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Time-variation of hydrothermal discharge at selected sites in the western United States: implications for monitoring

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Abstract

We compiled time series of hydrothermal discharge consisting of 3593 chloride- or heat-flux measurements from 24 sites in the Yellowstone region, the northern Oregon Cascades, Lassen Volcanic National Park and vicinity, and Long Valley, California. At all of these sites the hydrothermal phenomena are believed to be as yet unaffected by human activity, though much of the data collection was driven by mandates to collect environmental-baseline data in anticipation of geothermal development. The time series average 19 years in length and some of the Yellowstone sites have been monitored intermittently for over 30 years. Many sites show strong seasonality but few show clear long-term trends, and at most sites statistically significant decadal-scale trends are absent. Thus, the data provide robust estimates of advective heat flow ranging from ~130 MW in the north-central Oregon Cascades to ~6100 MW in the Yellowstone region, and also document Yellowstone hydrothermal chloride and arsenic fluxes of 1740 and 15–20 g/s, respectively. The discharge time series show little sensitivity to regional tectonic events such as earthquakes or inflation/deflation cycles. Most long-term monitoring to date has focused on high-chloride springs and low-temperature fumaroles. The relative stability of these features suggests that discharge measurements done as part of volcano-monitoring programs should focus instead on high-temperature fumaroles, which may be more immediately linked to the magmatic heat source. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: monitoring; time series; hot springs; fumaroles

1. Introduction

In this paper we review what is known about the natural time-variation in mass and heat discharge from selected hydrothermal systems in the western United States. Our focus on the United States in part reflects the geographic emphasis of the U.S. Geological Survey (USGS), but also the fact that much of the reliable data on time-variation of hydrothermal discharge derives from studies done in the western United States during the latter part of the 20th century. These data were collected for various purposes, including basic understanding of water–rock interaction (e.g. Fournier et al., 1975; Ingebritsen et al., 1994), environmental-baseline monitoring (e.g. Norton et al., 1989; Sorey and Colvard, 1994, 1997; Sorey et al., 1994; Friedman and Norton, 2000), volcano monitoring (e.g. Farrar et al., 1985; Sorey et al., 1998), and water-quality monitoring as part of the USGS National Assessment of Water Quality. Much of the data collection was driven by mandates to collect environmental-baseline data in anticipation of geothermal development. The data provide a quantitative basis for assessing the seasonal to multi-decadal variability of some types of hot-spring discharge. The

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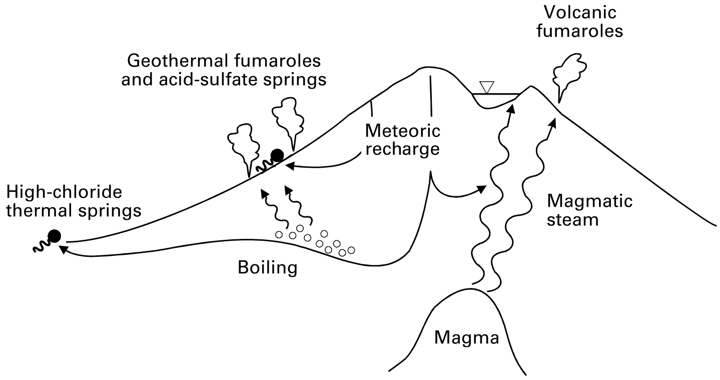


Fig. 1. Conceptual model of the three types of thermal-discharge features discussed in the text: high-chloride springs; geothermal fumaroles and acid-sulfate springs; and volcanic fumaroles.

findings are relevant to: (1) the design of environmental-baseline monitoring programs intended to protect geothermal features; (2) the value of hydrothermal-discharge monitoring as a component of comprehensive volcano-monitoring programs; and (3) heat-budget studies of areas with significant advective heat loss (e.g. Ingebritsen et al., 1989, 1994; Manga, 1998; James et al., 1999), which often assume that heat discharge from spring systems is constant.

Two electronic spreadsheets are an integral part of this report. These are accessible through a link labeled 'hydrothermal discharge in the western United States' under http://water.usgs.gov/nrp/proj.bib/ingebritsen.html. They include full details of all measurements from high-chloride spring (highclspringdat.PDF) and fumarolic areas (acid-sulfatedat.PDF), metadata with complete descriptions of the sites and methods, and basic time-series plots for each site (Skoustad et al., 1979; Farrar et al., 1989; Fishman and Friedman, 1989; Friedman et al., 1993; Howle and Farrar, 1996). We cite these spreadsheets in support of some particular points in the report; interested readers can use the spreadsheets to do their own complementary analyses. An index at the beginning of each spreadsheet facilitates cross-referencing with text, figures, and tables.

1.1. Scope

The scope of our discussion is mainly limited to the western United States and further limited to natural (non-anthropogenic) variations in mass and heat discharge. Geothermal development commonly leads to rapid and profound changes in thermal features — typically large and semi-permanent reductions in liquid discharge (Henley and Stewart, 1983; Turner, 1985; White, 1992; Glover and Hunt, 1996; Glover et al., 1996) and more temporary increases in steam discharge (Allis, 1981; Henley and Stewart, 1983; Ingebritsen and Sorey, 1985). Human impact on geothermal features in New Zealand and the western United States has recently been reviewed by Glover and Hunt (1998); Sorey (2000), respectively.

1.2. Types of thermal features

Three types of thermal features are commonly distinguished: high-chloride springs; geothermal fumaroles and associated acid-sulfate springs; and volcanic fumaroles (Fig. 1).

High-chloride springs emerge from liquid-dominated hydrothermal systems, generally have a near-neutral pH, and are commonly high in silica as well as chloride. Well-known examples in the western United States include the springs of Upper and Lower Geyser Basins, Yellowstone. Such features often occur in valleys near streams that eventually capture most of the thermal fluid, so their total discharge can often be gauged by measuring the solute flux in these adjacent streams. Discharge time-series from high-chloride spring systems in the western United States are relatively detailed and abundant.

Geothermal fumaroles (steam vents) and associated *acid-sulfate springs* are derived from steam up-flow generated by boiling of meteoric waters. Evolution of CO_2 and H_2S with the steam, and subsequent partial condensation and dissolution in shallow groundwater, leads to acid-sulfate springs with low pH, typically high dissolved CO_2, and variable sulfur (White et al., 1971). Well-known examples in the western United States include the thermal features of Lassen Volcanic National Park and Yellowstone's Mud Volcanoes. The total fumarolic and 'steam-heated' (acid-sulfate-spring) discharge is best measured by using total heat discharge as a proxy. Measurement of the multiple modes of heat discharge is time-consuming and difficult, and time series of fumarole and steam-heated discharge are sparse and rare, both in the United States and worldwide.

Volcanic fumaroles and associated springs occur where magmatic fluids reach, or nearly reach, the land surface. Whereas geothermal fumaroles are limited to $< \sim 160°C$ — the temperature obtained by adiabatic decompression of saturated steam of maximum enthalpy — volcanic fumaroles often reach temperatures of hundreds of degrees Celsius. The pH and composition of any associated springs depends upon the degree of interaction between the magmatic steam, meteoric water, and wallrock (Hedenquist, 1995; Reed, 1997). Along the western rim of the Pacific Ocean, high-temperature volcanic fumaroles are relatively common, and associated springs are often high in both chloride and sulfur. In the United States, volcanic fumaroles occur along the Aleutian arc, on the Alaska Peninsula, and in the Cascade Range, but they are relatively rare and weak, and associated springs are usually low in chloride. Many volcanic fumaroles are relatively inaccessible and their surroundings potentially dangerous. There are few reliable measurements of total mass or heat flux except in special circumstances where magmatic steam condenses into crater lakes, which also act as calorimeters (e.g. Brantley et al., 1993; Varekamp and Rowe, 2000).

The same monitoring methods apply to both geothermal and volcanic fumaroles. Therefore, for purposes of the discussion that follows, we will consider only two broad categories of hydrothermal features: high-chloride springs (highclspringdat.PDF) and fumaroles (acidsulfatedat.PDF). We will touch on the distinction between geothermal and volcanic fumaroles again in Section 4 that concludes this paper.

1.3. Historical background

The history of some developed hot springs in Europe indicates no dramatic changes in discharge, temperature, or chemical composition over millennia of casual human observation. Waring (1965), who published a global inventory of thermal springs, wrote that "(m)any hot springs have been described as remarkably uniform in temperature, flow, and mineral content," and noted that "(a)s most observations of the temperature and flow of thermal springs have been made at intervals of many years, no trends in their changes have been established."

In his monumental effort, Waring (1965) inventoried 1185 thermal-spring localities in the United States, mainly in California, Idaho, Nevada, Oregon, and Wyoming. Though he listed flow rates for most of the springs, a general lack of methodological detail leads us to regard the flow-rate data as non-quantitative. We will use Breitenbush Hot Springs, Oregon, as an example: for Breitenbush, Waring (1965) listed a flow rate equivalent to 60 l/s, and cited Langville et al. (1903) as a reference. This value, along with many others from Waring (1965), was repeated in USGS Circulars that assessed the geothermal resources of the United States in 1975 and 1978 (White and Williams, 1975; Muffler, 1978). However, Langville et al. (1903) listed no flow-rate data for Breitenbush; Waring (1965) provided no independent support for the 60 l/s value; and solute-inventory measurements made in the Breitenbush River in 1984–99 documented a flow rate of 12.5 ± 1.5 l/s ($n = 6$). The constancy of the data in the 1980s and 1990s, combined with the lack of supporting detail in Waring (1965), prompts us to discount the much higher reported value.

