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Geothermometric evaluation of geothermal resources in southeastern Idaho

G. Neupane[1,2], E. D. Mattson[1], T. L. McLing[1,2], C. D. Palmer[3], R. W. Smith[3], T. R. Wood[4], and R. K. Podgorney[1]

[1]Idaho National Laboratory, Idaho Falls, Idaho, USA
[2]Center for Advanced Energy Studies, Idaho Falls, Idaho, USA
[3]Office of Research & Economic Development, University of Idaho, Moscow, Idaho, USA
[4]Department of Geological Sciences, University of Idaho – Idaho Falls, Idaho Falls, Idaho, USA

Correspondence to: G. Neupane (ghanashyam.neupane@inl.gov)

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. We have estimated reservoir temperatures from chemical composition of thermal waters in southeastern Idaho using an inverse geochemical 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 °C) to over 175 °C. Specifically, hot springs near Preston, Idaho, resulted in the highest reservoir temperature estimates in the region.

1 Introduction

The state of Idaho in the US has high potential of geothermal energy. The US Geological Survey has estimated that there is up to 4900 MWe of undiscovered geothermal resources and 92 000 MWe of enhanced geothermal potential within the state (Williams et al., 2008). Southern Idaho has been regarded to have high geothermal potential for conventional as well as for enhanced geothermal system (EGS) development (Tester et al., 2006). Geologic evidence such as the passage of the Yellowstone hotspot, Pleistocene basaltic flows, young volcanic features, and warm to hot springs (Mitchell, 1976a, b; Ralston et al., 1981; Souder, 1985) in southern Idaho indicate that the area may have economically viable geothermal resources. 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 (BHTs) such as the King 2-1 well (202 °C, Table 1). Despite this geologic evidence and high BHTs, 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, 1976a, b).

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 used them to estimate reservoir temperatures using an inverse geochemical modeling tool (Reservoir Temperature Estimator, RTEst; Mattson et al., 2015). In this paper, we present results of RTEst applied to southern Idaho thermal water measured at a number of wells and springs.

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 (Fig. 1a) in the area is a part of Sevier fold–thrust
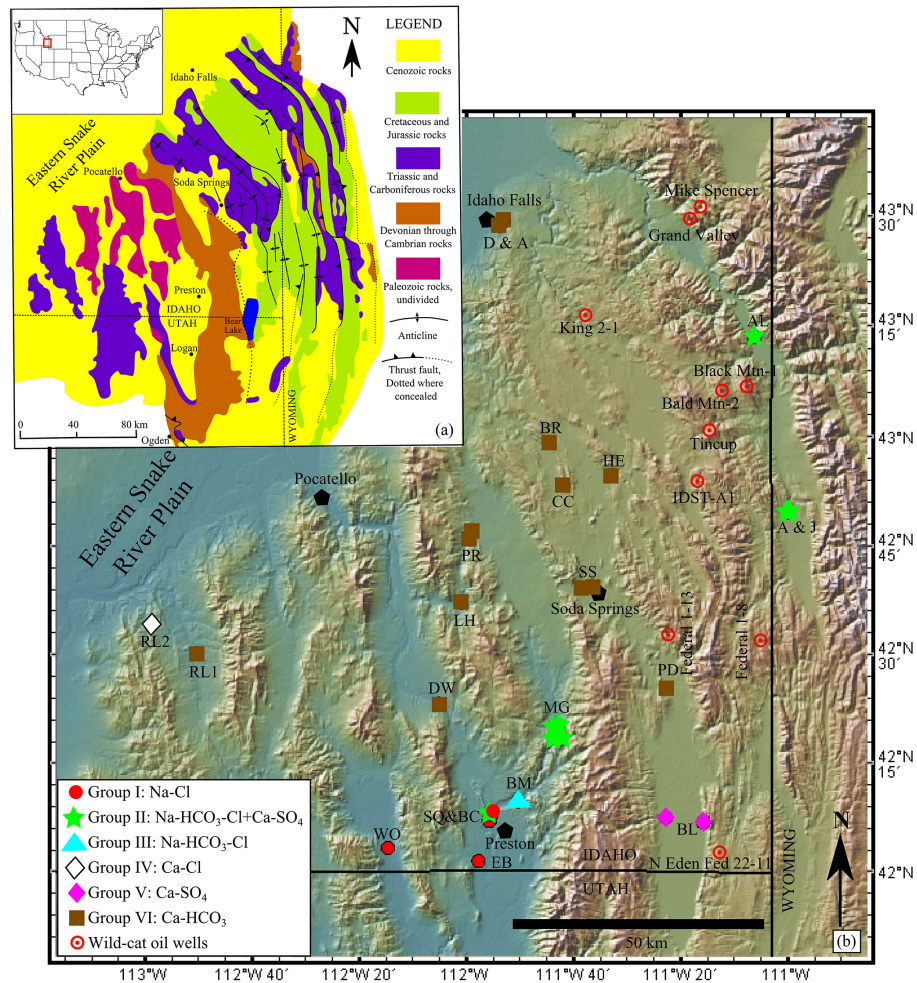
Figure 1. (a) Simplified geologic map of Idaho–Wyoming fold–thrust belt (Armstrong and Oriel, 1965). **(b)** Selected springs/wells and general water composition types in southeastern Idaho. The springs/wells codes correspond to the map code given in Table 2. The shaded relief map prepared from NASA 10 m digital elevation model data in GeoMapApp.

