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Chemical Geology 173 (2001) 271–291

CHEMICAL GEOLOGY
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Non-magmatic fracture-controlled hydrothermal systems in the Idaho Batholith: South Fork Payette geothermal system

Gregory K. Druschel *, Philip E. Rosenberg

Department of Geology, Washington State University, Pullman, WA 99164, USA

Received 15 June 1999; accepted 7 April 2000

Abstract

Non-magmatic, fracture-controlled hydrothermal systems located within and near the Idaho batholith have been examined and evaluated for potential application as natural analogues to high-level nuclear waste repositories. Detailed geochemical, petrologic, and structural studies of hot springs along the South Fork of the Payette River (SFPR) in central Idaho, USA, have been combined to assess the nature of a hydrothermal aquifer system along a major brittle shear zone.

Geochemical modeling of thermal spring water chemistry with SOLVEQ [Reed, M.H., 1992. SOLVEQ: Users Manual. Unpublished Report, Dept. Of Geological Sciences, University Of Oregon, Eugene, OR.] indicates reservoir temperatures of 85°C to 160°C, and an alteration assemblage consisting of several clays, chlorite, serpentinites, quartz, albite, microcline, and micas. Geothermometry calculations, coupled with independent estimates of geothermal gradients based on heat flow and regional geology, allow estimation of reservoir depths ranging from 2.4 to 6.7 km. Comparison of trace element, rare earth element, and SOLVEQ alteration assemblages supports the hypothesis that waters equilibrated with granitic rocks at depth and ascended to the surface in a short time span. Stable isotope data ($\delta D = -138.4‰$ to $-148.4‰$ and $\delta^{18}O = -17.5‰$ to $-19.4‰$) indicate that thermal waters of meteoric origin have interacted with country rocks at a very high water/rock (W/R) ratio. Systematic variations in δD values and the amount of $\delta^{18}O$ shift for thermal waters suggest that recharge probably occurs at much higher elevations than discharge. Fracture orientations, petrologic relations, and geochemical differences between adjacent springs delineate separate hydrothermal systems along the major structure defining the course of the SFPR.

Hydrothermal systems along the SFPR exist primarily as small circulation cells which are probably influenced by minor changes in geothermal gradients associated with different rock types. These subsystems are dependent on periodic seismic or microseismic activity to maintain flowpaths. A working model of deep, fracture-controlled hydrothermal convection systems based on the SFPR Geothermal System may be used to evaluate the long-term behavior of artificial hydrothermal systems associated with high-level nuclear waste repositories.

Keywords: Hydrothermal; Nuclear waste repositories; Geochemical models

1. Introduction

Hydrothermal systems worldwide have been studied in order to better understand hydrothermal ore deposits, geothermal energy resources, and potential

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* Corresponding author. Present address: Department of Geology and Geophysics, University of Wisconsin, 1215 W. Dayton St., Madison, WI 53706, USA. Fax: +1-608-262-0693.

E-mail address: druschel@ice.geology.wisc.edu (G.K. Druschel).

PII: S0009-2541(00)00280-1

high-level nuclear waste storage facilities. Studies of the geochemistry, hydrology, and structure of natural hydrothermal systems may provide information allowing more accurate prediction of the short and long-term behavior of artificial hydrothermal systems that would be generated by the subsurface emplacement of spent nuclear fuel. Studies of volcanogenic and non-magmatic hydrothermal systems worldwide have focused on the structure of hydrothermal systems and on chemical reactions associated with the transport of several trace elements and radionuclides of concern in possible release scenarios (Nordstrom et al., 1989; Michard et al., 1991; Olsson and Gale, 1995; Wood et al., 1997). Because of the requirement to mitigate the release and transport of radionuclides from a nuclear waste facility for tens of thousands of years, the long-term behavior of hydrothermal systems is an important concern in assessing the feasibility of potential sites (Buscheck and Nitao, 1993). The Department of Energy considers the long-term importance of potential fast pathways for water and radionuclide movement through a repository's engineered and natural barriers to be an important and incompletely understood process (DOE-YNP-0022, 1997). The use of deep granitic environments as nuclear waste repositories has been investigated by several countries, including: Canada, Sweden, France, Switzerland, Japan, and Finland (Olsson and Gale, 1995). The Idaho batholith has been proposed as a natural laboratory for the assessment of the geochemical response of lanthanides and actinides to interaction with hydrothermal solutions (Wood et al., 1997; van Middlesworth and Wood, 1998).

In this study a flow model for the hydrothermal system along the South Fork of the Payette River has been constructed based on analyses of thermal spring water chemistry and previous studies of petrology and structural geology in the vicinity of hot springs (Reed, 1986; Dingee, 1987). Temperatures at depth in this hydrothermal system are comparable to those predicted due to the decay of spent nuclear fuel over 1000 to 20,000 year periods (Buscheck and Nitao, 1993).

Hydrothermal systems in the Idaho batholith have probably been operating for many millions of years; stable isotopic evidence of altered Idaho batholith rocks indicates major hydrothermal activity following emplacement of Tertiary plutons (Criss and Taylor, 1983). Considering the nature of the heat source (Reed, 1986; Dingee, 1987), the observed activity of faults in the area through time (Kiilsgaard and Lewis, 1985), and the presence of a number of ore bodies of hydrothermal origin (Bennett, 1980), it is likely that these hydrothermal systems have been active for millions of years. In addition, the heterogeneous heat flow field caused by variations in U-, Th-and ^{40}K-content in different plutonic bodies of the Idaho batholith may be analogous to the localized perturbation of a hydrodynamic system in response to the emplacement of spent nuclear fuel (Swanberg and Blackwell, 1973). The long-term behavior of engineered barriers to potential fast pathways for fluid migration may also be analogous to the behavior of relatively impermeable dikes in the South Fork Payette Geothermal System. The effects of seismic/microseismic activity, petrology, structure, heat flow, hydrology, water/rock interaction, and alteration mineralogy on the geochemistry of thermal water in these systems may reveal the overall behavior of a potential repository. Studies of hydrothermal systems in the Idaho batholith should, therefore, provide insights into the long-term and large-scale behavior of hydrothermal systems that may develop as a result of the emplacement of spent nuclear fuel in a geologic medium.

2. Geological setting

The South Fork Payette Geothermal System (SFPGS) is located in the west-central portion of the Atlanta lobe of the Idaho batholith, a 25,000 km^2 granitic pluton in central Idaho (Fig. 1). Seven geothermal springs have been studied in this area (Fig. 1, Table 1), with outflow temperatures ranging from 37.1°C to 83.9°C. These thermal springs issue directly from fractured granitic rocks and are associated with crossing brittle structures and dikes.

In the Idaho batholith, Cretaceous plutonic rocks primarily intrude Belt and pre-belt basement schists and gneisses at depths between 13 and 25 km (Hyndman, 1983). Hornblende–biotite granodiorite, porphyritic granodiorite, and a biotite granodiorite form a 10–20-km-wide border zone around most of the

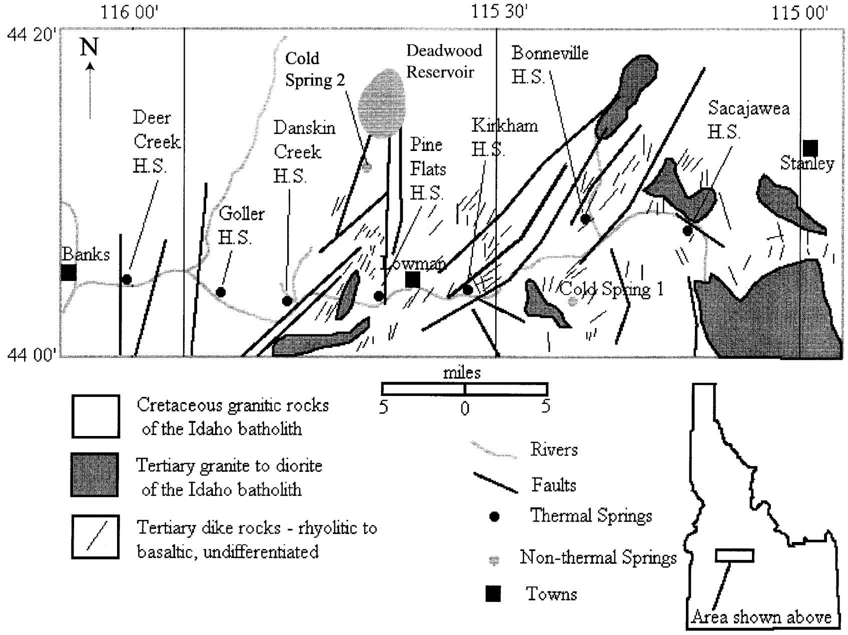

Fig. 1. Simplified geologic map of the SFPGS. Adapted from Reed (1986), Dingee (1987), and Lewis et al. (1992).

batholith. Leucocratic granite intruded the other Cretaceous granitic rocks shortly after their emplacement (Kiilsgaard and Lewis, 1985; Clarke, 1990). The area then underwent a period of substantial uplift and erosion; at least 12 km of material was eroded after emplacement of the Cretaceous rocks of the Idaho batholith (Hyndman, 1983).