Most flow rates reported in the early literature were likely based on non-quantitative visual observation, sometimes combined with direct measurement of individual orifices. We place much more confidence in flow rates determined more recently by solute-inventory or heat-flux measurements. A few of the measurements reported by Day and Allen (1925) and Allen and Day (1935) in the early 20th century are sufficiently detailed to provide an earlier point of comparison.

2. Methods of measurement

In most cases high-chloride spring discharge is best measured by the solute- (chloride) inventory method pioneered in New Zealand (Ellis and Wilson, 1955) and fumarolic discharge is best measured by the heat-flux methods pioneered in New Zealand (Dawson, 1964; Dawson and Dickinson, 1970) and Japan (Yuhara, 1970; Sekioka and Yuhara, 1974), and later refined for use in the western US (Sorey and Colvard, 1994). Because many hydrothermal-discharge areas include numerous vents, some of which may be beneath streams or lakes or otherwise inaccessible, measurements of individual vents can rarely succeed in capturing the total discharge. Further, the flow from individual vents can vary greatly over time due to such superficial factors as human engineering, erosion and sedimentation, or near-surface water-rock reaction — none of which are of particular interest for purposes of volcano- or environmental-baseline monitoring. The total mass and heat discharge from a hydrothermal area is of greater interest, is less vulnerable to near-surface disruptions, and can often be measured reasonably well by using solute and/or heat fluxes as proxies.

2.1. Point measurements

Where it is concentrated in well-defined outflow channels, liquid discharge from individual vents can be calculated from velocity–area measurements done with a current meter, or by emplacing weirs or flumes. Steam discharge from well-defined and accessible vents may also be calculated from velocity–area measurements; in this case velocities are measured with an anemometer or a pitot tube. Such point measurements are often an important component of campaigns to measure the total discharge.

2.2. Solute-flux method to measure high-chloride spring discharge

Most high-chloride hot springs in non-arid regions occur near perennial streams that eventually capture most of the thermal fluid. Thus the total discharge from hot-spring areas can often be monitored on the basis of downstream increases in the solute loads of nearby streams. Chloride is the most commonly used indicator, because it behaves conservatively in solution and thermal waters are usually, though not always (e.g. Olmsted et al., 1997), much higher in chloride than nearby surface water and/or shallow groundwater. Other ions present in elevated concentrations in thermal waters (e.g. As, B, Na, SO_4) are sometimes used in solute inventories, but are much more likely to be affected by reactions in streams or the shallow subsurface.

The chloride-inventory method was first used to measure hot-spring discharge at Wairakei, New Zealand, prior to the geothermal development there (Ellis and Wilson, 1955). The discharge rate of a hot-spring group (Q_t) is calculated from the chloride concentration upstream (Cl_u) and downstream (Cl_d) of the hot springs, the chloride concentration in the thermal water (Cl_t), and the discharge rate of the stream (Q_s):

$$Q_t = [Q_s(Cl_d - Cl_u)]/[Cl_t - Cl_{bkgd}], \tag{1}$$

where Cl_{bkgd} is the 'background' chloride concentration upstream of any thermal source and assuming that $Q_t \ll Q_s$ and $Cl_t \gg Cl_u$. If other chemical species that are abundant in the thermal waters behave quasi conservatively, they will give similar results (Fig. 2).

In the western United States, downstream sample sites are often selected to be near USGS stream-gauging stations ($\pm 5\%$ error in Q_s). Away from established USGS gauging stations, Q_s is determined by standard wading- or bridge-measurement techniques ($\pm 10\%$ error in Q_s or better; e.g. Buchanan and Somers, 1969). Chloride values reported in the electronic spreadsheets were determined by a variety of methods, as described in the metadata for each site; for previously published data we generally report the same number of significant digits as were reported in the original reference. Ion chromatography methods that were widely employed beginning in about 1990 significantly increased the accuracy of low-chloride determinations. Error in the chloride determinations is an important issue with respect to some of the earlier (pre 1990) measurements in the Oregon Cascade Range, where both upstream and downstream chloride concentrations tend to be low. The low, constant 'background' chloride value of 0.7 mg/l that we invoke for Yellowstone sites is based on Norton and Friedman's (1991) non-thermal chloride estimate of ~0.5 mg/l from rock weathering and ~ 0.2 mg/l from precipitation.

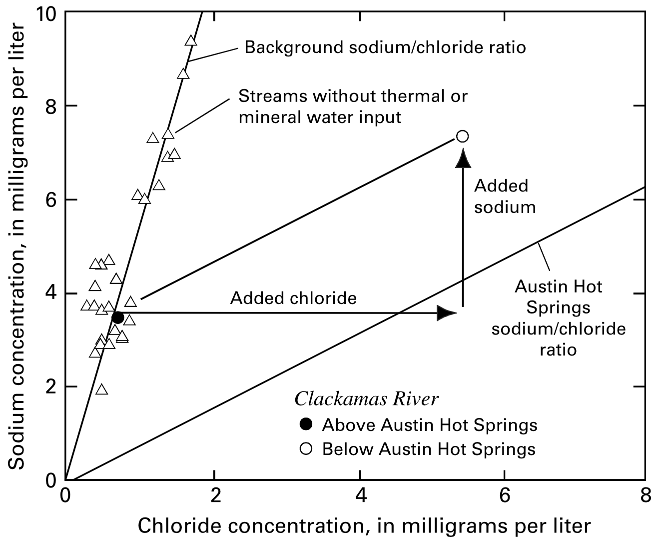


Fig. 2. Example showing how hot-spring discharge is calculated by solute-inventory methods. Here the discharge of Austin Hot Springs, Oregon (long. 122°00′30″, lat. 45°01′18″) on 8/15/85 was calculated based on the downstream increases in chloride and sodium and by using a two-component mixing model. In the mixing-model approach, the Na/Cl ratio of the thermal component was assumed to be that of the nearest hot spring, and the Na/Cl ratio of the non-thermal component was assumed to be 5.4 based on a linear least-squares fit to stream-chemistry data. Austin Hot Springs vents are located in and near the Clackamas River, which was flowing at 9400 l/s at the time of sampling. All three variants of the solute-inventory method gave $Q_t = 120$ l/s. After Ingebritsen et al., (1994).

2.3. Heat-flux method to measure fumarolic discharge

In areas of fumaroles and acid-sulfate (steam-heated) springs there are significant modes of discharge that cannot be captured by a simple solute inventory. The total heat flux must be measured and can be divided by steam enthalpy to arrive at a mass-discharge rate.

Significant heat loss from fumarolic areas occurs by direct discharge from fumaroles (H_{FUM}); by direct discharge from hot springs (H_{HS}) and lateral seepage in the subsurface (H_{LAT}); by evaporation, radiation, conduction, and molecular diffusion from water surfaces (H_{WS}); and by conduction, advection, and evaporation from warm or steaming ground (H_{GR}). Thus:

$$H_{TOT} = H_{FUM} + H_{HS} + H_{LAT} + H_{WS} + H_{GR}, \qquad (2)$$

where H_{TOT} is the total heat loss from the thermal area.

Heat advected from fumaroles (H_{FUM}) is the product of mass flow rate and steam enthalpy, and the mass flow rate in turn is the product of mean steam velocity, steam density, and vent area. For small vents, the velocity field is best measured with a pitot tube, whereas for large vents an anemometer can also be used (Sorey and Colvard, 1994). In many thermal areas only a small fraction of the fumaroles are safely accessible and have well-defined vents suitable for direct measurement. The mass discharge from the more numerous immeasurable features is sometimes estimated by visual comparison with a few measured vents.

Heat advected by direct discharge from individual hot springs (H_{HS}) can be measured by installing weirs or flumes. Discharge from many vents and lateral seepage in the subsurface (H_{LAT}) can be captured by solute-inventory methods or by measuring the temperature and flow rate of streams that pass through thermal areas. Large amounts of heat are also lost by evaporation, radiation, conduction, and molecular diffusion from free water surfaces (H_{WS}). Each of these processes is enhanced by increased water temperature; evaporative heat loss, in particular, is greatly enhanced by boiling (Dawson, 1964).

Empirical formulae developed in New Zealand (Dawson, 1964) relate the total heat loss by conduction, advection, and evaporation from warm or steaming ground (H_{GR}) to the ground temperature at 15 cm depth. These formulae have been applied in the

Table 1
Chloride-flux data from selected high-chloride hot-spring systems in the western United States. Italicized type indicates sites having long-term trends in excess chloride-flux that are statistically significant at the 5% level

Location	Date	Number of samples	Excess Cl flux, mean ± standard deviation (g/s)	Linear correlation between stream discharge and excess Cl flux, r^2
Yellowstone				
Yellowstone River, MT	1966–2000	521	567 ± 250	0.449
Gardner River, WY	*1966–1998*	*447*	*107 ± 19*	*0.021*
Hot River, WY	*1983–1994*	*154*	*84.4 ± 8.0*	*0.305*
Obsidian Creek, WY	*1988–1990*	*25*	*1.42 ± 0.57*	*0.274*
Madison River, WY	*1966–1999*	*390*	*744 ± 109*	*0.219*
Gibbon River, WY	*1966–1994*	*327*	*143 ± 32*	*0.264*
Firehole River, WY	*1966–1994*	*328*	*542 ± 62*	*0.204*
Snake River, WY	1982–1999	477	213 ± 104	0.690
Fall River, ID	1982–1999	499	175 ± 41	0.181
Boundary Creek, WY	1984–1995	66	37.7 ± 7.4	0.305
Warm River, ID	1984–1992	8	37.5 ± 14.7	0.962
Cascades				
Clackamas River, OR	1984–1999	6	42.0 ± 3.6	0.098
Breitenbush River, OR	1984–1999	6	14.8 ± 1.1	0.002
McKenzie River, OR	1984–1999	6	26.3 ± 1.2	0.432
Horse Creek, OR	1984–1999	6	5.6 ± 4.0	0.902
Warm Springs River, OR	1984–1999	6	12.1 ± 2.2	0.023
Mill Creek, CA	1983–1994	46	42.0 ± 5.2	0.105
Long Valley				
Hot Creek, CA	*1972–2000*	*163*	*51.7 ± 5.4*	*0.000*

western United States with minor modifications to account for the change in boiling point with elevation (Sorey and Colvard, 1994). Recent studies in areas of shallow steam condensation have successfully used both soil-temperature-gradient measurements and CO_2-flux measurements to determine heat loss from warm ground (Brombach, 2000; Severne, 2000).

