Flux measurements of explosive degassing using a yearlong hydroacoustic record at an erupting submarine volcano

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[1] The output of gas and tephra from volcanoes is an inherently disorganized process that makes reliable flux estimates challenging to obtain. Continuous monitoring of gas flux has been achieved in only a few instances at subaerial volcanoes, but never for submarine volcanoes. Here we use the first sustained (yearlong) hydroacoustic monitoring of an erupting submarine volcano (NW Rota-1, Mariana arc) to make calculations of explosive gas flux from a volcano into the ocean. Bursts of Strombolian explosive degassing at the volcano summit (520 m deep) occurred at 1–2 min intervals during the entire 12-month hydrophone record and commonly exhibited cyclic step-function changes between high and low intensity. Total gas flux calculated from the hydroacoustic record is $5.4 \pm 0.6 \, \text{Tg} \, \text{a}^{-1}$, where the magmatic gases driving eruptions at NW Rota-1 are primarily $\text{H}_2\text{O}$, $\text{SO}_2$, and $\text{CO}_2$. Instantaneous fluxes varied by a factor of $\sim 100$ over the deployment. Using melt inclusion information to estimate the concentration of $\text{CO}_2$ in the explosive gases as $6.9 \pm 0.7 \%$, we calculate an annual $\text{CO}_2$ eruption flux of $0.4 \pm 0.1 \, \text{Tg} \, \text{a}^{-1}$. This result is within the range of measured $\text{CO}_2$ fluxes at continuously erupting subaerial volcanoes, and represents $\sim 0.2$–$0.6\%$ of the annual estimated output of $\text{CO}_2$ from all subaerial arc volcanoes, and $\sim 0.4$–$0.6\%$ of the mid-ocean ridge flux. The multiyear eruptive history of NW Rota-1 demonstrates that submarine volcanoes can be significant and sustained sources of $\text{CO}_2$ to the shallow ocean.

**Theme:** Assessing Magmatic, Neovolcanic, Hydrothermal, and Biological Processes along Intra-Oceanic Arcs and Back-Arcs

### 1. Introduction

[2] Most of Earth’s volcanic activity occurs on the ocean floor, unobserved and undetected, at mid-ocean ridges where it is mostly effusive and at subduction zones where it is more likely explosive [e.g., Rubin et al., 2012]. Records of explosive activity longer than a few weeks are rare even for subaerial volcanoes [Garcés et al., 1998; Hilton et al., 2002; Johnson, 2003; Aiuppa et al., 2008] and nonexistent for submarine volcanoes. From February 2008 to February 2009, we recorded the first long-term, continuous hydroacoustic and volcanic plume record of an exploding submarine volcano, NW Rota-1. NW Rota-1 is a conical edifice rising ~2200 m from the surrounding seafloor [Embley et al., 2006] within the recently established Marianas Trench Marine National Monument in the western Pacific Ocean (Figure 1a). A hydrophone and turbidity/temperature sensor, both moored ~150 m from the volcanic vent and 100 m off the seafloor (Figure 1b), provided an unprecedented yearlong view of the dynamic behavior of a Strombolian-style eruption on the deep seafloor.

[3] In this paper we use hydroacoustic explosion records at this submarine volcano to estimate the total explosive flux of gas from the summit vent. Our method is based on studies of infrasound arrivals at subaerial volcanoes [Johnson, 2003], but compensates for the different density, sound velocity, and sound pressure levels in the ocean. A correlation between acoustic signals and explosive degassing has been shown to exist at several subaerial, open-conduit volcanoes [Firstov and Kravchenko, 1996; Vergniolle and Brandeis, 1996; Oshima and Maekawa, 2001]. At Stromboli volcano, large (~1 m in diameter) gas bubbles exiting the magma column produced 10–30 Pa infrasound waves recorded at microphones a few hundred meters distant, and a near one to one correspondence between modeled and synthetic acoustic waveforms was achieved [Vergniolle and Brandeis, 1996]. The volatile bubbles at Stromboli are typically 80% H₂O, 10% CO₂, 5% H₂S and 5% Cl₂. Similarly, Oshima and Maekawa [2001] and Firstov and Kravchenko [1996] were able to generate highly accurate synthetic models of the infrasound (acoustic) records from the eruptive acceleration of compressed volatiles from the summit vents of Unzen and Klyuchevskoi, respectively.

