	241	179	207	10	11	167
S (wt%)									
Sb									
Sc	8		4	3	6		30	28	
Sm		14.1				13.8			11.6
Sn		8				7			6
Sr	132	29.8	41	34	117	37.5	288	295	116
Ta		3.4				3.4			2.5
Tb		2				2.1			1.7
Th	30	35	47	39	29	34	2	1	26
Tl		0.6				0.5			<0.5
Tm		0.9				1.1			0.8
U		9.1				8.3			6.5
V	22	8	3	3	18	8	235	215	24
W		3				3			3
Y	69	53.2	93	90	66	74.2	33	27	57.6
Yb		5.8				7.3			5.6
Zn	92	58	101	92	91	85	94	88	90
Zr	590	420	525	478	582	462	116	109	509

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	04WL022	04WL024	04WL025	04WL027	04WL028	04WL030	04WL031	04WL032	04WL033
Unit	Tk	Tbd	Tk	Tfg	Tds	Tk	Tk	Tk	Tk
Ag	1	<1	<1	<1	<1	<1	<1	<1	<1
As									
Au_{ppb}									
Ba	1185	342	1115	618	1460	1370	1135	1150	1155
Be									
Bi									
Cd									
Ce	152	31.1	147.5	181	159	155.5	147.5	151.5	147.5
Co	4.1	48.6	3.7	0.5	1.8	4.7	4.8	4.6	4.3
Cr	30	350	50	20	50	30	70	70	30
Cs	3.1	0.1	3.3	4.1	3.6	3.2	3.1	3	3.1
Cu	9	82	6	<5	22	7	7	7	6
Dy	10.1	4.4	9.9	12.2	9.5	10	9.6	10.1	9.6
Er	6.2	2.6	6.3	7.7	5.8	6.3	6.1	6.3	6
Eu	2	1.5	2	1.3	1.4	2	1.9	2	2
Ga	21	19	21	21	19	22	21	21	22
Gd	11.8	4.6	11.6	14.2	11.3	11.8	11.4	11.8	11
Hf	14	3	15	12	12	15	15	15	16
Hg									
Ho	2.1	0.9	2	2.5	1.9	2.1	2	2.1	2
La	83.7	14.9	77	103	86.2	82.5	77.5	80	77.7
Lu	0.9	0.4	0.9	1.1	0.9	0.9	0.9	0.9	0.9
Mn									
Mo	5	<2	4	5	5	5	5	6	4
Nb	40	9	39	46	43	41	39	40	40
Nd	63.5	18.1	58	75.7	60.5	61.6	58	60.7	57.4
Ni	<5	124	<5	<5	<5	<5	5	27	<5
P									
Pb	22	5	20	23	29	24	19	22	24
Pr	17.7	4.2	15.9	20.9	17.2	17.2	16	16.2	15.8
Rb	165	8.5	163	210	208	174	160	163	164.5
S (wt%)									
Sb									
Sc									
Sm	12.4	4.4	11.1	14.2	11.4	11.7	11.4	11.6	11.2
Sn	6	1	6	8	7	6	6	6	6
Sr	115.5	298	111	26.5	51.2	124.5	115.5	117.5	120.5
Ta	2.5	0.5	2.5	3.1	2.7	2.5	2.5	2.5	2.6
Tb	1.8	0.8	1.7	2.2	1.6	1.8	1.7	1.8	1.7
Th	26	1	26	33	31	27	26	26	26
Tl	<0.5	<0.5	<0.5	0.5	0.5	<0.5	<0.5	<0.5	0.5
Tm	0.9	0.3	0.9	1.1	0.8	0.9	0.9	0.9	0.8
U	6.5	<0.5	6.4	8.2	7.6	6.6	6.5	6.5	6.5
V	24	244	24	6	6	24	25	26	25
W	4	1	4	3	3	3	3	3	3
Y	60.7	25	58.2	72	57.1	61.2	57.4	59.8	56.7
Yb	5.8	2.2	5.8	7.1	5.3	5.8	5.6	5.7	5.6
Zn	90	107	87	82	98	95	88	91	94
Zr	507	104	511	368	404	549	521	538	566