Quantitative assessment of the error associated with measurement of each of these modes of heat loss is problematic. In general, we believe that the uncertainty associated with each mode is significantly larger than the uncertainty in chloride-flux determinations from high-chloride systems.

Comprehensive heat-loss studies done in fumarolic areas at Wairakei, New Zealand, Poas, Costa Rica, and Lassen Volcanic National Park, California indicate which modes of heat loss are likely to be most significant. These three sites represent a broad spectrum of hydrothermal systems: Wairakei (in the natural state) included both high-chloride springs and fumaroles; Lassen Volcanic National Park includes several areas of fumaroles with associated steam-heated, acid-sulfate-spring discharge; and at Poas, fumaroles discharge magmatic fluids in and adjacent to a very active crater lake. Though fumaroles are highly visible at each of these localities, they accounted for only approximately 3% of the 430 MW natural heat loss measured at Wairakei in the 1950s (Dawson and Dickinson, 1970); approximately 10% of the 115 ± 9 MW measured at Lassen, California in 1984–93 (Sorey and Colvard, 1994); and 5% of the 265 ± 100 MW measured at Poas, Costa Rica in 1988 (Brown et al., 1989). Heat loss from open water surfaces consistently emerged 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 was significant both at Wairakei (40%) and Lassen (17%).

3. Results

3.1. High-chloride springs

We compiled 3481 chloride-flux measurements of high-chloride hydrothermal discharge from a variety of publications and from USGS files. Some aspects of these data are summarized in Table 1, and the entire data set is contained in (highclspringdat.PDF). We chose to consider 18 sites in the western United States for which there are at least six measurements over a multiyear period (Fig. 3). Most of these (11), and most of the individual measurements (>3200), are from the Yellowstone region; the others are from the Cascade Range and Long Valley (Hot Creek), California. The 18 sets of measurements reflect hydrothermal discharge at a variety of scales: the four major rivers draining the Yellowstone region (Yellowstone, Madison, Snake, and Falls Rivers) each include thermal waters from scores to hundreds of hot springs, whereas many of the other measurement sites were designed to isolate a single hot-spring group along a few-km-long stream reach. The relatively comprehensive Yellowstone database owes mainly to the sustained efforts of Friedman and Norton (2000) and Norton and Friedman (1991), augmented by significant contributions from Fournier et al. (1975) in 1966–7 and Sorey et al. (1991) and Sorey and Colvard (1997) in the early 1990s.

The current data set allows us to evaluate temporal variability of hydrothermal discharge at a seasonal to multi-decadal scale. Many of the sampling sites are at USGS stream-gauging stations where stream discharge is (or was) measured at 15-min intervals for decades, and for which historical mean-daily discharge values are available on the internet at http://water.usgs.gov/nwis/sw. However, Cl-sampling frequency — generally weekly at best — is insufficient to clearly resolve shorter-term variations. At a few sites, and for relatively brief periods, specific conductance was recorded at 15-min intervals, and relations were established between Cl concentration and conductance (e.g. Farrar et al., 1985). However, the conductance data are relatively scarce and have larger uncertainties than the actual chloride-flux data.

For sites that capture discharge from vents with similar chloride (Cl) concentrations it would be straightforward to translate the excess (hydrothermal) Cl-flux defined by $[Q_s(Cl_d - Cl_u)]$ into an apparent hydrothermal discharge using the Cl content of nearby hot springs (Eq. (1)). However, a number of the sites capture discharge from vents with a wide range of Cl concentrations, rendering a conversion to discharge values difficult. Here we focus on the excess (hydrothermal) Cl flux as a proxy for hot-spring discharge.

3.1.1. Variance and seasonality

The excess Cl flux measured at most sites is fairly constant; at 13 of the 18 sites its standard deviation is <25% of the mean value (Fig. 4). The largest variability occurs at those sites where excess Cl flux is strongly correlated with stream discharge (Fig. 4) and there is a distinct seasonality to the excess Cl flux (see examples in Fig. 5). We believe that there is two distinct causes for the strong seasonality at a subset of the sites: (1) transient storage of hydrothermal Cl in lakes and (2) transient storage of hydrothermal Cl in the shallow subsurface. The former effect may be dominant at the Yellowstone (Fig. 5a) and Snake River sites (Fig. 5b). These sites show a much stronger correlation between stream discharge and excess Cl flux than the other Yellowstone sites (Table 1), and both are downstream from large lakes that store substantial amounts of hydrothermal Cl. As Fournier et al. (1975) pointed out, about 1% of the water in Yellowstone Lake is geothermal, and the Cl flux out of the lake varies directly with its level, having no relation to direct flow from the hot springs. At other sites that capture only local hot-spring discharge, and where much of the hot-spring discharge occurs at some distance from the gauged stream (e.g. Table 1: Horse Creek, Warm River), transient accumulation of geothermal Cl in the shallow subsurface, and its seasonal flushing, may control the correlation between excess Cl flux and stream discharge. At these sites, Cl may accumulate near the hot springs during low-stream-flow periods, and be flushed to the streams during wetter periods.

Sampling was sufficiently detailed to assess seasonal variability in excess Cl flux for only 10 of the 18 sites. Eight-year time series from a subset of these show clear seasonality at the Yellowstone (Fig. 5a) and Snake River (Fig. 5b) sites, where excess Cl flux is well-correlated with stream discharge; more subdued seasonality at the Firehole River (Fig. 5c), where there are no lakes for transient storage of Cl, and where Cl flux is affected by hydrothermal events such as the infrequent eruptions of large geysers; and no seasonality at Hot Creek, California (Fig. 5d).

The relation between excess Cl flux and stream discharge seems to be mildly hysteretic; there is a

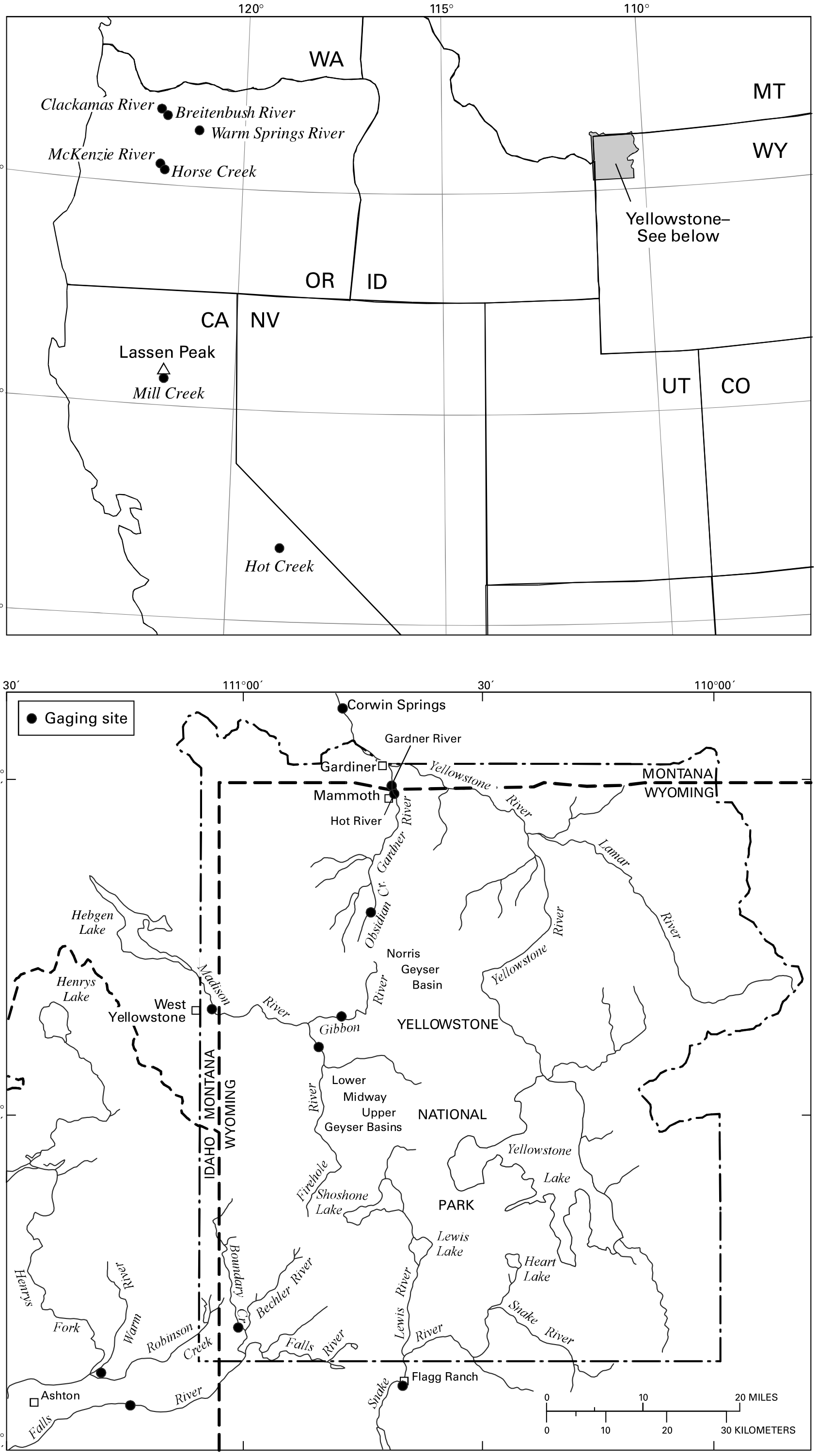


Fig. 3. Location of sampling sites in the western United States.

