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Crater Lakes Reveal Volcanic Heat and Volatile Fluxes

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ABSTRACT

Aqueous lakes situated at the top of active but quiescent volcanoes serve as gas condensers and calorimeters that provide long-term integrated release rates of volatiles and heat during passive degassing of subsurface magma. Some crater lakes contain the most acid natural water on Earth ($\text{pH} < 0$). Analysis of hydrogeology of the acid lake at Volcán Poás, Costa Rica, reveals volatile release rates into the hydrosphere of 0.78 Gg/yr fluorine, 15 Gg/yr chlorine, and 13 Gg/yr sulfur (1 Gg = 10^9 g) and a power output of 200 MW during passive degassing in 1988–1989. An equivalent flux of sulfur may be precipitating as chemical sediments in the crater lake. After magma intrusion or hydrofracturing events in the subsurface, these fluxes were observed to double (F, Cl) or even increase tenfold (S) for short periods of time. At Grímsvötn volcano, Iceland, the database for flooding from the lake suggests volatile release rates of 0.058 Gg/yr fluorine, 6.8 Gg/yr chlorine, 3.1 Gg/yr sulfur, and 39 Gg/yr carbon integrated over 15, 32, 32, and 15 yr, respectively, and 4000–5000 MW power output. Volatile flux ratios calculated over several years at Poás and over 15 yr at Grímsvötn are observed to be 0.87 and 0.71 (S/Cl) and 17 and 48 (S/F); the C/S flux ratio averaged 14 over 15 years at Grímsvötn. Power output for each volcano can also be used to calculate rate of magma cooling in the subsurface; the volatile release rate divided by the rate of magma cooling yields estimates for the decrease in volatile content of subsurface magma during cooling. For the shallowly buried calc-alkalic basalts and andesites at Poás, the magma lost 175 ppm F, 3400 ppm Cl, and 3000 ppm S; for the deeply buried



Figure 2. Photograph looking north into the crater of Volcán Poás, Costa Rica. Visible are the Main Crater walls, the pyroclastic cone of 1953–1954 (center), and the greenish yellow waters of the crater lake, from which steam clouds emanate. The crater is about 800 m in diameter, and its floor lies at 2300 m above sea level.

basalts at Grímsvötn, the magma lost <1 ppm F, 75 ppm Cl, 34 ppm S, and 1600 ppm CO_2 . Integrated estimates for release of F, Cl, and S at Poás are roughly within an order of magnitude of published estimates for most passively degassing volcanoes. However, F, Cl, S, and C fluxes at Grímsvötn are smaller than observed at noneruptive volcanoes. These low values suggest either that there are unaccounted-for sinks in the system or that other degassing estimates, based on short-term sampling, are overestimates. Only a few of the 80 or so crater lakes of the world have been analyzed; these lakes may reveal the long-term power and volatile release rates for these volcanoes.

INTRODUCTION

To assess the effects of anthropogenic input of critical components on the Earth system, we need to quantify the rates of natural release and cycling of these components into and within Earth's hydrosphere, biosphere, geosphere, and atmosphere. For example, to model the geochemical carbon cycle, we need to quantify degassing rates of CO_2 from volcanoes. Similarly, degassing rates of F, Cl, and S are needed to quantify the geochemical cycles of these elements. Most global volatile budgets rely on short-term measurements of volatile release at individual volcanoes, which are then extrapolated to longer time frames and multiplied by the number of passively degassing or actively erupting volcanoes (e.g., Stoiber et al., 1987). Gas emission rates and gas chemistry are measured by remote correlation spectrometry, direct fumarole sampling, treated filter analysis, sublimate analysis, MIRAN infrared spectrophotometry, Raman spectrometry, satellite remote sensing, incrusta-

tion sampling, ice-core-inclusion analysis, and melt-inclusion gas analysis. Most of these techniques do not provide long-term emission rates, unless permanent monitoring stations have been established. However, for at least 12% of the 714 Holocene-age or younger volcanoes listed in the *Catalog of Active Volcanoes of the World*, aqueous lakes in the crater condense volcanic volatiles. These lakes also change in volume and temperature, reflecting the power output of the volcano. Acting as condensers and calorimeters, these crater lakes thus integrate the long-term volatile and heat output of some active volcanoes.

Almost 40 crater lakes are reported to have above-ambient temperatures and to contain acidic, sulfur-rich water. Some of these lakes maintain the lowest pH of any natural waters on Earth ($\text{pH} < 0$). A few of these lakes have been sampled and analyzed: e.g., El Chichón, Mexico; Kusatsu-Shirane, Japan; Mt. Ruapehu, New Zealand; Volcán Poás, Costa Rica; Rincón de la Vieja, Costa Rica. Other crater lakes contain neutral-pH water that represents condensed volcanic gases diluted by meteoric influx. The pH of the lake water is determined by the rates of the natural titration reaction: acid volcanic gas + water

+ acid-neutralizing silicate rock. In many cases, these closed-basin lakes concentrate dissolved solutes until chemical sediments (e.g., gypsum) precipitate.

