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ABSTRACT

Southeastern Idaho exhibits numerous warm springs, warm water from shallow wells, and hot water from oil and gas test wells that indicate a potential for geothermal development in the area. Although the area exhibits several thermal expressions, the measured geothermal gradients vary substantially (19 – 61 ºC/km) within this area. We have estimated reservoir temperatures from chemical composition of thermal waters in southeastern Idaho using an inverse modeling technique (Reservoir Temperature Estimator, RTEst) that calculates the temperature at which multiple minerals are simultaneously at equilibrium while explicitly accounting for the possible loss of volatile constituents (e.g., CO_2), boiling and/or water mixing. The temperature estimates in the region varied from moderately warm (59 ºC) to over 175 ºC. Specifically, hot springs near Preston, Idaho resulted in the highest reservoir temperature estimates in the region.

1. INTRODUCTION

Southeastern Idaho has potential geothermal resources as suggested by geologic evidence such as Pleistocene basaltic flows, young volcanic features, and warm to hot springs (Mitchell, 1976; Ralston et al., 1981; Souder, 1985). More direct evidence of a high-temperature regime at depth in the area is provided by a limited number of deep wells with high bottom-hole temperatures such as King 1-2 well (a temperature of 249 ºC, Table 1). Despite this geologic evidence and hight bottom-hole temperatures, estimates of reservoir temperature based on traditional geothermometers applied to the chemistry of waters from springs in the region generally suggest a moderate temperature (Mitchell, 1976). As a part of an effort to assess the geothermal potential of southern Idaho, we assembled chemical composition of waters measured from numerous springs and wells in the region and applied a multicomponent equilibrium geothermometry (MEG) technique to estimate reservoir temperatures in this area.

2. GEOLOGY AND GEOTHERMAL SETTING OF THE AREA

2.1 Geology

The study area is located in both the Basin and Range and Rocky Mountains provinces. Specifically, the western part of the area has geographic characteristics of the Basin and Range such as wide and sediment filled basins separating fault-bound ranges, whereas the eastern part consists of several thrust-bound narrow sub-parallel ridges with thinly filed basins (Mabey and Oriel, 1970). Geologically, the fold-thrust belt in the area is a part of Sevier fold-thrust zone, locally known as the Idaho-Wyoming fold-thrust belt (Armstrong and Oriel, 1965).

Geology of the area (Figure 1) includes thick sequences Paleozoic and Mesozoic carbonate-rich sedimentary sequences deposited in Cordilleran miogeocline (Armstrong and Oriel, 1965). During the Jurassic-Cretaceous periods these sedimentary sequences were deformed by compressive stresses associated with the Sevier orgony resulting in numerous west-dipping low-angled thrust faults (Armstrong and Oriel, 1965). Starting in the Eocene and continuing to the recent, extensional activities resulted in Basin and Range type topography with normal faults bounding ranges and wide valleys (Armstrong and Oriel, 1965; Dixon, 1982). Quaternary volcanic activity in some areas in the region (McCurry et al., 2011) resulted in volcanic features such as the Blackfoot Volcanic Field (BVF) with dominant olivine tholeiite lava flows and occasional rhyolitic lava domes (McCurry et al., 2008; Pickett, 2004).

2.2 Geothermal Setting

The presence of several hot springs and warm springs indicate potential geothermal resources in southeastern Idaho. The western part of study area represents the amagmatic Basin and Range type geothermal system where convective upwelling dominates the thermal discharge along the extensional faults. The discharge of hot/warm water from springs and seeps in eastern and northern parts of the study area are also reported to be controlled by deep normal faults (Dansart et al., 1994). However, some recent works (e.g., McCurry et al., 2011; Welhan et al., 2014) also suggest a deep magmatic geothermal resource in this area. The conceptual model of magmatic-sourced geothermal setting in the fold-thrust belt in southeastern Idaho considers a magmatic geothermal resource at a depth of 12-14 km in an area beneath a 58 ka rhyolite domes at China Hat located within the BVF (Welhan et al., 2014). According to this hypothesis, the deep-sourced magmatic hydrothermal fluid from this zone migrates eastwards along the thrust faults and permeable Paleozoic and Mesozoic layers into a shallower (3-5 km) reservoir. The high-temperature and high-salinity (sodium-chloride) thermal fluids encountered at depth in some deep wildcat petroleum wells (e.g., King 2-1 well in Table 1) in the region are reported to be associated with these migrated magmatic fluids (Welhan et al., 2014).

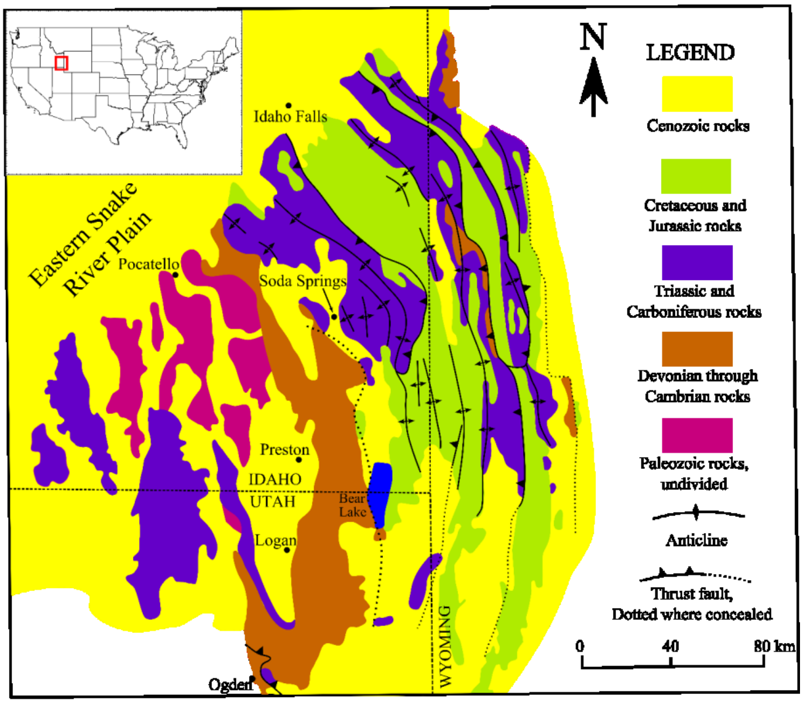

Figure 1: Simplified geologic map of Idaho-Wyoming fold-thrust belt (Armstrong and Oriel, 1965).