[4] Two previous experiments have also attempted to correlate infrasound records of volcanic SO₂ gas flux with the flux observed using spectrometers coupled to telescopes and UV cameras [McGonigle et al., 2009; Dalton et al., 2010]. While McGonigle et al. [2009] observed good correlation between the SO₂ fluxes observed spectroscopically and irradiance captured by a thermal camera, these measurements were not well correlated with the infrasound data which was attributed to the strong directionality of the acoustic signals at Stromboli. Dalton et al. [2010] also used SO₂ emission data measured from a UV camera and coincident infrasound recordings over a two-hour period at Pacaya volcano (Guatemala) to calculate gas masses from both data sets. However, they found that the infrasound explosion fluxes had a high correlation (R² of 0.7) with the camera data during short duration (2–3 min) bubble-burst degassing events.

[5] Once an estimate of the explosive gas flux is made, flux estimates for a specific gas, such as CO₂, could be determined if its concentration in the exploded gas was known. However, quantitative
sampling of such gas is a daunting challenge at an erupting volcano. For this study we instead infer the CO\textsubscript{2} concentration from samples of melt inclusions from deep and seafloor summit rocks at NW Rota-1.

Although this technique is admittedly first-order, even approximate CO\textsubscript{2} concentrations are a useful beginning for comparing the explosive CO\textsubscript{2} flux
from a submarine volcano to the global database of
degassing subaerial volcanoes.

2. NW Rota-1 Volcanic Explosions

[8] The eruption style observed at NW Rota-1 can be
classified as submarine Strombolian, [Chadwick et al., 2008a] where gas exists as a separate phase
from the melt, effusive to explosive gas-driven
eruption blasts last from 10 s of seconds to several
minutes, and the produced volume of erupted
tephra is relatively small (1–100 m³/hr). The funda-
mental source of sound at Strombolian volcanoes
is almost certainly the rapid release of pressurized
gas at the free surface [Vergniolle et al., 1996;
Rowe et al., 2000; Johnson and Lees, 2000]. This
volcanic sound source arises from atmospheric
pressure perturbations caused by the explosive out-
flow of volcanic volatiles, with gas fluxes on the
order of 10¹ kg s⁻¹ [Newhall and Self, 1982]. This
eruption style has been confirmed at NW Rota-1 by
both short-term video observations [Chadwick et al., 2008a] and the long-term, continuous hydro-
phone recordings presented here. Moreover, in
Strombolian eruptions the exsolving magmatic gas
bubbles rise in the conduit, interacting and coales-
cing to form gas-rich and gas-poor zones. Between
explosive bursts, seawater interacts with the top of
the magma column to form a solid quench cap
[Chadwick et al., 2008a; Deardorff et al., 2011].
Explosion bursts occur when the gas-rich zones
reach the vent and the building gas pressure blows
apart the solid quench cap. The violently expanding
gas has sufficient velocity to entrain and transport
molten and solid ejecta into the water column
[Deardorff et al., 2011]. Previous ROV dives at NW
Rota have shown that the vast majority of degassing
occurs at the eruptive vent, and is relatively minor at
the diffuse hydrothermal vents elsewhere at the sum-
mit where the effluent is highly diluted [Chadwick et al., 2008a; Butterfield et al., 2011].

