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Ongoing hydrothermal heat loss from the 1912 ash-flow sheet, Valley of Ten Thousand Smokes, Alaska

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Abstract

The June 1912 eruption of Novarupta filled nearby glacial valleys on the Alaska Peninsula with ash-flow tuff (ignimbrite), and post-eruption observations of thousands of steaming fumaroles led to the name 'Valley of Ten Thousand Smokes' (VTTS). By the late 1980s most fumarolic activity had ceased, but the discovery of thermal springs in mid-valley in 1987 suggested continued cooling of the ash-flow sheet. Data collected at the mid-valley springs between 1987 and 2001 show a statistically significant correlation between maximum observed chloride (Cl) concentration and temperature. These data also show a statistically significant decline in the maximum Cl concentration. The observed variation in stream chemistry across the sheet strongly implies that most solutes, including Cl, originate within the area of the VTTS occupied by the 1912 deposits. Numerous measurements of Cl flux in the Ukak River just below the ash-flow sheet suggest an ongoing heat loss of ~250 MW. This represents one of the largest hydrothermal heat discharges in North America. Other hydrothermal discharges of comparable magnitude are related to heat obtained from silicic magma bodies at depth, and are quasi-steady on a multidecadal time scale. However, the VTTS hydrothermal flux is not obviously related to a magma body and is clearly declining. Available data provide reasonable boundary and initial conditions for simple transient modeling. Both an analytical, conduction-only model and a numerical model predict large rates of heat loss from the sheet 90 years after deposition.
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Keywords: Valley of Ten Thousand Smokes; heat loss; ash-flow

1. Time-history of hydrothermal phenomena

The Aleutian Range, southwestern Alaska, is a highly active 2600-km-long volcanic chain resulting from rapid (~7 cm/yr) subduction of the Pacific plate beneath the American plate (Kienle and Swanson, 1983). On June 6–8, 1912, a new vent on the Alaska Peninsula, Novarupta (Fig. 1), released the largest eruption of the 20th century, and the largest rhyolitic eruption of the last millennium. The 1912 eruption from Novarupta apparently drew magma from beneath Mount Katmai, 10 km to the east (Fig. 1), causing it to

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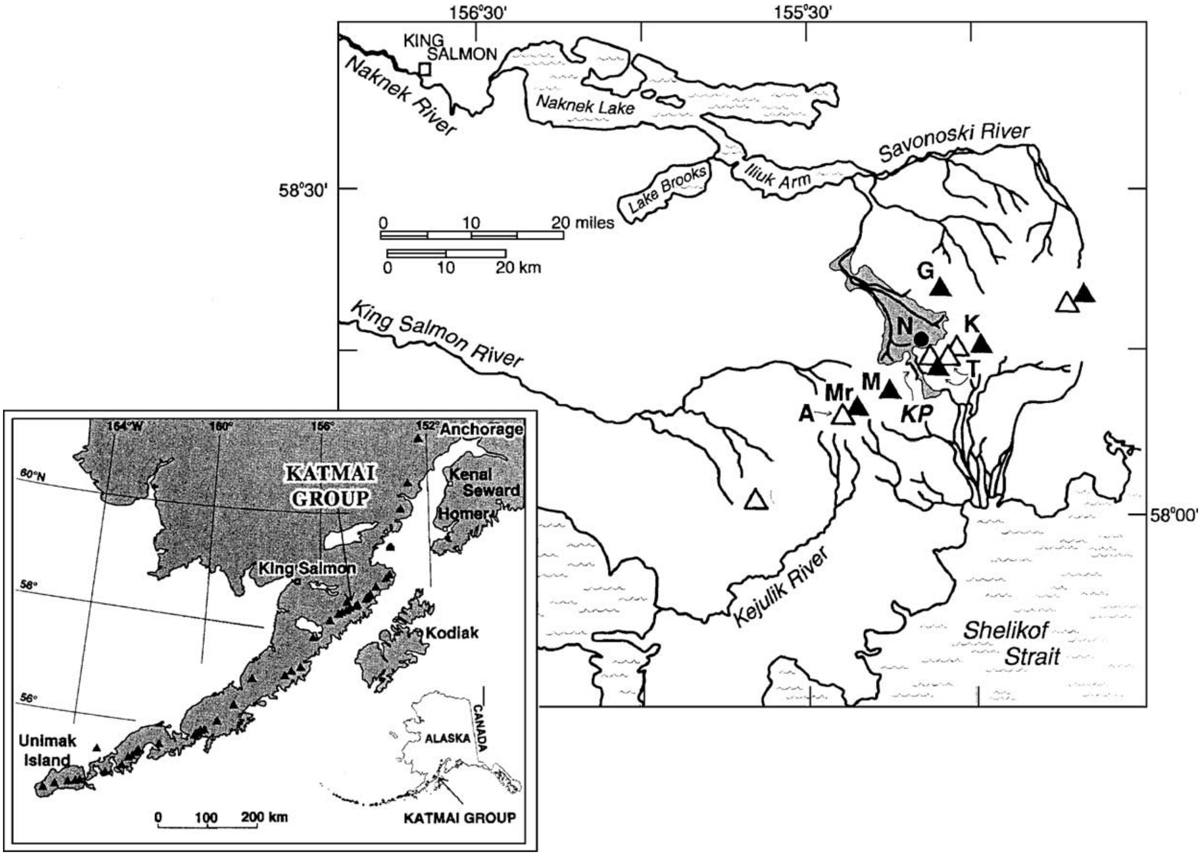

Fig. 1. Location maps showing part of the Aleutian Range volcanic chain. The smaller inset map shows the position of the Katmai cluster along the chain of Alaska Peninsula arc volcanoes. Solid triangles in the larger map indicate cones active during the Holocene; open triangles indicate Pleistocene cones. The Katmai cluster includes Alagogshak (A), Martin (Mr), Mageik (M), Griggs (G), Trident (T; three extinct cones and one recently active cone), and Mount Katmai (K). The solid circle (N) indicates Novarupta and KP indicates Katmai Pass. The gray-shaded area northwest of Novarupta is the 1912 ash-flow sheet. From Hildreth and Fierstein (2000).

collapse and creating a caldera 3 km in diameter and 1 km deep (Curtis, 1968; Hildreth and Fierstein, 2000). During 60 h of activity, approximately 13 km^3 of magma erupted from Novarupta, inflating to approximately 30 km^3 of tephra (Fierstein and Hildreth, 1992). Roughly half of the tephra filled nearby glacial valleys as ash flow deposits (Figs. 1 and 2), whereas the other half occurs as widespread air-fall deposits (Curtis, 1968; Fierstein and Hildreth, 1992).