Granite and quartz monzonite intruded Cretaceous rocks at shallow levels between 42 to 50 Ma (Bennett and Knowles, 1985; Reppe, 1997). Each plutonic rock type has an associated hypabyssal equivalent, present as a number of dike swarms throughout the batholith. Dikes were emplaced along pre-existing structures, but have subsequently been deformed. Joint densities in dike rocks are greater than in surrounding plutonic rocks (Reed, 1986), but have a strong orientation which may make them relatively impermeable to fluid flow oriented perpendicular to the fracture planes. Eocene Challis volcanics cover most of the eastern margin of the batholith, and are cogenetic with Tertiary plutonic and hypabyssal dike rocks.

Chemical analyses of hundreds of Cretaceous and Tertiary plutonic and dike rocks from the Atlanta Lobe have been published by Bennett and Knowles (1985), Kiilsgaard and Lewis (1985), Reed (1986), Reed and Rosenberg (1986), Dingee (1987, 1988), and Clarke (1990). F^- contents of the granitic rocks are as much as 100 times higher than surrounding sedimentary or volcanic rocks (Foley and Street, 1988). The Cretaceous and Tertiary granitic rocks are characterized by significant differences in Rb–Sr ratios and in U, Th, and ^{40}K concentrations (Swanberg and Blackwell, 1973). These differences create much higher geothermal gradients within and near Tertiary plutons compared to those within Cretaceous plutons (32°C/km and 25°C/km, respectively), resulting in a highly heterogeneous heat flow field in the area.

Seismic activity is required to maintain flow through fracture-controlled hydrothermal systems. The United States Geological Survey Earthquake database http://wwwneic.cr.usgs.gov/) lists three earthquakes of magnitude 4.0 on the Richter scale within 30 miles of the of the SFPGS since 1976; focal depths for two events are 10 km and 33 km for the third. The magnitude 7.3 Borah peak earthquake of October 28, 1983 was located along a northwest-

Table 1
Major element data for thermal and non-thermal springs in the SFPGS. Data for Cold Springs 1 and 2 from Young (1985)

	Temperature (C)	pH	Conductivity (umhos)	Ca^{2+} (mg/l)	Na^{+} (mg/l)	K^{+} (mg/l)	F^{-} (mg/l)	Cl^{-} (mg/l)	NO_2^{-} (mg/l)	PO_4^{3+} (mg/l)	SO_4^{2+} (mg/l)	Field alkalinity (as mg/l $CaCO_3$) (mg/l)	CO_3^{2+} (mg/l)	HCO_3^{-} (mg/l)	SiO_2^{0} (mg/l)
Deer Creek H.S.	78.2	8.0	1300	3.0	149	5.4	11.5	31.4	1.6	1.9	77.1	133	1.4	162.2	102.6
Goller H.S.	42.5	9.2	475	1.2	84.8	1.3	14.7	7.0	1.9	2.3	28.8	91	7.8	110.9	64.8
Danskin Creek H.S.	37.1	9.4	400	1.3	79.7	0.9	13.1	5.2	2.5	2.3	37.2	74	7.8	90.2	52.4
Pine Flats H.S.	59.4	9.2	556	1.1	88.0	1.1	16.1	6.0	0.9	2.0	32.5	75.3	5.0	91.8	61.9
Kirkham H.S.	60.3	9.2	500	1.3	76.4	1.1	13.5	3.7	2.4	2.4	33.2	57	3.1	70.7	59.4
Bonneville H.S.	83.9	9.1	975	1.2	92.8	2.7	16.4	6.0	3.5	2.4	45.0	74	0.9	90.2	98.2
Sacajawea H.S.	65.1	9.2	600	1.2	89.8	2.6	14.5	8.6	1.9	2.3	41.9	89	5.9	108.5	85.1
Cold spring 1	4.5	6.6	52	3.8	3.1	0.6	0.1	0.4	0.1	0.0	1.9	24	0	29.0	14.0
Cold spring 2	5.0	6.2	22	2.1	2.0	0.3	0.1	0.3	0.1	0.1	5.0	9	0	11.0	11.0

trending fault to the north of the study area; significant changes in the flow rates of thermal springs were observed near the epicenter (Wood et al., 1985). Monitoring of microearthquakes to the west of the study area suggests frequent microseismic activity; events were recorded in swarms of several per day (Applegate and Donaldson, 1977).

The predominant rock types exposed in the SFPGS are Cretaceous units of the Idaho batholith, which have been described as several types of micaceous granites, leucogranites, and granodiorites (Fig. 1; Reed, 1986; Dingee, 1987). Tertiary plutons, ranging in composition from granite to diorite, are also located throughout the SFPGS (Fig. 1). A large number of Tertiary dikes, typically less than 20 m in width, strike in a northeast–southwest direction (Fig. 1; Reed, 1986); their composition ranges from rhyolitic to basaltic (Reed, 1986).

Thermal springs in the SFPGS are located along the major shear zone, which defines the course of the South Fork of the Payette River, where the shear zone is intersected by crossing structures and/or dikes (Fig. 1; Reed, 1986; Dingee, 1987). Fault zones in the study area range from 50 to 150 m wide. All have very steep dips with both dip-slip and strike-slip components and no ductile features. Fracture density increases towards the center of these fault zones (Reed, 1986; Dingee, 1987).

3. Methods

3.1. Field

The temperature, pH, and conductivity of each thermal spring were measured in situ as close to the discharge point as possible. Temperature and pH were measured with an Orion Research model 230 temperature/pH meter with a standard thermistor and a VWR glass electrode. Conductivity was measured with an YSI model 33 S-C-T meter and reported in μmhos. Samples for alkalinity titrations were collected in a 1-l cylinder and analyzed using a digital titrator and a methyl red–bromocresol green indicator.

Two water samples were collected as close as possible to the discharge point at each spring vent in 250-ml glass bottles sealed with Teflon caps. Each sample was filtered with a vacuum filtration system through a 0.45 μm Millipore filter immediately after collection. After filtration one of the samples was acidified to a pH of less than two using a 50% trace metal grade nitric acid solution for major cation, silica and trace element analyses; the other sample was left unacidified for anion and stable isotope analyses. Unfortunately, no acidified field blanks were collected, and there may be some error (a few ppb at most) in the absolute trace metal concentrations of the samples. The sampling procedure was identical for all samples, and the relative distribution of trace metals between the samples may be considered accurate.

3.2. Laboratory

All samples were refrigerated and analyzed within 6 months of collection. Sodium, potassium, and calcium were analyzed by flame atomic absorption spectrophotometry. Samples were analyzed for sodium and potassium using CsCl and $SrCO_3$ additions to suppress matrix interferences. The error associated with atomic absorption spectrophotometry using similar procedures has been estimated to be between 4% and 8% (Fishman and Friedman, 1989). Cl^-, F^-, Br^-, SO_4^{2-}, NO_3^-, and PO_4^{3-} were analyzed by ion chromatography (IC) using an isocratic elution with a sodium bicarbonate–sodium carbonate eluent over a run time of 10 min. Analyses of several samples during the course of many runs suggests that the error is less than 5%. Silica analysis was carried out by the molybdate blue colorimetric method (Fishman and Friedman, 1989). Error associated with this method has been estimated at between 4% and 10% (Fishman and Friedman, 1989).

Trace element (Li, B, Mg, Mn, Fe, Co, Zn, Rb, Sr, Mo, Sb, Ba, Pb, U, As, and W) analysis was performed by inductively coupled plasma mass spectrometry (ICPMS). Blanks composed of nanopure (18 MΩ) deionized water were run with each set of analyses to monitor the extent of contamination. The average blank signal was subtracted from the sample signal and concentrations determined by reference to a prepared standard. Analyses of several samples during the course of the run indicate that the estimated error associated with the analytical data is less than 5% of the measured concentrations for samples above 5 ppb, and less than 10% for measured concentrations below 5 ppb.

Stable isotope analyses were performed by gas mass spectrometry using an OzTech isotope ratio reference gas WASH-329C carbon dioxide standard. Hydrogen/deuterium isotope samples were prepared by reacting the sample with 100% phosphoric acid and zinc and heating to 500°C before analysis (Coleman et al., 1982). Oxygen isotope samples were prepared by extracting CO_2 equilibrated with purified samples according to the method established by Epstein and Mayeda (1953). Isotope values are reported in δ per mil notation relative to VSMOW (Faure, 1986). Maximum error estimated for this procedure is 0.2‰ δD and 2‰ $\delta^{18}O$.

