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PART *1* ONE

EDITED BY PAUL KARL LINK AND BART J. KOWALLIS

V O L U M E 4 2 • 1 9 9 7

PROTEROZOIC TO RECENT STRATIGRAPHY, TECTONICS, AND VOLCANOLOGY, UTAH, NEVADA, SOUTHERN IDAHO AND CENTRAL MEXICO

Edited by
Paul Karl Link and Bart J. Kowallis

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Editor

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Cover photos taken by Paul Karl Link.

Top: Upheaval Dome, southeastern Utah.

Middle: Lake Bonneville shorelines west of Brigham City, Utah.

Bottom: Bryce Canyon National Park, Utah.

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Preface

Guidebooks have been part of the exploration of the American West since Oregon Trail days. Geologic guidebooks with maps and photographs are an especially graphic tool for school teachers, University classes, and visiting geologists to become familiar with the territory, the geologic issues and the available references.

It was in this spirit that we set out to compile this two-volume set of field trip descriptions for the Annual Meeting of the Geological Society of America in Salt Lake City in October 1997. We were seeking to produce a quality product, with fully peer-reviewed papers, and user-friendly field trip logs. We found we were bucking a tide in our profession which de-emphasizes guidebooks and paper products. If this tide continues we wish to be on record as producing "The Last Best Geologic Guidebook."

We thank all the authors who met our strict deadlines and contributed this outstanding set of papers. We hope this work will stand for years to come as a lasting introduction to the complex geology of the Colorado Plateau, Basin and Range, Wasatch Front, and Snake River Plain in the vicinity of Salt Lake City. Index maps to the field trips contained in each volume are on the back covers.

Part 1 "Proterozoic to Recent Stratigraphy, Tectonics and Volcanology: Utah, Nevada, Southern Idaho and Central Mexico" contains a number of papers of exceptional interest for their geologic synthesis. Part 2 "Mesozoic to Recent Geology of Utah" concentrates on the Colorado Plateau and the Wasatch Front.

Paul Link read all the papers and coordinated the review process. Bart Kowallis copy edited the manuscripts and coordinated the publication via Brigham Young University Geology Studies. We would like to thank all the reviewers, who were generally prompt and helpful in meeting our tight schedule. These included: Lee Allison, Genevieve Atwood, Gary Axen, Jim Beget, Myron Best, David Bice, Phyllis Camilleri, Marjorie Chan, Nick Christie-Blick, Gary Christenson, Dan Chure, Mary Droser, Ernie Duebendorfer, Tony Ekdale, Todd Ehlers, Ben Everitt, Geoff Freethy, Hugh Hurlow, Jim Garrison, Denny Geist, Jeff Geslin, Ron Greeley, Gus Gustason, Bill Hackett, Kimm Harty, Grant Heiken, Lehi Hintze, Peter Huntoon, Peter Isaacson, Jeff Keaton, Keith Ketner, Guy King, Mel Kuntz, Tim Lawton, Spencer Lucas, Lon McCarley, Meghan Miller, Gautam Mitra, Kathy Nichols, Robert Q. Oaks, Susan Olig, Jack Oviatt, Bill Perry, Andy Pulham, Dick Robison, Rube Ross, Rich Schweickert, Peter Sheehan, Norm Silberling, Dick Smith, Barry Solomon, K.O. Stanley, Kevin Stewart, Wanda Taylor, Glenn Thackray and Adolph Yonkee. In addition, we wish to thank all the dedicated workers at Brigham Young University Print Services and in the Department of Geology who contributed many long hours of work to these volumes.

Paul Karl Link and Bart J. Kowallis, Editors

Bimodal Basalt-Rhyolite Magmatism in the Central and Western Snake River Plain, Idaho and Oregon

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ABSTRACT

The purpose of this trip is to examine Miocene to Pleistocene basalt and rhyolite flows, ignimbrites and hypabyssal intrusions in a transect from the western Snake River Plain graben across the older part of the Snake River Plain "hot-spot-track." The earlier, dominantly explosive rhyolitic phase of volcanism will be examined primarily in the Cassia Mountains, near Twin Falls, Idaho. The second day of the field trip will focus on the Graveyard Point intrusion, a strongly differentiated diabase sill in easternmost Oregon. This late Tertiary sill is well exposed from floor to roof in sections up to 150 m thick, and is an example of the type of solidified shallow magma chamber that may be present beneath some Snake River Plain basalt volcanoes. The field trip will conclude with an examination of the diverse styles of effusive and explosive basaltic volcanism in the central and western Snake River Plain.

INTRODUCTION AND OVERVIEW

by Bill Bonnichsen and Martha Godchaux

The Snake River Plain (SRP) volcanic province, including the Yellowstone Plateau, extends from southwest to northeast across southern Idaho and adjoining parts of Oregon, Nevada, and Wyoming (Fig. 1). During the past 14 m.y. this elongate zone has been the site of extensive, bimodal, basalt-rhyolite volcanism. Although many exceptions exist, the general location of active volcanism has progressed from southwest to northeast along this zone and, at any particular place or time, the type of volcanism has changed from silicic ignimbrites to rhyolite lava flows to basalt lava flows. The major linear, time-transgressive component of the volcanism was the initiation of large-volume ignimbrite eruptions that swept from west to east. This

phase of silicic volcanism appears to have started in the Owyhee-Humboldt eruptive center near the junction of Idaho, Oregon, and Nevada (Bonnichsen and Kauffman, 1987), and perhaps at other sites to the north near the Idaho-Oregon state line. After the voluminous ignimbrite eruptions had ceased in a given portion of the SRP, widespread basaltic eruptions occurred and, even later in the history of the evolving SRP, sporadic, small-volume, silicic, dome-forming eruptions and additional basaltic eruptions occurred at various localities, but did not particularly follow the linear time-transgressive pattern shown by the ignimbritic eruptions from the large eruptive centers shown in Figure 1 (Bonnichsen et al., 1989).

The SRP volcanic province should be distinguished from the physiographic SRP. The volcanic province lies on either side of a nearly straight SW-NE-trending axis about 700

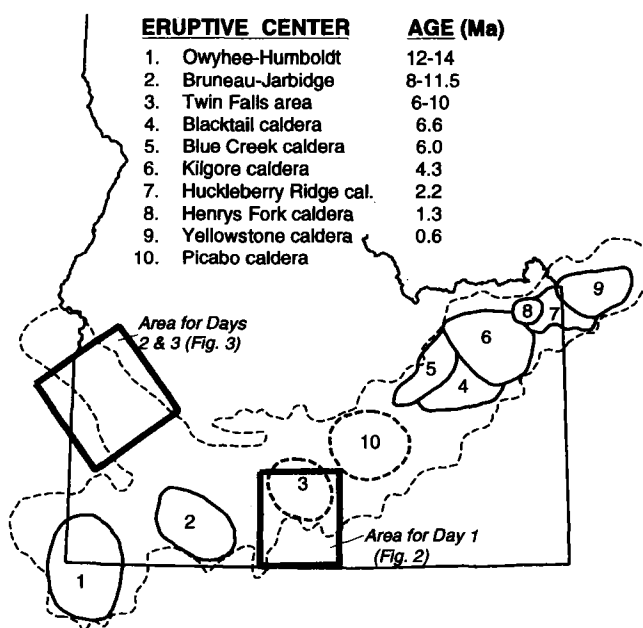


Figure 1. Map showing location of field trip areas in the central (day 1) and western (days 2 and 3) Snake River Plain, and the locations of the various eruptive centers and calderas in the SRP.

km long that trends from the Owyhee-Humboldt eruptive center near the Idaho-Oregon-Nevada junction to Yellowstone National Park in northwestern Wyoming (Fig. 1). Both topographically-subdued plains and plateau areas, and uplifted mountainous areas, are contained within the volcanic province. The physiographic SRP, on the other hand, includes the plains-and-plateau topographic zone stretching southwestward from eastern Idaho, then following the western SRP graben northwestward to the vicinity of the Idaho-Oregon state line, forming an arcuate zone partly within the volcanic province and partly to the north. Thus, in eastern Idaho the physiographic SRP coincides with the volcanic province, but in southwestern Idaho the physiographic SRP diverges northwestward from the axis of the volcanic province.

The SRP is, in effect, a boundary zone between differing tectonic provinces that would be encountered on east-west traverses at latitudes to the north and south of the SRP. Probably the most notable difference between a traverse south of the SRP and one to the north is the presence of the wide zone of tectonically-extended basin-and-range terrane to the south. The crust does not appear to be nearly as stretched by tectonic extension across central Idaho as in Nevada. The SRP volcanic province developed during and after the period of most active extension in the northern Basin and Range region and forms much of the northern margin of that large tectonic and physiographic province. Accordingly, the SRP volcanic province probably repre-

sents an accommodation zone that developed in the upper, brittle portion of the crust as the Basin and Range province was extended in an east-west direction (cf. Christiansen and McKee, 1978). During approximately the last 14 million years, this SW-NE-oriented tectonic accommodation zone, commonly known as the Yellowstone hot spot, was the site of the voluminous outpourings of rhyolite and basalt that accompanied the stretching and incipient fragmentation of the crust and upper mantle (Bonnichsen et al., 1989).

These observations are consistent with an interpretation that numerous batches of basaltic magma formed in the upper mantle and rose into the crust. This influx of basaltic magmas triggered the melting of large volumes of crustal material to form the rhyolitic magmas that erupted; subsequent basaltic eruptions occurred as the crust became more dense and rigid (Bonnichsen 1982b, Leeman 1982a).

Several million years ago the western SRP, which is a broad region northwest of the Twin Falls area, started sinking as the Earth's crust was stretched and thinned. The stretching direction generally was SW-NE, approximately parallel to the elongation of the SRP. This extension led to the formation of the somewhat complex western SRP graben over a period of several million years. Extensive silicic volcanism accompanied the development of the western SRP graben from about 12 to 10 Ma (Ekren et al., 1981, 1984). Following this were two major episodes of basaltic volcanism, the first from about 9 to 7 Ma, and the second from about 2.2 Ma to as recent as about 0.1 Ma (Othberg et al., 1995b). The crustal stretching created a topographic basin that forms the present western SRP region of Idaho and its extension into eastern Oregon. This basin previously held Lake Idaho, which at its maximum was about the size of today's Lake Ontario (e.g., Jenks and Bonnichsen, 1989). Lake Idaho was drained completely between 1 and 2 Ma ago, after Hells Canyon of the Snake River was cut. The terrane around the lake was fairly flat before it was drained, since the Snake River and its tributaries had not yet eroded into the volcanic rock layers. Canyon cutting started once the lake had drained; this has led to the formation of the narrow, deep canyons in SW Idaho and adjoining portions of Nevada and Oregon.

PLAN OF THE TRIP

The 1997 GSA field trip has been organized into two parts. The first day will be devoted to examining and discussing the silicic volcanism in the south-central part of the SRP, and the second and third days will be devoted to examining and discussing the results of basaltic volcanism in the western SRP graben, mainly where magma interacted with Lake Idaho or wet sediments left after the lake had drained. Accordingly, this field guide is organized into two

sections. The first part by Mike McCurry, Bill Bonnicksen, and Scott Hughes, discusses the rhyolitic units that were erupted from the Twin Falls eruptive center. Stops for Day 1 are enumerated on Figure 2. The second part, by Bill Bonnicksen, Craig White, and Martha Godchaux, discusses the basaltic volcanic and shallow plutonic units exposed in the western SRP graben. Stops for Days 2 and 3 are enumerated on Figure 3. For many of the stops discussed in this field-trip guide, additional information can be found in previously published field-trip guides by Bonnicksen et al., (1988, 1989) and Jenks and Bonnicksen (1989).

RHYOLITIC VOLCANISM, TWIN FALLS ERUPTIVE CENTER

by Mike McCurry, Bill Bonnicksen
and Scott S. Hughes

The Twin Falls Eruptive Center

The Twin Falls area lies within the SRP, a region of lava plains and steep canyons extending across southern Idaho. Many geologists think that the SRP formed as North America drifted southwestward over the Yellowstone hot spot (e.g., Pierce and Morgan, 1992; Smith and Braile, 1993). Leeman (1989) presents a critical summary of several alternate ideas. According to the hot-spot model, the hot zone is deeply rooted in the mantle; it has resulted in intense volcanism in southern Idaho for the past 14 Ma and now is situated under Yellowstone Park. As the region which became the SRP volcanic province passed over this hot zone, a series of large, generally circular, structural depressions were formed, from which copious volumes of silicic ignimbrites and lavas were erupted. These structural depressions have been referred to as eruptive centers (Bonnicksen, 1982a, Bonnicksen and Kauffman, 1987, and Pierce and Morgan, 1992), a complex of several intersecting caldera-like depressions that still are buried by their ejected volcanic products. These eruptive centers define the linear zone constituting the SRP volcanic province (Fig. 1). The existence of the Bruneau-Jarbridge eruptive center, located southwest of the proposed Twin Falls eruptive center, has been well documented on the basis of its structural features, volcanic products, and geophysical expression (Bonnicksen, 1982a). Likewise, the existence of the Owyhee-Humboldt eruptive center (Bonnicksen and Kauffman, 1987) seems to be on sound grounds, although it is not as well known, or the rocks in it as deeply exposed, as in the Bruneau-Jarbridge eruptive center. The existence of the proposed Twin Falls eruptive center (Pierce and Morgan, 1992) is even more speculative because of much more limited downcutting by rivers in the central SRP, in contrast to the situation farther west. However, given the voluminous quantities of silicic volcanic rocks both north

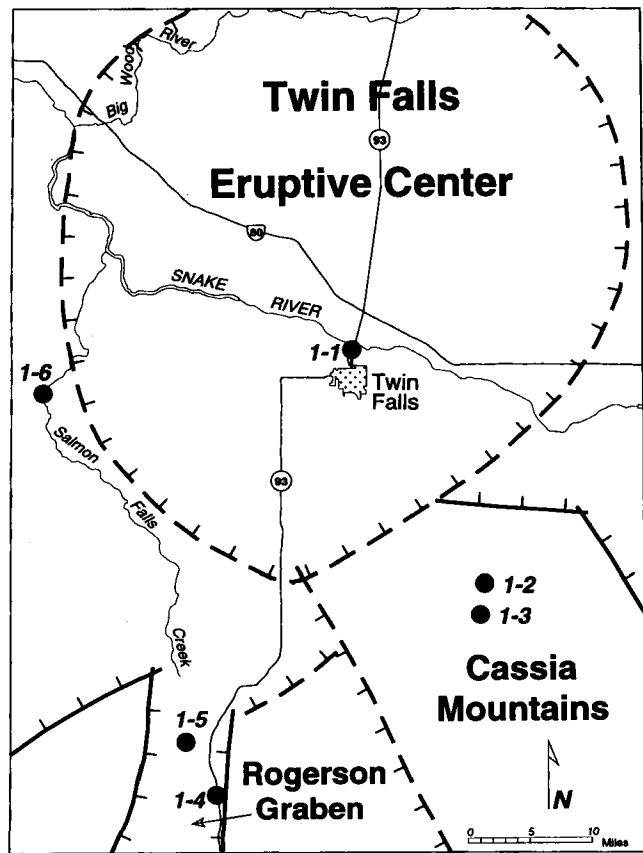


Figure 2. Map of south-central Idaho showing the locations of stops for day 1, and the approximate outline of the Twin Falls eruptive center.

and south of the central SRP, and the suggestion that appropriate structural features are present, it seems reasonable to believe that this center exists. This view is further supported by the thinning of the welded tuff sheets in the Cassia Mountains away from the SRP and the well developed flow lineations in that area that fan away from the central SRP (McCurry et al., 1996). Thus, we suggest that the circular feature outlined on Figure 2 is the approximate location of a complex of buried calderas that constitute the Twin Falls eruptive center, and that the large-volume ignimbrites that are magnificently exposed in the Cassia Mountains immediately south of the SRP were erupted from this zone.

Major ignimbrite volcanism principally occurred in the Twin Falls eruptive center from 10 to 8 Ma ago (Perkins et al., 1995) when this region passed over the hot spot. Basalt magma moved up from the Earth's mantle into the crust, where it melted and mixed with granitic crustal rocks and formed large buried chambers of molten rhyolite. Hundreds of cubic miles of this hot material erupted violently in repeated explosions many times more powerful than the

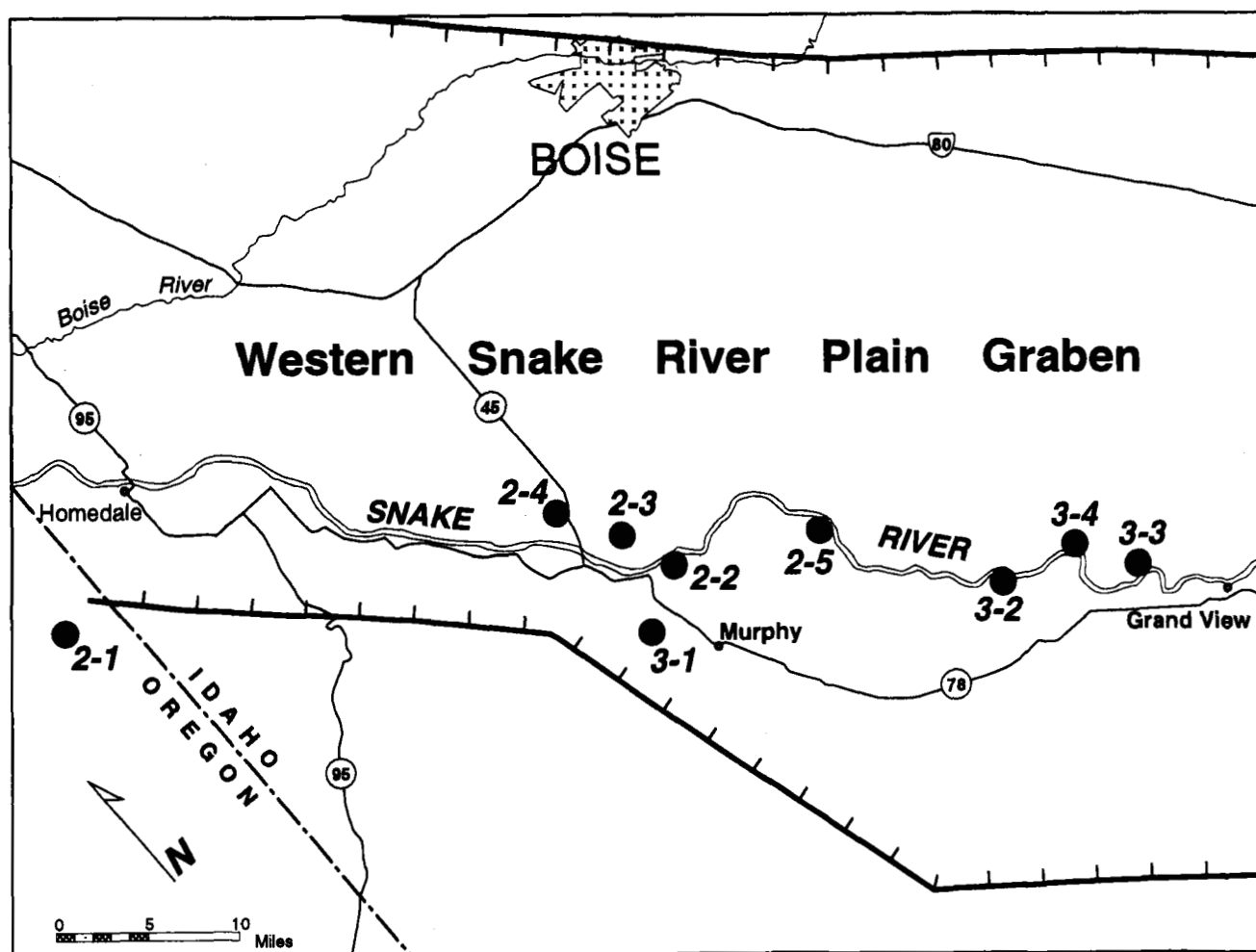


Figure 3. Map of the western Snake River Plain area showing locations of stops for days 2 and 3, and the boundaries of the western SRP graben.

1980 Mount St. Helens explosion. Each eruption spread a layer of welded tuff over hundreds of square miles and excavated a caldera that later was filled by lava flows. At least seven successive sheets of welded tuff, which together measure up to 1,300 feet thick, were laid down in the hills a few miles south of Twin Falls (Mytton et al., 1990; Williams et al., 1990 and 1991; McCurry et al., 1996). After the large-volume ignimbrites, rhyolite lava flows partially filled the structural depression that had developed. An example of this type of lava flow is the Shoshone Falls rhyolite. It is several hundred feet thick and can be traced for about 6 miles along the bottom of Snake River canyon near the city of Twin Falls.

After the rhyolitic volcanism ended in the Twin Falls eruptive center about 6 Ma ago (Armstrong et al., 1975), basaltic lava from deeper in the earth started to erupt onto the surface. These basalt eruptions were not as explosive

or voluminous as the earlier rhyolite eruptions; thus, calderas were not formed. Instead, most of the basalt issued from fractures to form broad, low-profile, shield volcanoes and lava sheets that flowed away from the shields. Such basaltic volcanism occurred episodically in the central SRP up to a few thousand years ago, and may occur again. Shield volcanoes visible east and south of Twin Falls include Skeleton, Hansen, Stricker, and Hub Buttes. The flows from these volcanoes appear in the walls of Snake River canyon. Shields southwest and west of Twin Falls are Berger Butte, others farther south, and two near Castleford. Many shields are prominent north of the Snake River. Interestingly, many of the shields south of the Snake River in the general Twin Falls region erupted ferrobasalt, whereas those north of the river erupted SRP olivine tholeiite, less rich in Fe and Ti. Also, the shields north of the river generally seem to be younger than those south of the river.

Neogene Volcanic Stratigraphy of the Cassia Mountains

Cassia Mountains are dominated by a layer-cake-like sequence of regionally distributed, high-grade (Branney and Kokelaar, 1992), silicic ignimbrite sheets interbedded with precursor fallout, coignimbrite ash, and reworked tephra (Williams et al., 1990; Struhsacker et al., 1983; Perkins et al., 1995; Hackett et al., 1989; McCurry et al., 1996). Ages and stratigraphic features are illustrated in Figure 4. Fourteen ignimbrites are identified. Most are simple cooling units, but there are also compound and composite cooling units present (e.g., members of tuffs of McMullen Creek and Steer Basin). Eight of the fourteen are grouped by Williams et al., (1990, 1991) and Mytton et al., (1990) into four major stratigraphic units: tuff of Big Bluff; tuff of Steer Basin; tuff of Wooden Shoe Butte; tuff of McMullen Creek. These four units dominate the volcanic section across most of the Cassia Mountains, reaching a maximum cumulative thickness of at least 400 meters in Rock Creek canyon (Williams et al., 1991).

Tuff of Big Bluff, lowermost of the four major units, is distinguished from the others by abundant quartz and alkali feldspar phenocrysts, as well as containing phenocrysts of plagioclase, pyroxenes and Fe-Ti oxides (Williams et al., 1990). On the basis of its distinctive mineral assemblage and age, Perkins et al., (1995) proposed a correlation between this unit and Cougar Point Tuff unit XIII, an ignimbrite produced by the Bruneau-Jarbridge eruptive center (Bonnichsen and Citron, 1982). This idea is supported by the occurrence of east trending pyroclastic-flow emplacement lineations documented by McCurry et al., (1996).

The upper three units, tuffs of Steer Basin, Wooden Shoe Butte and McMullen Creek, are remarkably similar to the tuff of Big Bluff in most of their volcanological features, and in their bulk major and trace element compositions (Hughes et al., 1996) and Nd-, Sr- and Pb-isotopic composition (McCurry et al., 1995). However they are distinguished by flow emplacement lineations suggesting a source to the north (Fig. 5; McCurry et al., 1996) rather from the west or east, as suggested by Hackett et al., (1989), and by their very low concentrations or absence of quartz and alkali feldspar phenocrysts. They are almost indistinguishable from each other in phenocryst abundance, assemblage and bulk chemical composition (Hughes et al., 1996, Parker et al., 1996, and Watkins et al., 1995, 1996).

All of the ignimbrites are interbedded with fallout vitric tuffs and, in some cases, water reworked tephra, a principal basis for subdividing the units by Williams et al., (1990) and Mytton et al., (1990). Several paleosol horizons are splendidly preserved, and examples can be seen at the Steer Basin Campground along Rock Creek (Stop 1-3 of this field guide).