The 1912 eruption had devastating consequences on local Native American communities, released fine atmospheric and stratospheric ash that caused haze as far away as North Africa, and influenced global climate (Griggs, 1922). The ash-filled proximal valleys were named ‘Valley of Ten Thousand Smokes’ (VTTS) in 1916, when members of a National Geographic Society party led by Robert F. Griggs crossed Katmai Pass (Figs. 1 and 2) from the southeast and saw the “most amazing vision ever beheld by mortal eye . . . The whole valley . . . was full of hundreds, no thousands – literally tens of thousands – of smokes, curling up from its fissured floor” (Griggs, 1922).

1.1. Fumaroles

The hot ash from Novarupta covered streams, snow, and marshes, creating secondary phreatic explosion craters and fumaroles. Snow and glaciers surrounding the VTTS melted due to the heat and

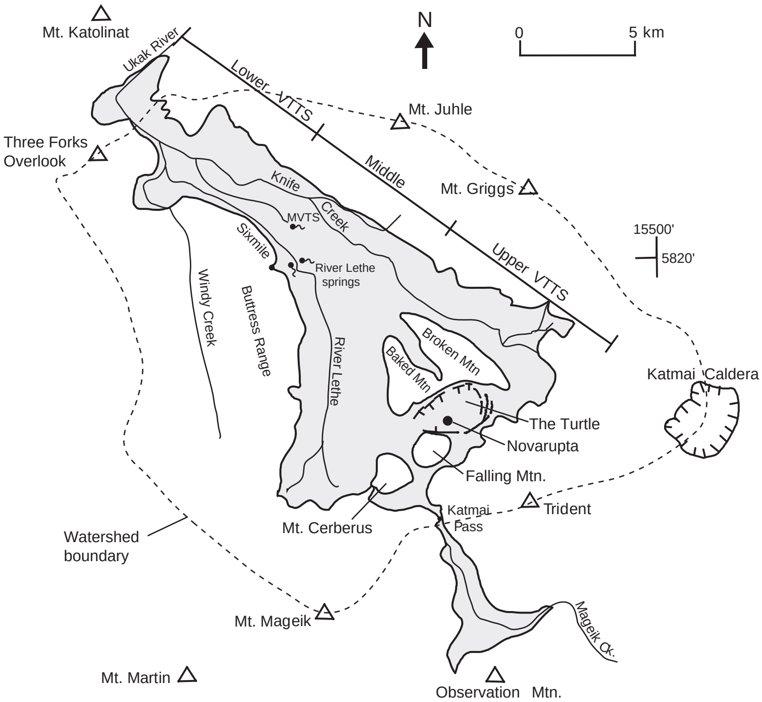

Fig. 2. Map showing the 1912 ash-flow sheet (gray) after Hildreth (1983); the mid-valley thermal springs (MVTS); the upper, middle, and lower VTTS regions; and other selected geographic features.

air-fall tephra, causing additional cold surface water to enter the hot deposits. Fumarolic activity was especially concentrated where large streams from the surrounding ranges enter the VTTS (for example around Sixmile Camp; Fig. 2) and above pre-1912 stream channels within the VTTS. By 1919, fumarolic activity had decreased but still constituted the dominant mode of heat loss from the ash sheet. Several hundreds of high-temperature fumaroles remained, though many fumaroles in the lower part of the valley had cooled completely and some had transformed to thermal springs. In the upper valley a maximum temperature of 645 °C was measured (Allen and Zies, 1923; Zies, 1924). The vapor emitted by the fumaroles was almost entirely (99.9%) water, and on a calm day maximum plume heights of 600 m were estimated (Zies, 1924). Calculations done in 1919 indicated that only 1/11 of all water entering the 1912 deposits exited as stream flow in the lower valley (Allen and Zies, 1923). The rest of the water was vaporized and discharged by fumaroles. In 1950, approximately 100 fumaroles were still active along fissures near Baked Mountain and Broken Mountain (Luntey, 1954). By the late 1980s, fumaroles in the Novarupta vent region still discharged at approximately 90 °C, but the only other active fumaroles in the VTTS were along fissures at the southwestern side of Baked Mountain, weakly discharging at ~30 °C (Keith, 1991). Due in part to conductive cooling of the ash-flow sheet, but mainly to advective cooling by the abundant inflow of precipitation and surface water, the sheet in the VTTS cooled rapidly.

1.2. Thermal springs

By the late 20th century fumarolic heat loss had become relatively insignificant, but in 1987 T.E.C. Keith discovered thermal springs issuing from the walls of a 15-m-deep gorge in the mid-valley (Fig. 2, MVTS). These mid-valley thermal springs (MVTS) discharged at approximately 20–30 °C over a 300-m reach of the gorge, becoming progressively warmer downstream and eventually forming the southwest fork of Knife Creek (Fig. 2; Keith et al., 1992). Chemical geothermometers indicate source temperatures of 30–60 °C for the MVTS, and their oxygen and deuterium isotopic data plot close to the World Meteoric Water Line (Keith et al., 1992).

The geomorphic setting of the MVTS was similar when revisited in 1991 and 1993, but by early June of 2001 the gorge had filled with at least 1 m of sand and pumice, leaving only 7 thermal springs issuing at the floor of gorge, partly hidden behind large blocks of snow. In 2001 the flow measured in the creek draining the MVTS was much reduced (Table 1) and, further down the gorge, the thermal waters percolated into the gorge fill rather than flowing on the surface into Knife Creek. Tritium contents of 6.7–6.9 TU in MVTS samples obtained in 2001 suggest fluid residence times, or recharge ages, of <20 years (Hogeweg, 2002).

Springs with a thermal component based on chemistry (<8.8 °C) also currently issue from vertical joints in the ash-flow sheet 1–2 m above the River Lethe (Fig. 2; Keith et al., 1992). Relatively high-temperature thermal springs (up to 80 °C at Three Forks) once occurred along the lower third of the River Lethe (Zies, 1924). These springs evolved from higher-temperature fumaroles as cooling of the ash-flow sheet progressed and liquid water accessed the fumaroles.