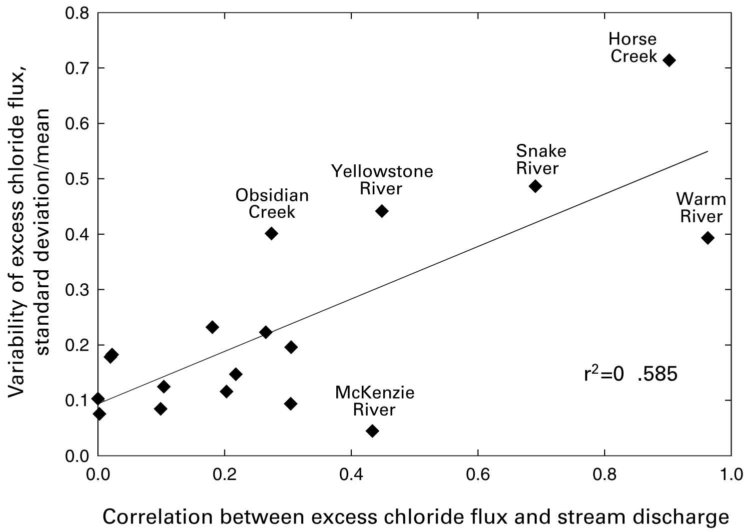


Fig. 4. Relation between variability in excess Cl flux (standard deviation/mean) and the strength of the correlation between excess Cl flux and stream discharge (r^2), showing that the largest variability tends to occur where Cl flux is well-correlated with stream discharge.

tendency for the Cl flux to be larger on the rising limb of the annual hydrograph than on the falling limb. The correlation between variability in excess Cl flux and variability in stream-flow is marginally significant ($r^2 = 0.151$, $n = 18$), and there is no correlation between the fraction of thermal water in the stream and variability of the Cl flux ($r^2 = 0.004$) (see highclspringdat.xcl under 'variance').

3.1.2. Long-term trends

The Cl-flux data set shows little evidence of decadal-scale trends in hydrothermal discharge. Records from only seven of the 18 sites show long-term trends in excess Cl flux that are statistically significant at the 5% level (Table 1), and some of these are problematic because of large data gaps (e.g. Fig. 6a and d). However, the declining trend in the Firehole River may be more than a statistical artifact, because regression-defined trends in the Firehole (−3.4 g/s Cl/yr), Gibbon (+0.8 g/s Cl/yr), and Madison River (−2.6 g/s Cl/yr) records are mutually consistent. (The Firehole and Gibbon rivers combine to form the Madison; see the lower panel of Fig. 3 for locations.) The other major rivers draining the Yellowstone system (Yellowstone, Snake, and Fall Rivers) show no statistically significant trends.

Other statistically significant trends observed in Yellowstone at Hot River (Fig. 6b) and Obsidian Creek (not shown; positive trend 1988–1990, $r^2 = 0.254$, $n = 25$) may represent relatively local and superficial variability. The Obsidian Creek site samples a very small amount of hydrothermal Cl (<2 g/s), and the strong trend observed at Hot River in 1983–1994 (−1.4 g/s Cl/yr) is not consistent with the less-pronounced trend in the Gardner River (−0.6 g/s Cl/yr) or the absence of any trend at the Yellowstone River site. (Both the Gardner and Yellowstone River sites eventually capture all discharge from Hot River; see the lower panel of Fig. 3 for locations.)

The relatively sparse ($n = 6$) long-term (1984–1999) records from the Oregon Cascade Range suggest some modest declines in excess Cl flux, most notably in the Clackamas River (Fig. 6c), which captures discharge from Austin Hot Springs, the largest hot spring in the Cascade Range. However, even the Clackamas River trend is marginally significant because of the small number of samples. Further, the Oregon Cascades trends depend mainly on the earliest (1984–5) measurements, which are subject to a large uncertainty due to the low Cl concentrations in the Cascades streams and the only marginally adequate analytical methods in use at the time (see details of analyses in highclspringdat.PDF). At the

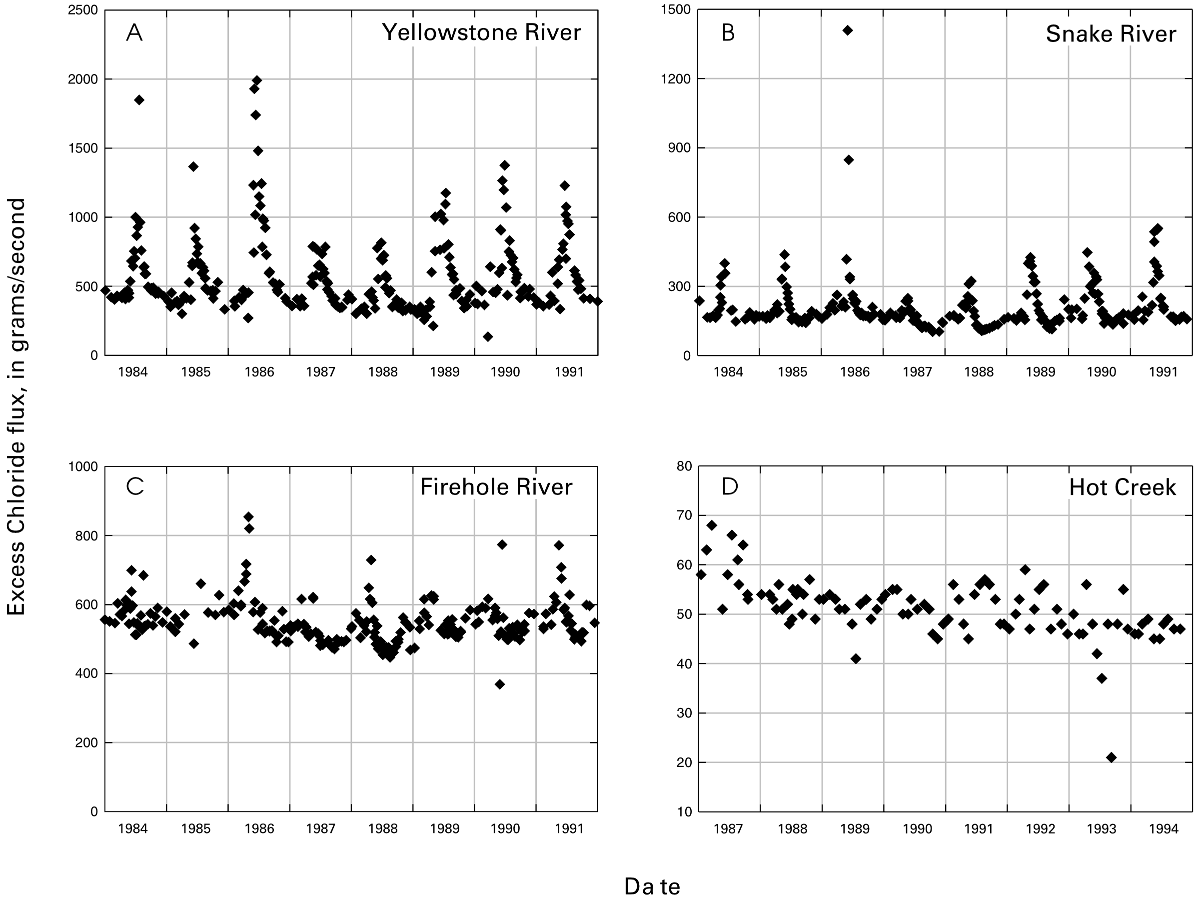


Fig. 5. Eight-year time series of excess Cl flux from sites exhibiting varying degrees of seasonality.

Oregon Cascades sites the difference in stream Cl concentrations above and below hot springs is typically only 0.3–5 mg/l, so that the improved analytical methods employed in the 1990s significantly decreased the uncertainty in excess Cl flux.

Trends at Hot Creek, California (Fig. 6d), are relevant to an ongoing local debate about whether nearby geothermal development will reduce the flow of thermal springs in Hot Creek Gorge, a popular recreation area. However, the Cl-flux observations to date are ambiguous. A long-term negative trend in excess Cl flux defined by all data collected since 1972 (1972–2000, $r^2 = 0.079$, $n = 163$) appears to predate the geothermal development, which began in 1985 and intensified in December 1990. Data collected after December 1990 actually show a statistically insignificant *positive* trend ($r^2 = 0.016$, $n = 87$). The excess Cl flux in Hot Creek Gorge appears to be affected by factors other than seasonality (Fig. 5d) and geothermal development.

