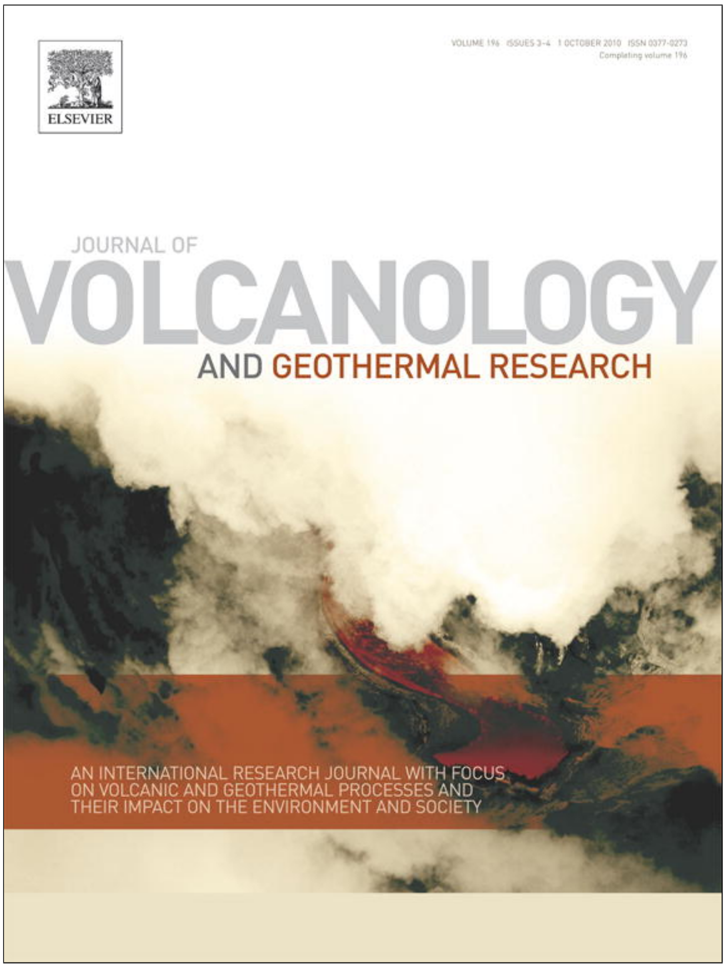
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Hydrothermal heat discharge in the Cascade Range, northwestern United States

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ABSTRACT

Hydrothermal heat discharge in the Cascade Range includes the heat discharged by thermal springs, by "slightly thermal" springs that are only a few degrees warmer than ambient temperature, and by fumaroles. Thermal-spring heat discharge is calculated on the basis of chloride-flux measurements and geothermometer temperatures and totals ~240 MW in the U.S. part of the Cascade Range, excluding the transient post-1980 discharge at Mount St. Helens (~80 MW as of 2004–5). Heat discharge from "slightly thermal" springs is based on the degree of geothermal warming (after correction for gravitational potential energy effects) and totals ~660 MW. Fumarolic heat discharge is calculated by a variety of indirect and direct methods and totals ~160 MW, excluding the transient mid-1970s discharge at Mount Baker (~80 MW) and transient post-1980 discharge at Mount St. Helens (>230 MW as of 2005). Other than the pronounced transients at Mount St. Helens and Mount Baker, hydrothermal heat discharge in the Cascade Range appears to be fairly steady over a ~25-year period of measurement. Of the total of ~1050 MW of "steady" hydrothermal heat discharge identified in the U.S. part of the Cascade Range, less than 50 MW occurs north of latitude 45°15′ N (~0.1 MW per km arc length from 45°15′ to 49°N). Much greater rates of hydrothermal heat discharge south of 45°15′N (~1.7 MW per km arc length from 40° to 45°15′N) may reflect the influence of Basin and Range-style extensional tectonics (faulting) that impinges on the Cascades as far north as Mount Jefferson but is not evident farther north.

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1. Introduction

In this paper we summarize the results of 25 years of intermittent measurement of hydrothermal heat discharge in the United States portion of the Cascade Range, a volcanic-arc segment that extends roughly 1000 km from latitude 40°N to the Canadian border at 49°N. A compilation of best-available data (Table 1) reveals pronounced along-arc variations in hydrothermal heat discharge. In particular, there is little hydrothermal heat discharge (~0.1 MW/km arc length) north of latitude 45°15′N, substantially more to the south (~1.8 MW/km arc length). We will first describe the hydrothermal data and how they are collected. We then relate these data to regional volcanic-vent distributions, conductive heat flow, and geologic structure. Finally, we consider transient variations in hydrothermal discharge, and relate the Cascade Range observations to observations made in other regions of focused volcanism.

This paper builds on a report by Mariner et al. (1990) that summarized heat discharge by thermal springs in the U.S. portion of the Cascades. Here we take a broader view of hydrothermal discharge, including also heat discharge by fumaroles and by "slightly thermal" springs only a few degrees warmer than ambient temperature. These slightly thermal springs appear to represent the largest single mode of hydrothermal heat discharge in the Cascade Range — a mode that was not yet quantified in 1990, but has attracted considerable attention since (*cf.* Nathenson et al., 1994; Manga, 1998; James et al., 2000; Nathenson et al., 2003; Manga and Kirchner, 2004). We also take advantage of detailed post-1990 studies at Mount St. Helens (Bergfeld et al., 2008; Edmonds et al., 2008; Gerlach et al., 2008) and Mount Hood (Bergfeld et al., 2004); in the north-central Oregon Cascade Range (Ingebritsen et al., 1994) and particularly in the Three Sisters region (Evans et al., 2004); at Crater Lake (Wheat et al., 1998); and in the Lassen region (Paulson and Ingebritsen, 1991; Sorey and Colvard, 1994; Sorey et al., 1994).

The first reproducible measurements of hydrothermal heat discharge in the Cascade Range were made at Lassen in the 1920s by Day and Allen (1925). Much more widespread measurement – mainly of fumarolic areas – was triggered by the availability of thermal-infrared remote-sensing technology in the 1970s. Early thermal-infrared data and complementary ground-based measurements (cf. Friedman and Frank, 1980; Friedman et al., 1982) remain the primary source of information on fumarolic heat discharge from some volcanoes, although such data have been superceded by intensive ground-based measurements at Mount St. Helens (Edmonds et al., 2008) and Lassen (Sorey and Colvard, 1994). Intermittent measurement of thermal-spring discharge began in 1984 under the auspices of the USGS Geothermal (1984 to mid-1990s) and Volcano Hazards Programs (mid-1990s to present). Recently (2002–present)

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Table 1
Current best estimates of hydrothermal heat discharge in the Cascade Range. Features discharging <1 MW heat are excluded.