zone, locally known as the Idaho–Wyoming fold–thrust belt (Armstrong and Oriel, 1965).

Geology of the area includes thick sequences Paleozoic and Mesozoic carbonate-rich sedimentary rocks deposited in a passive margin setting (Armstrong and Oriel, 1965). During the Jurassic–Cretaceous periods these sedimentary sequences were deformed by compressive stresses associated with the Sevier orogeny 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 resources (McCurry et al., 2008; Pickett, 2004).

2.2   Geothermal setting

The presence of several hot springs and warm springs indicates 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 discharges of thermal water from springs and seeps in eastern and northern parts of the study area are also 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 58 ka rhyolite domes at China Hat located within the Blackfoot Volcanic Field (BVF) (Welhan et al., 2014). According to this hypothesis, the deep-sourced magmatic hydrothermal fluid

Table 1. Depth and corrected bottom-hole temperatures (BHTs) of several wildcat oil exploration wells in southeastern Idaho (Ralston et al., 1981; Souder, 1985; Blackwell et al., 1992; Welhan and Gwynn, 2014).

Wells	Depth (m)	BHT (°C)
King 2-1	3927	202
Grand Valley	4931	140
Mike Spencer Canyon	4259	112
Bald Mountain-2	3830	140
Black Mountain-1	4158	120
Federal 1-8	5105	167
Big Canyon Federal 1-13	3551	172
IDST-A1	4952	183
Tincup	5059	180
N Eden Federal 21-11	2618	84

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 (such as, the King 2-1 well in Table 1) are possibly associated with these migrated magmatic fluids (Welhan et al., 2014).

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 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; McLing et al., 2002). A database is compiled of publicly available data for southeastern Idaho springs/wells. From a larger database, 50 water compositions (Table 2, Fig. 1b) were selected for the assessment of deep geothermal temperatures in southeastern Idaho.

4 Geothermometry

4.1 Approach

Geothermometry is a low-cost but useful geothermal exploration tool that uses the chemical compositions of water from springs and wells to estimate reservoir temperature. The application of geothermometry requires several assumptions: (1) the reservoir minerals and fluid attain chemical equilibrium, and (2) the water that moves from the reservoir to the sampling location retains its chemical signatures (Fournier et al., 1974). The first assumption is generally valid (provided

there is a sufficiently long residence time); however, the second assumption is more likely to be violated. As reservoir fluids move toward the surface, the pressure on the fluid decreases resulting in boiling and subsequent loss of volatiles (e.g., CO_2). The fluid temperature will decrease as a result of the associated heat of vaporization as well as thermal conduction. Boiling will increase the concentrations of non-volatile components in the liquid phase. The loss of acid volatiles such as CO_2 and H_2S will alter the pH while the loss of redox active species such as H_2S can shift the ratios of redox pairs (e.g., HS^-/SO_4^{2-}). These changes in temperature and solute concentrations may result in re-equilibration of the liquid phase with minerals in the zone above the main reservoir. In addition, this altered thermal water may mix with non-thermal waters which will further alter solute concentrations. These processes mask the initial geochemical signature of the reservoir fluid resulting in temperature estimates from traditional geothermometers being diverse and often being inaccurate or inconclusive.