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	04WL034	04WL035	04WL036	04WL037	04WL038	04WL039	04WL040	05WL041	05WL043
Unit	Tk	Tk	Tk	Tfg	Tfg	Tfg	Tfg	Tbw	Tcort
Ag	<1	<1	<1	<1	<1	<1	<1	<1	<1
As									
Au_{ppb}									
Ba	1205	1230	1255	593	626	823	760	565	1070
Be									
Bi									
Cd									
Ce	149.5	151.5	151	182	91.8	183	177.5	50.8	142.5
Co	3.7	3.8	4.2	0.5	0.5	0.9	0.5	57.3	6
Cr	40	30	40	20	20	20	10	130	30
Cs	2.8	3	3	4.3	3.3	4.4	4.3	0.2	2.8
Cu	6	6	6	<5	<5	<5	<5	34	8
Dy	9.6	10.2	9.9	12.1	11.4	11.5	11.8	7.2	9.4
Er	5.8	6.3	6.1	7.7	7.5	7.1	7.5	4.1	5.7
Eu	2	2	2.1	1.2	1.2	1.5	1.2	2.5	2.1
Ga	22	22	22	21	22	23	22	22	21
Gd	11.4	11.8	11.8	13.6	11.6	13.2	13	7.9	10.8
Hf	16	16	16	13	14	15	14	5	15
Hg									
Ho	1.9	2.1	2	2.5	2.4	2.3	2.4	1.4	1.9
La	83.5	82.9	83.7	96.9	80.6	97	93.3	24.4	74.7
Lu	0.9	0.9	0.9	1.1	1.1	1	1.1	0.6	0.8
Mn									
Mo	4	5	4	7	4	5	6	<2	4
Nb	42	41	41	48	50	51	48	21	42
Nd	62	63.4	63.2	70.7	63.5	71.3	68.2	31.5	57.2
Ni	5	<5	5	<5	5	<5	<5	117	6
P									
Pb	21	22	23	26	25	29	24	5	22
Pr	17.2	17.1	17.2	19.8	17.8	20	19.4	7.1	15.6
Rb	166.5	170	168.5	218	212	205	221	11.8	151
S (wt%)									
Sb									
Sc									
Sm	11.6	11.8	11.8	13.6	12.2	13.7	13.2	7.6	11
Sn	6	6	6	8	7	8	8	2	6
Sr	115	115.5	119	25.1	29.1	44.4	35.9	298	138.5
Ta	2.6	2.6	2.6	3.3	3.3	3.3	3.2	1.3	2.7
Tb	1.7	1.8	1.8	2.1	1.9	2	2	1.2	1.6
Th	27	27	27	34	36	34	33	1	24
Tl	<0.5	0.5	0.5	<0.5	0.5	0.6	0.5	<0.5	<0.5
Tm	0.8	0.9	0.8	1.1	1	1	1.1	0.6	0.8
U	6.8	7	6.9	8.4	8	7.7	8.1	<0.5	6.1
V	25	27	27	6	7	8	8	324	40
W	3	3	3	4	3	3	4	2	3
Y	54.5	59.8	55.8	70.4	69.2	64	69.2	39.1	53.7
Yb	5.5	5.6	5.5	7	7	6.7	6.9	3.6	5.2
Zn	88	91	97	85	87	89	83	163	81
Zr	547	541	563	396	417	478	432	186.5	519