Locality	Thermal springs[a]	"Slightly thermal" springs[b]	Fumaroles
Mount Baker	1 MW Mariner et al. (1990)	Possible (Sulphur Cr.) but not quantified	10 MW 1972 81 MW 1975 Friedman and Frank (1980)
Glacier Peak	4 MW Mariner et al. (1990)	N/A	N/A
Mount Rainier	7 MW Mariner et al. (1990)	Possible (Winthrop, Paradise: Frank (1995)) but not quantified	10 MW Frank (1985)
Mount Adams	N/A	N/A	N/A
Mount St. Helens	80 MW (transient) Bergfeld et al. (2008)	Large transients from 1980–present	>230 MW (transient)[c] Edmonds et al. (2008)
Mount Hood	3 MW Mariner et al. (1990)		10 MW Friedman et al. (1982)
	5 MW Nathenson (2004)	Included in "thermal" total	4 MW Bergfeld et al. (2004)
North-central Oregon	85 MW Austin HS 27 MW Kahneeta HS 9 MW Breitenbush HS Ingebritsen et al. (1994)	140 MW Lower Opal Spr. Manga and Kirchner (2004) 20 MW Metolius headwaters 30 MW Spring River (James et al., 2000, after GPE correction)	N/A
Three Sisters	20 MW Ingebritsen et al. (1994)	16 MW Separation Cr. sprs. Evans et al. (2004)	N/A
Upper Willamette River	4 MW Mariner et al. (1990)	N/A	N/A
Newberry	13 MW Mariner et al. (1990)	N/A	N/A
Umpqua River	3 MW Mariner et al. (1990)	N/A	N/A
Crater Lake	30 MW lake-floor sprs. Wheat et al. (1998)	87 MW Wood R. Sprs. Nathenson et al. (1994)	N/A
Medicine Lake	N/A	360 MW Fall R. sprs. Manga and Kirchner (2004)	1–2 MW (authors' field survey, 2009)
Shasta	N/A	0 to >>18 MW Shasta V. sprs. Manga and Kirchner (2004)	1–10 MW (T_g~210 °C)
Big Bend	2 MW Mariner et al. (1990)	N/A	N/A
Lassen	26 MW Sorey et al. (1994)	>1 MW Domingo Sprs. *etc.* (Paulson and Ingebritsen, 1991; Sorey et al., 1994)	120 MW (Sorey and Colvard, 1994)

[a] Heat output based on geothermometer temperatures (T_g), except at Mount St. Helens which assumes 60 °C.
[b] Heat output based on degree of geothermal warming (cf. Manga and Kirchner, 2004).
[c] Based on 7200 t/d H_2O (140 t/d CO_2) at ~2800 kJ/kg on 31 August, 2005. Gerlach et al. (2008) suggest that the median H_2O emission rate in 2004–5 was actually ~30,000 t/d.

the intermittent thermal-spring measurements have been complemented by semi-continuous (hourly) monitoring at a few locations. The important role of "slightly thermal" springs began to be quantified in the 1990s (Nathenson et al., 1994; Manga, 1998).

Our data and interpretations are confined to the U.S. portion of the Cascade Range. However, there are geological and geophysical similarities between the U.S. Cascade Range north of about 48 °C and the Canadian segment, which together comprise the Garibaldi Volcanic Belt (Guffanti and Weaver, 1988; Hildreth, 2007). Thus we speculate that hydrothermal conditions in the Canadian Cascades may be similar to those in the northernmost United States.

2. Methods of measurement

In this paper we are concerned with hydrothermal heat discharge from three categories of features — thermal springs, fumaroles, and "slightly thermal" springs that are only a few degrees warmer than ambient temperature. Here we describe measurement methods that apply to nearly all the data summarized in Table 1.

2.1. High-chloride thermal springs

Many thermal-spring discharge areas include numerous vents, some of which may be beneath streams or lakes or otherwise inaccessible, so that measurements of individual vents can rarely succeed in capturing the total discharge. However, most thermal springs occur in valleys, near streams that eventually capture most of the thermal fluid. Thus their total discharge can often be gauged by measuring the solute flux in adjacent streams (Ellis and Wilson, 1955). Chloride is the most commonly used indicator of thermal-spring discharge, because it behaves conservatively in solution and because thermal waters are usually much higher in chloride than nearby surface water and/or shallow groundwater. In the Cascade Range, chloride concentrations in surface waters are typically in the range of local precipitation (0.2–0.6 mg/L) unless there is some thermal- or mineral-spring input. Chloride concentrations in the thermal waters themselves are in the range of 100s to 1000s of mg/L (Mariner et al., 1990), providing a strong contrast. Although other ions present in elevated concentrations in thermal waters (*e.g.*, As, B, Na, SO_4) can be used in solute inventories, they are much more likely to be affected by reactions in streams or the shallow subsurface. As depicted in Fig. 1, the discharge rate of a thermal-spring group is calculated from the chloride concentration upstream and downstream of the thermal springs, the chloride concentration in the thermal water itself, and the discharge rate of the stream. The measure of thermal-spring heat discharge that we adopt here is

$$Q_{\mathrm{thermal}} = Cl_{\mathrm{flux}} c_w \left(T_{\mathrm{geo}} - T_{\mathrm{rch}}\right) / Cl_t, \qquad (1)$$

where Cl_{flux} is the hydrothermal chloride flux (Fig. 1), c_w is the heat capacity of the fluid, T_{geo} is the maximum fluid temperature at depth,

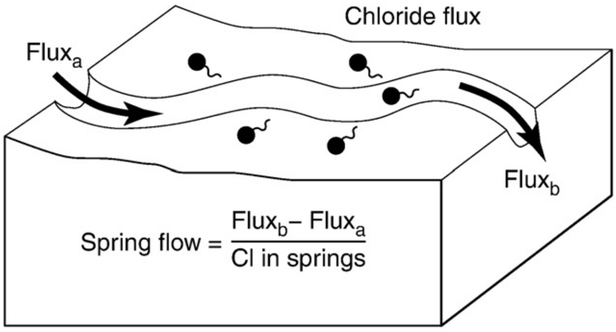

Fig. 1. Block diagram illustrating chloride-flux method of measuring thermal-spring discharge. The difference between chloride flux upstream (Flux$_a$, g/s) and downstream (Flux$_b$) of a thermal-spring group is divided by the chloride concentration in the thermal-spring waters (g/L) to determine thermal-spring discharge (L/s). That is, $D_{ts} \sim (D_s[Cl_d - Cl_u]/Cl_t)$, assuming that $D_{ts} << D_s$ and $Cl_t >> Cl_d$ or Cl_u, where D_{ts} is thermal-spring discharge, D_s is stream discharge, Cl_u and Cl_d are stream chloride concentrations above and below the thermal springs, respectively, and Cl_t is chloride concentration in the thermal springs themselves.

as determined by chemical geothermometry or other means, T_{rch} is the recharge temperature (0–5 °C in the Cascade Range), and Cl_t is the chloride concentration in the thermal springs themselves. As thus defined, $Q_{thermal}$ is a measure of the heat advected away from a deep heat source, rather than heat discharged directly by the thermal springs; thermal-spring discharge temperatures (≤100 °C) are often $<<T_{geo}$, due to conductive (or other) heat loss as the fluid moves toward the thermal-spring orifices. Reported geothermometer temperatures are generally in the range of 100–240 °C (*e.g.* Mariner et al., 1990), implying that the thermal waters have circulated to depths of several km. We rely mainly on SO_4-H_2O isotope geothermometers, which agree remarkably well with anhydrite-saturation temperatures for many Cascade Range thermal waters (Mariner et al., 1993).

The chloride-flux method depicted in Fig. 1 and Eq. (1) was used to determine most of the thermal-spring heat-discharge values in Table 1. The major exception is the lake-bottom thermal springs at Crater Lake, where difficulty of access requires less-direct methods. For Crater Lake, mass-balance and geothermometry calculations (Wheat et al., 1998) indicate a hydrothermal heat output of 30 ± 5 MW, which is the value that we adopt in Table 1. This is similar to the Crater Lake values obtained independently from lake-bottom heat-flow measurements (15–31 MW: Williams and Von Herzen (1983)) and deep-lake temperature and salinity profiles (23 ± 8 MW: McManus et al. (1993)).

