Controls on the distribution of thermal springs in the southern Canadian Cordillera[1]

Stephen E. Grasby and Ian Hutcheon

Abstract: Thermal springs examined in southern British Columbia are restricted to six major Eocene or later brittle fault systems. These faults provide a high-permeability flow path that allows for deep circulation of meteoric water. The depth of circulation, and thus temperature, is largely influenced by fault plane geometry. Calculated circulation depths of up to 4.8 km are consistent with models for hydrothermal ore deposits that invoke the mixing of mineralizing fluids with sulphur-rich meteoric water at depth. Mass-flux calculations indicate that a relatively small spring (5 L/s) can transport large amounts of bacterially reduced sulphur to ore-forming depths over a short geological time scale and can easily account for the sulphur associated with Pb–Zn deposits in the southern Canadian Cordillera. Modern thermal springs may be good analogies for the upper-crustal flow systems in hydrothermal ore deposit models.

Résumé :

Les sources thermales étudiées au sud de la Colombie-Britannique sont limitées à six systèmes de failles cassantes datant de l'Éocène ou plus tard. Ces failles fournissent un chemin d'écoulement à haute perméabilité qui permet aux eaux météoriques de circuler en profondeur. La profondeur de la circulation et donc la température sont grandement influencées par la géométrie du plan de faille. Des profondeurs de circulation calculées jusqu'à 4,8 km concordent avec les modèles de gisements de minerai hydrothermaux qui ont recours au mélange en profondeur de fluides minéralisateurs et d'eaux météoriques riches en sulfures. Des calculs de débits massiques indiquent qu'une source relativement petite (5 L/s) peut transporter de grandes quantités de sulfure réduit par les bactéries à des profondeurs de formation de gisements de minerai sur une courte échelle de temps géologique et peut facilement tenir compte des sulfures associés au gisements de minerai de Pb–Zn dans le sud de la Cordillère. En ce qui concerne les modèles de gisements de minerai hydrothermaux, les sources thermales modernes peuvent servir de bonnes analogies pour les systèmes d'écoulement dans la partie supérieure de la croûte.

[Traduit par la Rédaction]

Introduction

The crustal-scale movement of fluids is an important factor in heat transfer and mass transport, which leads to diagenesis, rock alteration, and the formation of ore deposits. In the southern Canadian Cordillera, the deep circulation of meteoric waters played a significant role in the precipitation of hydrothermal Ag–Pb–Zn deposits and vein-forming minerals (e.g., Nesbitt et al. 1989; Beaudoin et al. 1992*a*; Nesbitt and Muehlenbachs 1995). Understanding the factors that control the deep circulation of meteoric waters may assist in understanding how fluids are focussed on a regional scale, and this in turn may aid in mineral exploration.

The most obvious modern expression of deeply circulating fluids is the occurrence of thermal springs. Typical of many areas, thermal springs in the southern Canadian Cordillera tend to occur along faults. However, why thermal springs occur along some faults and not others, and why some springs have higher temperatures than others, particularly in areas where magmatic activity is minimal or nonexistent, is not certain. By examining the modern-day distribution of thermal springs, it is possible to understand better the controls on their occurrence, and thus the factors that control the deep circulation of meteoric water. This paper examines active thermal springs in the southern Canadian Cordillera (Fig. 1*a*). We focus on the area between 49° and 51°N due to the relatively well constrained regional geology (e.g., Gabrielse and Yorath 1991; Cook et al. 1992; and references therein) and the large number of springs. Nearly one-third of the over 130 thermal springs documented in Canada occur in this region (Fairbank and Faulkner 1992; Woodsworth 1998, and unpublished locations).

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S.E Grasby.[2] Geological Survey of Canada, 3303 – 33rd St. N.W., Calgary, AB T2L 2A7, Canada.
I. Hutcheon. Department of Geology and Geophysics, The University of Calgary, Calgary, AB, T2N 1N4 Canada

[1]Geological Survey of Canada Contribution 2000070.
[2]Corresponding author: (e-mail: sgrasby@gcs.nrcan.gc.ca).

Geological setting

The Canadian Cordillera developed in response to the collision from Jurassic to Tertiary of island-arc terranes against the western margin of North America (Coney et al. 1980; Monger et al. 1982; Parrish et al. 1988; Gabrielse 1985; Gabrielse and Yorath 1991). The Cordillera is divided into five morphogeological belts (Fig. 1). These can be roughly defined as deformed sedimentary strata of either North American (Foreland belt) or island-arc affinity (Insular and

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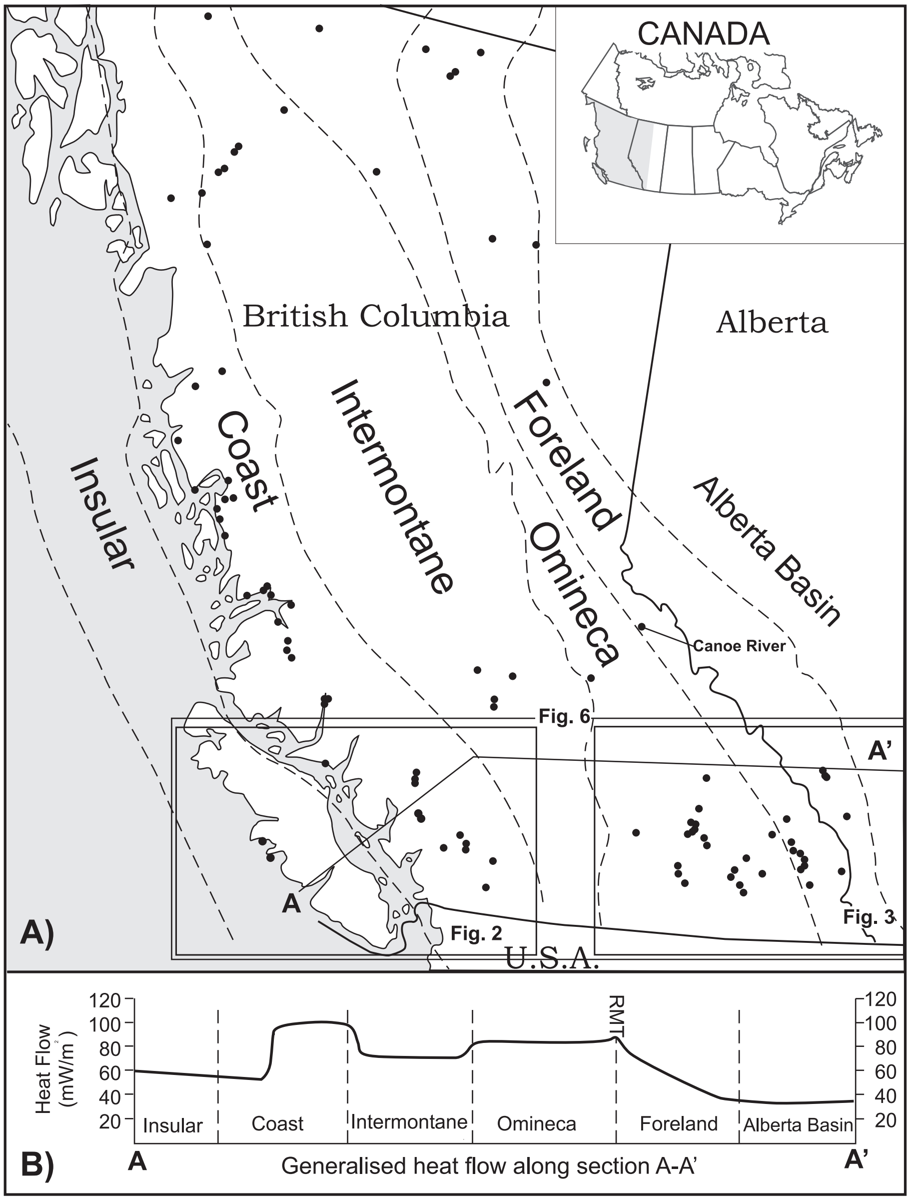
Fig. 1. (*a*) Regional map of western Canada showing locations of known thermal springs of the Canadian Cordillera. Springs occur in all five geomorphological belts. (*b*) Generalized heat-flow profile across the southern Canadian Cordillera along transect A–A′. Heat-flow data from Hydman and Lewis (1995), Majorowicz et al. (1984), and Bachu and Burwash (1994). Note that springs occur in both high and low heat-flow settings.



