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Assessing eruption column height in ancient flood basalt eruptions

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Abstract: A buoyant plume model is used to explore the ability of flood basalt eruptions to inject climate-relevant gases into the stratosphere. An example from the 1986 Izu-Oshima basaltic fissure eruption validates the model's ability to reproduce the observed maximum plume heights of 12 – 16 km above sea level, sustained above fire-fountains. The model predicts maximum plume heights of 13 – 17 km for source widths of between 4 – 16 m when 32% (by mass) of the erupted magma is fragmented and involved in the buoyant plume (effective volatile content of 6wt%). Assuming that the Miocene-age Roza eruption (part of the Columbia River Basalt Group) sustained fire-fountains of similar height to Izu-Oshima (1.6 km above the vent), we show that the Roza eruption could have sustained buoyant ash and gas plumes that extended into the stratosphere at ~ 45°N. Assuming 5 km long active fissure segments and 9000 Mt of SO_2 released during explosive phases over a 10-15 year duration, the ~ 180 km of known Roza fissure length could have supported ~ 36 explosive events/phases, each with a duration of 3 - 4 days. Each 5 km fissure segment could have emitted 62 Mt of SO_2 per day into the stratosphere while actively fountaining, the equivalent of about three 1991 Mount Pinatubo eruptions per day. Each fissure segment could have had one to several vents, which subsequently produced lava without significant fountaining for a longer period within the decades-long eruption. Sensitivity of plume rise height to ancient atmospheric conditions is explored. Although eruptions in the Deccan Traps (~ 66 Ma) may have generated buoyant plumes that rose to altitudes in excess of 18 km, they may not have reached the stratosphere because the tropopause was substantially higher in the late Cretaceous. Our results indicate that some flood basalt eruptions, such as Roza, were capable of repeatedly injecting large masses of SO_2 into the stratosphere. Thus sustained flood basalt eruptions could have influenced climate on time scales of decades to centuries but

the location (i.e., latitude) of the province and relevant paleoclimate is important and must be considered.

Keywords: flood basalt; climate; Roza; sulfur dioxide; Columbia River Basalt Group; plume heights

1. Introduction

There is an intriguing age correlation between several continental flood basalt (CFB) provinces emplaced in the past 300 Ma with mass extinction events (e.g., Wignall, 2001; Courtillot and Renne, 2003; Kelley, 2007). The link between CFB volcanism and mass extinctions may be due to gas release from the magmas or magma-sediment interactions potentially leading to environmental changes (Self et al., 2014, Schmidt et al., 2014). Recent studies have shed light on variations in eruption rate over time and the spatial distribution of eruptive vents for the 14.7 Ma Roza Member of the Columbia River Basalt Group (CRBG) (e.g., Brown et al., 2014). These new data allow more robust assessment of magmatic gas contributions to the atmosphere by CFB volcanism.

The effects of volcanism on climate are complex and occur over a range of time scales, from days to possibly centuries (Robock, 2000; Timmreck, 2012). Sulfur species (sulfur dioxide, SO_2, and hydrogen sulfide, H_2S) are the primary volcanic volatiles known to impact climate. Most basaltic eruptions release SO_2 (Sharma et al., 2004) and most is known about its relevance for climate when injected into the stratosphere. Sulfuric acid (H_2SO_4) aerosols derived from SO_2 and H_2S can have long residence times (1 – 3 years) in the stratosphere, where these particles scatter and absorb incoming solar and thermal infrared radiation, resulting in a net cooling at the surface (e.g., Robock, 2000). Release of volcanic CO_2 may have a limited effect on climate (Self, et al., 2006). However, magmatic interaction with sediments, may also play a role (Self et al., 2014).

Examples of large, historic, silicic explosive eruptions that have impacted climate on the 1–2 year timescale through the injection of large amounts of SO_2 into the stratosphere include Tambora (1815), Krakatau (1883), Agung (1963), El Chichón (1982), and Pinatubo (1991). The high-latitude (65^{o}N) basaltic fissure eruption of Laki in 1783-1784 (Thordarson and Self, 1993; Schmidt et al. 2010) injected SO_2 into the upper troposphere and lower stratosphere through repeated sub-Plinian explosive phases over a period of eight months, affecting climate in the northern hemisphere for up to two years (Thordarson et al., 1996; Highwood and Stevenson, 2003; Oman et al., 2006a; 2006b; Schmidt et al., 2012).