[7] During the 2008–2009 hydrophone deploy-
ment, the eruptive vent near the volcano’s summit,
at 520 m depth, exhibited near constant explosion
signal packets (Figure 2a). These explosion
packets lasted 60–120 s separated by quiescent
intervals of 10–30 s over the entire 12 month record
(Figure 2a). Each packet comprised hundreds of
individual (100–200 ms duration) explosion pulses
totaling ~12.7 million discrete pulses recorded
during the year (Figure 2b, bottom). The acoustic
explosion packets were broadband: 1–80 Hz with a
peak at 30 Hz (Figure 3). The loudest explosions
occurred during February to early June 2008 with a
typical sound source level of 192 dBₘₛ re µPa²/Hz
@ 1 m, equal to ~100 W of acoustic power. This
level is equivalent to the sound produced by an
oceanic supertanker, and would be detectable at a
range of ~100 km even during a Beaufort sea state
of 6 [Urick, 1975]. The hydrophone record also
documented a 1000-fold decrease in explosion inten-
sity (amplitude) over the course of the year, marked
by a sharp reduction after six months (in September
2008). This reduction in intensity of the acoustic
explosion pulses likely reflects a decrease in
explosivity at the vent, which may have been caused
by a decrease in the rate of eruption and degassing,
burial by accumulated ejecta, or other factors.

[8] Explosions at the summit vent also produced
counts of individual plumes that carried ash and
hydrothermal precipitates into the surrounding
water column [Resing et al., 2007; Butterfield et al.,
2011; Deardorff et al., 2011], with no long periods
of quiescence regardless of the explosion intensity
(Figure 3). Most of the short-duration spikes
recorded by the turbidity sensor, especially in the
latter half of the record, were precursors to longer
intervals of increased turbidity that lasted on the
order of a day. The frequency of intense light-
scattering pulses (Nephelometric Turbidity Units
>4 [American Public Health Association, 1985])
increased sharply between August 15 and 30 2008,
then lowered precisely when the explosion amplitude
abruptly decreased at the beginning of September.
This correlation is consistent with the hypothesis that
an interval of higher eruptive output in late August
changed the acoustic output of the eruptive vent
by effectively burying it in debris. The connection
between explosive intensity, acoustic amplitude, and
plume production is apparent from ROV video
recorded at the site in 2006 [Chadwick et al., 2008a],
and our data are the first to confirm the frequent
creation and dispersal of submarine volcanic plumes
on a yearlong time scale.

3. Hydroacoustic Explosion Analysis

[9] We use the hydroacoustic explosion records to
calculate the total explosive flux of gas from the
summit vent (Figures 2b, 4, and 5). Our method is
based on studies of infrasound arrivals at subaerial
volcanoes [Johnson, 2003], but compensates for the
different density, sound velocity, and sound pres-
sure levels in the ocean. Cumulative gas flux is
important because the total gas emission can be
used as a proxy to characterize the cumulative
magnitude of eruptive activity [Johnson, 2003;
Ripepe et al., 2007].
Ocean sound consists of a regular motion of molecules in the water medium. Because water is a fluid, particle motion is communicated to adjacent particles and a sound wave is thereby propagated outward from a source at a speed equivalent to the sound velocity in water (c). The particle motion is parallel to the overall direction of propagation, and since ocean water is compressible, the particle motion causes changes in pressure that can be detected by a pressure sensitive hydrophone [Urick, 1975]. Thus the acoustic pressure recorded on the hydrophone is a time history of ocean pressure perturbations (ΔP) relative to background ocean pressure at the recording water depth. These excess ocean sound pressures are usually small (~10^{-3} Pa), and it has been the standard in the ocean acoustic literature to express sound pressure in decibels (dB) relative to a reference pressure (P_o) of 1 μPa [Urick, 1975]. Moreover, it is convention to use 1 m as the reference distance (r_o) when calculating the acoustic pressure of the source (written as dB re 1 μPa).

The hydrophone used in this study records sound pressure changes in the deep ocean as proportional voltages. These signals are amplified, a 220 Hz low-pass anti-alias filter is applied, and the waveforms are digitized at a sample rate of 500 Hz before being stored on hard drives within the instrument pressure case.