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	05WL044	05WL045	05WL052	05WL053	05WL055	05WL056	05WL056A	05WL057	05WL058
Unit	Tcort	Tgs	Tpv	Tgs	Tbw	Tt	Tt	Tbm	Tgs
Ag	<1	<1	<1	<1	<1	<1	<1	<1	<1
As									
Au_{ppb}									
Ba	1045	1195	1200	1155	365	1185	1110	200	1240
Be									
Bi									
Cd									
Ce	140	172.5	173.5	176	33.1	147	155.5	24	158
Co	6.2	2.2	2.5	1.9	54.2	5	4.8	54.5	1.8
Cr	30	20	50	40	420	30	40	230	30
Cs	2.8	3.4	3.6	3.4	<0.1	2.3	2.8	<0.1	3.1
Cu	8	<5	6	<5	57	6	7	63	<5
Dy	9.5	11.2	11.2	11.7	5.3	10	10.2	5	10.9
Er	5.6	7	6.9	7.1	3.2	6	6.2	3.1	6.6
Eu	2	1.7	1.7	2	1.7	2.1	2	1.4	2
Ga	21	20	21	21	18	23	21	19	22
Gd	11	13	13	13	5.5	11.3	11.8	4.7	12.2
Hf	15	16	17	16	3	16	15	2	18
Hg									
Ho	1.9	2.3	2.2	2.4	1.1	2	2	1.1	2.2
La	74.4	90	91.2	89.9	19	80.2	82.2	12.4	88.4
Lu	0.8	1	1	1	0.4	0.8	0.8	0.4	0.9
Mn									
Mo	5	5	6	5	<2	4	5	<2	4
Nb	42	50	51	55	10	49	46	8	58
Nd	56.9	67	68.9	69.2	20.9	61.5	62.8	15.8	67.1
Ni	6	<5	10	<5	160	6	6	140	<5
P									
Pb	23	26	26	25	6	24	23	14	25
Pr	15.3	18.8	19	19.3	4.8	17	17.2	3.5	18.4
Rb	150	188.5	191	175	3.3	166	152	3.7	177.5
S (wt%)									
Sb									
Sc									
Sm	10.8	12.8	12.8	13.5	4.9	11.6	11.8	4.1	12.9
Sn	6	7	7	7	1	6	6	1	8
Sr	129.5	63.8	68	69	238	126	122.5	189.5	79.1
Ta	2.8	3.2	3.3	3.5	0.6	3.1	2.9	0.5	3.7
Tb	1.6	2	2	2	0.9	1.8	1.8	0.8	1.9
Th	24	31	32	28	1	26	25	<1	29
Tl	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Tm	0.8	1	1	1	0.4	0.8	0.9	0.4	0.9
U	6.2	7.6	7.9	7	<0.5	6.5	6.1	<0.5	7.3
V	44	9	11	6	295	27	28	256	9
W	3	4	3	4	2	3	3	2	3
Y	54	64.3	65.9	69.9	32.7	59.6	59.6	29.8	62.1
Yb	5.1	6.4	6.5	6.8	2.8	5.4	5.6	2.8	6.2
Zn	82	73	76	72	110	77	77	100	78
Zr	522	535	552	542	102.5	553	522	82.7	579