Estimation of the heat, mass, and/or solute budgets of a few crater lakes has revealed mechanisms of fluid flow and heat exchange above shallow magma chambers (Shepard and Sigurdsson, 1978; Björnsson, 1988; Hurst et al., 1991; Brown et al., 1989; Brantley et al., 1992; Rowe et al., 1992a). We summarize the implications of the heat and mass balance studies at two volcanoes: Volcán Poás, a convergent-plate volcano in Costa Rica, and Grímsvötn, a hot-spot volcano in Iceland.

VOLATILE AND HEAT BUDGETS

To estimate the volatile and heat budget of a volcanic crater lake, we define the outputs and inputs of the system (Fig. 1): R_e (evaporation), R_{out} (river or floodwater flow), R_s (bottom seepage), R_{in} (low-temperature influx), R_{gv} (geothermal influx as vapor), R_{gw} (geothermal influx as water). The flux, R_{in} , can be composed dominantly of rain (R_r) as in Volcán

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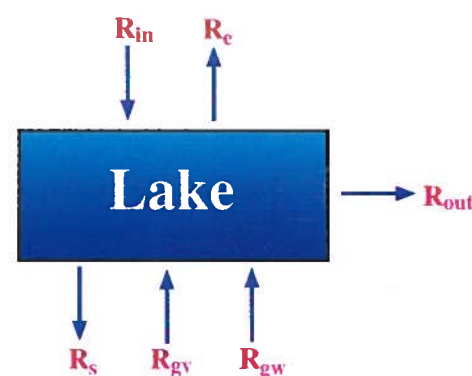


Figure 1. A box model for crater lakes summarizing the water fluxes: R_{in} (low-temperature meteoric input), R_{gv} and R_{gw} (geothermal input as vapor and liquid water, respectively), R_e (evaporative loss), R_s (seepage loss), and R_{out} (outflow as rivers or floodwaters).

Editor's Note:

Each year the David and Lucile Packard Foundation awards 20 Fellowships for Science and Engineering in a national competition to promising young scientists and engineers working in fields that receive relatively less popular attention than high-energy physics, space, and medicine. Each Packard Fellowship provides \$100,000 per year for five years to the Fellow's institution, \$90,000 of which is for use of the Fellow to support his/her research work. These young researchers are truly among the "best and brightest" in the United States. The science article in this issue is one of several in which Packard Fellows in earth science report on research in their field.

Poás, or dominantly of meltwater from ice (R_i) as in Grímsvötn.

Water mass balance is simply the balance between mass inputs, $R_{in} + R_{gv} + R_{gw}$, and outputs, $R_e + R_s + R_{out}$. Heat balance is simply the balance between heat inputs, $R_{gv}L_v + (R_{gv} + R_{gw})(h_h - h_i)$ and outputs, $R_eL_e + R_iL_i + R_{in}(h_i - h_{in}) + R_s h_i + R_{out} h_i$. Here, L_v is the latent heat of condensation at the temperature (T_h) at which the geothermal vapor enters the lake; L_e is the latent heat of evaporation at lake temperature (T_l); L_i is the latent heat of melting of ice; h_h is the specific enthalpy of water at T_h ; h_i is the specific enthalpy of water at T_l ; and h_{in} is the specific enthalpy of water at the temperature at which rain or meltwater enters the lake. We can also write a mass balance on a dissolved solute of concentration c which is non-volatile and does not dissolve or precipitate in the lake. Solute mass balance

equates the solute inputs due to low-temperature water ($c_{in}R_{in}$), geothermal vapor ($c_{gv}R_{gv}$), and geothermal water ($c_{gw}R_{gw}$), to the solute outputs due to evaporation (c_eR_e), seepage (c_sR_s), and river or flood output ($c_{out}R_{out}$).

Slightly modifying the approach of Björnsson (1988), we define the ratio of geothermal input to meteoric input as $k = (R_{gv} + R_{gw})/R_{in}$. Noting that seepage or river output is chemically identical to lakewater, $c_{out} = c_s = c_l$, we can solve the heat and mass balance equations and constrain k by noting that either R_{gv} or R_{gw} can equal zero:

$$\frac{K_2 - K_1 h_l}{h_h - 2h_l + L_v} < k < \frac{K_2 - K_1 h_l}{h_h - 2h_l}$$

In this equation, $K_1 = (R_e + R_{out})/R_{in} - 1$ and $K_2 = (R_e L_e + R_i L_i + (h_i - h_{in})R_{in} + R_{out} h_i)/R_{in}$. This value of k can also be used to calculate the ratio of geothermal influx to total water influx: $k/(1 + k)$. A further constraint can be placed on k by estimating c_{gw} , the

concentration of solute in the liquid component of geothermal influx (see Björnsson, 1988).

HEAT AND WATER BUDGET OF VOLCÁN POÁS

Volcán Poás, along with adjacent volcanic centers, forms the Cordillera Central of Costa Rica. The northern half of the main crater is occupied by a pit crater, which contains an acidic lake (Fig. 2); the southern edge of the lake consists of a 30-m-high pyroclastic cone that formed during the last phreatomagmatic eruption (1953–1954) and which represents the locus of sub-aerial fumarolic activity. The lake varied between 38 and 91 °C and between pH values of -0.9 to 0.5 between 1978 and 1991. Gravity measurements suggest that the top of the cooling magma body lies at a depth of ~500 m (Rymer and Brown, 1987).