3.3. Data processing

Two different approaches, activity diagrams and SOLVEQ modeling, have been utilized to investigate the equilibrium state of thermal waters in the SFPGS and to define the equilibrium mineral assemblages of the hydrothermal systems at depth. Logarithmic activity diagrams are often used to predict the mineral assemblages associated with geothermal systems (Giggenbach, 1981).

Several computer codes were used to define the speciation of the thermal fluids, geothermometry, mineral equilibrium assemblages, theoretical water/rock reactions, and reaction pathways (Table 2). The program SOLVEQ (Reed, 1982, 1992) was used to speciate thermal waters, calculate partial heterogeneous equilibrium, and calculate saturation indices of several hundred minerals as a function of temperature. In all SOLVEQ simulations Al^{3+} concentrations were determined by forcing equilibrium with maximum microcline according to the method established by Pang and Reed (1998). SOLVEQ calculates saturation indices for over 300 minerals. Those that cross the log $Q/K=0$ line over a narrow temperature range define a possible equilibrium assemblage and equilibrium temperature. The minerals output by SOLVEQ include minerals that are

Table 2
Geochemical modeling programs used in this study, their respective thermodynamic databases, and the information derived from their use with respect to the geothermal systems in the Idaho Batholith

Geochemical modeling program	Thermodynamic database (reference)	Information required	Application of program results
SOLVEQ (Reed, 1992)	SOLTHERM database (Reed, 1992) based on rigorous evaluation of several thermodynamic data sources	Concentrations of all aqueous species involved in possible water–rock equilibrium reactions	Speciation and identification of possible equilibrium phases at depth with site-specific geothermometry
CHILLER (Reed, 1992)	SOLTHERM database (Reed, 1992) based on rigorous evaluation of several thermodynamic data sources	Initial water composition (analogous to recharge water) and mineral phases reacting with that water	Forward calculation of equilibrium water composition and secondary mineral phases and volume changes
Act2 (Geochemist's Workbench 3.0, Bethke, 1998)	LLNL thermochemical database (Delany and Lundeen, 1990) based in large part on SUPCRT92 (Johnson et al., 1991)	Activities of aqueous species included in all possible speciation reactions concerning the system of interest	Identification of possible equilibrium phases given a specific water composition
React (Geochemist's Workbench 3.0, Bethke, 1998)	LLNL thermochemical database (Delany and Lundeen, 1990) based in large part on SUPCRT92 (Johnson et al., 1991)	Initial water composition (analogous to recharge water) and mineral phases reacting with that water	Forward calculation of equilibrium water composition and secondary mineral phases
NetPath 2.0 (Plummer et al., 1994)	WATEQ4F database (Ball and Nordstrom, 1991)	Initial and final water compositions, including all aqueous species involved in possible water–rock equilibrium reactions	Identification of possible mineral reactions responsible for changes in fluid chemistry between two points

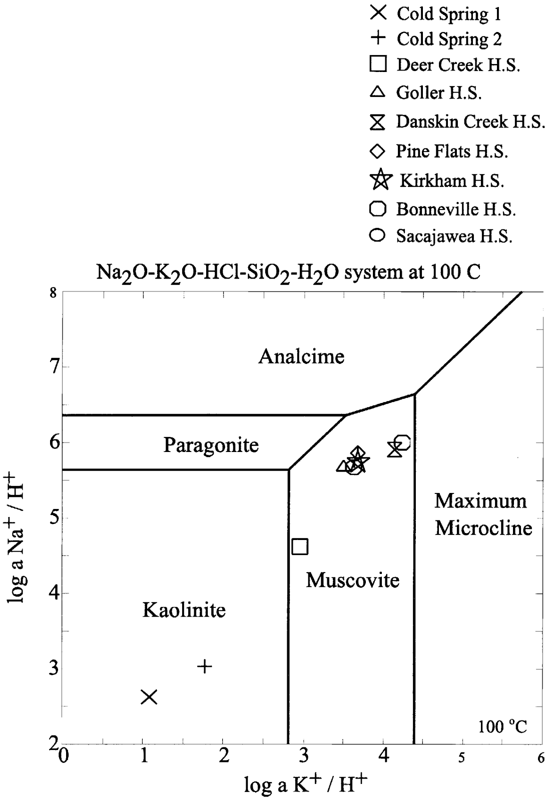

Fig. 2. Activity diagram for the system Na_2O–K_2O–SiO_2–HCl–H_2O at 100°C and 150°C calculated using the geochemical modeling code Act2 (Bethke, 1998). SiO_2 activity was set at quartz equilibrium. The activity of Na^+, K^+, and H^+ for each thermal spring (designated H.S.) was calculated using the geochemical modeling code React (Bethke, 1998).

metastable or unstable at the pressure and temperature defined in the system, due to a mathematical artifact of the SOLVEQ modeling procedure. The user must summarily evaluate this assemblage; the stability of each mineral must be considered before a decision can be made on an appropriate equilibrium assemblage. Tremolite, for example, is present in the SOLVEQ output for Kirkham Hot Springs (Fig. 3) and is supposedly saturated at 105°C. However, it is probably unstable below 500°C (Deer et al., 1992) and is thus excluded from the final diagram.

The convergence of selected minerals over a narrow temperature interval at log $Q/K = 0$ yields an estimate of equilibrium temperature specific to the chemistry of each spring. This circumvents some of the assumptions in choosing an accurate geothermometer from one or several of the traditional calculated geothermometers such as quartz, Na–K, and Na–K–Ca (Fournier and Truesdell, 1973; Giggenbach, 1988; Truesdell and Singers, 1974). A geothermometer based on several minerals which are probably part of the alteration assemblage provides a more accurate and realistic estimate of reservoir temperature (Michard and Roekens, 1983; Reed and Spycher, 1984; Michard et al., 1986; Bethke, 1996).

The forward reaction modeling programs CHILLER (Reed, 1992) and React (Bethke, 1998) were used to evaluate theoretical water/rock reactions between hypothetical initial waters and rocks compositionally similar to those in the field area.

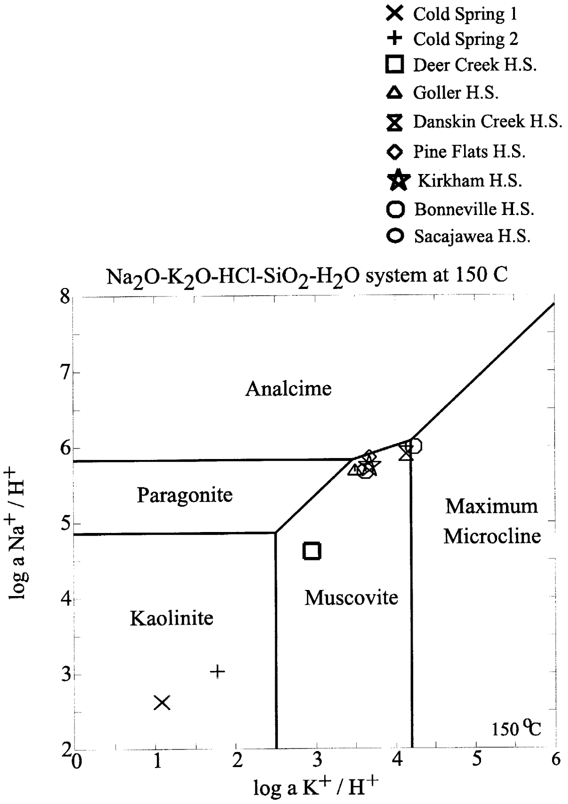

Fig. 3. Activity diagram for the system Na_2O–K_2O–SiO_2–HCl–H_2O at 150°C calculated using the geochemical modeling code Act2 (Bethke, 1998). SiO_2 activity was set by equilibrium with quartz. The activity of Na^+, K^+, and H^+ for each thermal spring (designated H.S.) was calculated using the geochemical modeling code React (Bethke, 1998) and plotted on the diagram.

Volumetric changes associated with these reactions were calculated by comparison of the change in net molar volume and reaction progress. The inverse reaction model NETPATH 2.0 (Plummer et al., 1994) was used to evaluate possible reaction pathways in order to explain observed changes in water chemistry between adjacent and hydraulically connected springs.

4. Results

4.1. Computer modeling

Activity diagrams for the system Na_2O–K_2O–SiO_2–HCl–H_2O at 100°C and 150°C and the aqueous activities of Na^+, K^+ and H^+ for the thermal and non-thermal springs were calculated using the geochemical modeling codes Act2 and React, respectively (Bethke, 1998). At approximate reservoir temperatures (100°C to 150°C), the thermal waters lie in the muscovite stability field (Figs. 2 and 3). The stability fields of smectite and illite, compositionally variable clay minerals, are thought to lie within or close to the stability field of muscovite on these diagrams, although suitable thermodynamic data does not exist to show their fields of (meta)stability (Savage et al., 1987). Non-thermal springs (Young, 1985) lie in the kaolinite stability field (Figs. 2 and 3). The activity diagrams suggest that the thermal waters are stable with respect to smectites (beidellite), illite and aluminum oxide minerals.