Masses of SO_2 injected into the atmosphere by historic explosive volcanic eruptions range from ~ 7 Mt (= 7 Teragram) by El Chichón in 1982 (Bluth et al., 1992), to ~ 20 Mt by Pinatubo in 1991 (Bluth et al., 1992), and ~ 60 Mt by Tambora in 1815 (Self et al., 2004). Up to 122 Mt of SO_2 was emitted by the Laki eruption (Thordarson, et al., 1996), with over 90 Mt injected into the upper troposphere and lower stratosphere over a 5 month period. Volcanic eruptions are also capable of redistributing large volumes of water from the lower atmosphere into the stratosphere (Glaze et al., 1997). Models indicate that a 25 km-high eruption column rising through a wet, tropical atmosphere can transport up to 4 Mt of H_2O per hour. (“Plume” describes the vertical eruption column and downwind ash/gas cloud.) In this case, ~30% of the water vapor in the plume at its maximum height is derived from the erupted magma and ~70% is entrained while passing through the moist lower atmosphere (Glaze et al., 1997).

Based on the atmospheric and environmental impacts observed following the Laki eruption, some have speculated that large CFB eruptions (e.g., CRBG, peak 16-15 Ma; Deccan Traps peak ~ 67-65 Ma) may have supplied large masses of SO_2 and other gases to the upper troposphere, or possibly even the stratosphere (e.g., Stothers et al., 1986; Woods, 1993a; Thordarson and Self, 1996; Chenet et al., 2005; Self et al., 2005). Early attempts by Stothers et al. (1986) and Woods (1993a) at modeling CFB plumes suggested that near-stratospheric heights could be attained by high-intensity basaltic eruptions but there were many simplifications. CFB lava flow-fields, each the product of one sustained eruption over decadal time-scales, are made up of multiple lava flows that are proposed to have erupted with volumetric flow rates similar to the maximum estimated for Laki (e.g., Self et al., 1998; Self et al., 2006). Long-duration effusive basaltic eruptions, such as the current >30 year eruption of Pu`u `Ō`ō, Hawaii, are consistent with much larger CFB events that may have been active over decades to possibly hundreds of years. These larger CFB events may have injected as much as 1000 Mt of SO_2 into the atmosphere annually (Self et al., 2005). Despite the long overall duration of CFB province emplacement (1 to a few Ma), it is likely that this activity was characterized by long periods of inactivity, punctuated by shorter duration eruptive phases lasting decades to possibly centuries.

We investigate the possible delivery of climate-relevant gases (SO_2 and H_2O) into the atmosphere from CFB eruptions by combining numerical modeling and volcanological datasets for the 14.7 Ma Roza flow of the CRBG (Thordarson et al., 1996; Brown et al., 2014). In particular, we place constraints on the mass of SO_2 that such eruptions could have injected into the stratosphere. This is the first application of a buoyant plume model to ash and gas plumes sustained above a fire-fountain.

2. Basaltic gas-ash plumes

The key to having an impact on climate for a basaltic eruption is the ability to loft climate-relevant gases into the stratosphere and to sustain the supply over many years, even intermittently. Although not common, sustained buoyant plumes associated with large basaltic fire-fountain events have been observed. An example is the Pu`u `Ō`ō eruption at Kilauea, Hawaii (Figure 1); early episodes of the eruption in 1983 and 1984 generated fire-fountains, with typical heights of 100–200 m, and occasionally as high as ~400 m (Wolfe et al., 1988), which sustained gas and ash plumes of 5-7 km height above sea level (ASL). The 1984 eruption of Mauna Loa, Hawaii, produced somewhat larger fire-fountains (up to 500 m high) along a 2 km-long active fissure that generated a buoyant plume estimated to rise to 11 km ASL (7.5 km above the vent) (Smithsonian Institution, 1984). Much larger fire-fountains were documented during the 1986 eruption of Izu-Oshima, Japan, and, indirectly, the 1783-1784 eruption of Laki, Iceland. At Izu-Oshima, fire-fountains 1.6 km high were observed to feed an ashy sub-Plinian plume that reached 16 km ASL (Endo et al., 1988; Sumner, 1998; Mannen and Ito, 2007). Miyakejima, Japan, also produced 12-km-high plumes during a basaltic fissure eruption in 1983, but the exact source of the plumes is not known (Aramaki et al., 1986). Fire-fountains at Laki, estimated to have reached 0.8 – 1.4 km in height, were observed from afar to sustain eruption columns of up to 15 km altitude (Thordarson and Self, 1993; Oman et al., 2006a). These historic eruptions can be used to evaluate plume-rise models for application to convecting, buoyant ash plumes sustained above fire-fountains.