Table 1: Depth and bottom-hole temperatures of several wild-cat oil exploration wells in southeastern Idaho (Ralston et al., 1981; Souder, 1985; Blackwell et al., 1992).

Wells	Depth (m)	Bottom-hole T (ºC)
King 2-1	3927	249
Grand Valley	4931	140
Mike Spencer Canyon	4259	112
Bald Mountain-2	3830	148
Black Mountain-1	4158	100
Big Elk Mountain-1	1545	103
Federal 1-8	5105	188
Big Canyon Federal 1-13	3551	161
IDST-A1	4952	180
Tincup	5059	160
N Eden Federal 22-11	2618	92

3. SOUTHEASTERN IDAHO WATER CHEMISTRY DATA

Chemical compositions of numerous water samples from southeastern Idaho were assembled to assess the potential geothermal reservoir temperatures in the region. Over the last several decades, water samples from springs and wells in the southeastern Idaho have been analyzed by several US government agencies and researchers for water quality and management, environmental remediation, and geothermal energy exploration (e.g., Young and Mitchell, 1973; Mitchell, 1976A,B; Ralston et al., 1981; Souder, 1985; Avery, 1987). A database has been compiled of publically available data from southeastern Idaho springs/wells. From a larger database, 50 selected water compositions (Table 2, Figure 2) were used for a preliminary assessment of the deep geothermal temperatures in southeastern Idaho.

4. GEOTHERMOMETRY

4.1 Approach

A newly developed geothermometry tool known as Reservoir Temperature Estimator (RTEst) (Palmer, 2013; Neupane et al., 2013, 2014) is used to estimate deep geothermal temperature in southeastern Idaho. The RTEst is an inverse geochemical tool that implements MEG with a capability of process optimization for secondary processes such as boiling, mixing, and gas loss. More detailed description about RTEst can be found elsewhere (e.g., Palmer, 2013; Neupane et al., 2014).

4.2 Missing Components

The MEG approach requires that measured water composition include all components present in the reservoir mineral assemblage (RMA). For aluminosilicate minerals, this requires measured values of Al that are often not available in historical data bases. For water compositions without measured Al, an Al-bearing mineral (e.g., K-feldspar) was used as a proxy for Al during geochemical modeling as suggested by Pang and Reed (1998).

Table 2: Water compositions of selected hot/warm springs and wells in southeastern Idaho used for temperature estimation. Elemental/species concentrations are given in mg/L. The pH was measured in the field.

Springs/Wells[a]	T (ºC)	pH	Na	K	Ca	Mg	$SiO_{2(aq)}$	HCO_3	SO_4	Cl	F	Map Code[b]	Water type[c]	Data source[d]
Woodruff WS	27	7.3	910	87	130	45	29	454	58	1600	0.6	WO	I	1
E. Bingham W	63	6.2	4600	770	320	36	68	930	48	7800	3.9	EB		1
Squaw HS-1	69	6.5	4184	708	135	23	126	816	27	6877	4.3	SQ & BC		2
Squaw HS-2	73	6.6	3844	533	241	26	126	866	23	6396	4.8			2
Squaw HS W-1	82	7.8	4300	880	250	23	130	733	54	7700	7			3
Squaw HS W-2	84	6.5	4368	782	279	24	124	791	35	7398	4.3			2
Squaw HS W-3	82	6.9	3996	694	261	21	139	725	35	7291	4.9			1
Battle Creek HS-1	43	6.7	3161	552	174	19	109	696	35	5241	6			2
Battle Creek HS-2	77	6.5	3071	535	166	15	107	697	29	5048	6			2
Battle Creek HS-3	81	6.5	3053	533	162	19	109	757	37	5034	6			2
Battle Creek HS-4	82	6.8	4184	686	215	24	97	610	33	6967	6.4			2
Wayland HS-1	84	7	3100	660	160	16	80	699	50	5400	12			3
Alpine WS	37	6.5	1500	180	560	100	40	880	1000	2800	2.7	AL	II	1
Wayland HS-2	77	6.9	499	77	82	22	64	454	323	585	1	SQ & BC		1
Treasurton WS-1	35	6.6	563	127	265	68	54	704	788	632	2.2	MG		2
Treasurton WS-2	40	6.4	542	110	336	48	54	726	735	629	2			1
Cleavland WS	55	6.2	444	90	259	41	62	565	517	574	1.7			1
Maple Grove HS-1	72	7.3	490	110	89	24	55	491	260	630	1.1			3
Maple Grove HS-2	60	6.8	501	82	93	29	85	495	261	601	1.1			2
Maple Grove HS-3	76	6.8	492	80	93	25	86	494	251	584	1			2
Maple Grove HS-4	71	7.8	494	76	69	31	52	424	255	595	0.9			4
Maple Grove HS-5	78	6.6	492	82	85	30	84	494	256	596	1.1			2
Maple Grove HS-6	75	6.3	550	71	132	24	66	466	282	586	0.3			1
Auburn HS	57	6.4	1327	162	509	76	68	822	996	1737	0.6	A & J		1
Johnson S	54	6.4	1494	176	454	45	88	973	1129	1947				1
Ben Meek W-1	40	7.4	348	20	23	5	90	526	5	321	11	BM	III	1
Ben Meek W-2	45	7.3	360	24	25	7	80	524	15	320	10			1
Ben Meek W-3	40	6.9	368	22	24	7	89	513	13	322	9.6			1
Rockland W-2	20	7.3	60	24	120	22	70	220	26	280	0.2	RL2	IV	5
Bear Lake HS-1	40	7	155	48	230	41	43	263	769	72	4.2	BL	V	1
Bear Lake HS-2	39	7.2	151	44	227	41	46	255	791	75	4.2			1
Bear Lake HS-3	33	7.1	163	43	227	41	40	271	758	74	4			1
Bear Lake HS-4	48	6.6	180	61	210	55	35	256	800	79	7.1			1
Downata HS	43	6.7	20	9	43	15	29	214	18	20	0.4	DW	VI	1
Black River WS	26	6.2	147	217	674	245	33	2357	1132	110	3.7	BR		6
Pescadaro WS	26	6.4	63	14	188	65	31	658	225	83	1.8	PD		1
Henry WS	20	6.4	25	8	284	44	40	870	145	32	1	HE		1
Steamboat HS	51	7	28	27	645	248	84	2380	472	8	0.3	SS		7
Soda Springs G	28	6.5	12	23	851	193	35	2613	801	6	1.6			1
Lava HS-1	45	6.6	170	39	120	32	32	542	110	190	0.7	LH		3
Lava HS-2	43	6.7	176	37	103	29	35	528	91	179	0.7			1
Portneuf R WS-1	34	6.2	81	62	280	64	38	1060	270	62	0.8	PR		8
Portneuf R WS-2	41	6.3	85	60	275	48	47	1060	259	53	0.7			1
Corral Creek W-1	42	6.5	101	237	701	263	28	2845	898	41	2.3	CC		6
Corral Creek W-12	41	6.8	97	242	620	246	30	2763	908	43	3.5			6
Corral Creek W-13	41	6.6	101	233	697	263	30	2723	896	40	2.4			6
Corral Creek W-14	36	6.6	99	233	649	253	30	2803	884	40	2.5			6
Dyer W	21	7.7	50	3	50	13	68	188	1	61		D & A		1
Anderson W	20	7.7	45	7	50	10	111	199	0	45				1
Rockland W-1	20	7.6	27	13	37	8	160	180	15	28	0.6	RL1		5