2.2. Fumaroles

In areas of fumaroles (steam vents) and associated acid-sulfate springs, there are significant modes of heat discharge that cannot be captured by a simple solute inventory. Significant heat loss from fumarolic areas occurs by direct discharge from fumaroles (Q_{fum}); by direct discharge from hot springs (Q_{hs}) and lateral seepage in the subsurface (Q_{lat}); by evaporation, radiation, conduction, and molecular diffusion from water surfaces (Q_{ws}); and by conduction, advection, and evaporation from warm or steaming ground (Q_{gr}). Thus

$$Q_{thermal} = Q_{fum} + Q_{hs} + Q_{lat} + Q_{ws} + Q_{gr}, \quad (2)$$

where $Q_{thermal}$ is the total heat loss from the thermal area. Measurement of the multiple modes of heat discharge is time-consuming and difficult, and evaluation of most terms is highly model-dependent (see Dawson, 1964; Dawson and Dickinson, 1970; Yuhara, 1970; Sekioka and Yuhara, 1974; and Sorey and Colvard, 1994). Associated uncertainties are large, and time series are sparse and rare, both globally and in the Cascade Range itself. Relatively comprehensive heat-loss studies have been done in fumarolic areas at Wairakei (New Zealand), Poas (Costa Rica), and, in the Cascade Range, in Lassen Volcanic National Park, California. These studies reveal that although the fumaroles themselves are highly visible, Q_{fum} is generally a minor component of $Q_{thermal}$, accounting for only approximately 3% of the 430 MW of natural heat loss measured at Wairakei in the 1950s (Dawson and Dickinson, 1970); 5% of the 265 ± 100 MW measured at Poas, Costa Rica in 1988 (Brown et al., 1989); and approximately 10% of the 115 ± 9 MW measured at Lassen, California in 1984–93 (Sorey and Colvard, 1994). Heat loss from-open water surfaces (Q_{ws}) consistently emerges as a dominant heat-loss mode, accounting for ~33% of heat loss at Wairakei, ~52% at Lassen, and ~83% at Poas. Heat loss from bare ground (Q_{gr}) is significant both at Wairakei (40%) and Lassen (17%).

2.3. Slightly thermal springs

Even low-temperature springs – those having temperatures within a few degrees of ambient – may discharge large amounts of geothermal heat, on the order of 10s to 100s of MW (Manga, 1998). In general, spring temperature varies with elevation roughly according to an adiabatic lapse rate of 4–6 °C/km. Spring temperatures that do not obey this relation may be affected by geothermal warming. For systems that are recharged at high elevations, an important caveat is that spring temperatures can also be significantly affected by the conversion of gravitational potential energy to heat (Domenico, 1972; Manga and Kirchner, 2004). The energy balance for water flowing through a simple spring system (Fig. 2) can be written as

$$\begin{aligned} &\textit{rate-of-change of thermal energy} = \textit{GPE dissipation} \\ &\quad + \textit{conductive heat loss to surface} + \textit{geothermal heating} \end{aligned}$$

where GPE indicates gravitational potential energy. In differential form,

$$\rho_w n b c_w \frac{dT}{dt} = \rho_w g b q_w \sin\theta + \frac{K_m(T-T_s)}{d} + q_h, \quad (3)$$

where ρ is density, n is porosity, b is the aquifer thickness, T and T_s are the temperatures in the aquifer and at the land surface, respectively; t is time; c is heat capacity; g is gravitational acceleration; q_w is the volumetric flow rate per unit area; θ and d are the slope of and depth to the aquifer, respectively; K_m is the thermal conductivity of the medium; q_h is geothermal heating; and the subscript w refers to the properties of liquid water (Manga and Kirchner, 2004). In some volcanic terranes, heat loss to the surface is negligible, because high rates of groundwater recharge cause the value of $(T-T_s)$ to be near-zero — a phenomenon well-documented in the Oregon Cascade Range

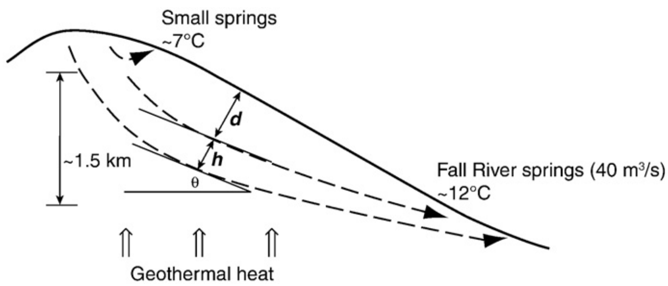

Fig. 2. Schematic diagram of the groundwater system at Medicine Lake volcano, California. Typical temperature for small, high-elevation springs (7 °C) is inferred to represent the recharge temperature for the regional-scale Fall River spring system. Because of the large discharge of the Fall River springs, the relatively modest warming between recharge and discharge areas amounts to several hundred megawatts of heat discharge. After Manga and Kirchner (2004).

(*e.g.* Ingebritsen et al., 1989, 1992). The heat generated by GPE dissipation can be calculated by assuming that all GPE loss between recharge and discharge elevations is ultimately converted to heat though viscous dissipation (friction). Heating due to this mechanism amounts to approximately 2.3 °C per kilometer of elevation (Domenico, 1972, p. 160, example 4.3). Thus for the Medicine Lake–Fall River spring system depicted in Fig. 2, the amount of heating due to GPE dissipation likely exceeds the geothermal heating, and the geothermal heating would be greatly overestimated if GPE dissipation were not taken into account.

3. Distribution of hydrothermal heat discharge

There are pronounced variations in hydrothermal heat discharge along the length of the Cascade Range (Fig. 3, Table 1). In this section we will be concerned only with the distribution of "steady" or average hydrothermal discharge. Transient hydrothermal discharge – notably at Mount Baker in the 1970s and Mount St. Helens from 1980 to present – will be discussed in Section 4.

Most obvious in Fig. 3 is the paucity of hydrothermal heat discharge north of 45°15′N. Of the total of ~1050 MW of "steady" hydrothermal heat discharge identified in the U.S. part of the Cascade Range (Table 1), less than 50 MW occurs north of latitude 45°15′ N (~0.1 MW per km arc length from 45°15′ to 49°N). Length-normalized rates of hydrothermal heat discharge are thus 15–20 times larger south of 45°15′N (~1.7 MW per km arc length from 40 to 45°15′N).

The general pattern of hydrothermal heat discharge shows some similarities to the distribution of Quaternary volcanic vents depicted in Fig. 4. In particular, the Quaternary vent distribution becomes discontinuous in Washington state, where hydrothermal discharge is low. Further, there is a break in the Quaternary volcanic arc near 42°N (near the California–Oregon border) that corresponds to the negligible hydrothermal discharge between the Medicine Lake highlands (~41°36′N) and the Wood River springs (~42°45′). However, the major change in the density and continuity of Quaternary vents

