CO₂ degassing in the Oregon Cascades
Elizabeth R. James, Michael Manga and Timothy P. Rose

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INTRODUCTION

The global volcanic CO$_2$ flux is commonly estimated by direct measurement of gas emissions from a few active volcanoes over relatively short time periods and extrapolation of these measurements to subaerial volcanoes worldwide (e.g., Marty et al., 1989; Gerlach, 1991; Marty and Le Cloarec, 1992; Williams et al., 1992; Sano and Williams, 1996; Marty and Tolstikhin, 1998). Although this method may provide a reasonable estimate of the CO$_2$ flux from actively degassing volcanoes (i.e., where a gas flux can be measured directly), it does not include diffuse CO$_2$ degassing from the flanks of volcanoes or magmatic CO$_2$ dissolved in ground water. Measurable diffuse CO$_2$ degassing has been recognized by several authors (e.g., Baubron et al., 1990; Farrar et al., 1995; Sorey et al., 1998). Recent studies have shown that the magnitude of diffuse emission of CO$_2$ from volcanoes can be similar to the flux at the crater plume (e.g., Allard et al., 1991). In some locations, the dissolved CO$_2$ in ground waters can be a significant component of the total volcanic CO$_2$ flux at a volcano (e.g., Rose and Davison, 1996).

Although the central Oregon Cascades have not had any volcanic activity for the past 1 k.y. (Wood and Kienle, 1990), it is possible to measure a dissolved flux of magmatic CO$_2$, which is assumed to be related to deep magmatic processes such as cooling and solidification of magma at depth (Harris and Rose, 1996). We show that by measuring the $^{14}$C activity in spring water, it is possible to estimate dissolved magmatic CO$_2$ emissions representative of large spatial areas and averaged over the residence time of ground water in the aquifer. Because magmatic CO$_2$ is devoid of $^{14}$C (Mook, 1980), measurement of this isotope in ground water can be used to estimate the dissolved flux of magmatic CO$_2$ (Rose and Davison, 1996).

GEOLOGIC AND HYDROGEOLOGIC SETTING

The Cascade Range in the Pacific Northwest extends 1200 km from southern British Columbia to northern California and is associated with the subduction of the Juan de Fuca plate beneath the North American plate. The High Cascades subprovince in central Oregon is located 250–300 km east of the convergent margin and forms the crest of the Cascades (Sherrod and Smith, 1990). Intense volcanic activity began ca. 8–10 Ma; more moderate activity lasted until the Holocene (Taylor, 1990). The High Cascades are composed primarily of Pliocene and Quaternary basalts and basaltic andesites. The Bachelor volcanic chain, located to the east of the central Oregon High Cascades, was emplaced between 18 and 7 ka (Scott and Gardner, 1992). The most recent volcanic activity in the central Oregon Cascades occurred 1.3 ka at Belknap Crater. South Sister last had volcanic activity nearly 2 k.y. ago, and Bachelor was last active 8 k.y. ago (Wood and Kienle, 1990). Other major volcanic centers include Mount Jefferson, which last erupted 70 ka, and Mount Washington, which last erupted 250 ka.

The permeable basalts and basaltic andesites of the High Cascades form a regional ground-water recharge area that feeds the many springs located to the east and west of the High Cascades (Ingebritsen et al., 1992). In this study we focus on the large volume cold springs to the east of the crest of the Cascade Range (Fig. 1). There are no hot springs or CO$_2$-rich soda springs in the study area.

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ABSTRACT

The carbon isotope content of dissolved inorganic carbon was measured for large cold springs in the central Oregon Cascades. Low $^{14}$C activities in some of the springs are interpreted to result from the dissolution of diffuse emissions of magmatic CO$_2$, even though volcanic activity has not occurred in this area for more than 1300 yr. On the basis of dissolved magmatic carbon concentrations in the springs, we infer a diffuse magmatic CO$_2$ degassing rate of 3.4 $\times$ 10$^5$ kg/yr per kilometer of arc for the central Oregon Cascades. The CO$_2$ flux calculated from estimates of the mean magmatic intrusion rate and experimentally determined values of CO$_2$ content in melts is consistent with that determined from our measurements of the dissolved CO$_2$ flux at springs.
ANALYTICAL METHODS

Waters were collected from nine cold springs in the central Oregon Cascades, and their dissolved inorganic carbon (DIC) was analyzed for $^{14}$C and $^{13}$C. Samples were treated in the field with HgCl$_2$ to prevent biological fractionation of the carbon isotopes and were stored in glass bottles with air-tight seals to prevent exchange with atmospheric CO$_2$.