[a]Wells/springs types – W: well, HS: hot spring, WS: warm spring, S : spring, G: geyser; [b] These map codes are used to define the springs/wells in Figure 2, [c]Water types are – I: Na-Cl (12 samples), II: $Na\text{-}HCO_3\text{-}Cl + Ca\text{-}SO_4$ (13 samples), III: $Na\text{-}HCO_3\text{-}Cl$ (3 samples), IV: Ca-Cl (1 sample), V: $Ca\text{-}SO_4$ (4 samples), and VI: $Ca\text{-}HCO_3$ (17 samples); [d] Data sources – 1: Ralston et al. (1981), 2: Mitchell (1976A), 3: Young and Mitchell (1973), 4: Dion (1969), 5: Parliman and Young (1992), 6: Mitchell (1976B), 7: Souder (1985), 8: Mitchell et al. (1980).

4.3 Reservoir Mineral Assemblage

Based on general lithology of the southeastern Idaho and literature assessment of secondary minerals for dominant rock and water types, we used reservoir mineral assemblages (RMAs) consisting of idealized clays, zeolites, carbonates, feldspars, and silica-polymorph (chalcedony) (Table 3) to determine temperatures from these waters.

5. RESULTS AND DISCUSSION

5.1 Southeastern Idaho Springs/Wells Waters

Compositions of waters from hot/warm springs and wells in southeastern Idaho are presented in Table 2. The pH of the southeastern Idaho thermal waters range from 6.2 to 8.1, with arithmetic mean, median, and standard deviation 6.87, 6.70, and 0.51, respectively. Similarly, the field temperature of southeastern Idaho springs/wells range between 20 to 84 ºC. The aqueous chemistry of these southeastern Idaho thermal waters shows a large range in total dissolved solids (TDS) from about 250 mg/L (Downata Hot Spring) to more than 14,000 mg/L (East Bingham Well).

Table 3: Weighting factors for minerals used in this study.

Minerals	Weighting factor (w_i)
Calcite	1/2
Chalcedony	1
K-feldspar	1/5
Mordenite-K	1/7
Clinochlore-14A	1/10
Paragonite	1/7
Saponite-K/Na	1/7.33
Disordered dolomite	1/4

The dominant cations in the southeastern Idaho thermal waters are Na and Ca with minor amounts of Mg (Figure 3). The thermal waters include samples dominated by Cl^-, HCO_3^-, or SO_4^{2-} while others appear to be dominated by more than one anion. Hierarchical cluster analysis using Ward's (1963) method as implemented in SYSTAT 13 (SYSTAT Software, Inc.) was performed using the 6 Piper diagram end members (Ca^{2+}, Mg^{2+}, Na^++K^+, Cl^-, HCO_3^-+CO_3^{2-}, SO_4^{2-}) for classifying water in the southeastern Idaho. Six compositional groups were identified within the 50 thermal water samples: Na-Cl (12 samples), Na-HCO_3-Cl + Ca-SO_4 (13 samples), Na-HCO_3-Cl (3 samples), Ca-Cl (1 sample), Ca-SO_4 (4 samples), and Ca-HCO_3 (17 samples) (Table 2). These groups likely reflect differences in sources of water, water-rock interactions, and structural control of the local geothermal systems.