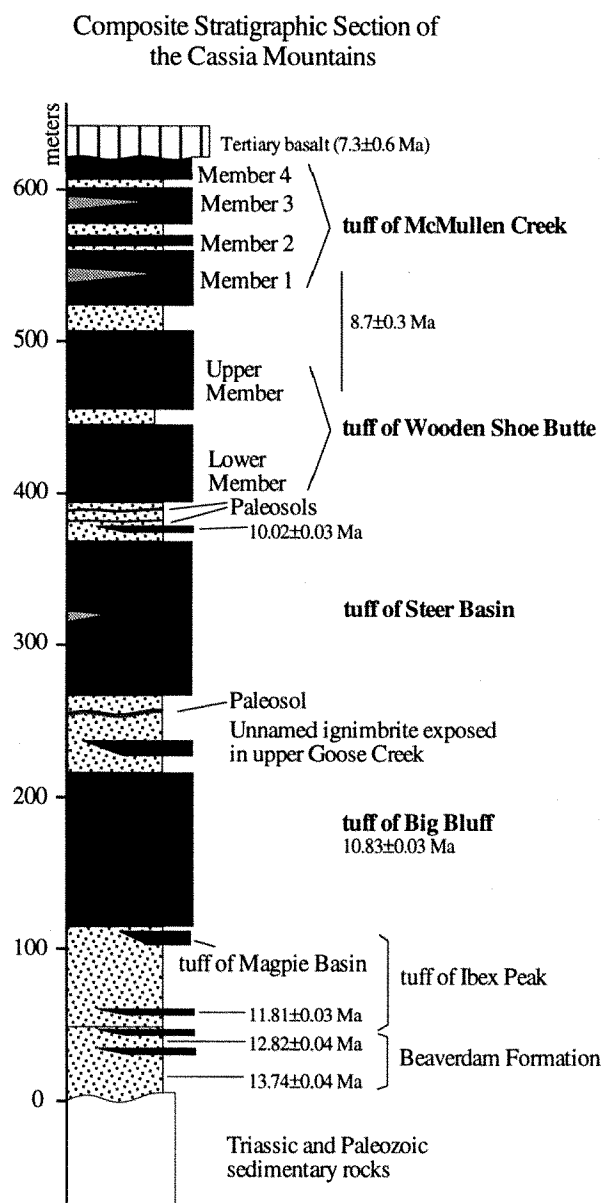


Figure 4. Diagrammatic stratigraphic section of the Cassia Mountains. Modified from Williams et al., (1990), Mytton et al., (1990), and Perkins (1995). Black pattern indicates ignimbrite cooling units; shaded pattern represents cooling breaks within composite cooling units (Smith, 1960); stipple pattern represents bedded fallout tuffs and water-reworked tuffs. Most radiometric dates are $^{40}\text{Ar}/^{39}\text{Ar}$ dates from Perkins et al., (1995); 8.7 and 7.3 Ma dates are K/Ar dates from Armstrong et al., (1975; 1980). Vertical arrow for the 8.7 Ma date indicates uncertainty as to whether Armstrong's analysis is from the tuff of McMullen Creek or from the tuff of Wooden Shoe Butte (Perkins et al., 1995).

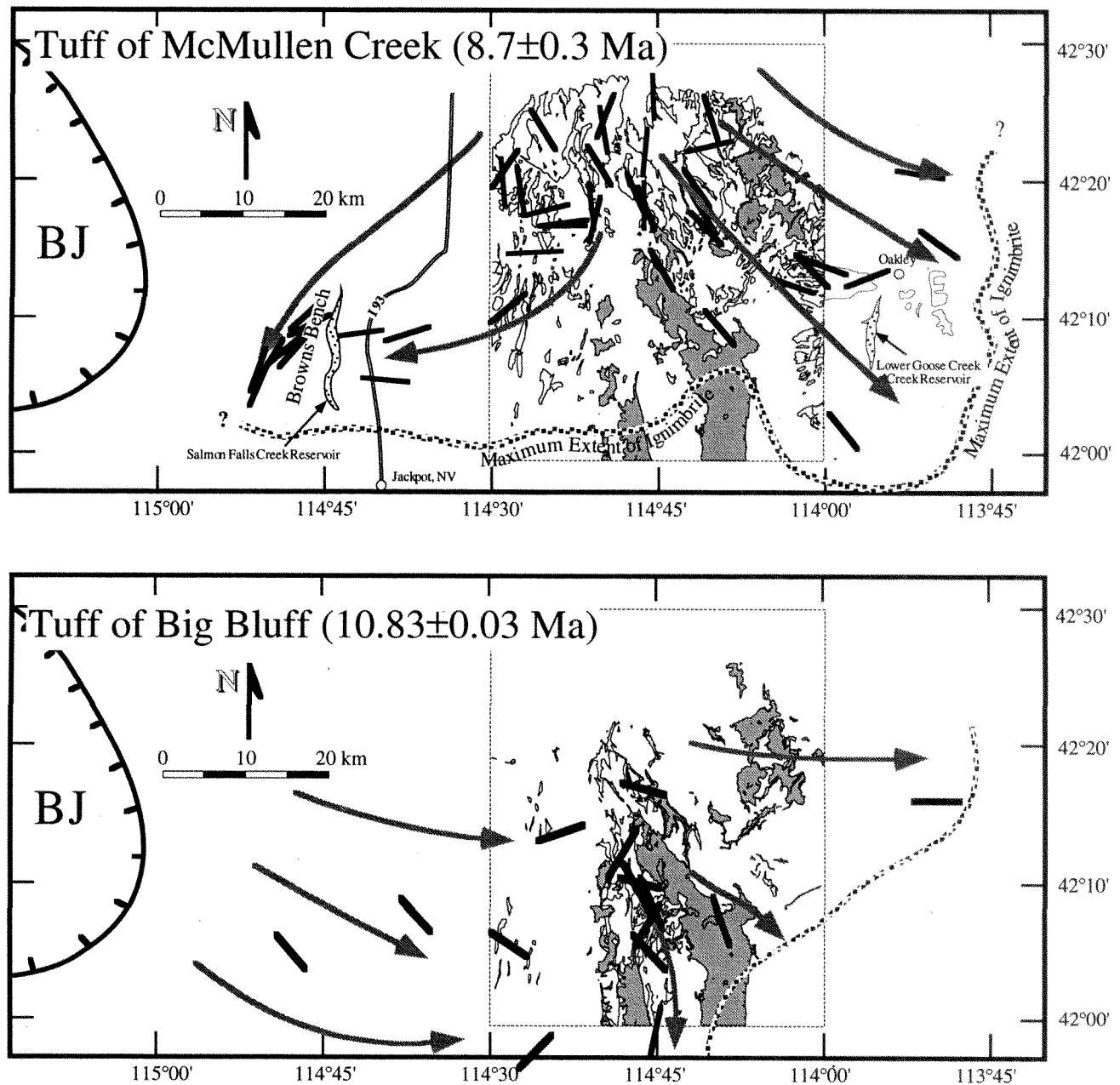


Figure 5. Rheomorphic flow lineations and exposures of the tuff of McMullen Creek (top), and tuff of Big Bluff (bottom). Lineations are illustrated by bold line segments; interpretations of original pyroclastic flow emplacement directions are shown with shaded arrowed lines, and maximum extents with bold stippled lines. Exposures of respective ignimbrites (outlined with no pattern) and undifferentiated pre-Tertiary rocks (stippled) are illustrated within the dashed rectangle, and are based upon recent detailed mapping of the four Stricker Quadrangles by Williams et al., 1990, 1991; Mytton et al., 1990; Williams, personal communication, 1996. Exposures of tuff of McMullen Creek are shown near Lower Goose Creek Reservoir, and are based upon our interpretation of unpublished maps by R. L. Armstrong (U.S. Geological Survey). BJ = Bruneau-Jarvis eruptive center (Bonnichsen, 1982a).

Petrology of Rhyolites from the Cassia Mountains— Implications for SRP Magmatism

Petrologic and geochemical characteristics of ignimbrites in the Cassia Mountains are summarized in order to compare the Twin Falls eruptive center with other volcanic centers of the SRP system. Approximately 235 chemical compositions of SRP rhyolites were compiled from the literature and from unpublished data in order to better define variations in tectonism, magma genesis and volcanic evolution along the track of the SRP system. These data are described in more detail by Hughes et al., (1996).

Mineralogy

Phenocrysts in most Cassia Mountains ignimbrites consist primarily of oligoclase to andesine plagioclase >> pyroxenes (pigeonite, augite and hypersthene), sanidine (minor or absent), Fe-Ti oxides, quartz, and accessory zircon and apatite. Crystals are typically euhedral to subhedral, and <0.3 to ~5 mm in size. Plagioclase is often the largest phenocryst and displays the greatest variety of textures. "Sieve textures," embayments, and rounded edges are common, and probably resulted from rapid crystallization followed by rapid decompression or volatile loss prior to eruption. Phenocrysts comprise varying proportions of the deposits, ranging from <2% (by volume) in basal bedded ash to >22% in surge layers.

Accessory apatite and zircon phenocrysts are found as inclusions within glass shards and essential minerals; rare biotite microlites (<0.03 mm) occur within incipiently crystallized glass shards associated with pyroxenes and apatite. Biotite appears to have formed during stages of incipient devitrification, rather than as an intratelluric phase. Skeletal grains of ilmenite and rutile often formed on glass shards and pumices.

Geochemistry

Cassia Mountains rhyolites are metaluminous to slightly peraluminous, based on molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ ratios (A/CNK) ranging from 0.9 to 1.1, and molar $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ ratios ranging from 1.0 to 1.6. Flow units generally do not display systematic vertical or lateral chemical zonation, suggesting that each ignimbrite either originated from a homogeneous source, or any chemical zonation in the original magma was disrupted during the eruption and emplacement process (Parker, 1996). However, with increasing SiO_2 contents, decrease in TiO_2 , MgO , Al_2O_3 , CaO , FeO , MnO , and Na_2O , and increase in K_2O are observed. Parker (1996) suggests that at least part of these variations resulted from crystal-glass segregation in the moving ash flow. Variations in Na_2O , K_2O , and SiO_2 concentrations resulted from glass hydration.

Trace element patterns (Fig. 6), normalized to average continental lower crust (Taylor and McLennan, 1985; Rudnick and Fountain, 1995), exhibit less enrichment in elements Ba, Sr, Eu, Ti and Sc that are fairly compatible in essential lower crust minerals, relative to less compatible elements Cs, Rb, Th, U, K, Zr, Hf and REE (rare-earth elements). Mineral-liquid equilibria during primary crustal anatexis, assimilation of partially-melted crust, or fractional crystallization of parental magma could account the relative abundances of these elements. The patterns suggest that the source restite (crystalline residue after partial melting) for the magma consisted of plagioclase, pyroxene, and Fe-Ti oxides, but did not include significant amounts of alkali feldspar, zircon, or other trace minerals such as apatite which, if originally present, became incorporated into the silicic magma. Glomerocrysts of plagioclase, pyroxene, and Fe-Ti oxides $\pm$ other minor minerals, may represent fragments of protolithic material derived from the magma source (Bonnichsen and Citron, 1982). The predicted restite assemblage and that of the glomerocrysts is consistent with the minerals found in granulite-facies rocks and models discussed above. The effect of subsequent mineral fractionation is also possible, as shown by depleted compatible elements, although this is not easily distinguished from a partial melt-restite relation.

Most significant is the uniformity among patterns, reflecting relatively minor magmatic evolution subsequent to magma generation. As with major elements, trace elements of Cassia Mountains ignimbrites display no systematic zoning either vertically through the deposit or horizontally on the scale of 10's of kilometers. It is therefore inferred, as with the major elements, that the spectrum of trace element abundances may be a product of crystal-glass segregation of the original pyroclastic flow.

Geochemical Comparison of SRP Volcanic Centers

A summary of locations, ages and lithology of SRP rhyolites obtained from various sources (Fig. 7) provides a basis for inferring petrogenesis and eruptive styles. Anhydrous intratelluric mineral assemblages and high-grade features predominate in the ignimbrites of southwestern SRP centers: Owyhee-Humboldt, Bruneau-Jarbridge, and Twin Falls. By contrast, rhyolites in the northeastern centers: Picabo, Heise, and Yellowstone more commonly contain hydrous intratelluric phases (biotite or amphibole), are poorly-to-densely welded, generally lack rheomorphic features and commonly contain well-developed vapor-phase zones (Morgan et al., 1984; Christiansen, 1984; Hackett and Morgan, 1988; Kellogg and Marvin, 1988). These observations indicate a southwest to northeast change in ignimbrite volcanism between the Twin Falls and Picabo eruptive centers, in which eruption temperatures decreased and volatile concentrations increased across the SRP province.

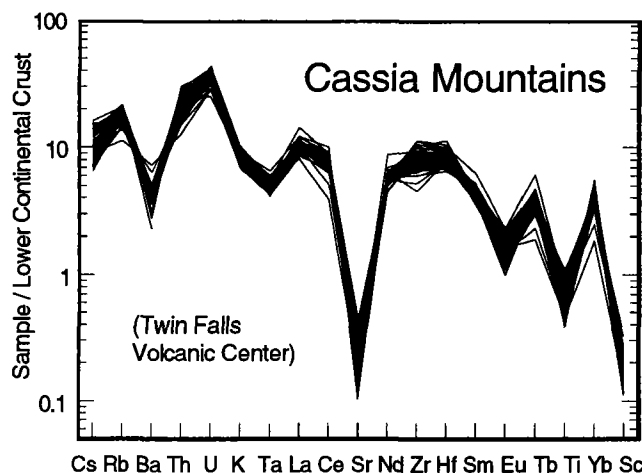


Figure 6. Trace element patterns of Cassia Mountains rhyolites (Twin Falls volcanic center) normalized to average lower continental crust values (Taylor and McLennan, 1985, Rudnick and Fountain, 1995). Elements are shown in order of overall decreasing incompatibility in lower crust and upper mantle regimes. (unpublished data from Hughes and McCurry).

Variation in selected trace elements reflect their compatibility in essential mineral phases under crustal conditions. Examples of such variation (Fig. 8) show separation of SRP volcanic centers, with ranges defining separate clusters for representative units from the northeastern and southwestern volcanic centers. Yellowstone and Heise volcanic centers exhibit relative depletion of these elements (Sr < 60 ppm, Ba < 1100 ppm, Zr = 140–480 ppm, Eu < 1.8 ppm) within the representative compositional ranges compared to higher values of these elements from the Twin Falls centers (Sr = 30–180 ppm, Ba = 600–1400 ppm, Zr = 400–800 ppm, Eu = 1.1–2.6 ppm). Bruneau-Jarbridge compositions roughly coincide with Twin Falls clusters, except for the higher Sr reported for several samples.

Chemical signatures of the Picabo volcanic center are represented by the Arbon Valley Tuff Member of the Starlight Formation (Kellogg et al., 1994) and associated rhyolites of Buckskin Basin, Two-and-a-half Mile Creek and Stevens Peak located near Pocatello (Kellogg and Marvin, 1988). Hughes et al., (1996) demonstrate that the Picabo Tuff, tuff of Thorn Creek, and other units located in and near the Mt. Bennett Hills north of Twin Falls (Leeman, 1982b; Honjo, 1990) yield different chemical signatures than the Arbon Valley compositions and plot in clusters defined for the Twin Falls volcanic center.

Current models of silicic magma genesis and evolution in the SRP system (e.g. Leeman, 1982a; Hildreth et al., 1991) argue in favor of a large heat source which caused extensive partial melting in the lithospheric mantle and subsequently in the lower crust. Rhyolitic ignimbrites in the

southwestern part of the SRP province erupted directly from deep-seated magma reservoirs below the Owyhee Mountains and did not produce significant ring fractures or caldera faults (Ekren et al., 1982, 1984). In contrast, ignimbrites in northeastern SRP systems are associated with ring fracture systems and caldera complexes such as those at Yellowstone (Christiansen, 1984), suggesting significant magma storage in mid- or upper-crustal reservoirs. The origin of the physiographic and chemical differences between the southwestern and northeastern SRP volcanic fields is not clear, but may be related to a west-to-east lithospheric transition from accreted to cratonic terranes (e.g., Leeman et al., 1992).

A schematic model of magma genesis and evolution in the two major regions of the SRP province is presented in Figure 9. Silicic magmas are derived at temperatures greater than 1000°C from deep-crustal sources where plagioclase and pyroxene dominate the residuum (K-feldspar and quartz probably are not restite phases). Generation of silicic magma is directly related to mafic magma underplating, which causes partial melting of the lower crust. Bulk lower crustal compositions are strongly modified prior to and during partial melting by hybridization with the mafic melts (e.g., Hildreth et al., 1991; McCurry et al., 1995). Concomitant formation of incipient silicic melts ensues, with segregation of silicic melts into separate lower crustal magma bodies. If these deep-seated magmas are tapped during lithospheric extension, high-temperature rhyolites are erupted without significant secondary evolution in the upper crust. However, if lithospheric extension is insufficient to allow magma ascent through fissures, or if the crust is more stable during lithospheric extension, magmas will remain as relatively small diapiric bodies that migrate upward and coalesce in upper crustal realms. This is more likely to occur in the northeastern segment of the SRP province where the crust is thicker, older, and compositionally different to crust in the southwestern segment.

FIELD TRIP STOPS

Stops for Day 1

Stop 1-1: Snake River Canyon at Perrine Bridge

The field trip begins at Perrine Bridge on U.S. Highway 93, about 1 mile north of Twin Falls (Fig. 2). At Perrine Bridge the Snake River canyon is 485 feet deep and about 1,500 feet across. The canyon exposes a vertical sequence of Late Miocene rhyolitic lava flows, overlain by late Miocene to Pleistocene basalt flows (Armstrong et al., 1975). The Shoshone Falls rhyolite lava flow is the lowest visible volcanic unit here and is well-exposed downstream where it forms the large bench that is the setting of the Blue Lakes golf course. The 150-foot-high cliff at the river's edge, a

Volcanic Center	Owyhee-Humboldt	Bruneau-Jarbridge	Twin Falls	Picabo	Heise	Yellowstone
location	42° N 116-117° W	42° 30' N 115-116° W	43° N 114° W	43° N 113° W	44° N 112° W	44° 30' N 111° W
age	13.8 - 12.0 Ma	12.5 - 11.3 Ma	10.02 - 8.6 Ma	10.2 Ma	6.7 - 4.3 Ma	2.0 - 0.6 Ma
ash-flow tuff units	tuff of Mill Cr., Swisher Mt. Tuff, upper and lower lobes of Juniper Mt. tuff of the Badlands	Cougar Point Tuff	tuffs of: McMullen Creek, Wooden Shoe Butte, and Steer Basin	Arbon Valley Tuff Member of the Starlight Formation	tuffs of: Blacktail, Blue Creek, and Kilgore	Huckleberry Ridge Tuff, Mesa Falls Tuff, Lava Creek Tuff
*essential minerals (trace minerals)	Qtz, Sa, Pl, Pig, Opx, Opq. (Zrn, Ap)	Pl, Sa, Qtz, Cpx, Pig, Fa, Mag, (Ilm, Ap, Mnz, Zrn)	Pl, Sa, Pig, Cpx, Qtz, Opq, (Ap, Zrn)	Qtz, Sa, Pl, Cpx, Bt	Qtz, Sa, Pl, Cpx, Opx, Mag, Bt (Zrn, Ap)	Qtz, Sa, Pl, Cpx, Cam, Fa, Mag, Ilm, (Zrn, Aln, Ap)
anhydrous assem.?	yes	yes	yes	no	no	no
lava-like rheomorphism?	yes	yes	in upper zones	no	no	no
temperature	930 - 960 °C	800-1050 °C	1040 - 1100 °C	not determined	not determined	> 850 °C

Figure 7. Summary of locations, ages, ash-flow tuff units, and eruptive mechanisms of six SRP volcanic fields after Bonnichsen, 1982 a,b; Bonnichsen and Citron, 1982; Embree et al., 1982; Christiansen, 1982, 1984; Ekren et al., 1984; Hildreth et al., 1984; Morgan et al., 1984; Neace, 1986; Bonnichsen and Kauffman, 1987; Kellogg and Marvin, 1988, Henry and Wolff, 1992; Honjo et al., 1992; Pierce and Morgan, 1992; Kellogg et al., 1994; McCurry et al., 1996. Abbreviations follow from Kretz (1983), except for Pig (pigeonite) and Opq (undifferentiated opaque minerals).

quarter mile downstream from the bridge, clearly reveals the interior of this gigantic rhyolite lava flow. The flow lobes at the margin of this flow can be seen downstream, farther to the west, near the golf course. From the Perrine Bridge area one can note the massive interior, the sheeted and folded rocks at the top of the massive section, and lenses of breccia at the top of the flow. Similar breccias are common at the tops of many SRP rhyolite lava flows. At several localities the upper breccia zone at the top of the Shoshone Falls rhyolite flow is especially thick and laterally extensive. However, it shows little evidence that it was fused by heat from the underlying lava, as is the case in many rhyolite lava flows. This flow may have run into standing water in ancient Lake Idaho (Jenks and Bonnichsen, 1989) when it was erupted about 6 Ma ago (Armstrong et al., 1975).

At Shoshone Falls, a few miles upstream, this same lava flow has an exposed thickness of more than 200 feet and holds up the high waterfall. Street and DeTar (1987) report that the Shoshone Falls rhyolite flow may be about 600 feet thick near Shoshone Falls, a few miles east of here, so

only about the upper half is exposed there. Drilling at that locality indicates this flow lies above tuffaceous and oolitic sediments of probable lacustrine origin, and these in turn lie on silicic welded tuffs similar to those exposed in the Cassia Mountains a few miles to the south. Published drill logs (Street and DeTar, 1987) suggest that a buried structural discontinuity probably occurs at the top of the silicic volcanic section south of the canyon just downstream of Perrine Bridge (Fig. 10). This discontinuity lines up with the margin of the Shoshone Falls rhyolite in the canyon downstream from the bridge. East of this discontinuity the Shoshone Falls lava flow lies below basalt and above welded tuffs, but to the west basalt lies on welded tuff that is about 1,000 feet higher, and there is no intervening section of Shoshone Falls rhyolite. Bonnichsen and Godchaux (1997, in press) have suggested a N-S-oriented caldera margin (part of the Twin Falls eruptive center) may be buried there. Interestingly enough, the fracture system associated with such a caldera-margin would project beneath the zone where geothermal wells are located on the College of Southern Idaho campus in Twin Falls.

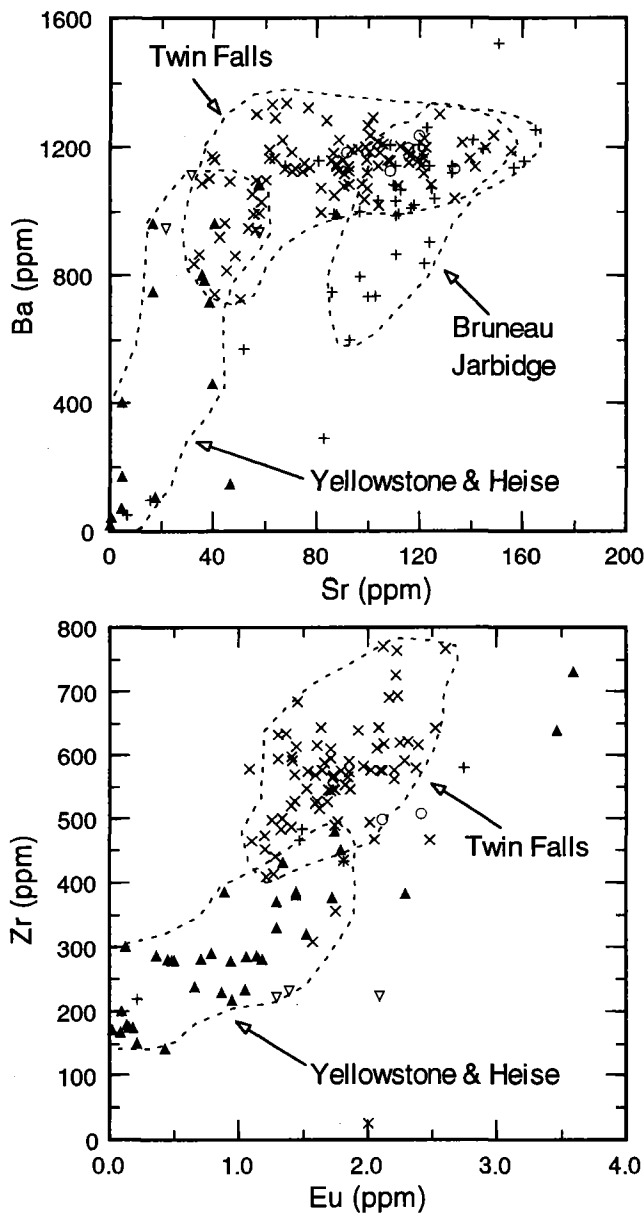


Figure 8. Co-variation plots of Ba vs Sr and Zr vs. Eu for SRP rhyolitic units. Data are from Bonnicksen *et al.*, (1994a, b), Ekren *et al.*, (1982), Leeman (1982b), Morgan *et al.*, (1984), Hildreth *et al.*, (1984, 1991), Kellogg and Marvin (1988), Honjo (1990), Manley (1995), and unpublished data by Hughes and McCurry. Symbols are as follows: plus signs (+) = Bruneau-Jarbidge, crosses (x) = Twin Falls, open circles = units associated with Mt. Bennett Hills and Arbon Valley Tuff, inverted open triangles = Heise, and solid triangles = Yellowstone.