3.1.3. Advective heat flow

One measure of advective heat transport by hot-spring systems is:

$$A = Q_{cl}\, c(T_g - T_r)/Cl_t, \quad (3)$$

where Q_{cl} is the excess Cl flux defined by $[Q_s(Cl_d - Cl_u)]$, c the heat capacity of the fluid, T_g the maximum fluid temperature at depth, as determined by chemical geothermometry or other means, and T_r is the temperature at the hot-spring recharge elevations. As thus defined, A is a measure of the heat advected away

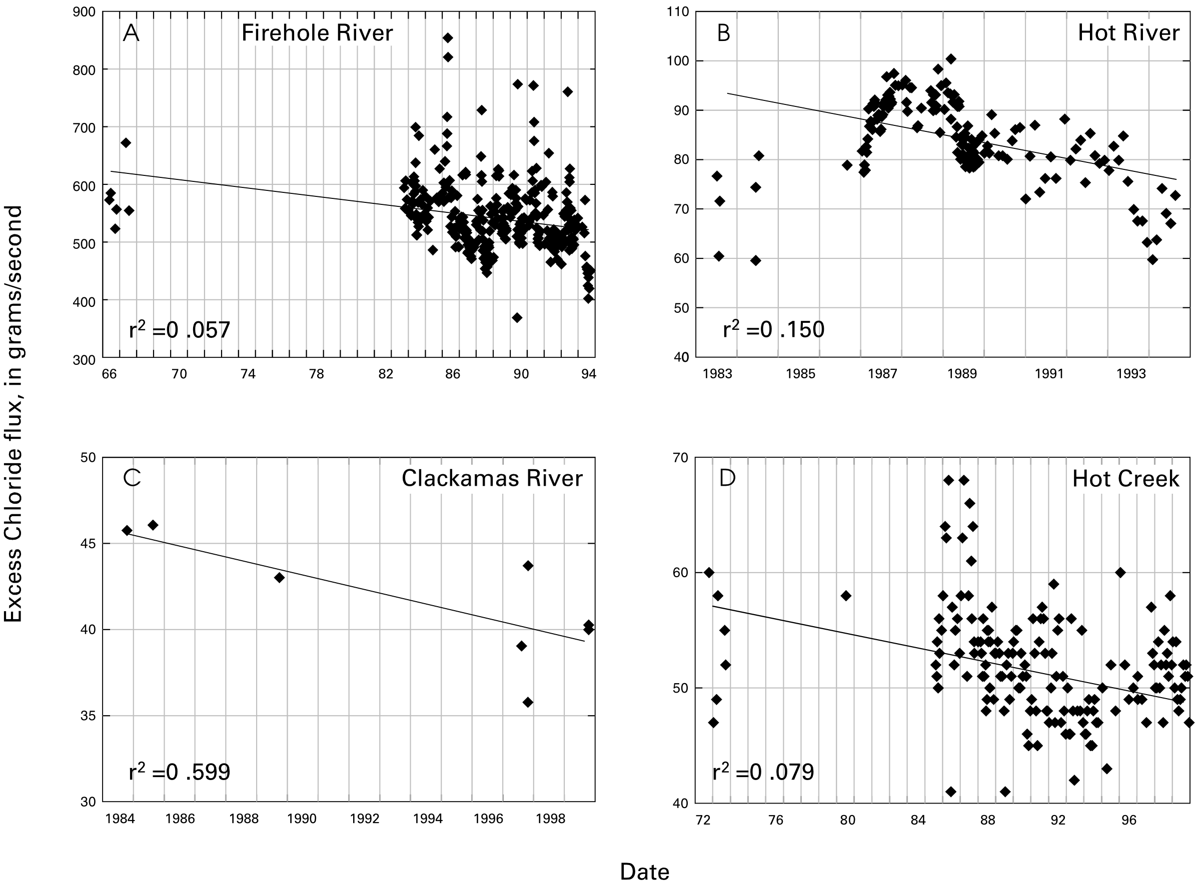


Fig. 6. Time series showing long-term trends in excess Cl flux.

from the deep heat source, rather than heat discharged at the hot springs; high-Cl hot-spring discharge temperatures (≤100°C) are often $\ll T_g$, due to conductive (or other) heat loss as the fluid moves toward the hot-spring orifices. We will assume that T_r is negligibly low except in the Oregon Cascades, where we use $T_r = 5$°C to facilitate comparison with the values reported by Ingebritsen et al. (1994). The assumption that $T_r \sim 0$ is reasonably good, because most of the systems considered here appear to be recharged mainly by high-elevation snowmelt.

The long-term average Cl-flux data provide robust estimates of the advective heat flow through some of the largest and most conspicuous hydrothermal systems in the western United States (Table 2). The advective heat-flow values from Yellowstone northeast of the Continental Divide, based on thousands of measurements from 1966–2000, are remarkably compatible with those reported by Fournier et al. (1975) based on only a few dozen measurements in 1966–67. Our value of 4.5 GW (Table 2: 263 + 4280 MW) northeast of the Divide is very close to the Fournier et al. value of 5.1 GW. We assumed a distinct Cl content and T_g value for waters originating at Mammoth Hot Springs (Table 2), whereas Fournier et al. (1975) assumed a single deep parent water, but this has only a minor effect on the total: if we apply their assumptions to the current data set we arrive at a similar value of 4.7 GW northeast of the Divide. The Fournier et al. (1975) value for the entire Yellowstone system (5.4 GW) is lower than that in Table 2 (6.1 GW)

Table 2
Mean advective heat flow through selected high-chloride hot-spring systems in the western United States. Values are averages for the periods of record indicated in Table 1. Subsurface temperatures are assumed constant

Location	Excess Cl flux (g/s)	Hot-spring Cl concentration (mg/l)	Subsurface temperature from geochemical indices (°C)	Advective heat flow (MW) (Eq. (3))
Gardner River, WY	107	170[d]	100[i]	263
Other Yellowstone NE of Continental Divide[a]	1204	400[e]	340[e]	4280
Yellowstone SW of Continental Divide[b]	426	400[e]	340[e]	1520
Yellowstone total	1737	various	various	6060
Oregon Cascades, lat. 44–45°15′ N	101	various[f]	various[f]	134
Lassen, CA[c]	42	2400[g]	240[j]	18
Hot Creek, CA	52	220[h]	180[k]	180

[a] Total of Yellowstone and Madison rivers minus contribution from Gardner River (Mammoth Hot Springs).
[b] Total of Snake, Fall, and Warm rivers.
[c] Mill Creek.
[d] Mammoth Hot Springs; Sorey and Colvard (1997).
[e] Cl concentration and temperature of Fournier's (1989) 'end-member deep thermal water'. There is substantial vapor discharge at Yellowstone, in addition to high-Cl liquid; by invoking this end-member (pre-boiling) temperature we are presumably accounting for the heat loss by vapor.
[f] Clackamas River (Austin Hot Springs) 390 mg/l Cl, 186°C; Breitenbush River (Breitenbush Hot Springs) 1200 mg/l Cl, 174°C; McKenzie River (Bigelow Hot Spring and Belknap Springs) 1200 mg/l Cl, 152°C; Horse Creek (Foley Springs) 1350 mg/l Cl, 100°C; Warm Springs River (Kahneeta Hot Spring) 240 mg/l Cl, 137°C (Ingebritsen et al., 1994).
[g] Morgan and Growler Hot Springs; Sorey et al. (1994).
[h] Farrar et al. (1985).
[i] Kharaka et al. (1991).
[j] Ingebritsen and Sorey (1985).
[k] Farrar et al. (1987).

because, lacking data southwest of the Continental Divide, they estimated too low a value of advective heat flow there (~0.25 GW vs. an actual value of ~1.5 GW).

Similarly, the advective heat-flow value reported here for a section of the Cascade Range volcanic arc in north-central Oregon between 44 and 45°15′N latitude (Table 2: 134 MW) is similar to the value of 140 MW reported by Ingebritsen et al. (1994) based on three sets of measurements in the mid-late 1980s. It should be noted that the high-Cl hot-spring contribution that Ingebritsen et al. (1994) reported for Horse Creek was erroneously high by about 3 MW owing to a diffuse upstream Cl source, since recognized, that contributes to the Cl-flux at the sampling site. The high-temperature advective heat flow in north-central Oregon, though high for the Cascades, is anomalously low compared to other volcanic arcs worldwide. In fact, the high-temperature advective heat flow between 44 and 45°15′N is comparable to the advective flux localized along a 1-km-long section of the Hot Creek Gorge in Long Valley caldera, California (Table 2: 180 MW) or in the vicinity of Lassen Volcanic National Park (~133 MW: 18 MW from high-Cl springs (Table 2) plus 115 MW from fumarolic areas (Sorey and Colvard, 1994)). Recent studies by Manga (1998) show that *low*-temperature advective heat transport is important in the Oregon Cascades.

