What makes hydromagmatic eruptions violent? Some insights from the Keanakāko’i Ash, Kīlauea Volcano, Hawai’i

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Abstract

Volcanic eruptions at the summit of Kīlauea volcano, Hawai’i, are of two dramatically contrasting types: (1) benign lava flows and lava fountains; and (2) violent, mostly prehistoric eruptions that dispersed tephra over hundreds of square kilometers. The violence of the latter eruptions has been attributed to mixing of water and magma within a wet summit caldera; however, magma injection into water at other volcanoes does not consistently produce widespread tephras. To identify other factors that may have contributed to the violence of these eruptions, we sampled tephra from the Keanakāko’i Ash, the most recent large hydromagmatic deposit, and measured vesicularity, bubble-number density and dissolved volatile content of juvenile matrix glass to constrain magma ascent rate and degree of degassing at the time of quenching. Bubble-number densities (9 × 10^4 – 1 × 10^7 cm^-3) of tephra fragments exceed those of most historically erupted Kīlauean tephras (3 × 10^3 – 1.8 × 10^5 cm^-3), and suggest exceptionally high magma effusion rates. Dissolved sulfur (average = 330 ppm) and water (0.15–0.45 wt.%) concentrations exceed equilibrium-saturation values at 1 atm pressure (100–150 ppm and ~ 0.09%, respectively), suggesting that clasts quenched before equilibrating to atmospheric pressure. We interpret these results to suggest rapid magma injection into a wet crater, perhaps similar to continuous-uprush jets at Surtsey. Estimates of Reynolds number suggest that the erupting magma was turbulent and would have mixed with surrounding water in vortices ranging downward in size to centimeters. Such fine-scale mixing would have ensured rapid heat exchange and extensive magma fragmentation, maximizing the violence of these eruptions. Published by Elsevier B.V.

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1. Introduction

The Keanakāko’i Ash (Fig. 1) is the most recent, best exposed and most thoroughly studied of several hydromagmatic tephra deposits erupted from the summit of Kīlauea volcano, Hawai’i, since the late Pleistocene. Originally thought to have been produced during a single eruption around A.D. 1790 (Dana, 1888), the number of eruptions represented by the deposit has been repeatedly revised (Stone, 1926; Powers, 1948; Christiansen, 1979; McPhie et al., 1990). It is currently thought to have resulted from at least a few eruptions over a period of perhaps a few centuries that ended around 1790 AD (Swanson et al., 1998; D. Swanson, written comm., 2001). The most

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extensive Keanakāko'i units that likely resulted from a single eruption (e.g. unit 2 of McPhie et al., 1990) cover a few hundred square kilometers (R. Christiansen, unpublished data, 2002), similar in area to the deposits of moderately large historical basaltic hydro-magmatic events (e.g. Taal, 1965; Moore et al., 1966). Late Pleistocene tephras at Kīlauea, however, are meters thick in exposures 10 km south of the caldera (loc. 1, Fig. 1a; Easton, 1987; Clague et al., 1995a) and may have originally covered thousands of square kilometers.

The violence of eruptions that produced these deposits is generally attributed to interaction of magma with water, as inferred from their fine grain size, wide range in vesicularity, cross-bedding, accretionary lapilli, and lithic components in the ejecta (Swanson and Christiansen, 1973; Decker and Christiansen, 1984; McPhie et al., 1990; Dzurisin et al., 1995; Clague et al., 1995a). Mastin (1997) hypothesized that water influx was caused by subsidence of the caldera floor below the water table (currently at ~ 490 m depth; Hurwitz et al., 2003). Since the early 1800s, the caldera floor has fluctuated in elevation by hundreds of meters and has come within less than ~ 100 m of the water table on at least three occasions (Mastin, 1997).

1.1. Explosivity

Subsidence of the eruptive vent below the water table would allow water to accumulate within the crater and would guarantee water-magma mixing during summit eruptions; It would not, however, guarantee that such eruptions be large and explosive. Along Kīlauea’s coast, lava commonly enters water without extensive fragmentation or dispersal of tephra (Mattox and Mangan, 1997). Non-violent effusion of lava into water was also observed at Mauna Loa in 1950 (Finch and Macdonald, 1953), at Soufrière of St. Vincent in 1972 (Shepherd and Sigurdsson, 1982) and Kick-’em Jenny in the West Indies in 1974–1978 (Devine and Sigurdsson, 1995), to name a few examples.