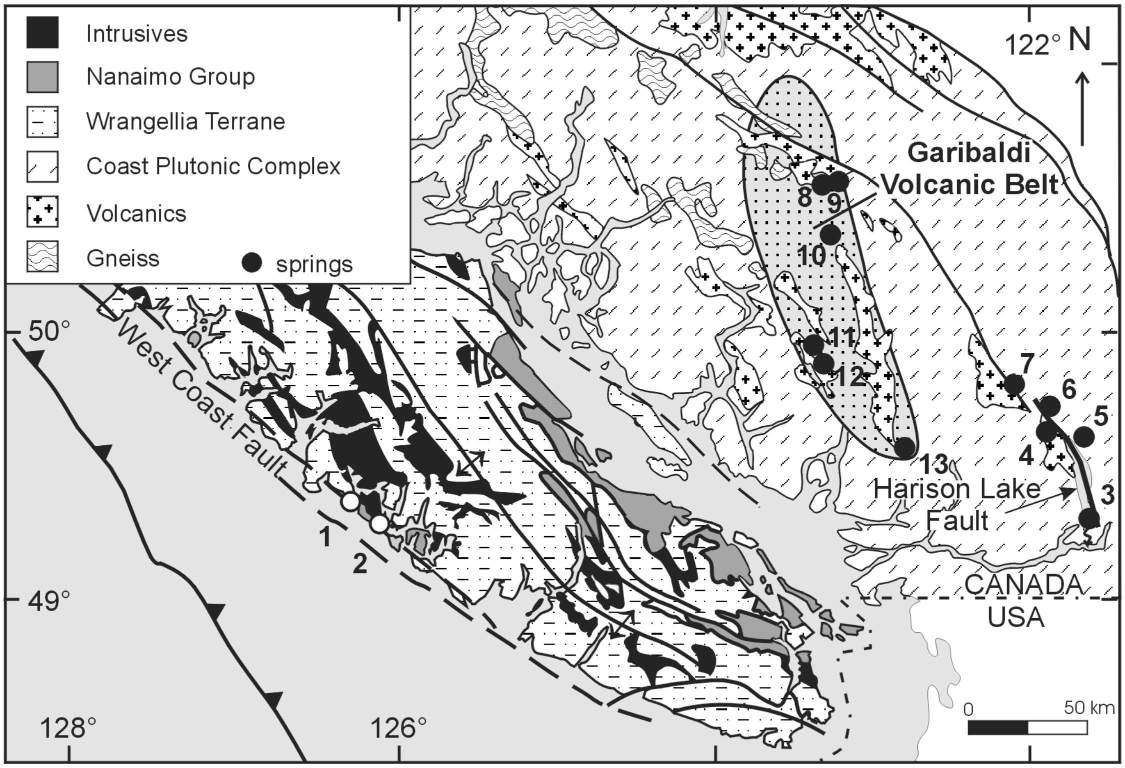
Intermontane belts) that are separated by belts of plutonic and high-grade metamorphic rocks (Coast and Omineca belts). Compressional deformation ended abruptly in the southeastern Canadian Cordillera during the Late Paleocene. At this time, crustal-scale extensional faults formed with associated plutonism and volcanism (Armstrong 1988; Parrish et al. 1988; Gabrielse and Yorath 1991). From Eocene to Recent, the southwestern Canadian Cordillera has been affected by right-lateral, strike-slip faulting (Gabrielse 1985). In addition, the Garibaldi Volcanic Belt (Fig. 2) developed from late Tertiary to Quaternary (Lewis and Souther 1978).

Description of springs

Before discussing the distribution, it is necessary to first clarify the definition of a thermal spring. White (1957) defines a spring as thermal if the temperature is significantly higher than the mean annual air temperature of the region in which the spring occurs, where 5°C is considered a significant difference. This definition is inadequate for the Canadian Cordillera, given that mean annual air temperatures are close to 0°C (Canadian Climatic Norms), and average groundwater temperature is typically around 5°C (i.e., all springs in the Canadian Cordillera would be thermal using White's (1957) definition). Using 10°C as a cut-off allows for over 130 thermal springs in the Canadian Cordillera (Fairbank and Faulkner 1992; Woodsworth 1998; this work, and unpublished locations). Of these, 48 occur in the study area (locations provided in Table 1). Two spring locations listed by Fairbank and Faulkner (1992) as "unconfirmed" are not included here.

Thermal springs in the southern Canadian Cordillera oc-

Fig. 2. Major geological features of southwestern British Columbia (after Gabrielse 1991). Locations of thermal springs (circles) are shown; numbers refer to Table 1. Thermal springs are associated with major Eocene faults and (or) the Garibaldi Volcanic Belt.



cur as both isolated springs and closely spaced groups of springs. Locations where several springs occur in close proximity (<100 m) are considered a single hydrothermal system with the highest temperature outlet taken to be the most representative. Discharge is usually low (1 to 50 L/s); however, larger flow rates occur at Meager Creek (up to 500 L/s) (Fairbank and Faulkner 1992). Chemical and stable isotope data for a number of thermal springs in the study area are reported elsewhere (Grasby et al. 2000). Temperatures for the springs we sampled are provided in Table 1. For the remaining springs in the study area, temperature and locations are obtained from Fairbank and Faulkner (1992) and Woodsworth (1998) (Table 1). Discharge temperatures range widely, from the defined minimum of 10°C to a maximum of 82°C for the Dewar Creek spring (29 in Table 1). The average temperature for thermal springs in the study area is 39°C. Some springs have associated tufa mounds that can cover large areas (hundreds of square metres). Extensive growths of cyanobacteria, algae mats, and *thiothrix* (a white filimentous bacteria) are common.

Distribution of thermal springs

An initial step in this study was to locate all known thermal springs in the southern Canadian Cordillera, through literature research, reports from geological surveys and forestry offices, and "word of mouth." The locations of all verified springs are listed in Table 1. The next step was to examine the spatial distribution of thermal springs in relation to geological features. An important feature we document here is that, despite the complex deformation leading to the formation of countless faults across the southern Cordillera, almost all of the thermal springs in southern British Columbia are restricted to only six major fault systems (Figs. 2, 3). From west to east they are West Coast, Harrison Lake, Okanagan Valley, Columbia River, Purcell Trench, and Rocky Mountain Trench faults. These faults are associated with some of the major river valleys in the southern Canadian Cordillera (Columbia, Kootenay, Okanagan, Lillooet, and Harrison Lake valleys). These major valley systems form regional topographic lows and are all bordered by high mountain ranges, with up to 2500 m of relief. In southern Alberta (Figs. 3, 4), thermal springs are largely restricted to localized zones of complex deformation along major thrust faults in the front ranges of the Rocky Mountains at the latitude of Banff (Rundle, Sulphur Mountain, and Bourgeau thrusts) and further south on the Misty Thrust (Figs. 3, 5). The structures associated with the occurrence of thermal springs in the southern Canadian Cordillera are described below.

