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Geothermal Potential of the Blackfoot Reservoir-Soda Springs Volcanic Field: A Hidden Geothermal Resource and Natural Laboratory in SE Idaho

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Errata: Co-author David W. Rodgers was inadvertantly omitted from the originally published version of this paper.

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Introduction

The majority of eastern Idaho’s hot springs and thermal wells occur within a broad belt stretching from Oregon to the Yellowstone plateau along the southern margin of the Eastern Snake River Plain (ESRP), an area believed to be favorable for developing sources of geothermal energy (Mitchell et al., 1980; Laney and Brizbee, 2003). The Pleistocene Blackfoot-Willow Creek-Grays Lake-Soda Springs-Gem Valley volcanic field (henceforth referred to as the Blackfoot-Soda Springs Volcanic Field or BVF) is located adjacent to the southeastern margin of the ESRP (Figure 1). Like the ESRP, it is a young bimodal volcanic system (e.g., McCurry et al., 2008; Ford, 2005) with a relatively thin carapace of fractured basalt that hosts a highly productive aquifer overlying much less permeable rocks (Dion, 1974). Unlike the ESRP, the BVF is situated in a fault-bounded Basin and Range graben and is underlain by Tertiary basin-filling sediments and Upper Paleozoic sedimentary bedrock (e.g., Oriel and Platt, 1980). Because it lacks the characteristic manifestations of a high-temperature resource, the BVF was originally designated a “Low-Temperature Geothermal Area” (Mitchell et al., 1980). Interestingly, as Coolbaugh et al. (2002) have pointed out, the BVF and the ESRP are the only two instances in the Basin and Range having young rhyolite extrusives but lacking known high-temperature geothermal systems, possibly because ground-water flow in their shallow basalt aquifers obscures any evidence of a high-temperature system at the surface (e.g., Brott et al., 1976). The BVF’s proximity to the ESRP hot-spot track, the presence of young rhyolite (China Cap dated at 58 ± 7ka; Heumann, 2004; Figure 1, 2) and numerous thermal springs (e.g., Huntspiller and Parry, 1985), and significant gravity and magnetic anomalies (Maybe and Oriel, 1970) make it a favorable prospect for a so-called “ESRP”-type of hidden geothermal resource.

For the purpose of assessing the characteristics of a potentially important hidden geothermal resource this paper documents geothermal-related features of the centermost of three main interconnected basins of BVF. Following Mansfield (1927), this region will be referred to as the Blackfoot Lava Field (BLF) (Figure 1, 2).

Figure 1. Location map for the Blackfoot Reservoir-Soda Springs volcanic field (BVF), and Blackfoot Lava Field (BLF). White color = basalt lavas; Red = rhyolite lava domes.
Geology and Hydrogeology

Regional Setting

The Blackfoot Lava Field (BLF) is located adjacent to the southeastern margin of the ESRP in the northeastern Basin and Range Province (Figure 1). Northwest trending basins contain 3 to 5 km of Tertiary intrabasinal clastic sediment, limestone, silicic tuffs and basalt lavas (Kellogg and Marvin, 1988; Bobo, 1991). Horst blocks, and deeper crustal rocks of BLF consist of late Proterozoic to Mesozoic supracrustal rocks of the Sevier fold and thrust belt (Oriel and Platt, 1980; Dixon, 1982). These have been stacked to a thickness of 10-15 km by numerous southwest dipping thrust faults (Dixon, 1982), producing a strong north striking, west dipping structural anisotropy to the upper crust. Mid- to lower crustal rocks consist of crystalline Archean Wyoming craton (Foster et al., 2006).

Quaternary Volcanism and Shallow Magmatic Activity

The Blackfoot Lava Field (BLF) consists mainly of Quaternary olivine tholeiite (basalt) lava flows and tephra deposits (Figure 1, 2; Mansfield, 1927; Fiesinger et al., 1982; Pickett, 2004). These basalts are coeval with and are mineralogically and geochemically similar to basalts erupted along the Eastern Snake River Plain segment of the Yellowstone-Snake River Plain volcanic track (McCurry et al., 2008; Kuntz et al., 1992). BVF is the largest basalt dominated volcanic field to have erupted marginal to the Snake River Plain. None of the BLF lavas have been dated, but they are at least in part coeval with lavas in Gem Valley that have yielded dates as young as 27.5±10ka (Reuben et al., 1960).