The chemical composition of the Shoshone Falls rhyolite is listed in Table 1 and is typical of many rhyolite lava flows in the SRP volcanic province. Phenocryst minerals in the Shoshone Falls rhyolite flow are plagioclase, augite, pigeon-

ite, and opaque oxides (Salma Monani, Mount Holyoke College, personal communication). Quartz, alkali feldspar, biotite, and amphiboles are absent—a petrographic characteristic of many SRP rhyolite lava flows. The Shoshone Falls rhyolite probably represents a late stage, intracauldron facies of the Twin Falls eruptive center.

Typical SRP basalt flows from subaerial eruptions can be seen in the walls of Snake River canyon above the Shoshone Falls rhyolite flow. At the south end of Perrine Bridge the basalt is from Hansen Butte, a large shield volcano located 14 miles to the SE, and perhaps from older sources. The top flow on the north side of the canyon is the Sand Springs basalt (Covington and Weaver, 1990). Much of it was erupted from Flat Top Butte located 9 miles to the north, and other portions of this unit may have come from shields located many miles east of here. Flat Top Butte is one of the largest shield volcanoes in the central SRP. Just downstream, basalt flows from Hub, Stricker, and Berger Buttes, among others, are found in the canyon walls (Bonnicksen and Godchaux, 1997, *in press*).

The largest of several short side canyons north of the river, Blue Lakes Alcove, is visible on the north side of Snake River canyon downstream from Perrine Bridge. This side canyon is believed to have been eroded back from the main canyon during the Bonneville Flood and, along with several similar blind canyons on the north side of Snake River canyon, marks the downstream end of the Rupert channel of the Bonneville Flood (Malde, 1968). Blue Lakes at the head of Blue Lakes alcove, like several other lakes near here in the canyon, is fed by large springs and occupies a deep plunge pool eroded during the flood. Blue Lakes constitutes the Twin Falls city water supply. Many large boulders litter the Snake River canyon floor downstream from the Blue Lakes area. These rocks were ripped from the canyon rims and deposited in gigantic gravel bars during the flood. The boulder deposits have been whimsically named the Melon Gravel after similar boulder deposits near Hagerman that resemble fields of watermelons (Malde and Powers, 1962).

The stone monument at the south end of Perrine Bridge was erected by the Twin Falls Chamber of Commerce to commemorate daredevil Evel Knievel's attempted rocket-propelled motorcycle jump across the Snake River canyon 2 miles east of this point in 1974. Unluckily (or luckily) the parachute on his flying machine opened prematurely, giving him a soft landing in the river, rather than on the rocks across the canyon. This may have been the most famous happening in the history of Twin Falls.

Perrine Bridge to Rock Creek Canyon

From Perrine Bridge, proceed east about 10 miles on Interstate Highway 84. Take Exit #182 towards Kimberly,

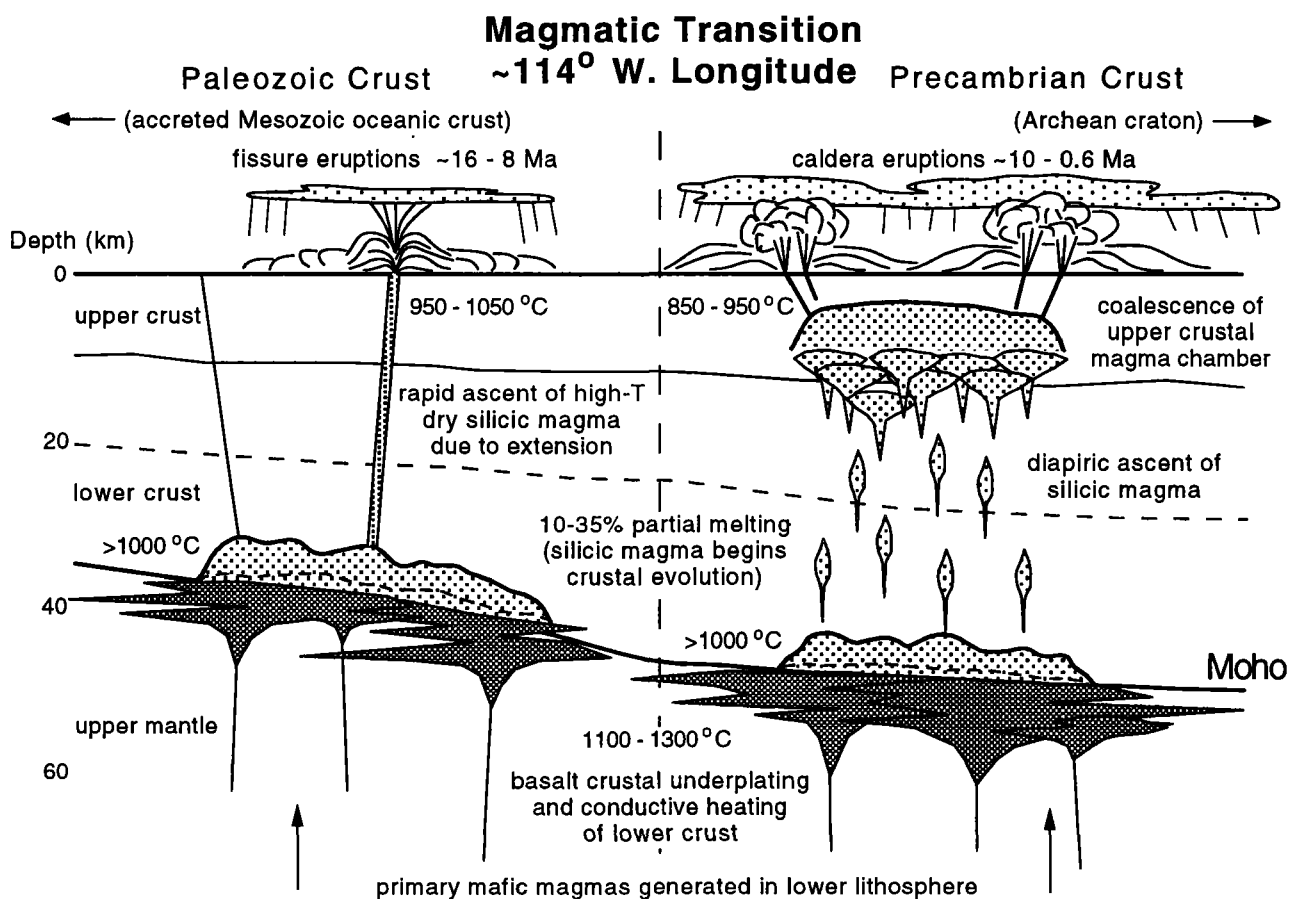


Figure 9. Schematic model of the magmatic transition in the SRP hotspot tract at ~114° W. longitude. West of the transition, high temperature rhyolites were erupted from regions in the lower crust directly along fissure systems in response to Basin and Range extension. East of the transition, granitic parental magmas diapirically ascended and coalesced into large upper crustal magma chambers capable of cataclysmic ash eruptions leading to caldera collapse (after Hughes et al., 1996).

and proceed southwest on State Route 50 for 1.1 miles. Turn left (south) onto County Road G-3 (Rock Creek Canyon Road) and proceed south towards the town of Hanson (1.9 miles). Continue south past Hanson towards a prominent canyon incised into the north flank of the Cassia Mountains. Over the next few miles there are many fine views of the Cassia Mountains. The mountains rise gently from the surrounding plains and valleys from ~4,500 feet to a maximum elevation of 8,060 feet. The Cassia Mountains are physiographically and structurally unusual for this region in that the rocks have been cut by numerous cross-cutting northeast-, north-, and northwest-trending Neogene-age normal faults. Tilting of the resultant fault blocks produced a broad domal topographic high rather than a more typical half-horst-type mountain range, producing dip slopes of ignimbrites, primarily the tuff of McMullen Creek (Fig. 4).

Approximately 7.5 miles south of Hanson, Rock Creek Canyon road intersects a paved road to Nat-Soo-Pah Hot Spring (you will return to this road between Stops 1-3 and

1-4). Continue south into Rock Creek Canyon. Over the next 7 miles you will be moving stratigraphically down section through the upper three of four major ignimbrites (Fig. 4).

Rock Creek Canyon is the largest of many spectacular structurally controlled canyons in the Cassia Mountains. Many originated by erosion along prominent faults. The upper part of Rock Creek Canyon is a narrow graben deepened by erosion. Numerous faults occur on both sides of the road, but are difficult to see, making correlation of the similar-looking ignimbrites difficult.

About 6 miles past the road to Nat-Soo-Pah Hot Spring, note distinctive black cliffs next to the east side of the road. These consist of densely welded glass of the lower part of the tuff of Wooden Shoe Butte (massive looking), and stratified rocks consisting of welded fallout, water-reworked vitric tuff and lapilli tuff (bottom). Welding of the stratified tephra occurred in response to heating from the overlying ignimbrite. The fact that these and similar tuffs elsewhere

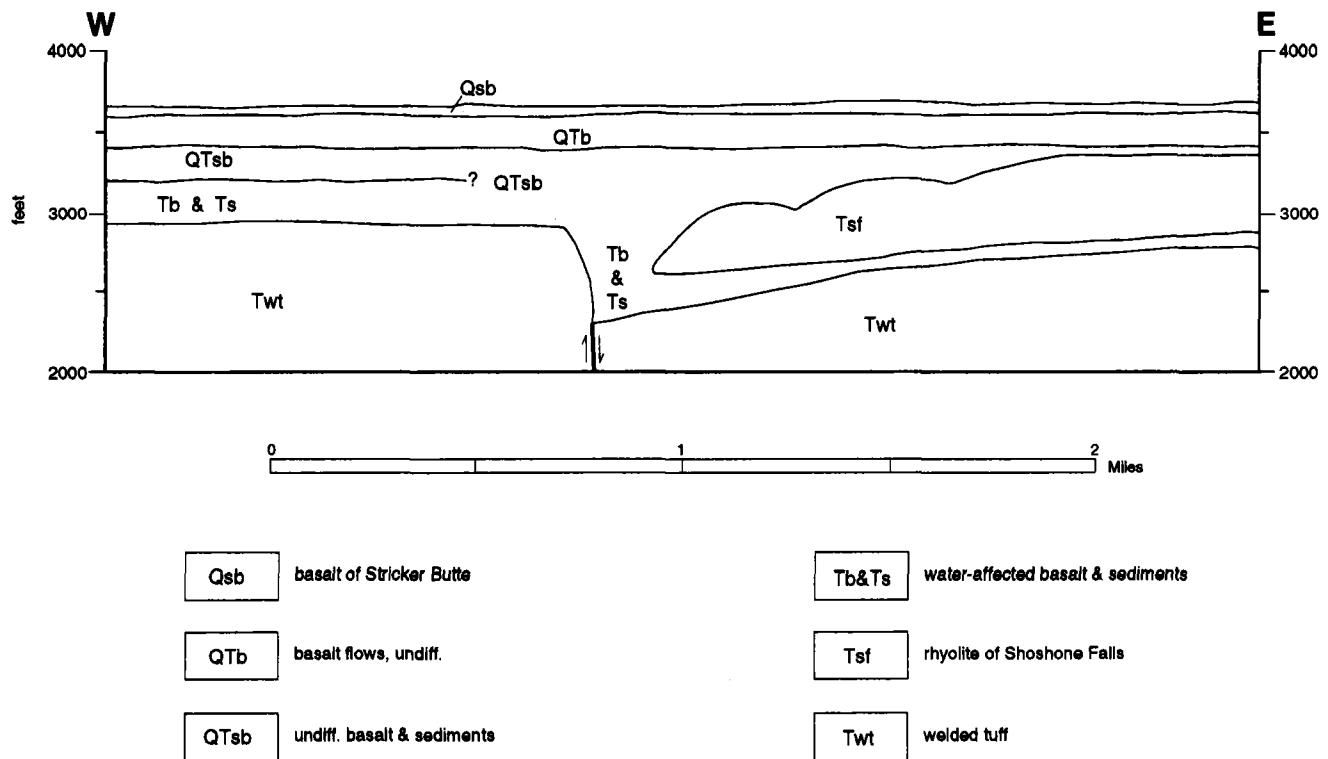


Figure 10. East-west cross-section at the north side of Twin Falls, about a mile south of Snake River canyon, showing possible buried caldera margin, from Bonnicksen and Godchaux, 1997 in press.

in the Cassia Mountains are commonly welded attests to the unusually high emplacement temperatures of the ignimbrites. Continue south on the Rock Creek Canyon road to Stop 1-2, approximately 7.3 miles south of the intersection to Nat-Soo-Pah Hot Spring.

Stop 1-2: Tuff of Steer Basin, Rock Creek Canyon

Objectives of this stop are: 1. to describe the volcanology and petrology of ignimbrites of the Cassia Mountains area; 2. observe volcanological features of the tuff of Steer Basin–Shipper section (Fig. 11).

Park in the large, paved pull-out on the right (west) side of the road. View to the east is of a layer-cake sequence of seven ignimbrite cooling units. These have been divided into three stratigraphic units by Williams et al., (1990) as follows: tuff of Steer Basin (lowest), tuff of Wooden Shoe Butte (upper and lower members); and tuff of McMullen Creek (4 members).

A prominent sequence ~15 meters thick of white to light gray unwelded fallout and water-reworked tephra occurs between the Steer Basin and Wooden Shoe Butte ignimbrites (referred to as the "Unnamed bedded tuff, unit 2" by Williams et al., 1990). The tephra can be seen high up on the valley wall, and also in a fault slice adjacent to the

east side of the Rock Creek Canyon Road. Walk across the road to observed the upper contact of the tuff of Steer Basin with the unnamed bedded tuff.

Lowermost exposures of the tuff of Steer Basin consist of strongly porphyritic (20%; dominantly plagioclase to 3 mm across), densely welded, devitrified tuff, and is also characterized by prominent flow banding. The devitrified welded tuff grades upward, through an undulatory zone ~1 meter thick, into perlitized vitrophyre. The vitrophyre is overlain, and in some places cut by, well-bedded ash-cloud surge deposits (e.g., Fisher, 1979). These in turn are unconformably overlain by 10–15 meters of white to light gray, well-bedded, water reworked tuff and lapilli tuff. A distinctive reddish soil profile occurs at the top of the ash-cloud surge deposit.

Return to the paved road and walk south about 1/4 mile, then scramble up the steep slope on the east side of the road to the base of the prominent cliffs. The lower part of slope is underlain by tuff of Big Bluff. This tuff has been dated at 10.83 ± 0.03 Ma by Perkins et al., (1995), and is inferred by them to be correlative with Cougar Point Tuff XIII (Bonnicksen and Citron, 1982). It is easy to distinguish from the younger ignimbrites of the Cassia Mountains by relatively abundant phenocrysts of alkali-feldspar

tuff of Steer Basin - Shipper Section

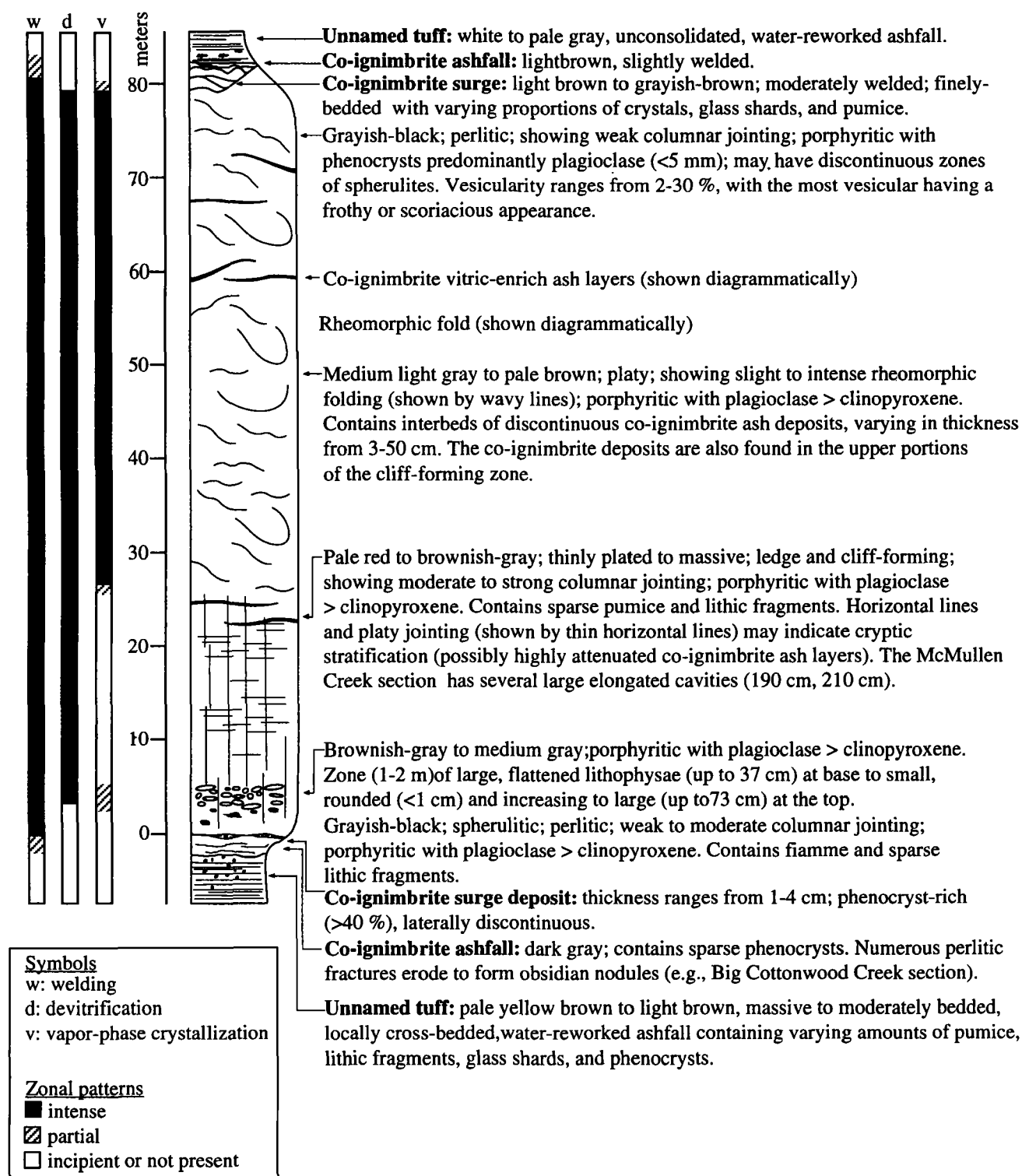


Figure 11. Simplified stratigraphic section of the tuff of Steer Basin—near Shipper Campground (Stop 1-2) (A. Watkins, personal communication, 1996). Columns on left indicate intensity of welding (w), devitrification (d), and vapor-phase crystallization (v).

and quartz (in addition to plagioclase, pyroxenes and Fe-Ti oxides). Outcrops here expose the uppermost part of the unit and consist of dark reddish gray, densely welded, devitrified tuff. Note the strong, near-horizontal sheeting joints, and hackly weathering. About 10 meters above the road there is a rapid upward gradation from devitrified tuff to glassy vitrophyre (poorly exposed).

Tuff of Big Bluff is overlain by about 10 meters of stratified tuffs. These are mostly covered, however small exposures occur about 6 meters above the vitrophyre. These consist of light to medium gray, to reddish brown, massive, unwelded, poorly sorted tuff and lapilli tuff. Weak, discontinuous bedding suggests the tephra was water reworked.

At the base of the prominent cliffs, tuff of Steer Basin (0 meters on Fig. 11) overlies ~1 meter of black, densely welded, well sorted, planar bedded to laminated fine-grained and medium grained vitric tuff and lapilli tuff. These are co-ignimbrite fallout deposits which immediately preceded and are comagmatic with the tuff of Steer Basin. Although originally unwelded, these welded deposits were subjected to intense heat and pressure from the overlying ignimbrite.

The base of the tuff of Steer Basin consists of a massive, dense black vitrophyre ~1.5 meters thick. The vitrophyre contains ~15% phenocrysts of plagioclase >> pyroxenes > Fe-Ti oxides, quartz (trace), and accessory phases (zircon and apatite). Lithics and pumice are sparse and small (generally < 1 cm). A well-developed layer 2a (McCurry et al., 1996), ~7–8 cm thick, occurs at the base of the ignimbrite, and is distinguished from the main mass of the ignimbrite (layer 2b) by a rapid, but gradational decline in phenocryst abundance and size near the base of the deposit. A discontinuous, thin (up to 4 cm) coignimbrite surge deposit occurs at the base of the ignimbrite, and is characterized by lens-like aggregates of plagioclase phenocrysts. Watkins et al., (1996) documents up to 80 cm of co-ignimbrite surge deposits for the tuff of Steer Basin at other locations.

About 1.5 meters above the base of the ignimbrite, vitrophyre gives way to devitrified tuff. The vitrophyre-devitrified tuff interface is sharp and characterized by relatively weak development of large spherulites. Lenticular lithophysal cavities occur in a zone 2–3 meters thick above the interface.

Return to the parking lot, then hike uphill around the left side of the prominent cliffs of tuff of Steer Basin. Skirt the left side of the cliffs until you reach to top, then proceed south about 200 meters to a small gully which exposes much of the upper part of the ignimbrite (25–85 m on Fig. 11). The upper part is commonly a slope former. The reason is not always clear, but apparently has to do with vertical changes in jointing and fracture patterns.

Hike up the gully; look for subtle discontinuous layers of densely welded, fine-grained, crystal-poor tephra (e.g.,

60 m on Fig. 11). These are often distinguished by smoother weathering surfaces than main mass (i.e. layer 2b type deposits) of the ignimbrite. The layers are intimately intermixed within the ignimbrite and are interpreted as vitric-enriched layer-3 (McCurry et al., 1996) ash-cloud surge deposits produced either by highly unsteady flow or by multiple flow emplacement (e.g., Branney and Kokelaar, 1992).

Further up the hill you will find patchy outcrops of the top of the ignimbrite (80 m on Fig. 11). In some places layer 2c (McCurry et al., 1996), co-ignimbrite ash cloud deposit, rests unconformably on the genetically associated layer 2b deposit. The erosional unconformity suggests that the original co-ignimbrite ash cloud (Fig. 12a) detached from the underlying pyroclastic flow (cf. Fisher, 1995), continuing its forward motion after the pyroclastic flow had “frozen.” Subsequently, the ash-cloud surge deposit was moderately welded by heat from the interior of the ignimbrite. Welding zonation of the tuff of Steer Basin therefore extends from co-ignimbrite ash (80 m on Fig. 11) through the middle and base of the ignimbrite, and also into underlying, originally “cold” tephra (to ~3 m on Fig. 11), features which attest to the extremely high emplacement temperature of the original pyroclastic flow deposit.

After returning to the vehicles, proceed south 4.4 miles to Steer Basin Campground.

Stop 1-3: Tephra Interbeds and Paleosols, Steer Basin Campground

At this stop we will discuss problems of regional stratigraphic correlations of Neogene volcanic rocks in the central and western Snake River Plain, and observe fallout and water-reworked tephra deposits.