3. Modeling plume rise above fire-fountains

Buoyant plume dynamics play an important role in the ability of explosive volcanic eruptions to supply climate-relevant gases to the atmosphere. Several models are available for buoyant plume rise from a central vent (e.g., Woods, 1988; Woods, 1993b; Glaze et al., 1997; Mastin, 2007) or linear vent (Stothers et al., 1986; Stothers, 1989; Woods 1993a; Glaze et al., 2011). The results of these studies are largely contained in two comprehensive models, *Plumeria* (Mastin, 2007) and *4C* (Glaze et al., 1997). However, the Woods (1988; 1993b) model, upon which *Plumeria* is based, includes two fundamental inconsistencies in the formulation, resulting in key differences with *4C* (Glaze and Baloga, 1996; Glaze, 1999). The net effect of these differences is a 4-7% overestimate of maximum plume height using the Woods (1988; 1993b) approach (see Appendix).

For plumes with large mass flux, the dynamics are almost completely dominated by the buoyancy driven region. Thus, the *4C* model as used here is focused on the buoyancy-driven regime and does not include a "jet region" near the vent, as defined by Woods (1988). This is consistent with modeling ash-gas plumes that are driven by fire-fountains (or even roiling lava lakes) as the starting point: the plume of interest is essentially a "hot-plate zone" of semi-consistent height (usually ≤ 1.5 km above the local surface for a big fountain).

The basic input data required for *4C* are the mass flux (defined by the source area, and eruption velocity), eruption temperature, and gas mass fraction of the erupting magma. Variations in the atmospheric temperature and pressure may also affect estimated maximum plume heights (Woods, 1993b; Glaze et al., 1997; Glaze, 1999; Glaze and Baloga, 1996). For applications to ancient eruptions, small variations in atmospheric composition, for example, differences in CO_2 concentrations between 65 Ma ago (if modeling the Deccan Traps eruptions) and now, may also influence atmospheric temperature structure, and thus predicted plume behavior (see Section 6.1). The primary *4C* model output is predicted maximum plume height along with mass fluxes of water and SO_2 at the maximum plume height altitude.

Buoyant plumes driven by basaltic fire-fountains differ from most other explosive volcanic plumes in that most of the solid material falls immediately back to the surface. The resulting plumes are, therefore, relatively gas-rich (i.e., ash-poor) in comparison to their more silicic counterparts. Typical values for bulk magma volatile contents range from 2–5wt%. However, gas fractions of 70–94wt% have been measured in explosive basaltic "bursts" at Stromboli (Chouet et al., 1974).

Figure 2 illustrates the importance of ash on maximum rise height. The simplest case is a steam plume with temperature of 100 $^{\circ}$C, and initial velocity of 50 m/s. The maximum plume height is increased 25-30% by increasing the eruption temperature to 925°C (thus increasing the density difference driving buoyancy). Addition of a relatively small amount of ash to the plume (50wt% compared with 95-98wt% in a typical explosive silicic eruption; where ash particles are 4-5 orders of magnitude more dense than gas) increases the mass flux at the source (keeping the vent size the same) and results in an overall increase in maximum plume height. In addition to increasing the mass flux (known to be correlated with maximum height, e.g., Sparks et al., 1997), solid particles have a much higher heat capacity than water vapor, resulting in the ability to keep the plume warm and buoyant, despite the higher initial bulk plume density.

One of the keys to modeling the behavior of buoyant plumes is to estimate the relative amounts of ash and volatiles in the plume. Previous studies (Stothers, 1989; Woods, 1988; 1993a; 1993b; Glaze et al., 1997; 2011) have used bulk magmatic volatile contents (~2-5wt%). However, for ash plumes sustained above fire-fountains, this approach will underestimate the ratio of volatiles to solid ash. To better estimate this ratio, there are two processes to consider.

First, because it is easier for volatiles to separate from the melt phase in basaltic magmas than from more silicic magmas (Sparks and Pinkerton, 1978), basaltic magmas may de-gas over time, producing a gas-rich layer trapped at the top of the magma body (Vergniolle and Mangan, 2000; Houghton and Gonnermann, 2008). Bonaccorso et al. (2011) indicate that the magma volume required to provide the SO_2 observed during a Mt. Etna fire-fountain event would have needed to be 10 times the amount of all the erupted magma (effusive plus explosive). They note that other styles of volcanic activity with longer repose times (months to years) have ratios more like 4 to 1 of degassed to erupted magma. This implies that gas erupted with the explosive phase was derived from a magma volume that is considerably larger than the combined volume of erupted lava and ash. Thus, models of buoyant plume rise from basaltic eruptions may need to allow for a volatile content that is greater than the bulk magmatic value.