On the basis of observations in 1987–1990 (Table 1), Lowell and Keith (1991) concluded that there was seasonal fluctuation of the temperatures in the MVTS and that there was no correlation

Table 1
Maximum temperatures, maximum chloride concentrations, and discharge values measured at the MVTS on various dates (Lowell and Keith, 1991; Keith et al., 1992; and previously unpublished USGS data)

Date	Maximum temperature (°C)	Maximum Cl concentration (mg/L)	Discharge (L/s)
24-Jul-1987	17	74	–
23-Aug-1989	17.8	71	–
4-Jun-1990	29.2	90	–
15-Mar-1991	27	79	–
7-Jun-1991	26	81	590
29-Jul-1993	22	66	590
10-Jun-2001	20.6	55	30–90

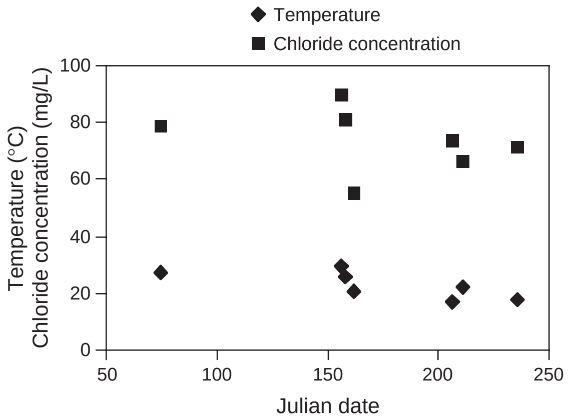

Fig. 3. Maximum observed temperatures (squares) and chloride concentrations (diamonds) from MVTS vents (Table 1) as a function of Julian date (day of the year; where January 1 is day one).

between the maximum observed chloride concentration and the maximum observed temperature. However, the extended (1987–2001) data set shows no clear seasonal pattern (Fig. 3) and does indicate a significant correlation between chloride concentration and temperature (Fig. 4). There is no temporal trend in the maximum observed temperature but there is a statistically significant decline in the maximum observed chloride concentration (Fig. 5). If chloride is an indicator of advective heat flux, as discussed in the following section, this trend suggests the possibility of a declining heat flux. However, the declining trend depends largely on data collected in 2001. The geomorphic changes observed at the site in 2001 suggest snowmelt mixing, perhaps even within the MVTS orifices, as

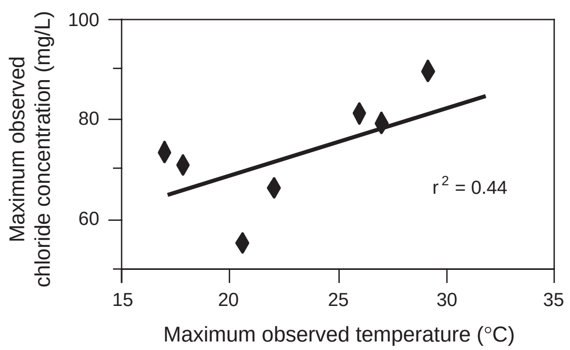

Fig. 4. Maximum observed chloride concentration as a function of maximum observed temperature from MVTS vents (Table 1).

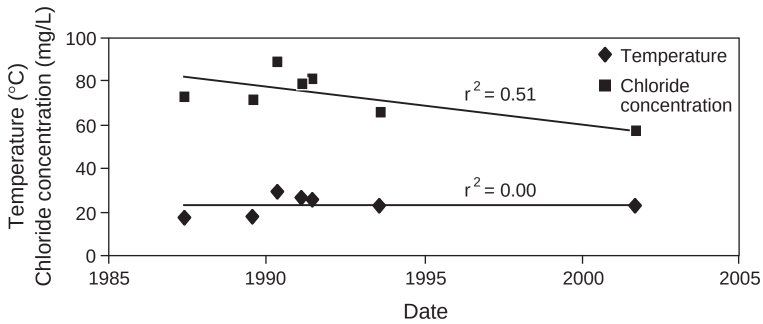

Fig. 5. Maximum observed chloride concentrations (+1 mg/L) and temperatures (+1 °C) from MVTS vents (Table 1) as a function of time, 1987–2001.

an alternative explanation of the decreased chloride concentrations.

2. Chloride-flux measurements

The chloride-inventory or chloride-flux method is often used to estimate hydrothermal discharge in igneous terrane (e.g. Ellis and Wilson, 1955; Ingebritsen et al., 2001). This method is predicated on the facts that thermal waters are usually higher in chloride (and many other ions) than nonthermal waters, and that chloride behaves conservatively in solution. Further, most liquid-phase hydrothermal discharge occurs near streams that eventually capture most of the hydrothermal fluid, so that

$$Q_t = [Q_s(Cl_d - Cl_u)]/[Cl_t - Cl_u] \quad (1)$$

where Q_t is the hydrothermal discharge (L/s), Q_s is stream discharge (L/s), Cl_d and Cl_u are the chloride concentrations downstream and upstream of the hydrothermal discharge (mg/L), Cl_t is the chloride concentration of the hydrothermal discharge (mg/L), and assuming $Q_t \ll Q_s$.

Two major streams, Knife Creek and the River Lethe (Fig. 2), flow through 20 to 30-m-deep gorges incised into the 1912 ash-flow deposit. Knife Creek has been observed to flow as early as mid-March, and its lower reaches may flow year round due to the hydrothermal input exemplified by the MVTS. River Lethe is dry and snow-filled most of the winter and, in the upper valley, as late as the end of June.

Streams in the VTTS originate mainly from glaciers in the upper parts of the surrounding watershed. Near the upstream edge of the 1912 deposits all streams are relatively low in chloride (1.2–5.1 mg/L) and other dissolved constituents, whereas in the lower valley all streams except Windy Creek (Fig. 2) are much higher in chloride (17–57 mg/L) and other constituents (Keith et al., 1992). Windy Creek has negligible contact with the ash-flow sheet because it has completely cut through the thin distal ash-flow tuff in the northwestern part of the VTTS and now flows mainly through the shallow marine, Upper Jurassic Naknek Formation and pre-1912 fluvial, glacial and volcanic deposits. The variation in stream chemistry across the sheet strongly implies that most solutes originate within the area of the VTTS occupied by the 1912 deposits (Keith et al., 1992). Chloride is abundant in the ash-flow deposits: up to 1500 ppm by weight can be found in rhyolite pumice, 2200 ppm in dacite pumice, and 1000 ppm in andesitic pumice (Lowenstern, 1993), which together make up most of the zoned deposit. Addition of chloride to the system by volcanic degassing is not likely, because the ash sheet has cooled substantially and is not connected with an actively degassing magma chamber (Hildreth and Fierstein, 2000). It is therefore reasonable to use chloride as an indicator for thermal waters leaching the 1912 deposits.