Although BLF basalts are similar to ESRP basalts in many respects they are distinguished in others. ESRP basalts are dominated by effusive eruptions from rift zones and widely scattered central vent systems resulting in development of thick piles of overlapping shield volcanoes (Greeley, 1982; Kuntz, 1992; Hughes et al., 1999). In contrast, in the BLF, shield volcanoes are less common, and evidence of more explosive volcanism, mainly formation of magmatic tephra cones, is more common. BLF lavas and tephra also more commonly exhibit greater degrees of crustal interaction (assimilation) than those of the ESRP (Pickett, 2004; McCurry, unpub. data).

Rhyolite Lava Domes

Three clusters of rhyolite lava dome overlap spatially with the BLF basalts. The oldest are located west of Greys Lake and

Figure 2. Map of salient geologic and geohydrologic features of the Blackfoot Lava Field.

Figure 3. View of China Hat lava dome looking north (photo by Mark Ford).

Figure 4. Representative view of dike-related, young volcanic rifts located west of China Hat.
as isolated islands in Blackfoot Reservoir. Both yield overlapping K/Ar dates of ~1.5 Ma (Luedke and Smith, 1983). The youngest cluster includes the 3 coeval and geochemically and mineralogically indistinguishable domes and 2 maars (‘China Hat Dome Field’ of Figure 1). A rhyolite sample from China Cap (middle of the three lava domes) yielded a 40Ar/39Ar date of 58±7 ka (Heumann, 2004). The domes and maars are aligned along a north-northeast direction, in contrast to northwest trending Basin and Range dominated tectonic features in the region.

Mineralogical and geochemical features of the rhyolites allow for determination of their depths and temperatures of origin (~13 km and ~760°C, respectively). Interestingly, preliminary melt inclusion analysis indicates that the rhyolites also were strongly hydrous at depth, subsequently degassing during slow ascent towards the surface (Ford, 2005; McCurry, unpublished data). In addition, the rhyolites contain common but volumetrically minor amounts of mafic magmatic enclaves. The enclaves are similar in composition to nearby olivine tholeiite lavas, and they indicate that the two magma types intermingled with each other just prior to their eruption.

Eruption of rhyolites in the central dome field was preceded by moderate to minor explosive activity, some of which was magmatic, but at least some of which was also phreatomagmatic (Ford, 2005; McCurry, unpublished data). The former occur in a quarry on the north side of China Hat (Figure 2, 3). These beautifully exposed deposits were emplaced from a combination of ash fall, ballistic fall and dilute density currents. They include a variety of essential rhyolite and accidental basalt lithics. Some of the latter exhibit weak palagonitization, suggesting that the eruption was in part phreatomagmatic.

Two overlapping maars (Burchett and Gronewell Lakes) occur between China Cap (center dome) and North Cone (northern dome) (Figure 2). Although the maar deposits are obscured by loess, they strongly suggest that abundant and shallow ground water occurred in the area at the time. Fall deposits that are potentially correlative with explosive deposits of the central dome field have been observed in borehole studies in the Greys Lake (K. Pierce, personal communication, 2005).

**Volcanic Rift Zone**

The Blackfoot Rift Zone (BRZ; Figure 2, 5) is a robust, young, 2.5 – 5 km wide zone of normal faults in the Blackfoot Lava Field (Figure 4; Polun et al., 2010). Faults typically strike north to north-northwest, and some exhibit a curvilinear strike. Most faults appear to dip vertically at the surface, but are interpreted to dip 50 - 70° at depth based on observations in other volcanic rift zones (e.g. Gudmundsson, 1992). Dip direction is both east and west, creating a complex pattern of graben, horsts, monoclines and half-grabens. Narrow grabens, for example, are common in the middle of the rift zone (see Cross Section B-B’; Figure 6) and along the east or west rift zone margins. Faults are manifested in the basalt lava as abrupt scarps, scarps with broad colluvial aprons, or as monoclines that presumably drape blind faults. Vertical offset across individual faults ranges from 1 – 50 m, and faults with the greatest throw (30 – 50 m) are located on one side or the other of the rift margin (Figure 2, 5).