The recorded acoustic signal levels shown in Figures 2a and 4 were derived using the sonar equation [Urick, 1975]:

\[
SL = RL + TL - IR,
\]

Figure 2. (a) Still image of magmatic explosion from summit vent in April 2006 [Chadwick et al., 2008a]. (b) Top diagram shows September 2008 hydrophone record of five explosion signal packets, each packet is comprised of 100 s of individual explosion pulses. Bottom diagram shows (left to right) typical explosion pulse (~200 ms duration), mass flux of pulse calculated from equation (5), and cumulative mass (gas) flux of the pulse calculated using equation (6).
Figure 3. Hydrophone and eruption plume records from NW Rota-1 volcano. (a) Long-term, daily average, spectrogram of the hydrophone data (frequency in Hertz, amplitude intensity in decibels relative to 1 micro-Pascal$^2$/Hz) for the entire 12-month record. (b) Corresponding turbidity measurements in nephelometric turbidity units (NTU). NTU data (black line) were recorded every 15 min. The steady rise in baseline turbidity after September is interpreted to result from cumulative fouling of the optical lens. Red line shows time history of number of intense turbidity pulses (NTU > 4), perhaps of elemental S precipitates (e.g., as seen in Embley et al. [2006]). Inset shows how these pulses are normally followed by a longer-lasting period of increased turbidity. Because the turbidity pulses typically saturated the light-scattering sensor (5 V full scale), we cannot determine if the maximum turbidity correlates with explosion intensity. Even with quieter explosions, enough magmatic volatiles may be released to create a full-scale light-scattering response.
where RL is the acoustic received level recorded at the hydrophone, calculated as $10 \log_{10} (\Delta P/P_0)^2$ with units of dB re 1 $\mu$Pa. IR is the frequency-dependent instrument response that accounts for the sensitivity of the hydrophone ($-194$ dB re 1 V/$\mu$Pa) and the gain of the amplifier. TL is the transmission loss (dB) over the acoustic propagation path. Since the hydrophone is very close to the sound source ($\sim 180$ m slant range), we assume the acoustic signal is dominated by the volcanic explosion source, the signal undergoes only spherical spreading with negligible acoustic attenuation, and any distortion of the signal because of source directionality should be minimal. Unlike seismic propagation in the earth, ocean acoustic waveforms are relatively undistorted during propagation because the ocean does not support shear waves and is largely devoid of structures that scatter, attenuate, or reflect acoustic waves. Treating the source as a point source and applying the spherical spreading loss, the TL is then estimated as $20 \log_{10} (r/r_0)$, where $r$ is the distance from the source to the receiver and $r_0 = 1$ m. The source level SL therefore represents the acoustic pressure of the signal at a distance of 1 m from the source (dB re 1 $\mu$Pa @ 1 m).

[13] Strombolian-style hydroacoustic signal generation at NW Rota-1 is relatively straightforward to model because: (1) a point source approximation is appropriate, because there is only one erupting vent, (2) vent diameter ($\sim 1$ m [Chadwick et al., 2008a]) is much less than the explosion sound wavelengths ($\sim 20$–1500 m), (3) sound sources are primarily generated at the magma conduit/ocean interface and not within the magma column, (4) the elastic waves from the volcanic explosions are generated by the volatile gas bubble collapse as it

Figure 4. (top) Typical gas explosion pulses (200 ms duration) recorded on NW Rota-1 hydrophone. (middle) Mass flux calculated from each of the pulses using equation (5). (bottom) Cumulative gas flux of the pulses calculated using equation (6).

Figure 5. (top) Sound levels, root-mean square (rms) pressures in 1-h bins, recorded 150 m from the summit vent. (bottom) Cumulative gas flux calculated using equation (6), total as of February 2009 is $5.4 \pm 0.6$ Tga$^{-1}$. Note that the acoustic amplitude axis in the top panel is a log scale.
exits the magma at the vent-ocean interface, thus the bulk of the elastic wave energy is produced prior to the mixing of magmatic gases into seawater, and (5) ejection velocities are below the ocean sound speed and no shockwave is produced. The NW Rota-1 acoustic signals (Figures 2a, 2b, and 4) are clearly explosions and not jets; jet signals produce sustained band-limited tremor, whereas explosions result in broadband “N-wave” shaped waveforms [Johnson and Lees, 2000]. Thus the NW Rota-1 submarine volcanic sounds can be modeled as a simple or monopole source in a homogeneous medium, where the restoring force in the ocean is proportional to particle displacement. Acoustic compressional waves propagate elastically according to the wave equation [Aki and Richards, 2002]:

$$\nabla^2 (\Delta P) - \frac{1}{c^2} \frac{\partial^2}{\partial t^2} (\Delta P) = -f(t) \delta(r), \quad (2)$$

where \( f(t) \) \( \delta(r) \) represents the source time force as a function of time \( t \) and the delta function \( \delta(r) \) is zero everywhere except at \( r = 0 \). For spherical waves, a solution to the wave equation can then be written as [Aki and Richards, 2002]:

$$\Delta P = \frac{-f(t - r/c)}{4\pi r}, \quad (3)$$

For a simple acoustic point source in a homogeneous medium, the effective force function is equal to the rate of change of flux (mass outflow) from the source. Following the linear theory of sound, the excess pressure \( (\Delta P) \) radiating into a half-space is given by [Lighthill, 1978]:

$$\Delta P = \frac{1}{2\pi} \left( dq(t - r/c) \right), \quad (4)$$

where \( q(t) \) is the mass flux from a point source at a distance \( r \) from the receiver, \( c \) is the speed of sound within the water column, and the \( 2\pi r \) term accounts for geometric spreading of the pressure wave within a spherical half-space. The acoustic impedance of the fluid medium is inherently accounted for through conservation of momentum, and only the excess pressure, not the absolute pressure, enters into the calculation of mass flow.

4. Gas Flux Estimates

[14] Several approximations must be employed when estimating the flux of mass (or essentially gas in these volcanic explosion) values: (1) instrument response and ocean acoustic propagation effects are deconvolved, (2) the location of the hydroacoustic source is a point fixed at the vent, and (3) pressure perturbations at the source are small enough that a linear relationship exists between excess pressure, particle velocity, and particle displacement in the acoustic medium. Deconvolving the instrument response and acoustic propagation effects is standard practice in waveform analysis to estimate the physical parameters of the source. Moreover, the assumption of a point source is valid because the size of the vent is small (~1 m) as compared to the distance the acoustic waves propagate from the vent to the hydrophone (~150 m). Also since the excess ocean sound pressures in the NW Rota-1 explosions are relatively small (~1 Pa), there is very likely a linear relationship between excess pressure \( (\Delta P) \) and particle velocity and displacement.

[15] From the recorded hydroacoustic pressure traces (Figures 2b, 4, and 5 (top)), the corresponding mass flux \( (\text{kg s}^{-1}) \) for a source of time duration \( (\tau) \) may be approximated by:

$$q(t) = 2\pi \int_0^\tau \Delta P(t + r/c) dt \quad (5)$$

The cumulative flux \( M(t) \) is then the time integral of the mass flux rate (Figure 5, bottom):

$$M(t) = \int_0^\tau \left[ \int_0^\tau (\Delta P(t + r/c)) dt \right] d\tau \quad (6)$$

The hydroacoustic records should adequately represent cumulative explosive gas flux at NW Rota-1 because gas flow out of the summit vent was constantly turbulent (non-laminar), continually producing pressure waves. The cumulative flux mirrors the acoustic amplitudes (Figure 5), clearly showing a rapid gas release from March to early June 2008, another short increase during August, and a leveling off of gas flux rates beginning in September through the end of the record. In calculating cumulative gas flux we selected individual explosion pulses that exceed 1 Pa in amplitude and were between 150 and 2000 ms in duration, using both the positive and negative portions of the pulse amplitude (Figures 2b and 4). Using these criteria, a total of ~12.7 million explosion pulses were detected from the NW Rota-1 hydrophone, and each discrete pulse was integrated using equation (5) and equation (6) to calculate the cumulative mass flux. As the hydrophone receiver is tethered above the seafloor, changes in the slant range between the receiver and vent, which are estimated to be on the order of ±20 m, contribute the largest (11%) uncertainty to the mass calculation, making the total mass estimate 5.4 ± 0.6 Tg a⁻¹.
The uncertainty in the calibration of the ITC 1032 hydrophone, as reported by the manufacturer, is ±1 dB re 1 μPa, and there is perhaps another ±1 dB re 1 μPa associated with the calibration of the instrument pre-amplifier. Hence the total uncertainty in the instantaneous pressure readings (σi) is on the order of 2 dB re 1 μPa (or 1.26 × 10−6 Pa). For the NW Rota data set this translates to a total uncertainty of ±0.014 Tg of mass—a value that is negligible relative to the uncertainty in slant range.