Figs. 4 and 5 are selected SOLVEQ diagrams for two of the seven thermal springs in the SFPGS, showing mineral assemblages and equilibrium temperature ranges. Similar diagrams for the rest of the springs in the SFPGS may be found in Druschel (1998). A summary of the assemblages of stable minerals calculated with SOLVEQ for all of the springs in the SFPGS is given in Table 3.

Inspection of SOLVEQ diagrams for all of the thermal springs in the SFPGS (Figs. 4 and 5; Druschel, 1998) reveals differences in mineral assemblages (Table 3) and in apparent reservoir temperatures which may be due to:

1. Variations in wall rock petrology or exposure of fresh mineral surfaces,
2. Presence of metastable minerals due to differences in flow rates and kinetic effects,

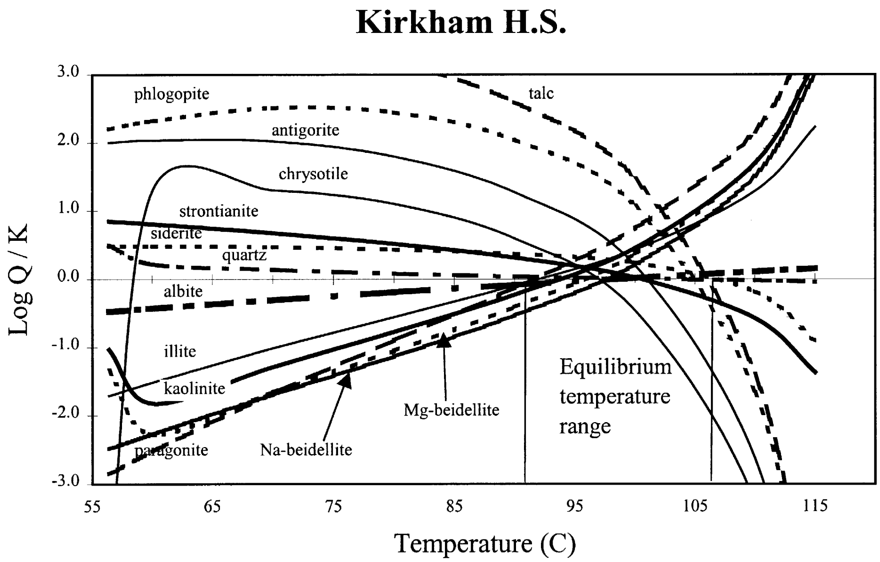

Fig. 4. Log Q/K — Temperature diagram for Kirkham Hot Springs calculated using SOLVEQ (Reed, 1992). The calculations were made assuming partial heterogeneous equilibrium with respect to microcline.

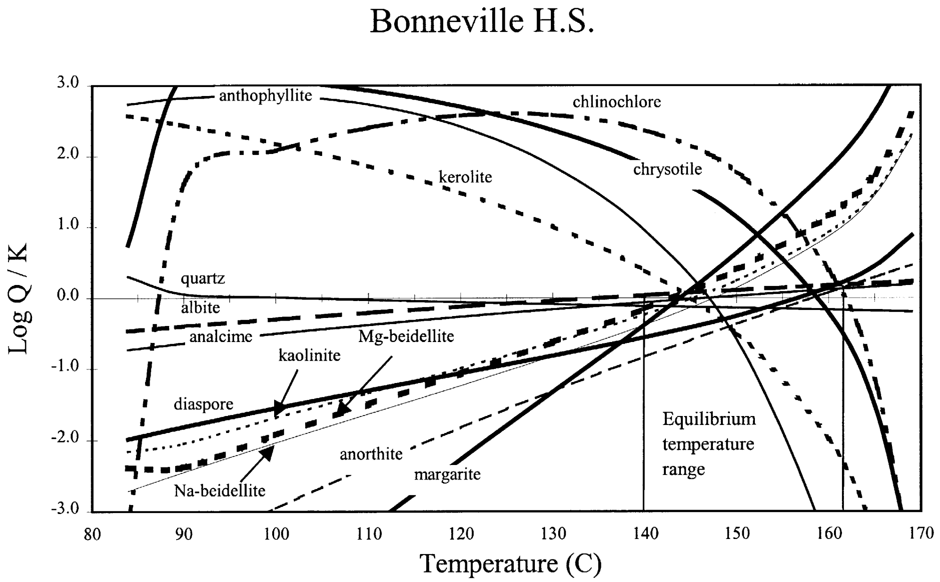

Fig. 5. Log Q/K — Temperature diagram for Bonneville Hot Springs calculated using SOLVEQ (Reed, 1992). The calculations were made assuming partial heterogeneous equilibrium with respect to microcline.

3. Possible variations in source water composition,
4. Re-equilibration of some waters at lower temperatures,
5. Differences in reservoir temperatures.

Apparent alteration assemblages based on SOLVEQ modeling include zoisite, clinozoisite, and diaspore at the three higher temperature springs (Deer Creek, Bonneville, and Sacajawea Hot Springs). These three minerals are absent from the other springs in the SFPGS (Table 3), suggesting they are not stable at lower temperatures in this geothermal system.

The alteration assemblages consisting of albite, quartz, microcline, and clay minerals (beidellite, illite, chlorite, serpentine) and, at higher temperatures, zoisite, are consistent with granite–water experiments conducted by Charles and Bayhurst (1983), Savage (1986), Savage et al. (1987), Zuddas et al. (1995), and Azoural and Fouillac (1997), in which the products of the reaction of granite, granodiorite, and quartz monzonite with water were identified at temperatures ranging from 80°C to 180°C. Smectites and chlorites have also been identified as products of intermediate argillic alteration in a number of lower temperature hydrothermal ore deposits (Hedenquist and Brown, 1989; Reed, 1997).

Hypothetical water compositions have been calculated from the products of a reaction of local groundwaters (Young, 1985) with local rocks (Zimmer, 1996; Zimmer and Rosenberg, 1997) using the React geochemical modeling code (Bethke, 1998). The results (example, Fig. 6) are in partial agreement with interpretations based on the activity diagrams (Fig. 2) and SOLVEQ calculations. Both of the approaches used in this study to determine alteration assemblages indicate that the thermal waters are in equilibrium with clay minerals. There are considerable differences, however, as to the mineral species which are in apparent equilibrium with these waters. Disagreement between these approaches suggests that the evolution of geothermal systems may not be accurately modeled by simple thermodynamic equilibrium models. Accurate prediction of specific mineral phases that are formed as a result of hydrothermal alteration may require a more rigorous kinetic formulation.

Forward reaction modeling with CHILLER (Reed, 1992) shows that the precipitation of clay minerals from the reaction of water with granitic minerals at

Table 3
Alteration minerals predicted from SOLVEQ modeling for the 7 thermal springs of the SFPGS. In each calculation, maximum microcline was assumed to be at equilibrium

Spring name	Albite	Analcime	Antigorite	Anorthite	Barite	Mg–Beidellite	Na–Beidellite	Calcite	Chrysotile	Clinochlore	Clinozoisite	Cuprite	Diaspore
Deer Creek H.S.	X				X	X	X				X		X
Goller H.S.	X		X			X	X		X				
Danskin Creek H. S.	X	X	X			X	X		X	X		X	
Pine Flats H. S.	X	X	X			X	X					X	
Kirkham H. S.	X		X			X	X		X			X	
Bonneville H. S.	X	X		X		X	X		X				X
Sacajawea H. S.	X	X	X			X	X	X	X	X	X		X

100°C results in a significant increase in volume, decreasing porosity as the reaction progresses. Thus, periodic seismic activity is required to keep flow-paths open over any significant length of time.