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	05WL059	05WL060	05WL061	05WL062	05WL063	05WL064	05WL065	05WL066	05WL067
Unit	Tt	Tk	Tbd	Tbd	Qb	Tbd	Tbw	Tgs	Tt
Ag	<1	<1	<1	<1	<1	<1	<1	<1	<1
As									
Au_{ppb}									
Ba	1370	1235	347	419	607	445	524	1140	1215
Be									
Bi									
Cd									
Ce	154	162	31.5	31.9	50.4	40.5	50.1	169.5	170.5
Co	3.9	5.4	54.4	49.1	51.1	47.5	58.6	1.5	3
Cr	50	30	390	410	200	260	140	20	30
Cs	2.7	3.3	<0.1	0.1	0.1	0.1	<0.1	2.9	3.2
Cu	11	9	81	67	27	55	38	<5	5
Dy	10.4	10.6	5.3	4.3	7.1	5	6.9	11.8	12.1
Er	6.3	6.5	3.1	2.5	4.1	3	4	7.4	7.3
Eu	2.2	1.7	1.6	1.4	2.6	1.5	2.4	2	2.5
Ga	22	20	20	19	22	19	21	21	23
Gd	12.4	11.9	5.5	4.5	7.7	5.1	7.6	13.2	13.3
Hf	16	17	3	3	6	3	5	17	17
Hg									
Ho	2.1	2.1	1.1	0.9	1.5	1.1	1.5	2.4	2.5
La	85	91.3	20.4	16.2	23.3	20.8	25.5	89.1	91.1
Lu	0.9	0.9	0.4	0.3	0.6	0.4	0.5	1	1
Mn									
Mo	5	5	<2	<2	2	<2	<2	4	5
Nb	47	46	12	9	18	11	20	55	52
Nd	66.1	65.1	23	18	30	21.2	30.8	69.5	70.9
Ni	6	6	156	132	78	126	119	<5	<5
P									
Pb	23	22	<5	22	10	9	14	23	25
Pr	18	18.2	5.5	4.2	6.7	5.1	7	19.1	19.4
Rb	150	175	4.7	9.8	8.6	11.8	9.7	172	171
S (wt%)									
Sb									
Sc									
Sm	12.7	12.3	5.2	4	7.2	4.7	7.1	13.4	13.8
Sn	5	6	1	1	1	1	1	6	7
Sr	155.5	101	256	319	351	297	313	75.5	128
Ta	2.9	3	0.7	0.5	1.2	0.6	1.2	3.6	3.3
Tb	1.9	1.9	0.9	0.7	1.3	0.9	1.2	2.1	2.1
Th	24	29	<1	1	1	1	<1	28	26
Tl	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Tm	0.9	0.9	0.4	0.3	0.6	0.4	0.6	1.1	1
U	6.5	7.2	<0.5	<0.5	<0.5	<0.5	<0.5	7.2	6.7
V	28	28	258	236	365	230	308	9	13
W	4	4	3	2	5	2	3	4	3
Y	63.9	64.4	30.3	25.5	40.2	29.5	40.3	69.2	70.4
Yb	5.9	6	2.8	2.2	3.5	2.7	3.4	6.9	6.5
Zn	78	73	104	98	172	104	150	71	89
Zr	522	565	121	112	222	124.5	183.5	538	550

I-C TRACE ELEMENT ANALYSIS BY SAMPLE

Sample	05WL068	121978 (04WL010)
Unit	Tbm	Tcvs
Ag	<1	<0.5
As		9
Au_{ppb}		<5
Ba	303	2230
Be		1.6
Bi		<2
Cd		<0.5
Ce	27	
Co	55.8	1
Cr	330	42
Cs	0.1	
Cu	55	1
Dy	4.9	
Er	3	
Eu	1.4	
Ga	18	
Gd	4.6	
Hf	2	
Hg		0.01
Ho	1.1	
La	13.2	
Lu	0.4	
Mn		124
Mo	<2	<1
Nb	9	
Nd	15.6	
Ni	154	3
P		310
Pb	<5	14
Pr	3.6	
Rb	5.5	
S (wt%)		0.02
Sb		<5
Sc		
Sm	3.8	
Sn	1	
Sr	206	447
Ta	0.5	
Tb	0.8	
Th	<1	
Tl	<0.5	
Tm	0.4	
U	<0.5	
V	266	16
W	3	<10
Y	29.4	
Yb	2.7	
Zn	98	22
Zr	88.6	