The hydroacoustic pulses integrated in this manner should adequately represent cumulative explosive gas flux at NW Rota-1, since the eruption is a series of gas explosions. The onsets of explosions at NW Rota-1 are highly impulsive, and because gas expansion accelerates dramatically at the onset, high-amplitude hydroacoustic excess pressures are generated. Thus these impulse signals may be used to determine a cumulative mass flux for the onset of an explosion. It is also important to note that the hydroacoustic record does not provide information on the rate that solid tephra debris is expelled into the water column since that process does not contribute to the acoustic signal recorded at the hydrophone.

In most cases, accurate location of the explosion source is particularly important since volcanic systems can possess multiple active vents, e.g., Stromboli [Ripepe et al., 2007]. However, seafloor observations before and after the hydrophone deployment confirm there is only one explosion vent at NW Rota-1 [Chadwick et al., 2008a], and thus the only visible gas emissions are from this one explosive eruptive vent. It is also important to note that equation (6) refers to the cumulative mass flux of gas emitted from the volcanic vent and not the mass of water displaced by the volumetric change of the source [Lighthill, 1978; Johnson, 2003]. Unfortunately, low-frequency contributions to the cumulative mass flux are often inadequately represented because laminar, steady state gas flow out of a vent should theoretically generate no sound [Johnson, 2003]. Because passive degassing is a common mechanism at most active volcanoes [ Sparks, 1997], cumulative gas flux values recovered from infrasonic pressure records should be considered a lower limit or an estimate of a transient contribution.

5. Estimating CO2 Flux

We next turn to the more challenging problem of estimating the concentration of CO2 in the explosive gas. It is known that the magmatic gases driving the eruptions at NW Rota-1 are primarily H2O, SO2, and CO2 [Resing et al., 2007; Chadwick et al., 2008a; Lupton et al., 2008; Butterfield et al., 2011], but we had no sampling capability to directly measure their exact proportions and quantities at the vent. Instead, we estimate the composition of gas emitted at NW Rota-1 by comparing volatile concentrations measured in melt inclusions trapped within olivine crystals at deep depths (≈5.2 km) to those trapped at shallow depths at or near the eruption site. H2O, CO2, and SO2 species are the main volatile phases that degas from the melt as it ascends, and their concentrations in the melt inclusions range from 1.4 to 2.8 wt%, 5–460 ppm, and 535–800 ppm, respectively [Shaw et al., 2006], where the associated errors are on the order of 10% [Shaw et al., 2010]. Thus by subtracting the lowest melt compositions (trapped at the seafloor) from the highest values (from a relatively deep undegassed source), we calculate that the expelled gas contains 3.1 mol % or 6.9 ± 0.7 wt % CO2, within the range (≈1–10%) found at other volcanoes along subduction zones [Symonds et al., 1994]. The similarity between previous measurements of the S/C mole ratio in the NW Rota-1 plume, 1.4 [Resing et al., 2007], and the S/C mole ratio derived from the melt inclusions, 0.8, supports our melt inclusion approach. S/C ratios can vary during an eruption cycle, but rarely more than an order of magnitude [e.g., Hobbs et al., 1991; Doukas, 1995; Aiuppa et al., 2008]. The average S/C ratio from high-temperature fumaroles across 11 volcanic arcs is 0.8 ± 1.3 [ Fischer, 2008], the same as our value from melt inclusions. This close agreement is somewhat fortuitous but does imply that our S/C value is reasonable.

As previously noted, the hydroacoustic record does not provide information on the rate that solid debris is expelled into the water column, most of which is ultimately moved down the steep south flank of the volcano by gravity flows [Walker et al., 2008; Deardorff et al., 2011; Chadwick et al., 2012]. Also, because the mass flux is a measure of gas emitted from the volcanic vent and not the mass of water displaced by the volumetric change, in our view it is not necessary to correct our mass flux estimate for the gas density of the three major volatiles comprising the total gas volume.