West Coast Fault

Figure 2 illustrates the major geological features in the southwestern Cordillera along with the distribution of thermal springs. The westernmost springs are Sharp Point and Ahousat on the west coast of Vancouver Island (1 and 2 in Table 1). Both springs are associated with the West Coast Fault (WCF) zone. The WCF is a crustal-scale feature characterized by a steeply dipping NW–SE-trending fault zone that locally exhibits a wide zone of crushed and shattered rocks (Muller et al. 1981). The fault zone consists of two or more parallel strands with a complex history of dip-slip and strike-slip motion. In addition to being associated with this complex fault zone, both springs are spatially related to the intersection of the WCF with the Tofino fault (Phillips 1994).

Table 1. Thermal springs in the southern Canadian Cordillera grouped by association with major geological features.

Spring	Lat.	Long.	T (°C)	max T (°C)	depth (km)
West Coast Fault					
1. Sharp Point[1]	49°21	126°16	58.5	95	4.5
2. Ahousat[1]	49°16	126°04	22	78	3.7
Harrison Lake Fault					
3. Harrison Lake.[1]	49°20	121°46	62.4	127	2.5
4. Sloquet[1]	49°45	122°20	60.8	116	2.3
5. Clear Creek[2]	49°41	121°45	43		
6. August Jacob's[2]	49°48	122°16	49		
7. Skookumchuck[1]	49°59	122°28	50	116	2.3
Garibaldi Volcanic Belt					
8. Meager Creek[1]	49°16	123°30	47	186	3.7
9. Peable Creek[2]	50°34	123°30	60		
10. Elaho River[2]	50°06	123°17	30		
11. Turbid Creek[2]	50°05	123°16	30		
12. Shovel Nose[2]	49°44	122°43	57		
13. Pitt River[2]					
Okanagan Valley Fault					
14. KLO[1]			22.7	137	4.7
Columbia River Fault					
15. Albert Canyon[1]	51°08	117°45	25.7	74	2.3
16. Wiskey Point[2]	50°30	117°55	?		
17. Halcyeon[1]	50°30	117°55	50.7	132	4.1
18. Halfway[1]	50°28	117°51	58.9	117	3.7
19. St.Leon[1]	50°26	117°51	46.5	123	3.8
20. Nakusp[1]	50°17	117°40	55.8	131	4.1
21. Wilson Lake[2]	50°13	117°36	30		
22. Taylor[2]	50°04	117°56	25		
23. Snowshoe[2]	49°55	118°11	?		
24. Jorden Ranch[2]	49°48	118°10	12		
25. Octopus Creek[2]	49°44	118°00	49		
Purcell Trench					
26. Fry Creek[2]	50°05	116°45	?		
27. Ainsworth[1]	49°44	116°54	45	163	4.8
28. Crawford Bay[2]	49°40	116°47	32		
29. Dewar Creek[1]	49°55	116°28	83	147	4.3
Rocky Mountain Trench					
30. Wolfenden[1]	50°50	116°16	27.7	35.6	1.7
31. Radium[1]	50°38	116°02	44	69	3.3
32. Fairmont[1]	50°20	115°54	46.7	62	3
33. Toby Creek[1]	50°25	116°19	11	102	4.9
34. Lussier[1]	50°08	115°35	43.2	67	3.2
35. Ram Creek[1]	50°02	115°35	36.5	42	2
36. Red Rock[2]	50°13	115°42	?		
37. Wildhorse[1]	49°48	115°30	31	58	2.8
Rocky Mountains					
38. Fording Mt.[1]	49°58	114°52	20.5	33	1.6
39. Mist Mt.[1]	50°33	114°54	33	48	2.3
40. Forty Mile	51°13	115°38	?		
41. Cave[1]	51°10	115°35	29.8	52	2.5
42. Basin[1]	51°10	115°35	31.8	59	2.8
43. Middle[1]	51°10	115°34	22	57	2.7
44. Kidney[1]	51°09	115°34	38.9		
45. Upper[1]	51°09	115°34	41.3	67	3.2
46.Vermillion[1]	51°11	115°38	22		
47. Canmore[1]	51°05	115°23	12		
48. Many[1]	51°04	115°07	11	13	0.6

Note: Outlet temperatures from [1]This study and [2]Fairbank and Faulkner (1992). Calculated maximun temperatures are based on chemical data (Grasby et al. 2000). Corresponding circulation depths are calculated based on local geothermal gradients (see text).

Harrison Lake Fault

A series of five springs occurs along the trend of the Harrison Lake Fault (HLF; Fig. 2). The HLF is a major dextral strike slip fault and has a complex history of ductile and brittle deformation (Monger and Journeay 1994; Monger and van Ulden 1998). This fault system connects further north to a series of faults that define a major valley system containing the Meager Creek springs. The Meager Creek springs are also spatially related to the Garibaldi Volcanic Belt (Fig. 2).