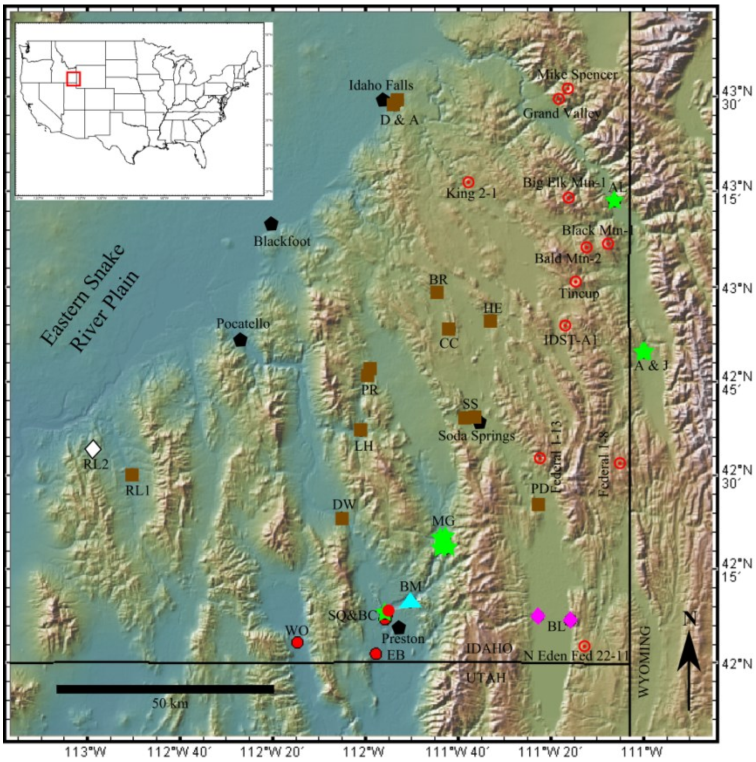

Figure 2: Shaded relief map of southeastern Idaho prepared from NASA 10-m DEM data in GeoMapApp. Water compositions of selected hot/warm springs and wells [water types - red circles (●) Group I; green stars (★): Group II; cyan triangles (▲): Group III; open diamond (◇): Group IV; magenta diamonds (◆): Group V; and brown squares (■): Group VI] in southeastern Idaho are used for temperature estimation (Table 4). The springs/wells codes correspond to the map code given in Table 2. The wild-cat petroleum wells (Table 1) are represented by red target (⦿) signs.

The Na-Cl and Ca-SO_4 type waters may have been originated with the water-rock interactions involving pockets of evaporites in the area. Oriel and Platt (1980) have reported the presence of evaporites (e.g., halite, gypsum, and alum) in Middle Jurassic sequences (Pruess Redbeds) in southeastern Idaho. Recently, Welhan et al. (2014) indicated that the high-salinity waters in some deep wild-cat petroleum wells may be related to magmatic waters from a zone as deep as 12-14 km under the BVF in the fold-thrust belt in southeastern Idaho. However, all Na-Cl types waters considered in this study are from the surface expressions (hot/warm springs) or from the rather shallow (compared to the deep wild-cat petroleum wells) wells located in the western part (Basin and Range Province). The Na-Cl type waters from this part of the study area may have been originated by the water-rock interactions involving evaporites. This type of water is also reported from the Raft River Geothermal Area (RRGA) located to the west of the present study area. Moreover, there is likely an additional source of evaporites in the Tertiary rocks (Ayling and Moore, 2013). All Ca-SO_4 type waters are from hot springs near Bear Lake, located near Idaho-Wyoming-Utah triple point. Deep sourced water from a nearby deep wild-cat petroleum well (N Eden Federal well with depth >2500 m) has very high SO_4 concentration; however, this water has low Ca concentration and high Na concentration (Souder, 1985). The Ca-SO_4 type waters that Bear Lake hot springs issue may have separate sources of Ca and SO_4 or there may have some ongoing cation exchange reaction involving Ca and Na along the flow path from depth to the surface.

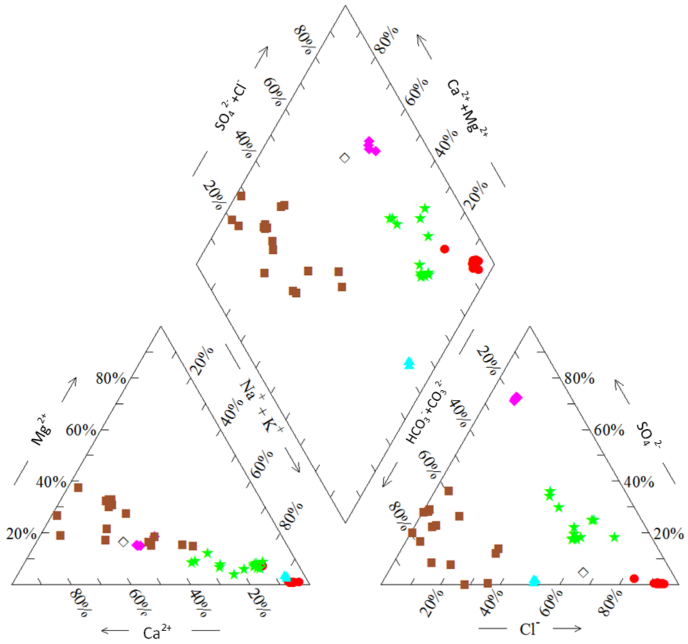

Figure 3: Reported chemistry of waters measured from several hot/warm springs and wells located in southeastern Idaho.

(Water types - red circles (●) Group I; green stars (★): Group II; cyan triangles (▲): Group III; open diamond (◇): Group IV; magenta diamonds (◆): Group V; and brown squares (■): Group VI.)

The Ca-HCO_3 type waters are scattered throughout the area. These waters typically exhibit low Cl concentrations (Table 2). With some exceptions (e.g., Black River Warm Spring, Corral Creek Wells, Soda Geyser, Pescadaro Warm Spring), these waters also have low SO_4 concentration. This type of water is generally regarded as a product of the interaction of groundwater with Ca-rich rocks at shallower depth. In the adjoining ESRP, the Ca-HCO_3 type water represents the water in the active part of the ESRP aquifer whereas the deeper waters in ESRP area are Na-HCO_3 type (Mann, 1986; McLing et al., 2002).

Only one sample that represents the Ca-Cl type water is from Rockland W-2 located in the westernmost part of the study area (Figure 2). In a previous study (Neupane et al., 2014), this type of water was not identified in the Eastern Snake River Plain (ESRP) geothermal system located to the west-northwest side of the present study area (Figures 1 and 2). The apparent lack of Ca-Cl waters in the ESRP system could be related to limited numbers of water samples used in that study (Neupane et al., 2014) or this type of water is not a common water in south Idaho (including ESRP and southeastern Idaho), and it represents an outlier in the present study. Although this water has some similarity with the Ca-HCO_3 and Ca-SO_4 types of water in terms of high Ca content compared to the Na + K concentrations, its high Cl concentration with low Na concentration makes it difficult to assign it as a direct product of a particular water-rock interaction.