Inputs and outputs of the summit hydrothermal system, including ground-water seepage to the watershed of the Agrio River to the northwest, are summarized in Figure 3. Rowe et al. (1992a) analyzed all available data over the period 1978–1990 and concluded that the baseline power output of the volcano was 200 MW, with fluctuations to 800 MW. Using appropriate values summarized by Rowe et al. (1992a) for Poás for the period March 22, 1978, to January 1, 1980, we calculate that k must lie between 0.77 (all vapor geothermal influx) and 8.5 (all liquid geothermal influx). Noting that lake and rain water at Poás contained 1400 ppm and 4 ppm Na, respectively, geothermal fluid must have had a concentration of dissolved Na at least 4% higher than the lake water. Rowe et al. (1992a) concluded that the lake thus represents the upper manifestation of a liquid-dominated convection cell (Fig. 3) and that virtually no vapor enters the lake: $k = 8.45$. Calculations for other periods of lake stability during the past decade yielded similar conclusions. A value of $k = 8.45$ suggests that 90% of the lake's water mass is geothermal input.

HEAT AND WATER BUDGET OF GRÍMSVÖTN VOLCANO

Grímsvötn, the most active volcano in Iceland (more than 50 eruptions over the past 1100 yr), is located in the interior of the Vatnajökull ice cap. Grímsvötn consists of evolved quartz-normative tholeiites, and it hosts a dilute, neutral-pH crater lake. The bottom of the lake is at 1050 m above sea level, and pressure at the lake bottom varies from about 3 MPa when the top of the lake surface is 1330–1350 m above sea level, to 4 MPa when the top of the lake reaches 1425 to 1450 m (Björnsson, 1988). At this critical level,

the water lifts the 7-km-wide ice barrier lying in the saddlepoint at the south-eastern edge of the caldera, and lake water floods for 50 km beneath the ice (probably through tunnels), and continues over 30 km of the Icelandic coast to the sea. These periodic floods, termed jökulhlaup, drain water from the bottom of the lake at maximum discharge rates since 1938 of 1000–8000 m³/s (Björnsson, 1988). Jökulhlaup are closely monitored; discharge, river chemistry, and ice cap elevation data are available back to 1954. Since 1974, jökulhlaup water samples were collected from the river Skeidará about 8 km from where the jökulhlaup water emerges from beneath the glacier.

Using these data, Björnsson (1988) estimated the rate of ice melting into the lake (R_i) at 400 Tg/yr (1 Tg = 10¹² g), and the rate of ablation at the surface of the glacier (R_a) at 80 Tg/yr water. Total water input, R_{in} , thus averages 480 Tg/yr. Assuming that the lake temperature (T_l) averages 0.1 °C, the melting temperature (T_o) is 0 °C, and the temperature at which the geothermal fluid enters the lake (T_h) is 235 °C (boiling temperature at 3 MPa), Björnsson (1988) concluded that the power output of the volcano has decreased from about 5000 to 4000 MW since 1860.

Between jökulhlaup, we can follow Björnsson (1988) and solve for k : $0.10 < k < 0.28$. Assuming that silica is a conservative component introduced to the lake only by geothermal input, we can use our observed value for the lake, $c_l = 68$ ppm (see data below), and again following Björnsson and Kristmannsdóttir (1984), we note that the upper limit for silica concentration in Icelandic geothermal waters is 700 ppm, which, assuming mass balance on silica, puts a lower constraint on k : 0.18. The liquid fraction of the geothermal influx must therefore be greater than 65%. These calculations, first completed by Björnsson and Kristmannsdóttir (1984) reveal that the lake at Grímsvötn, like Poás, is primarily receiving liquid geothermal input. The ratio of geothermal to total water input is less than 22%, explaining the wide difference in chemistry between Grímsvötn and Poás.

VOLATILE BUDGET OF VOLCÁN POÁS

Rowe et al. (1992a) summarized several periods of relative constancy in Poás crater lake volume and chemistry. During these periods, the only volatile losses from the system consisted of ground-water losses from the liquid-dominated convection cell at the top of the volcano, the cinder-cone fumarole flux, and volatilization from the