We surmise that violence (or non-violence) of these and other hydro-magmatic eruptions is affected not only by the injection of magma into water but also the circumstances under which they mixed. At least, two mechanisms for explosive magma-water mixing have been described during central-vent eruptions: (1) drawdown of magma in a conduit below the water table, followed by influx of water, and (2) jetting of magma through surface water or through a wet crater into which water is entering through the porous walls. Mechanism (1) was noted at Ukinrek Maars, Alaska, when a hydromagmatic explosion, observed from the air on 3 April, 1977, was preceded a few minutes earlier by the draining of a lava lake and collapse of the crater walls (Kienle et al., 1980). The sequence
was similar to well-described non-juvenile steam explosions at Hālemaʻumaʻu Crater in 1924, which took place after a long-lived lava lake drained and hot conduit walls collapsed (Stearns, 1925; Decker and Christiansen, 1984).

The jetting of magma through water was described during continuous-uprush phases of Surtsey Volcano, Iceland (Thorarinsson, 1964; Moore, 1985). These eruptions were characterized by cylindrical columns 100–250 m in diameter at their base (Moore, 1985), that blasted fragmental debris upward at velocities of ~100 m/s and persisted for hours. They occurred when seawater access to the vent was “partially or wholly blocked” (Thorarinsson, 1964), and were generally preceded by intermittent hydromagmatic explosions that repeated with increasing frequency until jetting became continuous. (The cause of intermittent explosions has been the subject of disagreement (Kokelaar, 1983) and may be a variant on mechanism (1)). Water was apparently supplied to the jet by seepage over and through the tephra pile that lined the vent crater. Continuous-uprush phases produced much more tephra and dispersed it more widely than Surtsey’s intermittent explosions (Thorarinsson, 1964, p. 44).

In principle, it should be possible to distinguish mechanisms (1) and (2) using vesiculation textures and dissolved-volatile contents of juvenile clasts. Under mechanism (1), if magma existed within a subaerial lava lake prior to the explosions, it should contain dissolved volatiles in equilibrium with 1 atm pressure. Magma quenched during withdrawal should contain low vesicularity and low bubble-number densities that typify low ascent rates or stagnation (Cashman and Mangan, 1994). Under mechanism (2), juvenile clasts may contain dissolved volatiles equal to or exceeding equilibrium values at 1 atmosphere depending on the pressure and degree of equilibration of the clasts at the time of quenching. Magma that quenched during rapid ascent should exhibit abundant, fine vesicles with high bubble-number densities similar to lava-fountain tephras in subaerial environments (Cashman and Mangan, 1994; Mangan and Cashman, 1996).

In this study, we examine vesiculation textures and dissolved-volatile concentrations in the lower unit of the Keanakākoʻi Ash to constrain magma-water conditions at the time of quenching. The Keanakākoʻi Ash is especially suited to this analysis because the solubility and pre-eruptive volatile content of Kīlauean basalt are well constrained (e.g. Gerlach, 1986; Dixon et al., 1991, 1995; Wallace and Anderson, 1998) and vesiculation textures have been extensively characterized (Mangan et al., 1993; Cashman et al., 1994; Mangan and Cashman, 1996).

2. The Keanakākoʻi deposit

The Keanakākoʻi Ash consists of three major units and several minor ones (Decker and Christiansen, 1984; McPhie et al., 1990). The major units include (1) well-bedded ash and lapilli of inferred hydromagmatic origin containing primarily fresh, juvenile sideromelane glass (the “lower, juvenile-rich beds” of McPhie et al., 1990); (2) massive and cross-bedded ash and lapilli, also of inferred hydromagmatic origin, containing both juvenile glass and lithic fragments (the “middle, mixed beds”); and (3) block fall and cross-bedded surge deposits composed almost exclusively of lithic debris (the “upper lithic beds”). These units are named II, III and V, respectively, in the nomenclature of Decker and Christiansen and 0–4, 6–10, and 11–16 in the nomenclature of McPhie et al. (1990). The units are underlain by a massive reticulite bed (unit I of Decker and Christiansen), which in turn is locally underlain by a gray vitric-lithic ash (D. Swanson, written comm. 2002). The three units are overlain on the northwest and southwest sides of the caldera by a pumice layer a few tens of centimeters thick (unit VI of Decker and Christiansen; the “Golden Pumice” of Sharp et al., 1987). Southwest of the caldera, along the contact between units III and V of Decker and Christiansen are scattered mats of Pele’s hair and tears (unit IV of Decker and Christiansen), inferred to have erupted from a fissure whose lavas are intercalated with the Keanakākoʻi Ash (unit 1790f of Neal and Lockwood, in press). In addition, at the base of the middle, mixed beds (according to McPhie et al., 1990), or in the upper part of unit II (according to Decker and Christiansen, 1984), is a scoria bed up to a few tens of