The remaining two types of waters – Na-HCO_3-Cl and Na-HCO_3-Cl + Ca-SO_4 are mixed waters. Although the cluster analysis did not classify a separate group of Na-HCO_3 type water, this water is a representative of the deep water in adjoining ESRP area. It is likely that these waters are Na-Cl type waters but interacted with carbonate sections with or without gypsum/anhydride layers.

5.2 Southeastern Idaho Geothermal Temperatures

5.2.1 Giggenbach Diagram

When plotted on a Giggenbach diagram (Giggenbach, 1988), the majority of the southeastern Idaho waters selected for this study plot in the immature zone with some waters lie in the zone of partial equilibration (Figure 4). The mature waters in Figure 4 are from hot springs and wells near Preston, Idaho (Battle Creek and Squaw hot springs), and these water could have interacted with rock at a temperature range of 260 – 300 °C. The lack of equilibrium (immaturity) in majority of southeastern Idaho waters could be related to low Na content, as suggested by Giggenbach (1988), as well as to their higher Mg content. The waters containing high Mg content are deemed to be unsuitable for some traditional solute geothermometry; although there have been some efforts made for implementing Mg correction in the estimated temperature (e.g., Fournier and Potter, 1979).

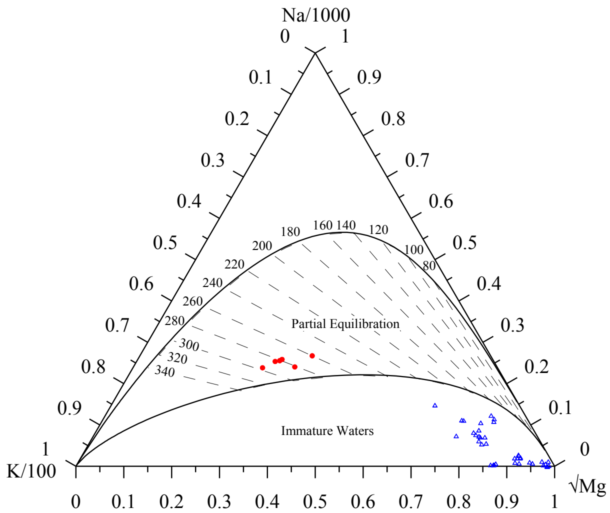

Figure 4: Southeastern Idaho waters from hot/warm springs and wells plotted on Giggenbach diagram (Giggenbach, 1988). The red and blue symbols represent mature and immature waters, respectively. All mature waters belong to Group 1 type waters.

5.2.2 Temperatures Estimated by MEG

Estimates of reservoir temperatures for southeastern Idaho thermal waters (Table 2) were made using RTEst. The RMAs used consisted of representative minerals (Mg bearing minerals – clinochlore, illite, saponite, disordered dolomite; Na bearing minerals – paragonite, saponite; K-bearing minerals – K-feldspar, mordenite-K, illite; Ca bearing minerals – calcite, disordered dolomite; and chalcedony) (Table 3). For the selected compositions of southeastern Idaho thermal waters that do not have measured Al concentration, a value determined by assuming equilibrium with K-feldspar was used in the geochemical modeling.

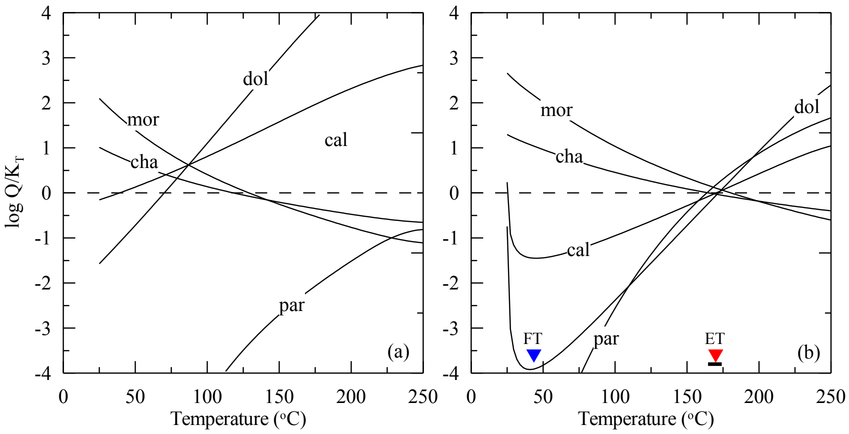

Figure 5: Temperature estimation for Battle Creek Hot Spring near Preston, Idaho. (a) The log Q/K$_T$ curves for minerals calculated using original water chemistry with K-feldspar used as proxy for Al, (b) optimized log Q/K$_T$ curves [FT: field temperature (43ºC); ET: estimated temperature (169 ºC), the dark horizontal bar below ET represents the ±standard error for the estimated temperature (±5 ºC); cal: calcite, cha: chalcedony, dol: disordered dolomite, mor: mordenite-K, and par: paragonite).

In MEG, the reservoir temperature is estimated by first selecting a reservoir mineral assemblage (RMA) with which it is believed the fluid in the reservoir is equilibrated. For a water sample from a spring or shallow well, the activities of the chemical species in solution are determined and the saturation indices [SI = log (Q/K$_T$), where Q is the ion activity product and K$_T$ is the temperature dependent mineral-water equilibrium constant) calculated using the laboratory measured temperature of the sample. This calculation is repeated as a function of temperature and the resulting SIs recalculated. Likely reservoir temperature is the one at which all minerals in an assemblage are in equilibrium with the reservoir fluid as indicated by near zero log Q/K$_T$ values of these minerals on a log (Q/K$_T$) versus temperature plot [log (Q/K$_T$) plot] (Reed and Spycher, 1984; Bethke, 2008). Alternately stated, reservoir minerals are expected to be in equilibrium with the fluid and they should yield a common equilibrium temperature with a near zero log (Q/K$_T$) value for each mineral; this common equilibrium temperature coincides with the reservoir temperature. If log

(Q/K_T) curves of minerals in a reservoir do not show a common temperature convergence at log (Q/K_T) = 0, then it suggests that there exists errors in analytical data, the selected mineral assemblage does not represent the actual mineral assemblage in the reservoir, or the sampled water must have been subjected to composition altering physical and chemical processes during its ascent from the reservoir to the sampling point.