APPENDIX II: THIN SECTION SAMPLE ANALYSIS

APPENDIX II-A: THIN SECTION SAMPLE ANALYSIS: BASALT

		Textures				Phenocrysts				Opaques			
Sample #	Unit	Crystals	Matrix	Diktytaxitic		pheno % of rock	Plag An content	% plag	% olivine	Magnetite	illmenite	Amorphous	Comments
WL018	Tbd	porphyritic	intergranular	poorly	stellate plag w/oliv	7%	An_{39}	65%	35%	minor	minor	minor	Euhedral Plag 2-3mm; Sub-anhedral 1-2mm olivine altered red; glomerocrysts
WL019	Tbd	porphyritic	subophitic	poorly	stellate plag w/oliv	15%	An_{40}	60%	40%	rare	minor	minor	euhedral plag 2-3mm; subhedral olivines 1-2 mm altered red with spinel inclusions; homogeneous matrix; glomerocrysts
WL057	Tbm	aphyric	subophitic	strongly	weakly trachyitic	<1%	An_{38}	100%	0%			very common	An content from matrix microlites- 0.5-1 mm plag with interstitial opaques;
WL061	Tbd	porphyritic	intergranular	strongly	stellate plag w/oliv	8%	An_{38}	65%	35%	rare	rare	common	2-3mm plag laths with 1-2mm olivine altered red, glomerocrysts
WL063	Qb	porphyritic	subophitic	strongly	weakly with plag	12%	An_{28}	90%	10%		common	minor	2-3 mm lpag 1-2 mm subhedral olivine altered red; glomerocrysts
WL064	Tbd	porphyritic	intergranular	poorly	weakly with plag	20%	An_{30}	80%	20%	rare	minor	minor	8mm plag; 1-2mm eu-subhedral olivines altered red; glomerocrysts
WL065	Tbw	porphyritic	subophitic	poorly	stellate plag w/oliv	7%	An_{33}	90%	10%	common	common	common	2-4 mm bent? plag; 1-2mm eu-subhedral embayed olivines altered red;

APPENDIX II-B: THIN SECTION SAMPLE ANALYSIS: RHYOLITE

Sample #	Unit	%Total Phenocryst	Plag	San	Qtz	Fe-Aug	Pig	Opx	Fay/oliv	Magn	Ilm	Zircon	Apatite	Chev/Allan	Comments/ Textures
			Phenocrysts												
			x=major; 0=not present			% total pyroxene						x=present; 0=not present			
WL001	Tk	25	x, An_{31}	rare	minor	25	65	10	0	minor	0	x	x	0	Seived plag; many plucked grains
WL007	Tfg	2	x, An undet	minor	x	40	60	0	0	minor	minor	x	x	0	Shardy matrix; weird balls
WL022	Tk	20	x, An_{29}	minor	0	43	52	5	0	minor	0	x	x	0	minor zoning of plag; sieve plag; plag/pyxglomero; randomly oriented microlite matrix
WL028	Tds	15	x, An_{30}	x	x	50	50	0	rare	minor	minor	x	x	0	cut along foliation
Wl037	Tfg	2	x, An_{35}	x	x	60	40	0	0	minor	rare	x	x	0	welded shardy texture; granophyric plag and K-spar
WL043	Tcr	25	x, An_{32}	0	0	46	21	33	0	minor	0	x	x	0	bright red mineral(ppl); glassy matrix
WL044	Tcr	25	x, An_{30}	0	0	44	19	37	0	minor	0	x	xx	0	weaky crystallized matrix with aligned microlites
WL045	Tgs	5	x, An_{26}	rare	minor	33	40	27	0	minor	rare	x	x	0	welded shardy matrix; 3% accessory lithics
WL052	Tpv	10	x, An undet	0	minor	45	50	5	0	minor	minor	x	x	0	shardy matrix; seived plag & pig, weird balls; lithics; x'stal aggregates of plag pig aug
WL067	Tt	20	An_{21}	0	0	40	60	0	0	minor	0	x	x	0	large plag with some staining?; x'stal aggregates of pig and plag; matrix starting to devit-0.5mm spherulites?; zoned plag

APPENDIX III: PALOEMAGENTIC ANALYSIS

Sample #	Unit	UTM Location:	Northing	Easting	**Polarity**
1	**Knob**		662530	4785750	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	negative	negative	Inconclusive
4	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
2	**Knob**	05WL047	662760	4786002	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	negative	Inconclusive
2	negative	negative	Inconclusive
3	negative	positive	Normal
4	negative	positive	Normal
5	negative	positive	Normal
6	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
3	**basalt of Davis Mtn**		663045	4780807	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	positive	negative	Reverse
4	positive	negative	Reverse
5	negative	positive	Normal
6	negative	positive	Normal
7	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
4	**Deer Springs**		665280	4785880	**Reverse**