Steady intrusion, crystallization, and cooling (800–300°C) of silicic magma at a rate of 1 km^3/Ma translates to a heat flux of only 0.06 MW (Ingebritsen and Sanford, 1998, pp. 181–184). Thus hydrothermal fluxes as large as those observed at Yellowstone, or even Lassen, imply emplacement and cooling of very large volumes of magma. A steady heat discharge of 100 MW corresponds to intrusion at a rate of 1700 km^3/Ma, and such volumes are roughly equivalent to the largest known silicic eruptive units (Hildreth, 1981). Though we see no clear trends over human time scales of 10^1–10^2 years, localized

Table 3
Heat-loss data from fumarolic areas in California. Sulphur Works, Little Hot Springs Valley, Bumpass Hell, and Devils Kitchen are 'steam-heated' geothermal areas in Lassen Volcanic National Park (see left-hand-side of conceptual model in Fig. 1). Mammoth Mountain fumarole is a low-temperature volcanic fumarole (right-hand side of Fig. 1). Values in parentheses in heat-loss column are percentages of total heat loss for that area under late-summer conditions; na denotes 'not applicable'

Location	Date	Number of samples	Heat loss, mean ± std. deviation (MW)	Linear correlation between stream discharge and heat loss, r^2
Lassen				
Sulphur Works				
Heat advected in West Sulphur Creek	1984–1993	16	4.1 ± 4.5 (~25%)	0.814
Little Hot Springs Valley				
Heat advected in East Sulphur Creek	1987–1994	9 (8)	6.9 ± 3.5 (40–50%)	0.102 (0.848)
Bumpass Hell				
Heat advected in Bumpass Creek	1983–1993	21	1.4 ± 1.8 (~5%)	0.98
Bumpass Hell				
Heat loss from five large 50–90°C pools	1986–1988	4	18.6 ± 2.4 (~70%)	na
Devils Kitchen				
Heat advected in Hot Springs Creek	1922–1996	15	13.4 ± 5.3 (~50%)	0.718[a]
Long Valley				
Mammoth Mountain fumarole				
Heat discharge from a single volcanic fumarole	1990–1998	20	0.0043 ± 0.0018	na

[a] Logarithmic correlation.

advective heat-discharge rates ≥100 MW are obviously likely to be transient over geologic time scales of 10^4–10^6 years.

3.1.4. Fluxes of chloride and arsenic

The data of Table 1 imply that fluxes of other solutes, such as arsenic (As), are potentially large enough to affect water quality, even when greatly diluted by non-thermal water and/or at large distances from the original geothermal source. For example, the As/Cl ratio in the Yellowstone thermal waters is ~0.01 (Ball et al., 1998), so that the mean geothermal Cl flux of 1740 g/s (Table 2) translates to an arsenic flux of about 15–20 g/s. This arsenic flux, if the arsenic were to behave conservatively, would need to be diluted by some 340 000 l/s of arsenic-free water to meet the current U.S. Environmental Protection Agency (EPA) drinking-water standard of 0.05 mg/l. The mean flow of the four major rivers draining the Yellowstone region is only approximately 180 000 l/s (see highclspringdat.PDF) and, although chemical and biological processes presumably remove substantial amounts of arsenic from solution, arsenic concentrations in Yellowstone-area streams commonly exceed drinking-water standards (e.g. Ball et al., 1998). The As/Cl ratio and arsenic flux at Lassen are relatively low (~0.004 and ~0.2 g/s, respectively), and arsenic loss in streams between the hot-spring orifices and nearby sampling sites is well-documented (Sorey et al., 1994). The As/Cl ratio in the Hot Creek Gorge, Long Valley, California is similarly low, approximately 0.005 (Farrar et al., 1985), but the ~0.3 g/s arsenic flux in Hot Creek poses significant challenges to the managers of the large Los Angeles–Owens River aqueduct (e.g. Eccles, 1976). The proposed (30 June 2000) new EPA drinking-water standard for arsenic of 0.005 mg/l, if enacted, would make the hydrothermal arsenic fluxes from all of these sites more problematic.

3.2. Fumaroles

Time series of mass and heat discharge from areas of fumaroles and acid-sulfate springs ('steam-heated areas') are much scarcer than time series of high-Cl hydrothermal discharge. We were able to find only six multi-year time series in the USGS archives: 5 from Lassen Volcanic National Park (LVNP) and one from Mammoth Mountain, Long Valley, California. Some aspects of these data are summarized in Table 3, and

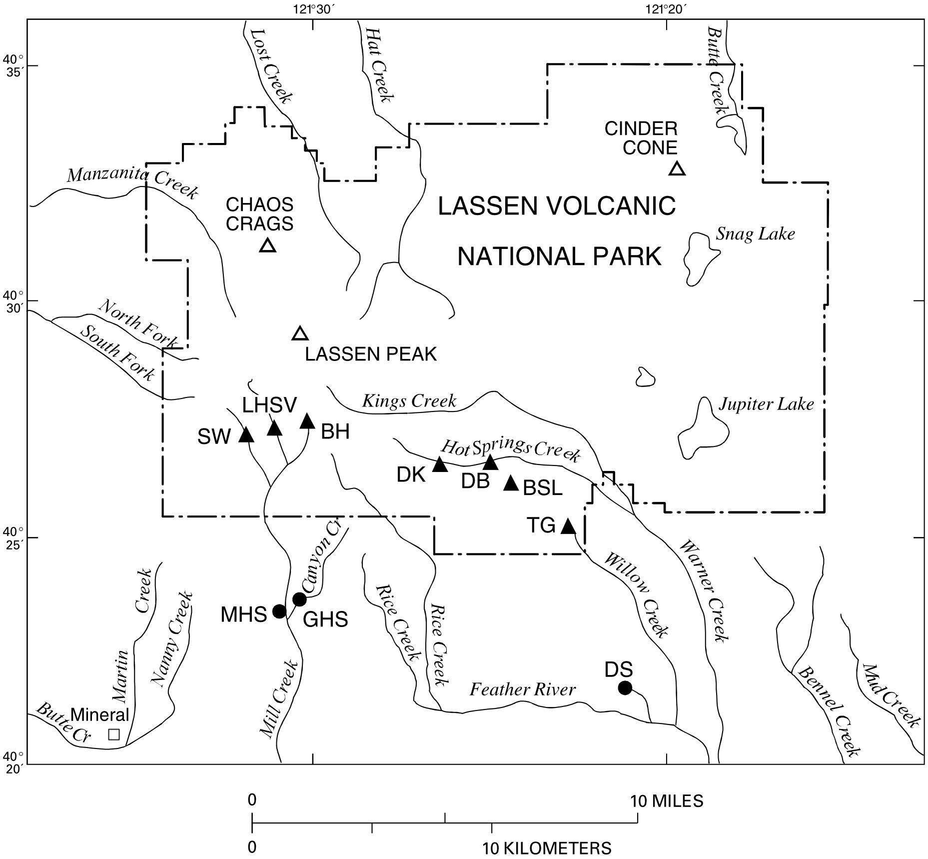


Fig. 7. Map showing areas of thermal-fluid discharge and major streams in the Lassen region. 'Steam-heated' areas with fumaroles, acid-sulfate springs, and/or low-chloride conductively heated springs are shown as triangles (SW, Sulphur Works; LHSV; BH, Bumpass Hell; DB, Drakesbad; DK; BSL, Boiling Springs Lake; TG, Terminal Geyser). Areas with high-Cl thermal-liquid discharge are shown as solid circles (MHS, Morgan Hot Springs; GHS, Growler Hot Spring; DS, Domingo Spring).

the entire data set in electronic form is contained in (acidsulfatedat.PDF).

The Lassen hydrothermal system includes steam-heated areas at relatively high elevations within LVNP and high-chloride hot springs at relatively low elevations approximately 10 km to the south (Fig. 7), with both sets of features connected to and fed by a single circulation system at depth (e.g. Ingebritsen and Sorey, 1985). The Lassen system is similar in this respect to many other high-temperature hydrothermal systems in regions of moderate to great topographic relief (as depicted schematically on the left-hand-side of Fig. 1). However, many young hydrothermal systems with great topographic relief, like the Mammoth Mountain system, lack high-chloride discharge.

At LVNP, all modes of heat loss at each steam-heated area (Eq. (2)) have been measured at least once under late-summer, low-stream flow conditions (Sorey and Colvard, 1994), and for some areas selected modes have been measured repeatedly. The most extensive time series are of heat advected by streams (corresponding to $H_{HS} + H_{LAT}$ in Eq. (2)), which is relatively easy to measure and, for some of the areas in LVNP, accounts for a substantial fraction of the total heat loss (Table 3). Heat losses from the steam-heated areas at Lassen can be converted to steam up-flow rates by assuming that the heat is originally provided by saturated steam of maximum enthalpy (~2800 kJ/kg; see, e.g. Muffler et al., 1982). At Mammoth Mountain, Sorey et al. (1998) made periodic measurements of temperature and mass flow rate in a single low-temperature (<90°C) volcanic fumarole from 1990 on (Table 3).

3.2.1. Variance and seasonality

The data from steam-heated areas at Lassen show heat loss in streams to be highly correlated with stream discharge (Table 3); the dominant control on stream discharge is the annual snowmelt. The physical basis for the heat loss-stream discharge correlation is that, during high-flow periods, a significant amount of the heat stored in the shallow subsurface, or in thermal pools, is advected into the streams. The Little Hot Springs Valley area is an apparent exception; however, the low correlation between heat loss and stream discharge ($r^2 = 0.102$) observed there is entirely due to a single anomalous measurement under high-flow conditions. This measurement was made on 6/15/94 when the reference stream reach — normally a 'gaining' reach that receives inflow from the underlying groundwater system — was instead losing substantial amounts of water to the groundwater system (see data in acidsulfatedat.PDF). Apparently, groundwater recharge by the stream either temporarily suppressed steam up-flow, or caused hydrothermal fluids to discharge farther downstream than under normal conditions. When this single measurement is omitted, the correlation between heat loss and stream discharge at Little Hot Springs Valley ($r^2 = 0.848$) is comparable to that observed at the other Lassen steam-heated areas (Table 3: range $r^2 = 0.718$–0.980).