Figure 5a shows log (Q/K_T) curves of the RMA (calcite, chalcedony, disordered dolomite, mordenite-K, and paragonite) used for the reported Battle Creek Hot Spring-1 water compositions. The log (Q/K_T) curves of these minerals intersect the log (Q/K_T) = 0 at a wide range of temperatures, ranging from 40 ºC (calcite) to over 250 ºC (paragonite), making the log (Q/K_T) curves derived from the reported water chemistry minimally useful for estimating temperature. The range of equilibration temperature for the assemblage minerals is a reflection of physical and chemical processes that may have modified the Battle Creek Hot Spring-1 water composition during its ascent to the sampling point.

Table 4: Temperature estimates for southeastern Idaho thermal waters RTEst and other geothermometers.

Springs/Wells[a]	$T^b \pm \sigma^c$	$M_{H_2O}{}^d \pm \sigma^c$	$\log f_{CO_2} \pm \sigma^c$	φ[e]	Quartz[f]	Chalcedony[g]	Silica[h]	Na-K-Ca[i]	Types[j]
Woodruff HS	97±3	0.61±0.03	1.04±0.12	3.51E-04	78	47	49	56	I
E. Bingham W	161±4	0.61±0.02	2.36±0.14	4.67E-04	117	88	88	193	
Squaw HS-1	179±9	0.45±0.08	2.22±0.3	2.09E-03	151	125	123	204	
Squaw HS-2	157±6	0.31±0.06	1.87±0.19	9.04E-04	151	125	123	183	
Squaw HS W-1	175±5	0.38±0.05	2.42±0.17	6.94E-04	152	127	125	229	
Squaw HS W-2	174±6	0.43±0.05	2.26±0.19	8.73E-04	150	124	122	217	
Squaw HS W-3	171±7	0.35±0.07	2.16±0.24	1.33E-03	156	132	129	216	
Battle Creek HS-1	169±5	0.46±0.04	2.14±0.15	5.72E-04	142	116	114	205	
Battle Creek HS-2	175±6	0.50±0.04	2.19±0.18	8.37E-04	141	115	113	215	
Battle Creek HS-3	170±5	0.47±0.03	2.12±0.15	5.24E-04	142	116	114	202	
Battle Creek HS-4	171±4	0.51±0.03	2.23±0.14	4.95E-04	136	109	107	204	
Wayland HS-1	175±5	0.54±0.03	2.5±0.15	5.80E-04	125	97	97	230	
Alpine WS	98±9	0.46±0.11	1.27±0.34	3.30E-03	92	61	63	92	II
Wayland HS-2	144±7	0.60±0.06	0.82±0.27	1.67E-03	114	85	85	84	
Treasurton WS-1	111±3	0.44±0.04	1.24±0.14	4.78E-04	105	76	77	78	
Treasurton WS-2	111±9	0.45±0.12	1.17±0.39	3.79E-03	105	76	77	113	
Cleavland WS	119±7	0.45±0.09	0.92±0.29	1.99E-03	112	83	84	106	
Maple Grove HS-1	126±4	0.55±0.04	1.43±0.16	6.29E-04	106	77	78	97	
Maple Grove HS-2	123±4	0.27±0.07	0.77±0.18	7.33E-04	128	101	100	73	
Maple Grove HS-3	124±3	0.28±0.05	0.78±0.14	4.78E-04	129	101	101	82	
Maple Grove HS-4	115±7	0.49±0.08	0.96±0.28	1.90E-03	104	74	75	54	
Maple Grove HS-5	126±6	0.31±0.09	0.76±0.25	1.43E-03	128	100	99	67	
Maple Grove HS-6	122±5	0.44±0.07	0.69±0.22	1.13E-03	115	86	87	97	
Auburn HS	107±9	0.26±0.17	0.99±0.39	3.78E-03	117	88	88	104	
Johnson S	116±13	0.18±0.26	1.08±0.52	6.64E-03	130	103	102	134	
Ben Meek W-1	106±7	0.02±0.18	-0.15±0.33	2.44E-03	131	104	103	86	III
Ben Meek W-2	106±4	0.12±0.09	0.01±0.2	9.00E-04	125	97	97	72	
Ben Meek W-3	109±4	0.06±0.1	-0.16±0.2	9.28E-04	131	103	102	73	
Rockland-W2	110±7	0.29±0.13	0.23±0.34	2.79E-03	118	90	90	93	IV
Bear Lake HS-1	113±7	0.57±0.06	0.86±0.27	1.84E-03	95	64	66	73	V
Bear Lake HS-2	111±7	0.53±0.07	0.75±0.28	1.98E-03	98	68	69	94	
Bear Lake HS-3	107±8	0.56±0.07	0.77±0.31	2.46E-03	92	61	63	92	
Bear Lake HS-4	121±4	0.64±0.02	1.02±0.12	4.41E-04	86	55	57	90	
Downata HS	97±3	0.62±0.03	0.17±0.17	6.61E-04	78	47	49	49	VI
Black River WS	103±3	0.48±0.04	2.17±0.13	4.53E-04	83	52	55	85	
Pescadaro WS	68±8	0.24±0.21	-0.11±0.43	4.59E-03	81	49	52	41	
Henry WS	60±16	-0.12±0.98	-0.41±0.94	2.18E-02	92	61	63	89	
Steamboat HS	96±11	-0.12±0.25	1.77±0.43	3.41E-04	128	100	99	46	
Soda Springs G	59±15	-0.09±0.59	0.91±0.72	1.31E-02	86	55	57	88	
Lava HS-1	94±6	0.56±0.07	0.84±0.27	1.81E-03	82	51	53	67	
LAVA HS-2	94±5	0.52±0.07	0.76±0.25	1.51E-03	86	55	57	64	
Portneuf R WS-1	100±6	0.53±0.08	1.35±0.28	1.93E-03	89	59	61	92	
Portneuf R WS-2	101±9	0.44±0.13	1.29±0.4	3.90E-03	99	69	70	111	
Corral Creek W-1	98±3	0.51±0.03	2.55±0.12	3.97E-04	77	45	48	98	
Corral Creek W-2	100±4	0.45±0.05	2.57±0.15	5.79E-04	79	48	51	97	
Corral Creek W-3	98±3	0.48±0.04	2.51±0.12	3.90E-04	79	48	51	99	
Corral Creek W-4	98±3	0.49±0.04	2.53±0.13	4.67E-04	79	48	51	100	
Dyer W	121±3	0.41±0.03	0.41±0.11	5.28E-04	117	88	88	57	
Anderson W	144±4	0.33±0.03	0.81±0.12	7.44E-04	143	117	115	74	
Rockland-W1	131±4	-0.13±0.09	-0.31±0.16	5.97E-04	165	142	138	88	