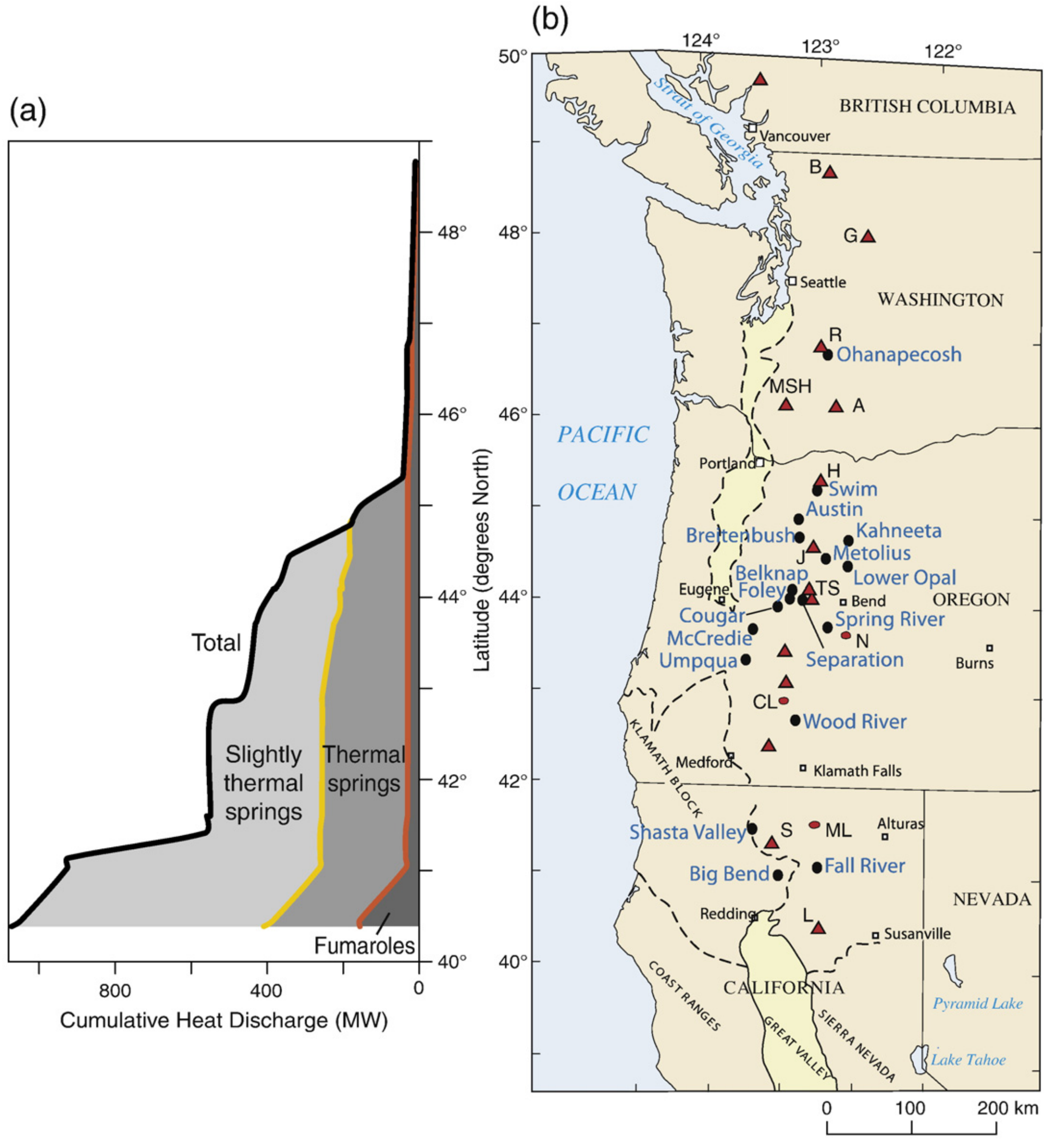

Fig. 3. (a) Cumulative heat discharge from Cascade Range hydrothermal features (summed north to south), showing that <5% of the total hydrothermal heat discharge occurs north of 45°15′ and (b) map showing location of selected Cascade Range hydrothermal features (solid black circles) relative to compositionally evolved major volcanic centers (red triangles and ovals): B, Mount Baker; G, Glacier Peak, R, Mount Rainier; A, Mount Adams; MSH, Mount St. Helens; H, Mount Hood; J, Mount Jefferson; TS, Three Sisters; N, Newberry; CL, Crater Lake; S, Shasta; ML, Medicine Lake; and L, Lassen. Dashed lines denote boundaries between physiographic provinces.

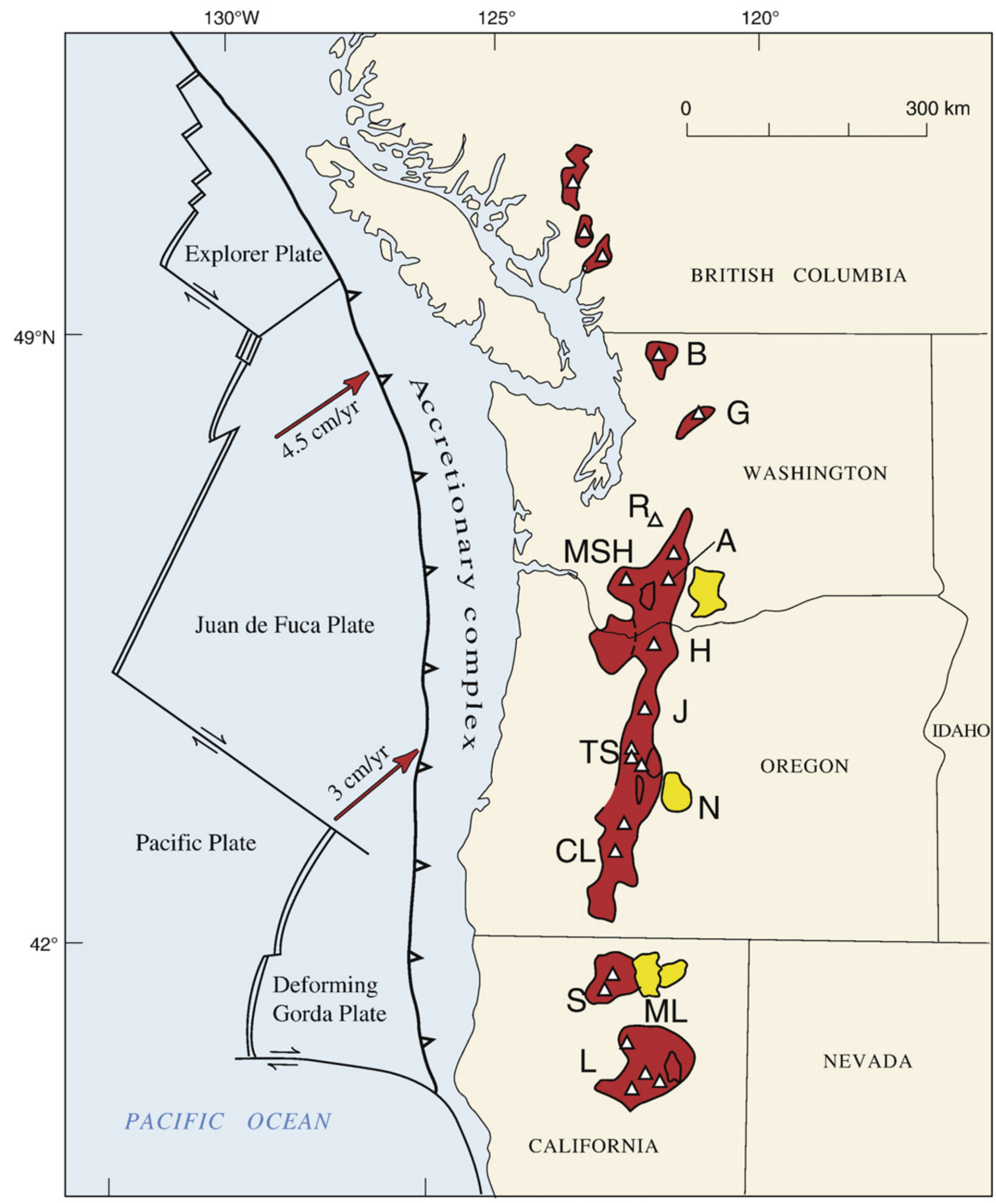

Fig. 4. Quaternary Cascades volcanic arc. Red areas encompass more than 2300 vents from more than 2000 independent volcanoes and yellow areas encompass extensive rear-arc volcanic fields. Despite continuity of offshore subduction, the breaks in the Quaternary arc are real (Hildreth, 2007). Compositionally evolved major centers listed in Table 1 and/or discussed in the text are B, Mount Baker; G, Glacier Peak, R, Mount Rainier; A, Mount Adams; MSH, Mount St. Helens; H, Mount Hood; J, Mount Jefferson; TS, Three Sisters; N, Newberry; CL, Crater Lake; S, Shasta; ML, Medicine Lake; and L, Lassen. After Hildreth (2007).

occurs in south-central Washington near 46°40′, over 150 km north of the distinct change in hydrothermal discharge in north-central Oregon at ~45°15′.