Geothermal temperature predictions using multicomponent equilibrium geothermometry (MEG) provide apparent improvement in reliability and predictability of temperature over traditional geothermometers. The basic concept of this method was developed in the 1980s (e.g., Michard and Roekens, 1983; Reed and Spycher, 1984), and some investigators (e.g., D'Amore et al., 1987; Hull et al., 1987; Tole et al., 1993) have used this technique for predicting geothermal temperature in various geothermal settings. Other researchers have used the basic principles of this method for reconstructing the composition of geothermal fluids and formation brines (Pang and Reed, 1998; Palandri and Reed, 2001). More recent efforts (e.g., Bethke, 2008; Spycher et al., 2011, 2014; Smith et al., 2012; Cooper et al., 2013; Neupane et al., 2013, 2014, 2015a, b; Peiffer et al., 2014; Palmer et al., 2014; Mattson et al., 2015) have been focused on improving temperature predictability of the MEG.

An additional advantage of MEG over traditional geothermometers is that it considers a suite of chemical data obtained from water analyses for temperature estimation. Although MEG has advantages over the traditional geothermometers, it is also subject to the same physical and chemical processes that can violate the basic assumptions of geothermometry. However, MEG also provides a quantitative approach to account for subsurface composition-altering physical and chemical processes through inverse geochemical modeling. Therefore, it is important to reconstruct the composition of geothermal water for estimation of reservoir temperature with a greater certainty.

A newly developed geothermometry tool known as Reservoir Temperature Estimator (RTEst) (Palmer et al., 2014; Mattson et al., 2015) is used to estimate geothermal reservoir temperatures in southeastern Idaho. The RTEst is an inverse geochemical tool that implements MEG with an optimization capability to account for processes such as boiling, mixing, and gas loss. A more detailed description about RTEst can

Table 2. Water compositions of selected hot/warm springs and wells in southeastern Idaho used for temperature estimation. Elemental/species concentrations are given in $mg\,L^{-1}$. The pH was measured in the field.

Springs/wells[a]	T (°C)	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
Pescadero 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] Well/spring 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 Fig. 2, [c] 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); [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).

be found elsewhere (e.g., Palmer et al., 2014; Neupane et al., 2014; Mattson et al., 2015).

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 composition database. 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).

4.3 Reservoir mineral assemblage

Based on the geology of southeastern Idaho and literature assessment of secondary minerals generally associated with the dominant rock and water types, we assumed reservoir mineral assemblages (RMAs) consisting of idealized clays, zeolites, carbonates, feldspars, and silica polymorphs (chalcedony) to estimate equilibrium temperatures using RTEst. Recently, Mattson et al. (2015) reported that the RTEst results in similar temperature estimates for the same water compositions when applied with slightly different RMAs. Selection of one or two unrepresentative minerals in the RMAs produced larger uncertainties in estimated temperatures than the estimated temperatures themselves because of poor convergence (Mattson et al., 2015). Therefore, while selecting the RMAs for RTEst, it is recommended to consider local geology, water chemistry (e.g., pH), and expected range of the reservoir temperatures. For more detailed information on selecting RMAs, see Palmer et al. (2014).

Using an appropriate RMA and measured water composition, RTEst estimates an equilibrium reservoir temperature (as well as a fugacity of CO_2 and mixing with non-thermal water or boiling) by minimizing an objective function (Φ) that is the weighted sum of squares of the saturation indexes for the selected equilibrium minerals:

$$\Phi = \sum (SI_i w_i)^2, \tag{1}$$

where $SI_i = \log(Q_i/K_{i,T})$ for the ith equilibrium mineral (Q_i and $K_{i,T}$ are the ion activity product and temperature-dependent equilibrium constant, respectively for the ith mineral) and w_i is the weighting factor for the ith mineral.

The weighting factors ensure that each mineral that contributes to the equilibrium state is considered equally and the results are not skewed by reaction stoichiometry or differences in analytical uncertainty. There are three options for weighting factors in RTEst: inverse of variance, normalization, or unit weights. They are discussed in more detail by Palmer et al. (2014). In this paper, we used a normalization method for weighting, which assumes that the analytical errors for all thermodynamic components expressed as basis species are equal and that the thermodynamic activity of
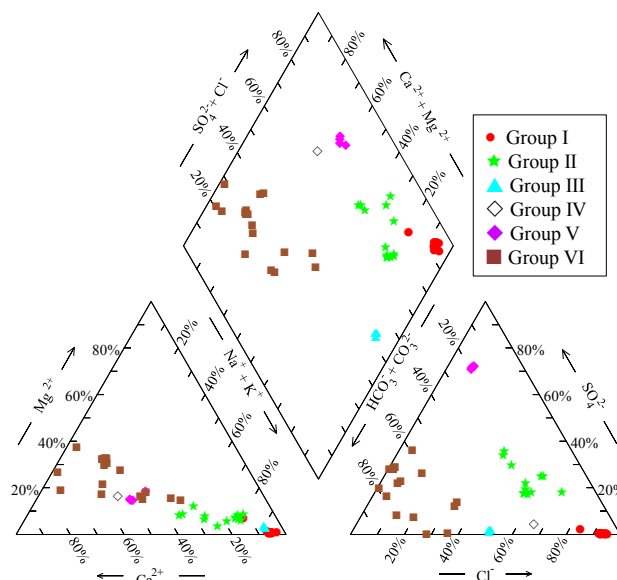