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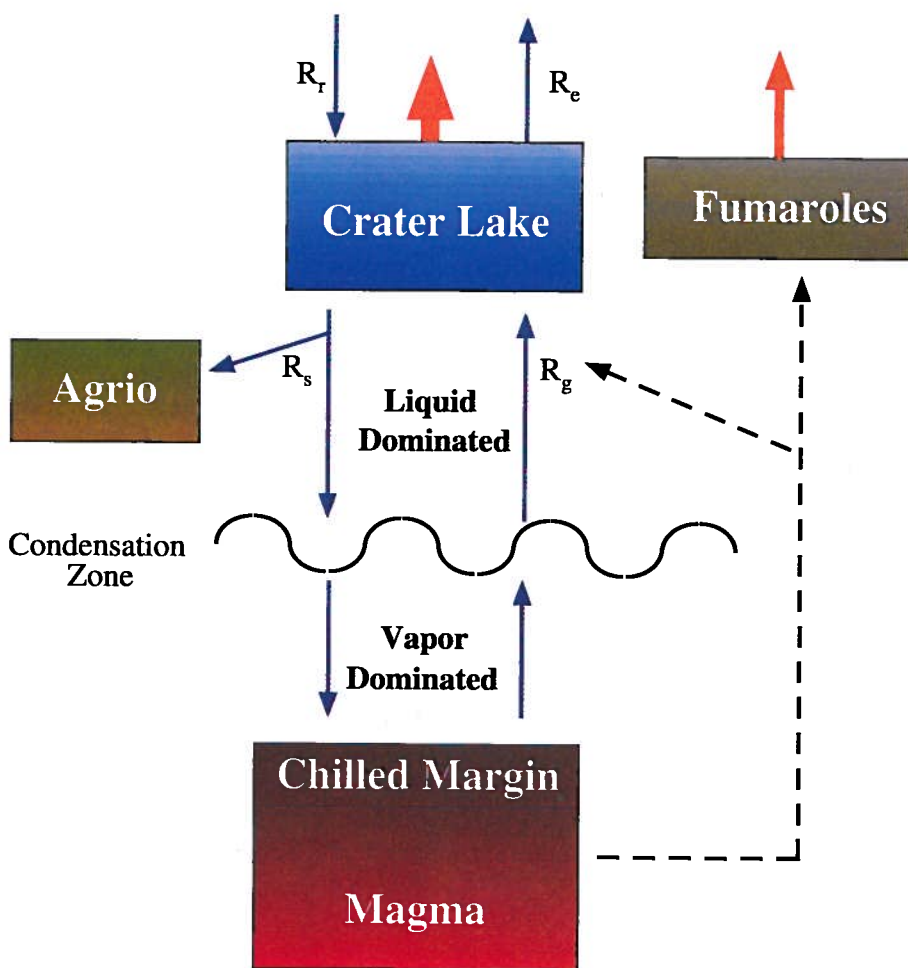


Figure 3. Box model for the summit hydrology of Volcán Poás. Inputs to the crater lake are rainfall (R_r) and geothermal fluid (R_g); outputs from the lake are evaporation (R_e) and seepage (R_s). Downward-seeping brine is heated until it rises again as geothermal influx to the lake, or escapes the summit system by flowing northwest to the Agrio River system. The simplest model that explains water mass, solute mass, and heat balance of the crater lake hypothesizes two convection cells in the volcano: an upper, liquid-dominated cell that is manifested at the surface as the lake, and a lower, vapor-dominated cell. High-temperature vapor escapes directly to the atmosphere as fumaroles emanating from the cinder cone in the Main Crater.

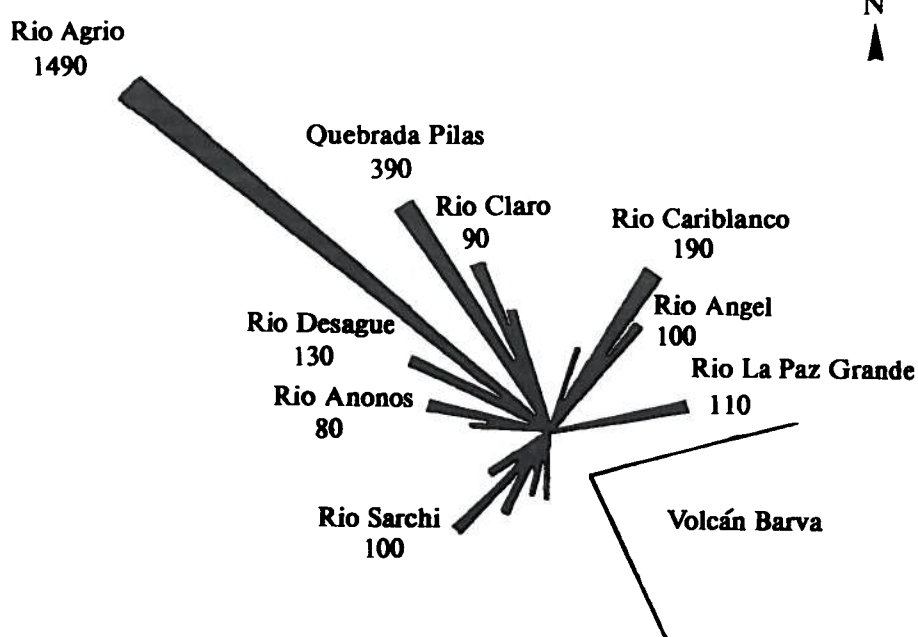


Figure 4. Schematic diagram illustrating the relative contribution of chemical weathering in rivers as a function of position around Volcán Poás. Triangular rays are scaled to indicate solute fluxes (metric tons of rock per year per km² of drainage basin) for each river as labeled. A second volcano, Volcán Barva, abuts Poás to the southwest.