The prominent prow-like cliff directly across Rock Creek from the Steer Basin Campground is tuff of Wooden Shoe Butte. It overlies beautifully exposed parts of the “Unnamed bedded tuff, unit 2” (Williams et al., 1990). From the east side of the creek note that the unnamed tuff contains two horizontal reddish bands, which are paleosols developed at the top of well-bedded sequences of light-gray fallout vitric tuff and water-reworked tephra.

A distinctive layer of light yellowish green crystal tuff and crystal-lithic lapilli tuff occurs beneath the lowermost paleosol—possibly the result of formation of authigenic clays in zones of high permeability. The paleosol itself is developed within a relatively massive tephra. This is atypical of the tephra sequences and may be a result of amalgamation due to bioturbation.

Several meters of fallout tuff overlie the upper paleosol; these exhibit no evidence of water reworking, and are probably fallout deposits which were precursors to the eruption which produced the tuff of Wooden Shoe Butte. The upper

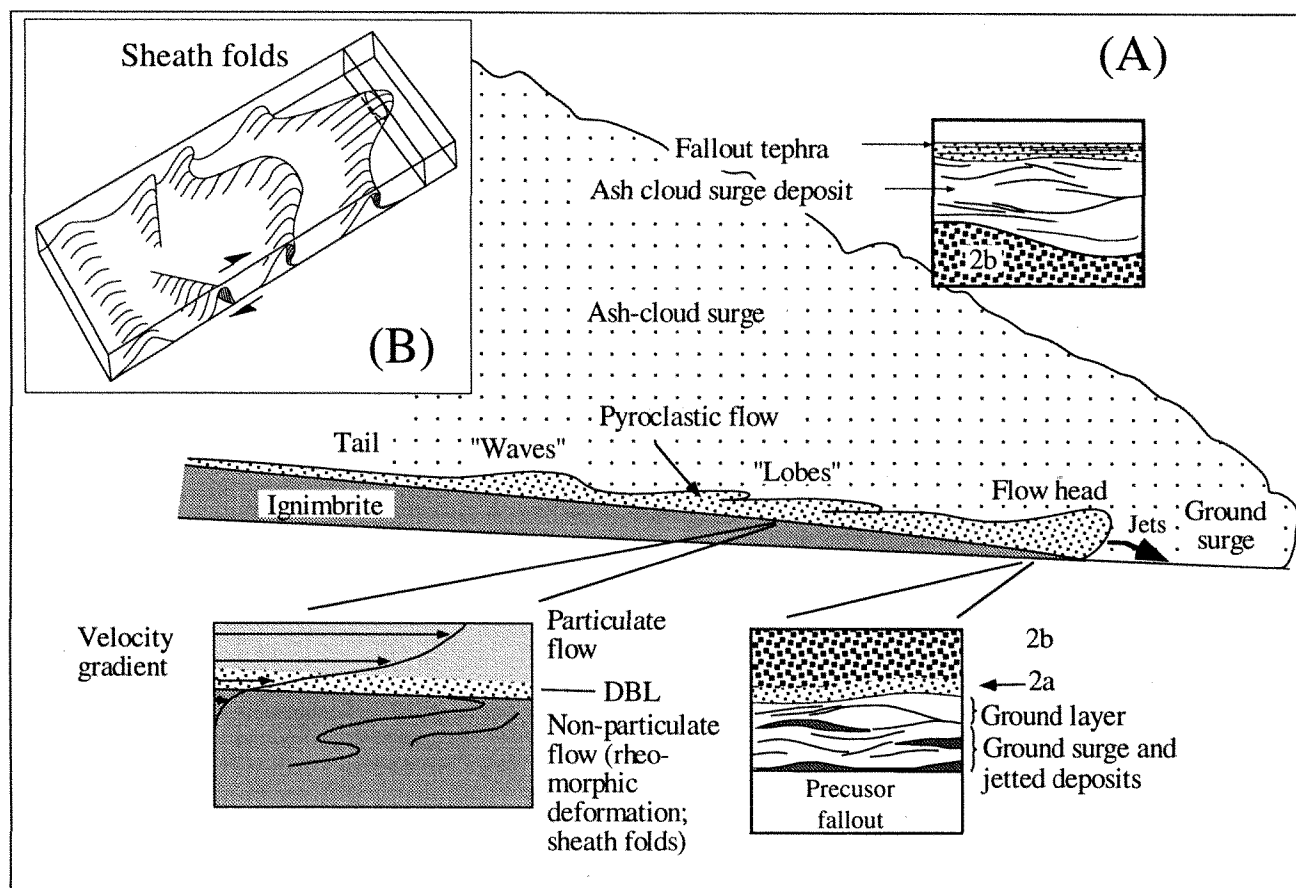


Figure 12. Illustrations of key depositional features, and an interpretation of depositional and rheomorphic processes, for a high-grade ignimbrite. Nomenclature for the depositional features follows Wilson and Walker (1982), Walker et al., (1981), Sparks et al., (1973) and Fisher (1979). (A) The pyroclastic flow is modeled as a dense particle flow from which deposition occurs by agglutination of particles through a narrow depositional boundary layer (DBL—following Branney and Kokelaar, 1992). Many of the “expanded-flow” deposits typically found at the margins of ignimbrites (i.e. ash-cloud surge, ground layer, ground-surge deposits and jetted deposits) may also occur within the body of the ignimbrite because of highly unsteady flow (Branney and Kokelaar, 1992; Walker et al., 1995). (B) An illustration of the geometry and formation of sheath folds. Such folds may form by progressive amplification, stretching and rotation into the direction of shear of fold axis which originally were perpendicular to the direction of shear (Cobbold and Quinquis, 1980; sketch after Hudleston, 1986).

parts of the fallout sequence are welded in a manner similar to that observed beneath the tuff of Steer Basin at our previous stop (1-2).

Rock Creek Canyon to Salmon Falls Creek Reservoir

Return to the SRP by driving north on the Rock Creek road, past the various welded tuff units. A short distance after exiting the canyon, turn west onto the paved road leading to Nat-Soo-Pah Hot Spring; continue west past the hot spring to US Highway 93. Along the way to Highway 93, the Cassia Mountains will be visible on the left (south) and various shield volcanoes, including Stricker and Hub Buttes, will be visible on the right. Turn left (south) once you reach Highway 93 just south of Hollister. The first hill

as you drive south on Highway 93, where the highway checking station is located, is another of the numerous basaltic shield volcanoes. As you drive south on Highway 93, the Cassia Mountains continue to be visible on the left. This portion also is mainly composed of welded-tuff sheets that erupted from the region around Twin Falls, and possibly from the west. Locally, Paleozoic rocks are exposed beneath the welded-tuff units. Several additional basalt shield volcanoes are visible along this part of the route. After passing the small town of Rogerson and continuing toward Jackpot, more welded-tuff units are visible on either side of the broad valley (the Rogerson graben) in which the highway is located. The escarpment to the east, 3–5 km away, is a continuation of the Cassia Mountains. The Browns Bench

escarpment, which was uplifted along a major fault, is visible 10–12 km to the west, and contains many welded-tuff sheets. The lower part of that section is thought to include many units erupted from the Bruneau-Jarbridge eruptive center several km farther west, whereas the upper part of the section is thought to consist of units erupted from the Twin Falls eruptive center.

Stop 1-4: Thin But Very Hot Ignimbrite North of Jackpot

Stop along Highway 93 across from the long exposure of welded and unwelded tuff on the east side of the highway about 11.5 miles south of Rogerson. The upper part of this exposure is a 2–3 m thick vitrophyre layer. This welded tuff unit was first mapped by Earl Cook (Alief, 1962), and probably correlates with the tuff of McMullen Creek (P. Williams, U.S. Geological Survey, personal communication, 1994). The vitrophyre overlies 2 or 3 m of friable air-fall ash. The top of this ash layer was fused by heat from the overlying ignimbrite. The ash contains pumice-rich layers and lies on a layer of tan, massive, nonwelded pyroclastic-flow deposit that resembles bedded silt.

The upper vitrophyre contains elongate vesicles and flow marks, lensoidal inclusions that probably started as pumice clasts, and spherulites. The densely-welded nature of this thin vitrophyre, the greatly flattened pumice inclusions and flow marks it contains, and the fusing of the underlying ash attest to the high emplacement temperature of the ignimbrite. Its preservation as glass with only scattered spherulites indicates rapid cooling once it was in place, as would be expected for a layer this thin.

Subparallel streaks or flow marks and elongate vesicles are common in many of the high-temperature, welded-tuff units in the Snake River Plain volcanic province. These streaks vary considerably in appearance and size. They probably were gas cavities that became extremely stretched during flowage in the final stage of emplacement. The parallelism of the flow marks and stretched vesicles presumably reflects the direction the hot ash was flowing when it became capable of containing and preserving vesicles. These flow marks would be termed “primary” structures in the sense of Chapin and Lowell (1979) because they were formed during flowage away from the erupting source, rather than being “secondary” structures that formed later, during rheomorphism of the sheet. The flow marks and stretched vesicles in the upper vitrophyre layer of this exposure are subparallel (N80W-S80E) suggesting that the ignimbrite source was to the west or northwest. Bonnicksen and Citron (1982) systematically measured flow marks within several Cougar Point Tuff units and demonstrated their regional consistency within individual welded-tuff units, and McCurry et al., (1996) have measured similar flow-mark data for several of the ignimbrite units from the Twin Falls

eruptive center. For the flow marks, dense welding, and extreme compaction to exist, it is probable that the ignimbrite was so hot when emplaced that it merged back to a liquid condition before it was quenched by rapid cooling.

Return about 3.5 miles north on Highway 93 to the Gray's Landing exit, then another 2 miles west to the boat landing on Salmon Falls Reservoir where there is plenty of parking space, and an outhouse, but no drinkable water.

Stop 1-5: Rheomorphic Welded Ignimbrite, Grays Landing

Objectives of this stop are to 1. discuss problems of ignimbrite deposition, in particular the formation of linear flow fabrics; 2. observe rheomorphic features of a high-grade ignimbrite.

Cliffs bordering the entry road and boat landing area consist of a densely welded, devitrified ignimbrite which is apparently correlative with the tuff of McMullen Creek. The tuff is rheomorphically deformed and exhibits a wide range of remarkably well developed flowage features including:

1. Sheath folds
2. S-type-folds
3. Striations and small grooves
4. Stretched lithophysal cavities

Most impressive among these are rheomorphic sheath and S-type folds; most of which verge to the south-south-west. The folds vary in size from a few centimeters to 10 meters across. Note that when viewed on vertical surfaces striking northwest the sheath folds, defined by flow banding within the ignimbrite, appear as ellipses, but when viewed at right angles to this direction, they appear as nappe-like folds (Fig. 12b).

Origin of such rheomorphic features in ignimbrites has been the subject of debate. Wolff and Wright (1981) argue that rheomorphism is a “secondary process,” i.e. one which occurs after the pyroclastic flow has ceased movement. They support a model of ignimbrite deposition in which the pyroclastic flow moves “en masse,” gradually losing energy, to the point of deposition (e.g., Sparks, 1976; Wright and Walker, 1981). In their view subsequent welding will produce a viscous mass which may, if deposited on an inclined plane, move much like a lava flow (e.g., Bonnicksen and Kauffman, 1987). Linear rheomorphic features, therefore, have little or nothing to do with the direction of movement of the original pyroclastic flow, but, instead, are controlled by preexisting topography.

In contrast, Branney and Kokelaar (1992; 1994) present a model, elaborating upon earlier work by Fisher (1966), in which ignimbrites accumulate incrementally through a depositional boundary layer (“progressive aggradation model”;

Fig. 12a). In cases where welding is sufficiently rapid, frictional coupling between the rapidly moving, overlying particle dispersion, and underlying welded part of the flow will transfer shear stresses into the welded deposit, and by viscous drag impart linear fabrics and folds in the direction of movement of the pyroclastic flow. Additional rheomorphic deformation may occur at any time in response to interactions between the welded tuff and underlying topographic features. They view rheomorphism as a continuum of processes related to original flow direction and differential stresses imparted from topographic barriers.

McCurry et al., (1996) describe several features of ignimbrites in the Cassia Mountains area which they believe supports the idea of Branney and Kokelaar, and conclude that most of the lineations are "primary." They also point to some prominent exceptions, where the ignimbrites are distinguished by seemingly randomly oriented lineations near partially exhumed paleotopographic highs, in which the lineations were apparently "secondary." Chapin and Lowell (1979) present a case-history of an ignimbrite which also exhibits these contrasting types of rheomorphic behavior.

Salmon Falls Creek Reservoir to Castleford Crossing

Return to U.S. Highway 93 and turn left (north). Follow Highway 93 to about 12 miles north of Hollister passing the junction with State Route 74 on the way, then turn left (west) on the country road located about 1 1/2 miles north of the junction with State Route 74, and proceed toward Castleford Crossing of Salmon Falls Creek canyon about 19 miles west of this turnoff. On the drive between Highway 93 and Castleford Crossing you will be crossing the flow field of the Sucker Flat ferrobassalt. This basalt erupted mainly from Berger Butte and an associated linear shield volcano situated a few miles north-northwest of Hollister, and flowed northward for many miles. It flowed into the western part of Lake Idaho north of Buhl, and disappears beneath younger basalt flows on the north side of Snake River canyon. It is by far the largest single ferrobassalt unit that we have found in the Snake River Plain. Its linear venting system, although discontinuous, is nearly 5 km long, and the volume of ferrobassalt erupted would seem to be between 10 and 100 cubic km.

A few miles before reaching Salmon Falls Creek canyon the road crosses a hill. This is Lookout Butte, another shield volcano. This shield, along with Sunset Butte located about 3 km to the NW, are the sources of the Lucerne Basalt, another Fe- and Ti-rich basalt which also flowed northward into Lake Idaho. Both the Berger Butte linear vent system and the Lookout-Sunset Buttes alignment are oriented NW-SE, which is parallel to, and on line with, the western SRP graben, which also contains much ferrobassalt.

Once Salmon Falls Creek canyon is reached, you will descend on the grade to the canyon bottom by driving down through the lower of two rhyolite units in the canyon, the Balanced Rock rhyolite lava flow.

Stop 1-6: Rhyolite Lava and Rheomorphic Ignimbrite Units at Castleford Crossing

Two rhyolite sheets are exposed in the Balanced Rock area. These were probably derived from the Twin Falls eruptive center. Balanced Rock is within the lower flow so it has been named the Balanced Rock rhyolite. It is believed to be a rhyolite lava flow. The upper rhyolite has been named the Castleford Crossing rhyolitic tuff, and it is provisionally believed to have been emplaced as an ignimbrite which was so hot that it underwent sufficient rheomorphic deformation to cause the unit to look very much like a rhyolite lava flow. The grade on the west side of Salmon Falls Creek affords an excellent cross section through the internal zones of these units. Two parallel northwest-trending faults have uplifted the zone immediately northeast of the road somewhat, so it may be confusing as to which portion of which unit one is looking at along the road. As a matter of fact it is so confusing that in previous guidebooks (Bonnichsen et al., 1988 and 1989) Balanced Rock was attributed to being in the upper unit rather than in the lower. However, systematic geologic mapping in 1996 (Bonnichsen and Godchaux, in prep.) has clarified the situation. Along the road climbing out of the west side of Salmon Falls Creek canyon, the rocks along the right (northeast) side of the road belong to the massive middle portion of the Balanced Rock (lower) rhyolite unit, and the rocks along the left (southwest) side of the road are the upper, brecciated portion of the Balanced Rock (lower) unit overlain by the basal portion of the Castleford Crossing (upper) unit.

The Balanced Rock (lower) rhyolite unit shows many features indicating it is a lava flow (Bonnichsen, 1982b, Bonnichsen and Kauffman, 1987). Its base, which is exposed just downstream from where the road crosses Salmon Falls Creek, consist of many feet of breccia in irregular contact with the overlying devitrified middle zone. The basal breccia sits on baked soil, with no airfall ash in evidence. The flow's internal features, including thickness variations, joint patterns, and abundant flow folding, are consistent with a lava-flow emplacement origin. By following the road out the west side of the canyon one gets an excellent view of internal features in the upper zone of this unit, and it has many features that are characteristic of a rhyolite lava flow. Especially abundant in the upper part of this flow are breccias, zones of alteration, spherulites, and other complex relations.

The Castleford Crossing (upper) rhyolite sheet, however, may be a pyroclastic unit (rather than a rhyolite lava flow

Table 1. Twin Falls Eruptive Center Rhyolite Analyses.

Number:	1	2	3	4	5	6
Stop:	1-1	1-6	1-6	1-6	1-6	1-6
Sample:	I-2940	I-3359	I-3396	I-3400	I-3394	I-3395
SiO ₂	69.83	70.43	70.20	75.27	70.72	69.81
TiO ₂	0.69	0.77	0.80	0.40	0.79	0.85
Al ₂ O ₃	13.42	13.22	13.32	12.48	13.18	13.45
Fe ₂ O ₃	4.62	4.64	4.66	2.15	4.65	4.90
MnO	0.08	0.09	0.08	0.06	0.08	0.08
MgO	0.64	0.67	0.78	0.45	0.70	0.80
CaO	2.10	2.10	2.27	1.00	2.11	2.46
Na ₂ O	2.85	3.30	2.97	2.05	3.31	2.87
K ₂ O	5.47	4.52	4.91	6.11	4.50	4.85
P ₂ O ₅	0.15	0.19	0.17	0.16	0.20	0.19
Total:	99.85	99.93	100.16	100.13	100.24	100.26
Mg# (At%)	21.5	22.2	24.9	29.3	23.0	24.4

Analyzed by Bill Bonnicksen at Ronald B. Gilmore x-ray fluorescence laboratory of the University of Massachusetts at Amherst. Total analyzed iron is expressed as Fe₂O₃.

1. Shoshone Falls rhyolite flow, upper vitrophyre.
2. Castleford Crossing rhyolite unit, devitrified material
3. Castleford Crossing rhyolite unit, lower vitrophyre.
4. Fused fallout ash immediately below the Castleford Crossing rhyolite unit
5. Balanced Rock rhyolite flow, devitrified material.
6. Balanced Rock rhyolite flow, upper vitrophyre.

as was previously suggested by Bonnicksen, 1982b; Bonnicksen and Kauffman, 1987 and Bonnicksen et al., 1988). The very base of the Castleford Crossing unit consists of a vitrophyre layer of fairly uniform thickness with a sharp, approximately level, contact with the overlying devitrified rhyolite in the middle zone of the unit. This type of base, which has been seen at the bottom of many large, hot, welded tuff units, has been traced along the canyon walls for several miles both south and north of the Castleford Crossing area. At the bottom of the basal vitrophyre is a partially welded zone a few cm thick, lying on probable ash-fall tuff. Thin sections show that the partially welded zone was emplaced as fragmental material, although no thin bubble-wall shards were noted. Immediately above this thin, partially welded zone the basal vitrophyre of the Castleford Crossing unit is completely welded, contains flow bands, folds, and similar features indicating that it became entirely molten and flowed.

Phenocryst minerals observed in these flows include plagioclase, augite, hypersthene, and Fe-Ti oxides. Analyses of both units are included in Table 1. They are essentially identical and about as femic as any of the rhyolite units in the Snake River Plain volcanic province.

Balanced Rock, located about one km up the grade west of Salmon Falls Creek, is the local scenic wonder, and is

well worth the visit. It is about 40 feet high and rests on a base only about a foot across, and which has been reinforced from the way it was before civilization came to this part of Idaho a few years ago. This monolith has been sculpted out of one of the large vertical columns in the middle part of the Balanced Rock rhyolite flow, probably by a combination of repeated freeze-thaw shrinkage and expansion acting on all the little joints in the rhyolite and guided by the original configuration of the vertical shrinkage joints that bounded the block from which it formed. Some have suggested that wind erosion may have helped form it, which also seems possible. Although the age of Balanced Rock is not known, its obviously precarious stance certainly is testament to a lack of significant seismic events in this part of the SRP for the last many thousands of years.

At the top of the hill above Balanced Rock the Castleford Crossing (upper) rhyolite unit is exposed; here it is only a few feet thick, consists mainly of vitrophyre, and rests on partially welded silicic fallout ash. Flow folds have been seen in this unit even where it is thin and, sporadically, zones of brecciated material occur in its lower part. Even though we lean toward the notion that this unit was distributed from its eruptive source by a pyroclastic-flow type of mechanism, the exact manner of its final emplacement remains somewhat of a mystery.

Castleford Crossing to Nampa

Return on the road back to the east side of Salmon Falls Creek and continue east and north to Buhl. On the way we will again be traveling across a basalt-covered portion of the Snake River Plain. At Buhl, join Highway 30 and head west toward Bliss. A few miles west and north of Buhl the highway descends into the Snake River canyon. Interesting items to be seen in the canyon include features formed during the Bonneville Flood, the Thousand Springs flowing from the east wall of the canyon, basalt flows that were altered when they ran into standing water, and remnants of phreatomagmatic basalt vents that were erupted into Lake Idaho. North of Hagerman, as the highway climbs out of the canyon, the world-famous Hagerman fossil beds are present within the Glenss Ferry Formation sediments exposed across the river (McDonald et al., 1996; Repenning et al., 1995; Sadler and Link, 1996). The Hagerman fossil beds, which are particularly famous for Pliocene horse fossils that were excavated, have recently been preserved as part of the National Park system. When driving out of Snake River canyon north of Hagerman, a pillow delta within the McKinney Basalt (Leeman and Vitaliano, 1976) is present in the same area where much evidence of land subsidence can be seen along the canyon wall. In 1993, a major landslide occurred near here, damming the Snake River, and cutting the road which had been placed across debris from the previous landslide. This road, of course, was immediately rebuilt and now patiently awaits the next slide.

At Bliss take Highway I-84 west toward Mountain Home, and on to Nampa. Between Bliss and Glenss Ferry the road again descends into Snake River canyon. Abundant exposures of lake sediments of the Idaho Group (Glenss Ferry Formation) that were deposited in Lake Idaho are present in this area. These sediments typically are capped by subaerial basalt flows. Close inspection of some of these flows reveals pillows, indicating the basalt had flowed into water.

A few miles west of Glenss Ferry, the highway again leaves the Snake River and climbs onto another basalt-covered plain. As you approach Mountain Home several basalt shield volcanoes can be seen. Also visible, to the northeast (right) at a distance of a few km, are the Mount Bennett Hills. These are composed of rhyolitic lava flows and welded-tuff units (Wood and Gardner, 1984) and are about the same age as, and very similar in character to, those observed on the south side of the SRP that were erupted from the Bruneau-Jarbridge and Twin Falls eruptive centers. During the drive to Nampa we will be going along the floor of the western SRP graben. One can get a feel for its width by noting the distance between the Owyhee Range to the southwest (left) and the Boise Front to the northeast (right).

The western SRP graben is about 60 km wide and extends about 160 km NW from the main SRP trend.

BASALTIC VOLCANISM, WESTERN SNAKE RIVER PLAIN

by Bill Bonnichsen, Craig White,
and Martha M. Godchaux

Western Snake River Plain Geology

Eastern and Western Snake River Plain Contrasts

The arcuate topographic lowland called the Snake River Plain that stretches like a smiling face across southern Idaho formed as a result of two principal tectonic events. The main, SW-NE-oriented SRP volcanic province is part of a trend of bimodal rhyolitic-basaltic volcanism that extends across southern Idaho and into adjoining portions of Nevada, Oregon, and the Yellowstone Park region. It has long been considered that this main SRP-Yellowstone trend, in which the first major volcanism consisted of the formation of a series of large calderas and eruptive centers (e.g. large complexes of intersecting calderas that now are buried by their own effusive products) forms a NE-younging progression (Fig. 1). Many geologists have speculated whether this linear trend represents: (1) the result of North America moving passively over an upwelling plume of exceptionally hot, less dense, material originating in the Earth's mantle, (2) an accommodation zone between the only slightly extended central Idaho terrane to the north and the more extended basin-and-range terrane to the south, in which late Cenozoic magmatism was more prevalent, or (3) the result of a SW-NE-oriented zone of crustal weakness, perhaps caused by shearing prior to Miocene time. Whatever is ultimately resolved about its origin, little evidence has appeared suggesting that NW-SE extension across the SRP is much of a factor. Rodgers et al., (1990), and Hackett and Smith (1991, 1992) have suggested that through time there may have been considerable NE-SW extension within the main SRP, parallel to its length. Their suggestion is based on the observation that dikes in the eastern SRP tend to be oriented NW-SE, trending across the SRP. We also have observed this at many localities in the central and SW portion of the main SRP trend.