Figure 2. Reported chemistry of waters measured from several hot/warm springs and wells located in southeastern Idaho.

water is unity and is invariant. Examples of weighting factors (normalization factors) for several minerals are given in Palmer et al. (2014).

5 Results and discussion

5.1 Southeastern Idaho spring/well waters

Compositions of waters from hot/warm springs and wells in southeastern Idaho are presented in Table 2. The pH values of southeastern Idaho thermal waters are circum-neutral, ranging from 6.2 to 8.1, with arithmetic mean, median, and standard deviation of 6.87, 6.70, and 0.51, respectively; the field temperatures range between 20 and 84 °C. The aqueous chemistry of these thermal waters shows a large range in total dissolved solids (TDS) from about 250 mg L^{-1} (Downata Hot Spring) to more than 14 000 mg L^{-1} (E. Bingham Well).

The dominant cations in the southeastern Idaho thermal waters are Na and Ca with minor amounts of Mg (Fig. 2). The thermal waters include samples dominated by Cl, HCO_3, or SO_4 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 (Fig. 2) end-members (Ca^{2+}, Mg^{2+}, $Na^+ + K^+$, Cl^-, $HCO_3^- + CO_3^{2-}$, SO_4^{2-}) for classifying water compositions. Six compositional groups were identified within the 50 thermal water samples – Group I: Na-Cl (12 samples), Group II: Na-HCO$_3$-Cl + Ca-SO$_4$ (13 samples), Group III: Na-HCO$_3$-Cl (3 samples), Group IV: Ca-Cl (1 sample), Group V: Ca-SO$_4$ (4 samples), and Group VI: 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.

The Na-Cl and Ca-SO$_4$ (Group I and V, respectively) type waters may have originated via 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 (Preuss Redbeds) in southeastern Idaho. Recently, Welhan et al. (2014) indicated that the high-salinity waters in some deep wildcat petroleum wells may be related to either dissolution of salts from the Preuss evaporites or magmatic waters from a zone as deep as 12–14 km under the BVF in the fold–thrust belt. However, all Na-Cl (Group I) type waters considered in this study are from the surface expressions (hot/warm springs) or from shallow wells and may have originated via 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 (Ayling and Moore, 2013). All Ca-SO$_4$ (Group V) type waters are from hot springs near Bear Lake, located near the Idaho–Wyoming–Utah border. Deep sourced water from a nearby deep wildcat petroleum well (N Eden Federal well with depth > 2500 m) has a very high SO$_4$ concentration; however, this water has low Ca concentration and high Na concentration (Souder, 1985). The Ca-SO$_4$ (Type V) type waters that the Bear Lake hot springs issue may have separate sources of Ca and SO$_4$, or there may be cation exchange reactions involving Ca and Na along the flow path.

The Ca-HCO$_3$ (Group VI) 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, Pescadero Warm Spring), these waters also have a 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 eastern Snake River Plain (ESRP), the Ca-HCO$_3$ (Group VI) 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 represents Ca-Cl (Group IV) type water, the Rockland W-2 well located in the westernmost part of the study area (Fig. 1). Although this water has some similarity with the Ca-HCO$_3$ (Group VI) and Ca-SO$_4$ (Group V) types of water in terms of high Ca content compared to its Na and K concentrations, its high Cl and low Na concentrations make it difficult to assign it as a direct product of a particular type of water–rock interaction.