The Mammoth Mountain fumarole also shows sensitivity to surficial hydrologic conditions: continuous monitoring of temperature in the Mammoth Mountain vent during 1991–96 revealed pronounced annual lows related to snowmelt infiltration, a behavior not captured by the relatively infrequent site visits for flow-rate measurements (Sorey et al., 1998, their Fig. 4). No correlation between temperature and mass flow rate was observed at the Mammoth vent, perhaps because it is a relatively weak low-temperature (~90°C) feature; in contrast, a strong correlation between temperature and discharge has been documented for high-temperature (350–550°C) fumaroles at Volcan Colima, Mexico (Connor et al., 1993). Mass and heat discharge from the Mammoth feature do show some sensitivity to volcanic unrest. Peak temperatures of about 90°C were measured in late 1989, during a 6-month period of earthquake swarms likely caused by dike emplacement in the upper crust (Sorey et al., 1998). Results of mass- and heat-discharge monitoring that began in mid 1990 suggest that discharge rates peaked nearly two years after the earthquakes, in mid 1991 (acidsulfatedat.PDF).

3.2.2. Long-term trends

Unlike the chloride-flux time series, the heat-flux time series from fumarolic areas are generally too short to document the presence or absence of decadal-scale trends. The single exception is Devils Kitchen, Lassen, 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 ($n = 1$) are available for comparison with measurements made in 1986–1996 ($n = 13$). On first inspection, the simple time series suggests a long-term declining trend, because two of the three largest values of heat loss were measured in the 1920s (acidsulfatedat.PDF). However, both of these measurements were made under high-flow conditions. Considered as a whole, the Devils Kitchen data show a regular and systematic relation between stream discharge and advective heat loss that encompasses the entire data set (Fig. 8), so we can conclude that there has been no obvious temporal trend in the heat advected in Hot Springs Creek during the 20th century. We lack data to assess the stability of other modes of heat loss from Devils Kitchen.

4. Discussion and summary

From numerous publications and the USGS archives we have been able to compile multi-year time-series consisting of 3481 measurements of high-Cl hydrothermal discharge at 18 sites (high-clspringdat.PDF) and 112 measurements of heat discharge at 6 sites in fumarolic areas (acidsulfatedat.PDF). These data are from selected systems in the western United States where the hydrothermal phenomena are believed to be as yet unaffected by human activity. Consideration of these data as a whole suggests strategies for identifying hydrother-

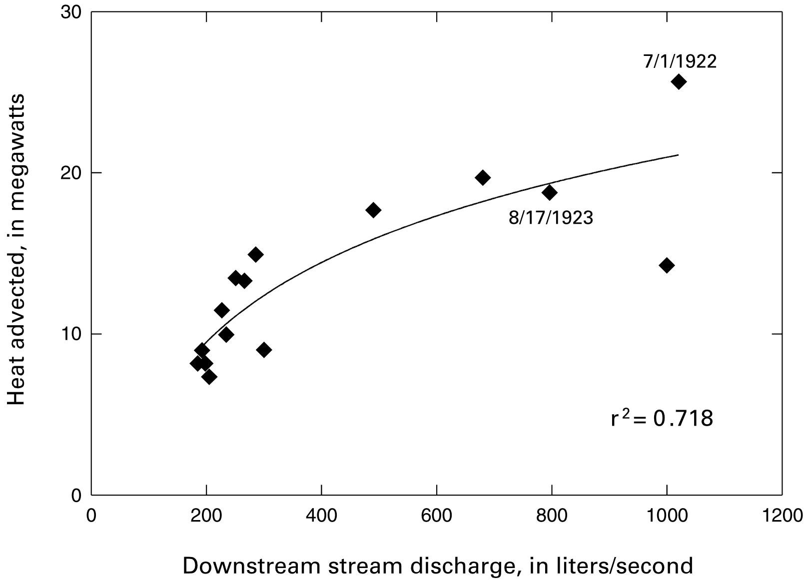


Fig. 8. Relation between heat advected from Devils Kitchen in Hot Springs Creek and discharge of Hot Springs Creek, Lassen Volcanic National Park, 1922–1996.

mal-discharge anomalies and for the design of environmental-baseline and volcano-monitoring programs.

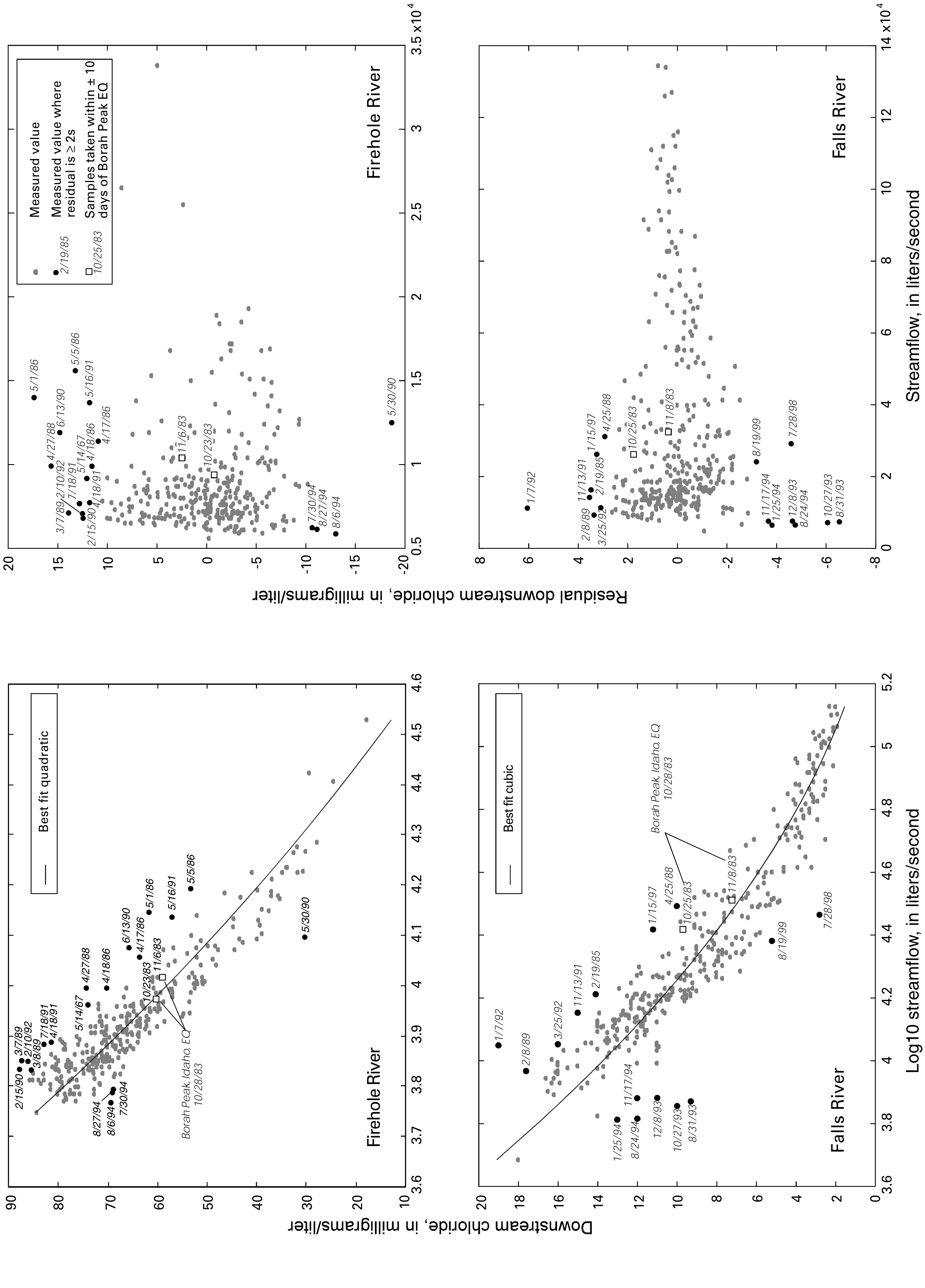
4.1. Detecting discharge anomalies

A major motivation for collecting time series of hydrothermal discharge is to develop the ability to detect changes due to stimuli such as tectonic events, geothermal-resource development, or volcanic unrest. Results from our compilation and analysis suggest that data sets collected across a wide range of hydrologic conditions will often be needed to diagnose anomalies caused by such stimuli, because both the excess chloride flux from high-chloride hydrothermal systems and the heat flux from fumarolic areas can be strongly influenced by local hydrology.

For instance, chloride flux time series for most of the major streams draining the Yellowstone region show pronounced seasonal trends related to the spring snowmelt (e.g. Fig. 5a–c). These seasonal hydrologic effects are difficult to remove by regression because they are highly variable in their timing and magnitude. An alternative is to examine the overall relation between Cl concentration and stream discharge (Fig. 9). Obvious outliers from concentration-discharge relations fitted to hundreds of data points may represent anomalously high or low hydrothermal discharge.

Observations from the Devils Kitchen fumarolic area at Lassen also demonstrate the importance of understanding the hydrologic context of hydrothermal-discharge measurements. The average of two heat-loss measurements made at Devils Kitchen in the 1920s is 23 MW, whereas 13 measurements in 1974–1996 average 12.0 ± 3.8 MW (acidsulfatedat.PDF). One might be tempted to conclude that heat loss from Devils Kitchen has declined since the 1920s, but the consistent relation between stream discharge and heat loss defined by the entire data set

Fig. 9. Relation between chloride concentration and stream discharge, Firehole River, Wyoming, and Falls River, Idaho. Left-hand panels show best polynomial fits to chloride vs. log streamflow ($r^2 = 0.84$ for Firehole River, 0.87 for Falls River). Right-hand panels show residual chloride concentrations for each site based on departures from the best-fit polynomial. Measurements for which the absolute value of the residual chloride concentration exceeds two standard deviations of the residuals population are labeled with measurement dates. Measurements immediately preceding and following the main shock of the Borah Peak, Idaho earthquake (10/28/1983, **M** 7.3) are also date-labeled.