3.1. Relation to compositionally evolved volcanic centers

Perfect correspondence between volcanic-vent distribution and hydrothermal discharge would be unexpected. Most high-temperature hydrothermal systems worldwide are related to silicic magmatism in the upper crust, and many large stratocones or major components of compound edifices can be constructed very rapidly without establishment of significant upper-crustal magma reservoirs (Hildreth and Fierstein, 1995). Mount St. Helens is a prime example of this postulate. For millennia, it has been the fastest-growing, most active volcano in the Cascade Range. The bulk of its pre-1980 edifice was constructed in the past 3900 years (Clynne et al., 2005). Yet, prior to its eruption in 1980, hydrothermal phenomena were nearly absent (Phillips, 1941; Korosec and Schuster, 1980).

As indicated in Table 1, most of the more conspicuous hydrothermal phenomena in the Cascade Range can be related to particular long-lived, compositionally evolved volcanic centers that have some component of silicic magmatism. Many Cascade Range hydrothermal phenomena are proximal to such centers — sufficiently proximal that they cannot be distinguished as separate symbols on Fig. 3b. Many that are less proximal (>5–10 km distant) are clearly connected to their host volcanic centers by topographic (and presumably hydraulic) gradients. Some that are much more remote from candidate volcanic centers (up to 50 km or more distant) nonetheless show the geochemical signatures of crustal magmatism (James et al., 2000).

3.2. Relation to regional heat flow

Although the general pattern of hydrothermal heat discharge (Fig. 3a, Table 1) can be related to the distribution of Quaternary volcanic vents (Fig. 4), the distinct increment in hydrothermal discharge near 45°15′N occurs ~150 km south of the primary change in vent distribution. There appears to be a change in regional conductive heat flow (Fig. 5) near 45°15′N that corresponds more

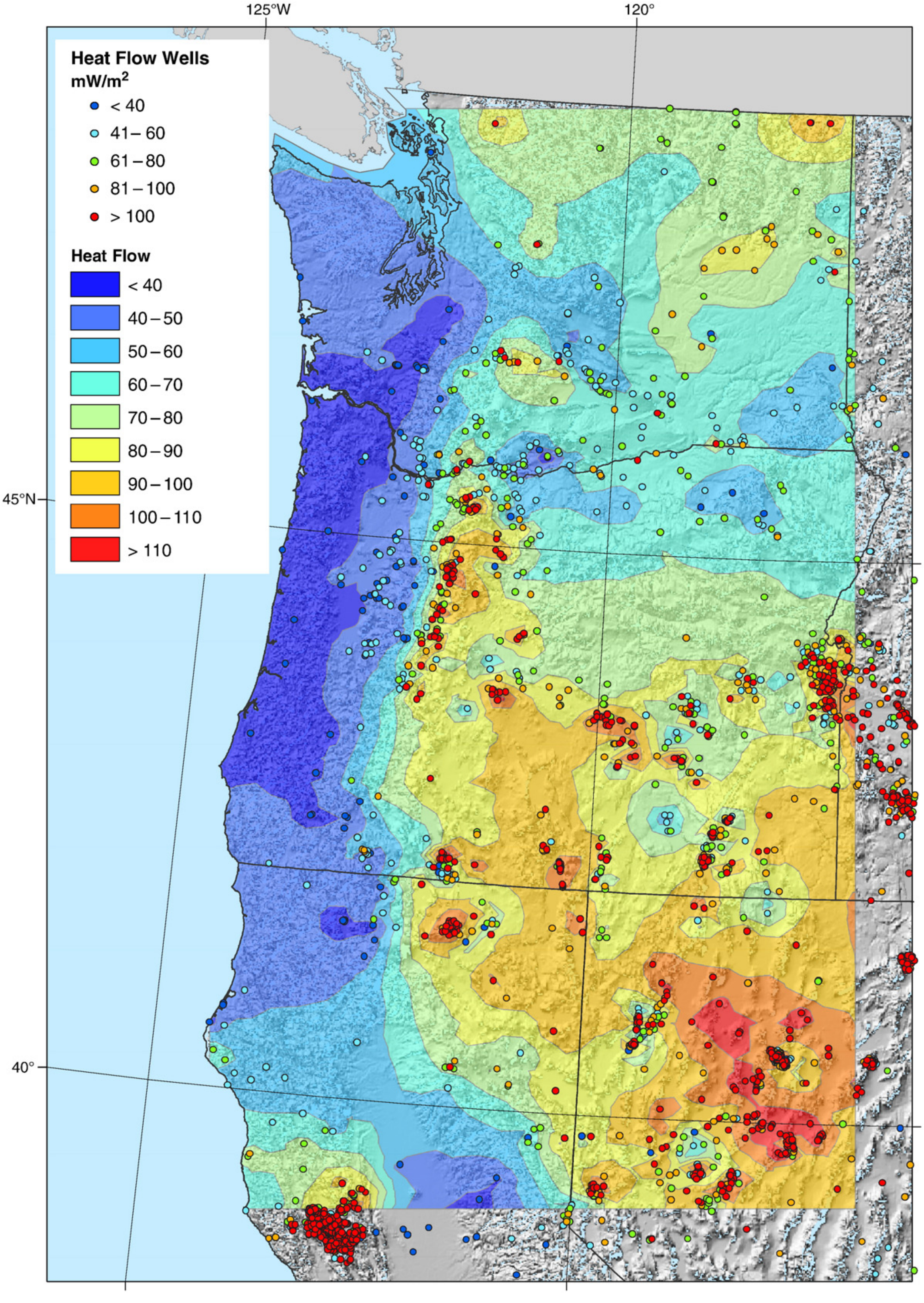

Fig. 5. Heat-flow contour map of the northwestern United States. Contours are based on data for the Cascade Range and adjacent regions from the USGS ArcGIS heat-flow database. From Williams and DeAngelo (2008).

exactly with the increment in hydrothermal discharge. Other than a small area near Mount Hood, there is little evidence for conductive heat flow ≥90 mW/m^2 in the Cascade Range north of 45°15′N. Conductive heat-flow data are sparse in the Washington Cascades (Fig. 5), but are sufficient to suggest that the conductive-heat-flow transition in the Mount Jefferson–Mount Hood region is real.

3.3. Relation to regional tectonics and structure

Both the relatively vigorous thermal-spring discharge (Fig. 3a) and the higher conductive heat flow (Fig. 5) south of approximately 45°15′N may reflect the influence of Basin and Range-style extensional tectonics (faulting). Basin and Range structures impinge on the Cascades as far north as Mount Jefferson (Fig. 6). They are not evident farther north, although Basin and Range-style seismicity (Jones and Malone, 2005) has occurred in the vicinity of Mount Hood, and volcanic-vent alignments (Hildreth, 2007, his Fig. 8) and the north–south trending folds of the Yakima Fold Belt (Hildreth, 2007, his Fig. 25) indicate that extension affects the Cascade Range axis as far north as the Mount Adams region.

Vigorous thermal-spring discharge requires both focused heat and relatively high vertical permeability that permits advective heat transport between deep, hot rocks and the land surface. In order for substantial hydrothermal discharge to occur, hydrothermal upflow must be connected to a deep heat source through a pathway with a time-averaged effective permeability of $\geq 1 \times 10^{-16}$ m^2 (Hurwitz et al., 2003). Subvertical normal faults are perhaps the best candidates for sustaining such connectivity.