4.2. Geothermometry

Reservoir temperatures can be estimated from the equilibrium temperature ranges observed in SOLVEQ diagrams (Table 4, Figs. 4 and 5). The Na–K–Ca geothermometer (Fournier and Truesdell, 1973) yields results consistently comparable to those obtained from SOLVEQ (Table 4). The minimum and maximum values of reservoir depths (Table 4) have been calculated using the minimum and maximum values for the SOLVEQ equilibrium temperatures and geothermal gradients (24°C/km to 32°C/km) based on concentrations of radioactive elements and

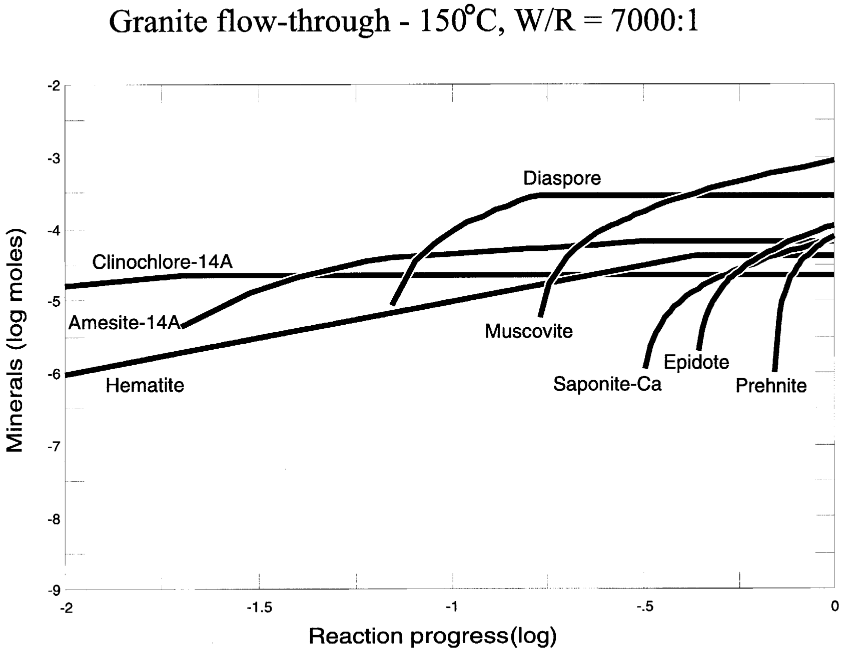

Fig. 6. Results of the minerals predicted for the reaction of a granite of local composition (Dingee, 1986) with a local groundwater (Parliman, 1986) calculated using the geochemical modeling code React (Bethke, 1998).

Illite	Kaolinite	Kerolite	Margarite	Microcline	Paragonite	Phlogopite	Prahnite	Pyrophyllite	Quartz	Sepiolite	Strontianite	Tremolite	Zoisite
X		X	X	X		X	X	X	X			X	X
X	X	X		X	X			X	X		X	X	
X	X	X		X	X				X	X			
X	X	X		X	X				X				
X	X			X	X				X		X		
	X	X	X	X					X				
X	X	X	X	X	X				X				X

heat flow characteristics (Druschel, 1998). The wide range in the estimated depths also includes uncertainty in defining the spatial distribution of the heterogeneous heat flow field for the area, and possible differences in equilibration depth due to hydrodynamic differences (effect of changes in fracture-induced permeability) throughout the system.

4.3. Stable isotopes

Oxygen and hydrogen isotope ratios provide an insight into the source of water and the extent of water/rock reaction in geothermal systems. Thermal waters around the world show a shift in $\delta^{18}O$ from their original meteoric source to higher values, which is attributed to the progressive equilibration of meteoric water with silicate and carbonate rocks (Craig, 1961). The small magnitudes of $\delta^{18}O$ shifts in thermal waters of the SFPGS (Fig. 7, Table 5) are indicative of a system with very high water/rock (W/R) ratios or a system which has not had time to equilibrate with the surrounding rocks. Smaller W/R ratios would result in large $\delta^{18}O$ shifts, inasmuch as igneous rocks in the Idaho batholith have $\delta^{18}O$ of $+9.3 \pm 1.5‰$ and δD of $-70 \pm 5‰$ (Criss and Taylor, 1983). ^{14}C dating of thermal waters in the Idaho batholith by Young (1985) indicates residence times that exceed 9000 years; therefore, it seems unlikely the small $\delta^{18}O$ shift is due to kinetic constraints. The $\delta^{18}O$ shifts of these thermal springs from the local meteoric water line (Craig, 1961) is variable, from 0‰ to just over 1‰, suggesting that seismic disturbances which expose fresh mineral surfaces are localized events.

δD values are often used to investigate water sources because there is very little hydrogen in rocks and, therefore, the δD of water equilibrating with those rocks should be effectively unchanged except

Table 4
Geothermometry results from SOLVEQ modeling and the Na–K–Ca geothermometer (Fournier and Truesdell, 1973). Depth range represents uncertainty in defining reservoir temperature (from SOLVEQ modeling) and geothermal gradients

	Deer Creek H.S.	Goller H.S.	Danskin Creek H.S.	Pine Flats H.S.	Kirkham H.S.	Bonneville H.S.	Sacajawea H.S.
Equilibrium temperature range from SOLVEQ models (°C)	130–150	95–108	82–95	92–105	90–102	142–160	122–142
Na–K–Ca geothermometer (°C) (Fournier and Truesdell, 1973)	144	107	94	98	102	134	133
Approximate reservoir depth (km)	4.1–6.0	3.0–4.3	2.6–4.2	2.9–4.2	2.8–4.1	4.4–6.4	3.8–5.7

at very low W/R ratios (Craig, 1961). Higher recharge elevation and/or cooler temperature will have a significant effect on the amount of δD depletion of these waters (Faure, 1986). Hobba et al. (1979) suggested values of −8‰ δD depletion per 1000 m increase in elevation and −4‰ δD depletion per degree for recharge waters in the Idaho batholith. Relief in the SFPGS exceeds 1000 m within a 5-km radius of each thermal spring. Furthermore, recharge temperatures probably decreased during Pleistocene glaciation when temperatures were approximately 3.5°C cooler than at present (Young, 1985). However, temperature declines cannot be accurately estimated during the Pleistocene because weather patterns then may not correspond exactly to those of today.

Examination of the δD and $\delta^{18}O$ values for the SFPGS (Table 5 and Fig. 7) reveals a decrease with respect to location and elevation which reflects the increasing elevation eastward across the system. The increasing $\delta^{18}O$ depletion eastwards may also be indicative of greater frequency or intensity of seismic activity. Using the equation given by Hobba et al. (1979), the elevation changes between springs in the SFPGS (Table 5) closely predict the observed δD values. This suggests that the primary cause of the shift in δD from meteoric to thermal waters is higher recharge elevation.

Springs with similar δD vs. $\delta^{18}O$ ratios are also spatially related (particularly Pine Flats–Kirkham Hot Springs and Bonneville–Sacajawea Hot Springs) suggesting that hydraulic communication exists within these groups (Figs. 1 and 7). The separation of groups (Fig. 7) implies that the system is probably composed of small convection cells that encompass one or two thermal springs, with little or no hydraulic communication between them.

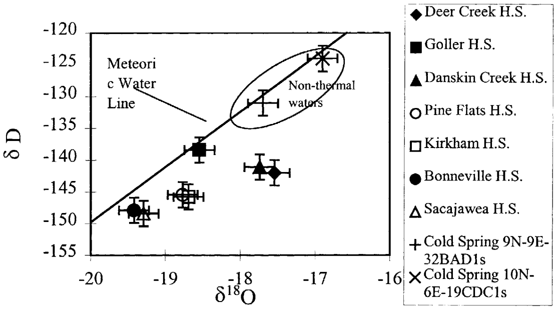

Fig. 7. δD and $\delta^{18}O$ values for thermal and non-thermal springs in the SFPGS presented in per mil notation relative to VSMOW. Meteoric water line from Craig (1961). Error bars represent maximum analytical errors of 2‰ δD and 0.2‰ $\delta^{18}O$. Data for the Cold Springs is from Young (1985).

Table 5
Stable isotope results for thermal and non-thermal springs in the SFPGS, in per mil notation relative to VSMOW. Data for Cold Springs 1 and 2 from Young (1985), listed as springs 9N-9E-32BAD1s and 10N-6E-19CDC1s, respectively

	$\delta^{18}O$	δD	Discharge elevation (ft)
Deer Creek H.S.	−17.5	−142.0	3200
Goller H.S.	−18.5	−138.4	3200
Danskin Creek H.S.	−17.7	−141.1	3280
Pine Flats H.S.	−18.8	−145.5	3600
Kirkham H.S.	−18.7	−145.8	4000
Bonneville H.S.	−19.4	−147.9	4700
Sacajawea H.S.	−19.3	−148.4	5000
Cold Spring 1	−17.7	−131.0	4000
Cold Spring 2	−16.9	−124.0	3800

4.4. Trace elements

Thermal springs in the SFPGS show marked uniformity in trace element concentrations, with the exception of Deer Creek Hot Springs, which has much higher concentrations of Fe and Ba (Table 6, Fig. 8). Differences in Rb/Sr ratios appear to provide a criterion that may be used to distinguish hydrothermal fluids equilibrated with Cretaceous granitic rocks from those equilibrated with Tertiary granitic rocks of the Idaho batholith (Druschel, 1998). Rb/Sr ratios of hydrothermal fluids may be affected by fractionation into a precipitating solid phase or by adsorption reactions (Michard, 1990). However, thermal waters in the Idaho batholith appear to reflect approximate Rb/Sr ratios of parent rocks, suggesting that these elements are not selectively fractionated into any of the minerals present. The Rb/Sr ratios of thermal waters in the SFPGS (Table 6, Fig. 8) are apparently not in equilibrium with Tertiary granitic rocks, inasmuch as these ratios do not approach the value expected in waters reacting with Tertiary rocks (Rb/Sr ≈ 1; Druschel, 1998).