In contrast to the main SRP-Yellowstone trend, the western SRP, which stretches NW from the Twin Falls area, past Boise, and into eastern Oregon, is demonstrably a complex graben, or continental rift (Mabey, 1982; Malde, 1991; Wood, 1994). There may have been several km of NE-SW-oriented widening across this rift during the last 10 to 12 Ma. Since this extension direction is nearly parallel to the length of the main SRP trend, it seems reasonable that the same stresses were responsible for producing both features, resulting in the general topographic configu-

ration (Fig. 1) in which the western SRP graben extends northwestward from the side of the main province. This fundamental difference in origin for the western and eastern SRP segments is reflected in the geology of the western SRP, especially by the fact that the western SRP held a large lake, commonly referred to as Lake Idaho, for most or all of the time between 11 and 1 Ma (Jenks and Bonnichsen, 1989). Also, the composition, style, and timing of volcanism in the western SRP graben differs somewhat from that in the main SRP trend.

Geologic History of the Western Snake River Plain Graben

The wide range of plutonic, volcanic, and sedimentary rocks that are exposed within and adjacent to the western SRP graben are summarized in the accompanying generalized stratigraphic column (Fig. 13). Southwest of the western SRP, principally in the Owyhee Mountains (Ekren et al., 1981) there are several older groups of igneous rocks that illustrate the complex geologic evolution of this region during pre-late Miocene time. Granitic rocks of late Cretaceous to early Tertiary age are extensively developed on both sides of the western SRP. Those to the north are part of the Idaho batholith and those to the south constitute the Silver City batholith. Eocene rhyodacitic to rhyolitic volcanic rocks occur in the Rough Mountain area on the SW margin of the western SRP graben. Although slightly younger than most of the Challis volcanic rocks of central Idaho, their composition and age suggest they are a south-westward extension of the Challis volcanic field. During the Oligocene a small field of generally andesitic rocks, the Salmon Creek volcanics (Norman and Leeman, 1989), was erupted near the SW margin of the western SRP. Prior to the inception of SRP volcanism and formation of the western SRP graben, between 17 and 14 Ma ago, there was an episode of bimodal basalt-rhyolite volcanism in the Owyhee Mountains immediately SW of the western SRP, and at other localities in northern Nevada and eastern Oregon. In SW Idaho this major pre-SRP episode of volcanism was not accompanied by the same style of structural deformation as that observed for the SRP system. North-south fault orientations suggest that in this 17–14 Ma time interval the crustal stretching direction was E-W. The development time interval of this pre-SRP volcanic episode is approximately the same as for the voluminous volcanism that formed the Columbia River basalt plateau.

Large-volume rhyolitic volcanism in the western SRP region started about when the Bruneau-Jarbridge eruptive center formed, 12–10 Ma ago (Ekren et al., 1981). These two areas lie side by side (Fig. 1). The silicic volcanism in the Bruneau-Jarbridge center was dominated by large, caldera-forming, explosive eruptions (Bonnichsen, 1982a,

GENERALIZED STRATIGRAPHIC COLUMN FOR THE WESTERN SNAKE RIVER PLAIN GRABEN

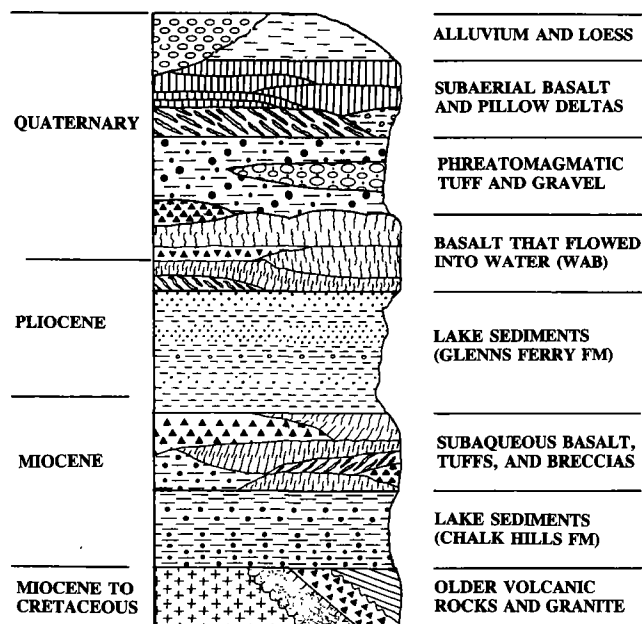


Figure 13. Generalized stratigraphic column for the western Snake River Plain graben.

and Bonnichsen and Kauffman, 1987), whereas the silicic volcanism that marked the inception of the western SRP graben seems to have produced voluminous rhyolite lava flows (Bonnichsen and Kauffman, 1987; Kauffman and Bonnichsen, 1990). These rhyolite flows are concentrated along the SW margin of the rift, although some have been noted on the NE side near Boise (Clemens, 1993), and many large rhyolite units occur a few miles east of Mountain Home, near the junction of the rift with the main SRP trend (Wood and Gardner, 1984). The first important NE-SW-oriented extension in the western SRP seems to have been synchronous with the formation of the Bruneau-Jarbridge eruptive center. Some of the largest and latest rhyolite lava flows within the area of the Bruneau-Jarbridge center may, in fact, be part of the western SRP graben development rather than part of the buried caldera-complex that constitutes the eruptive center (Bonnichsen, 1982b; Bonnichsen and Kauffman, 1987). By the end of the 12–10 Ma silicic volcanism episode the western SRP graben was established. Even then it evidently contained a lake, as suggested by the way the low-elevation portions of several of the large rhyolite lava flows are fragmented and altered.

There were two basaltic volcanism episodes in the western SRP graben after the silicic volcanism was finished (Fig. 13). The first, at about 9 to 7 Ma (Ekren et al., 1981;

Othberg et al., 1995a; Amini et al., 1984; Hart and Aronson, 1983), was mainly basaltic lava with olivine tholeiite compositions like the basalts from the Bruneau-Jarbridge and Owyhee-Humboldt centers. In the western SRP graben this episode involved many subaqueous eruptions, or eruptions that occurred near enough to Lake Idaho so that the flows ran out into the lake (Jenks and Bonnicksen, 1989; Jenks et al., 1993). The time limits of this volcanism are not well established, but it appears to coincide with the construction of many of the shield volcanoes within the Bruneau-Jarbridge center, and probably was partly synchronous with the development of the Twin Falls eruptive center. This 9 to 7 Ma basaltic episode in the western SRP graben occurred while the Chalk Hills Formation lake beds (Malde and Powers, 1962) were accumulating. Consequently, these basalts commonly are intercalated with, and even locally invasive into, the lacustrine sediments. The basalts that erupted in Lake Idaho at some localities seem to have been extruded into water that might have been several hundred feet deep, as inferred from high-stand shoreline features noted all around the lake basin (Jenks and Bonnicksen, 1989), and from the lack of vesicles in many of the subaqueous basalt flows. Near Mountain Home, a geothermal well penetrates nearly 3000 feet of such basalt, beneath 1000 feet of lake sediments, near the center of the rift (Lewis and Stone, 1988).

After this first western SRP basaltic episode, there was a hiatus of nearly 5 Ma, during which basalt was not erupted. Whether this hiatus affected adjacent areas, such as the Bruneau-Jarbridge or Twin Falls eruptive centers, is not yet known. During this hiatus up to several hundred feet of additional lacustrine sediments (the Glenss Ferry Formation) were deposited in Lake Idaho. Toward the end of Glenss Ferry deposition substantial downcutting of the outlet of Lake Idaho occurred, as the Hells Canyon of the Snake River was eroded deeper and deeper. Eventually Lake Idaho drained permanently through Hells Canyon, leaving the lake bottom exposed. Afterwards, additional downcutting by the Snake River further lowered the base level for its tributary streams, permitting the cutting of the deep canyons in central and southwestern Idaho, and better exposing interior portions of the SRP volcanic system.

Toward the end of Lake Idaho's existence, a second basaltic volcanic episode commenced in the western SRP graben (Fig. 13). It started about 2.2 Ma ago, and includes eruptions as late as about 0.1 Ma ago (Othberg et al., 1995a). The first eruptions were in the lake, probably under fairly shallow conditions, or were on land in areas adjacent to the lake, so that the lava flowed into the lake to form pillow deltas or actually just flowed out into the lake and resulted in water-affected basalt (WAB) sheets. As time passed and as the lake shrank the eruptions changed from dominantly subaqueous to phreatomagmatic. Numerous maars, tuff

rings, and tuff cones formed during this stage, largely due to subaerial conditions or only very shallow lake water. At the same time copious amounts of groundwater were available in lake sediments, in alluvial stream gravels, and in previously-formed basaltic pillow deltas. As Lake Idaho drained and erosion of the lake bottom commenced, the groundwater table sank to greater depths. Thus, the latter phase of this second basaltic episode consists principally of subaerial eruptions and flows, either from shield volcanoes or from small spatter and cinder cones. During this second basaltic episode there was some redistribution of unconsolidated lake deposits. This led to the deposition of the Bruneau Formation sediment (Malde and Powers 1962), which is mainly fine white silt intercalated between lava flows and tuffaceous layers.

After the basaltic volcanism ended, continued erosion and deepening of the canyons led to the present SW Idaho landscape. This process was enhanced along the course of the Snake River about 14,500 years ago, when the Bonneville Flood passed through (O'Connor, 1993 and Malde, 1968). After plucking numerous basaltic blocks from the canyon walls this torrent redeposited them as imposing boulder bars along the Snake River canyon. Finally, the most recent deposits in the area are sand dunes, loess blankets, redistributed alluvium along water courses, and landslide deposits on canyon walls.

Style of Basaltic Volcanism in the Western Snake River Plain

The western SRP rift provides classic exposures for studying the effects of water on the products of basaltic eruptions. Significant variations in water depth, from a large deep lake to an elevated water table, produced exposures with a wide range of characteristics. In addition, eruptions which ranged from small dike-fed flows to large, multi-phase volcanoes, created a variety of volcanic products including flows that ran into the lake, pillowed basalt flows that formed lake-margin deltas, pyroclastic tuffaceous units deposited beneath varying water depths, and hydrovolcanic constructions including tuff cones, tuff rings, and maars.

In the western SRP there are both effusive and explosive basaltic vents. The effusive vents are either shield volcanoes or small spatter and cinder constructs that were formed in the absence of sufficient meteoric water to cause phreatomagmatic explosions. There are many explosive basaltic vents of phreatomagmatic origin that resulted from the interaction of basaltic magma with groundwater or shallow surface water. The phreatomagmatic volcanoes in the western SRP can be grouped into three basic types: emergent, subaqueous, and subaerial (Godchaux et al., 1992). Emergent volcanoes began erupting under water and built themselves

up above the lake level; subaqueous volcanoes never built up above the lake level, or did so only briefly; subaerial volcanoes formed when magmas intercepted buried aquifers and interacted explosively with water (Fig. 14). Most emergent and subaqueous volcanoes are tuff cones and most subaerial volcanoes are maars or tuff rings. Each type of volcano has three components: basalt massive deposits, overlying bedded tuffs, and overlying or cross-cutting late magmatic deposits.

Emergent volcanoes typically are large, relatively symmetrical, tuff cones with complex walls exhibiting multiple unconformities; collapse sectors are generally limited to the central portions of these edifices and, in many cases, to the upper parts of the stratigraphic sections. Proximal, medial and distal facies are all equally well-developed, though they are not equally well-preserved. Phreatomagmatic planar-bedded deposits and late magmatic deposits greatly predominate over basal massive deposits. Some of these volcanoes have broad flat tops and others have irregular tops where late-stage eruptions built small cinder cones on earlier phreatomagmatic edifices, or because thick coatings of welded spatter were draped over the tuff cone walls. Some emergent volcanoes have late-stage dikes arranged in radial patterns and a few of the larger edifices fed lava flows to their surroundings. All the emergent volcanoes probably resulted from relatively prolonged voluminous eruptions, or they started in shallow water, or both.

Subaqueous volcanoes are the least complicated edifices in the western SRP hydrovolcanic field. They never fully emerged, or in any case did not reach heights at which their eruptive conduits were sealed off from lake water. Apparently they had such low eruption rates or brief eruptive periods that they could not build up out of even fairly shallow water. Also, they may have started out in much deeper water or collapsed repeatedly. They are small asymmetrical tuff cones, or more commonly are accumulations of material preserved within craters. They have relatively simple interior deposits that exhibit only proximal facies materials. Basal massive deposits greatly predominate over other kinds of deposits. Bedded tuffs are rare and are commonly steeply dipping and broken by a multitude of small faults. Late magmatic products include many small dikes and thin, discontinuous, steeply-tilted, columnar-jointed lava flows. The underlying lake sediments commonly are profoundly disturbed.

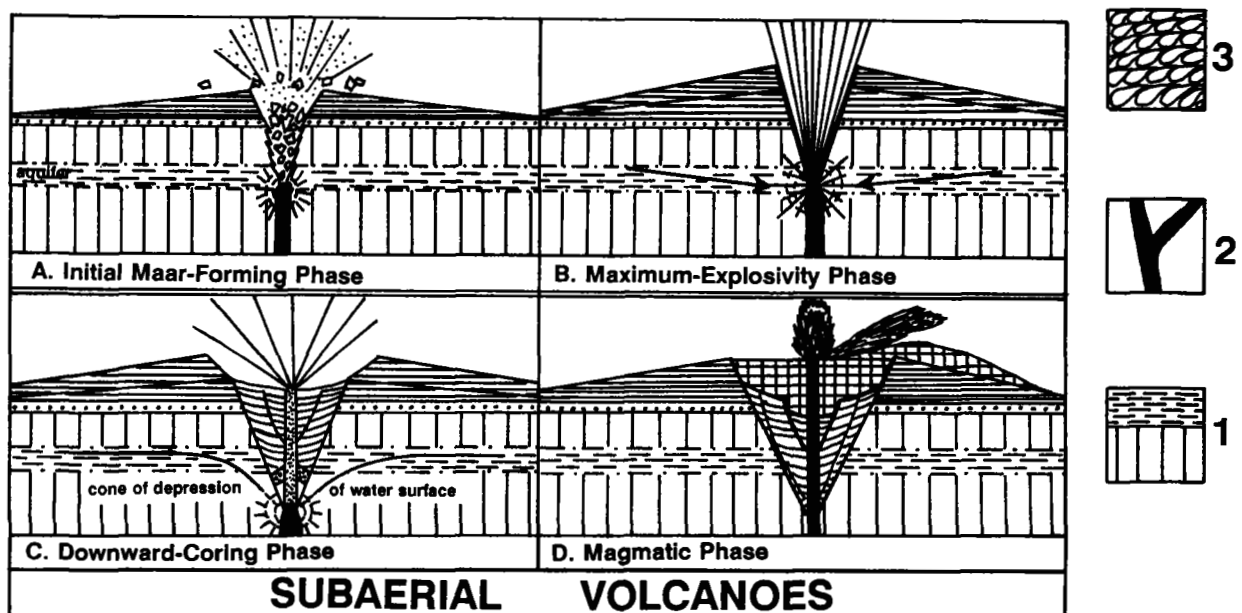
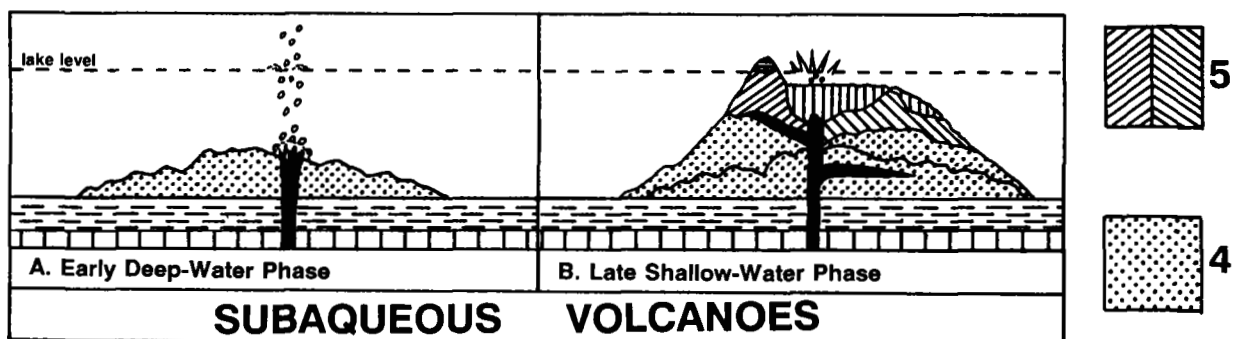
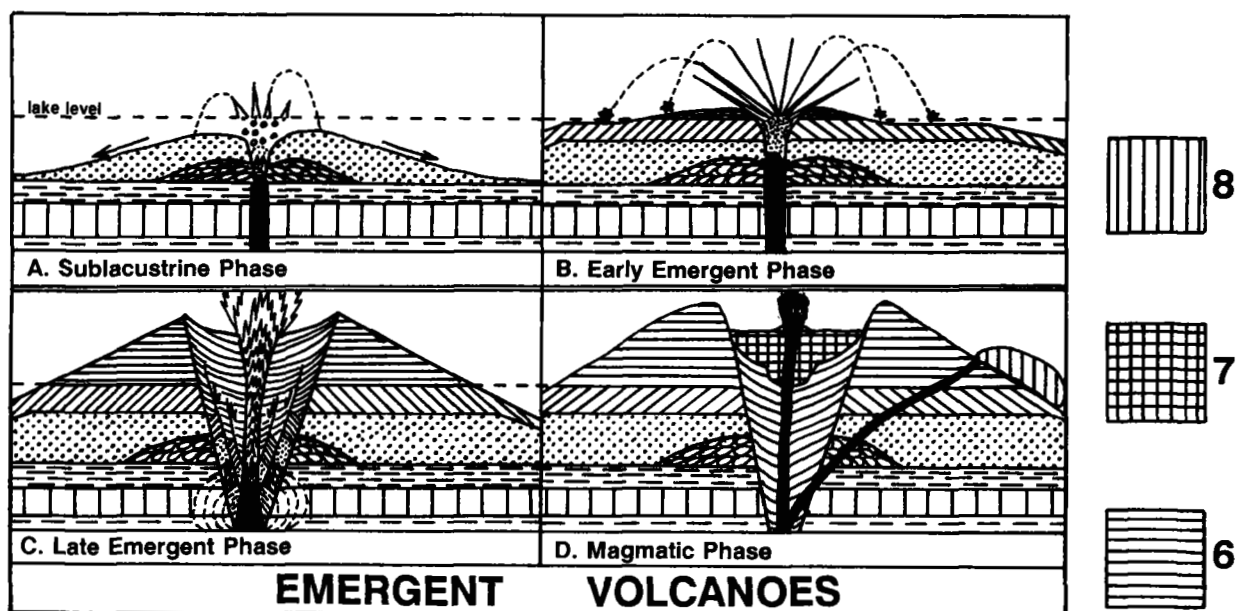
Subaerial volcanoes in the western SRP hydrovolcanic field are maars or tuff rings with deep central craters. Some are almost completely filled by post-eruption sediments. In general, the edifice walls did not collapse, but they are locally offset by small antithetic normal faults. The proximal facies tends to be well-preserved but the medial and distal facies are rarely preserved. Basal massive deposits form wedges or tongues extending outward from the craters in all directions and bedded tuffs make up most of the edifices.

Figure 14. Evolution of emergent, subaqueous and subaerial volcanoes in the western Snake River Plain hydrovolcanic field, from Godchaux et al., 1992. Map patterns: 1 = lake deposits (sediments and lava flows); 2 = dikes; 3 = pillow lavas, hyaloclastite breccias and accidental materials; 4 = basal massive deposits, characterized by slumping and debris-flow morphology; 5 = sublacustrine bedded tuffs (map pattern does not represent dip of beds); 6 = subaerial bedded tuffs (map pattern does not represent dip of beds); 7 = welded spatter and/or cinders; 8 = subaerial lava flows.

Phases: Emergent volcanoes. (A) Sublacustrine Phase—Pillow lavas and hyaloclastite breccias build up a small mound on lake floor. As slopes of mound become steeper, slumps and debris flows occur. (B) Early Emergent Phase—Basal massive deposits are planed off to wave base, and sublacustrine bedded tuffs are deposited atop wave-cut bench. Subaerial deposits form low, incomplete or easily breached cones, and lake water continues to have free access to vent. Intermittent explosions occur. (C) Late Emergent Phase—A large, steep-walled tuff cone is built, sealing vent off from direct contact with lake; water percolating downward triggers energetic explosions; minor downward-coring occurs, and early-formed tuffs subside into crater. (D) Magmatic Phase—Water is entirely excluded, magma rises into crater, producing fire-fountaining; radial dikes feed subaerial lava flows.

Subaqueous volcanoes. (A) Early Deep-Water Phase—Initial eruption produces irregular pile of slumped hyaloclastite breccia. (B) Late Shallow-Water Phase. Basal massive deposits build up within a few tens of meters of the lake surface; explosive eruptions produce subaqueous bedded tuffs and very minor amounts of subaerial bedded tuffs; small sills and dikes are intruded into deeper parts of volcano; small spatter accumulations on walls of crater flow back down toward vent and/or down sides of volcano.

Subaerial volcanoes. (A) Initial Maar-Forming Phase. Magma rises into confined aquifer, producing initial "wet" explosions and building small tuff ring. (B) Maximum-Explosivity Phase—Steady-state flow through aquifer supplies water to magma in near-ideal proportion; continuous energetic explosions keep vent clear of debris and prevent subsidence of early-formed tuffs. (C) Downward-Coring Phase—Water is drawn down faster than the recharge rate; explosion focus migrates downward, and early formed tuffs subside, forming larger maar with narrower central conduit for explosively ejected materials. (D) Magmatic Phase—Hot, relatively volatile-rich magma rises to the surface, producing fire fountains which are higher and denser than those of emergent volcanoes; vent blockages produce inclined collimated jets; welded spatter and/or cinders fill crater and cover portions of outer slopes of volcano.



Near-vertical basalt dikes are common and many are arranged tangential to the volcanoes. Variable amounts of welded and non-welded cinders and spatter form the upper portions of many of the eruptive sequences. Several maars have small lava or spatter ponds in their centers, indicating that the supply of magma outlasted the water supply during eruptions.

Composition of Western Snake River Plain Basalt

The first basalt eruptive episode (9 to 7 Ma) mainly produced basalts with olivine tholeiite compositions, similar to basalts within the Bruneau-Jarbridge and Owyhee-Humboldt eruptive centers. Basalts erupted during the second episode (2.2 to 0.1 Ma) vary more widely, from compositions like those of the first episode to basalts that are highly enriched in Fe, Ti, and P, and somewhat depleted in Si, and Al. Altogether, the basalts in the second episode are quite evolved in comparison to ordinary SRP olivine tholeiites. Figure 15 compares 71 analyzed basalts, mostly from the second (2.2–0.1 Ma) episode in the western SRP graben with 71 basalts from the Bruneau-Jarbridge and 56 from the Owyhee-Humboldt eruptive centers. This plot clearly shows the more evolved nature of the basalts from the western SRP graben. Basalts belonging to the second episode with more than 18% Fe_2O_3 and more than 4% TiO_2 have been found in the western SRP graben. Within the second basalt episode there is a tendency for the younger eruptions to have higher K concentrations (Fig. 16), and also higher Si and Na concentrations (Othberg et al., 1995a). It has not yet been clarified if this effect is due to contamination of the basalt by granitic material or silicic volcanic ash from the lake beds through which the magmas rose, or was due to compositional variations in the mantle basalt source (Vetter and Shervais, 1992). Major element analyses of basalts from most of the localities that are discussed in this field guide are compiled in Table 2.