Knife Creek, River Lethe, and Windy Creek join at Three Forks to form the Ukak River (Fig. 2), which has a high and variable chloride concentration. The chloride concentration in the Ukak River depends partly on which tributary is seasonally dominant. Windy Creek runs high in early spring, whereas River Lethe and Knife Creek contribute more once their headwaters start melting in late spring.

We measured the total chloride flux in the Ukak River a number of times in 1991, 1993, and 2001 (Table 2). The June 1991 value is an estimate; the discharges of Windy Creek (4700 L/s) and Knife Creek (10,500 L/s) were measured by standard wading methods (Buchanan and Somers, 1969), and the discharges of River Lethe and Ukak River were calculated using mixing models that employed these discharges and the chloride contents of all four streams (1.8, 50, 19, and 28 mg/L, respectively). Aside from Windy Creek and the upper reaches of the River Lethe, the streams in the VTTS are not wadeable, and in 1993 and 2001 Ukak River discharge was measured by the dye-dilution method (Kilpatrick and Cobb, 1985). Measured volumes of rhodamine (around 125 mL) were injected near Three Forks and, from 1 to 4 min after the injection, 10 water samples per minute were obtained 200 m downstream of the injection point. Rhodamine contents of the water samples were analyzed within 20 min after collection in order to avoid sorption of rhodamine by inorganic particles (Smart and Laidlaw, 1977).

The Ukak River discharge data collected in 1993 and 2001 (Fig. 6) document large diurnal variations that are presumably related to snowmelt. The minimum discharge observed around 6 AM (t=30 h on Fig. 6) in both of the 2001 discharge time-series may reflect the transit time between the upper drainage area, where most snowmelt occurs, and the sampling site at Three Forks.

Windy Creek chloride concentrations were used to adjust Ukak River chloride-flux values (Table 2). The corrected chloride flux and discharge of the Ukak River are highly correlated (Fig. 7; r^2=0.89 when the estimated 1991 value is omitted). The corrected chloride flux in the Ukak River (>400 g/s) is much larger than that of the MVTS (<50 g/s; Table 1). Additional thermal waters, flowing within or at the

Table 2
Chloride-flux measurements, Ukak River at Three Forks. Windy Creek chloride concentrations of 1.2 mg/L in 1991 and 1993 and 2.0 mg/L in 2001 were used to adjust raw chloride-flux values (from Keith and Ingebritsen, 1991, and previously unpublished USGS data)

Date	Time (ADT)	Ukak River discharge (L/s)	Ukak River chloride concentration (mg/L)	Corrected chloride flux (g/s)
4-Jun-91	19:10	24,000 *e*	28	660 *e*
27-Jul-93	13:15	28,900	21	570
28-Jul-93	10:45	31,200	20	590
31-Jul-93	05:00	32,900	20	620
	09:00	33,900	21	670
	13:00	33,100	21	660
	21:30	39,500	19	700
8-Jun-01	12:15	21,000	22	420
12-Jun-01	08:45	32,000	19	540
	19:15	30,000	21	570
13-Jun-01	06:45	23,200	20	420

e=estimate made by combining discharge of Knife Creek, River Lethe, and Windy Creek.

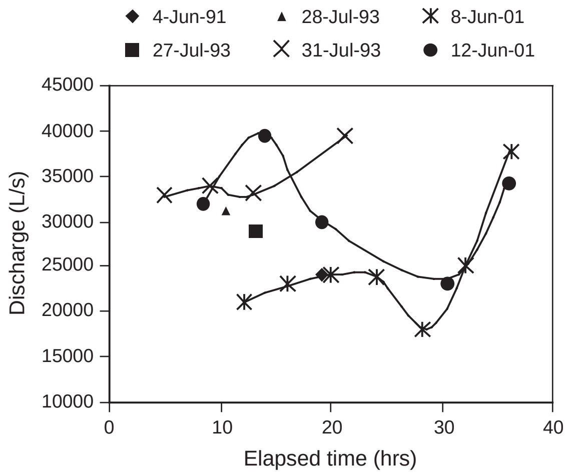

Fig. 6. Ukak River discharge as a function of elapsed time referenced to midnight of first day of measurement.

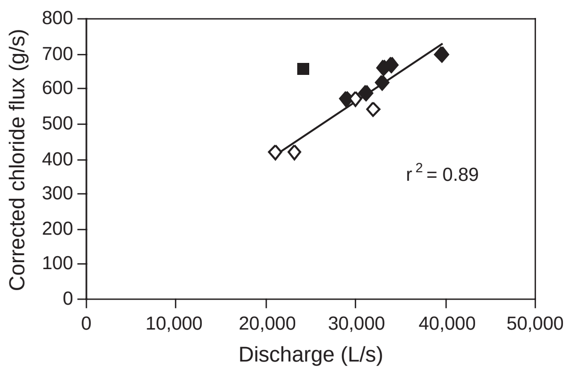

Fig. 7. Chloride flux in the Ukak River as a function of discharge (Table 2). Square denotes the estimated 1991 data point (Keith and Ingebritsen, 1991) and is omitted in the calculation of r^2. Solid diamonds are 1993 data and open diamonds are 2001 data.

base of the ash-flow sheet, must enter the streams, by submerged springs and from subaerial cold springs with a thermal component similar to those feeding River Lethe, perhaps not visible from the sides of the steep and unstable gorges.

3. Current hydrothermal heat loss

Assuming that the MVTS are typical of the hydrothermal input to the stream system in the VTTS, we can use the mean chloride flux in the Ukak River at Three Forks to calculate current hydrothermal heat loss from the 1912 deposits. In order to accomplish this we must also estimate the mean annual discharge of the Ukak River, which is likely lower than any of the point values measured to date. All of the discharge measurements have been made during the early summer, when snowmelt is near maximum. Discharge decreases in fall and in winter, when many streambeds are dry and snow filled.