We thus use the total flux of explosively released gas from NW Rota-1, 5.4 ± 0.6 Tg a−1, to calculate an explosive CO2 flux of 0.4 ± 0.1 Tg a−1, using the errors associated with the calculations of total gas flux and the CO2 fraction of the gas. This estimate is a minimum, as some CO2
degassing could have occurred prior to melt inclusion entrapment (i.e., the original source could have had even higher CO$_2$). A highly conservative approach to the error assessment is to calculate the range of CO$_2$ flux using the observed range of ~1–10% CO$_2$ in volcanic gases from other arc volcanoes [Symonds et al., 1994]. This calculation yields a CO$_2$ flux range of 0.05–0.54 Tg a$^{-1}$.

[22] Our estimate does not include CO$_2$ vented by non-eruptive hydrothermal discharge. At NW Rota-1, water column measurements indicate two CO$_2$ plumes originating from the summit region: a high-rising plume rich in CO$_2$ associated with explosive eruptions, and a deeper plume with lower CO$_2$ apparently derived from passive fluid discharge from low-temperature vent fields [Resing et al., 2007].

There is no evidence for gas venting on the volcano’s flanks [Walker et al., 2008]. No hydrothermal tracers (e.g., $^3$He, dissolved Mn and Fe) have been detected in waters around the flanks, and measured ΔCO$_2$ below the summit is ~±1 μM (measured plume values range up to 50 μM [Resing et al., 2007]). The balance between eruptive and passive flux has not been quantified at NW Rota-1, but vertical profiles through both plumes find substantially higher CO$_2$ concentrations in the eruption plume. This assumption is consistent with measurements at subaerial volcanoes, where diffuse degassing typically contributes <~10% of the CO$_2$ flux at erupting (i.e., open conduit) arc volcanoes [e.g., D’Alessandro et al., 1997; Carapezza and Federico, 2000; Wardell et al., 2001; Varley and Armienta, 2001].

Thus we consider explosive eruptions to be the major source of CO$_2$ from NW Rota-1.

6. Discussion

[23] Our estimated explosive CO$_2$ flux from NW Rota-1 falls within the broad range, <0.01 to >30 Tg a$^{-1}$, established from measurements at individual arc or near-arc subaerial volcanoes (Table 1). This compilation shows that eruptive basaltic/andesitic volcanoes, such as NW Rota-1, are distinguished by CO$_2$ fluxes that are consistently higher (>~0.5 Tg a$^{-1}$) and more variable than quiescent volcanoes. During an eruption cycle, CO$_2$ fluxes can easily vary by one or more orders of magnitude (e.g., Etna, Redoubt, Augustine, NW Rota-1 (Table 1)), so occasional discrete measurements may provide misleading information about CO$_2$ output on a long-term basis. At Etna, for example, a rare continuous time series of CO$_2$ emissions during a several month eruption cycle found fluxes repeatedly fluctuating between ~0.4 and >24 Tg a$^{-1}$ (Figure 6a), a cycling only crudely approximated by concurrent discrete sampling [Aiuppa et al., 2008]. Highly variable gas flux has also been documented at erupting volcanoes such as St. Augustine [Symonds et al., 1992], Spurr [Doukas, 1995], Stromboli [Allard et al., 1994], Galeras [Zapata G et al., 1997] and others. Our observations at NW Rota-1 (Figure 6b) display a similar scale of variability. Published exceptions to such variability occur only when the number of observations is small and the volcano is not in an eruptive phase (e.g., White Island and other quiescent volcanoes). The absence of variability estimates from studies at many volcanoes (Table 1) underscores the difficulty of obtaining representative long-term measurements. Uncertainties about the CO$_2$ flux from volcanoes thus arise directly from the challenge of conducting continuous, long-term, and integrative gas flux measurements.