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2 Water could not truly have been “wholly” blocked as Thorarinsson describes, since he also characterizes these events as hydromagmatic.
centimeters thick (unit 6 of McPhie et al.) whose axis of dispersal extends to the southeast, in contrast to the southwestward dispersal of most units in the Keana-kako’i deposit.

2.1. Evidence for depositional breaks

In one of the first written discussions of the deposit, Dana (1888) assumed it resulted from a single eruption in 1790 AD that was known from oral accounts (Ellis, 1827; Dibble, 1843). Early 20th century authors attributed only the upper lithic unit to the eruption in 1790 and lower units to earlier eruptions (Hitchcock, 1909; Powers, 1916; Stone, 1926; Stearns and Clark, 1930; Wentworth, 1938; Finch, 1947; Powers, 1948). Later, Christiansen (1979) and Decker and Christiansen (1984) proposed that all members between the basal reticulite and the Golden Pumice erupted during a single sequence in or around 1790 AD. McPhie et al. (1990) concurred but noted three horizons of erosion and reworking that suggested pauses of unknown and possibly long duration. Most recently, Swanson et al. (1998) cited organic horizons, water-erosion features, archaeological artifacts and carbon dates as evidence for at least few large eruptive sequences, beginning shortly after appearance of the present-day caldera in the late 15th century and ending around 1790 AD (D. Swanson, written comm., 2001). Horizons that show the strongest evidence for non-depositional periods lie immediately above the basal reticulite, slightly below and above unit 6 of McPhie et al., and at the base of the upper lithic beds (McPhie et al., 1990).

2.2. Field locations

Our study concentrates on unit II of Decker and Christiansen, using samples collected in Sand Wash (locations 61 and 84, Fig. 1b) and at a fissure a kilometer north of Sand Wash (location 62, Fig. 1b). At these locations, unit II contains three subunits: two coarsening-upward sequences of well-bedded ash and lapilli (units IIA and IIB) and a sequence of fine ash beds with sparse lapilli horizons (unit IIC; Figs. 2 and 3). Conformable bedding within subunits IIA and IIB suggests that each was deposited during a single eruptive pulse or rapid series of pulses with no significant pauses. The contacts at the top of units IIA and IIB are sharp and truncate underlying beds at a low angle (Fig. 3b), suggesting some centimeters of erosion, perhaps by surges or wind, between depositional pulses. The upper contact of unit IIA lacks recognizable evidence of water erosion, suggesting the pause was short and without rain. The upper contact of unit IIB exhibits scarce water-erosion rivulets and locally overlies cross-bedded and apparently wind-reworked debris from underlying units (McPhie et al., 1990), suggesting a longer pause of unknown duration.

McPhie et al. (1990, p. 351) speculated that juvenile-rich beds resulted from repeated explosions that began with vesiculation-driven magma ascent and ended with drainback of degassed magma. Mastin (1997), used quantitative vesicularity and bubble-number density measurements to argue that some horizons in units IIA2 and IIB2 involved sustained, high magma discharge. Whether other beds involved sustained high discharge was not investigated.

3. Sample collection and analysis

We collected samples by trowel from the outcrop (Tables 1 and 2) and measured the volume percent of fresh glass, altered glass, olivine and lithic fragments by point-counting grains mounted in thin sections (≈ 400 grains per section; Table 3). We analyzed dissolved-volatile concentrations using the electron microprobe for sulfur and Fourier transform infrared (FTIR) spectrometry for water and CO2. Juvenile glass in the deposit ranges in vesicularity from 0% to >80%, requiring us to prepare samples for FTIR analysis using two methods: (1) slightly-to-moderately vesicular clasts, 1–2 mm in diameter were embedded in a 2.5-cm diameter wafer of epoxy that was then ground and polished, flipped, remounted on a glass slide, then ground to a thickness of 70–165 μm, and

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3 Unpublished grain-size analyses give $md_f = -0.64$ to 3.5, $\sigma_f = 1.22–1.86$ for ash in the lower half of these sequences and $md_f = -0.38–0.62$, $\sigma_f = 1.51–2.10$ for ash and lapilli in the upper half.