Firehole River
Falls River
Best fit quadratic
Best fit cubic
Downstream chloride, in milligrams/liter
Log10 streamflow, in liters/second
Residual downstream chloride, in milligrams/liter
Streamflow, in liters/second
Measured value
Measured value where residual is ≥ 2s
Samples taken within ± 10 days of Borah Peak EQ
Borah Peak Idaho, EQ
10/28/83

(Fig. 8) indicates that the 1920s values were larger simply because they were measured during larger stream-flows.

4.2. Response to regional tectonic events

Some of the chloride flux time series are long enough, and complete enough, to inspect for the possible effects of regional tectonic events such as major earthquakes or uplift/subsidence cycles. One might expect to see significant responses to major earthquakes, in particular, because these have been observed to greatly increase the activity of individual hydrothermal vents (e.g. Sorey and Clark, 1981; Hutchinson, 1985). However, the Cl-flux records show no obvious sensitivity to major earthquakes. We speculate that the large effects sometimes observed at individual vents are due to near-surface permeability changes such as those documented after the Loma Prieta, California, earthquake (Rojstaczer and Wolf, 1992; Rojstaczer et al., 1995), whereas the Cl-flux time series, which integrate flow from many vents, are controlled by the relatively steady up-flow from deeper in the system.

As an example, let us consider the response of the Yellowstone hydrothermal system to the M7 + Borah Peak earthquake on October 28, 1983. Many individual hydrothermal features at Yellowstone became more active following the Borah Peak event (Hutchinson, 1985). Further, both Norton and Friedman (1985) and Fournier (1989), relying on less complete data sets, indicated that Cl flux anomalies in the major streams preceded the Borah Peak event, culminating shortly thereafter. However, the Cl fluxes measured within 11 days of the Borah Peak event do not appear anomalous in the context of the concentration-discharge relations defined by the entire Yellowstone data set (Fig. 9). In fact, the obvious outliers from the Yellowstone concentration-discharge relations occur on a wide variety of dates (Fig. 9), with little consistency in timing between sites. The lack of consistent timing between sites suggests that the outliers are more likely related to local hydrologic conditions or hydrothermal activity than to regional earthquakes.

The Yellowstone Cl flux time series also show no obvious correlation with the well-documented cycles of uplift/subsidence of the Yellowstone caldera, which can be roughly summarized as uplift from 1923–1984; subsidence from 1984–1995; and renewed uplift from 1995 on (e.g. Wicks et al., 1998). The relative steadiness of the Cl flux in the major Yellowstone streams since ~1966 argues that these cycles have had little effect on hydrothermal through-flow. Rates of volume change invoked to explain the uplift/subsidence range from 0.01 to 0.028 km^3/yr (Wicks et al., 1998). The hydrothermal flux is ~0.14 km^3/yr, so that if the volume change were to be expressed purely as a hydrothermal flux it would probably be detectable. The fact that correlative changes in the hydrothermal flux have not been detected argues that the uplift/subsidence is driven by deeper magmatic processes that do not immediately affect the near-surface hydrothermal flux.

4.3. The steadiness of advective heat discharge from high-chloride systems

The 24 hydrothermal-discharge time series that are the basis for this report range from 2 to 74 years in length and have a mean length of about 19 years. Nearly all of the records from fumarolic areas are too sparse and/or brief to evaluate for long-term trends. The more complete records from high-Cl systems show little evidence of long-term trends, with the exception of the Firehole/Madison River at Yellowstone and a few sites that capture very localized discharge.

The general absence of multi-decadal-scale variability seems sensible in the context of the longevity of the probable heat sources and the likely spatial and temporal scales of the high-Cl hydrothermal flow systems. The likely magmatic heat sources for most of the systems considered are large enough to retain significant heat for 10^4–10^6 years, likely flow-path lengths range from a few kilometers to tens of kilometers, and likely fluid travel times are on the order of 10^2–10^4 years. Continental hydrothermal systems differ in these (and other) respects from the more dynamic subsea hydrothermal systems associated with the mid-ocean ridge, which seem to exhibit much more short-term variability (e.g. Von Damm et al., 1997; Wilcock, 1997; Johnson et al., 2000). On the continents, similar dynamism in mass and heat flux seems to be observed only in fumarolic areas on recently active volcanoes (e.g. Christenson, 2000; Martinez et al., 2000; Shevenell and Goff, 2000) and in the short-lived, moderately high-Cl hydrothermal systems associated with recent eruptions, such as Loowit Hot Springs at Mount Saint Helens (Shevenell and Goff, 1995) and the mid-valley thermal springs in the Valley of Ten Thousand Smokes (Lowell and Keith, 1991).

4.4. Implications for monitoring

Monitoring of hydrothermal discharge is a key element of resource assessment prior to geothermal development and is often a component of volcano-monitoring programs. We will continue to focus on mass and heat discharge, rather than chemical characteristics, but it is important to recognize that chemical indices can be more responsive than mass and heat discharge. There is repeated evidence for chemical changes during volcanic unrest (e.g. de la Cruz-Reyna et al., 1989; Giggenbach et al., 1990; Lopez and Williams, 1993; Tedesco, 1994; Christenson, 2000; Martinez et al., 2000; Shevenell and Goff, 2000). However, most geochemical monitoring requires repeated visits to the source and fairly sophisticated analyses. Mass and heat discharge — or proxies thereof such as Cl in outflow streams — can sometimes be monitored remotely, using relatively simple equipment.

4.4.1. Environmental-baseline monitoring

Environmental-baseline monitoring typically focuses on high-Cl hot springs, which are valued for recreation and, in some cases, are culturally significant to indigenous peoples. Many high-Cl springs worldwide have diminished or ceased to flow as a result of nearby geothermal development (e.g. Henley and Stewart, 1983; Turner, 1985; White, 1992; Glover and Hunt, 1996; Glover et al., 1996).

The data compiled for this report demonstrate the importance of multi-year, year-round sampling to establish the mean flow of high-Cl spring systems. For some of the systems considered, several years of regular Cl-flux measurement were needed to establish a stable mean and standard deviation. If year-round sampling is not feasible, an alternative strategy is to sample repeatedly under similar stream-flow conditions. Experience at Yellowstone, where mean Cl-flux values from a few dozen samples collected in 1966–67 are consistent with the results from thousands of samples collected since, indicates that this strategy can be successful.

4.4.2. Volcano monitoring

High-Cl-spring discharge is relatively unlikely to show short-term response to volcanic unrest. Especially in terrain with substantial topographic relief, high-Cl hot springs are commonly located far from their likely magmatic heat sources (Fig. 1) —about 10 km lateral distance at Lassen (Fig. 7), for example, and up to 30 km in the Oregon Cascades (Ingebritsen et al., 1994), in addition to a several-kilometer vertical distance. Transmission of pressure and thermal pulses over such distances is likely to be too slow and damped to allow changes to be observed during periods of volcanic unrest. Thus mass- and heat-discharge measurements done as part of volcano-monitoring programs should generally focus on fumarolic areas. As we have shown, time series of *total* heat and mass discharge from fumarolic areas — whether geothermal or volcanic — are essentially nonexistent, because of the difficulty of repeatedly measuring the multiple modes of heat loss (Eq. (2)). At a few localities worldwide, total fumarolic mass and heat fluxes from active volcanoes have been inferred from changes in the composition, volume, and temperature of volcanic crater lakes, which act as condensors and calorimeters (Brantley et al., 1993). In such cases a clear link between hydrothermal discharge and volcanic processes often emerges (Rowe et al., 1992a, b; Christenson and Wood, 1993; Agustdottir and Brantley, 1994; Ohba et al., 1994; Badrudin, 1994). However, because access is often problematic, and the necessary measurements are difficult and time-consuming, volcanic crater-lake measurements have been infrequent and semi-quantitative.

In fact, for purposes of routine monitoring, the difficulty of measuring total heat flow in fumarolic areas is nearly prohibitive, so we advocate: (1) doing a one-time survey of all modes of heat loss, and (2) selecting a readily measured principal mode for (nearly) continuous monitoring. Strong candidates for continuous monitoring include advective heat loss via streams and heat loss from pools, because these modes commonly comprise a large fraction of the total heat loss. Monitoring heat loss in streams requires continuous measurement of upstream and downstream temperatures and downstream stream stage, along with sufficient direct stream flow measurements to define a relation between stage and stream discharge. Monitoring heat loss from pools requires continuous measurement of pool temperature, stage, and meteorological data, along with detailed knowledge of the bathymetry and topography. Temperature, stage, and meteorological parameters can all be recorded by relatively robust on-site instruments and monitored remotely through satellite telemetry.

Though fumarolic discharge is rarely a dominant mode of heat loss, continuous monitoring of individual fumaroles also holds promise. Connor et al. (1993) demonstrated the importance of frequent, automated sampling, and pointed out that, although thermocouple measurements provide an indirect measure of mass flow, they are inexpensive and resilient. At Volcan Colima, Mexico, Connor et al. (1993) monitored temperatures in five individual high-temperature (350–550°C) volcanic fumaroles at 20-min intervals for one year (May 1991–May 1992), and used a numerical model to relate fumarole temperature to mass flow rate. They found temperature and mass flow rate to be strongly influenced by diurnal variations in barometric pressure, particularly for the cooler (lower flow rate) fumaroles, but also reported a 'broad correlation between volcanic activity and fumarole temperatures'.

Most long-term discharge monitoring to date has focused on high-Cl springs and relatively low-temperature fumarolic areas. The relative stability of these features, as documented in this report and the complementary electronic datasets, suggests that discharge measurements done as part of volcano-monitoring programs should focus instead on high-temperature fumarolic areas, which are more responsive to volcanic unrest.

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