The second consideration is that much of the erupted lava from a fire-fountain returns to the ground and does not participate in the buoyant plume. Kaminski et al. (2011b) propose a mechanism for partitioning erupted magma between that which is lofted by the buoyant plume and that which is not (effusive flows and fire-fountain material that immediately falls back to the surface) by defining a parameter, f, the "fraction of magma finely fragmented and injected into the plume". The f parameter has a single value that determines the mass flux boundary condition for solid material into the plume at the source. For magma with a bulk volatile content of n_o=3wt%, and all the associated magma explosively released concurrently with the gas forming a buoyant ash plume, f=1 and the mass fraction of solids in the plume is (1 - n_o)=97wt%. If, however, only a fraction of the erupted magma contributes to the buoyant plume, the new "effective" volatile content is (Kaminski et al., 2011b),

$$n_f = \frac{n_o}{n_o + f(1 - n_o)}$$

For example, if the mass of lava erupted as lava flows and fire-fountains is twice that which makes up the buoyant plume (lava+fire-fountain=66.66%, plume=33.33% of total erupted mass), then f=33.33%. For n_o=3wt%, the effective volatile content becomes, n_f =8.5wt%, almost three times higher than the bulk volatile content of 3wt%. The observations of Chouet et al. (1974) are consistent with ash-poor buoyant plumes associated with intermittent basaltic fire-fountains and even higher values of n_f. Based on the observations reported by Bonaccorso et al. (2011), if the ratio of degassed to erupted magma is 4 to 1, and the mass fraction of solid material in the plume to the total erupted mass is 3%, then f=3% * 25%=0.75%, and n_f=80wt% (for n_o=3wt%). This is consistent with the Chouet et al. (1974) observations of 70–94wt% volatiles in Strombolian fire-fountains. Values for f can range from 0–100%, however, for a fire-fountain to sustain a buoyant ash plume a substantial fraction of solid material must be incorporated into the plume. A detailed study of the minimum f to sustain a buoyant ash plume is reserved for future study. It can be estimated, however, that the 1986 Izu-Oshima and the Laki fire-fountain eruptions, which both sustained large ash plumes, erupted ~30-35% of the magmatic mass as tephra (f=30-35%, see discussions below).

4. Model validation for 1986 Izu-Oshima eruption

The 1986 eruption of Izu-Oshima, Japan, produced some of the largest documented basaltic fire-fountains. Fire-fountains from the Izu-Oshima "B" fissure reached a maximum height in excess of 1600 m (above the vent) (Endo et al., 1988; Sumner, 1998). At their most energetic, the fire-fountains sustained a sub-Plinian buoyant ash plume that reached 12–16 km ASL

(Mannen, 2006; Mannen and Ito, 2007). The vent system for this eruption was made up of eight vents distributed along four en echelon fissure segments (Endo et al., 1988; Sumner, 1998). At the climax of the B eruption, lava fountains were observed from all vents between B3 and B8 (~1 km of fissure length).

Assuming ballistic physics of a vertical projectile, the maximum height of fire-fountain material can be found from $y=y_o+(v^2-v_o^2)/(2g)$, where y_o and y are the initial and final altitudes, v_o and v are the initial and final velocities, and g=-9.8 m/s^2 is gravitational acceleration (e.g., Resnick and Halliday, 1977). For a fire-fountain with $(y - y_o)$=1600 m and velocity at the apex of v=0, this implies an ejection velocity (at the vent, v_o) of ~175 m/s. At some point within the fire-fountain, the fine-grained ash and gas components dynamically separate from the ballistic material (which falls back to the surface). Lacking data on where this dynamic separation occurs, we assume that the gas and fine-grained ash that form the buoyant plume separate from the fire-fountain about 1000 m above the vent (~2/3 of the maximum fountain height), and begin modeling the buoyant convective plume at this point. Using the same expression for ballistic physics, the velocity is ~100 m/s at y=1000 m. The Izu-Oshima B-vents extend from ~500–600 m elevation (Smithsonian Institution, 1986). Thus, the buoyant plume model is initiated at an altitude of 1500 m (1000 m above the vent). A velocity of 100 m/s is assumed at the starting point of buoyant plume rise. Fire-fountains are generally still "red hot" at the apex. In the absence of data on lava temperatures during the Izu-Oshima eruption, the temperature of the buoyant plume material at the point where it separates from the fire-fountain is assumed to be ~1075 $^{\circ}$C. The model is not sensitive to choice of initial velocity and reducing the starting temperature by as much as 150 $^{\circ}$C has minimal effect on the overall plume rise.

Entrainment dynamics in the lower part of the buoyant plume are more efficient for a linear vent than a central vent (Sparks et al., 1997; Glaze et al., 2011), and can have an effect on overall plume rise. Figure 3 shows the maximum height achieved by buoyant plumes from three combinations of vent size and geometry. The x-axis shows a range of effective gas contents from very ash-poor, n_f = 80wt% (f=0.5%, n_o=2wt%) to ash-rich examples, n_f = 10wt% (f=18%, n_o=2wt%).