4 $md_f = -0.67–3.71$, $\sigma_f = 1.31–2.53$ based on unpublished grain-size analyses.
polished on the other side; (2) highly vesicular lapilli were ground with a mortar and pestle and flakes ~ 1 mm in diameter were handpicked, mounted in epoxy and doubly polished as described above.

3.1. Dissolved volatile analysis

Before removing the polished grain mounts from their glass slides, each fragment was analyzed for sulfur by electron microprobe using the JEOL JXA-8900 Electron Probe Microanalyzer at the USGS Menlo Park Laboratory. These samples were a subset of more than 23 thin sections containing tephra fragments analyzed for sulfur during 6 microprobe sessions of 1–2 days each. For reference, we also analyzed clasts from non-hydromagmatic units I, IV, VI and the 1959 Kilauea Iki tephra (Table 1). Values of S and K2O in Table 4 and the major-element data (see supplementary data in the online version) represent averages of two to four analyses per clast (exact numbers are listed online). During each session, we used an acceleration voltage of 15 keV, a beam current of 25 nA, a beam diameter of 10–15 μm, a barite standard for sulfur, and a peak counting interval of 80 s. Results for elements other than sulfur, and standards used, are provided in the online supplement. Of 343 analyses, 330 gave totals between 97% and 101%.

The precision of sulfur concentrations (ppm) is estimated from counting statistics and reproducibility of USNM basaltic glass standard VG-2 (Jarosewich et al., 1979), which is similar in composition to the analyzed glasses. The standard deviation in sulfur values of VG-2 glass measured during individual microprobe sessions is 57 ppm (the half-width of the error bar in Fig. 2); however, mean VG-2 sulfur analyses for a given session range from 1080 to 1350 ppm, suggesting some drift between sessions. Sulfur calibration numbers were therefore adjusted so that the average VG-2 analysis in each session equals 1340 ppm, the value obtained by Dixon et al. (1991).

After microprobe analysis, the grain mounts were removed from their glass slides and bubble-and-crystal-free regions of glass were targeted for FTIR analysis using apertures above and below the sample. Measurements were performed using a Nicolet 5 Magna 750 spectrometer with an attached SpectraTech® Analytical-IR microscope that utilizes a liquid-N2-cooled MCT-A detector. We collected 512 scans per analysis on each sample and on backgrounds following each sample analysis. The measured absorbance of infrared radiation is directly proportional to the concentration of H2O in the glass, adjusted for sample density and thickness. Weight percent H2O was calculated according to Eq. (15) of Ihinger et al. (1994) for the OH– stretch at 3570 cm⁻¹, assuming a glass density of 2700 ± 100 g l⁻¹, an extinction coefficient (ε) of 63 ± 3 l mol⁻¹ cm⁻¹ and sample thickness as measured by a Mitutoyo® Digital Micrometer. Uncertainty for ε, sample thickness and density are all about 5%, so that propagated errors are all between 10% and 15% relative.

We detected no molecular H2O (peak at 1630 cm⁻¹) or CO3⁻ (1515 and 1430 cm⁻¹); we analyzed two clasts in sample 517B twice, collecting 1024 scans during the second analysis, to verify the absence of these peaks. The detection limit of dissolved CO3⁻ is ~ 50 ppm, suggesting that the melt degassed to less than ~ 5 MPa (using solubility relations of Dixon et al., 1995), or about 600 m depth (assuming a pressure gradient of 25 MPa/km). The absence of a molecular water peak is consistent with the experimentally measured speciation of dissolved water in basaltic glass at this concentration (Dixon et al., 1995). Meteoric water adsorbed into the clasts at magmatic temperature could also speciate of OH⁻, but water absorbed at lower temperature would likely remain in molecular form (P. Wallace and J. Dixon, written comm., 2001). If water were absorbed into the clasts during or after the eruption, we would expect it to be absorbed over a range of temperatures and to leave at least some water in molecular form. The complete absence of molecular H2O in these clasts therefore suggests to us that all dissolved water is magmatic.