Geology and Petrology of the Graveyard Point Sill

by Craig White

The Graveyard Point sill is a differentiated tholeiitic intrusion of late Miocene age (6.7 ± 0.4 Ma; Ferns, 1989) that was emplaced into middle Miocene tuffs and tuffaceous sedimentary rocks over an area of about 20 km² in eastern Malheur County, Oregon. It is a recognizable unit on the regional geologic map of Kittleman et al., (1967) and is a prominent feature of the 1:24,000-scale geologic map of Ferns (1989). Chemical analyses of the chilled margin (Table 3) indicate the initial sill magma was a quartz-normative tholeiite similar to basalts erupted along the northern edge of the Owyhee Mountains in Idaho. Hart (1985) classified these lavas as "transitional basalts" (TB) and noted

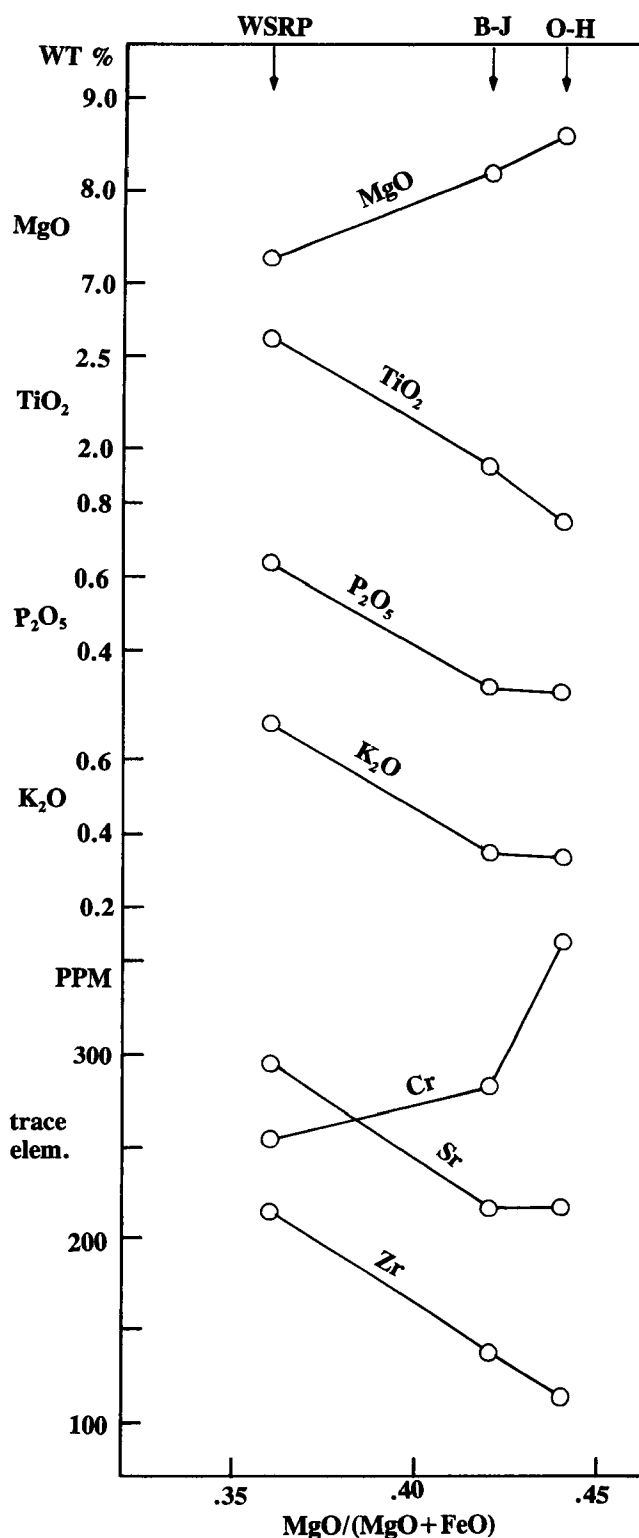


Figure 15. Comparison of average chemical compositions of 71 basalts from the western Snake River Plain graben (WSRP), with 71 basalts from the Bruneau-Jarbridge eruptive center (BJ) and 56 basalts from the Owyhee-Humboldt eruptive center (OH).

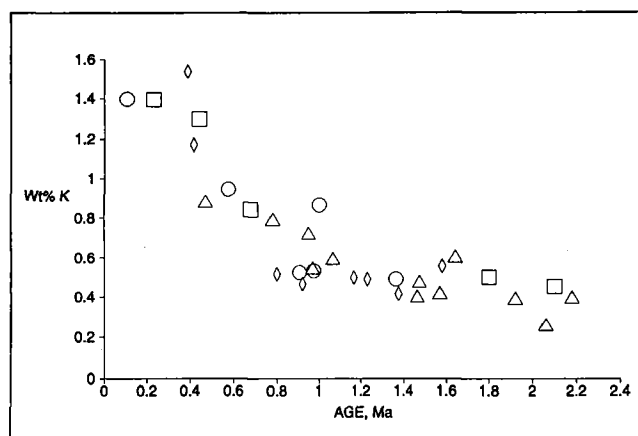


Figure 16. Percent of potassium plotted against age of dated basalts from the second basaltic volcanism episode in the western Snake River Plain, from Othberg et al., 1995a. Data from following sources: circles, K values and K/Ar dates from Othberg et al., 1995a; diamonds, unpublished. K analyses by B. Bonnicksen and Ar/Ar dates from Othberg et al., 1995a; squares, K values and K/Ar dates from Howard et al., 1982; triangles, K values and K/Ar dates from Amini et al., 1984.

that they are intermediate in composition between the MORB-like high-alumina olivine tholeiites (HAOT) of the northwestern Basin and Range and the chemically and isotopically evolved olivine tholeiite basalts of the Snake River Plain (SROT).

Igneous Stratigraphy of the Sill

The sill is exposed in a series of discontinuous outcrops in the highlands between Succor Creek and Sage Creek in the southern part of the Graveyard Point 7 1/2 minute quadrangle (Fig. 17). The most complete sections through the sill are located near its eastern end where both the upper and lower intrusive contacts are exposed in several places. In this area the sill is wedge shaped and ranges in thickness from greater than 150 m to less than 25 m. A 27-m-thick measured section consists of medium-grained olivine diabase that is uniform in composition except for a slightly greater abundance of olivine crystals near the middle. Geochemical profiles through this section are D-shaped (i.e., increasing concentrations towards the middle of the sill) with regard to MgO and Ni, and the calculated bulk composition is slightly more mafic than the chilled margins. These features suggest that olivine crystals were concentrated by flowage differentiation during intrusion. In contrast, sections greater than 100 m thick are strikingly inhomogeneous in texture and contain a variety of iron- and silica-rich differentiated rocks. The calculated bulk composition of a 155-m-thick measured section is substantially

more differentiated than the chilled margins, suggesting that either large volumes of crystals and/or liquid were transported laterally after emplacement, or that the intrusion was filled in stages, with evolved melt following an initial influx of less differentiated liquid.

The igneous stratigraphy of the 155-m-thick measured composite section is described below and geochemical profiles are shown in Figure 18. Chilled borders are well developed next to most contacts and there are no significant differences between the upper and lower margins of the sill or among samples collected at different locations. In thin section, the chilled rocks consist of fine-grained, intergranular microgabbro with small phenocrysts of plagioclase and olivine. The lower chilled border grades upward into fine- to medium-grained diabase in which olivine and plagioclase are partly or completely enclosed by augite. About 10 m above the floor the rocks become slightly coarser grained and augite first appears as discrete subhedral crystals. The diabase below this level is similar in composition to the chilled border, but the rocks above it have much lower Mg-numbers and Ni contents and slightly greater amounts of incompatible elements. There is no layer of olivine cumulate next to the floor, so the most likely explanation for the abrupt reduction in MgO and Ni is the influx of a second, more chemically evolved magma into the sill.

About 30 m above the floor the rocks take on a moderate to strong planar igneous lamination caused by the shape-preferred orientations of plagioclase and augite crystals. The laminated gabbros are cumulates according to the criteria proposed by Irvine (1982): their textures consist of a relatively compact framework of touching crystals and their chemical compositions indicate they formed by small to moderate degrees of crystal accumulation. They typically contain veins and small dikes of granophyre which cut across the igneous lamination.

Near the middle of the intrusion, about 60 m above the floor, laminated gabbro grades upward into a patchy textured medium- to coarse-grained ferrogabbro. In thin sections this rock consists of a porous framework of plagioclase, monoclinic pyroxene, olivine, apatite and skeletal magnetite surrounding pools of fine-grained interstitial granophyre. Within the ferrogabbro zone there are commonly one or more lens shaped layers in which interstitial granophyre comprises as much as 50% of the total rock. These exceptionally granophyre-rich rocks, which have the bulk composition of a ferrodiorite, are particularly distinctive because they weather to a reddish brown color in outcrop and contain numerous small miarolitic cavities. The interstitial granophyre commonly contains swallow-tailed or hollow-centered crystals indicative of rapid crystallization. The presence of quench textures and miarolitic cavities in these rocks suggests that the sill may have vented to the surface late in its crystallization history, causing the volatile pres-

Table 2. Western Snake River Plain Basalt Analyses.

Number:	1	2	3	4	5	6	7	8	9	10	11	12
Stop:	2-2	2-2	2-2	2-3	2-4	2-5	3-1	3-1	3-2	3-3	3-4	3-4
Sample:	I-2730	I-2732	I-2844	I-2706	I-2712	I-2435	I-3053	I-3054	I-2831	I-2982	I-2985	I-2986
SiO ₂	46.51	46.80	48.09	47.33	48.24	45.43	47.47	47.08	47.94	46.12	46.34	47.56
TiO ₂	3.25	1.92	3.10	3.41	1.50	3.53	1.76	2.10	3.53	3.09	3.12	3.04
Al ₂ O ₃	16.39	15.61	13.75	13.72	16.95	13.88	16.46	15.74	13.82	14.22	14.60	13.97
Fe ₂ O ₃	14.58	12.57	15.45	16.00	11.31	16.82	13.29	13.64	15.23	15.73	15.10	14.61
MnO	0.19	0.19	0.21	0.23	0.18	0.23	0.20	0.21	0.21	0.21	0.27	0.21
MgO	4.89	9.04	6.50	5.80	6.85	6.67	7.93	7.71	5.05	6.97	6.03	6.00
CaO	10.24	10.23	9.10	9.20	11.90	9.42	9.82	10.18	10.18	9.62	11.00	10.50
Na ₂ O	2.87	2.65	2.79	2.56	2.71	2.49	2.80	2.62	2.49	2.55	2.63	2.45
K ₂ O	0.62	0.74	0.73	0.78	0.37	0.62	0.37	0.28	0.71	0.61	0.50	0.61
P ₂ O ₅	0.69	0.43	0.64	0.71	0.21	0.82	0.23	0.31	0.63	0.79	0.74	0.77
Total	100.23	100.18	100.36	99.74	100.22	99.91	100.33	99.87	99.79	99.91	100.33	99.72
Mg#	39.9	58.8	45.5	41.8	54.5	44.0	54.2	52.8	39.7	46.8	44.2	44.9

Analysed by Bill Bonnicksen at Ronald B Gilmore x-ray fluorescence laboratory of the University of Massachusetts at Amherst. Total analyzed iron is expressed as Fe₂O₃, Mg numbers are expressed in atomic percent.

1. I-2730 Guffey Railroad Bridge basalt flow
2. I-2732 Guffey Table basalt flow.
3. I-2844 Basalt dike in Guffey Butte volcanic complex
4. I-2706 Basalt of White Butte.
5. I-2712 Basalt flow from Grouch Drain volcanic complex.
6. I-2435 Sinker Butte basalt flow.
7. I-3053 Basalt flow in Teapot subaqueous volcanic field.
8. I-3054 Basalt dike in Teapot subaqueous volcanic field.
9. I-2831 Welded basaltic spatter from Castle Butte.
10. I-2982 Brooks Ranch basalt flow.
11. I-2985 Basalt flow between tuff units at Rabbit Creek.
12. I-2986 Welded basaltic spatter above tuff units at Rabbit Creek

sure to drop suddenly and initiating both rapid crystallization and vesiculation in remaining interstitial melts.

The iron-rich rocks occur at the base and within the lower part of a series of much less differentiated poikilitic olivine gabbros that typically consist of small (1 mm) subhedral crystals of plagioclase surrounded by much larger (1 cm) poikilitic crystals of olivine and augite. The textures of these rocks and their position in the upper part of the sill suggest they comprise an upper border series which crystallized *in situ* from the roof downward. On average, they contain about 10% more modal plagioclase than the cumulus textured gabbros in the lower part of the sill, which is consistent with the slow settling rate of plagioclase phenocrysts compared to that of comparably sized crystals of olivine and augite. The uppermost 8–10 m of the sill consists of dark gray, fine-grained diabase similar to the MgO- and Ni-rich rocks adjacent to the floor.

Differentiation of the Sill Magma

Despite its relatively small size, the Graveyard Point intrusion contains a remarkably well differentiated suite of

rocks. Isotopic ratios of strontium and oxygen are nearly uniform throughout the entire range of rock compositions, indicating there was little or no assimilation of the silicic wall rocks after the sill was emplaced (White et al., 1995). Computer-mixing programs support a fractional crystallization model for the differentiation of the sill; however, geochemical modeling does not shed much light on how or where the fractional crystallization took place. It is our present opinion that many features of the sill are consistent with the model proposed by Marsh et al., (1990) and Marsh (1996) for the differentiation of shallow magma chambers filled by phenocryst-bearing melts. In this model, the initial phenocrysts are redistributed by sinking or floating but crystals nucleated near the roof after emplacement do not escape capture by the downward growing solidification front. Interstitial liquids may migrate upward through the cumulate pile, resulting in small to moderate shifts in both bulk-rock and mineral compositions. Highly differentiated rocks are confined to segregation veins or lens-shaped layers that form in the middle and upper parts of the intrusion by "sag filter pressing," a process by which intercumulus melts

Table 3. XRF Analyses of Selected Samples from the Graveyard Point Sill.

Stop No. Field No. Wt % oxide	LCB* GP303	2-1A(1) GP7	2-1A(2) GP338	2-1A(3) GP234	2-1A(4) GP328	2-1A(5) GP324	2-1A(6) GP323	2-1A(7) GP321	2-1B GP17	2-1C GP293
SiO ₂	49.34	48.79	49.87	50.20	50.74	54.10	49.11	56.62	64.25	67.96
Al ₂ O ₃	15.87	15.66	13.94	13.58	14.30	11.22	15.95	11.96	12.19	12.77
TiO ₂	1.86	1.90	2.67	2.40	1.90	2.75	2.50	2.31	1.25	0.90
FeO	10.91	11.41	12.80	11.16	15.52	16.25	11.73	14.72	9.95	6.93
MnO	0.19	0.19	0.22	0.20	0.25	0.28	0.19	0.25	0.14	0.13
CaO	11.22	11.20	11.22	12.70	7.26	6.75	11.71	6.01	3.69	2.61
MgO	7.57	7.58	5.58	6.49	5.12	1.95	5.28	1.27	0.37	0.56
K ₂ O	0.37	0.43	0.67	0.47	0.94	1.93	0.54	2.08	3.22	3.68
Na ₂ O	2.40	2.54	2.61	2.47	3.46	3.60	2.68	3.87	4.63	4.26
P ₂ O ₅	0.28	0.29	0.41	0.32	0.51	1.19	0.32	0.90	0.31	0.19
ppm										
Ni	90	83	24	28	26	1	28	1	4	8
Cr	267	255	241	327	3	1	213	1	0	4
Sc	41	44	56	64	30	34	45	41	19	17
V	311	301	402	433	106	49	363	42	0	2
Ba	311	277	431	276	509	1016	415	1144	1673	1677
Rb	5	4	9	7	13	30	7	35	58	65
Sr	240	233	229	217	253	211	253	214	167	121
Zr	111	118	158	115	205	445	125	478	745	748
Y	32	31	45	35	50	113	33	119	143	135
Nb	11	13	17	12	21	46	14	48	60	55

Note. Analyses were performed at the Geoanalytical Laboratory in the Geology Department at Washington State University and are recalculated to 100% (water free). Total analyzed iron is expressed as FeO.

*LCB is an analysis of the lower chilled border.

are drawn into sub-horizontal tears in the rigid, but not completely crystallized, upper solidification crust.

Figure 19 shows the compositions of rocks from various parts of the Graveyard Point sill plotted on the normative Ol-Di-Qz ternary diagram. The chilled margin and most of the rocks adjacent to it plot below the Ol-Di cotectic, reflecting the composition of the initial pulse of magma. Diabase from the middle of the 27-m-thick measured section and the lower border of the 155-m-thick measured section plot slightly closer to the Ol corner owing to the presence of small amounts of accumulated olivine. Laminated cumulates in the lower-middle part of the sill plot above the cotectic, within the primary phase field of pyroxene, presumably because they accumulated crystals of augite. Lenses and dikes of granophyre represent samples of differentiated liquid extracted from the cumulate pile and plot directly on the cotectic. It is noteworthy that both the petrography of the cumulates and the geochemical trend defined by the rocks indicate that augite was an important fractionating phase, although it is conspicuously absent as a phenocryst in all samples of the chilled margin. If the

Marsh model is correct and only pre-existing phenocrysts are effectively removed by settling, then the sill must have been emplaced either in several stages or as a compositionally zoned body.

Stops for Day 2

Nampa to Graveyard Point

Drive west from Nampa to the town of Homedale, ID, near the Idaho-Oregon border (Fig. 3). Beginning from downtown Nampa, at the intersection of S12th Avenue and Nampa-Caldwell Blvd., drive northwest on Nampa-Caldwell Blvd. for 2.8 miles. Turn left (west) onto State Route 55. Proceed west on State Route 55 for 9.0 miles, then turn right at the sign for the tiny community of Huston and drive north on Pride Lane. After 1.0 mile turn left (west) on Homedale Road and drive for about 7.2 miles to the intersection with State Route 95. Turn left (south) on Route 95 and immediately cross the Snake River and enter the town of Homedale. From downtown Homedale continue south on State Route 95 for 3.0 miles, then turn right (west) on

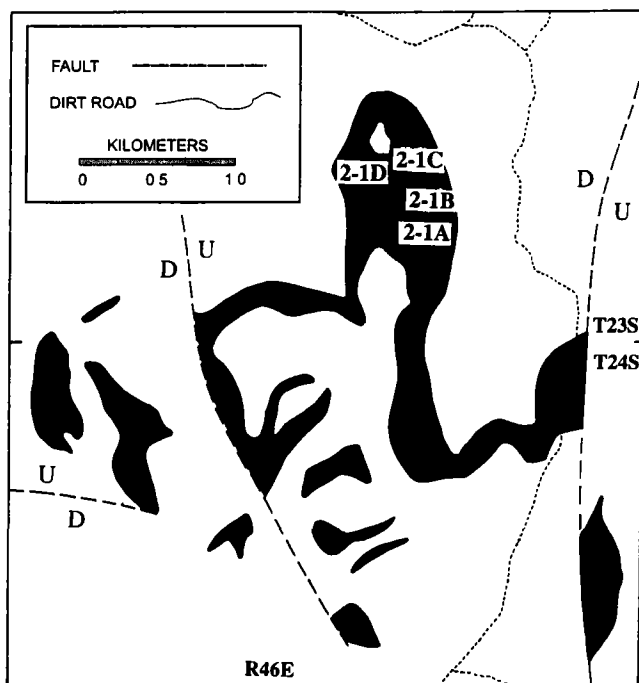


Figure 17. Map showing the areal extent of exposures of the Graveyard Point sill (after Ferns, 1989) and the general locations stops 2-1A through 2-1D.

Graveyard Point Road. This intersection is just north of where State Route 95 makes a bend to the southeast. Go west on Graveyard Point Road for 3.8 miles to the T-intersection with the paved Sage Road (Graveyard Point Road continues but is gravel beyond this intersection). Turn left (south) on Sage Road and drive for 0.8 miles to a Y-intersection where Sage Road veers to the left and a dirt road goes right, crossing a small bridge over an irrigation canal. Take the right turn across the bridge. CAUTION: the dirt road from here to the sill is deeply rutted and requires a high clearance vehicle; it may be impassable after heavy rains.

The following mileages are measured from the beginning of the dirt road; however, they may vary somewhat depending on which of the several tracks are taken. After crossing the irrigation ditch, drive west on the deeply rutted "main" dirt road, passing beneath power lines at about 1.0 mile. At about 1.6 miles the road curves to the northwest, goes around the nose of a hill, then heads south-southwest. The area between here and the sill is a favorite agate-digging locality and a number of pits have been excavated by local rock and mineral clubs. Dirt tracks head off in all directions; generally keep to the left on the most well traveled road and continue driving south. At approximately 4.5 miles the road begins to pass just east of cliffs in which the brown weathered gabbro of the Graveyard Point sill is well

exposed. At about 4.7 miles turn right onto a dirt track that leads up hill to the mouth of a dry gully cut into the east-facing escarpment. A level spot just east of the rock exposures provides parking space for 3 or 4 vehicles. We will spend about 3 hours examining the sill; all outcrops will be accessed by foot from here.

Stop 2-1A. Measured Section Through the Graveyard Point Sill

Walk to the outcrop at the mouth of the narrow gully just west of the parking area. This will be our first stop. From here we will hike up the gully for about 0.3 mi, examining the lower two-thirds of the section illustrated in the geologic column in Figure 18. Because the freshest exposures are relatively small and are located at the bottom of a narrow gully, this traverse is designed to be self-guided and self-paced. The general route is marked with orange flagging; the numbered outcrops described below are identified with blue flags.

2-1A(1): The small smooth outcrop at the mouth of the gully consists of fine-grained olivine diabase that is probably only 1-2 m above the lower contact of the sill. It contains a few small phenocrysts of olivine (Fo_{79}) and plagioclase (An_{81}) and has a chemical composition that is typical of the Ni- and MgO-rich rocks exposed in the first 9-10 m of the section. The lower contact and chilled border of the sill are exposed about 0.25 miles south of here.

2-1A(2): This outcrop marks the first appearance of gabbro in which pyroxene is present as distinct subhedral crystals (primocrysts) rather than anhedral poikilitic plates. The rocks from here upward also are more chemically evolved than the diabbases in the lower 9-10 m of the sill. No textural evidence for a recharge event has been found in outcrop; however, in thin section this rock contains distinctive sector-zoned augite and strongly zoned plagioclase suggesting disequilibrium crystallization.

As you walk up the gully look for boulders of black obsidian-like rock and flat pieces of float containing unusual mud-crack-like fractures. These are fragments of the contact-metamorphosed roof that have washed down from above. X-ray diffraction has shown that some of these rocks contain tridymite and cordierite (Ferns, 1989).