Fire-fountains during the early days of the Pu`u `Ō`ō eruption ranged from 200–400 m high. Images in Wolfe (1988) indicate these fire-fountains had widths of ~50–100 m. A circular source with diameter of 50 m is shown in Figure 3. It can be seen that for this 'typical' fire-fountain geometry, ash-poor plumes are only capable of rising to about 6 km, consistent with buoyant plumes observed above Pu`u `Ō`ō fire-fountains. Increasing the relative amount of ash released with the gas drives the plume higher, but the plume does not reach the minimum 12 km ASL observed at Izu-Oshima, even if all erupted magma is incorporated into the rising plume.

Doubling the source diameter to 100 m (not shown) can produce a sustained plume that almost reaches 12 km ASL, but only for n_f <<10wt%, where essentially all of the erupted material is incorporated into the buoyant plume. Assuming f≈100% is not consistent with observations of the Izu-Oshima fire-fountains where Sumner (1998) reports clastogenic lava flows, indicating that a substantial proportion of the fire-fountain material was still hot and fluid enough to remobilize after returning to the ground. Thus, circular-vent boundary conditions are not able to produce a sustained buoyant plume of the scale observed during the eruption.

Figure 3 also shows estimated plume heights for a linear source that is 1 km long (as observed at Izu Oshima) and 4 m wide or 16 m wide. For the 16 m wide linear source, the effective volatile contents that result in a sustained plume between 12-16 km range from relatively ash-poor (n_f=50wt%) to ash rich (n_f=10wt%). The Izu-Oshima eruption phases from the B-vents produced 3.0×10^{10} kg of lava (in the form of lava flows and scoria cones) and 1.4×10^{10} kg tephra (transported by the buoyant plume) (Endo et al. (1988). This results in f=32% and n_f=6wt% (for n_o=2wt%). The two linear source widths shown in Figure 3 result in predicted plume heights between 13.1 – 17.4 km ASL for n_f=6wt%, consistent with the range of observed buoyant plume heights.

5. Results and implications for the Roza eruption

The 14.7 Ma Roza flow (CRBG) comprises 1300 km^3 of lava erupted from a 180-km-long vent system, and has the best described fissure flood-basalt vent system in the world. Thordarson and Self (1998) suggest that the Roza eruption involved multiple phases over a period of ten to several tens of years. Evidence from pyroclastic deposits indicates multiple vents along the 180 km fissure system (Swanson et al., 1975), with a minimum of 11 identified vents in the northern-most 32 km (Brown et al., 2014).

Brown et al. (2014) show that eruptive activity at the northerly Roza vents began with a more explosive phase, followed by effusion of large-volume lava flows with less or little attendant explosive activity. Observations of historic basaltic fissure eruptions indicate that active linear vents are generally limited to lengths of no more than a few kilometers (e.g., Walker et al., 1984; Thordarson and Self, 1993), and generally contract down to centralized vents within a short period of time (Bruce and Huppert, 1989). Fissure lengths of 1 - 5 km are considered here. Although the active-fissure length does not affect the predicted plume heights (Glaze et al., 2011), it will have a substantial influence on the mass of material delivered to the plume top.

To estimate the likely plume rise height for the Roza eruption we assume parameters analogous to Izu-Oshima (fissure width=4-16 m, eruption temperature=1075 $^{\circ}$C, starting velocity=100 m/s). Based on glass inclusion data (Thordarson and Self, 1996), a bulk Roza volatile content of 2wt% is assumed. As for Izu-Oshima, we assume the buoyant column separates from the fire-fountain ~1000 m above the vent. The current Roza vent elevation is ~ 550 m ASL. However, incision of the Snake River Canyon indicates that the region has likely been uplifted by several hundred meters since 15 Ma ago. Thus, an initial vent elevation of 200 m is assumed here, resulting in a plume source altitude of 1200 m. The 300 m difference in initial altitude (compared to Izu Oshima) results in a small difference of <0.1 km in overall rise height.

Because of the similarity to the Izu Oshima example, the range of maximum plume heights for the Roza eruption is 13.1–17.4 km ASL for fissure widths of 4-16 m (as shown in Figure 3). At Roza's ~45°N paleo-latitude the tropopause is between 10–13 km ASL, depending on season. Thus, a plume from a 16 m wide linear source ($f \sim 32\%$, n_f=6wt%), can easily drive a plume into the stratosphere (Figure 3), even for more ash-poor cases. In order to assess the possible atmospheric impact of the Roza eruption, we can examine the volumes of magmatic volatiles that are injected into the stratosphere at the maximum plume height. Lacking specific total magmatic volatile information for Roza, we have assumed an example bulk magmatic volatile composition based on compositions of Kilauea lavas from Gerlach (1980), with 42% H_2O; 15% CO_2; 43% SO_2.