How deeply might the thermal-spring waters circulate? Let us consider for instance Lassen, California, where extensional faulting is clearly evident (Fig. 6). There, shallow earthquake clusters (3.5–5.5 km depth) are believed to result from interaction between deeply circulating meteoric recharge and hot but brittle rock (Janik and McLaren, 2010). Other considerations imply that the magma–hydrothermal interface at Lassen must be relatively thin. The magma body (or bodies) themselves are too small to be resolved by seismic surveys, so that the heat-transfer area is restricted to no more than a few km^2. If we take the heat-transfer area to be <5 km^2, the average conductive heat flux over that area must be >20 W/m^2. If we then assume a reasonable thermal conductivity of 2 W/(m-K) and a temperature difference of 500 °C between magma (800 °C) and circulating groundwater (300 °C), then the “conductive boundary layer” between the magma and the hydrothermal system must be <50 m thick in order to transfer the >100 MW of heat (Table 1) that eventually emerges at the land surface. We can thus infer robust hydraulic connection between the land surface and depths of up to 5.5 km. Similarly, Cascade Range geothermometer temperatures in the range of 100–240 °C (*e.g.* Mariner et al., 1990) imply that most thermal waters have circulated to depths of several km.

Although the thickness of the postulated “conductive boundary layer” at Lassen is purely hypothetical, an exploration well at Kakkonda, Japan, penetrated an entire hydrothermal system and part of the underlying neo-granitic pluton. The temperature profile at Kakkonda was boiling-point-controlled to a depth of 3.1 km and conduction-dominated at greater depths (Muroaka et al., 1998, their Fig. 7; Ikeuchi et al., 1998, their Fig. 3). The inflection point of the temperature profile at 3.1 km depth represents the transition from advection-dominated to conduction-dominated heat transfer and, at about 380 °C, may also represent the brittle–ductile boundary.

3.4. Relation to volcanic stratigraphy

Because thermal-spring waters must circulate to depths of several km to attain the requisite temperatures, the impingement of Basin and Range structure offers an attractive explanation for the relative vigor of thermal-spring discharge south of 45°15′. However, the step-like change in total hydrothermal discharge at about 45°15′N (Fig. 3) also reflects the apparent absence of “slightly thermal” springs at more northerly latitudes.

In contrast to the thermal springs, many of the “slightly thermal” springs identified in Table 1 may be regarded as stratigraphically controlled. They are large springs or spring groups that emanate from areally extensive, unconfined aquifers that can capture meteoric recharge (from above) and regional heat flow (from below) over large areas (Fig. 2). Whereas thermal-water temperatures require fluid circulation to several km depth, the aquifers feeding “slightly thermal” springs may be quite shallow (perhaps 100–500 m thick: Manga (1996), Manga and Kirchner (2004)).

With one exception, all of the “slightly thermal” springs identified as discharging substantial amounts of heat are very large-discharge springs. In fact, inventories of large springs (≥~100 cfs, or ~3000 L/s) done by the USGS in the early 1900s showed Lower Opal Springs, the Metolius headwaters, Spring River, the Wood River springs, and the Falls River springs (Fig. 3b) to be among the 65 largest springs in the entire conterminous United States (Meinzer, 1927). The Shasta Valley springs are also identified as unusually large (up to ~600 L/s) in early USGS inventories. The slightly thermal springs in the Separation Creek watershed are the single exception; there, many small springs up to 5 °C above ambient temperature (Evans et al., 2002, 2004) in an area of about 20 km^2 may reflect upward leakage from an underlying high-temperature flow system (Evans et al., 2004, their Fig. 3).

With the exception of the springs in the Separation Creek watershed, the “slightly thermal” springs of Table 1 and Fig. 3 are very large springs fed by areally extensive, shallow, permeable volcanic aquifers. Although the carapace of young, permeable volcanic rocks is fairly continuous as far north as Mount Adams (Fig. 4), no large, “slightly thermal” springs have been identified north of Mount Jefferson. It is possible that the apparent absence of slightly thermal springs between Mount Jefferson and Mount Adams reflects incomplete reconnaissance. However, we note that none of the 26 “large” (≥~3000 L/s) Cascade Range springs identified in the early-1900s USGS inventories occur north of Mount Jefferson (Meinzer, 1927, his Fig. 1 and associated text).

3.5. Concentration of hydrothermal discharge near Mount Jefferson

Perhaps most enigmatic in terms of their relation to major, compositionally evolved volcanic centers (Section 3.1) are the group of springs in north-central Oregon in the general vicinity of Mount Jefferson: Austin Hot Springs, Breitenbush Hot Springs, Kahneeta Hot Springs, the Metolius headwater springs, and Lower Opal Springs (Fig. 3b). Together, these spring groups account for ~300 MW of hydrothermal heat discharge (Table 1), more than ¼ of the total identified in the entire U.S. portion of the Cascade Range. Yet the Mount Jefferson volcanic center itself is relatively small (*cf.* Hildreth, 2007) and inactive (*cf.* Ewert et al., 2005). Further, the waters at Kahneeta (δD ~ −119 to −118: Ingebritsen et al. (1994)) and Lower Opal Springs (δD equivalent ~ −112: James et al. (2000)) are too isotopically depleted to be sourced by modern recharge on Mount Jefferson (δD −109 to −94‰: Ingebritsen et al. (1994)) and are better matched by meteoric recharge on the Newberry or Three Sisters highlands >50 km to the south (δD −117 to −107 and −116 to −99‰, respectively: Ingebritsen et al. (1994)).

3.6. Limited data and associated uncertainty near Mount Adams

The most obvious data gap in the U.S. portion of the Cascade Range is the vicinity of Mount Adams, a compositionally evolved volcanic center in southern Washington (Fig. 3). The best-known aspect of the Mount Adams hydrothermal system is the extensive hydrothermal alteration on and near the summit (Fowler, 1935; Finn et al., 2007) that results from persistent, largely subglacial, solfataric emission