The remaining two types of waters are mixed waters – Na-HCO$_3$-Cl (Group III) and Na-HCO$_3$-Cl + Ca-SO$_4$ (Group II). The cluster analysis did not identify a separate Na-HCO$_3$ type water, as it is a representative of deep waters in the adjoining ESRP. It is more likely that these waters are Na-Cl type waters that have interacted with carbonate with or without gypsum/anhydride.
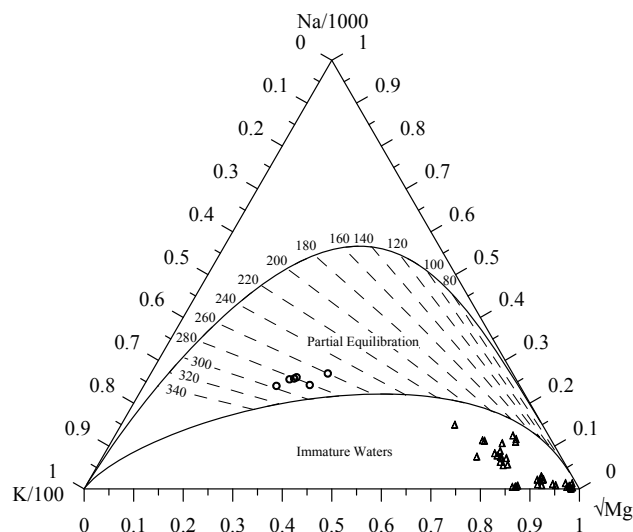

Figure 3. Southeastern Idaho waters from hot springs and wells plotted on Giggenbach diagram (Giggenbach, 1988). All partially equilibrated waters are of Na-Cl (Group I) type waters.

5.2 Southeastern Idaho geothermal reservoir temperatures

5.2.1 Giggenbach diagram

When plotted on a Giggenbach diagram (Giggenbach, 1988), the majority of the southeastern Idaho waters in this study are in the immature zone with some waters in the zone of partial equilibration (Fig. 3). The partially equilibrated waters in Fig. 3 are from hot springs and wells near Preston, Idaho (Battle Creek and Squaw hot springs), and distribution of these waters on Giggenbach diagram indicates that these waters could have interacted with rock at a temperature range of 260–300 °C. The immaturity reflects their low Na content as well as their higher Mg content (Giggenbach, 1988). Waters with high Mg are deemed to be unsuitable for some traditional solute geothermometry, although there have been efforts to develop a Mg correction (e.g., Fournier and Potter, 1979).

5.2.2 Temperature estimated by multicomponent equilibrium geothermometry

Estimates of reservoir temperatures for southeastern Idaho thermal waters shown in Table 2 were made using RTEst and several conventional geothermometers (Table 3). The RMAs that were used in RTEst 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). For thermal waters that do not have measured Al

Table 3. Temperature (°C) estimates for southeastern Idaho thermal waters RTEst and other geothermometers.