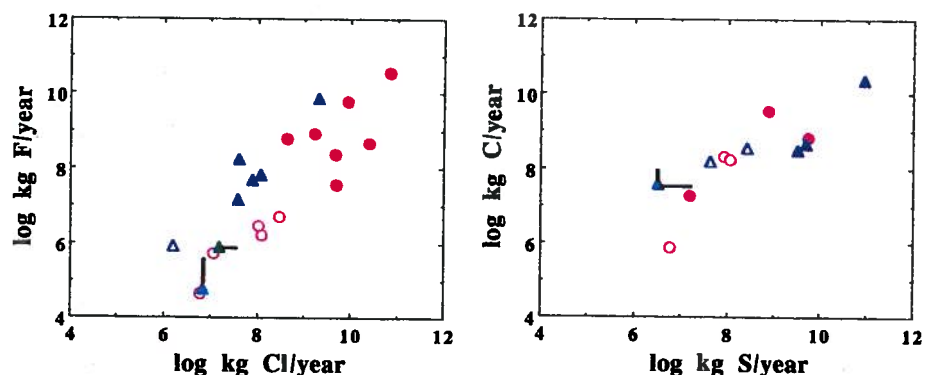


Figure 5. A compilation of published estimates of volatile fluxes, where both volatile components (C and S or Cl and F) are available, for erupting (solid symbols) and nonerupting (open symbols) volcanoes. Volatile release rates for Poás and Grímsvötn are indicated in green and aqua (no estimate for C release was made at Poás). Circles refer to volcanoes at convergent plates (Agung, Augustine, Etna, Katmai, Masaya, Merapi, Poás, Satsuma Iwojima, St. Helens, Tambora, Tarawera, Vulcano, White Island); triangles refer to intraplate or hot-spot volcanoes (Erebus, Grímsvötn, Kilauea, Öraefajökull); and the triangle indicating the largest C and S release refers to the estimate for the global mid-ocean ridge system. Literature references are in Ágúst-dóttir (1993).

Crater Lakes continued

lake (Fig. 3). Quantification of these fluxes reveals overall volatile flux from the volcano. No attempt was made to assess CO₂ flux.

Numerical modeling of the summit ground-water system indicates that seepage from the crater lake, constrained by a high water table associated with older volcanic cones to the north and south, flows dominantly to the Rio Agrio watershed to the northwest (Sanford et al., 1993). Rowe et al. (1993) have shown that acid seepage causes dissolution around the volcano to be asymmetric (Fig. 4); dissolution along the northwest flank aquifer is approximately 6.7 Gg/yr (Rowe et al., 1992b). Observed fluxes of the conservative elements F, Cl, and S in waters exiting the Rio Agrio watershed during 1988–1989 were 0.69, 14, and 11 Gg/yr, respectively, and the residence time of fluids in the aquifer is estimated to be between 3 and 17 yr (Rowe et al., 1993).

The fluxes through the dome fumaroles fluctuated greatly: during the high-temperature period of 1981–1984, observed F, Cl, and S fluxes were 0.72, 22, and 110 Gg/yr, respectively; fluxes during low-temperature periods (e.g., 1985–1989) were on the order of 0.036, 0.22, and 0.36 Gg/yr (Rowe et al., 1992a; see also Casadevall et al., 1984). Rowe et al. (1992a) argued that these heat and volatile bursts were caused by hydrofracturing of the cooled magma carapace, or by influx of magma in the subsurface.

Analysis of lake chemistry reveals that some HCl gas escapes from the lake, especially when the lake temperature increases. Calculated chloride concentrations in the vapor phase during high-temperature periods were mirrored by observed chloride concentrations in acid rain falling downwind (Rowe et al., 1992b). On the other hand, the concentration of sulfate in acid rain collected during periods of low fumarole emission was generally low, and significant volatilization of HF from the lake was also not observed (Rowe et al., 1992b).

To quantify the loss of components such as Cl by volatilization from the lake during moderate-temperature periods (e.g., 1978–1988), we can calculate the flux of these components in flank rivers affected by acid rain (Anonos, Desague, Gata, Claro, Pozo Azul, Angel; see Fig. 4). By first correcting these compositions by subtracting concentrations of a nonacidified river (Gorrión), and then multiplying corrected concentrations by flow rates measured in 1988 and 1990, we calculate fluxes of 0.058, 1.0, and 1.8 Gg/yr F, Cl, and S, respectively (Rowe et al., 1993).

Summing these fluxes out of the volcano (fluxes through Agrio + fumaroles + lake volatilization), we estimate that steady-state passive degassing at Poás releases 0.78 Gg F, 15 Gg Cl, and 13 Gg S per year into the surrounding atmosphere and hydrosphere (Fig. 5). After magma intrusion or hydrofracturing events in the subsurface, these fluxes were observed to double (F, Cl) or increase tenfold (S) for short periods of time.

One sulfur sink we have neglected is precipitation of sulfur and gypsum in the lake bottom. Lake sediment averaged 4.7 wt% native S and 33 wt% CaSO₄ (Rowe, 1992), and its volume was 2.4×10^6 m³, if we assume a cylinder of sediment with area 6×10^4 m² and height 40 m. A sediment density of 1500 kg/m³ suggests that 0.45 Tg of sulfur accumulated over a period of 27 years, representing a rate of sulfur precipitation of 17 Gg/yr, roughly equi-

valent to the flux out of the Agrio watershed.

VOLATILE BUDGET OF GRÍMSVÖTN VOLCANO

In collaboration with the Icelandic Glaciological Society, in June 1991 we drilled through the Grímsvötn ice cap and collected lake-water samples at various depths at two different boreholes. Water chemistry and temperature varied with depth (Ágústsdóttir, 1993), but the deepest samples contained 15.7, 0.22, 68, 12.7, and 811 ppm Cl, F, SiO₂, total S, and total carbonate as CO₂, respectively. Lake temperature averaged 0.15 °C. Samples of the jökulhlaup that occurred 5 months after the ice-drilling expedition revealed concentrations of 14.9, 0.14, 67.9, 11.0, and 492 ppm of Cl, F, SiO₂, total S, and total carbonate, respectively.