Mean annual precipitation and drainage area can be used to roughly constrain the mean annual discharge of the Ukak River, assuming negligible interbasin groundwater flow. The larger watershed (Fig. 2) is in the mountainous terrain of the Aleutian volcanic arc. There is likely great variability in mean annual precipitation, but available precipitation maps are coarse and there are few precipitation stations in the vicinity. King Salmon (Fig. 1), the closest permanent station, lies in the rain shadow of the volcanic arc and has an average annual precipitation of 0.49 m/yr. Precipitation data collected intermittently in Brooks Camp, between Brooks Lake and Iliuk Arm (Fig. 1), suggest that mean annual precipitation in the lower VTTS may be comparable to that in King Salmon (Amanda Austin, National Park Service, personal communication, 2001). Kitoi Bay, at sea level on Kodiak Island (Fig. 1), lies on the rainy side of the arc and has an average annual precipitation of 1.6 m/yr. Precipitation on the Katmai group of volcanoes is likely higher than either of these values. Regional-scale precipitation maps are compatible with these point measurements, suggesting that the mountain range to the east of the VTTS receives 0.8–2.0 m/yr and the valley itself 0.6–0.8 m/yr (Western Region Climate Center, http://www.wrcc.dri.edu/).

Rates of evapotranspiration and sublimation from glacial ice are likely modest. There are no local evapotranspiration data, and few data for Alaska in general, but mean pan evaporation at seven Alaskan sites is <0.4 m/yr (http://www.wrcc.dri.edu/). Multiplying this value by 0.75 to more closely estimate evaporation from moist natural surfaces yields a value of <0.3 m/yr (and even this value may be an overestimate for the overcast conditions typical of the Alaska Peninsula).

It thus seems reasonable to assume that the streamflow yield (precipitation minus evapotranspiration) of the Ukak River basin above Three Forks is approximately 1 m/yr, or ~2.5×10^8 m^3/yr (8×10^3 L/s) over the ~250 km^2 area of the basin (Fig. 2). If this is the case, then the mean annual discharge at Three Forks is roughly a factor of three less than the lowest measured value (Table 2). Further assuming that the linear relation between chloride flux and discharge holds – an assumption supported by the fact that both the 1993 and 2001 data fit the same trend (Fig. 7) – the mean annual chloride flux at Three Forks is ~200 g/s. This translates to ~3000 L/s of MVTS-equivalent (~60 mg/L Cl) thermal water.

The associated hydrothermal heat flux can be calculated as

$$\Phi_{\text{heat}} = Q_t \rho c T \tag{2}$$

where Φ_{heat} is the excess heat flux in Ukak River (W), Q_t is the inferred thermal-water component in the Ukak (L/s), ρ is the density of water (kg/L), c is the specific heat of water (J/kg/°C), and T is the excess temperature of the thermal-water component (°C). Taking q_{thermal}=3000 L/s, ρ=1 kg/L, c=4200 J/kg/°C, and T=20 °C, based on the temperature of the MVTS (Table 1), gives Φ_{heat} ~250 MW. This is much less than the heat flux (800 MW) calculated by Keith and Ingebritsen (1991) based on a single chloride-flux estimate for the Ukak River (660 g/s: Table 2), but still appears to represent one of the largest hydrothermal discharges in North America (cf. Lassen, CA, 130 MW; north-central Oregon Cascades, 130 MW; Hot Creek, CA, 180 MW; greater Yellowstone, 6100 MW: Ingebritsen et al., 2001). The VTTS hydrothermal flux is related dominantly to mining of heat from the 1912 ash-flow deposits, and has clearly declined over the period of historical observation, whereas other large hydrothermal discharges in North

America are related to mining of heat from silicic magma bodies, and are quasi-steady on a multi-decadal time scale (Ingebritsen et al., 2001).

4. Thermal and hydraulic constraints from simple models

The discovery of the MVTS in 1987 was immediately followed by questions of the location, dimensions, and hydrological properties of the heat source for the thermal waters. Keith et al. (1992) discussed three possibilities. A genetic relation with thermal springs south of Katmai Pass and beneath Katmai–Trident–Mageik volcanoes (Fig. 2) seems unlikely because the chemistry of these waters is very different from that of the MVTS. A second possibility is leakage from a deep hydrothermal system of unknown dimensions and temperatures underlying the Novarupta vent region; however, because the isotopic composition of the MVTS waters plots very close to the Meteoric Water Line, a high-temperature hydrothermal origin seems unlikely. The third and perhaps most likely heat source is a lens of incompletely cooled welded tuff in the upper VTTS. Subsequent quantitative analyses of hydrothermal activity have been based on this premise.

Lowell and Keith (1991) suggested that any part of the 1912 ash sheet which remains incompletely cooled must be a densely welded tuff with no jointing, to have prevented the abundant surface water from cooling the tuff more quickly and completely. The most likely location for an incompletely cooled lens of tuff is the upper River Lethe area (Fig. 2), where the total thickness of the ash sheet may be as much as 200 m (Curtis, 1968). The last known fumaroles in the upper River Lethe area were 10–13 km upvalley from the MVTS, and an incompletely cooled lens of tuff in this area could also feed weak (<32 °C) residual fumaroles along the southwestern edge of Baked Mountain (Keith, 1991). Lowell and Keith (1991) used steady-state conductive cooling models for a hypothetical flow path between the upper River Lethe area and the MVTS to explain the difference between the observed spring temperatures and the source temperature inferred from chemical geothermometry.

Some insight into the transient nature of the system can be gained from conduction-only solutions, such as that for temperatures in a semi-infinite medium following instant emplacement of a rectangular parallelepiped of anomalous temperature (Simmons, 1967):

$$T(x,y,z,t) = \frac{T_0}{8}E(x,x_1,x_2)E(y,y_1,y_2)[E(z,z_1,z_2) + E(z,-z_1,-z_2)] \quad (3)$$

where

$$E(\alpha,\alpha_1,\alpha_2) = \operatorname{erf}\left[\frac{\alpha-\alpha_1}{(4\kappa t)^{\frac{1}{2}}}\right] - \operatorname{erf}\left[\frac{\alpha-\alpha_2}{(4\kappa t)^{\frac{1}{2}}}\right] \quad (4)$$

$$\operatorname{erf}\beta = \frac{2}{\pi^{\frac{1}{2}}}\int_0^{\beta} e^{-x^2}\mathrm{d}x \quad (5)$$

t is time since emplacement of the body of anomalous temperature, T_0 is the initial temperature of the parallelepiped (°C), and κ is thermal diffusivity (m^2/s).