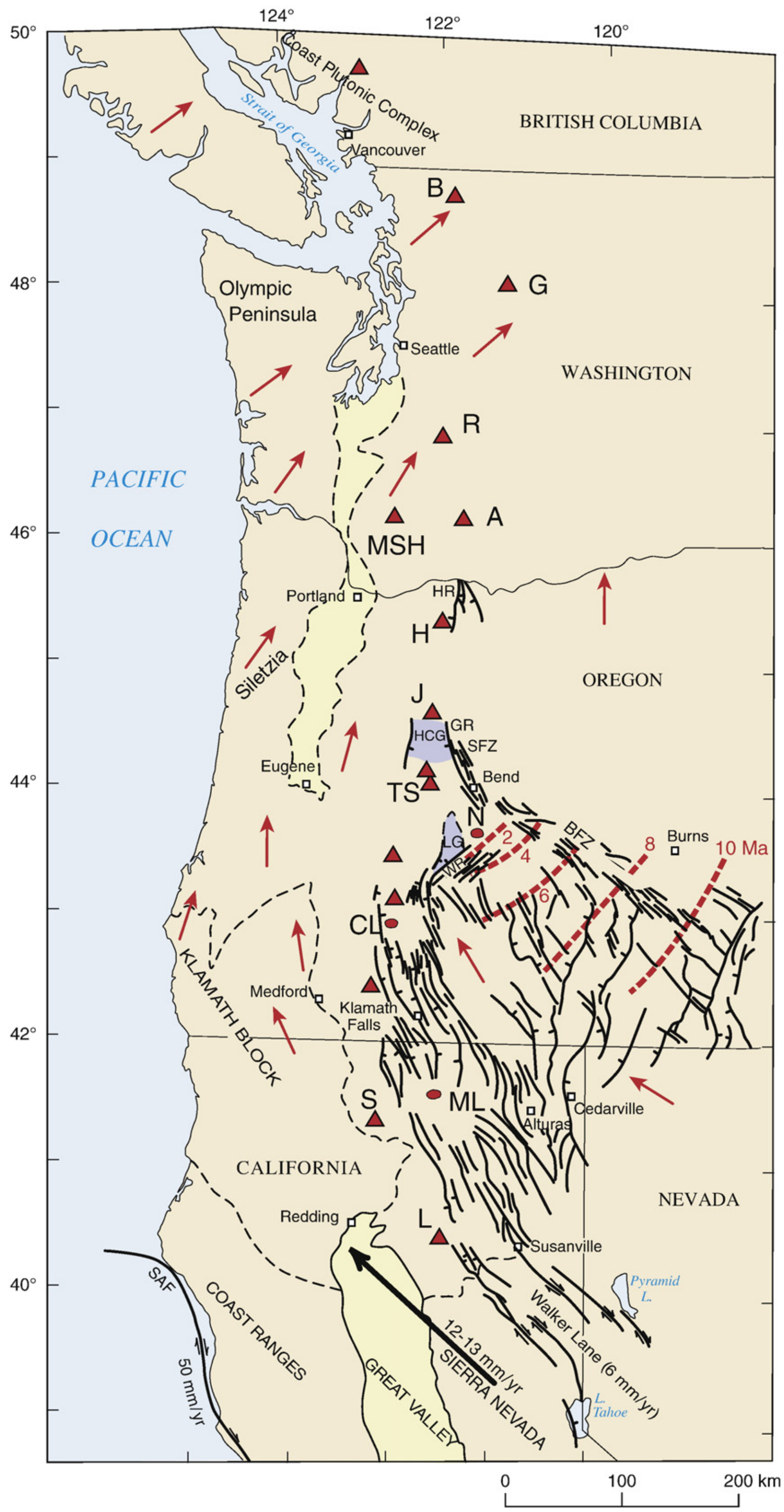

Fig. 6. Tectonic setting of the Quaternary Cascades volcanic arc. Basin and Range extension has expanded westward since late Miocene to overlap the volcanic arc in California and Oregon; black lines (with ticks on downthrown sides) indicate main faults. Clockwise rotation of the Oregon forearc block (red arrows) contributes to intra-arc extension along its trailing edge. Labeled faults are HR, Hood River; GR, Green Ridge; SFZ, Sisters Fault Zone; BFZ, Brothers Fault Zone; HCG, High Cascades Graben; LG, La Pine Graben. Red dashed lines in south-central Oregon indicate westward progression of rhyolitic volcanism across the High Lava Plains. Compositionally evolved major volcanic centers are labeled as in Figs. 3 and 4; see Fig. 4 for generalized distribution of all Quaternary vents. After Hildreth (2007).

surmised to reflect focusing of a weak gas flux from deep melt zones (Hildreth et al., 1983). The summit fumaroles are diffuse and difficult to access. Summit gas data obtained by one of the authors (RHM) in 2005 support the inference of magmatic-volatile discharge ($^3He/^4He = 4.4\ R_A$). A fumarolic temperature of 65 °C was reported by Fowler (1935) – perhaps induced by digging (Hildreth and Fierstein, 1995) – but such temperatures have not been encountered since. Heat discharge from summit fumaroles on other quiescent Cascade volcanoes is typically <10 MW (Table 1). Many other Cascade Range summit fumaroles are hotter, and some cause more extensive melting of summit ice. Thus we suggest that hydrothermal heat discharge from the Mount Adams fumaroles is likely <10 MW, perhaps <<10 MW.

Evidence for lateral flow of a Mount Adams hydrothermal fluid towards thermal or "slightly thermal" springs does not exist, although there are five or six thermal or mineral springs within ~40 km of the summit: Orr Creek Warm Springs, Klickitat Meadow Soda Spring, McCormick Meadow Soda Spring, Soda Spring Creek Soda Spring, Fish Hatchery Warm Spring (Korosec et al., 1981), and Klickitat River weir spring (Hildreth and Fierstein, 1995, their Table 2). These are all relatively low-temperature (<22 °C), low-discharge features, and it is unlikely that any single one accounts for >1 MW of heat.

Because of the lack of reliable data from Mount Adams, we do not include it in our Cascade Range totals (Table 1). The likely hydrothermal heat discharge in the vicinity of Mount Adams is <10 MW, or approximately 1% of the Cascade Range total, insufficient to affect our conclusions about the overall distribution of hydrothermal heat loss.

4. Time-variation of hydrothermal heat discharge

Major hydrothermal transients in the Cascades have been observed only in conjunction with the volcanic unrest at Lassen in the early 20th century (Day and Allen, 1925), at Mount Baker in the 1970s, and at MSH from 1980 to present. At Mount Baker, fumarolic heat discharge temporarily increased from ~10 MW in 1972 to ~80 MW in 1975 during a period of volcanic unrest (Friedman and Frank, 1980). And although pre-1980 hydrothermal discharge at Mount St. Helens was negligibly small, recent (post-2004) hydrothermal discharge there amounts to several hundreds of MW (Table 1). In fact, current (2005–present) rates of hydrothermal heat discharge at Mount St. Helens (>300 MW, Table 1) are comparable to other short- and long-term cooling indices. For instance, the rate of progressive magnetization of the cooling Mount St. Helens lava dome in 1984–1986 indicates roughly 125 MW of heat loss (Dzurisin et al., 1990), and the long-term growth rate of the Mount St. Helens edifice (~0.2 m^3/s) translates to roughly 270 MW, given a latent heat of crystallization of 420 kJ/kg for basaltic rocks (Stakes and Taylor, 2003), a density of 2500 kg/m^3, a heat capacity of 1 kJ/(kg-K), and cooling from 1200 °C to an ambient temperature near 0 °C.

Available data spanning several decades indicate fairly steady hydrothermal discharge in the Cascade Range under conditions of volcanic quiescence (*cf.* Ingebritsen et al., 2001). This general "steadiness" is in marked contrast to mid-ocean ridge systems, which seem to exhibit much more short-term variability (*cf.* Von Damm et al., 1997). The USGS record from Austin Hot Springs (Fig. 7) – which at ~85 MW is by far the largest single hydrothermal discharge in the Cascade Range – is typical of those from high-chloride thermal springs. Relatively frequent measurement beginning in 2002 revealed a distinct seasonality to the hydrothermal flux at Austin. However, there is no evidence for a change in the mean behavior during the 22-year (1984–2006) period of record (Fig. 7). This general absence of multidecadal-scale variability seems sensible in light of the longevity of the probable heat sources and the likely spatial and temporal scales of the high-chloride thermal-spring systems. The likely magmatic heat sources for most of the systems are large enough to retain significant heat for 10^4 to 10^6 years (Smith and Shaw, 1975, 1979; Hayba and Ingebritsen, 1997), likely flow-path lengths range from a few kilometers to tens of km, and likely fluid travel times are on the order of 10^2 to 10^4 years (Ingebritsen et al., 1994). Cascade Range hydrothermal systems differ in these respects from the more dynamic subsea hydrothermal systems associated with relatively shallow magmatism along the mid-ocean ridge.