Timing of the BRZ is partially constrained by geologic field relations. The 58 ka China Hat lava dome was deposited over the
top of one large rift fault scarp (Figure 3, 5) and it is in turn cut by another fault (Figure 5). Although none of the BLF basalts are radiometrically dated they are mainly of normal magnetic polarity (Maybe and Oriel, 1970), and are moderately to non-eroded. They appear to be at least in part coeval with late Pleistocene lavas (≥ ~100 - 25 ka) of the Gem Valley lava field. All basalt lavas and at least some scarp s are overlain by late Pleistocene (<14 ka?) loess. Finally, no faults cut other faults and several faults splay into others, implying that the BRZ faults are of approximately similar age or at least formed in the same stress regime. A preliminary best guess for the development of the scarp s is estimated to be ~50 – 150 ka (although it could vary from as young as 25 ka to as old as ~780 ka).

A series of 33 east trending topographic transects spaced 500 m apart were created and analyzed to measure the amount of horizontal extension and its variation along strike within the BRZ. Topographic offset for each fault was measured and converted to horizontal extension by assuming a fault dip of 50 - 70°.

As shown in Figure 7, cumulative rift zone extension increases from 10 – 200 m northward to a maximum in the vicinity of China Hat. Measurements north of China Hat are complicated by the Blackfoot Reservoir which conceals central faults within the rift zone. However, a linear relationship between rift zone width and cumulative fault scarp offset is observed to the south, and this relation can be used to estimate cumulative rift zone extension north of China Hat (Figure 7). Given 75 – 200 m of extension near China Hat and a maximum faulting duration of 780 ka., a minimum surface extension rate of 0.1 – 0.26 mm/yr is interpreted for the BRZ over the 25 km width of the containing valley, which is comparable to strain rates observed in Central Idaho via geologic, seismic and geodetic methodologies (Eddington et al., 1987; Parsons et al., 1998; Payne et al., 2008). Considering that the BRZ is located in a broad valley, 10 km from the nearest range front, and has < 50 m topographic and structural relief, such an extension rate is incompatible with regional Basin and Range extension but compatible with other rift zones associated with dike injection (Parsons et al., 1998; Rowland et al., 2007).

It is unlikely that the surface deformation in the BRZ results from a single massive dike, but instead from numerous dikes intruded as a swarm. To calculate the width and depth to the top of this dike swarm, field measurements of surface deformation were input into an elastic dislocation model (Okada, 1985). This method approximates the dike swarm as a single, rectangular dislocation at some depth in a homogenous, isotropic elastic half space. Depth to dike tip was inverted from rift zone width, and with a fixed depth to dike tip, dike width was inverted from cumulative surface extension amounts. The preferred result using this method yields a depth to dike tip of 1 – 2 km, shallowing to the south, and a dike width roughly 2.5 – 3 times surface extension. Cross-sections in Figure 6 show the modeled dike depth, width, and location relative to surface faulting. Collectively, approximately 17 km$^2$ of magma may have been intruded between depths of 1 and 5 km along the BRZ.

**Blackfoot Reservoir**

Blackfoot River Reservoir (Figure 2) was commissioned in 1907 to control the upper reach of the Blackfoot River and supply the Fort Hall Indian Reservation on the Snake River Plain with irrigation water. It was completed in about 1910 and has a usable storage capacity of 413,000 acre-feet at a designed maximum water surface elevation of 6124 ft amsl. However, due to high seepage losses from the southern end of the reservoir, the operating level has been limited to an elevation of about 6120 ft amsl, or about 340,000 acre-feet of storage capacity. Post-reservoir groundwater levels south of the reservoir (Figure 2) illustrate strong southward down gradient flow apparently channeled in part by the BRZ. Due to this leakage, the reservoir has had a major influence on the hydrology of the valley south to Soda Springs.