The total mass flux of material (gas plus ash) into the buoyant plume from a 16 m wide linear source is 5,567 kg/s per meter of fissure length. For n_f=6wt%, and a volatile content analogous to Kilauea, the mass flux of SO_2 is 144 kg/s per meter of fissure length. Thus, this plume could release 12.4 Mt of SO_2 per day for each kilometer of active fissure. If an eruptive episode began with an explosive phase along a 5 km segment of fissure, that phase could inject 62 Mt per day of SO_2 into the stratosphere, the equivalent of about three 1991 Pinatubo eruptions each day.

Thordarson and Self (1996) report a total of >12,400 Mt of SO_2 potentially released by the Roza eruption, with >9000 Mt of SO_2 emitted during explosive phases. At a rate of 62 Mt of SO_2 per day (per 5 km segment), it would take ~ 145 days of explosive activity (distributed throughout the decades-long eruption) to release 9000 Mt of SO_2. Assuming 5 km-long active fissure segments, the 180+ km of known fissure length could thus have supported ~ 36 explosive events, each with duration of 3 - 4 days of intense fire-fountaining.

6. Discussion

The model applications above assume that the atmospheric composition and temperature/pressure profiles 15 Ma ago (Miocene) were similar to the modern atmosphere. However, much older CFB provinces, e.g., Deccan Traps, India (~66 Ma), likely erupted into a substantially different atmosphere. A sensitivity study is provided here to assess the influence of atmospheric temperature structure and tropopause height, based on increased CO_2 concentrations, on predicted plume rise heights. We also explore an alternative to convective rise above fire-fountains where hot volcanic gases released from massive sheet flows covering large areas combine with warmed air above the flows forming a convective plume (Kaminski et al., 2011a).

6.1 Implications for plume rise in ancient atmospheres

To test the sensitivity of plume rise heights to atmospheric conditions representative of the Late Cretaceous and Miocene we have used climate simulations based on the HadAM3 Atmosphere-only climate model (Pope et al., 2000). This model has been extensively used and evaluated in Cretaceous (including Craggs et al., 2012; Hunter et al., 2013) and Miocene studies (Lunt et al., 2008; Pound et al., 2011). The model has 2.5°x3.75° horizontal resolution and 19 unevenly spaced layers in the vertical (up to 39 km). For the Maastrichtian we use the 71.3-65.0 Ma time-slice, 1120 ppmv CO_2 control experiment of Hunter et al. (2013). The Miocene experiment represents a 11.6-7.3 Ma time-slice with 395 ppmv CO_2 (Pound, 2012). Both experiments assume a modern orbit and pre-industrial atmospheric composition (except for CO_2). Model integration length is sufficient to ensure surface climatology has attained equilibrium. Vertical profiles of pressure, temperature and specific humidity are averaged from the final 30 model years and linearly interpolated to 500 m grid cells. Based on the standard World Meteorological Organization (WMO) lapse-rate criterion, we calculate the tropopause height as the "lowest level at which the lapse-rate decreases to 2°C/km or less, provided that the average lapse-rate between this level and all higher levels within 2 km does not exceed 2°C/km" (WMO, 1957).

4C has been run using the Roza boundary conditions (Section 5), but replacing the atmosphere with Miocene definitions for temperature (Figure 4). The Miocene pressure profile does not differ substantially from the modern atmosphere. The resulting range of maximum plume heights is 12.2 – 16.5 km (fissure widths of 4 and 16 m, respectively). Both ends of this range are shifted downward by 0.9 km relative to results using the modern atmosphere. Although the effect is relatively small, this systematic behavior is attributed to the warmer troposphere in the mid-Miocene. In the HadAM3 model, the tropopause height of about 13 km at 45°N for the

Miocene is ~2 km higher than the pre-industrial/modern atmosphere. Thus, the ability of Roza-like plumes to reach the stratosphere is somewhat sensitive to the choice of temperature profile.