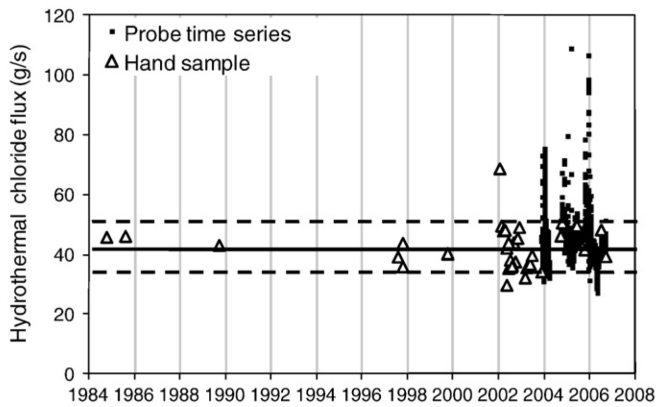

Fig. 7. Hydrothermal chloride flux from Austin Hot Spring 1984–2006, based on intermittent late-summer measurements 1984–1999, monthly measurements in 2002–2003, and continuous (twice daily to hourly) measurement in 2004–2006. Open triangles denote site visits, and solid squares are based on data from conductivity-temperature-pressure probes. Hydrothermal chloride flux is determined by the method shown in Fig. 1, with the value of Q_s at USFS streamgage at Big Bottom (14208000) estimated on the basis of measured Q_s at the USGS streamgage at Three Lynx (14209500). Solid/dashed lines are mean ± standard deviation for the entire period of record. The mean hydrothermal chloride flux of ~43 g/s can be converted to a mean thermal-spring discharge of 110 L/s and a hydrothermal heat discharge of 85 MW on the basis of the thermal-spring chloride concentration of 390 mg/L and a geothermometer temperature of 186 °C. The frequent measurements in 2004–2006 reveal a strong seasonal signal, but there is no clear trend in mean discharge over a period of observation of ~22 years.

Most of the heat-flux time series from fumarolic areas are insufficient to document the presence or absence of trends. A major exception is Devils Kitchen, Lassen, California, where measurements made in Hot Springs Creek by Day and Allen (1925) in the early 1920s ($n = 2$) and Friedman and Frank (1978) in the 1970s ($n = 1$) appear to be compatible with measurements made in 1986–1996 ($n = 13$) (13.4 ± 5.3 MW; Ingebritsen et al., 2001, their Fig. 8). At Mount Hood, Oregon, the fumarolic heat discharge of ~10 MW measured in the 1970s (Friedman et al., 1982) is somewhat larger than the heat discharge that would be inferred from the CO_2 discharge of 6–7 tonnes/day measured in 2003 (Bergfeld et al., 2004), assuming a H_2O:CO_2 weight ratio of ~20:1 (Symonds et al., 2003) (6.5 t/day = 0.075 kg/s $CO_2 \times 20 = 1.4$ kg/s $H_2O \times 2800$ kJ/kg ~ 4 MW).

5. Discussion

The most prominent single feature of hydrothermal heat discharge in the Cascade Range is the step-like change at about latitude 45°15′N (Fig. 3a), from a length-normalized rate of ~0.1 MW per km arc length between 45°15′ and the Canadian border to a length-normalized rate of ~1.7 MW per km arc length in northern California and most of Oregon. Prominent second-order features include the absence of significant hydrothermal discharge between the Wood River springs (~42°45′) and the Medicine Lake highlands (~41°36′N), which corresponds to a break in the Quaternary volcanic arc near the California–Oregon border (Fig. 4), and the concentration of ~300 MW of hydrothermal heat discharge within about 50 km of Mount Jefferson in north-central Oregon.

The step-like change at about 45°15′N may in part reflect Basin and Range impingement that has the dual effects of enhancing crustal heat flow (Fig. 5) and, through extensional tectonics (Fig. 6), providing deep permeability for fluid circulation. The concentration of hydrothermal discharge in the general vicinity of Mount Jefferson remains enigmatic and warrants further investigation. Although Mount Jefferson itself is relatively small (*cf.* Hildreth, 2007) and inactive (*cf.* Ewert et al., 2005), it lies within an andesite–dacite anomaly — a 20 km × 8 km axial strip where mafic magmas have been excluded throughout the Quaternary (Conrey, 1991; Conrey et al., 2001). A zone of northwest–southeast-trending faults may connect Austin Hot Springs (Figs. 3 and 7) to the Mount Jefferson area (Sherrod and Conrey, 1988).

Over the time scale of human observation, the discharge and temperature of the large "slightly thermal" springs of the Cascade Range are remarkably constant, based on comparison between early-20th century data (*cf.* Meinzer, 1927) and more recent measurements. Excluding periods of volcanic unrest, the actual thermal springs show a strong seasonal signal superimposed on relatively constant mean behavior (*e.g.* Fig. 7). Although substantial fumarole time series are lacking, we speculate that ongoing measurements in fumarolic areas will eventually define a strong seasonal signal. In general, we expect that most Cascade Range hydrothermal features can be characterized as "steady with seasonality" over observational time scales. Over longer time scales (>~10^2 years), we would expect temporal variations that reflect the waxing and waning of magmatic heat sources. This is to be expected because the large heat discharges from thermal springs are ultimately sustained by magmatic heat input. A heat discharge of ~100 MW – as seen at Austin Hot Springs and at Lassen (Table 1) – equates to crystallization and cooling of silicic magma at a rate of ~0.05 m^3/s, given a latent heat of crystallization of 270 kJ/kg (Harris et al., 1970), a density of 2500 kg/m^3, a heat capacity of 1 kJ/(kg-K), and cooling from 800 °C to an ambient temperature of 300 °C.

The length-normalized heat-discharge rate for the entire 1000-km length of the U.S. Cascade Range is slightly greater than 1 MW/km arc length. This is more than 10 times greater than the rate reported by Mariner et al. (1990), who focused more narrowly on thermal springs and evaluated thermal-spring heat discharge on the basis of discharge temperatures (rather than geothermometer temperatures). The larger rate reported here is superficially similar to that reported for Japan (1.2 MW/km arc length: Horii (1985)). However, the Japanese figure includes non-volcanic-arc terrane and does not include the slightly thermal springs that comprise 60% of the Cascade Range total. The length-normalized Cascade Range rate is much lower than those estimated for the Taupo volcanic zone (16 MW/km arc length: Hedenquist (1986)) or the mid-ocean ridge (10–120 MW/km ridge: Fisher (2003)).

The Quaternary volcanic output of the entire 1250-km length of the U.S. and Canadian Cascade Range amounts to ~6400 km^3 of eruptive products (Hildreth, 2007), or ~3 km^3/km arc length/Ma. This translates to a volcanic heat output of about 0.1 MW/km arc length, given a latent heat of crystallization for mafic products of 420 kJ/kg (Stakes and Taylor, 2003), a density of 2500 kg/m^3, a heat capacity of 1 kJ/(kg-K), and cooling from 1200 °C to an ambient temperature near 0 °C. Thus the current hydrothermal heat output is about 10 times larger than the average Quaternary volcanic heat output. The hydrothermal heat output may be comparable to the heat supplied by magmatic intrusion; heat-budget-based intrusion-to-extrusion ratios for the central Oregon Cascade Range are in the range of 1.5 to 10 (Ingebritsen et al., 1989, 1994). Volatile budgets for the Kamchatka–Kuril and Central America volcanic arcs imply comparable intrusion-to-extrusion ratios of about 7 (Taran, 2009).

Any comparison of volcanic and hydrothermal heat discharge for the entire length of the arc ignores the fundamental difference in hydrothermal heat discharge north and south of about 45°15′N. The hydrothermal circulation of ~1.7 MW/km arc length south of 45°15′N is sufficient to have a major, perhaps locally dominant influence on the shallow thermal structure of the volcanic arc, whereas the hydrothermal circulation of ~0.1 MW/km arc length north of 45°15′N must have much less effect on the overall thermal structure. We attribute the hydrothermal differences between the northern and southern part of the arc primarily to differences in the permeability structure. To the south, an extensive carapace of permeable volcanic products (Fig. 4) facilitates capture of regional heat flow by shallow aquifer systems that feed large, "slightly thermal" springs. This permeable carapace is either absent or more areally restricted in the northern part of the Cascade Range. South of about 45°15′ the tectonic and structural effect of Basin and Range impingement (Fig. 6) facilitates relatively permeable deep circulation systems that feed thermal springs. The limited thermal-spring discharge to the north likely reflects limited deep permeability that is a consequence of the weakly extensional to compressional tectonic regime.

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