The containment of the Blackfoot River and flooding of a substantial portion of the northern valley has had a profound influence on the hydrology of the southern valley, all the way to Soda Springs, and has contributed to the masking of the deeper ground water system by the highly permeable basalts that overlie the Tertiary basin fill and Paleozoic section. Prior to the reservoir, Five Mile Meadow was a dryland grazing range; within six months of the reservoir’s filling, Five Mile Meadow was inundated by ground water that leaked from the impoundment and drastically
changed the shallow ground water flow system. Figures 2 and 8 summarize some of the information that is available on the shallow basalt-hosted aquifer (after Dion, 1974). The specific capacity data available from a few irrigation wells (up to 3500 gpm/foot of drawdown) and resulting transmissivity estimates indicate that the hydraulic characteristics of these basalts are very similar to the eastern Snake River Plain aquifer. Hydraulic gradients south of the Blackfoot Reservoir indicate that ground water flow velocities (30-50 ft/day) are also like those seen in the ESRP aquifer.

**Subsurface Geology**

Maybe and Oriel (1970) document regional patterns of Bouguer gravity and aeromagnetic anomalies for BVF. Graben basin geometry is inferred primarily from gravity data, and indicates that BLF is developed within a northwest trending basin that has a maximum depth of about 1.5 km. The basin is asymmetrical having a shallow western platform that deepens rapidly near the western margin of AVZ. We infer that this is a result of an inactive Tertiary normal fault that has been buried by the Pleistocene basalts.

Hubbard 25-1 Borehole: This well was drilled in 1980-81 by Phillips and Hunt Oil as a wildcat geothermal exploration well. Located about 1 km south of China Hat, the borehole penetrates ~250 m of basalt lavas, and another ~1.5 km of intrabasinal limestone and clastic sediment, and interlayered basalt lavas (or dikes/sills). Those in turn overlie early Mesozoic clastic sediment and Paleozoic limestone. The well encountered water up to 96 °C in a zone between 6000 and 7200 ft depth before bottoming in 70 °C water at 7835 ft TD (Figure 9). Due to hole instability and other problems, no thermally equilibrated measurements were made and the well was abandoned without any wireline logs being run. Interestingly, the zone of warmer water coincides with downward projection of the inferred buried, east-dipping normal fault.

**Geothermal Activity**

**Springs and Travertine Terraces**

Springs and travertine terraces are widely distributed around the margins of the BLF (Figure 2). Active and inactive terrace development records evolving patterns of spring activity (Figure 2). Terraces that cluster along the margins of the graben-filling basalt suggest control of subsurface flow along tectonic range bounding faults. No dating has been done on the carbonates to determine how long-lived the spring complexes are or whether the terraces formed coevally or in response to thermally evolving subsurface conditions. The springs issuing on these terraces are Ca-Mg-HCO₃-type waters that range from ambient ground water temperature (8-10 °C) to a maximum of 40 °C; all are slightly acidic due to high CO₂ contents.

Besides the Ca-Mg-HCO₃ water type of fluid whose temperatures are slightly above ambient ground water and that occur throughout the valley (e.g., Woodall Spring and Hooper Spring, Table 2.a), two other types of fluid chemistries are present: a cold, acid-sulfate type (Sulfur Springs; Figure 2) and a Ca-Mg-HCO₃-SO₄ type that is associated with the hottest springs (Corral Creek Springs and Soda Geyser).

All geothermometry evidence indicates that these fluids equilibrated at a temperature no higher than 100 °C. These fluids may have been heated during descent along deep faults at the prevailing regional geothermal gradient or they may represent mixing (and re-equilibration?) of hotter fluids with shallow groundwater. Based on the work of Hutsinpillar and Parry (1985), water-rock reactions at shallow depth likely plays the biggest role in determining the chemical composition of these fluids.