In this model the parallelepiped – representing a lens of low-permeability welded tuff – is emplaced instantaneously at zero depth (top at z=0) in a medium with a uniform and constant thermal diffusivity (0.7×10^{-6} m^2/s) and conductivity (1.4 W/m/K). The initial surrounding temperature (T_0) at t=0 is 0 °C and the temperature at the surface is held at 0 °C (due to the last term in Eq. (3)). These boundary conditions represent the surrounding and overlying higher-permeability tuff, efficiently cooled by circulation of snowmelt-derived water at temperatures close to 0 °C. The heat flux at the top surface of the parallelepiped (z=0) can be calculated using:

$$Q_z|_{z=0} = \frac{-K_{\mathrm{th}}T_0}{4(\pi\kappa t)^{\frac{1}{2}}}E(x,x_1,x_2)E(y,y_1,y_2)\left(e^{\frac{-z_1^2}{4\kappa t}} - e^{\frac{-z_2^2}{4\kappa t}}\right) \quad (6)$$

where K_{th} is thermal conductivity (W/m/K).

On the basis of valley geometry (Fig. 2) and the seismic-refraction profiles of Kienle (1991), the maximum dimensions of an incompletely cooled lens of welded tuff in the upper River Lethe branch are approximately 3 km by 13 km by 60–80 m thick.

Solution of Eqs. (3)–(6) indicates that such a lens would have a heat-loss rate of 180–270 MW 80–90 years after deposition, sufficient to support the current hydrothermal heat loss (~250 MW) inferred from the Ukak River chloride-flux measurements.

Reasonable boundary and initial conditions can be estimated for transient, numerical models of the hydrothermal system in the VTTS. The initial temperature and density of the ash-flow sheet, the current temperature of the water in the MVTS (~20 °C), and current rate of hydrothermal heat loss (~250 MW) in Ukak River are useful constraints. Here we will describe results obtained for a two-dimensional (vertical-slab) model with a width of 3 km, consistent with the width of River Lethe branch of the VTTS (Fig. 2). A longitudinal profile along the River Lethe branch, through the MVTS, can be constructed on the basis of valley profiles (Curtis, 1968), seismic profiles (Kienle, 1991), and U.S. Geological Survey topographic maps (Fig. 8).

Variable degrees of welding, sintering, and fracturing exert a controlling influence on the hydraulic properties of the ash-flow sheet, but the extent to which these processes have affected various parts of the sheet is very poorly known. Thus, for purposes of modeling, we invoke a simple model of two tuff layers above the shallow-marine Naknek Formation, a flat to gently dipping, well-bedded conglomerate sandstone and siltstone that is locally up to 2.5 km thick (Berg et al., 1967). The Naknek Formation has developed large northwest trending vertical fractures, parallel to the direction of main tectonic compression (Wallmann et al., 1990), but otherwise appears to have very low porosity and permeability.

The simulation domain (Fig. 8) is a 13-km-long parallelepiped extending upvalley from the location of the MVTS. The sloping upper boundary approximates the slope of the valley. The bottom and lateral boundaries are taken to be no-flow boundaries, on the basis of the assumed low permeability of the Naknek Formation at depth and of the bounding mountain ranges relative to the 1912 tuffs. The upper boundary is assigned a constant temperature of 4 °C and a fluid pressure of 0.1013 MPa (1 bar). The temperature of the ash when erupted was about 800 °C (Hildreth, 1983). Allowing for cooling of the ash particles in transit, an initial temperature of 700 °C was assigned in the simulations. The underlying Naknek Formation is assigned an initial temperature of 5 °C (see Hogeweg, 2002 for

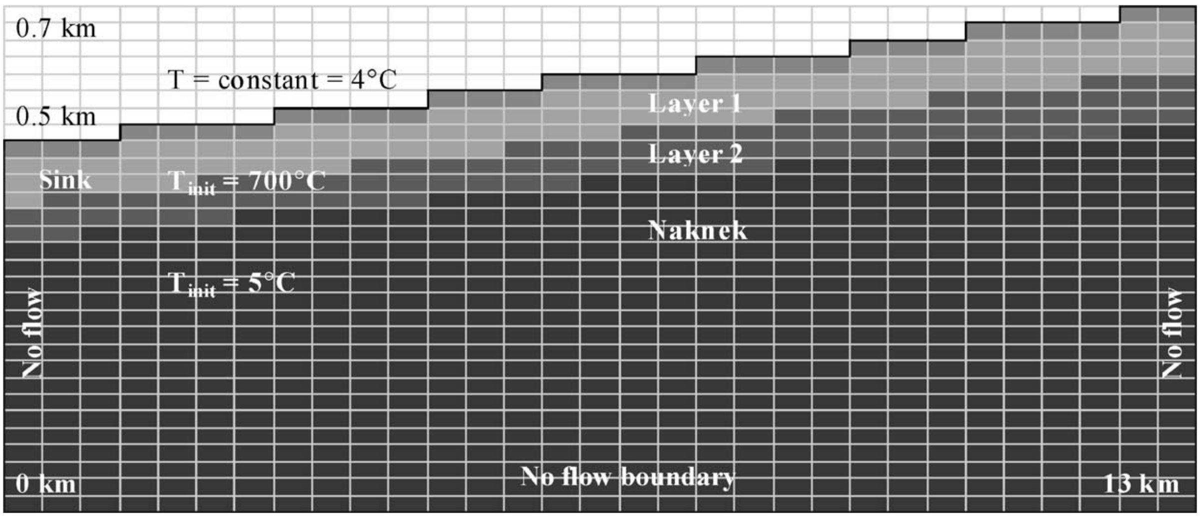

Fig. 8. Geometry, boundary, and initial conditions used in numerical modeling of hydrothermal system in the VTTS. Our "basic model" invokes the following hydraulic and thermal parameters (ρ=density, n=porosity, k=isotropic permeability, k_x=horizontal permeability, k_z=vertical permeability, K_{th}=thermal conductivity): Layer 1—ρ=1500 kg/m^3, n=0.2, k_x=10^{-11} m^2, k_z=10^{-12} m^2, K_{th}=1.2 W/m/K; Layer 2—ρ=1750 kg/m^3, n=0.1, k=10^{-13} m^2, K_{th}=1.4 W/m/K; Naknek Formation—ρ=2500 kg/m^3, n=0.01, k=10^{-17} m^2, K_{th}=2.2 W/m/K. Specific heat for all units is 1000 J/kg/K. Density data from Curtis (1968) and Kienle (1991); porosity values inferred from bulk densities; permeability values weakly constrained by model results (see text); and thermal-conductivity values based on measurements made in similar rocks.

sensitivity to initial conditions and other details of these simulations).