To estimate the volatile release rates of Grímsvötn over the period 1948–1991, we used river concentration data and jökulhlaup discharge rates for eight jökulhlaup (Ágústsdóttir et al., 1992). Assuming the crater lake was at steady state, we estimated the volatile release rates, $M_j c_j / \Delta t$ where M_j is the mass of jökulhlaup water, c_j is the concentration of volatile component in jökulhlaup water, and Δt is the time between floods. We also calculated release rates integrated over the entire length of time for which data are available: 0.058 Gg/yr fluorine (15 yr), 6.8 Gg/yr chlorine (32 yr), 3.1 Gg/yr sulfur (32 yr), and 39 Gg/yr carbon (15 yr) (Fig. 5).

The calculation rests on the assumptions that (1) the lake chemistry maintains steady state, (2) there are no other outlets for volatiles from the volcano, (3) all dissolved volatiles in lake water are volcanic in origin, and (4) no significant loss or dilution of volatiles occurs during transport to the river sampling point.

The assumption of steady state is confirmed by the observation that volatile budgets calculated for each jökulhlaup—the volatile budgets calculated over the ~5 yr between each jökulhlaup—are all within a factor of 2, with a slight trend of decreasing release rate with time, in agreement with the interpretation of Björnsson (1988) that the power output has decreased. If volatiles collected between jökulhlaup were not flushed with each event, the concentration in the lake and apparent release rates would increase.

Furthermore, some of the variability in release rates is explained by eruptive activity. The average ratio of release of F vs. Cl for floods not associated with eruptions is 0.013 ± 0.004 , whereas the average of the same ratio for the two jökulhlaup associated with eruptions (1953 and 1983) is 0.04. Similarly, the ratios of release rates of S and Cl for noneruptive vs. eruptive periods is 0.53 and 1.4, respectively.

Fumaroles on the nunataks at Grímsfjall mountain (the southern caldera rim) constitute a volatile leak. However, gas collected over several hours at four fumaroles revealed insignificant F, Cl, S, and C release rates (Ágústsdóttir, 1993).

Because the lake is undersaturated with respect to all oxidized gas phases at depth, minimal degassing of these components is expected. However, some degassing of H₂S occurs at the edge of the ice cap, where lake water is ephemerally exposed (area of ice hole in summer 1991 was only 20 m²); H₂S degassing is presumably also accompanied by CO₂ degassing.

Calculations of chemical speciation in the lake water suggest that no precipitation of volatile-containing

minerals is likely to occur, with the exception of alunite [KAl₃(SO₄)₂(OH)₆]. However, no alunite was discovered in sediments.

We have also assumed that contributions from snow melt and dissolution of bedrock or sediment are minimal. The concentrations of dissolved volatiles in glacial snow from Vatnajökull are well below 1 ppm. Correction of the volatile fluxes for calculated contribution from dissolution of basalt decreases the fluxes only slightly (Ágústsdóttir, 1993).

We expect little degassing to occur during the several-hour transport through ice tunnels 50 km to the edge of the glacier, but some degassing must occur in the 8 km between ice and sampling sites, especially for CO₂ and H₂S. We can compare concentrations of the relevant components in lake water sampled in June 1991 and jökulhlaup river water sampled in November 1991 (Ágústsdóttir, 1993). Because lake chemistry varied as a function of depth and distance to Grímsfjall, it is impossible to predict average lake concentration—or predicted jökulhlaup concentration if no degassing occurred—from just three sampling localities. For three components—fluoride, sulfur, and total carbonate as CO₂—the observed jökulhlaup concentrations (0.14, 11.0, and 492 ppm, respectively) were less than the observed highest concentrations in the lake water (1.05, 35, and 1109 ppm, respectively). All other components (except Ca) were observed to be identical between jökulhlaup and most concentrated lake waters. This suggests that, in the worst case, errors could be as large as a factor of 7.5, 4.9, and 2.2 for F, S, and C fluxes, respectively (Fig. 5).

VOLATILE RELEASE VS. POWER OUTPUT

Björnsson (1988) calculated the rate of magma cooling in the subsurface at Grímsvötn by calculating the heat released per kilogram of cooled magma: the specific heat of crystallization (419 kJ/kg) plus the specific heat capacity ($1.046 \text{ kJ} \cdot \text{kg}^{-1} \cdot \text{°C}^{-1}$) multiplied by the drop in temperature

(1300 to 200 °C). The power output of the volcano divided by the heat release per kilogram of magma reveals the rate of magma cooling in the subsurface: at Grímsvötn, 4500 MW power requires 9×10^{10} kg/yr. Using the same thermodynamic constants for Poás magma cooling, but assuming a lower temperature cooling range (1150 to 200 °C), we estimate that 200 MW power requires 4.4×10^9 kg/yr.

Assuming these magma cooling rates, we can also calculate the volatile content released from the magma during degassing. We assume that the depth of magma intrusion controls the extent of degassing, but that cooling of magma and associated cracking and water infiltration allows release of gas to the surface. At Grímsvötn, <1 ppm F, 75 ppm Cl, 34 ppm S, and 1600 ppm CO₂ must be released to account for the observed fluxes. At Poás, 180 ppm F, 3400 ppm Cl, and 2950 ppm S must be lost.