This study can also be extended to explore possible buoyant plumes associated with fire-fountains during the emplacement of the Deccan Traps that were coincident with the K-Pg mass extinction (e.g., Wignall, 2001; Keller et al., 2012). The Maastrichtian atmospheric temperatures differ significantly from the modern atmosphere, in part due to the increased levels of CO_2 (4 times pre-industrial concentration assumed here). Using the Maastrichtian atmosphere temperature profile, and keeping all other boundary conditions the same, the predicted range of maximum plume heights is 11.9 – 18.3 km (fissure widths of 4 and16 m, respectively). The low end of this range is lower and the high end is higher than for the modern atmosphere. The broader range of maximum plume heights is primarily owing to differences in the shape of the temperature profile compared with the US Standard temperature profile (Figure 4). At low altitudes, the Maastrichtian atmosphere is warmer than the US Standard, meaning the density difference (driving buoyancy) between the plume and ambient air is smaller and the plume will stop rising sooner (a lower maximum rise height for 4 m fissure width). Above ~13 km altitude, the US Standard atmosphere is significantly warmer than the Maastrichtian, which reduces the density difference for the US Standard atmosphere. Plumes in the Maastrichtian atmosphere will continue to rise a little higher in the cooler troposphere. Despite a higher buoyant rise in the Maastrichtian atmosphere, at about 21°S (the paleo-latitude of the Deccan Traps) the tropopause is shifted upward significantly to an altitude of about 20.5 km compared to about 16 km in the modern atmosphere. Thus, based on the climate model predictions of tropopause height, it seems unlikely that plumes originating from volcanic vents in the tropics of the size examined here would have reached the stratosphere during the Maastrichtian. These results imply that the chemical processing and the lifetime of volcanic gases and aerosol particles may differ for different CFB provinces depending on atmospheric temperatures at the time of emplacement.

The influence of atmospheric composition on plume buoyancy was also considered. *4C* assumes an atmospheric mol mass of 28.966 g/mol. Increasing atmospheric CO_2 concentrations by a factor of four above pre-industrial levels, and keeping all other constituents the same, results in a mol mass of 28.990 g/mol. This slight increase results in a bulk atmosphere gas constant for the Maastrichtian of 286.8 J/kg/K, indistinguishable from 287.0 J/kg/K used in *4C*.

6.2 Rise of plumes above lava lobes

To model the rise of plumes above CFB flow fields, Kaminski et al. (2011a) considered penetrative convection driven by cooling of large basaltic lava flows. While penetrative convection is likely present and may somewhat increase predicted maximum plume heights, we question whether there would be a sufficient area of hot lava to support a vigorous convective plume above a flood basalt sheet lobe, or group of active lobes.

One issue not accounted for by Kaminski et al. (2011a) is that most (~ 75 % by mass) of the magmatic gas released to the atmosphere comes from the vent, with the remainder released from the lava flows (see summary in Thordarson et al. (2003)). Thus, by far the biggest release of gas into the atmosphere is the plume above the vent. Moreover, even near the vents, the CFB lava flow-fields are predicted to grow very gradually over years to decades (Self et al., 2005) and seldom will there be areas with near-magmatic-temperatures bigger than a few 100 m^2 due to the rapid development of a strongly insulating crust (Hon et al., 1994).

Another issue is the geometry with the vent in the middle of the CFB lava flow field (Kaminski et al., 2011a). This does not occur often in CFB provinces, as the lava flows away from the vent area, exemplified by the smaller Laki 1783-1784 analog (Thordarson and Self, 1993) and the distribution of the Roza flow (Brown et al., 2014). Further, the Van Dop et al. (2005) experimental tank upon which the Kaminski et al. (2011a) hypothesis is partly based has finite width and length. In contrast, the "well mixed" atmospheric boundary layer is infinitely wide compared to the areal extent of the lava flow. Kaminski et al. (2011a) do not discuss the importance of heat source area required to make this approach valid. The heat source areas covered by flood basalt lava flows are small relative to those considered by Van Dop et al. (2005), such as oceans.

Forest-fire pyro-cumulus clouds/plumes (pyro-Cbs) also provide an interesting comparison with flood basalt eruption plumes, as they can reach the stratosphere without any "gas-thrust" phase (Fromm et al., 2005; 2010). While forest-fire burn temperatures (~800-1000 $^{\circ}$C) are a little lower than magmatic basalt temperatures, they usually provide wide-area plume bases. Many of the fires that generate pyro-Cbs cover 10s – 100s of thousands of hectares, one to two orders of magnitude larger than the 3 – 20 km^2 of individual sheet lobes, or groups of co-emplaced lobes, that make up the Roza member. With development of insulating upper crusts (Self et al., 1998), this small area would not have been at "magmatic" temperatures. Thus it is unlikely that the surface of flood basalt sheet lobes are hot enough to sustain vigorous penetrative convection to great heights as suggested by Kaminski et al. (2011a).

7. Conclusions

Based on historical observations of large basaltic eruptions, buoyant plumes of gas and ash could have been sustained above fire-fountains along the 14.7 Ma Roza fissure vent system. Two modeling approaches are evaluated for predicting plume rise heights. Maximum plume heights estimated by the *4C* model (Glaze et al., 1997; 2011) are comparable to those predicted by the publicly available *Plumeria* model. However, *Plumeria* consistently overestimates maximum plume heights by 4-7% owing to two inconsistencies in the Woods (1988; 1993b) model upon which it is based: (1) discontinuous numerical solutions across the jet-buoyancy transition, and (2) the thermal energy conservation term is not consistent with momentum conversation.