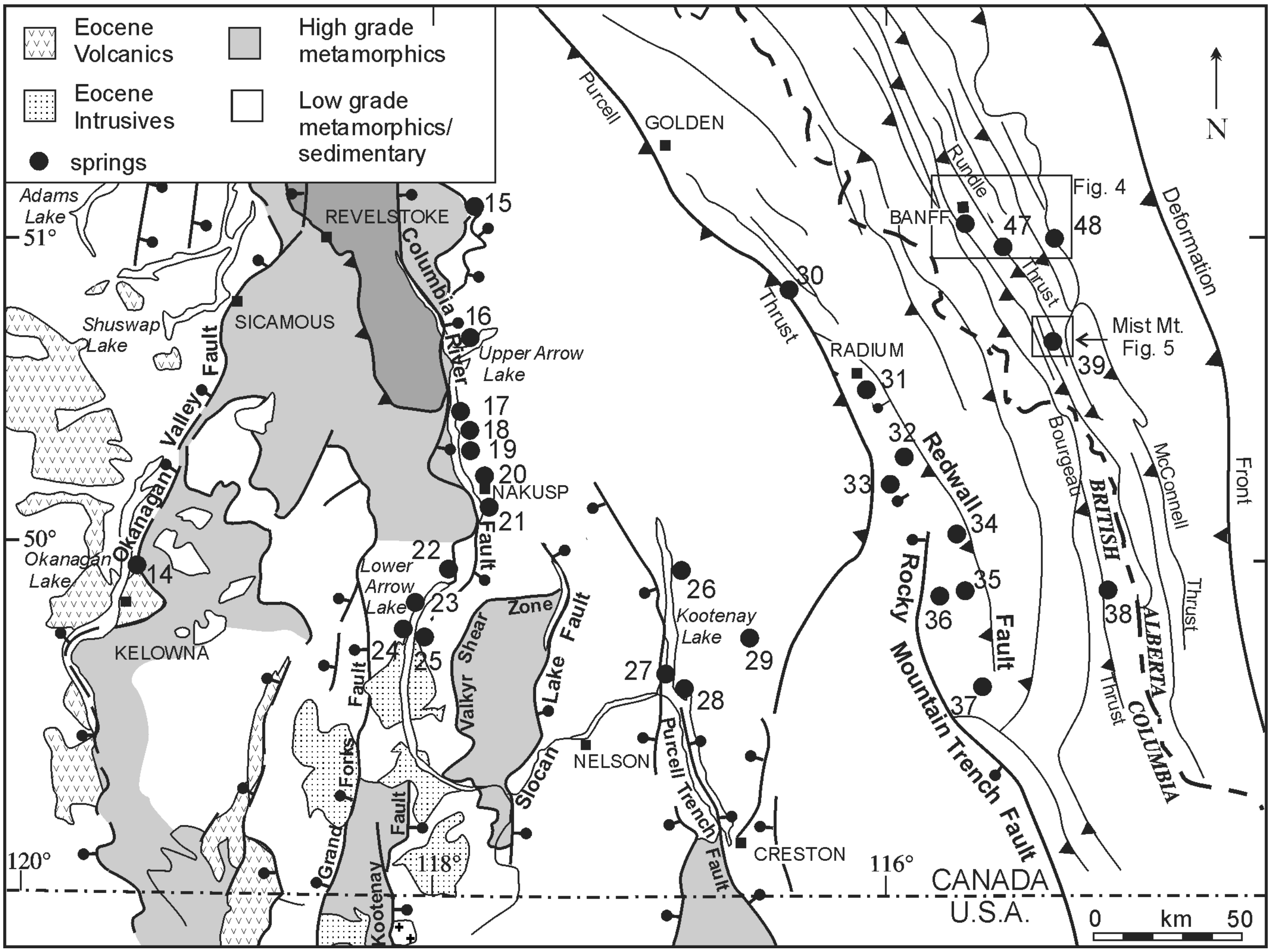
Okanagan Valley Fault

Figure 3 illustrates the major geological features and the distribution of thermal springs in the southeastern Cordillera. West of the Rocky Mountain Trench, the southeastern Cordillera was affected by a regional extension in the Eocene that exposed a core zone of high-grade metamorphic rocks (Parrish et al. 1985; Gabrielse 1991) (Fig. 3). Normal faults within this core zone are high angle and have displacements of hundreds of metres. In contrast, the bounding faults on the eastern and western margins tend to be low angle (20–30°), and have kilometres of displacement (Parrish et al. 1988; Parrish 1991; Cook et al. 1992). Within this regional extension complex, thermal springs occur only along the outer bounding faults. The Okanagan Valley Fault (OVF), a bounding fault on the western margin, is a 1 to 2 km thick shear zone characterized by mylonite and microbreccia. Only one known thermal spring (KLO, 14 in Table 1) is associated with the OVF.

Columbia River Fault

The Columbia River Fault (CRF; Fig. 3) occurs on the eastern margin of the regional extension complex discussed above. It is characterized by a mylonite zone up 1 km wide that has been intensely fractured and folded (Lane 1984). Eleven springs are associated with the CRF, along Upper Ar-

Fig. 3. Generalized geological map of southeastern British Columbia and southwestern Alberta, showing major extension faults, as well as main thrust faults in the Rocky Mountains (after Gabrielse 1991). Locations of thermal springs (circles) are shown; numbers refer to Table 1.



row Lake, as well as further south near the terminus of the fault, along Lower Arrow Lake. The Albert Canyon spring (15 in Table 1) occurs in the hanging-wall splay of the CRF in a zone of N–S-trending brittle normal and strike-slip faults (J. Crowley, personal communication, 1997).

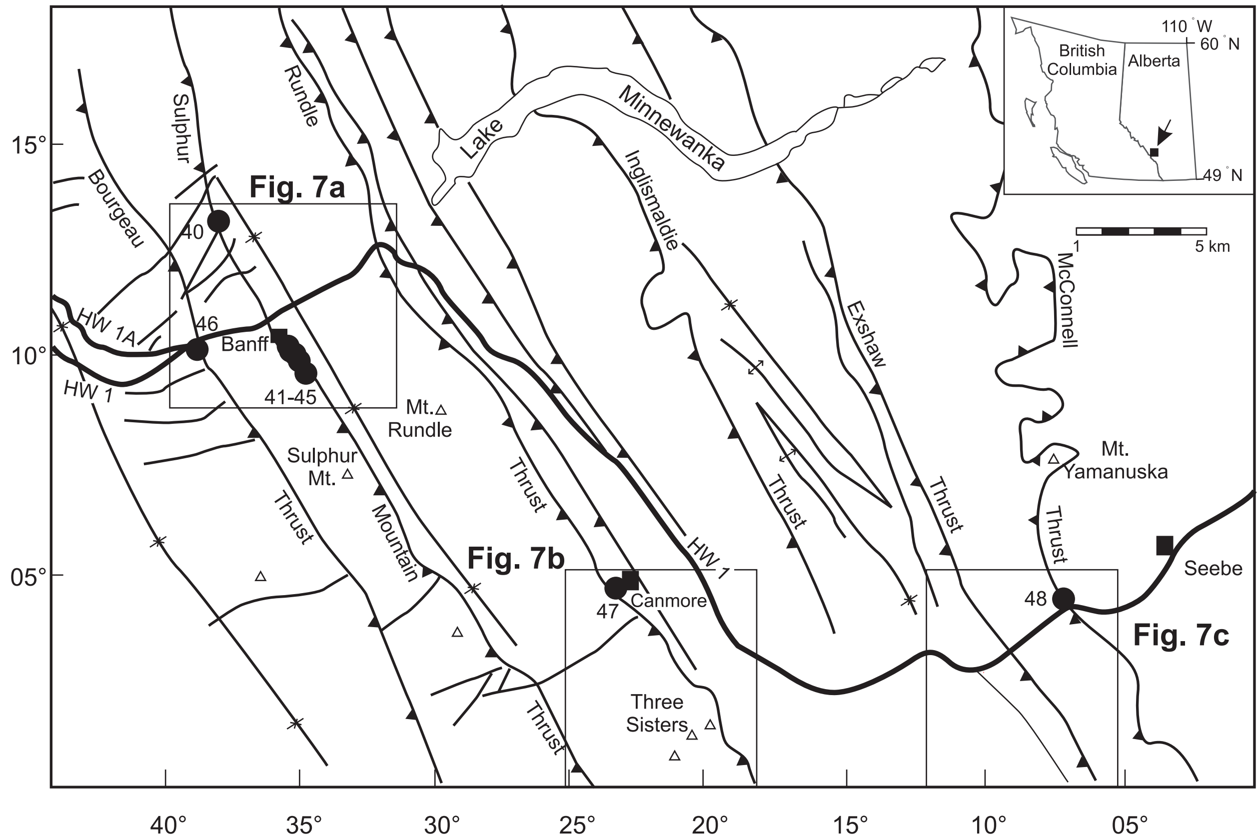
Purcell Trench Fault

At least three springs occur near the northern termination of the Purcell Trench Fault (PTF; Fig. 3). The PTF is a crustal-scale extension fault similar to the CRF discussed above. The trace of the PTF is not exposed, being largely covered by Kootenay Lake. However, cataclastic deformation is observed along what are considered subsidiary faults near the southern end of Kootenay Lake (Barrett 1983). The hottest recorded thermal spring in the southern Canadian Cordillera (Dewar Creek, 29 in Table 1) occurs east of the Purcell Trench. This spring does not appear to be directly related to a major structure; however, it occurs along a steeply dipping unnamed fault between the PTF and the Hall Lake Fault (another major crustal feature) (Cook et al. 1992; Reesor 1996).

Rocky Mountain Trench Fault zone

In the southern Rocky Mountain Trench (RMT, Fig. 3), several springs occur in a linear trend paralleling, and in between, the traces of the Southern RMT and Redwall faults. Some springs, and nonactive tufa deposits, occur within the RMT itself. The Southern RMT Fault zone is characterized by a series of Eocene and younger, en echelon, southwest-dipping normal faults (Price 1962; Bally, et al. 1966; Clague 1974), with over 10 km of displacement (van der Velden and Cook 1994). The Redwall Fault is not well described in the literature, however it is a significant structural boundary between the Porcupine Creek fan structure and the Purcell Anticlinorium (Foo 1979). The fault is described by Foo (1979) as an east-verging, back-rotated thrust fault. Leech (1954) indicates that it has a component of late-stage, near-vertical normal motion that is characterized by zones of brecciated carbonate. The springs that occur between the Redwall Fault and the RMT are related to small scale normal faults, thrust faults, and bedding contacts (van Everdingen 1972). However, the fact that the distribution of springs is restricted to the zone between the RMT and

Fig. 4. Major faults of the Bow Valley corridor (after Ollerenshaw 1975). Locations of thermal springs (circles) are shown. Numbers refer to Table 1.



Redwall faults suggests that these features are the primary controls on the occurrence of springs in this area.