Trace element patterns in the thermal waters are similar to those observed for local granites and gran-

Table 6
Trace element concentrations in ppb for thermal springs in the SFPGS

	Li	B	Mg	Al	Mn	Fe	Co	Cu	Zn	Rb	Sr	Mo	Sb	Ba	Pb	U	As
Deer Creek H.S.	131	240	22	53	6.5	4927	0.3	2.4	6.3	26.2	226	13	1.4	136.5	0.8	0.3	7.9
Goller H.S.	108	110	22	76	7.1	125	0.2	0.8	2.0	7.1	82	6.7	0.5	18.7	0.5	0.3	7.1
Danskin Creek H.S.	43	57	58	65	1.9	105	0.9	2.4	4.2	1.7	40	17	0.2	2.1	0.3	0.2	4.3
Pine Flats H.S.	86	80	14	98	6.7	111	0.2	4.3	6.5	7.0	95	20	0.4	3.2	0.4	0.3	8.3
Kirkham H.S.	71	50	11	59	1.1	95	0.1	2.0	1.5	7.6	99	50	0.3	3.1	0.3	0.1	2.2
Bonneville H.S.	116	70	9	93	1.1	89	0.1	0.7	1.6	24.8	163	44	0.2	2.0	0.3	0.1	1.1
Sacajawea H.S.	91	90	9	59	0.9	94	0.1	0.5	1.0	19.4	240	60	0.2	1.5	0.2	0.7	1.4

odiorites (Fig. 8) suggesting that the thermal waters of the SFPGS are equilibrating with granites at depth. This similarity is surprising because all other evidence supports the hypothesis that this system is both mineralogically well developed and geochemically dynamic. Changes in mineralogy can create significant differences in selective adsorption behavior, which may have a notable effect on selected trace element ratios (Stumm and Morgan, 1996).

4.5. Hydraulic communication

Thermal spring waters that are part of a common aquifer should have common chemical signatures. Chemical and physical parameters are rarely, if ever, homogeneous, and differences in water chemistry may be expected in a system due to changes in rock-type, temperature, and recharge water compositions along a flowpath. Distinguishing the chemical characteristics of springs due to heterogeneity in a single system from differences due to those of separate systems requires a detailed investigation of petrologic and structural relations as well as water / rock equilibria. Stable isotope evidence bearing on the hydraulic connection between springs has been discussed previously (see Section 4.3).

Thermal springs in the study area are spatially associated with, and possibly linked by, the fracture

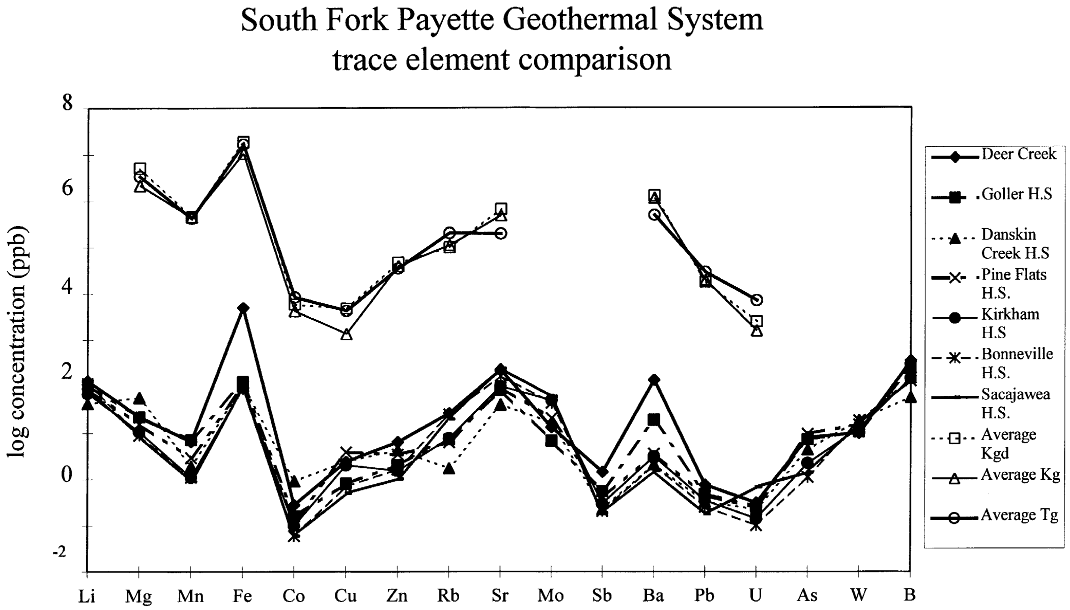

Fig. 8. Trace elements concentrations of the thermal springs in the SFPGS and average values of common rock types in the area (Clarke, 1990).

network defining the course of the South Fork of the Payette River (Reed, 1986; Dingee, 1987). Examination of solution compositions and stable isotope ratios along this possible flowpath might be expected to show variations in the chemistry of aligned springs which could be related to water/rock equilibria. Figs. 8 and 9 show selected major element and trace element chemistry, respectively, for each of the springs along an east-west transect in the study area.

The major element chemistry of Deer Creek Hot Springs is different than those of other springs in the SFPGS. Compositional differences cannot be correlated with temperature, suggesting that this spring is hydraulically isolated and may be equilibrating with different rocks or source waters. Chloride is the most conservative element in geothermal waters, and, thus, is often used as an important diagnostic solute (Nicholson, 1993). Chloride concentration in Deer Creek Hot Springs is about 4 times higher than in any other spring, and its F^- concentration is the lowest of any of the thermal waters in the SFPGS (Table 1). The large difference in Cl^- between Deer Creek Hot Springs and the other thermal springs of the SFPGS suggests little or no hydraulic communication. The slightly lower fluoride concentration at Deer Creek Hot Springs may be due to a lower abundance of micas in the rocks equilibrating with this thermal water, inasmuch as fluorine concentration is often affected by the presence of micas in which it occurs as a trace constituent (Nicholson, 1993). The remainder of the thermal springs in the SFPGS have similar major element concentrations, suggesting that these waters are either in hydraulic communication or equilibrating with similar rock-types and source waters.

Small differences in the Ca^{2+}, Na^+, K^+ and SiO_2 contents (Fig. 9) of the other thermal spring waters may be due to temperature-dependent clay re-equilibration reactions. Inverse reaction modeling with the computer code NETPATH 2.0 indicates that chemical differences between springs can be modeled by a series of clay re-equilibrations. The modeling exercises identified reactions between different clays (Na–, Ca–, K–beidellite, illite, kaolinite), chlorite, serpentine and primary granitic minerals (quartz, plagioclase, mica) (Druschel, 1998).

Cretaceous granitic rocks in the study area exhibit a marked degree of heterogeneity (Reed, 1987; Dingee, 1987; Fisher et al., 1992) and, therefore, it is probable that waters along an extended flowpath are in contact with more than one type of rock. However, forward reaction modeling with the geochemical modeling code React (Bethke, 1998) indicates that very small differences are to be expected due to

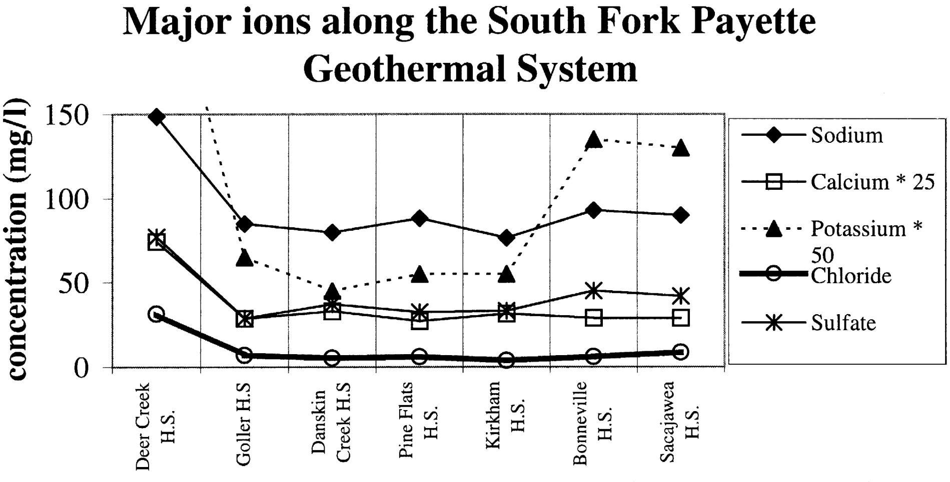

Fig. 9. Comparison of selected major ions in thermal spring waters along a west to east (left to right on the diagram) transect of the SFPGS. Concentrations of selected elements multiplied by $*x$.