[a] HS: Hot spring, WS: Warm spring, W: well; [b] RTEst estimated temperature; [c] σ is standard error in each RTEst optimized parameter (temperature, mass of water, and fugacity of CO_2); [d] Positive and negative numbers indicate the fraction of cold water and steam-loss per kilogram of sampled water, respectively; [e] φ is objective function of RTEst; [f] Quartz no steam loss, Fournier (1977); [g] Fournier (1977); [h] Arnórsson et al. (1983); [i] Truesdell and Fournier (1973), Mg correction applied according to Fournier and Potter II (1979); [j] Water types are – I: Na-Cl (12 samples), II: Na-HCO_3-Cl + Ca-SO_4 (13 samples), III: Na-HCO_3-Cl (3 samples), IV: Ca-Cl (1 sample), V: Ca-SO_4 (4 samples), and VI: Ca-HCO_3 (17 samples).

Two common composition altering processes are the loss of CO_2 due to degassing and the gain/loss of water due to mixing/boiling. Particularly, the loss of CO_2 from geothermal water has direct consequence on pH of the water, and it is often indicated by the oversaturation of calcite (Palandri and Reed, 2001). Similarly, dilution of thermal water by mixing with cooler water or enrichment of constituents (chemical components) by boiling is indicated by lack of convergence of log (Q/K_T) curves over a small temperature range at log (Q/K_T) = 0. Although, in principle, these composition altering processes can be taken into account by simply adding them into the measured water composition and looking for convergence of the saturation indices of the chosen mineral assemblage, a graphical approach becomes cumbersome even for two parameters (e.g., temperature and CO_2).

To account for possible composition altering processes, RTEst (Palmer, 2013) was used to simultaneously estimate a reservoir temperature and optimize the amount of H_2O and the fugacity of CO_2 (Table 4). The optimized results for Battle Creek Hot Spring-1 are shown in Figure 5b. Compared to the log (Q/K_T) curves calculated using the reported water compositions (Figure 5a), the optimized curves (Figure 5b) converge to log (Q/K_T) = 0 within a narrow temperature range (i.e., 169±5 ºC).

The optimized temperatures and composition parameters for the other southeastern Idaho waters reported in Table 2 were similarly estimated using RTEst. The estimated reservoir temperatures, mass of water lost due to boiling or gained due to mixing, and fugacity of CO_2 along with the associated standard errors are presented in Table 4.

5.2.3 Temperature Estimates with Traditional Geothermometers

In addition to RTEst, some traditional geothermometers were also used for reservoir temperatures estimation (Table 4). Because most of the waters from hot/warm springs and wells in southeastern Idaho are issuing immature waters (Figure 4), the use of traditional geothermometers to estimate their temperatures is not very reliable. Temperatures obtained with silica polymorphs and Na-K-Ca geothermometers appear somewhat comparable with the RTEst temperatures. Mean and standard deviation of estimated temperatures for each group of waters with RTEst, chalcedony, and Na-K-Ca geothermometers are presented in Table 5. Group-wise mean chalcedony calculated reservoir temperatures are consistently cooler than the mean RTEst calculated reservoir temperature for each group.

Chalcedony reservoir temperatures were calculated using the silica concentrations assuming that the sample waters completely represent the thermal water. On the other hand, RTEst reservoir temperatures are calculated with MEG using optimized (reconstructed) waters. Table 4 provides the RTEst optimized fraction of cold water (positive numbers) or steam-lost (prior to sampling) per kilogram of sample water from hot/warm springs and wells in southeastern Idaho. Whenever RTEst indicates that the sample water has appreciable fraction of cold water, in general, a higher RTEst temperature is calculated for that sample.

Table 5: Mean and standard deviationa of estimated temperature for each group of water.

Geothermometer	Group 1[b]	Group 2[c]	Group 3[d]	Group 4[e]	Group 5[f]	Group 6[g]
RTEst	165±22	119±11	107±1	110	113±6	98±22
Chalcedony[h]	110±24	85±13	102±4	90	62±5	67±28
Na-K-Ca[i]	196±46	91±21	77±8	93	87±10	79±21

[a] Standard deviation for RTEst temperatures are calculated using RTEst temperatures of each group without incorporating standard error associated with estimated temperature of individual sample; [b] Na-Cl type water (n = 12); [c] $Na-HCO_3-Cl + Ca-SO_4$ type water (n =13); [d] $Na-HCO_3-Cl$ type water (n = 3); [e] Ca-Cl type water (n = 1), since this water type is represented by one sample, no standard deviations were calculated; [f] $Ca-SO_4$ type water (n = 4); [g] $Ca-HCO_3$ type water (n = 17); [h] Fournier (1977); [i] Truesdell and Fournier (1973), Mg correction applied according to Fournier and Potter II (1979).