Springs/wells[a]	$T^b \pm \sigma^c$	Quartz[d]	Chalcedony[e]	Silica[f]	Na-K-Ca[g]	Types[h]
Woodruff HS	97 ± 3	78	47	49	56	I
E. Bingham W	161 ± 4	117	88	88	193	
Squaw HS-1	179 ± 9	151	125	123	204	
Squaw HS-2	157 ± 6	151	125	123	183	
Squaw HS W-1	175 ± 5	152	127	125	229	
Squaw HS W-2	174 ± 6	150	124	122	217	
Squaw HS W-3	171 ± 7	156	132	129	216	
Battle Creek HS-1	169 ± 5	142	116	114	205	
Battle Creek HS-2	175 ± 6	141	115	113	215	
Battle Creek HS-3	170 ± 5	142	116	114	202	
Battle Creek HS-4	171 ± 4	136	109	107	204	
Wayland HS-1	175 ± 5	125	97	97	230	
Alpine WS	98 ± 9	92	61	63	92	II
Wayland HS-2	144 ± 7	114	85	85	84	
Treasurton WS-1	111 ± 3	105	76	77	78	
Treasurton WS-2	111 ± 9	105	76	77	113	
Cleavland WS	119 ± 7	112	83	84	106	
Maple Grove HS-1	126 ± 4	106	77	78	97	
Maple Grove HS-2	123 ± 4	128	101	100	73	
Maple Grove HS-3	124 ± 3	129	101	101	82	
Maple Grove HS-4	115 ± 7	104	74	75	54	
Maple Grove HS-5	126 ± 6	128	100	99	67	
Maple Grove HS-6	122 ± 5	115	86	87	97	
Auburn HS	107 ± 9	117	88	88	104	
Johnson S	116 ± 13	130	103	102	134	
Ben Meek W-1	106 ± 7	131	104	103	86	III
Ben Meek W-2	106 ± 4	125	97	97	72	
Ben Meek W-3	109 ± 4	131	103	102	73	
Rockland-W2	110 ± 7	118	90	90	93	IV
Bear Lake HS-1	113 ± 7	64	66	73		V
Bear Lake HS-2	111 ± 7	98	68	69	94	
Bear Lake HS-3	107 ± 8	92	61	63	92	
Bear Lake HS-4	121 ± 4	86	55	57	90	
Downata HS	97 ± 3	78	47	49	49	VI
Black River WS	103 ± 3	83	52	55	85	
Pescadero WS	68 ± 8	81	49	52	41	
Henry WS	60 ± 16	92	61	63	89	
Steamboat HS	96 ± 11	100	99	46		
Soda Springs G	59 ± 15	86	55	57	88	
Lava HS-1	94 ± 6	82	51	53	67	
Lava HS-2	94 ± 5	86	55	57	64	
Portneuf R WS-1	100 ± 6	89	59	61	92	
Portneuf R WS-2	101 ± 9	99	69	70	111	
Corral Creek W-1	98 ± 3	77	45	48	98	
Corral Creek W-2	100 ± 4	79	48	51	97	
Corral Creek W-3	98 ± 3	79	48	51	99	
Corral Creek W-4	98 ± 3	79	48	51	100	
Dyer W	121 ± 3	117	88	88	57	
Anderson W	144 ± 4	143	117	115	74	
Rockland-W1	31 ± 4	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 temperature; [d] quartz no steam loss, Fournier (1977); [e] Fournier (1977); [f] Arnórsson et al. (1983); [g] Fournier and Truesdell (1973), Mg correction applied according to Fournier and Potter II (1979); [h] 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).
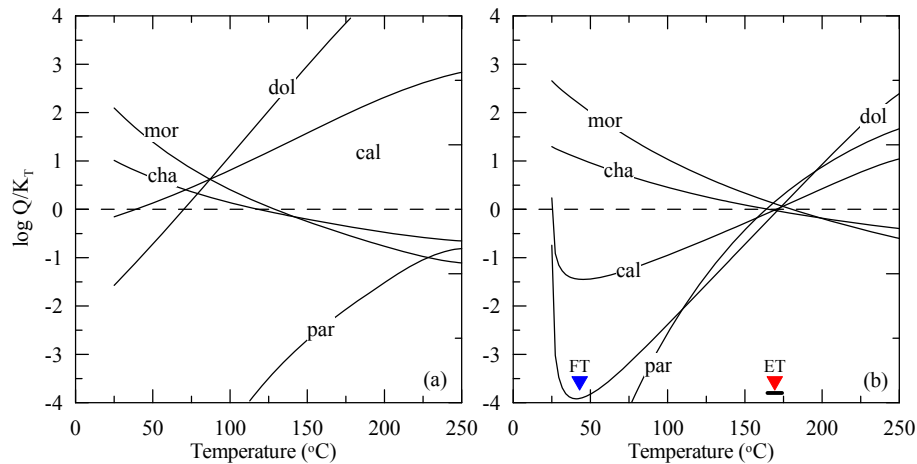
Figure 4. Temperature estimation for Battle Creek hot spring near Preston, Idaho. **(a)** The $\log Q/K_T$ curves for minerals calculated using original water chemistry with K feldspar used as a proxy for Al, **(b)** optimized $\log Q/K_T$ curves (FT: field temperature ($43\,^\circ$C); ET: estimated temperature ($169\,^\circ$C), the dark horizontal bar below ET represents the $\pm$ standard error for the estimated temperature ($\pm5\,^\circ$C); cal: calcite, cha: chalcedony, dol: disordered dolomite, mor: mordenite K, and par: paragonite).

concentration, a value determined from assumed equilibrium with K feldspar was used in the calculations.

In MEG, the reservoir temperature is estimated by assuming a representative RMA with which the fluid in the reservoir is believed to have equilibrated. Next, the activities of the chemical species in solution are determined and the saturation indices [$SI = \log(Q/K_T)$] calculated using the laboratory measured temperature of the sample. This calculation is repeated for a range of temperatures and the resulting SIs recalculated. The apparent reservoir temperature is the one at which all minerals in the RMA are in equilibrium with the reservoir fluid as indicated by near-zero $\log Q/K_T$ values on a $\log(Q/K_T)$ versus temperature plot [$\log(Q/K_T)$ plot] (Reed and Spycher, 1984; Bethke, 2008). In other words, reservoir minerals in equilibrium with the fluid at depth should yield a common equilibrium temperature with a near-zero $\log(Q/K_T)$ value for each mineral at the same temperature; this common equilibrium temperature is assumed to be the reservoir temperature. If all the $\log(Q/K_T)$ curves do not show a common temperature convergence, 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.