2-1A(3): Gabbros first become distinctly laminated at about this level owing to the alignment of lath-shaped crystals of plagioclase and elongate prisms of augite. The rock here is a cumulate containing primocrysts of plagioclase (An_{62}), augite (En_{44}, Fs_{15}), olivine (Fo_{52}) and magnetite, and small amounts of interstitial apatite and granophyre. Granophyre veins up to a few cm wide are well exposed on north side of the gully. The next 25 m of section consists of laminated

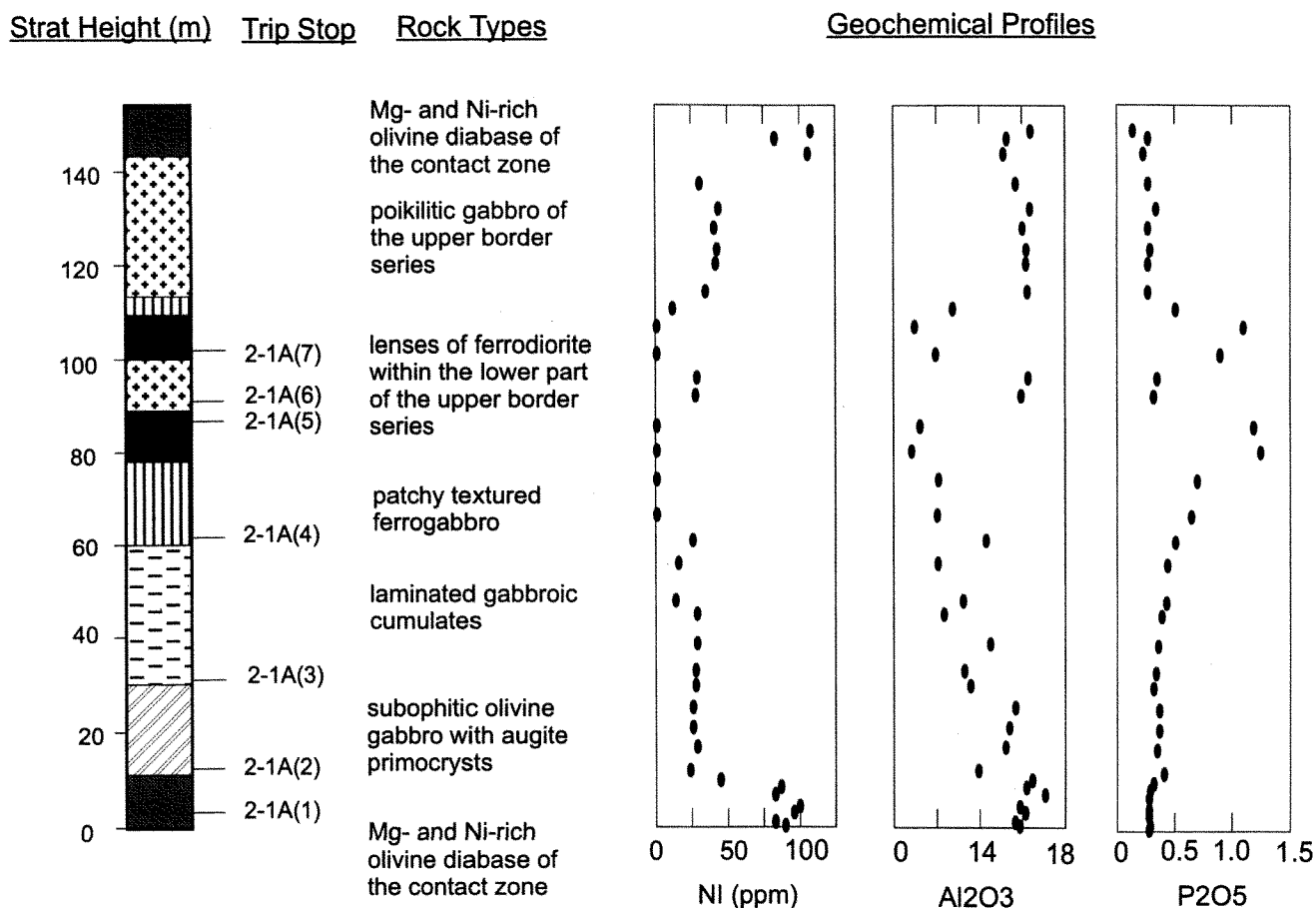


Figure 18. Schematic geologic column showing the igneous stratigraphy and three whole-rock geochemical profiles through the 155-m-thick measured section of the Graveyard Point sill.

gabbro containing increasingly greater amounts of interstitial granophyre.

2-1A(4): At this level, near the middle of the sill, the relatively uniform sequence of laminated gabbros grades into a texturally inhomogeneous series of ferrogabbros with iron contents (FeO^T) greater than 15 percent. Pegmatoidal clots containing crystals of plagioclase and augite up to 4 cm long are present in the outcrop at this location

2-1A(5): The rock here is typical of the granophyre-flooded ferrodiorite that forms laterally discontinuous lenses within the upper half of the intrusion. In outcrop the rock has a brecciated appearance that is best seen on weathered surfaces. Thin sections contain elongate crystals of plagioclase (An_{40}), augite ($\text{En}_{35}\text{Fs}_{28}$) and apatite, and skeletal crystals of Fe-Ti oxide, within a matrix of fine-grained granophyre and isotropic limonite.

2-1A(6): The rounded and polished exposure at this stop contains poikilitic gabbro typical of the upper border se-

quence. The rock here contains large, optically continuous crystals of augite and olivine enclosing smaller, subhedral laths of plagioclase.

2-1A(7): A second layer of ferrodiorite is exposed at this stop. Analyses of several samples from this zone show that it differs in composition from the ferrodiorite layer below, indicating that we are not just encountering a structural repetition of the granophyre-rich unit below. We will leave the gully at this point and walk up the hill to the right (north). If you were to continue walking up the gully you would pass back into poikilitic gabbro of the upper border series. The roof and uppermost rocks of the sill are well exposed on east-facing cliffs about 0.2 miles to the south.

Climb out of the gully to the north. Exposures of ferrodiorite on the side of the gully provide a good opportunity to examine the interesting textures of these rocks. Continue to the top of the hill and walk north along the ridge for about 0.2 mi, keeping slightly to the left (west) side of the ridge crest. Look for an irregularly shaped pod of grano-

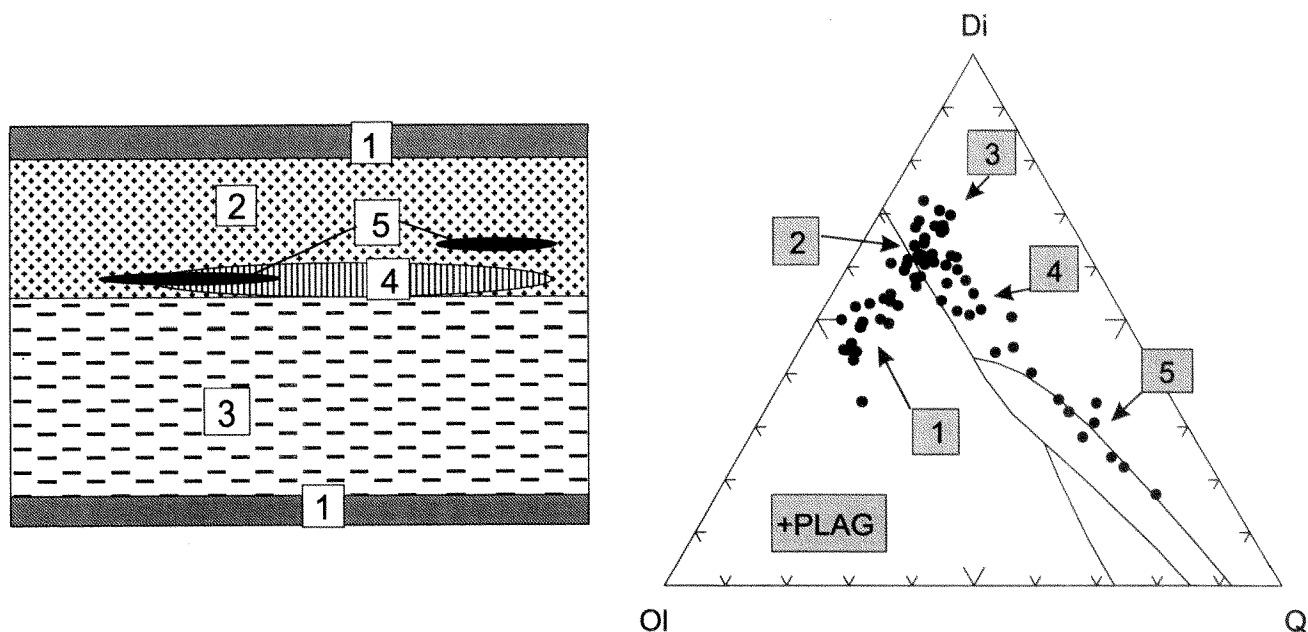


Figure 19. Analyses of samples from the Graveyard Point sill are plotted on the low-pressure, plagioclase-saturated pseudoternary diagram (Grove and Baker, 1984). The general positions of samples within the intrusion are indicated on the schematic section: (1) border zones within 10 m of the contacts (includes samples from the 27-m-thick measured section); (2) poikilitic-textured gabbros from the upper part of the sill; (3) laminated gabbros from the lower part of the sill; (4) ferrogabbros near the middle of the intrusion; (5) ferrodiorites and granophyres from small lenses and sheets.

phyre that crops out over an area of approximately 6 X 9 m. The exposure can be recognized from a distance by its dark reddish purple color which contrasts with the browner colored rocks around it.

2-1B: Granophyre Pod in the Upper Part of the Sill.

This is the largest continuous body of granophyre identified within the sill. Granophyre more typically occurs as interstitial patches in ferrodiorite or as 1–30-cm-thick veins and dikes. This rock has a silica content of 64% and consists of small crystals of sodic plagioclase, ferroaugite, magnetite, and apatite in a fine-grained matrix of intergrown quartz and alkali feldspar.

Walk northwest and slightly down hill for about 0.1 mile across nearly continuous exposures of ferrodiorite and ferrogabbro to a dry gully that heads down hill to the southwest. A spectacular meshwork of granophyre dikes is exposed in the walls of the gully.

2-1C: Meshwork of Granophyre Dikes

The granophyre dikes exposed here and further down the gully are segregation veins that formed when late-stage interstitial liquids were drawn into fractures within rigid, but not completely solid, crust. Note that many of the dikes

have gradational contacts with their wall rocks and some appear to be just solidified sheets of crystal mush rather than intrusions of melt. Minerals in the granophyre dikes include sodic plagioclase, ferroaugite, magnetite, apatite, pigeonite, and fayalitic olivine (Fo₁₀), all surrounded by quartz and alkali feldspar, commonly in micrographic intergrowths. Silica contents in the dikes at this locality range from 49–68%.

Walk down the gully about 0.1 mi, crossing two 0.3 m-thick granophyre sheets, to a small, flat, sandy area formed at the confluence of three dry washes.

2-1D: Contact Metamorphosed Drop Block or Roof Pendant

Olivine gabbro at this locality surrounds a 30-m-long drop block or roof pendent of contact metamorphosed tuffaceous sediment. A chilled border of fine-grained diabase extends outward for several meters from the contact. Note that the prominent amygdale trains in the chilled rock appear to be oriented parallel the contacts with the block. There is no evidence in either the outcrop or the bulk rock analyses to suggest the block melted to any significant degree; however, the amygdale trains indicate that volatiles were driven out of the sediment and into the magma during contact metamorphism.

This is the last stop. Retrace our route back to the parking spot.

Graveyard Point to Guffey Butte Area

After visiting the Graveyard Point sill, return to State Route 95 via the Graveyard Point road (Figs. 3, 20). At State Route 95 turn south and, at the junction with State Route 55, take 55 to Marsing. At Marsing take State Route 78, which generally follows the course of the Snake River upstream toward the southeast, to its junction with Highway 45 near Walters Ferry. Along the way, the Owyhee Range can be seen to the southwest. The northern portion of these mountains contains several of the older rock units discussed earlier, including late Cretaceous granitic rocks, the Oligocene Salmon Creek volcanics, Miocene pre-SRP rhyolite and basalt, and rhyolite lava flows that formed during the first stage of western SRP graben development. The northeastern margin of the range is downfaulted, but most of the faults have been buried by detritus eroded from the mountains. The valley fill is mainly lake beds and post-lake valley fill that contains some interspersed, but buried, basalt flows. Across the river to the northeast, all along this route, there is a prominent scarp in which Pliocene and Pleistocene lake beds (the Glens Ferry Formation) are capped by various basalt flows. This escarpment may have resulted from a buried NW-SE-trending fault or series of faults. Or, this escarpment simply may be the result of the resistant basalt cap protecting the sediments beneath it from erosion, while unprotected sediments were eroded out of the valley. Abundant hot water at depth along this zone, and even surface hot springs such as those around Givens Hot Springs, generally support the idea that buried faults might be important in this area.

At the junction of Highways 78 and 45, take 45 toward Nampa. Proceed north across the river, to just past the Walters Ferry store, and turn right onto Ferry Road and follow it eastward. After 2.0 miles, Ferry Road ends in a T-intersection with Hill Road. Turn right (south) on Hill Road and proceed for 2.2 miles to Sinker Road (somewhere along here Hill Road becomes Warren Spur). Turn right on the gravel Sinker Road and drive for 3.8 miles to the old Guffey Railroad Bridge. The prominent reddish-brown colored butte just east of the junction of Ferry and Hill Roads is Walters Butte. It is a multistage, partially eroded or sector-collapsed, tuff cone, one of several discussed by Godchaux et al., (1992). It is capped by welded basaltic spatter, and sits on a base of lake sediments that are visible part way up the construct. As Hill Road is followed south we will pass another small butte, White Butte, which will be stop 2-3.

Sinker Road follows the route of the railroad that once connected Nampa and Murphy. Once you have driven past

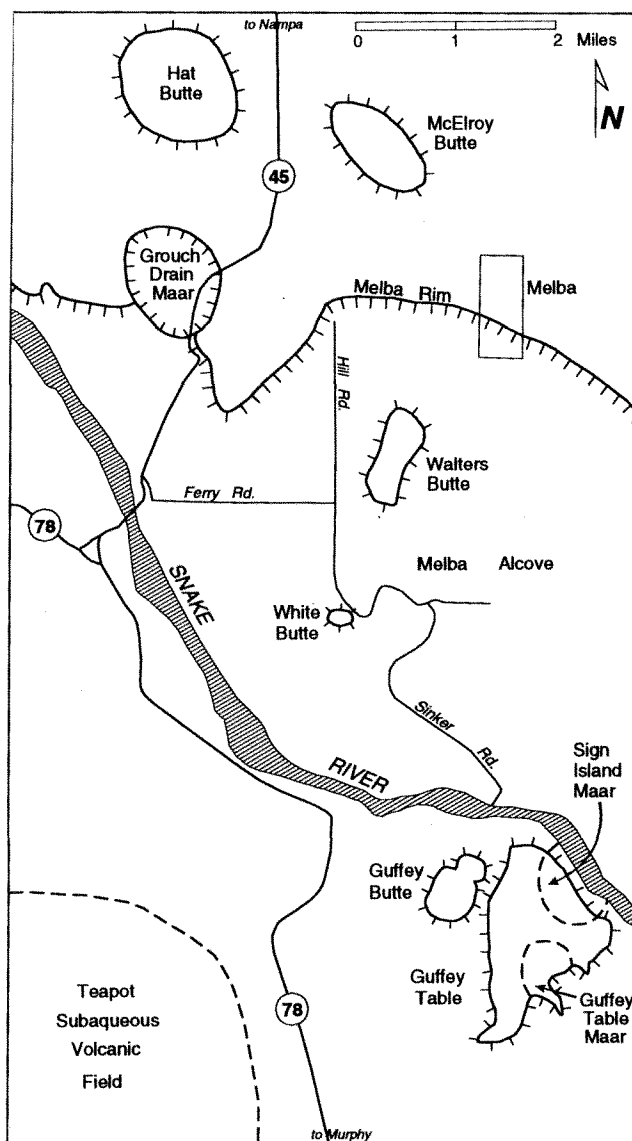


Figure 20. Map showing the location of geographic and volcanic features in the Guffey Butte-White Butte-Grouch Drain area.

White Butte, for the last couple of miles before reaching the river, the road crosses a very large boulder and gravel bar that was deposited during the Bonneville Flood (O'Connor, 1993 and Malde, 1968). Near the river, the road is in a cut through this boulder bar, the sides of which give a very clear impression of the size of particles the Bonneville Flood was able to transport. Guffey Butte is just across the Snake River from Celebration County Park via the old Guffey Railroad Bridge. This bridge, built in 1897, has been converted into a pedestrian and bicycle bridge and provides easy access to Guffey Butte and to Guffey Table, the large flat-topped hill farther east (Fig. 20).

Stop 2-2: Guffey Butte Area

Many geological aspects of the western SRP can be read from the volcanic and sedimentary rock layers at Guffey Butte and Guffey Table (Fig. 20). Before about 2 Ma ago Lake Idaho waxed and waned in size as the SW Idaho climate varied and as tectonic events caused the sinking of the basin and uplift of the adjacent mountainous areas. During this time the lake continued to receive sediments from streams and rivers around its margins. Silt beds, accompanied by sand lenses and clay layers typical of sediments deposited in Lake Idaho, are exposed in the slopes of Guffey Butte. During this early part of Lake Idaho's existence, while the lake stood at a high level, basalt was erupted onto the lake floor and built piles of underwater pillow basalt and formed underwater lava flows. Material of this sort is widespread a few miles west and south of Guffey Butte, and small amounts of underwater flows occur near the base of Guffey Butte on its north side, and at the east end of Guffey Table. As the lake drained for the last time, starting perhaps about 2 million years ago, the present western SRP volcanic landscape began to form. The sediments being deposited in the Guffey Butte area became coarser, changing from clay and silt to sand and gravel beds, as the lake shrank and flowing streams extended further into the lake basin.

Basaltic volcanism resumed as the Guffey Butte area was changing from submerged under water to newly emergent land. The landscape probably was an uneroded nearly flat plain crossed by slow-moving streams and estuaries of swampy ground that drained into the remnant of the lake. The lake and stream sediments would have been completely saturated with groundwater at that time. Basalt lava erupted and then spread out across this swampy land, damming some of the streams and even flowing into the lake. A basalt flow from this period is visible as the lowest cliff on the northern end of Guffey Table. This flow has a blotchy appearance with many relatively soft, easily eroded, internal zones rather than being fresh, solid basalt. Its appearance is the result of the basalt soaking up water while it was still hot, which caused softer, more easily eroded, minerals to form than those in ordinary air-cooled basalt. This water-affected basalt, or WAB, flow probably is about 2 Ma old and has been named the Guffey Railroad Bridge basalt.

As Lake Idaho drained, streams and rivers began to erode gullies and valleys in the exposed lake sediments. This gradually lowered the groundwater table. During volcanism at this time basalt magma came up as dikes along vertical fractures and encountered groundwater that was trapped by the impervious overlying materials a few tens of feet below the surface. Some of this water was superheated to steam and reached relatively high pressures. Eventually, the confining strength of the overlying material was exceeded.

This caused explosive blowouts of the near-surface sediments and the formation of craters, or maars, excavated in the bedrock.

Explosions that occurred when magma was injected into the limited supply of groundwater trapped beneath the older basalt-flow cap excavated at least three major craters in the Guffey Butte area. One is Guffey Butte itself, and the others are the Sign Island and Guffey Table maars that have been partially covered by the Guffey Table basalt flow. Both the Sign Island and Guffey Table maars were partially filled by spatter cones, as was also the case for Guffey Butte itself. The tuff rings surrounding these maars consist of sediment and older lava fragments blown out of the crater, and are accompanied by a small amount of fragmented juvenile lava. The explosive eruptions started with the blowing out of the material lying above the steam accumulation. Some of the earliest ejected rocks adjacent to the craters are several feet across. Once a crater was formed, lava was injected from below while groundwater percolated into it from the sides. This caused a series of explosions in which small lava blobs, probably lapilli-sized, were mixed with sediment and blocks of the overlying basalt layer and were hurled out of the crater onto the developing tuff ring. Many of these deposits are distinctly layered; some even have crossbeds generated by deposition from the passage of sediment-bearing steam surges. Such bedded tuff deposits lie above the basal massive deposits from the initial blowouts and beneath the deposits formed largely from later lava-spatter eruptions. One of the most conspicuous of these bedded zones is visible southwestward from Guffey Railroad Bridge on the steep area on the northwest part of Guffey Butte. (Binoculars are helpful for this view.) As more lava entered the craters and consumed all the available water in producing the last of the explosive surges, the nature of the eruptions changed. Larger lava blobs were emitted less explosively and built a ring of coalesced spatter on the tuff. Such spatter deposits are nearly 100 feet thick on the west side of Guffey Butte.

As the canyons throughout the western SRP were deepened, the top of the groundwater zone was lowered. For basaltic magma that rose at this time the greater depth of the groundwater kept it from forming steam pockets with enough pressure to form blowouts. Instead the basalt pushed its way to the surface, where subaerial lava flows and spatter cones, cinder cones, and shield volcanoes formed. At Guffey Table, a small lava flow erupted at the south end of the area after the groundwater level had been lowered. This eruption formed the Guffey Table basalt flow about 1.0 Ma ago.

Guffey Butte is a well-preserved maar with one of the deepest present-day central craters of any maar in the area. The maar shape was considerably modified before the final Strombolian phase of the eruption began, however, and

the inner walls are liberally coated with welded spatter. A late-stage fire fountain can be recognized high on the southwest inner-crater wall. The basal massive deposits contain a spectacular array of blocks and bombs in a matrix of extremely finely comminuted recycled lake sediment. The bedded tuffs exhibit surge cross-bedding, providing further evidence that this eruption was extremely violent. The relative thickness and large aerial extent of the welded spatter layer implies that the water was used up before magma stopped rising in the conduit, so that the late-magmatic phase of the eruption was fairly long-lived (Godchaux et al., 1992).

Spatter that ponded within the crater apparently caused extensive melting of silicic volcanic-rich sediments beneath and adjacent to it (Table 4). A felsic dike of mixed solid sediment fragments and melted silicic material, in the shape of a thin cone sheet, was intruded along the inner crater wall. Flow streaks suggest that it may have formed at the base of the spatter pond and was displaced upward by the weight of the overlying pond. This silicic dike may have vented at one place to form a tiny lava accumulation along one side of the crater. Petrographic studies of the dike materials have shown that glasses of various compositions became commingled, but did not thoroughly mix, before they were quenched. Some of these quenched glasses have hybridized basaltic compositions (Morrow, 1996).

Melba alcove, the wide valley that stretches almost all the way northward from Guffey Butte to the town of Melba, was eroded after the Guffey Table lava flow erupted. When volcanism resumed about 0.4 Ma ago (Othberg et al., 1995a), it produced lavas that flowed from the north and east into the alcove and extended all the way to Guffey Butte. These flows primarily came from Initial Point volcano, Kuna Butte, Powers Butte, and several smaller eruptive points lying within a few miles of Kuna Butte. They cascaded down the Melba rim onto the alcove floor. They must have dammed up the ancestral Snake River for a time, until it cut through, making a new canyon where the Snake River is located today. The cliffs just north of Celebration Park are carved in these 0.4 Ma old lavas and represent the final stage of Snake River incision.

The Bonneville Flood was the last major geologic event to affect the Guffey Butte area. When Lake Bonneville overtopped and eroded down its outlet in SE Idaho, a short-lived, but very high-volume, torrential flood resulted, that followed the course of the Portneuf, Snake, and Columbia Rivers to the ocean. This flood lasted only a few weeks but had a maximum flow rate greater than that of the Amazon River. Since its channel is fairly steep, the water flowed rapidly and carried large boulders for long distances. One of the most impressive examples of this is Walters bar, the enormous boulder and gravel bar that extends northwestward from Guffey Railroad Bridge for several miles. Many large boulders several feet across are well displayed just

Table 4. Analyses of melted sediments at Guffey Butte.

Number:	1	2
Sample:	I-2835	I-2842
SiO ₂	79.13	77.80
TiO ₂	0.23	0.22
Al ₂ O ₃	11.06	11.41
Fe ₂ O ₃	1.58	1.46
MnO	0.02	0.02
MgO	0.45	0.61
CaO	1.53	2.14
Na ₂ O	2.51	2.80
K ₂ O	3.65	3.40
P ₂ O ₅	0.09	0.10
Total	100.25	99.96
Mg# (At%)	36.1	45.3

Analyzed by Bill Bonnicksen at Ronald B. Gilmore x-ray fluorescence laboratory of the University of Massachusetts at Amherst. Total analyzed iron expressed as Fe₂O₃.

1. I-2835 Glassy rock from felsic dike cutting crater fill.

2. I-2842 Vesicular flow-like felsic rock from crater margin.

north of the bridge. The largest boulder measured on this bar has an intermediate diameter of 4.6 m and an estimated mass of 140 tons (Othberg et al., 1995b).

Stop 2-3: White Butte

White Butte is a runty little volcano; the entire edifice measures only 400 m across and it stands only 63 m high. White Butte stood directly in the path of the Bonneville Flood. Much of its original, softer outer-flank deposits were stripped off by the flood and now are scattered as detritus in the streamlined pendant bar that extends northwestward from the butte. The depression that lies immediately southeast of the butte, now occupied by Jensen Lake, is a scour hole formed during the flood. The rapidly flowing and boulder-laden flood rushing past White Butte was able, especially on its southern side, to remove the volcano's flanks and expose the dike and sill complex in its core. White Butte has several interesting features, including inward-dipping crudely-bedded tuffs unconformably overlain by thick lenses of massive block-and-bomb-rich tuff. Its west side has poorly-sorted non-palagonitized tuff grading upward to water-affected and columnar-jointed magmatic deposits that may be small subaqueous lava flows. These deposits dip in many directions, predominantly toward the center of the butte, and they thicken and thin over short distances, as if subaqueous spatter was remobilized and flowed down to pond within the crater. White Butte had a short-lived eruption, possibly in fairly deep water (Godchaux et al., 1992). White Butte basalt is quite Fe- and Ti-rich (Table 2); it is

similar to the Walters Butte basalt just to the north. It is probable that these two constructs are nearly the same age.