We used the U.S. Geological Survey computer code HYDROTHERM (Hayba and Ingebritsen, 1994) to simulate the cross section depicted in Fig. 8. HYDROTHERM is a finite-difference model that simulates multidimensional, multiphase flow of water and heat by solving mass- and energy-balance equations in terms of pressure and enthalpy. An extensive lookup table accurately defines fluid density, viscosity, and temperature as a function of pressure and enthalpy over a temperature range of 0 to 1200 °C and a pressure range of 0.5–10,000 bars.

A controlled sink representing the MVTS (Fig. 8) begins abstracting 600 L/s (Table 1) from the system at t=25 yr, a point in time at which most fumarolic activity had ceased. To prevent all water in the MVTS sink from being obtained directly from the overlying boundary, a constant-flux boundary (0.8 m/yr) is assigned immediately above the sink at t>25 yr. This prescribed flux represents the estimated precipitation-recharge rate in mid-valley.

Various observations and inferences can constrain the permeability of the model units to a limited extent. The permeability of the uppermost model unit (Fig. 8, Layer 1) must be sufficient to accommodate the annual recharge in the area of the VTTS (~1 m/yr) and lateral flow towards the sink representing the MVTS (600 L/s). A horizontal permeability of 10^{-11} m^2 and a vertical permeability of 10^{-12} m^2 are sufficient. The permeability of the Naknek Formation beneath the ash-flow tuff is taken to be 10^{-17} m^2, sufficiently low that heat transport in that unit is nearly entirely by conduction.

The permeability assigned to the lower tuff unit in the model (Fig. 8, Layer 2) affects the cooling history of the sheet, including temperatures within the deeper parts of the 1912 deposit and simulated temperatures at the MVTS (Fig. 2, sink). However, the surficial data collected over the past decade are not clearly diagnostic. For example, the simulated MVTS temperatures at t=90 yr (present day) are not highly sensitive to large variations in the permeability of the lower tuff unit (Fig. 9). Only low values, such that heat transport is conduction-dominated (e.g. k=10^{-16} m^2), give MVTS temperatures (41 °C) that appear incompatible with the range of maximum temperatures observed in 1987–2001 (17–29 °C: Table 1).

Simulated temperatures within the lower tuff unit are strongly controlled by permeability, and actual temperature-depth data – currently unavailable – might constrain its permeability. If the lower tuff unit is as permeable as the upper unit (k_x ~10^{-11} m^2, k_z ~10^{-12} m^2, then maximum temperatures in the 1912 deposits at t=90 yr are predicted to be only ~40 °C. In contrast, if all or part of the lower tuff has permeability as low as 10^{-16} m^2, then temperatures as large as 200 °C may persist (Hogeweg, 2002). Chemical geothermometry indicates that temperatures in the waters discharging at the MVTS have not exceeded 30–60 °C since 1987 (Keith et al., 1992), and there are no associated mineral deposits, but these observations do not preclude higher temperatures in conduction-dominated parts of the 1912 deposits.

All permeability structures considered in the numerical simulations (tuff $k \geq 10^{-16}$ m^2) result in hydrothermal heat loss in the range of 70–130 MW at t=90 yr (Hogeweg, 2002). This is less than the value inferred from the Ukak River chloride-flux measurements (~250 MW) or from the conduction-only solution (Eqs. (3)–(6)) for a lower-tuff unit with similar dimensions of 3 km by 13 km by 60 m (180 MW). The discrepancy between the analytical and numerical results in this regard is mainly due to advective heat loss, which is significant even for the lowest value of permeability considered in the numerical simulations.

There are various possible reasons for the discrepancy between the simulated heat loss and the value inferred from field observations. Most obvious is the large difference between simulated hydrothermal throughflow rates (up to 1000 L/s, including 600 L/s representing the MVTS) and the rate inferred for the entire VTTS system based on the Ukak River measurements (3000 L/s). In the context of the numerical model, simulating a larger hydrothermal discharge would require: (1) higher permeabilities in parts of the system; (2) introducing fluid sources or constant-pressure lateral boundaries to represent recharge from the larger watershed surrounding the VTTS; and/or (3) increasing the size and complexity of the model to allow for possible hydrothermal input from the Knife Creek branch or the Novarupta vent area (Fig. 2)). In light of the relatively high temperatures (90 °C fumaroles) that persisted in the Novarupta vent area until at least the late 1980s, the

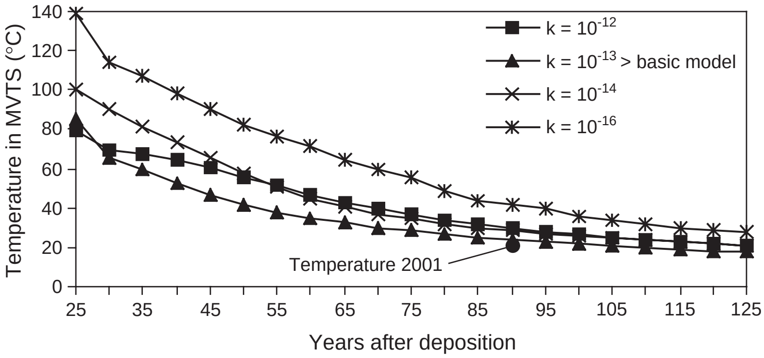

Fig. 9. Effect of permeability in the lower tuff unit (Fig. 8, Layer 2) on simulated temperature at the MVTS, 25–125 years after 1912. 'Temperature 2001' is the temperature measured in the MVTS in June of 2001. Permeability (k) in m^2.

latter possibility may be the most likely. Shallow underflow into the upper Lethe directly from the still-hot Novarupta area might feed MVTS-like springs, supplementing the hydrothermal discharge.