Measurement	Result	Reverse result	Polarity
1	Positive	negative	Reverse
2	Positive	positive	Inconclusive
3	Positive	negative	Reverse
4	Positive	negative	Reverse
5	Positive	negative	Reverse

Sample #	Unit	Location:	Northing	Easting	**Polarity**
5	**McHan basalt**		666767	4778600	**Reverse**

Measurement	Result	Reverse result	Polarity
1	Positive	negative	Reverse
2	Positive	negative	Reverse
3	Positive	negative	Reverse
4	Negative	positive	Normal
5	Positive	negative	Reverse

Sample #	Unit	Location:	Northing	Easting	**Polarity**
6	**Gwin Springs**		667947	4778700	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	negative	positive	Normal
4	negative	positive	Normal
5	negative	negative	Inconclusive

Sample #	Unit	Location:	Northing	Easting	**Polarity**
7	**Thorn Creek**		667501	4779059	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	negative	Inconclusive
3	negative	positive	Normal
4	negative	negative	Inconclusive
5	positive	negative	Reverse
6	negative	positive	Normal
7	positive	positive	Inconclusive

Sample #	Unit	Location:	Northing	Easting	**Polarity**
8	**basalt of Davis Mtn**		668761	4783060	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
9	**basalt of Davis Mtn**		672034	4789551	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
10	**basalt of Burnt Willow**		669648	4782534	**Normal**

Measurement	Result	Reverse result	Polarity
1	negative	positive	Normal
2	negative	positive	Normal
3	negative	positive	Normal

Sample #	Unit	Location:	Northing	Easting	**Polarity**
11	**Fir Grove**		665297	4786928	**Reverse**

Measurement	Result	Reverse result	Polarity
1	positive	negative	Reverse
2	positive	negative	Reverse
3	negative	positive	Normal
4	negative	positive	Normal
5	negative	positive	Normal
6	positive	negative	Reverse
7	positive	negative	Reverse
8	positive	negative	Reverse

Sample #	Unit	Location:	Northing	Easting	**Polarity**
12	**Fir Grove**	Fir Grove Mountain			**Reverse**

Measurement	Result	Reverse result	Polarity
1	positive	negative	Reverse
2	positive	negative	Reverse
3	positive	negative	Reverse
4	positive	negative	Reverse

APPENDIX IV: DETRITAL ZIRCON DATA

Summary of SHRIMP U-Pb zircon results for sample 1HGM99.

							Total Ratios				Radiogenic Ratios							Age (Ma)				
Grain spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{238}U/^{206}Pb$	±	$^{207}Pb/^{206}Pb$	±	$^{206}Pb/^{238}U$	±	$^{207}Pb/^{235}U$	±	$^{207}Pb/^{206}Pb$	±	ρ	$^{206}Pb/^{238}U$	±	$^{207}Pb/^{206}Pb$	±	% Disc
1.1	5643	4903	0.87	6.6	0.000321	0.87	732.6	10.3	0.0531	0.0012	0.0014	0.0000						8.7	0.1			
2.1	175	76	0.43	0.3	0.004890	4.50	587.1	17.3	0.0818	0.0066	0.0016	0.0001						10.5	0.3			
3.1	87	35	0.40	0.1	0.010111	17.27	518.2	18.5	0.1827	0.0302	0.0016	0.0001						10.3	0.7			
4.1	111	52	0.46	0.2	0.010341	13.79	606.2	21.2	0.1551	0.0150	0.0014	0.0001						9.2	0.4			
5.1	437	332	0.76	0.6	0.004592	6.21	625.0	13.2	0.0953	0.0075	0.0015	0.0000						9.7	0.2			
6.1	616	350	0.57	3.5	0.000670	0.96	149.8	2.1	0.0544	0.0014	0.0066	0.0001						42.5	0.6			
7.1	131	82	0.63	0.2	0.010702	13.24	634.0	21.2	0.1508	0.0147	0.0014	0.0001						8.8	0.4			
8.1	101	63	0.62	0.2	0.000213	6.50	432.0	14.1	0.0977	0.0157	0.0022	0.0001						13.9	0.6			
9.1	87	46	0.53	0.1	-	17.28	587.9	23.8	0.1827	0.0166	0.0014	0.0001						9.1	0.5			
10.1	158	61	0.38	0.2	0.003028	4.16	608.0	18.0	0.0791	0.0066	0.0016	0.0000						10.2	0.3			
11.1	114	50	0.44	0.2	0.012548	21.22	589.6	19.6	0.2139	0.0272	0.0013	0.0001						8.6	0.6			
12.1	122	56	0.46	0.2	0.004201	9.35	549.8	18.1	0.1201	0.0101	0.0016	0.0001						10.6	0.4			