Occurrences in the Rocky Mountains

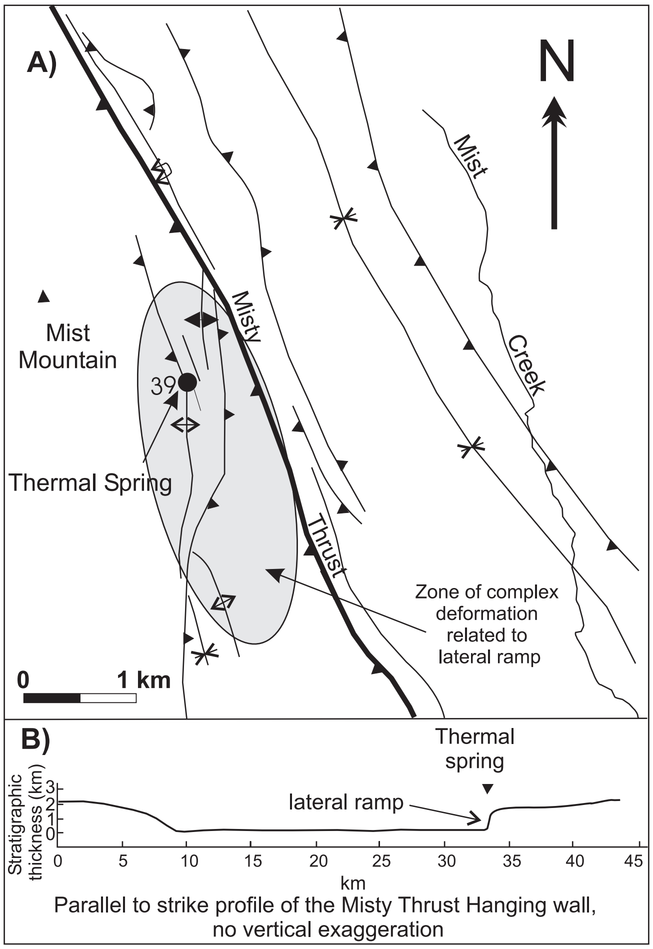
Six thermal spring systems occur in the southern Rocky Mountains (Foreland Belt, Fig. 1). The majority occur in the Bow Valley corridor, which forms a major topographic depression cutting approximately east–west through the southern Rockies. Springs occur along the major thrust faults in the Front Ranges at the latitude of Banff; from east to west they are the Rundle, Sulphur Mountain, and Bourgeau thrust faults (Fig. 4). For most of their length, these thrust sheets are relatively uncomplicated homoclinal SW-dipping panels. However, in the Banff area the Sulphur Mountain thrust is crosscut by a swarm of steeply dipping, northeast-trending dextral faults, with up to 200 m offset. Price (in press) indicates that these faults are related to along-strike stretching of the Sulphur Mountain thrust sheet in response to the development of a transverse monocline in the underlying Rundle Thrust sheet. Seven springs (40 to 46 in Table 1) occur in this area, along the Sulphur Mountain Thrust and a footwall splay, in an anomalous deformation zone (up to 200 m wide) that is characterized by complex faulted and folded strata. The Canmore spring (47 in Table 1) occurs on the Rundle Thrust near the point where it is cut by another vertical transverse fault (Fig. 4).

A similar situation is seen further south. Here the Mist Mountain spring (39 in Table 1), occurs at the southern end of the Misty Range (Fig. 5*a*) at over 2500 m a.s.l., 350 m above the valley floor and 300 m below the top of Mist Mountain. This is the highest occurrence of a thermal spring in Canada. The spring occurs along the Misty Thrust, a steeply dipping sub-planar 50 km long fault (Castonguay and Price 1995). Near the southern terminus of the thrust, there is a steep oblique hanging wall ramp that cuts abruptly upwards through 2 km of strata (Figs. 5*a*, 5*b*; Castonguay 1993, Castonguay and Price 1995). A zone of complex deformation that is characterized by oblique folds, faults, and back thrusts developed over the ramp. The Mist Mountain spring occurs within this zone of complex deformation, above the hanging-wall ramp (Fig. 5).

Controls on the distribution of thermal springs

Numerical models of deep fluid flow in mountain systems indicate that the bulk-rock permeability of the host rock and high-permeability faults, acting as conduits for preferential fluid flow, have the greatest influence on the development of thermal systems (e.g., Forster and Smith 1988*a*, 1988*b*; Lopez and Smith 1995, 1996); with infiltration rate, slope profile, and basal heat flux playing important, but lesser,

Fig. 5. (*a*) Geology of the southern Misty Range (after Castonguay and Price 1995), showing complex deformation associated with the lateral ramp in the Misty Thrust and the location of the Mist Mountain thermal spring. (*b*) Cross-section parallel to strike of the Misty Thrust, illustrating the lateral ramp in the hanging wall (after Castonguay and Price 1995). Location of this figure is shown in Fig. 3. The Mist Mountain springs is associated with the complex deformation along the lateral of the Misty Thrust.



roles. Smith and Forster (1990) and Lopez and Smith (1995) illustrate that there is a critical range of permeabilities (10^{-16} to 10^{-14} m^2) for the host rock that favors the development of thermal springs. At lower permeabilities, conductive heat flow will dominate, whereas at higher permeabilities the ground-water flux is sufficiently high to depress the local geothermal gradient. The range of permeabilities that favors the development of thermal springs falls within the range of permeabilities of fractured rock (Freeze and Cherry 1979), suggesting that deformation belts are conducive to the development of thermal systems. Measuring the bulk permeability of mountain massifs would be a near impossible task; however, the existence of thermal springs indicates that the bulk permeabilities are sufficient for their development. Slope profiles for the glaciated terrain of the Cordillera tend to be concave, and therefore do not show significant variability. Given this, we examine the role that the basal heat flux, infiltration rates, and fault zones play in the distribution of thermal springs.

Basal heat flux

Thermal springs in Canada tend to be restricted to the Cordillera. However, there have been reports of springs with temperatures greater than 10°C in the continental interior, along the eastern edge of the Western Canada Sedimentary Basin (Tyrell 1892; Cole 1915), related to discharge of saline waters from the Williston Basin (Bachu and Hitchon 1996). Upon inspection, springs with higher temperatures tend to have very low discharge rates and form shallow pools. As reported temperatures were recorded in the summer, it is likely that these waters were affected by solar warming. This is supported by the observation that nearby saline springs with higher discharge rates all have temperatures below 10°C. Given this and our cut-off of 10°C, the only thermal springs in Canada that we are aware of are in the Cordillera. Overall, the Cordillera has elevated heat flow compared to the Canadian Shield and sedimentary basins of the continental interior (Fig. 1B). This would suggest that the relatively high heat flow in the Canadian Cordillera may be a factor in the development of thermal springs. However, within the Cordillera, the importance of a heat source in controlling the distribution of thermal springs is less clear.

The thermal regime of the southern Canadian Cordillera is well established (e.g., Hyndman and Lewis 1995; Lewis 1991; Lewis et al. 1985; 1992; Davies and Lewis 1984). A generalized heat-flow profile (Fig. 1*b*) of the southern Cordillera is characterized by five heat flow regimes (Hyndman and Lewis 1995; Marjorowicz et al. 1984; Bachu and Burwash 1994). From east to west they are (1) low heat flow in the fore-arc region of Vancouver Island (Insular Belt), (2) very high heat flow associated with the Pliocene to Recent Garibaldi Volcanic Belt (within the Coast Belt), (3) moderate to high heat flow in the Intermontane and Omineca belts, (4) moderate to low heat flow across the Foreland belt, and (5) low heat flow in the Alberta Basin (Fig. 1). Within the study area, the only springs associated with anomalously high heat flow are those in the Garibaldi Volcanic Belt (Fig. 2).