3.2. Vesicularity and bubble-number density

To obtain bubble-number densities, all of the dense, hand-picked clasts analyzed by FTIR were photographed in thin section under reflected light. Printed photomicrographs were overlain with transparencies and the outlines of the clasts, and of internal bubbles, were traced in ink. The inked transparencies

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5 Use of trade names in this document does not imply endorsement by the USGS.
were then digitally scanned and the clast diameter, area, and the number and size distribution of bubbles were computed using the software Scion Image®. The number of bubbles per clast ranged from zero to several dozen; too few to derive meaningful size-distribution statistics but enough to estimate the order-of-magnitude bubble-number density.

The highly vesicular clasts were destroyed when ground by mortar and pestle and could not be analyzed for bubble-number density. Prior to grinding, however, their vesicularity was measured by weighing the samples in and out of water following the method of Houghton and Wilson (1989). We calculated clast density \( \rho \) using the formula:

\[ \rho = \rho_w(\omega - \omega_0) \]

where \( \rho_w \) is the density of water (1000 kg/m\(^3\)), and \( \omega \) and \( \omega_0 \) are the measured weights in air and in water, respectively. The volume-fraction gas \( v_g \)—i.e. the fraction of the clast volume composed of bubbles—was calculated using the formula:

\[ v_g = \frac{1 - \rho/\rho_g}{1 - \rho_g} \]

where \( \rho_g \) is the density of basaltic glass, taken as 2700 kg/m\(^3\). We calculated the equivalent radius \( r \) of each lapillus (assuming a spherical shape) from the formula:

\[ r = \left( \frac{3v_g}{4\pi} \right)^{\frac{1}{3}} \]

Densities are generally repeatable to ±10%.

During the course of our analysis, we observed that vesicularity varied systematically from one stratigraphic unit to the next and from ash to lapilli-sized clasts within a given unit. To quantify these variations, we compared \( v_g \) from 3000 clast density measurements, published earlier (Mastin, 1997), with \( v_g \) from ~3000 ash-sized grains, estimated visually in thin section by comparing the abundance of bubbles to diagrams of modal percentage (Compton, 1985, Appendix 3). Diagrams in Compton (1985) are given for percentages of 0.5, 1, 1.5, 3, 5, 7, 10, 15, 10, 25, 30, 35, 40, 45 and 50. We consider those intervals to indicate the resolution of the estimation technique. Results of these measurements, as well as microprobe data and lapilli clast density, are posted online.

4. Results

Sulfur values in matrix glass from unit II (Fig. 2) are on average 330 ppm with a standard deviation of 80 ppm—significantly above the 100–150 ppm typically measured for Kilauean glass that has equilibrated to atmospheric pressure (Swanson and Fabbri, 1973; Mangan and Cashman, 1996; Thornber, 2001). Glass in non-hydromagmatic units I, IV, VI and the Kilauea Iki tephra, all contain dissolved sulfur values consistent with equilibrium degassing to 1 atm pressure. Dissolved sulfur in glass inclusions in olivine ranges from about 250 to 1530 ppm, with most around 1000–1200 ppm (Fig. 2). Considering the middle of this range to represent undegassed inclusions and 100 ppm to be totally degassed, most hydromagmatic glass fragments are about 70–80% degassed in sulfur.

The concentration of dissolved water in juvenile glass also measurably exceeds equilibrium saturation under atmospheric conditions (Fig. 2). Sulfur and water values (Table 4) are moderately correlated \( (r^2 = 0.69) \) supporting the view that the water is magmatic in origin. Measured H\(_2\)O concentrations correspond to quench pressures (under equilibrium H\(_2\)O saturation) of ~0.3–2.3 MPa (Fig. 2), equivalent to 10–100 m depth at lithostatic pressure (assuming a pressure gradient of 25 MPa/km), or ~25–
250 m at hydrostatic pressure (100 MPa/km). The amount of H$_2$O lost to degassing can be assessed from the ratio of H$_2$O/K$_2$O (Table 4, Fig. 4). Both of these components are incompatible in a crystallizing Kilaeuan melt and, in the absence of degassing, their ratio should remain constant at about 1.3 (Dixon et al., 1991; Wallace and Anderson, 1998). Average K$_2$O (0.44 ± 0.05 wt.%) in clasts from hydromagmatic units imply pre-eruptive water content of 0.57 ± 0.06 wt.%. Measured H$_2$O/K$_2$O ratios (0.39–1.02) suggest that the melt had lost 25–80% of its H$_2$O by the time it quenched.