Notes :

1. Uncertainties given at the one σ level.
2. Error in FC1 Reference zircon calibration was 1.37% for the analytical session.
   (not included in above errors but required when comparing $^{206}Pb/^{238}U$ data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. For areas older than ~800 Ma correction for common Pb made using the measured $^{204}Pb/^{206}Pb$ ratio.
5. For areas younger than ~800 Ma correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ratios
   following Tera and Wasserburg (1972) as outlined in Williams (1998).
6. For % Disc, 0% denotes a concordant analysis.

Summary of SHRIMP U-Pb zircon results for sample 01PL05.

G spot	U ppm	Th ppm	Th/U	^{206}Pb* ppm	^{204}Pb/^{206}Pb	f_{206} %	^{238}U/^{206}Pb	±	^{207}Pb/^{206}Pb	±	^{206}Pb/^{238}U	±	^{207}Pb/^{235}U	±	^{207}Pb/^{206}Pb	±	ρ	Age ^{206}Pb/^{238}U	±
1	459	233	0.51	3.0	0.001152	2.59	129.9	2.8	0.0675	0.0034	0.0075	0.0002						48.1	1.0
2	762	165	0.22	5.1	0.000453	1.32	128.7	2.3	0.0575	0.0023	0.0077	0.0001						49.2	0.9
3	505	592	1.17	3.2	0.000813	1.76	134.9	2.8	0.0609	0.0031	0.0073	0.0002						46.8	1.0
4	532	207	0.39	3.5	-	1.92	132.3	2.7	0.0621	0.0030	0.0074	0.0002						47.6	1.0
5	510	411	0.81	3.2	0.000485	1.87	138.8	2.9	0.0617	0.0033	0.0071	0.0002						45.4	1.0
6	1603	1278	0.80	10.6	-	<0.01	129.6	1.8	0.0464	0.0014	0.0077	0.0001						49.6	0.7
7	821	734	0.89	5.3	-	1.12	133.3	2.3	0.0559	0.0033	0.0074	0.0001						47.6	0.8
8	549	355	0.65	3.6	0.002869	4.36	130.7	2.5	0.0815	0.0031	0.0073	0.0001						47.0	0.9
9	1017	1380	1.36	6.7	-	1.08	130.6	2.1	0.0555	0.0019	0.0076	0.0001						48.6	0.8
10	953	844	0.89	6.3	0.001285	0.94	129.7	2.1	0.0544	0.0020	0.0076	0.0001						49.1	0.8
11	1189	677	0.57	8.1	0.001209	1.64	125.6	1.9	0.0600	0.0025	0.0078	0.0001						50.3	0.8
12	807	463	0.57	5.2	0.001813	0.51	132.2	2.5	0.0510	0.0021	0.0075	0.0001						48.3	0.9
13	180	101	0.56	1.2	0.005214	4.61	125.2	3.8	0.0835	0.0061	0.0076	0.0002						48.9	1.5
14	1716	1250	0.73	11.0	0.000107	0.42	133.6	1.9	0.0503	0.0017	0.0075	0.0001						47.9	0.7
15	1164	1558	1.34	7.6	0.000248	1.11	130.9	2.1	0.0558	0.0019	0.0076	0.0001						48.5	0.8
16	1076	525	0.49	7.3	0.000276	0.85	127.0	2.0	0.0537	0.0018	0.0078	0.0001						50.1	0.8
17	951	595	0.63	6.4	0.000470	1.25	127.7	2.1	0.0569	0.0020	0.0077	0.0001						49.6	0.8
18	389	170	0.44	2.5	0.002080	3.11	133.1	3.0	0.0716	0.0045	0.0073	0.0002						46.8	1.1
19	131	71	0.54	0.9	-	5.11	130.4	4.8	0.0874	0.0079	0.0073	0.0003						46.7	1.8
20	621	524	0.84	3.8	0.009382	1.30	138.6	2.7	0.0572	0.0035	0.0071	0.0001						45.7	0.9
21	832	685	0.82	22.9	0.001258	1.33	131.8	1.8	0.0575	0.0019	0.0075	0.0001						48.1	0.7
22	865	2059	2.38	15.5	0.001949	1.42	129.6	1.8	0.0582	0.0018	0.0076	0.0001						48.8	0.7
23	193	153	0.79	19.9	-	0.54	129.6	2.9	0.0512	0.0047	0.0077	0.0002						49.3	1.1
25	976	594	0.61	43.5	0.000821	0.52	128.8	1.7	0.0511	0.0020	0.0077	0.0001						49.6	0.7