[24] Hydroacoustic monitoring of submarine volcanoes makes such measurements practical. Our best estimate of the mean explosive CO$_2$ flux at NW Rota-1, 0.4 ± 0.1 Tg a$^{-1}$, was at the low end of subaerial eruptive volcanoes but substantially greater than almost all measurements at quiescent volcanoes. Its low flux relative to erupting subaerial volcanoes is not surprising given that most subaerial measurements target volcanoes with large calderas and correspondingly large mafic heat sources. Although the instantaneous explosive CO$_2$ flux at NW Rota-1 varied by a factor of ~100 over the deployment, our continuous monitoring provided a robust estimate of the cumulative flux with a relatively small error estimate (25%).

[25] On the global scale, estimates of the total eruptive output of CO$_2$ from subaerial arc volcanoes range from ~70–250 Tg a$^{-1}$ [Hilton et al., 2002; Williams et al., 1992; Mörner and Etiophe, 2002; Fischer, 2008]. The total volcanic supply of CO$_2$ to the ocean is also uncertain. The flux from the entire mid-ocean ridge system has been estimated at ~70–100 Tg a$^{-1}$ [Shaw et al., 2010; Sano and Williams, 1996; Marty and Tolstikhin, 1998], but there is no comparable estimate from submarine arc volcanoes. Our best estimate of CO$_2$ from NW Rota-1, 0.4 ± 0.1 Tg a$^{-1}$, represents ~0.2–0.6% of the annual estimated output of CO$_2$ from all subaerial arc volcanoes, and ~0.4–0.6% of the mid-ocean ridge flux. The only previous estimate of CO$_2$ flux at a submarine arc volcano comes from non-erupting NW Eifuku (in the northern Mariana arc), where liquid CO$_2$ was observed leaking as buoyant droplets at a rate of 0.035 Tg a$^{-1}$ [Lupton et al., 2006]. This flux is comparable to that from the quiescent volcanoes in Table 1 (median =
0.095 Tg a\(^{-1}\)). The CO\(_2\) output from NW Eifuku has a measurable ecological impact as it has been shown to hinder shell development of local chemosynthetic mussels [Tunnicliffe et al., 2009]. The almost continuous eruption of NW Rota-1 since at least 2003 [Embley et al., 2006; Chadwick et al., 2012] injects an even larger and sustained source of CO\(_2\) into the shallow ocean that could have similarly significant biological impacts. Two other submarine volcanoes in the Mariana arc (Nikko and Daikoku) and two in the Tonga-Kermadec arc (Giggenbach and Volcano-1) have recently been shown to also be discharging CO\(_2\)-rich bubbles [Lupton et al., 2008]. Although CO\(_2\) flux has not yet been estimated from these volcanoes, their presence indicates that carbon fluxes from submarine arcs may be significant. Thus although the explosive CO\(_2\) flux from NW Rota-1 is only a small fraction of the flux from subaerial arc volcanoes, the total CO\(_2\) flux from all submarine volcanoes may be a substantial fraction of the flux from subaerial volcanoes and mid-ocean ridges.

Remote hydroacoustic measurements make the monitoring of many volcanoes a quantitative and economically feasible approach to the study of volatile output from submarine arcs. Two other submarine volcanoes with long-lived eruptive activity are already known. Hydroacoustic \(T\) waves have been detected at Monowai, at 25\(^\circ\)45'S on the Kermadec arc, for more than 30 years [Kibblewhite, 1966; Talandier and Okal, 1987; Wright et al., 2008; Chadwick et al., 2008b; Watts et al., 2012]. Lava eruptions and magmatic gas explosions were observed at West Mata, at 15\(^\circ\)5.7'S on the Tonga arc in 2009 [Resing et al., 2011]. More erupting
submarine volcanoes certainly lie undiscovered on other, less explored, arcs. Our results from NW Rota-1 demonstrate that quantifying the global contribution of CO$_2$ from submarine volcanic arcs depends not only on the enumeration of all hydrothermal sources, but also on the discovery of sites of volcanically controlled gas discharge.

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References

Chiодини, Г., и Ф. Фронтини (2001), Carbon dioxide degassing from the Albian Hills volcanic region, Central Italy, Chem. Geol., 177, 67–83, doi:10.1016/S0009-2541(00)00382-X.