It is also possible that the heat- and fluid-flux values are overestimated from available field data. We have assumed that all anomalous chloride in Ukak River results from MVTS-like thermal waters, and estimated the mean annual discharge of the Ukak on the basis of a basinwide precipitation yield of 1 m/yr. If thermal waters with lower temperatures than the MVTS are contributing significantly to the chloride flux, or the precipitation yield is much less than assumed, then the actual hydrothermal discharge is <250 MW.

5. Implications for ash-sheet hydrology and cooling history

The hydraulic properties of ash flows are of interest both from a water-resources perspective (Winograd, 1971) and because cooling rates and mechanisms are not well understood. The ongoing cooling of the 1912 deposits in the VTTS offers a novel opportunity to constrain the large-scale permeability structure of an ash sheet. However, this opportunity cannot be fully realized because available data are insufficient to constrain detailed, quantitative hydrogeologic models.

In general the hydrologic properties of welded and nonwelded tuffs differ considerably (Winograd, 1971). Welding is defined as the process that promotes the union or cohesion of the still hot, viscous glassy shards and pumice under the weight of an overburden. Pumice and glass fragments are flattened during welding, decreasing the porosity of the tuff and its primary permeability. Welded tuff develops and sustains vertical fractures easily, however, often resulting in significant secondary permeability. In many thick tuff units the secondary permeability of the welded tuff is greater than the primary permeability of the nonwelded tuff, which is generally too soft and non-cohesive to sustain fractures (Winograd, 1971).

Preliminary numerical-model results for the VTTS, only partially constrained by field observations of hydrothermal discharge, fail to distinguish between a model with uniformly high ash-flow-sheet permeability and a model with lower permeability at depth. The key to better-constrained hydraulic properties would appear to be temperature-depth data, such as might have been obtained by the scientific drilling program proposed in the early 1990s (Eichelberger et al., 1991). Though current models are nonunique in this regard, it seems likely that there is actually significant heterogeneity in hydraulic properties within the VTTS ash-flow deposits. Welded, partially welded, nonwelded, and sintered tuff all exist, though the welded tuff is seldom observed in situ. Sintering is caused by chemical reactions of fumarolic fluids with the ash. Sintered tuff is therefore abundant in the VTTS, especially near the Novarupta vent, where sintering makes the tuff strong enough to form columnar joints. The sintered tuff in the upper and middle valley is heavily fractured, either perpendicular or parallel to the direction of ash-flow. The fractures perpendicular to the flow are due to a small component of downvalley creep, together with cooling contraction in the upper part of the sheet, whereas the fractures (sub-) parallel to the flow are likely due to a greater axial compression in areas where the tuff is thicker (Keith, 1991). Welded tuff is rarely exposed in outcrops in the VTTS, but phreatic explosion craters in the Knife Creek branch are surrounded by blocks of welded tuff that have been thrown to the surface (Curtis, 1968). Assuming the same thermal conditions in the River Lethe branch, where the sheet is thicker, welded tuff is almost certainly present at depth there as well.

Both field observations and model results provide some insight into the cooling history of the 1912 ash-flow sheet. Assuming that 6–7 km^3 of erupted magma (Fierstein and Hildreth, 1992) was emplaced as tephra at about 700 °C, the initial heat content of the sheet was about 1×10^{19} J. Initial rates of heat loss were extremely high. In 1919, heat loss was on the order of 20,000 MW, based on our estimate of the mean annual streamflow (2.5×10^8 m^3/yr, or 8 m^3/s) and Allen and Zies' (1923) calculation that >90% of the water entering the VTTS was vaporized. Such large rates of heat loss would have entirely depleted the initial heat content of the sheet in about 15 years and could not be sustained. Similarly, numerical-modeling results show that – regardless of permeability in the deeper parts of the sheet – cooling during the first 5 years after emplacement will be rapid, because of the influx of precipitation recharge and the very large temperature gradients between the ash-flow sheet and the upper boundary (4 °C) and the Naknek Formation (initial temperature of 5 °C).

Numerical results further indicate that thermal conditions after 90 years of cooling are not particularly sensitive to the early cooling history. For instance, differences in initial ash-flow-sheet temperature of 200 °C (500–700 °C) result in maximum temperature differences of only 25 °C in the simulation domain, and a difference of only 2 °C in the simulated temperature of the MVTS (Hogeweg, 2002). (The actual emplacement temperature was presumably greater than 500 °C, because ash would not weld at such a low temperature.)

Both analytical and numerical calculations suggest that the current heat content of the ash-flow sheet is only about 1/10 of the 1912 emplacement value, or ~1×10^{18} J. However, given the current low rate of thermal evolution (e.g., Fig. 9), hydrothermal heat losses of >100 MW can be expected to persist for many more decades.

6. Conclusions

By the late 1980s most fumarolic activity had ceased in the VTTS, but the discovery of thermal springs in mid-valley (MVTS) in 1987 suggested continued cooling of the ash-flow sheet. Data collected at the MVTS in 1987–2001 show a statistically significant correlation between maximum observed Cl concentration and temperature and also show a statistically significant decline in the maximum Cl concentration. The chloride-flux method and an estimated mean annual discharge of Ukak River just below the ash-flow terminus can be used to constrain the current rate of discharge of hydrothermal waters from the VTTS at approximately 3000 L/s. This quantity of MVTS-equivalent thermal water translates to an associated hydrothermal heat flux of ~250 MW, much less than the heat flux (800 MW) calculated by Keith and Ingebritsen (1991) based on a single chloride-flux estimate for the Ukak River, but still one of the largest hydrothermal discharges in North America. Available data provide reasonable boundary and initial conditions for simple transient modeling.

Both an analytical, conduction-only model and a numerical model predict large, though declining, rates of heat loss from the ash sheet 90 years after deposition. Various field observations suggest that the 1912 deposits are likely quite heterogeneous in terms of their hydraulic properties, but existing data fail to constrain the permeability structure of the ash sheet, except that large precipitation-recharge rates (~0.6–0.8 m/yr) require large near-surface permeabilities (k_x ~10^{-11} m^2, k_z ~10^{-12} m^2). Experimentation with the numerical model indicates that the key to better-constrained hydraulic properties may be temperature-depth data, such as might have been obtained by the scientific drilling program proposed in the early 1990s.

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