**Paucity of Surficial Hydrothermal Alteration**

Little or only minor hydrothermal alteration of host rocks has been documented in previous literature, or in reconnaissance observations at springs, travertine deposits or within or adjacent to rhyolite lava domes. Some palagonitized basalt lithics occur within tuff cone deposits at China Hat recording weak hydrothermal interaction, although that likely occurred during or just prior to its eruption. Additionally, no evidence of hydrothermal alteration or mineral deposition was recorded in borehole cuttings reports from well Hubbard 25-1. The well was drilled to a depth of ~2.5 km, and is located only 1 km from China Hat, and near the axis of BRZ, where our structural analysis suggests shallow intrusions of mafic magma. Further south a prominent magnetic anomaly, possibly representative of mafic intrusion or sulfide-rich mineralized rocks occurs near the center of the BRZ (Figure 10). However additional geophysical characterization and modeling of the anomaly is required.
thermometry indicators range from 100-115°C, relict spring deposits are found near the margins of the basin-filling younger shallow intrusions, and voluminous young shallow intrusions suggest that BVF may have magmatically heated hot rock at shallow depths. To justify exploration for a possibly deeper, hotter, possibly enhanced geothermal system (EGS), or related hydrothermal systems that are veiled by geologic and hydrogeologic features of the system, a compelling exploration conceptual model is needed. We believe that one exists, and that it is testable:

First, the BVF is characterized by young rhyolite domes whose roots may retain significant latent heat. The lateral and cross-sectional geometry of youthful rifting in the surrounding volcanic rift system (Kuntz, pers. comm., 2009), the China Hat area alone may retain more than $2 \times 10^{20}$ joules of igneous thermal energy, equivalent to that of Newberry Crater and one-sixth that of the Coso Mountains. If the volcanic rift system encompassing China Hat developed from intrusion of rhyolite dikes then the geothermal potential of the area could be substantially greater than $2 \times 10^{20}$ joules.

**Why is BVF a Hidden Resource?**

We suggest that three main factors dominate suppression of surficial expressions of geothermal activity for BLF: 1. a distinctive shallow hydrogeology (highly hydrologically conductive basalts) via mixing, dilution, and redirection of upward directed geothermal waters; 2. mid-level cooling effects (on heat flow; heat sources and related geothermal fluids) of robust downward infiltration/convection of cold meteoric water via numerous, closely spaced rift-related normal faults; 3. deeper level redirection of upward convecting geothermal waters by major thrust faults and B&R faults. These hypotheses are elaborated on in a companion paper by Autenrieth et al. (this volume).

**Summary and Discussion**

The BLF is located in a fault-bounded graben, so its minor surficial thermal energy could originate from deep circulation along basin-bounding faults into a conventional warm-water geothermal reservoir. Such a hypothesis is supported by geothermometry of local hot spring fluids (indicating maximum reservoir temperatures of 100-115°C) and supported by the maximum temperatures tapped by the Hubbard 25-1 well (in excess of 95°C; Figure 9) in a restricted zone between 2000 and 2200 meters below China Hat. On the other hand, the presence of young rhyolite volcanism, and voluminous young shallow intrusions suggests that the BLF may have magmatically heated hot rock at shallow depths. To justify exploration for a possibly deeper, hotter, possibly enhanced geothermal system (EGS), or related hydrothermal systems that are veiled by geologic and hydrogeologic features of the system, a compelling exploration conceptual model is needed. We believe that one exists, and that it is testable:

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The characterization and interpretation of the BVF’s thermal, geologic and hydrologic features will be critical to targeting drillable subsurface heat sources. A fundamental understanding of the basin’s thermal manifestations, their spatial relationships and the mechanisms that control their surface expression will contribute to enhancing the exploration for, and development of, hidden geothermal resources in other young volcanic terrains such as the ESRP. The robust geologic, hydrologic, geochemical and geophysical features of the Blackfoot–Soda Springs volcanic field indicate that it is an ideal natural laboratory for improving our understanding of, and hone methods for effective geothermal exploration of a common type of hidden geothermal resource.

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