Fig. 5 shows an inverse relationship between dissolved water and $v_g$, which is consistent with a
scenario in which clasts were quenched during active vesiculation. Dissolved H$_2$O is lowest in the highly vesicular clasts of unit IIB2. Nearly all data lie below the line that represents closed-system gas exsolution with total water (dissolved plus exsolved) equal to 0.55 wt.%. Large, highly vesicular lapilli appear to have suffered more gas loss than smaller, less vesicular ash.

These data clearly show that hydromagmatic glass did not fully degas in sulfur or water. In contrast, non-hydromagmatic clasts are completely degassed in sulfur. We did not measure dissolved water in non-hydromagmatic tephra, but tephra from phase 1 of the Kilauea Iki eruption, analyzed by Wallace and Anderson (1998, fig. 1), are in equilibrium with 1 atm pressure (phase 1 tephra erupted before degassed lava

### Table 1

Summary of samples collected and analyzed for this paper

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<td>67</td>
<td></td>
<td>mp, d</td>
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Abbreviations in the right column include clast density measurements (“d”), vesiculility measurements from thin section (“ves”), microprobe glass analyses (“mp”), FTIR analyses (“FTIR”), bubble-number density (“bnd”) and point counts (“pc”).
from lava ponds entered the vent). We therefore infer that water, like sulfur, is elevated due to rapid quenching by external water. Whether these clasts quenched at elevated pressure or at 1 atm before they could degas to equilibrium is the subject of later discussion.

4.1. Vesiculation textures

In general, pyroclasts of a given size are less vesicular in units IIA1, IIB1, more vesicular in units IIA2 and IIB2. Within individual units, average vesic-
ularity increases with average grain size (Fig. 6). Although $v_g$ was estimated for the ash-and lapilli-sized clasts using two different methods, we do not think that sample bias in one or both methods is responsible for the variation. Measurements made using both methods in the same size range (for example, in unit IIIB2, 2–3 mm diameter clasts) yield the same range in $v_g$. Moreover, estimates by visual inspection are similar to results from the more accurate image processing technique (triangles, units IIA2, IIB1 and IIC1).

For clast diameters larger than a few tenths of a millimeter, lower values of $v_g$ do not appear to result simply from fragmentation at a scale smaller than the average bubble diameter. Many clasts of this size (e.g. Fig. 7) contain bubbles smaller than their own diameter. The lower average dissolved water concentration of the larger, more vesicular clasts (Fig. 5) and the incomplete degassing of all hydromagmatic clasts suggest that the smaller clasts were at an earlier stage of vesiculation.

Bubble-number densities in unit II (Table 4, Fig. 8) range from $9 \times 10^4$ to $1 \times 10^7$ cm$^{-3}$ similar to those measured in unit IIB2 by Mastin (1997). Though the earlier data were obtained only from highly vesicular lapilli, the new data come from ash-sized clasts with vesicularity ranging down to about 20%. All these values exceed number densities obtained from vent samples of effusive eruptions (3000–5000 cm$^{-3}$; Mangan et al., 1993) and most exceed values measured in lava-fountain tephra from Pu‘uO‘o (19,000–180,000 cm$^{-3}$; Mangan and Cashman, 1996).

Unlike our Keanakāko‘i clasts, the lava-fountain tephra rose many seconds in a hot plume before cooling, raising the question of whether bubble coalescence and escape could account for their lower number densities. We do not think so. Such processes recognizably alter texture by increasing vesicularity in clasts interiors and reducing it on margins (Walker and Croasdale, 1972). Mangan and Cashman (1996) avoided this complication by selecting clasts that were texturally homogeneous. From plots of log bubble number versus bubble diameter, they inferred that only about 13–40% of bubbles coalesced from two or more smaller ones—not enough to account for the <100 $\times$ lower number densities of lava-fountain clasts relative to Keanakāko‘i clasts.

5. Implications

These high bubble-number densities of unit II suggest rapid magma ascent and eruptive activity

![Fig. 4. Dissolved H$_2$O versus K$_2$O in matrix glass from Keanakåko‘i Ash. Melt that has not lost H$_2$O to degassing plots along the line H$_2$O/K$_2$O = 1.3, while completely degassed melt would plot along a line having zero slope and an intercept at H$_2$O = 0.09 wt.%. Intermediate degrees of degassing are represented by lines having intermediate slopes, intersecting the H$_2$O/K$_2$O = 1.3 line at H$_2$O = 0.09 wt.%,](image)