Figure 4a shows typical $\log(Q/K_T)$ 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 $40\,^\circ$C (calcite) to over $250\,^\circ$C (paragonite), making the $\log(Q/K_T)$ curves derived from the reported water chemistry minimally useful for estimating tem-

perature. The range of equilibration temperature for the assumed RMA 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.

Three common composition-altering processes are degassing, mixing, and boiling. In particular, the loss of CO_2 from geothermal water due to degassing has a direct effect on pH, and it is often indicated by oversaturation of calcite (Palandri and Reed, 2001). Similarly, dilution of thermal water by cooler water or enrichment of solutes by boiling is indicated by lack of convergence of $\log(Q/K_T)$ curves over a small temperature range. 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, but such a graphical approach becomes cumbersome even for two variables (e.g., temperature and CO_2).

To account for possible composition-altering processes, RTEst was used to simultaneously estimate a reservoir temperature and optimize the amount of H_2O (to account for mixing or boiling) and the fugacity of CO_2 (to account for degassing) (Palmer et al., 2014). The optimized results for Battle Creek hot spring-1 are shown in Figure 4b. Compared to the $\log(Q/K_T)$ curves calculated using the reported water compositions (Fig. 4a), the optimized curves (Fig. 4b) converge to $\log(Q/K_T) = 0$ within a narrow temperature range (i.e., $169 \pm 5\,^\circ$C).

The optimized temperatures and composition parameters for the other southeastern Idaho waters reported in Table 2 were estimated using RTEst in the same manner. The estimated reservoir temperatures and associated standard errors are summarized in Table 3.

Table 4. Mean and standard deviation[a] 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] Fournier and Truesdell (1973), Mg correction applied according to Fournier and Potter II (1979).

5.2.3 Temperature estimates with traditional geothermometers

In addition to RTEst, other traditional geothermometers were also compared (Table 3, Fig. 5). Because most of the waters from thermal springs and wells in southeastern Idaho are geochemically immature (Fig. 3), the use of traditional geothermometers to estimate their temperatures is unreliable. Temperatures obtained with silica polymorphs and Na-K-Ca geothermometers can be quite variable, compared with the RTEst temperatures. As shown in Table 4, group-wise mean chalcedony-based reservoir temperatures are consistently cooler than the mean RTEst-calculated reservoir temperature.

Chalcedony-based reservoir temperatures were calculated using the observed silica concentrations. On the other hand, RTEst reservoir temperatures were calculated with reconstructed solute concentrations (i.e., optimized for water gain/loss). For the majority of samples, the chalcedony-based temperatures are lower than the RTEst-estimated temperatures (Fig. 5a). In general, wherever RTEst indicates that the sample contains an appreciable fraction of additional water, the RTEst temperature is higher than the chalcedony-based temperature for that sample.

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 geothermometry predicts in cooler temperatures in the lower temperature range and hotter temperatures in the upper temperature range compared to the RTEst temperatures (Fig. 5b). The main weakness of Na-K-Ca geothermometer is poor reliability in waters with a significant amount of Mg. Even the Mg-corrected Na-K-Ca temperature estimates have poor reliability if the Mg concentration in thermal waters is high and is controlled by nonchlorite minerals. In southeastern Idaho, the Mg concentration in thermal waters is likely to be controlled by carbonate minerals (limestone/dolomite) as these waters may have interacting with carbonate sequences in the reservoir or along the flow path. Compared to the RTEst temperatures, Na-K-Ca temperatures are lower for all but Group 1 waters. The overprediction of temperature for Group 1 waters is likely
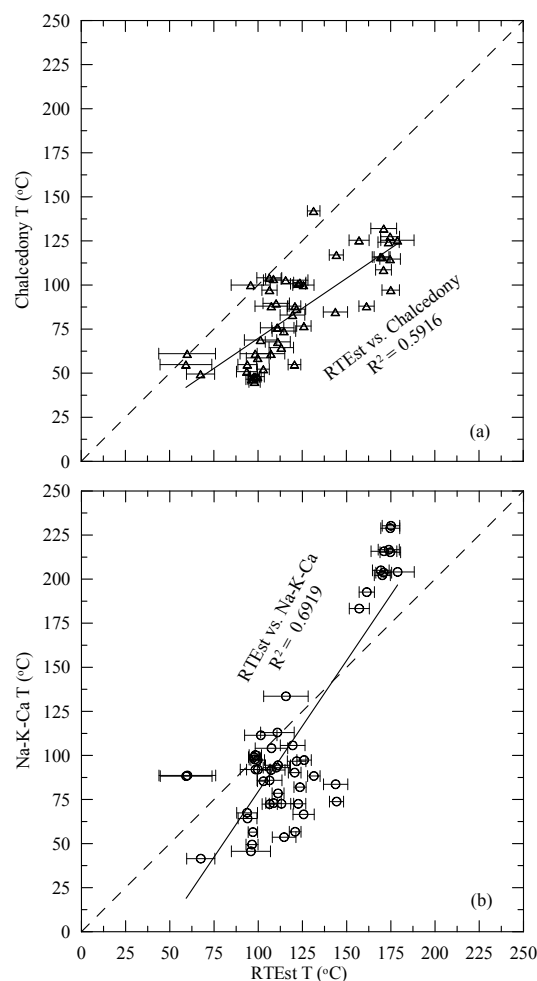