26	802	671	0.84	1.8	0.006612	16.67	110.9	1.5	0.1790	0.0038	0.0075	0.0001						48.3	0.7
27	138	99	0.71	15.5	0.000002	<0.01	4.278	0.056	0.0917	0.0010	0.2338	0.0031	2.955	0.051	0.0917	0.0010	0.759	1354	16
28	164	83	0.51	16.9	0.003879	2.93	133.9	3.9	0.0702	0.0063	0.0072	0.0002						46.6	1.4
29	149	96	0.64	8.9	0.004272	2.96	129.7	3.1	0.0704	0.0047	0.0075	0.0002						48.1	1.2
30	805	343	0.43	8.6	0.000898	0.30	133.5	1.9	0.0494	0.0017	0.0075	0.0001						47.9	0.7
31	250	176	0.70	10.1	0.000427	0.90	128.3	2.5	0.0542	0.0046	0.0077	0.0002						49.6	1.0
32	876	1346	1.54	3.7	0.002669	3.54	129.8	1.8	0.0750	0.0021	0.0074	0.0001						47.7	0.7
33	532	353	0.66	16.4	0.010718	21.48	105.0	1.6	0.2171	0.0078	0.0075	0.0002						48.0	1.0
34	775	1527	1.97	3.7	-	0.58	131.5	1.8	0.0516	0.0026	0.0076	0.0001						48.6	0.7
35	1856	1061	0.57	14.0	0.000292	0.74	122.9	1.5	0.0529	0.0011	0.0081	0.0001						51.9	0.6
36	6849	3244	0.47	18.1	0.000142	0.61	112.0	1.2	0.0520	0.0005	0.0089	0.0001						57.0	0.6
37	1249	1022	0.82	71.3	0.000785	0.14	131.3	1.7	0.0481	0.0013	0.0076	0.0001						48.9	0.6
38	958	608	0.63	11.8	0.000605	1.01	125.8	1.6	0.0550	0.0015	0.0079	0.0001						50.5	0.7
39	2281	2718	1.19	10.2	0.000446	0.74	133.3	1.5	0.0528	0.0010	0.0074	0.0001						47.8	0.6
40	296	198	0.67	7.7	0.000980	1.70	128.5	2.4	0.0605	0.0030	0.0076	0.0001						49.1	0.9

Notes :

1. Uncertainties given at the one σ level.
2. Error in FC1 reference zircon calibration was 0.36% for the analytical session.
   (not included in above errors but required when comparing $^{206}Pb/^{238}U$ data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. For areas older than ~800 Ma correction for common Pb made using the measured $^{204}Pb/^{206}Pb$ ratio.
5. For areas younger than ~800 Ma correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ratios
   following Tera and Wasserburg (1972) as outlined in Williams (1998).
6. For % Disc, 0% denotes a concordant analysis.