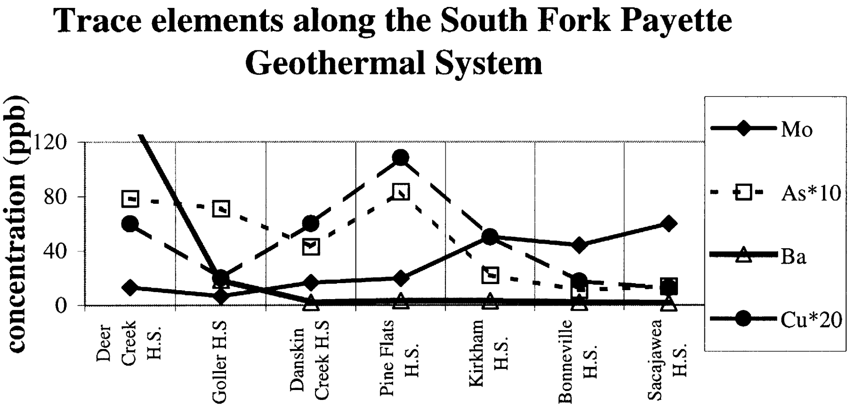

Fig. 10. Differences in expected major ion fluid compositions expected between water reacting with either a local Cretaceous granite or Tertiary granodiorite calculated using the geochemical modeling code React (Bethke, 1998).

reaction with the different types of granitic rock in the area (Fig. 10). Therefore, most of the differences in the chemical composition of hot spring waters are probably due to temperature-dependent clay re-equilibration reactions and/or the existence of isolated reservoirs with different source waters.

Inasmuch as the Mg^{2+} concentrations of thermal and non-thermal waters in the Idaho batholith are significantly different, Mg^{2+} concentrations may be used as an indicator of the extent of mixing (Druschel, 1998). Non-thermal waters in the Idaho batholith typically have Mg^{2+} values ranging from 2 to 20 mg/l (Parliman, 1986), whereas SFPGS thermal waters have values between 9 and 58 ppb. Danskin Creek Hot Spring thermal water has a higher Mg^{2+} concentration than the other thermal springs in the area (58 ppb, Table 6). An increase in the Mg^{2+} concentration of thermal waters from 10 to 58 ppb Mg^{2+} would require the mixing only 3% of local meteoric water with 2 ppm Mg^{2+}. Thus, it is reason-

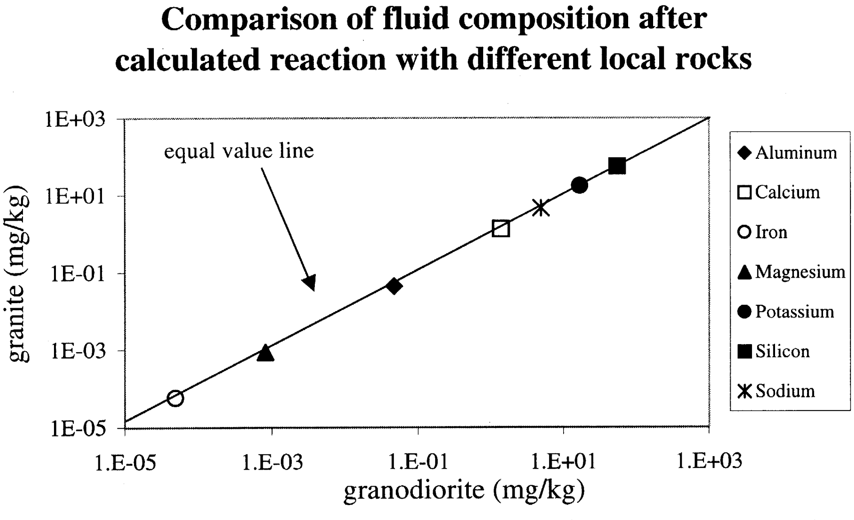

Fig. 11. Comparison of trace element in thermal spring waters along a west to east (left to right on the diagram) transect of the SFPGS. Concentrations of selected elements multiplied by $*x$.

able to assume that mixing of non-thermal waters with thermal waters at Danskin Creek Hot Springs or any of the other springs in this system is minimal and has had a negligible effect on water chemistry.

There is a much greater variation in the trace element-contents of the springs in the SFPGS (Fig. 11) than is observed for the major elements (Fig. 9). Molybdenum concentrations increase from east to west; Sacajawea Hot Springs exhibits the highest Molybdenum concentration of any of the springs in the SFPGS (60 ppb). Bennett and Knowles (1985) note Molybdenum concentrations in stream sediment samples from Tertiary granitic terrains that are much higher than in stream sediment samples from Cretaceous granitic terrains. Krahmer (1994, 1995) reported a correlation of Mo concentration with the proximity of springs to the Sawtooth Batholith and its associated alteration halo. However, it may be impossible to quantify changes in trace element concentrations without consideration of the effects of sorption and changes in sorption behavior due to temperature-dependent mineralogical changes.

Deer Creek Hot Springs shows the same extent of variation in trace element and major element chemistry, supporting the hypothesis that the spring is hydraulically isolated and/or equilibrating with a different rock-type or mineral assemblage.

4.6. Hydrology

The connection of fracture networks along the South Fork of the Payette River and the intersection of other brittle structures with these networks establishes a framework for hydrothermal water flow. Flow in unfractured crystalline rocks is assumed to be negligible due to permeability differences in comparison to fractured rock (Kassoy and Zebib, 1978; Freeze and Cherry, 1979). Flow of geothermal waters in the SFPGS can be separated into three physically distinct areas: recharge, lateral flow, and upflow zones. Waters are contained in the geothermal system for more than 9000 years (Young, 1985) and probably spend most of that time in lower permeability downward and lateral flowpaths at depths exceeding 2.5 km. Fluids equilibrate with the largely granitic rocks in the study area, and flow is concentrated in fracture networks which may extend over large distances (≥ 150 km) at great depths (up to 6 km). Dikes may be important in the horizontal flow system over a distance and in the model for upflow zone permeability.

Recharge of thermal springs in the SFPGS occurs along fractured conduits in high elevation areas peripheral to the South Fork Payette River fracture network (Reed, 1986; Dingee, 1987). Values of δD compared to those of current meteoric water (Fig. 7) suggest that the recharge water is probably from a much higher elevation. Water recharging along a fracture at different elevations will reach different depths, a function of overpressurized head, bulk permeability, and fluid flux (Forster and Smith, 1988b). Distribution of water at depth in a vertical fracture network could be envisioned as (1) a confined fluid flow conduit similar to a pipe system, or (2) a very tall and thin sheet of moving water across the entire depth and width of the structure.

Physical evidence of isolated, channelized fluid flow in deep granitic systems at the German Continental borehole (KTB) and the International Stripa Project (SKB) suggests that fluid flow occurs through a number of isolated fracture conduits at different depths (Olsson and Gale, 1995; Durham, 1997). It is reasonable to assume that the fracture-controlled hydrothermal fluid conduits in the SFPGS may share similar characteristics. This interpretation of the structure of deep thermal water flow and the lack of significant mixing with shallow meteoric water suggests separate high-elevation thermal waters and lower elevation local groundwater systems. Forster and Smith (1988a,b) suggest that the topography of mountainous terrains can have a significant impact on recharge and groundwater flow; it is possible that this causes the separation of thermal and non-thermal flow systems in the Idaho batholith. High elevation recharge at flow divides will drive water deeper into a system (Forster and Smith, 1988b), with downward movement eventually halted by changes in hydraulic head caused by heating and corresponding changes in density (Kassoy and Zebib, 1978).

Upflow and discharge of thermal waters occurs along fracture networks that have much greater permeability than the rest of the hydrothermal system. The thermal cooling of waters from their equilibration temperature to discharge temperature occurs faster than the equilibration of that fluid with respect to phases that are saturated at equilibrium temperature but supersaturated at discharge temperature. This behavior determines the accuracy of geothermometers and suggests that water comes up from depth on a time scale of hours or days (Reed, 1997). Further consideration of the kinetics of secondary mineral precipitation and conductive cooling may be used to better constrain the timing of the fluid ascent. The zone of greater permeability may be due to intersecting brittle structures (Dingee, 1987; Krahmer, 1995; Zimmer, 1996), or the greater joint density observed in dikes (Reed, 1986). Upflow may also occur when fluids intersect a relatively impermeable dike; hydraulic overpressure at that point will drive fluids to the surface (Kuhns, 1980).