Stop 2-4: Grouch Drain Maar

Grouch Drain maar is traversed by Highway 45 where this road reaches the plateau north of the Snake River. It is about one km across and one of the best-exposed maars in the western SRP hydrovolcanic field. It formed about 0.7 Ma ago (C.M. White, unpublished Ar/Ar date) when basalt magma intersected groundwater in sediments beneath the Hat Butte basalt. When Grouch Drain maar formed the Snake River valley was already partly developed, so that the maar was excavated high on the valley side rather than on flat ground. The succession of materials that were erupted from the Grouch Drain maar were massive, blocky explosion breccia, then bedded tuffaceous materials rich in recycled sediments, and finally basaltic cinders and spatter deposited from Strombolian activity that occurred as the water supply ran out. At the end, a lava pond formed in the maar. It probably was fed from a lava fountain located where the hump remains today in the center of the crater. As this lava pond developed, the southern margin of the maar failed or was undermined by lava, and because it was at the top of a slope instead of on flat ground, the lava flowed down the slope into the valley. At about a km from its source this lava reached the valley bottom, which then was several meters above the present river level. This is indicated by the fact that the terminus of the Grouch Drain lava flow is markedly water-affected, as if the flow encountered swampy ground or even standing water. The Grouch Drain basalt is quite Mg-rich in comparison to many other western SRP basalts, even though it was one of the latest to be erupted (Table 2).

Stop 2-5: Swan Falls Dam-Sinker Butte Area

(This is an alternate stop in case the unimproved road to Stop 2-1, the Graveyard Point sill, is impassable.)

Travel from Nampa to the Swan Falls-Sinker Butte area by first going to the town of Kuna. To do this proceed east from Nampa on Highway 84. Take exit 44 and proceed 6.5 miles south on State Route 69 to the town of Kuna. Continue south on Route 69 for another 0.5 mile to the Kuna Road. Turn right and proceed west on the Kuna Road for 1 mile to the Swan Falls Dam Road. Turn left and proceed south on the Swan Falls Dam Road for about 17 miles to the Swan Falls Dam. When traveling south from Kuna you can see Kuna Butte to the west of Swan Falls Dam Road. This volcano was a hill eroded in sediments when basalt eruptions occurred on it about 0.4 Ma ago (Othberg et al., 1995a). Some of the basalt erupted from near the top of the hill, converting it into a shield-like construct. Abundant sediment exposures on its northeast flank, however, show that

it is not just a simple shield. In addition to eruptions from the top of Kuna Butte other basalt eruptions formed a number of satellite cinder cones scattered around the south and east sides of Kuna Butte. Just to the southwest, Powers Butte, with a basalt composition very much like that at Kuna Butte, also was built at this time. Basalt flows from this group of eruptions went southward, cascaded into the Melba alcove, and flowed as far as the Snake River where they now form the north canyon wall near Celebration County Park. Farther along, the Swan Falls Dam road passes by Initial Point, which can be seen to the east. Initial Point is mainly a broad shield, capped by a late eruptive cone. Basalt from Initial Point also erupted at about 0.4 Ma ago and lies above the basalts erupted from the Kuna Butte area. Like the basalt from the Kuna Butte area, Initial Point lavas cascaded into the Melba alcove and flowed almost as far south as the Snake River.

An excellent view of Sinker Butte is available from the top of the grade that descends to Idaho Power Company's Swan Falls Dam and power plant. Sinker Butte is a topographically imposing, symmetrical remnant of an even larger tuff cone (Fig. 21). It has a broad flat welded spatter cap in the center of which is a central mound whose structure suggests that it was a late-stage fire-fountain deposit. The volcano has well-exposed radial dikes and extensive associated lava flows which moved away from its base in all directions. It is the largest phreatomagmatic edifice in the western SRP; its summit stands 370 m above the Snake River at Swan Falls Dam, and its present height above its base on the old lake bottom (the Swan Falls Reservoir Basalt) is about 270 m. Before erosion removed its outer walls, it may have stood 370–400 m above the lake bottom, and 240–270 m above the lake surface. This volcano may have commenced erupting into water which was 100 or more m deep. The reconstruction shown in the figure is based on dips of tuff beds in the north and south erosional alcoves, and uses the 970 m elevation of the subaerial-to-pillowed transition in the flow to the south of the butte as the lake level. Sinker Butte is one of the youngest volcanoes in the field; its age is between 0.8 and 1.5 Ma (Amini et al., 1984).

Godchaux et al., (1992) noted the following stratigraphic-upwards trends in the Sinker Butte tuff deposits:

(1) Massive black, poorly sorted deposits containing relatively little juvenile material, and exhibiting crude polygonal cracks at several scales, form the base of the volcano. They are overlain by about 100 m of heterogeneous thin, planar-bedded tuff with abundant juvenile sideromelane lapilli. These lapilli typically have palagonitized rims, but some beds contain completely unaltered lapilli in a palagonitized ash matrix.

(2) Where original shard morphology has not been obscured by palagonitization, it is apparent that there

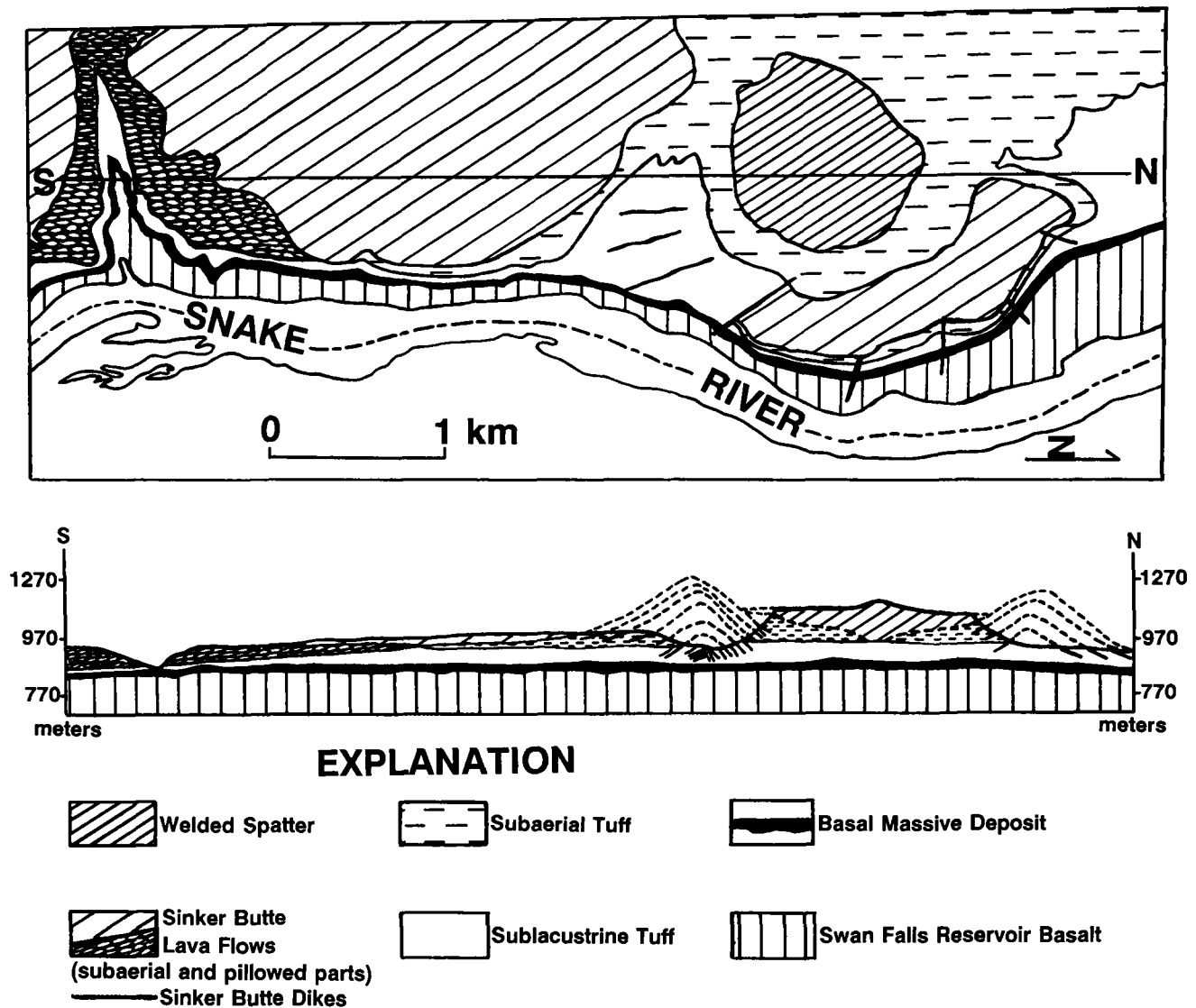


Figure 21. Map and north-south cross-section of Sinker Butte, from Godchaux et al., 1992.

is a general trend upward in the bedded tuffs from non-vesiculated or poorly-vesiculated shards to highly vesiculated shards.

(3) Block and bomb sags are abundant and pronounced in the lower layers, but minor to absent in the upper layers.

(4) Blocks and accidental lapilli change upward in abundance and type in a fashion that suggests a continually deepening explosion focus as the eruption progressed. Thus, each lithologic type first appears in the tuff beds in the reverse of the stratigraphic order of the underlying layers. The fact that the highest tuff beds contain some blocks of the earliest-cored layers suggests that periodic collapse of the

growing cone carried both early tuff deposits and possibly also slices of bedrock back down into the active vent.

(5) The bedded Surtseyan palagonite tuffs are interbedded with relatively well-sorted Strombolian cinder beds near the top of the bedded section, and the bedded section as a whole is overlain by welded spatter.

These observations are consistent with a Surtseyan eruption that began beneath at least 100 m, and perhaps even 150 m., of standing water and ended as a subaerial Strombolian eruption. The apparent scarcity of surge cross-bedding, even in the upper part of the bedded section, may be misleading. The proximal facies in the high-standing cone

walls, where the sandwave facies would presumably have been best developed, has mostly been eroded away, and what is left is mostly obscured by slope debris. The excellent exposures in the canyon walls are mostly medial and distal facies. Deposition of these thin tuff beds probably involved both low-concentration pulsatory surges and air-falls. The transition from sublacustrine to subaerial deposition of the Sinker Butte tuffs has been interpreted (Godchaux et al., 1992) as the height in the bedded section where secondary minerals, including calcite and gypsum, cease to be the cementing agent in the indurated layers. This occurs at about 970 m elevation, in good agreement with the subaerial to pillowed transition in the associated lava flows.

Stops for Day 3

Stop 3-1: Teapot Subaqueous Volcanic Complex

Follow Highway 45 south from Nampa, and turn southeastward on Highway 78 after crossing the Snake River at Walters Ferry (Fig. 3). After passing Guffey Butte, turn off the highway onto the unimproved road that follows West Rabbit Creek, and stop about where the road and West Rabbit Creek become the same. This area is in the informally-named Teapot volcanic field, named for the shape of the basalt-capped hill located just east of where the road becomes the stream bed. The basalt in the Teapot field appears to have been erupted mainly under fairly deep water, between about 9 and 7 Ma ago (Amini et al., 1984; Hart and Aronson, 1983; Othberg et al., 1995a; Ekren et al., 1981). This local area will serve as a microcosm of the complexities that arise under such conditions. Here, we can see abundant amounts of water-affected basalt interspersed with generally massive tuffs. Some of the basalt here appears to be material that flowed along the bottom of the lake, at places becoming invasive into the soft sedimentary material that was present. These sediments were baked by the basalt at several locations, and have become flint-like. Some of the basalt actually may be the dikes that fed basaltic lava into the lake bottom environment. Age-dates in this area are 7.85 ± 0.19 and 7.92 ± 0.19 Ma (C.M. White, unpublished Ar/Ar dates), and the composition of typical samples are given in Table 2.

Also present in the Teapot field, but occurring at some distance from where we'll stop along West Rabbit Creek, are underwater venting areas in which pillows that were forming in the vent were being pushed upward by more basalt pillows and, in the process, were being fragmented to a breccia consisting of small, sometimes glassy, angular fragments. Such deposits have formed several square kilometers of brecciated, pillow-basalt, volcanic agglomerate. Also present at other places in the Teapot field are shallowly-emplaced diabasic sills that have textural features similar

to those in the Graveyard Point Sill, although not as coarse-grained. At present, our understanding of these underwater basalts is not very refined. All these deposits simply have been lumped together in a single map unit and a few chemical analyses and age determinations made.

Teapot Volcanic Field to Castle Butte

Return to Highway 78 and turn right (southeastward), toward Murphy and Grand View (Fig. 3). Follow Highway 78 to the left-hand turnoff at Castle Creek. Follow this road to the Castle Butte outlier for Stop 3-2. Along the way, between the Teapot area and Castle Butte, several things of interest will be encountered. The Owyhee Range will continue to be prominent to the southwest. All along this stretch the pediment gravels rise up to the margin of the range at elevations typically between 3600 and 3800 ft; this elevation very likely represents the high-stand of Lake Idaho, probably achieved sometime during the late Miocene. As you travel by the town of Murphy, which is the county seat of Owyhee County and displays its famous single parking meter in front of the court house, you'll be able to catch glimpses of intensely water-affected basalt mixed in with sediments. This WAB forms the dark-colored areas along the highway. The altered basalt in most of these exposures has disaggregated to a crumbly, grus-like material. These basalts are the Murphy basalt (Ekren et al., 1979) and belong to the first western SRP basaltic volcanism episode. They flowed to the bottom of Lake Idaho, probably while it was quite deep, but their source is unknown.

Farther on, where Highway 78 crosses Sinker Creek, more WAB is exposed in the road cut on either side of the creek. Note how altered it is, compared to subaerial basalt flows seen previously. This is the Sinker Creek basalt and its source is believed to be Hill 3337, located a short distance north of Highway 78 and east of Sinker Creek. The Sinker Creek basalt is also thought to be late Miocene and part of the first western SRP episode. Up Sinker Creek, about a mile from Highway 78, an extensive pillow delta occurs where the Sinker Creek basalt flowed southward into standing water. It is probable that by the time the Sinker Creek basalt erupted, the lake level was somewhat lower than when the Murphy basalt erupted, and it is likely that Hill 3337, the source volcano, had emerged above the lake's surface.

A few miles southeast of Sinker Creek, north of the highway about a mile, is Fossil Butte, a prominent flat-topped hill. It too, was formed during the first episode of western SRP basaltic volcanism. The basalt that caps Fossil Butte is spatter, and it overlies several tens of meters of poorly-exposed, generally palagonitized, basaltic tuff. It seems likely that Fossil Butte is a partially eroded tuff cone, in which the outer flanks have been eroded away so that only

the core of the construct remains, protected by the resistant basaltic spatter cap. This construct, like Hill 3337, probably emerged out of Lake Idaho. It also was formed later than the WAB that we passed near Murphy.

After passing Fossil Butte, travelers will be treated to extensive panoramas of white silt and associated sediments that collected in Lake Idaho; some of this material accumulated when the older stage of basaltic volcanism was occurring (Chalk Hills Formation) and part of it was deposited later (Glenns Ferry Formation), between the two episodes of basaltic volcanism. During mapping of this and adjacent areas we found the lithologic differences between these two formations to be negligible. Thus, if one is interested in clearly distinguishing them, datable chronostratigraphic markers, such as volcanic ash beds, and fossil assemblages that are indicative of age, must be used to date the sediments.

Stop 3-2: Basaltic Tuffs at Castle Butte Outlier

Castle Butte volcano, a noted landmark along the South Alternate Oregon Trail, consists of two preserved remnants of what probably was originally a larger construct. The main spatter-capped butte with a summit elevation of 2700 feet, and a lower outlier about half a mile to the northwest with a summit elevation of 2500 feet, are the two remnant sectors of an original asymmetric tuff ring or cone in which the south side probably stood highest. Inasmuch as Castle Butte stood in the Bonneville Flood path, it is likely that an unknown amount of it was removed. Also, a maar? crater, which may have formed to the northeast of the remains of this construct, is now entirely covered by later flood and other alluvial deposits.

An unusually vigorous late magmatic stage at Castle Butte resulted in the building of at least three spatter cones atop the high south wall. The water depth at the time of the eruption is not well constrained, but it was probably shallow. The level of Lake Idaho probably was no higher than 2450 feet, the lowest elevation at which the base of one of the late-stage cinder cones is exposed, and it may have been even lower. Basal massive deposits have an uncommonly high proportion of finely comminuted sediment in the matrix. Cored bombs are also common. Well-bedded tuffs are less abundant and layers of such tuffs are less continuous and more lensoid on Castle Butte than on other emergent volcanoes; however, surge bedding is locally well-developed.

Distal-facies tuffs exposed in a small outcrop on the north side of the outlier exhibit spectacular soft-sediment deformation and huge sediment rip-ups. Monolithic erosional remnants on the south side of the outlier preserve the steep inward-dipping face of the original explosion crater where phreatomagmatic debris and interbedded sediment

lenses were deformed during collapse into the vent. Cinders and a thick spatter layer, with the largest and most exquisitely flattened and baked sediment clasts that we have observed, occur at the top of the butte. The Castle Butte phreatomagmatic complex appears to be a good example of an emergent volcano characterized by a high eruption rate, at least at the end of the eruption, resulting in the summit morphology for which the butte is named.

The bedded tuff that is preserved on top of the Castle Butte outlier contains numerous, excellent examples of volcanic and sedimentary structures and consists of two major zones. The lower zone overlies structureless, fine-grained white sediments and consists of thinly bedded to layered sediments and basaltic ash and cinders. The most striking features within this lower zone are the numerous and complex flame, load, and rip-up clast structures within the thickest bed. The upper zone has a lower part that is thinly bedded and consists predominantly of fine-grained, white sediments. These lower beds have been disrupted into decollement folds and large rip-up clasts by the emplacement of the overlying, more massive and structureless, basaltic tuff. Within this upper layer of massive basaltic tuff are numerous vertical dewatering channels. This tuff may be from Castle Butte, a source further east, or from a hidden source to the north.

The Castle Butte basalt is modestly Fe- and Ti-rich (Table 2). The absolute age of Castle Butte has not been determined, but the construct clearly belongs in the second episode of basaltic volcanism in the western SRP graben. Castle Butte may be relatively young, since there is no sign of younger materials having been deposited on its top. Even though it stands at a level lower than the partially eroded lake sediments exposed north of the Snake River, there is a good chance that it formed after some, or even most, of the erosion of the Snake River channel had occurred. Alternatively, if it were older than the sediments and had been exhumed, one would expect to find preserved remnants of sediments that were deposited upon it.

Castle Butte Area to Jackass Butte Area Via Grand View

Return to Highway 78, turn left, and continue southeast toward Grand View. At Grand View, turn off onto Highway 67, and follow this route across the Snake River. At about 2 miles north of the river, turn off the main highway and follow the country roads northwestward, toward the Black Butte-Jackass Butte area, downstream along the Snake River. On the way there you can see that the bluffs north of the river are a mixture of white lake sediments and reddish, yellowish, and brownish volcanic materials. That area contains several phreatomagmatic eruptive points. They probably were subaqueous or emergent, and gave rise to a basaltic tuff layer that can be followed westward into the Jackass

Butte-Black Butte area. On the way to that area, where the road again comes up next to the river, there is a small area of pillow delta exposed, which will be stop 3-3. Downriver, on past the pillow delta, the low bench adjacent to the river is capped by the Brooks Ranch basalt. This is the same basalt flow that forms the high canyon rim farther north. The segment by the road appears to have been downfaulted to the south. This is one of the few places along the Snake River in the western SRP graben where the faults that gave rise to the escarpment that follows the north side of the river are actually in evidence. On past the downfaulted bench the road follows the east side of the river as it goes into the canyon segment between Jackass Butte west of the river, and Black Butte on the east side. We will park in this canyon and hike up a trail into the bluffs alongside Rabbit Creek for a view of the eruptive points, basalt flows, and tuff layers, as stop 3-4.

Stop 3-3: A Pillow Delta Complex

Along the side of the downfaulted Brooks Ranch basalt bench is a pillow-delta occurrence. Pillow deltas similar to this one are widely developed in the western SRP graben, especially in association with many of the younger lava flows. These lavas were erupted subaerially and were spreading across relatively flat land when they encountered bodies of water. As is the case here, once a flow entered water it tended to extrude pillows at its advancing front. These pillows extended themselves down the lake margin and became quite elongate. As flow continued, new pillows formed and lapped over the earlier ones, forming deltas of pillows prograding into the lake. In many instances, as the pillow delta advanced, additional lava formed sheet-like subaerial flows right on top of the pillow delta. Where such boundaries between pillow deltas below, and subaerial flows above, are encountered they clearly mark the level of the lake surface when the basalt was deposited. As can be observed at this locality, much of the basalt flow material lying above the foreset pillows has well developed columnar joints, another feature commonly associated with situations where basalt flows interacted with water. The factors that appear to have favored pillow-delta development, rather than the formation of extensive sheets of water-affected basalt, probably are: (1) steeper shoreline slopes would favor pillow delta formation whereas shallower slopes would favor the development of sheet-like flows that became altered, and (2) high effusion rates would favor the development of lava sheets, whereas lower rates would favor pillow-delta development.

Stop 3-4: Black Butte-Jackass Butte Area

Jackass Butte is an erosional remnant consisting mainly of Glenns Ferry sediments and is cut and capped by phreatomagmatic tuff and basalt. Forming a layer that extends through Jackass Butte, sandwiched between the sediments

below and the phreatomagmatic deposits above, is a layer of water-affected basalt, the Brooks Ranch basalt. It evidently came from a source now buried north of the Snake River, and flowed across the lake bottom, perhaps even while remnants of lake were still there, about 2.2 Ma ago (Anini et al., 1984). Black Butte, although it stands topographically above its surroundings and looks like another shield volcano, is actually a faulted anticline in which the capping folded unit is the Brooks Ranch basalt. This is one of several occurrences of relatively young deformation of the lava flows in this portion of the western SRP graben. Cutting the side of Black Butte is a phreatomagmatic construct, the Rabbitt Creek eruptive point. It probably is best thought of as a phreatomagmatic tuff and spatter cone built up in an earlier-formed explosion crater, or maar. Two additional phreatomagmatic points, approximately the same age as the Rabbit Creek tuff cone, are visible on Jackass Butte, across the river.

At this stop we will walk along the old wagon road that goes up the side of the canyon. Along the way one can clearly see into the interior of the Rabbitt Creek tuff cone and see how the color changes from light to dark upwards, as the abundance of basalt in the erupting crater-fill material gradually increased at the expense of the sediments. Near the top of the trail the stratigraphic succession of several units is clearly exposed. The Brooks Ranch basalt flow is overlain by two different sediment-rich, phreatomagmatic tuff units. Another basalt flow is sandwiched between the two tuff units and clearly shows they were deposited at different times. It is likely that these two tuff units came from the different eruptive points in the Black Butte-Jackass Butte area. Above the tuffs is a considerable amount of coalesced basalt spatter; it may have erupted from a source hidden in the hillside north of the canyon. Basalt analyses for this stop are given in Table 2.

Grand View to Twin Falls Area

Return to Highway 78 at Grand View and turn to the east (left; Fig. 3) and proceed on to Hammett. Along the way we will pass through the town of Bruneau. East of Bruneau, the road climbs up through a section of lake sediments, capped by gravels. These are lake beds deposited in Lake Idaho, probably as part of a delta complex distributed by the ancestral Bruneau River, during Glenns Ferry time. The gravels at the top of the sediment section are part of the Tuana Gravel, they represent the stream-deposited material that prograded farther and farther out into the lake as its level was drawn down, near the end of its existence. Farther along we will pass by Bruneau Dunes State Park. These large sand dunes are purported to be the highest stationary dunes, at 400 feet above their base, in the United States. They are located in a protected zone of the Snake River

canyon, where wind-blown sand has accumulated since the Bonneville Flood passed this way. Farther along, where the canyon narrows, is an interesting occurrence where the rim basalt was being undercut by the Bonneville Flood at a locality where the water velocity was dropping as the channel abruptly widened. Here, one can see that the rim basalt was broken up and dropped to a lower level as soft sediment was washed out from beneath it. Further on, where Highway 78 crosses back to the north side of the Snake River, there are good exposures of the Glens Ferry Formation sediments; here they consist of interbedded silts and sands. These beds may have been deposited in an alternating fluvial and lacustrine environment. Just past Hammett the road joins Highway I-84. From here we will retrace our path back to the Twin Falls area, and then on to Salt Lake City.

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