![Fig. 5. Weight percent dissolved water versus volume fraction gas in melt. Squares are measurements (symbol size is proportional to clast diameter). Lines give volume-fraction gas versus dissolved water for closed-system degassing for a total water content (dissolved plus exsolved) of (top to bottom) 0.55, 0.44, 0.33 and 0.22 wt.%, respectively. The lines represent loss of 0%, 20%, 40% and 60% of the original gas, respectively.](image)
more akin to continuous-uprush events at Surtsey than to drainback-induced explosions of Ukinrek or Halema’uma’u. Moreover, bubble-number data suggest that all unit II subunits erupted rapidly, even those containing only moderately or poorly vesicular tephra. Whether all hydromagmatic phases during this period were violent is less clear; less violent phases may have failed to eject debris out of the caldera where it could be preserved. The cause of repeated, high-flux rate eruptive pulses is not known but could be related to caldera subsidence.

High ascent rates could explain some but not all of the fragmentation and dispersal that we associate with

Fig. 6. Volume fraction gas ($v_g$) versus clast diameter for ash and lapilli fragments collected from most juvenile-bearing units in the Keanakāko’i Ash. Dots represent $v_g$ estimated by visual comparison of clasts with modal area plots (e.g. Compton, 1985, pp. 366–367). Crosses are $v_g$ of lapilli published earlier (Mastin, 1997). Triangles are obtained from digital image processing of photomicrographs.

Fig. 7. Moderately vesicular ash from unit IIA2.
explosivity. Dry Kīlauean high-flux eruptions (lava fountains) certainly produce more fragmental debris than low-flux eruptions, but the fragmentation process in dry fountaining is primarily gas expansion. In the Keanakākoʻi Ash, many juvenile clasts are poorly vesicular and blocky in shape, implying fragmentation by thermal fracture rather than bubble expansion. We surmise that thermal fractures develop on clast surfaces when they contact water or steam. Fracture spacing, which determines the final grain size, is influenced by the temperature contrast at the interface. Rapidly formed surfaces are hotter and likely to host more finely-spaced thermal cracks than surfaces formed by slow, taffy-like stretching. The glassy rinds between these fractures probably shed as the interface strains and deforms (Fig. 9a); the more rapid the deformation, the more extensive and fine the shedding of debris. Fragmentation by this mechanism should be finest and most extensive, and heat-transfer rates from magma greatest, when ascent rates are high and when melt is violently torn and deformed as it enters a lake or the atmosphere. Fine particles are likely to originate on the clast margins, which cool before clast interiors and should, as our results suggest, contain the highest dissolved volatile concentration.

5.1. The importance of turbulence

We hypothesize that the degree of fragmentation is controlled by the turbulence of mixing. Photographs of continuous-uprush phases at Surtsey (Thorarinson, 1964) show fully turbulent jets with exit velocities exceeding 100 m/s (Moore, 1985). Turbulence occurs when perturbations along the jet margins are not damped by viscous forces and develop into eddies and vortices (White, 1991, p. 471). In circular jets, the jet margin becomes unstable at Reynolds numbers \( Re \) as low as 4 (Kaplan, 1964), though full turbulence at the vent is not ensured unless flow is turbulent in the upper conduit (i.e. \( Re > \sim 2300 \), where \( Re = \rho u D/\mu \), \( \rho = \) fluid density, \( u = \) velocity, \( D = \) conduit diameter and \( \mu = \) fluid viscosity. During basaltic lava-fountain eruptions, flow in the upper conduit may be either laminar or turbulent (Fig. 10), with laminar or unstable flow during less vigorous events and full turbulence during more vigorous ones.

Engineering tests have long established that jet turbulence increases rates of heat transfer and chemical reaction between mixing fluids by orders of magnitude (e.g. Burmeister, 1983, p. 394). Higher rates of turbulent shear cause more efficient mixing as eddies progress to ever smaller scales. The finest-scale vortices dampen out at the so called Kolmogorov length scale \( \kappa \approx \delta U/3 \), where \( \delta = \) boundary-layer thickness and \( U = \) mean velocity, Kolmogorov, 1941; White, 1991, p. 471). For magma–steam–water mixtures moving at tens of meters per second (based on observed lava-fountains; Mangan and Cashman, 1996) within a boundary layer a meter or two wide, the Kolmogorov scale would be decimeters or smaller depending on whether the viscosity and density of gas, water or magma are used in the calculation. The scale of mixing is fully sufficient to incorporate centimeter-to decimeter-sized volumes of water into melt as required to generate molten fuel-coolant interactions (Buettner and Zimanowski, 1998).