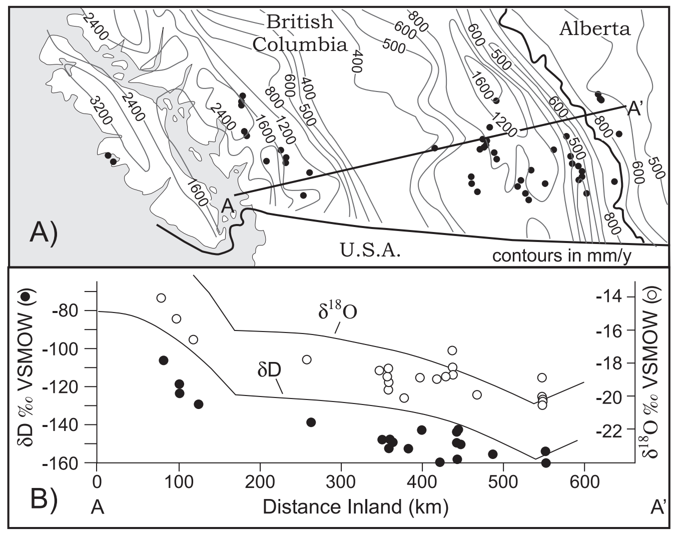
The Garibaldi Volcanic Belt consists of late Tertiary to Quaternary basaltic to rhyolitic volcanics (Lewis and Souther 1978) that form the northern extension of the Cascade Chain. Several thermal springs (8 to 13 in Table 1) are associated with anomalously high heat flux (up to 339 mW/m^2; Lewis et al. 1992) at the Meager Mountain and Mount Cayley volcanic centres (Fig. 2), which form part of the Garibaldi Volcanic Belt (Clark et al. 1982; Souther and Delleochaie 1984; Ghomshei and Clark 1993).

Elsewhere, the thermal spring distribution does not appear to be a function of heat flow. Although there tend to be more thermal springs in the higher heat flow regions of the Canadian Cordillera (Coast and Omineca belts), springs occur in all heat-flow provinces within the deformed belt, even in the low heat-flow settings of Vancouver Island and the Foreland Belt (Fig. 1*b*). Thus, the normal geothermal gradients of the Cordillera, even in low heat-flow settings, are sufficient for thermal springs to develop. Additional heat sources, such as

Table 2. Stable isotope data for thermal springs in the southern Cordillera.

Spring	Distance inland (km)	δD (‰)	δ^{18}O (‰)
3. Harrison	80	–106	–14.1
4. Sloquet Creek	100	–120	–15.3
7. Skookumchuck	120	–129	–16.6
8. Meager Creek	100	–123	–15.3
14. KLO	260	–136	–17.8
15. Albert Canyon	400	–142	–18.9
17. Halcyon	360	–145	–19.2
18. Halfway	360	–146	–18.4
19. St. Leon	360	–151	–19.7
20. Nakusp	360	–145	–18.9
27. Ainsworth	350	–145	–18.5
29. Dewar Creek	380	–149	–20.1
30. Wolfenden	425	–147	–19.1
31. Radium	430	–154	–18.8
32. Fairmont	430	–145	–18.8
33. Toby Creek	420	–160	–19.0
34. Lussier Canyon	440	–145	–18.3
35. Ram Creek	440	–141	–17.3
37. Wildhorse River	440	–138	–18.7
38. Fording Mt.	470	–154	–19.9
39. Mist Mountain	550	151	–20.2
41. Cave	550	–160	–20.5
42. Basin	550	–163	–20.2
43. Middle	550	–160	–20.1
45. Upper	550	–179	–20.3
46. Vermillian	550	–148	–18.9

Fig. 6. (*a*) Average annual precipitation, in mm, across the southern Canadian Cordillera (after Halstead 1991). Springs occur in both high- and low-precipitation areas, indicating that infiltration rates do not influence the development of thermal systems. (*b*) Trend line showing variation in δD of meteoric water (after Yonge et al. 1989) and δD of thermal springs (circles), plotted as distance inland from the Pacific coast along transects paralleling A–A′.



radiogenic heat production or magma, are not essential for the development of a geothermal system in the Cordillera.

Infiltration rate

The stable isotope compositions of waters from thermal springs across the Canadian Cordillera are consistent with a meteoric origin (Grasby et al. 2000, Phillips 1994, Clark et al. 1982). Our data (Table 2) indicate that the δD and δ^{18}O values of thermal springs in the southern Canadian Cordillera show a general eastward depletion in D and ^{18}O (Fig. 6) and vary consistently with the inland decrease in modern precipitation (Yonge et al. 1989). This suggests that, on a regional scale, recharge-zones are not significantly distant from the spring outlets. Given this, we can use mean annual precipitation as a proxy for available infiltration rates, as suggested by Forster and Smith (1988*b*). Contours of mean annual precipitation are shown along with the locations of thermal springs in Fig. 6. Thermal springs occur in both high and low precipitation regions of the southern Canadian Cordillera, suggesting that the occurrence of thermal springs is not constrained by a water source or, in turn, infiltration rate. However, there is evidence that thermal springs require a minimum amount of precipitation to maintain sufficient water tables, and thus hydraulic head, to drive the system.

Where records are available (mainly for springs developed for commercial use), thermal springs flow throughout the year. However, there are isolated incidents of springs intermittently running dry. In the winters of 1923 and 1998, the Upper Hot Spring (the topographically highest in a group of springs in Banff National Park) went dry for over two months (Warren 1927; D. Lepinski, personal communication, 1999). These flow stoppages followed the years with the first (1922) and fourth (1997) lowest recorded total annual precipitation on record (Environment Canada unpublished records for 1887 to 1999). As well, following the year with the second lowest recorded precipitation, 1970, there was an extreme reduction in flow. This suggests that in low rainfall years the water table drops below that necessary to maintain flow.

Fault zones

Brittle faults in the upper crust are known to have enhanced permeability that allows preferential fluid flow (e.g., Caine et al. 1996). Within the Canadian Cordillera, thermal springs are restricted to a small number of faults. West of the Rocky Mountains, most thermal springs occur along six major faults that share the common features of (1) penetrating deeply into the crust and (2) having a relatively recent (Eocene or later) component of brittle deformation. These major Eocene, and later, structures postdate peak metamorphism (Greenwood et al. 1991). In the horizontally extended metamorphic core complex of the southeastern Canadian Cordillera, we observe springs only along the deep-cutting and highly fractured outer bounding faults. The high-angle, low-displacement faults within the metamorphic core complex do not have any associated springs. The importance of deep-cutting brittle faults on the distribution of thermal springs is also illustrated in the Rocky Mountain Trench. The zone of thermal springs in the RMT ends just north of Radium (Fig. 3). This is coincident with the transition from the southern RMT, characterized by discrete normal faults with significant displacements (e.g., van der Velden and Cook 1994) to a wide zone of low-displacement normal faults in the Golden area (e.g., Kubli and Simony 1994; P. Simony, pers. com. 1998) (Fig. 3). Normal faulting in the RMT becomes more discrete and has more significant offset (over 2 km) further north, near Bush Arm on Kinbasket Lake (approximately 75 km NW of Golden) (e.g., Gal and Ghent 1990). North of Bush Arm another thermal spring occurs (Canoe River spring, Fig. 1), whereas there are no recorded springs between Bush Arm and Golden. This suggests that only the portions of the Rocky Mountain Trench that have normal faults with significant offsets can develop thermal systems.