Mg-corrected Na-K-Ca temperatures are relatively similar to the RTEst temperatures; however, the trend between mean RTEst and Na-K-Ca temperature varies with groups. In general, Na-K-Ca resulted in cooler temperature at lower temperature range and hotter temperature in the upper temperature range compared to the RTEst temperatures (Figure 6). The main weakness of this geothermometer is its less reliability for waters with significant amount of Mg. Compared to the RTEst temperatures, Na-K-Ca temperatures are lower for all but Group 1 waters. The cooler Na-K-Ca temperatures for most of the waters are resulted in due to the large Mg-correction factor because of the high Mg content. Furthermore, the Mg concentration in southeastern Idaho waters seems to be controlled by minerals other than chlorite. Since southeastern Idaho waters have traversed through or are in contact with thick carbonate (limestone/dolomite) sequences, concentration of Mg in these waters appears to be controlled by disordered dolomite. Similarly, for some springs/wells (e.g., Anderson/Dryer wells) that issue water from non-carbonate terrain, concentration of Mg is controlled by smectite-type clays. The overprediction of temperature for Group 1 waters is reported to be caused by the disproportionate (relative to Na and K) loss of Ca due to calcite precipitation (Young and Lewis, 1981). If there were no other pre-sampling consequences (such as mixing/loss of CO_2) that might have happened to the water that Battle Creek Hot Spring in Preston Idaho issues, it is reasonable to assume that this water might have lost Ca due to calcite precipitation such that calcite is oversaturated in this water at field temperature or above (Figure 5a). The Na-K-Ca geothermometer may not produce a reliable temperature for a system where the Ca concentration is independently controlled by non-Ca feldspar minerals (e.g., carbonates) without causing corresponding changes in concentration of K and Na (Fournier and Truesdell, 1973). However, the supersaturation of calcite in this water (Battle Creek Hot Spring) is caused by loss of CO_2 due to degassing. The RTEst modeling estimates temperature using reconstructed water where calcite and other assemblage minerals are at equilibrium at reservoir temperature (Figure 5b). For southeastern Idaho waters, the Na-K-Ca geothermometer may fail to estimate a reliable temperature because most of the assumptions on which it is based are likely to be violated here.

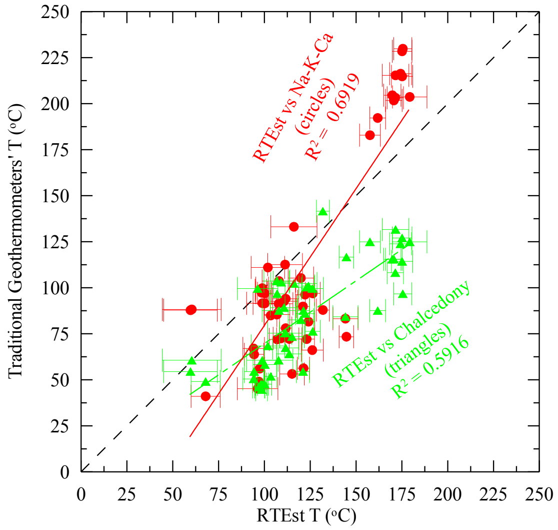

Figure 6: RTEst temperatures versus chalcedony [green triangles (▲)] and Na-K-Ca [red circles (●)] temperatures for the southeastern Idaho thermal waters.

5.2.4 Estimated Temperatures Versus Bottom-hole Temperature of Wild-cat Petroleum Wells

In general, RTEst calculated reservoir temperatures appear positively correlated with nearby bottom-hole temperatures, supporting the argument that MEG can be used to predict deep geothermal reservoirs. As reported in Table 1, some of the wild-cat petroleum wells in the fold-thrust belt in southeastern Idaho provide measured temperature at depth. Although these wells are located several kilometers away from the springs/wells that are used in RTEst temperature estimates (Figure 2), the bottom-hole temperatures at these wells could be compared with the RTEst temperatures in the region. It is important to note that even the bottom-hole temperatures of nearby wells are sometimes varied significantly. For example, Bald Mountain-2 (3830 m) has a reported temperature of 148 ºC whereas as the nearby well Black Mountain-1 (4158 m) has a bottom hole temperature of 100 ºC. Such variation in temperature at depth in nearby wells may suggest that the deep temperatures could be related to their vicinity with the thermal water flow paths controlled by deep discontinuity such as faults. Nevertheless, for some springs, the temperature estimates are close to the bottom-hole temperatures in nearby deep wells.

The North Eden Federal 22-11 well (2618 m) is located east of the Bear Lake, near the triple junction of Idaho, Utah, and Wyoming. This well has slightly lower bottom-hole temperature (92 ºC) than the RTEst temperature estimates (107-121 with standard error ±4 to ±8) for nearby hot springs (Bear Lake Hot Springs represented by letter code BL in Figure 2). Similarly, RTEst temperature estimate for Alpine Spring (letter code AL in Figure 2) (98±9 ºC) is very similar to the bottom-hole temperatures of nearest deep wells, (Big Elk Mountain-1 (1545 m, 103 ºC) and Black Mountain-1. On the other hand, the bottom-hole temperatures at Federal 1-8 (188 ºC) and Federal 1-13 (161 ºC) are significantly higher than the estimated temperature (68±8 ºC) for the closest spring (Pescadaro Warm Spring with PD letter code in Figure 2). The highest bottom-hole temperature was recorded for King 2-1 (3927 m, 249 ºC) well; however, there is no RTEst temperature estimates in the vicinity of this well. Similarly, there is no deep measured temperature in the vicinity of Battle Creek and Squaw Hot Springs near Preston so that we could not directly compare our the highest estimated temperature in southeastern Idaho. However, these hot springs themselves are issuing rather hot waters (up to 84 ºC); and some of the recent shallow wells in the area are reportedly producing water with temperature over 100 ºC.

6. OBSERVATIONS

Geological setting coupled with the direct evidences of thermal expressions such as hot/warm springs in the area suggest that southeastern Idaho has good potential for geothermal resources. Our temperature estimates using RTEst with thermal water compositions measured from southeastern Idaho indicate the presence of relatively hotter zones at depth. Specifically, thermal waters of Battle Creek Hot Springs and Squaw Hot Springs provided a promising geothermal prospect near Preston, Idaho. In other areas, however, the moderately high temperature estimates might reflect the mixing of local groundwater to the deeper thermal water or re-equilibration of high temperature thermal waters at lower temperature zone near surface. Several factors, such as, use of pure water during modeling and overall quality and completeness of the reported water chemistry, might have also contributed in underestimating the true temperature at depth.

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