5.2. Water depth and volatile equilibration

Turbulent entrainment relations can be used to constrain water depth from the deposit characteristics. Immediately above the jet exit ("A", Fig. 9b), jets entrain ambient fluid within a turbulent boundary layer...
Fig. 9. (a) Schematic illustration of the fragmentation sequence of a magma blob entrained in turbulent flow. Fragmentation is assumed to occur by extension and shear of the droplet, supplemented by growth of cooling fractures and shedding of the glassy rind on clast margins. (b) Velocity profiles and boundary layers around an axisymmetric turbulent jet exiting into ambient fluid.
whose thickness increases linearly until, at a height \( h \) of several times the vent diameter \( D \), it has penetrated into the potential core of the jet (“B”, Fig. 9b). Further downstream (“C”, Fig. 9b), the jet diameter \( d \) increases linearly and centerline velocity decreases inversely with distance. Jet momentum remains constant with distance but total mass flux increases as the jet continues to entrain surrounding fluid.

In experiments using jet and ambient fluids of similar properties (e.g. water or saline solution injected into water), the rate of entrainment follows the relation (Eq. (6) of Wilson et al., 1995):

\[
M_j = \pi \tilde{D} h \rho_j u_j \sqrt{\frac{\rho_j}{\rho_a}}
\]

where \( M_j \) is the mass per unit time of ambient fluid added to the jet between the jet exit and height \( h \); \( \tilde{D}, \rho_j \) and \( u_j \) are the average jet diameter, density and velocity between the exit and \( h \); \( \rho_a \) is density of the ambient fluid; and \( x \) is an empirical entrainment coefficient, which is roughly 0.08 in the self-similar region (“A”), Fig. 9b and somewhat less in the initial region (“A”).\(^6\) The mass influx calculated by this method may be compared with the mass flux of the jet, \( M_j = \pi \tilde{D}^2 \rho_j u_j / 4 \).

For a jet of magma and gas exiting into water, realistic values for \( D (\sim 5 \text{ m}, \text{ based on conduit modeling of observed flow rates}; \text{Mastin, 2002}), \rho_a (1000 \text{ kg/m}^3), \rho_j (500–1500 \text{ kg/m}^3 assuming a moderately to highly vesicular magma), u_j (50–100 m/s based on observed lava-fountain velocities; \text{Mangan and Cashman, 1996}) \) and \( x (0.06–0.08) \), suggest that the mass flux of water entrained will equal the mass flux of the magma–gas mixture when water depth is a few to several vent diameters. A magma:water ratio of 1:1 is much less than the optimal ratio \(( \sim 3–5.5:1) \) for converting thermal to kinetic energy (\text{Wohletz, 1986}), and would leave most (60–80\%) of the water in liquid from (\text{Mastin, 1995}), very likely causing deposits to remobilize into watery slurries. Optimal magma:water ratios would likely require water depths on the order of a single vent diameter, perhaps less. Significantly, at such shallow depths, water would be entrained only into the boundary layer and would mix with magma in the core of the jet only after escaping into the atmosphere. This inference is consistent with observations from Surtsey, in which the margins of continuous-uprush jets contained black ash and white steam, but the jet core glowed a visible red color at night (\text{Thorarinsson, 1964; Moore, 1985}).

Although accretionary lapilli are common in fine-grained beds of unit II, coarse-grained beds of units IIA2 and IIB2 contain none, and there is no evidence for soft-sediment deformation or other remobilization. We therefore suspect that these coarse-grained beds erupted through water depths on the order of a vent diameter. Their elevated volatile concentrations

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\(^6\) Magma injection into water involves phase changes and viscosity contrasts that may result different entrainment coefficient than the \( x = 0.08 \), which is derived from jets of similar fluids. Therefore, the implications of these relations are only presented semi-quantitatively here. Experimental studies using magma and water (or analogue fluids) are required for more quantitative analysis.
therefore probably reflect disequilibrium rather than elevated quenching at pressure in a deep lake. Could the disequilibrium have resulted from high ascent rates alone? We suspect not, as non-hydromagmatic lava-fountain tephras (Fig. 2) appear to have equilibrated before cooling. High volatile concentrations in these glasses reflect both rapid ascent and rapid quenching by water at the surface. In this sense, these deposits provide a freeze-frame of degassing at the time of eruption that is unattainable in non-hydromagmatic tephra.

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