The restriction of thermal springs to the major Eocene and younger extension faults is consistent with patterns of hydrothermal alteration observed across the Canadian Cordillera. Work by Magaritz and Taylor (1986) shows evidence for meteoric hydrothermal alteration of batholiths across southern British Columbia, with the most intense alteration localised along the major Eocene extension faults. Similarly, Nesbitt and Muehlenbachs (1995) show evidence for prominent depletions of $\delta^{18}O$ values in quartz and carbonate veins along the Columbia, Slocan, and Okanagan faults, as well as the Valhalla Shear Zone, and suggest “most vein-forming fluids originated as meteoric waters” (p. 1048). Beaudoin et al. (1992*b*) give mineralization ages for hydrothermal alteration along the Slocan Lake Fault of 44–59 Ma, indicating that the hydrothermal systems associated with these faults initiated during the early stages of deformation (47–59 Ma; Parrish et al. 1988).

Although there is geologic evidence for hydrothermal alteration in the past, not all of the major Eocene extension faults in the southern Canadian Cordillera have associated modern-day thermal springs (Fig. 3). Thermal springs are likely transient features. Thermal springs commonly develop large tufa deposits, suggesting that they may seal themselves off through time. We have observed extensive tufa mounds at numerous locations that have no associated spring discharge.

In contrast to southern British Columbia, thermal springs in the Front Ranges of the Rocky Mountains are associated with localized zones of complex deformation along major front range thrust faults. For springs in the Banff area and at Mist Mountain, transverse steep-dipping structures appear to have favored the localized development of thermal springs.

In summary, the primary control on the distribution of thermal springs in the southern Canadian Cordillera appears to be the existence of brittle faults that provide a high-permeability flow path for deep-circulating meteoric water. Neither infiltration rate nor heat flux appear to have a significant influence on the distribution of springs.

Estimate of circulation depth

Circulation depths may be estimated by comparing the temperature of thermal springs with local geothermal gradients. However, temperatures measured at the surface outlet of springs are unlikely to be representative due to differences in flow rates and near-surface cooling. Only direct physical measurements can provide reliable subsurface temperatures. Since this is not feasible, an alternative is to use aqueous geothermometers to estimate the maximum temperatures reached in the thermal systems (Kharaka and Mariner 1988).

Grasby et al. (2000) examined a number of chemical and stable isotope geothermometers for thermal springs in the Canadian Cordillera to obtain estimates of subsurface temperature (values are provided in Table 1). Due to high Ca concentrations, precipitation of Ca minerals at surface, and relatively low temperatures (<180°C), Grasby et al. (2000) argue that the chalcedony geothermometer (Arnorsson 1975) provides the most reasonable temperatures for carbonate-hosted springs. In silicate- and volcanic-hosted springs, it is uncertain which phase (quartz or chalcedony) controls silica activity (Arnorsson 1975). Quartz temperatures tended to be fairly consistent with the Na–K–Ca geothermometer, which is favored over Na–K due to more rapid equilibration (Kacandes and Grandstaff 1989). Therefore, an average of quartz and Na–K–Ca temperatures was used for silicate and volcanic hosted springs. Reequilibration along the flow path and dilution by near surface mixing, would lower the calculated temperatures. Thus, the geothermometers give a minimum estimate of the highest temperature reached in the system.

In a strong flow system, geothermal gradients are likely perturbed. However, without detailed information about the flow path, it would be difficult to accurately predict the true gradient. Numerical models suggest that if advective heat flow is dominant, the geothermal gradient in the recharge zone will tend be lowered (Forster and Smith 1988*a*, 1988*b*). Thus, as with potential errors in the chalcedony geothermometers, errors in the geothermal gradient will tend to minimize depth estimates. Local geothermal gradients were obtained from a variety of sources. Gradients used were 21°C/km for the Rocky Mountains and the Rocky Mountain Trench (Hitchon 1984); 32°C/km along the Columbia River Fault; 34°C/km along the Purcell Trench (Lewis et al. 1992); 29°C/km for the Okanagan Valley Fault (Davies and Lewis 1984); 20°C/km for Vancouver Island (Lewis et al. 1992); and 50°C/km along the Harrison Lake Fault. The calculated minimum depth estimates range from 1.6 to 4.8 km, with an average of 3 km (Table 1). However, these results are minimum estimates and do not preclude the possibility of deeper flow systems.

The circulation depths estimated here confirm that geological features of a significant scale are required to provide pathways for waters from up to 5 km depth to reach the surface.

Influence of fault plane geometry on circulation depth

Although faults act as effective conduits for fluid flow along the fault, they may also be effective barriers to fluid flow across the fault (e.g., Caine et al. 1996). This suggests that the circulation depth of geothermal systems may be a function of the depth to the fault plane below the recharge area. We test this hypothesis by examining the relationship between calculated circulation depth and local geology. We

Fig. 7. Simplified cross-sections across the (*a*) Sulphur Mountain, (*b*) Rundle, and (*c*) McConnell thrusts, after Price (1970) and Price and Mountjoy (1972). Estimated circulation depths (see text) are indicated. The estimated depth roughly corresponds to the depth of the fault plane below the topographic high of the mountain massif.

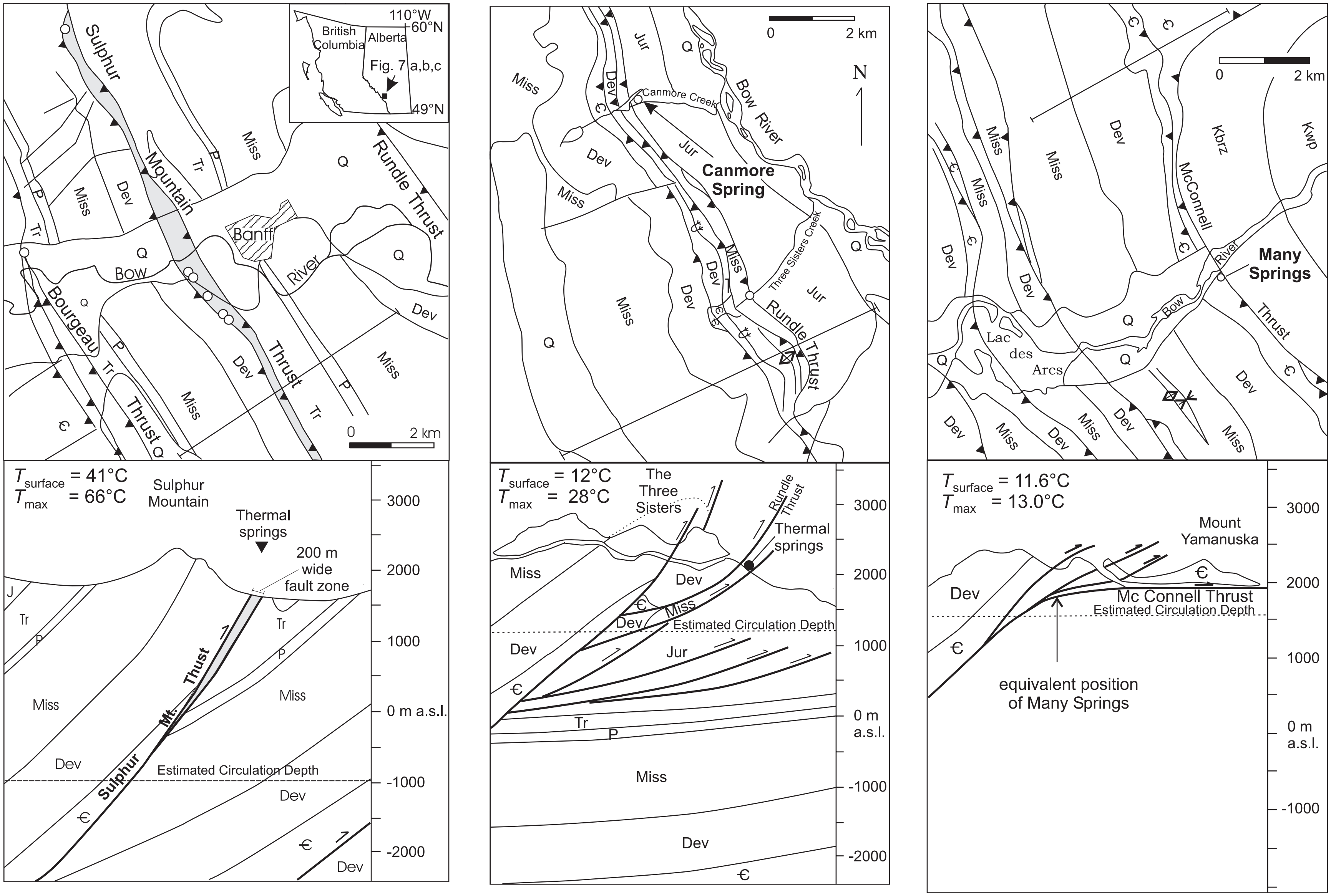


Table 3. Discharge and sulfur concentrations (SO_4 and HS) for thermal springs where both are available.