At Grímsvötn, we can assume that initial volatile contents of the tholeiitic basalts are approximately equal to that of Kilauean basalts: 350 ppm F, 87 ppm Cl, 1000 ppm S, and 3000–6000 ppm CO₂ (Gerlach and Graeber, 1985; Moore and Calk, 1991). Comparison of calculated and expected compositions suggests that emission of F and S is strongly suppressed at Grímsvötn, while a significant proportion of Cl and CO₂ is released.

Comparison of volatile release at Poás to that expected for andesitic magmas (Anderson, 1974) indicates suppression of F release, but significant Cl and S release. These observations can be qualitatively explained by the observation that S exsolution from magma is much more depth-dependent than exsolution of CO₂ (Gerlach, 1986). Anderson (1974) has also argued that Cl is much more efficiently transferred to surface reservoirs than is S. Enhanced release of volatiles at Poás is expected, given the shallower depth of magma intrusion (~500 m) compared to that of Grímsvötn (probably greater than 2 km).

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COMPARISON TO ESTIMATES IN THE LITERATURE

Release rates of F, Cl, and S at Poás are approximately within an order of magnitude of estimates for most passively degassing volcanoes, especially if the accumulation of sulfur sediments in the lake is included (Fig. 5). Including the maximum error estimates based on lake- vs. jökulhlaup-water concen-

trations, the F, S, and C flux at Grímsvötn is within about an order of magnitude of other fluxes. Although we saw no evidence of the existence of alunite in sediments, S loss as alunite is theoretically predicted, and it may also help explain the low S flux. However, the C flux, although within a factor of 10 of other estimates, is lower than all but one of the seven subaerial volcanoes summarized by Gerlach (1991). This, combined with the low value for Cl (and possibly F and S) release at

Grímsvötn may indicate that our estimates are minima, or that published fluxes for other volcanoes are overestimates based on spot measurements limited by logistics and weather. Sampling bias may also emphasize volcanoes with large volatile release rates.

CONCLUSIONS

At 200 MW, Volcán Poás is comparable to a moderately small coal-fired power plant in the United States; at 4000–5000 MW, Grímsvötn is much larger than the average coal-burning power plant (1000 MW). However, we note that only 102 volcanoes were known to be degassing in 1981–1982 (Stoiber et al., 1987), and if the power output for these volcanoes is between 200 and 5000 MW (see also Glaze et al., 1989), we calculate a rough global power output between 2×10^4 and 5×10^5 MW for nonerupting volcanoes. This global power output is small compared to the overall heat flow of Earth (4×10^7 MW) or commercial power consumption by humankind worldwide (1×10^7 MW, according to the Worldwatch Institute).

At 13 Gg/yr S release, Poás is comparable to a 200 MW coal-fired power plant burning coal that has 2 wt% S. The local effects of this volcanic S release were documented by Brantley et al. (1992). Storage of S at the top of this volcano could have important implications if a large-scale explosive eruption released the S. Globally, release of S to the atmosphere from passively degassing volcanoes amounts to about 3.4 Tg/yr (Stoiber et al., 1987). This is a very small flux compared with the total anthropogenic emission (about 100 Tg S/yr).

Ratios of S/Cl and S/F release integrated over several years at Poás (0.87 and 17), and integrated over 15 yr at Grímsvötn (0.71 and 48), are within the range 0.5 to 50 estimated by Symonds et al. (1988) for most active volcanoes. Symonds et al. (1988) used such ratios to calculate global release rates of F and Cl based on an assumed global release of 3.4 Tg S/yr. Our long-term integrated ratios may be especially useful in predicting global volatile release rates by this method; however, more long-term estimates of volatile ratios are needed. The two long-term estimates summarized here would suggest global passive release rates of approximately 0.07 to 0.2 Tg/yr F and 3.9 to 4.8 Tg/yr Cl, comparable to the anthropogenic fluxes due to coal-burning and industrial production of halocarbons: 0.44 Tg/yr F and 4.0 Tg/yr Cl (Symonds et al., 1988).

The ratio of C/S release at Grímsvötn is quite large, presumably because of the depth of the magma body. If the average global C/S ratio lies between the estimated minimum of 2.3 (Gerlach, 1991) and the value at Grímsvötn (14), then the global CO₂ flux would lie between 7.8 and 44 Tg/yr C, much smaller than the global anthropogenic flux (approximately 6000 Tg/yr C). However, more such long-term estimates, based on crater lakes or spectrophotometric monitoring, are needed to determine more accurately the global flux of all the volatile elements.

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REFERENCES CITED

Ágústsdóttir, A., 1993, Heat and volatile fluxes integrated over four decades at Grímsvötn volcano, Iceland [M.S. thesis]: University Park, Pennsylvania State University.

Ágústsdóttir, A., Brantley, S. L., Gudmundsson, M. T., and Björnsson, H., 1992, Volatile release rates from the Grímsvötn volcano, Iceland, in Kharaka, Y., and Maest, A. S. eds., Proceedings of the 7th International Conference on Water-Rock Interaction: Rotterdam, Balkema, p. 465–468.

Anderson, A. T., 1974, Chlorine, sulfur, and water in magmas and oceans: Geological Society of America Bulletin, v. 85, p. 1485–1492.