Figure 5. RTEst temperatures versus chalcedony **(a)** and Na-K-Ca **(b)** temperatures for the southeastern Idaho thermal waters. Each horizontal bar represents 1 standard error on each side in RTEst temperature estimate.

caused by the disproportionate (relative to Na and K) loss of Ca to calcite precipitation as suggested by Young and Lewis (1981).

5.2.4 Estimated temperature versus bottom-hole temperature of wildcat petroleum wells

As reported in Table 1, some of the wildcat petroleum wells in the fold–thrust belt in southeastern Idaho have measured BHTs. In general, RTEst-calculated reservoir temperatures appear positively correlated with nearby BHTs, supporting the argument that MEG can be used to predict deep geothermal reservoir temperatures. The North Eden Federal 21–11 well (2618 m) is located east of the Bear Lake, near the border of Idaho, Utah, and Wyoming. This well has slightly lower BHT (84 °C) than the RTEst temperature estimates (107–121 with standard error ±4 to ±8) for nearby Bear Lake hot springs (represented by letter code BL in Fig. 1). Similarly, the RTEst temperature estimate for Alpine Spring (letter code AL in Fig. 1) (98 ± 9 °C) is similar to the BHT of the nearest deep well, Black Mountain-1 (4158 m, 120 °C). However, there are some thermal features that have different RTEst-calculated reservoir temperatures than the BHTs of some nearby wells. For example, the BHTs measured for Federal 1–8 (167 °C) and Federal 1–13 (172 °C) are significantly higher than the estimated temperature (68 ± 8 °C) for the closest spring (Pescadero Warm Spring with PD letter code in Fig. 1). In this case, it is important to note that BHTs of some nearby wells also varied slightly. For example, Bald Mountain-2 (3830 m) has a BHT of 140 °C whereas the nearby Black Mountain-1 well (4158 m) has a BHT of 120 °C. Such variation in temperature at depth in nearby wells may suggest that some deep temperatures measured in wells are not equilibrated (i.e., disturbed by drilling) or could reflect variable proximity to the thermal water flow paths along fault or other heterogeneities.

The highest BHT was recorded for the King 2-1 well (3927 m, 202 °C), but there are 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. However, these hot springs discharge hot waters (up to 84 °C), and some of the recent shallow wells in the area are reportedly producing water with a temperature over 100 °C. New initiatives (e.g., Wood et al., 2015) would help further assess the geothermal potential of the Preston area system.

6 Conclusions

The geological setting coupled with the direct evidence of thermal expressions such as hot/warm springs in the area suggests that southeastern Idaho has good potential for geothermal resources. Our temperature estimates using RTEst with water compositions from southeastern Idaho thermal springs and wells indicate the presence of geothermal reservoirs at depth. Specifically, thermal waters of the Battle Creek hot springs and the Squaw Hot Springs suggest a promising geothermal prospect near Preston, Idaho. The US Department of Energy-sponsored new initiative in the Preston area

with geological, geochemical, and geophysical approaches is expected to further assess the geothermal potential. In several other areas, oil and gas wildcat wells indicate presence of high temperature at depth; however, the moderate RTEst temperature estimates from nearby thermal springs and shallow wells might reflect mixing of local groundwater with deep thermal water and/or re-equilibration of high temperature thermal waters in a shallow low temperature zone.

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