The DIC was extracted from the water by using a dynamic vacuum-line acid-stripping technique that uses H$_3$PO$_4$ to acidify the water. The resulting CO$_2$ gas is liberated from the sample by using a nitrogen carrier gas, and the CO$_2$ is trapped cryogenically. CO$_2$ splits were made for $^{13}$C and $^{14}$C analyses, and samples intended for $^{14}$C analysis were reduced to graphite at 570 °C by using hydrogen gas and a cobalt metal catalyst. The $^{14}$C analyses were made on the accelerator mass spectrometer at Lawrence Livermore National Laboratory. Results are reported as a percent of modern carbon (pmc) relative to a National Bureau of Standards oxalic acid standard (Stuiver and Polach, 1977). The $^{13}$C analyses were made with an isotope-ratio mass spectrometer, and results are reported in the conventional delta notation relative to the Peedee belemnite (PDB) standard. Analytical precision is ±1 pmc for $^{14}$C and ±0.3‰ for $^{13}$C. Data are reported in Table 1. Data for Lower Opal Springs are from Caldwell (1998).

RESULTS

Three end-member DIC compositions were previously recognized in ground waters in the central Oregon Cascades: (1) DIC equilibrated with...
soil CO₂, (2) DIC equilibrated with atmospheric CO₂, and (3) dead carbon (¹⁴C = 0 pmc) of magmatic origin (Rose and Davison, 1996). Atmospheric- and soil-equilibrated DIC have ¹⁴C values >100 pmc and have δ¹³C values that range from ~−6‰ to +1‰, and from ~−18‰ to −14‰, respectively (Clark and Fritz, 1997). Magmatic CO₂ has typical δ¹³C values around −7‰ (Craig, 1963), although δ¹³C values for mantle carbon are more variable (Javoy et al., 1986). Fumarolic CO₂ gas emissions at Lassen Peak and Mammoth Mountain in California have δ¹³C values near −9.5‰ and −4.5‰, respectively (Janik et al., 1983; Sorey et al., 1998). Our study area does not contain any known fumaroles or soda springs; however, Minnehaha Soda Spring near Crater Lake, Oregon, contains 5100 ppm DIC with ¹⁴C = 0.2 pmc and δ¹³C = −7.4‰ and represents the Oregon magmatic DIC end member in our study. Figure 2 is a plot of ¹⁴C vs. δ¹³C for samples from this study showing the inferred end-member carbon isotope compositions (boxes) and mixing relationships between these end members. Similar mixing trends have been observed for other geographic locations (e.g., Rose and Davison, 1996; Sorey et al., 1998).

Open symbols in Figure 2 represent large cold springs with modern (bomb-pulse) ¹⁴C values, and filled circles represent samples with ¹⁴C values <100 pmc. The modern ¹⁴C values observed in six of the nine large cold springs are interpreted to indicate discharge from shallow, local ground-water flow (Fig. 3A). This interpretation is consistent with inferences based on spring temperatures (Manga, 1998). Four of the springs with modern ¹⁴C values have δ¹³C values indicative of open-system equilibration with soil CO₂, whereas two other samples show limited mixing between the soil- and atmosphere-equilibrated end members.

Three spring waters have δ¹³C-¹⁴C pairs that plot on a dead carbon addition trend in Figure 2, indicating possible mixing between biogenic soil DIC and the Oregon magmatic CO₂ end member, represented by Minnehaha Soda Spring. Sample MG has a δ¹³C value that suggests mixing between the three end-member compositions, probably as a result of contributions from atmosphere-equilibrated runoff at locations between the headwater spring and the gauging station. The springs with ¹⁴C values <100 pmc are warmer and at lower elevations and might reasonably be expected to discharge relatively deep, regional ground water (Fig. 3B) that has interacted with magmatic CO₂. The presence or absence of dead carbon in spring water in the central Oregon Cascades may therefore be used to infer the spatial scale of ground-water flow (Rose et al., 1996).