Spring	Discharge (L/sec)	SO_4 (mg/L)	HS (mg/L)	Total reduced S	S flux (t/a)
1. Sharp Point	8	36	8.2	11	2.7
2. Ahousat	2	11.6	13.9	15	0.9
3. Harrison	250	547		36	284.6
4. Sloquet	100	375		25	78.1
8. Meager	500	125		8	130.1
15. Albert	7	21		1	0.3
17. Halcyon	5	396	12.2	38	6.0
18. Halfway	3	490	1	33	3.2
19. St. Leon	2	560	3.8	41	2.6
20. Nakusp	1	290	5.2	24	0.8
27. Ainsworth	7	58		4	0.8
31. Radium	23	356		23	17.0
32. Fairmont	8	929		61	15.5
34. Lussier	4	148	32	42	5.3
35. Ram Creek	3	56		4	0.3
37. Wildhorse	7	1038		69	15.1
38. Fording	5	1482	123	221	34.8

Note: Total reduced S is estimated as 0.2(S in SO_4) + S in HS (see text). The mass flux of S is total reduced S multiplied by discharge. Discharge data from Fairbank and Faulkner (1992), chemistry from Grasby et al. (2000).

examined only the springs along the Bow Valley because (1) the geology is relatively simple and well defined; (2) the rock type and deformation history are the same, tending to neutralize effects of variations in permeability structure on the flow path; and (3) heat flow and precipitation rates are relatively constant.

Figure 7 illustrates structural cross-sections for the three spring systems in the Bow Valley (Many Springs, Canmore, and the Banff Springs). These springs are associated with the McConnell, Rundle, and Sulphur Mountain thrusts, respectively. For each cross-section, the estimated circulation depth, based on aqueous geothermometry and local geothermal gradients, is marked. Depths are measured relative to the spring outlet. In each case, the estimated circulation depth approximates the depth of the fault plane beneath the topographic high of the mountain. A general model can be considered then, where a mountain massive sits above an associated fault plane. Precipitation that falls on the highland percolates down through the mountain along fractures and porous networks until reaching the fault plane. The fault acts as a barrier to fluid flow deeper into the crust, while acting as a preferential flow path along the fault to surface. As the outcrop trace of faults tends to occur along valley bottoms, there is an inherent topographic drive to the thermal spring system. Therefore, the dip of the fault appears to play a significant role in the depth to which water circulates.

Relationship to ore models

Hydrothermal systems played an important role in the formation of Au–Ag–Pb–Zn deposits in southern British Columbia (e.g., Zhang et al. 1989; Beaudoin et al. 1992*a*). Beaudoin et al. (1992*b*) indicate that Eocene extensional faults in southern British Columbia (the same as some modern thermal springs are associated with) were first-order controls on channeling deep-seated mineralizing fluids towards upper crustal levels. Numerous Au–Ag–Pb–Zn deposits occur in association with these major structures. Beaudoin et al. (1992*b*) also suggest that mixing of these deep-seated fluids with relatively shallow, circulating, sulphide-rich meteoric water led to the precipitation of ore minerals in the mixing zone. They estimate that mixing-zone Ag–Pb–Zn deposits formed around 6 km depth. Modern thermal systems are consistent with these ore models in that they (1) show preferential deep circulation of meteoric water along Eocene extension faults and (2) have estimated circulation depths, up to 5 km, similar to those proposed for mineralization.

Interestingly, the $\delta^{34}S$ values of sulfide ore minerals reported by Beaudoin et al. (1992*a*), 0 to –13‰, are consistent with the $\delta^{34}S$ values of H_2S in thermal spring along the Columbia River Fault, –0.8 to –12.9‰ (Grasby et al. 2000). The H_2S in modern thermal springs is produced by bacterially mediated sulfate reduction (BSR; Krouse et al. 1970; Smejkal et al. 1971; Grasby et al. 2000). The coincidence in $\delta^{34}S$ values and the modern distribution of thermal springs suggest that BSR in thermal system may have played an important role in the formation of sulphide-rich meteoric water that Beaudoin et al. (1992*b*) suggest led to formation of ore deposits in southern British Columbia. We test this hypothesis with a mass-flux calculation. Table 3 lists thermal springs where both chemical analyses and flow rates are available. Sulfur occurs as both oxidized (SO_4^{2-}) and reduced species in the thermal waters (the exact form of reduced sulphur depends on the pH of the water, but HS^- is used here as a general term). The amount of HS^- is the important variable in terms of forming hydrothermal ore deposits. The HS^- measured at surfaces will be less than in the subsurface, as HS^-

oxidizes rapidly to SO_4^{2-}. Grasby et al. (2000) estimate that up to 50% of SO_4^{2-} measured at surface occurs as HS^- at depth. As a conservative estimate, we take the total concentration of HS^- at depth to be $0.2(SO_4^{2-}) + HS^-$. This gives estimated mass-flux rates of reduced sulfur in thermal spring systems of 0.3 to 285 t/a (Table 3).

It is constructive to compare the mass flux of reduced sulfur in thermal springs along the Eocene Columbia River fault to the size of hydrothermal ore deposits associated with similar Eocene extensional faults in the Kokanee Range. Approximately one million t of Pb–Zn have been produced from the deposits along the Slocan Lake Fault to date, with an associated 0.3 million t of sulfur. That amount of sulfur could be transported to depth in 50 000 years by a single, relatively low discharge spring (e.g., the 5 L/s Halcyeon spring; 17 in Table 3), with a sulfur flux of 6 t/a. Thus, modern thermal springs are capable of transporting large amounts of reduced sulphur to ore-forming depths on a short geological time scale and may form good analogies for the upper-crustal flow systems in hydrothermal ore models that invoke mixing of meteoric and deep-seated mineralizing fluids (e.g., Beaudoin et al. 1992*b*).

Conclusions

The distribution of thermal springs in the southern Canadian Cordillera is not random; instead thermal springs tend to be restricted to major crustal-scale brittle faults. These brittle faults are a primary control on the development of thermal systems in that they provide a permeable pathway that allows hot water at depth to flow quickly to the surface before losing heat. Normal geothermal gradients in the Cordillera are sufficient for thermal springs to develop, provided there is a suitable flow path. High heat flow, magmatic activity, or radiogenic heat sources are not necessary conditions for the occurrence of thermal springs in the Canadian Cordillera.

Calculated circulation depths for meteoric water involved in thermal springs systems range from 1.6 to 4.8 km. Circulation depths are probably controlled by the geometry of the fault plane that acts as a conduit to the surface.

The modern-day distribution of thermal springs and their estimated circulation depths are consistent with models for formation of hydrothermal ore deposits associated with Eocene extensional features in southern British Columbia. Mass-flux calculations indicate that thermal springs can transport significant amounts of reduced sulfur (up to 285 t/a) to depth over a relatively short period of geological time. Thus modern thermal springs form a good analogy for the upper crustal circulation system invoked in hydrothermal ore models.

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