The permeability and orientation of dike rocks in shear zones determine their ability to act as physical barriers to fluid flow. Reed (1986) measured joint density in dikes and surrounding plutonic rocks in the SFPGS, and found the Cretaceous granodiorite to be generally lower in fracture density than adjacent fine-grained dike rocks. This suggests that the dikes are not impermeable, and may act as fluid conduits. If so, then it is reasonable to expect fluids to discharge from dike rocks, but that was not observed in any of the 40 + springs examined in the Idaho batholith by Druschel (1998). Although joint density measured by Reed (1986) is greater in dike rocks at the surface, the geometry of the fractures (predominantly oriented NW–SE) and the nature of the material in the fractures may render them relatively impermeable; thus, dikes may serve to inhibit the flow of thermal waters (Fetter, 1980; Reed, 1986). Dikes in the Idaho batholith generally have a NE–SW to N–S orientation (Bennett and Knowles, 1985). They are 0.3–20 m wide and are traceable up to 1 km along the strike (Reed, 1986). Dikes that are located in north or northeast trending shear zones often do not fill the entire width of the shear zone, therefore, may not serve as barriers to flow, regardless of the joint density. On the other hand, relatively impermeable dikes which cross NW-trending shear zones are perpendicular to these shear zones and truncate their width. These dikes may serve as barriers to hydrothermal water flow, physically separating the system into smaller subsystems.

A model incorporating the geology, structures, seismicity, hydrology, and geochemistry of the SFPGS is illustrated in Fig. 12. The main fracture system, which defines both the course of the South Fork of the Payette River and the topographic lows

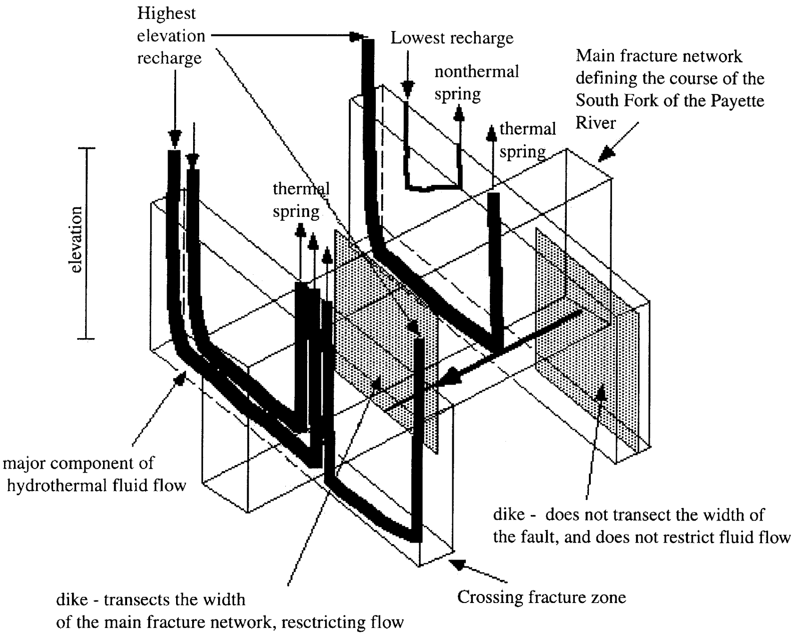

Fig. 12. Hydrothermal flow model for the SFPGS.

where thermal springs generally discharge, is the central component of the model. Recharge occurs at high elevations along the intersecting brittle structures, consistent with models of deep, fracture-controlled systems in mountainous terrains (Kassoy and Zebib, 1978; Forster and Smith, 1988a,b), and with observed stable isotope ratios. Several brittle structures intersect the main fracture system, and the highly fractured and permeable zone of intersection forms the primary conduit through which thermal waters rise quickly to the surface from depth. Deep hydrothermal fluid flow comes to the surface over a matter of hours or days with minimal mixing between hydrothermal waters and shallow groundwaters. Dikes may serve as barriers to hydrothermal fluid flow depending on their orientation and relative permeability. A combination of the hydrodynamics involved with deep, fracture-controlled fluid flow and local geological barriers creates a system which is divided into several smaller convection cells, with little or no hydraulic communication between them.

5. Discussion and conclusions

5.1. SFPGS

The SFPGS hot spring waters equilibrate with an alteration assemblage of granitic origin, and flow primarily in small, separate convection cells located within fracture networks peripheral to the river. Source areas for recharge water are at high elevations, as suggested by hydraulic considerations of thermal water flow and stable isotope data. Thermal waters infiltrate fractures to a depth of 2.4–6.3 km, and flow laterally at depth; residence times are over 9000 years.

Physical evidence in similar, deep, granitic terrains at the German Deep Continental borehole (KTB) and the International Stripa Project indicates that fluid flow in fractures is primarily contained in isolated channels (Nordstrom et al., 1989; Olsson and Gale, 1995; Durham, 1997). Thermal waters come to the surface in a very short time (hours to days); upflow is initiated by the intersection of a zone of crossing structures and high permeability or a relatively impermeable dike. Maintenance of the water flowpaths is dependent on periodic seismicity, which is required over time due to volumetric changes resulting from the precipitation of clay minerals. The periodic disturbance of these hydrothermal systems by seismic activity is reflected by varying $\delta^{18}O$ shifts at different springs in the system

SOLVEQ modeling indicates an apparent equilibrium assemblage of quartz, albite, microcline, clay minerals, chlorites, serpentinites, micas, and at higher temperatures, zoisites. These results are consistent with experimentally determined mineral assemblages and thermal water compositions from the reaction of a granitic material with water at elevated temperatures. The equilibration with granitic rocks is further supported by similarities in trace element patterns between local granitic rocks and thermal waters. Deer Creek Hot Springs exhibits marked differences in pH, Cl^- and alteration assemblage minerals, indicating it is equilibrating with a different rock type and/or with a source water of different composition

5.2. Comparison with models for high-level nuclear waste repositories

The hydrothermal system located in the SFPGS may be used as a natural laboratory for the study of long-term behavior of an artificial hydrothermal system such as those that will be initiated by the heat released due to the emplacement of spent nuclear fuel. Hydrothermal systems in this area have probably been active for many million years and, thus, reflect the long-term development and behavior of a potential post-closure artificial hydrothermal system. Furthermore, residence times on the order of 9000 + years serve as a window on the possible cumulative effects of extended water/rock interactions and seismic disturbances over extended time periods.

There are several common characteristics of the hydrothermal system in the SFPGS and potential deep geologic repositories being considered worldwide. The hydrothermal system in the SFPGS is hosted by granitic rock and fluid flow is completely fracture-controlled. In most of the potential sites worldwide, the preferred host rock is crystalline, and, therefore, the fluid transport pathway of greatest concern would be fracture-controlled. The reservoir temperatures of the SFPGS are between approximately 80°C and 160°C; similar temperatures have been modeled by Buscheck and Nitao (1993) for the thermal response of a system to the heat generated by spent nuclear fuel. SFPGS hydrothermal waters are also heated at depth by a radioactive heat source, and a heterogeneous heat flow field also influences the hydrology of the system.

The SFPGS could serve as a model to validate forward reaction modeling codes, by comparison of the output of these models with calculated equilibrium assemblages of the waters and by physical comparison of calculated reaction products with actual samples from depth. However, use of the later criteria will require deep drillholes into active hydrothermal systems, which do not currently exist in the Idaho batholith.

Deep fracture-controlled lateral flowpaths in the Idaho batholith have stayed open for millions of years, contrary to what one may expect as a function of positive molar volume changes due to granitic alteration reactions, because of seismic activity. Understanding the transport of potentially hazardous nuclear products in water as dissolved ions or as particulate material is also very important in properly predicting failure scenarios. In a study of Idaho batholith thermal waters near the SFPGS, van Middlesworth and Wood (1998) observed that rare earth element concentrations in unfiltered samples vary much more with time than do filtered samples. This suggests significant particulate transport of rare earth elements, which were investigated because they are potential products in a nuclear waste repository (van Middlesworth and Wood, 1998). The temporal variations in particulate transport over a period of a few years further suggests that the magnitude of events which perturb these systems and maintain fast fluid pathways may be microseismic. The magnitude of seismic events that maintain fluid conduits over extended periods of time is an important consideration in assessing the long-term behavior of an artificial hydrothermal system.

Acknowledgements

Trace element, stable isotope, and silica analyses were carried at the Washington State University Geoanalytical Laboratories. Flame AA and ion chromatography were carried out at the Washington State University Soils Laboratory and the University of Idaho, respectively. Peter Larson, C. Kent Keller, Scott Wood, Regan Patton, Charles Knaack, and John Watkinson provided a wealth of directed discussion, technical assistance, and access to field and laboratory equipment. Thanks to Mark Reed for providing copies of SOLVEQ and CHILLER, and for assistance in their use. GKD is very grateful to Martin Schoonen at the State University of New York at Stony Brook for support, discussion, and use of the Geochemical Workbench 3.0 programs during my fruitful stay at Stony Brook. E. Oelkers and an anonymous reviewer are thanked for their helpful and insightful comments and suggestions.

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