Björnsson, H., 1988, Hydrology of ice caps in volcanic regions: Reykjavík, Iceland, University of Iceland, Societas Scientiarum Islandica.

Björnsson, H., and Kristmannsdóttir, H., 1984, The Grímsvötn geothermal area, Vatnajökull, Iceland: Jökull, v. 34, p. 25–50.

Brantley, S. L., Rowe, G. L., Konikow, L. F., and Sanford, W. E., 1992, Toxic waters of Poás Volcano: National Geographic Research and Exploration, v. 8, p. 328–337.

Brown, G. C., Rymer, H., Dowden, J., Kapadia, P., Stevenson, D., Barquero, J., and Morales, L. D., 1989, Energy budget analysis for Poás crater lake: Implications for predicting volcanic activity: Nature, v. 339, p. 370–373.

Casadevall, T. J., Rose, W. I., Jr., Fuller, W. H., Hart, M. A., Moyers, J. L., Woods, D. C., Chuan, R. L., and Friend, J. P., 1984, Sulfur dioxide from Poás, Arenal, and Colima volcanoes, Costa Rica and Mexico: Journal of Geophysical Research, v. 89, p. 9633–9641.

Gerlach, T. M., 1986, Exsolution of H₂O, CO₂, and S during eruptive episodes at Kilauea Volcano, Hawaii: Journal of Geophysical Research, v. 91, p. 12,177–12,185.

Gerlach, T. M., 1991, Present-day CO₂ emissions from volcanoes: Eos (Transactions, American Geophysical Union), v. 72, p. 249–255.

Gerlach, T. M. and Graeber, E. J., 1985, Volatile budget of Kilauea volcano: Nature, v. 313, p. 273–277.

Hurst, A. W., Bibby, H. M., Scott, B. J., and McGuinness, M. J., 1991, The heat source of Ruapehu Crater Lake, New Zealand: Journal of Volcanology and Geothermal Research, v. 46, p. 1–20.

Moore, J. G. and Calk, L. C., 1991, Degassing and differentiation in subglacial volcanoes, Iceland: Journal of Volcanology and Geothermal Research, v. 46, p. 157–180.

Rowe, G. L., 1992, The acid crater lake system of Poás Volcano, Costa Rica: Geochemistry, hydrology, and physical characteristics [Ph.D. thesis]: University Park, Pennsylvania State University, 309 p.

Rowe, G. L., Brantley, S. L., Fernandez, M., Fernandez, J. F., Borgia, A., and Barquero, J., 1992a, Fluid-volcano interaction in an active stratovolcano: The crater lake system of Poás volcano, Costa Rica: Journal of Volcanology and Geothermal Research, v. 49, p. 23–51.

Rowe, G. L., Ohsawa, S., Takano, B., Brantley, S. L., Fernandez, J. F., and Barquero, J., 1992b, Using crater lake chemistry to predict volcanic activity at Poás Volcano, Costa Rica: Bulletin of Volcanology, v. 54, p. 494–503.

Rowe, G. L., Brantley, S. L., Fernandez, J. F., and Borgia, A., 1993, The chemical and hydrologic structure of Poás Volcano, Costa Rica: Journal of Volcanology and Geothermal Research (in press).

Rymer, H., and Brown, G. C., 1987, Causes of microgravity changes at Poás Volcano, Costa Rica: An active but non-erupting system: Bulletin of Volcanology, v. 49, p. 389–398.

Sanford, W. E., Konikow, L. F., Rowe, G. L., and Brantley, S. L., 1993, Ground-water transport of crater-lake brine at Poás Volcano, Costa Rica: Journal of Volcanology and Geothermal Research (in press).

Shepard, J. B. and Sigurdsson, H., 1978, The Soufrière crater lake as a calorimeter: Nature, v. 271, p. 344–345.

Stoiber, R. E., Williams, S. N., and Huebert, B., 1987, Annual contribution of sulfur dioxide to the atmosphere by volcanoes: Journal of Volcanology and Geothermal Research, v. 33, p. 1–8.

Symonds, R. B., Rose, W. I., and Reed, M. H., 1988, Contribution of Cl- and F-bearing gases to the atmosphere by volcanoes: Nature, v. 334, p. 415–418.

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GSA Officer and Councilor Nominees for 1994

Council announces the following officer and councilor candidates. Biographical information on all candidates will be mailed with the ballot to all voting members in August.

For Councilor (1994–1995) and President (1994)
William R. Dickinson; Tucson, Arizona

For Councilor and Vice-President (1994)
David A. Stephenson; Scottsdale, Arizona

For Councilor and Treasurer (1994)
David E. Dunn; Richardson, Texas

For Councilor (1994–1996)—Position 1
Charles G. Groat; Baton Rouge, Louisiana
John A. Cherry; Waterloo, Ontario, Canada

For Councilor (1994–1996)—Position 2
Grant Garven; Baltimore, Maryland
Leigh H. Royden; Cambridge, Massachusetts

For Councilor (1994–1996)—Position 3
Keros Cartwright; Champaign, Illinois
C. Blaine Cecil; Leesburg, Virginia

For Councilor (1994–1996)—Position 4
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H. Catherine W. Skinner; New Haven, Connecticut

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