DISCUSSION

Figure 4 shows the relationship between DIC (in mg/L as HCO₃⁻) and ¹⁴C for the Oregon springs. The curve is a theoretical curve representing the addition of magmatic DIC (Cₘₐₓ) to a system with an initial DIC concentration of C₀. We choose a value of 36 g/m² for C₀, which is the average DIC for the Oregon samples that have ¹⁴C values >100 pmc. The total DIC concentration in the system is equal to Cₘₐₓ + C₀ and ¹⁴C (pmc) = 115(1 + Cₘₐₓ/C₀), where 115 pmc represents a typical ¹⁴C value for a modern ground water in this region. Our data are close to this theoretical curve, which suggests that measured ¹³C activities result from addition of dead carbon rather than from aging. If this added carbon were a product of calcite dissolution, we would see low ¹⁴C values in all of the springs because the dissolution of calcite occurs predominantly in the unsaturated zone during recharge. Furthermore, high aquifer permeabilities, heat-flow arguments, and mass-balance calculations suggest that aquifer residence times in this area cannot exceed the time-scale resolution of the ¹⁴C dating method (Manga, 1997).

The ¹³C activities, DIC measurements, and spring-discharge measurements can be used to calculate the dissolved magmatic CO₂ flux from a spring. Calculated fluxes are 2.4, 8.0, 1.8, and 4.4 × 10⁶ kg/yr for MH, MG, SP, and LO, respectively, the four samples in our study area observed to contain magmatic carbon. The dissolved magmatic CO₂ fluxes from MH, MG, SP, and LO represent the dissolved fluxes from Mount Washington and Belknap Crater, Mount Jefferson, Mount Bachelor, and Three Sisters, respectively. Our flux estimates provide lower bounds for the drainage areas because the springs do not discharge all ground water in the study area. The Metolius (MG) flux may be close to the total dissolved CO₂ flux for the drainage area, however, because the various springs along the Metolius River discharge at the eastern edge of the High Cascades, where a large fault scarp serves as a barrier to ground-water flow.

Dissolved fluxes of magmatic CO₂ are small in comparison to typical CO₂ fluxes from craters plumes (Brantley and Kopeckien, 1995). For example, Mount Etna contributes 2.6 × 10¹⁰ kg/yr to global volcanic CO₂ emissions (Allard et al., 1991), and Kilauea emits CO₂ at a rate of 1.3 × 10⁹ kg/yr (Greenland et al., 1985). Nevertheless, our measured dissolved CO₂ fluxes are comparable to dissolved fluxes from recently active volcanoes. The dissolved CO₂ flux from Lassen in northern California is 7.5 × 10⁶ kg/yr, which is 20% of the total geothermal CO₂ flux from this volcano (Rose and Davison, 1996). The dissolved CO₂ flux from Mammoth Mountain in California is 1.1 × 10⁹ kg/yr, which is 10% of the total diffuse CO₂ flux (Sorey et al., 1998). Lassen was last active in 1915 and Mammoth Mountain has undergone seismic activity over the past decade that is attributed to magmatic activity in the upper crust (Sorey et al., 1993). It is noteworthy that the springs in our study area are recharged at volcanic centers that have not produced any recent volcanic or seismic activity, yet the measured flux of magmatic CO₂ is still comparable to measurements for other regions with more recent volcanic activity.

If we assume that half the magmatic CO₂ is discharged to the east of the north-trending crest of the Cascades, CO₂ flux measurements from samples LO and MG can be used to estimate the total dissolved magmatic CO₂ flux from the Cascade arc. The total dissolved CO₂ flux for samples LO and MG is 12.4 × 10⁶ kg CO₂/yr, and represents ~75 km of Cascade arc. The dissolved CO₂ flux for both sides of the Cascades is therefore 3.4 × 10⁹ kg CO₂/yr per kilometer of arc. Extrapolating to the 1200 km length of Cascade arc gives a dissolved CO₂ flux of 4 × 10¹⁰ kg/yr (and for the ~40,000 km of arc worldwide, we extrapolate simply to the global dissolved CO₂ flux of 1.4 × 10¹¹ kg/yr). Marty et al. (1989) estimated the CO₂ flux from subduction zones to be between 4.4 × 10⁹ and 2.2 × 10¹⁰ kg/yr. Thus, the dissolved CO₂ flux—a contribution that has been commonly overlooked—might be a significant component of the total volcanic CO₂ flux.

Our estimated magmatic CO₂ fluxes for the central Oregon Cascades cold springs are consistent with estimates of mean magmatic intrusion rates...