Buoyant gas and ash plumes above an active fire-fountain are modeled using *4C* by assuming, based on observations, that ~2/3 of the erupted magma returns to the surface. To validate the model application to a buoyant plume fed by a fire-fountain, an example from the 1986 Izu-Oshima fissure eruption is examined. Linear source widths of 4 m and 16 m produce maximum plume heights of 13.1 and 17.4 km ASL, respectively, that bracket the range of observed heights (12 – 16 km ASL) for an effective volatile content of n_f=6wt%, equivalent to a bulk magma volatile content of n_o=2wt% and 32% of the erupted magma participating in the buoyant plume as fragmented ash. A basaltic fissure eruption of the scale of Izu-Oshima can easily drive a buoyant plume into the stratosphere, consistent with earlier work of Stothers et al. (1986).

If the Roza eruption sustained fire-fountains of similar height to Izu-Oshima (1.6 km above the vent), buoyant gas and ash plume heights of 13.1–17.4 km reaching the stratosphere at 45°N could also be sustained. The primary scale differences between Izu-Oshima and Roza are the possible lengths of active fissure and the duration of explosive activity. If an eruptive episode at

Roza began with an explosive phase along a 5 km fissure segment, that phase could inject 62 Mt per day of SO_2 into the stratosphere; equivalent to SO_2 emissions from three 1991 Mount Pinatubo eruptions each day. Assuming 5 km active fissure segments, the 180 km of known fissure length could have supported ~36 explosive events each with duration of 3-4 days. Small changes in the atmospheric temperature profile for the Miocene (relative to the modern atmosphere) result in slightly lower predicted plume heights. Plumes erupted in the Maastrichtian could have reached much higher into the atmosphere, but may not have extended into the stratosphere owing to the higher tropopause in the late Cretaceous.

The plume heights and mass flux estimated here can be used as input for 3-D climate models to quantify the effect on surface temperatures from eruptions such as Roza that consistently injected SO_2 into the stratosphere on a periodic basis over many years. The Roza eruption represents only one of ~ 120-150 members of the CRBG. Taking into account other eruptions in the CRBG, it is possible that large-scale flood basalt eruptions could have had a substantial influence on the chemistry of the atmosphere on timescales of decades to hundreds of years, and consequently, on climate.

Acknowledgements

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Figure Captions

Figure 1: Lava fountain from Pu`u `O`o on September 19, 1984. Fire fountain extends to 450 m high and sustains an ash and gas plume. Note how ash plume separates from fountain before apex. See text for discussion. Photograph taken by C. Heliker.

Figure 2: Maximum predicted plume heights as a function of fissure width using the Glaze et al. (2011) model for a linear source geometry. In all cases shown, water vapor is assumed to be the only volatile, and the initial velocity at the vent is 50 m/s. Steam plume (long dash) is water vapor only at a temperature of 100 °C. Super-heated steam (short dash) is water vapor only at near magmatic temperatures of 925 °C. 50% ash plume (solid line) assumes that the plume is made up of 50wt% water vapor and 50wt% fragmented ash. This is relatively ash-poor compared to typical bulk magmatic volatile amounts of 2 - 5wt%.

Figure 3: Maximum predicted plume heights as a function of effective gas content, n_f, in the buoyant ash and gas plume. Horizontal dashed lines indicate the range of observed buoyant ash plume heights (12 – 16 km ASL) during the explosive phase of the 1986 Izu-Oshima basaltic fissure eruption. All cases assume an initial temperature of 1075 °C and initial velocity of 100 m/s at the buoyant plume source, 1000 m above the fissure vent (point of gas separation from the fire-fountain, approximate 2/3 of the maximum fire-fountain height). Curves indicate buoyant plumes from a circular source with diameter of 50 m (solid circles), linear source with width of 4 m (open diamonds), and linear source with width of 16 m (solid line). For n_f=6wt%, plumes from linear sources 4 – 16 m wide rise to

heights of 13.1 – 17.4 km ASL. Assuming that the 14.7 Ma Roza eruption sustained fire-fountains of similar height to Izu-Oshima, analogous buoyant plumes would have reached the stratosphere at 45°N.

Figure 4: Comparison of temperature profiles used as input for the *4C* plume rise model for the modern atmosphere (US Standard), Miocene at 45°N (with atmospheric CO_2 concentrations set to 395 ppmv), and Maastrichtian at 21°S (assuming atmospheric CO_2 concentrations of 1120 ppmv, which is about four times that of the pre-industrial atmosphere). Arrows indicate the height of the tropopause based on the World Meteorological Organization (WMO) lapse-rate criterion (WMO, 1957) for the Maastrichtian (20.5 km), Miocene (13 km), and US Standard (11 km) atmospheres.

Figure 1.

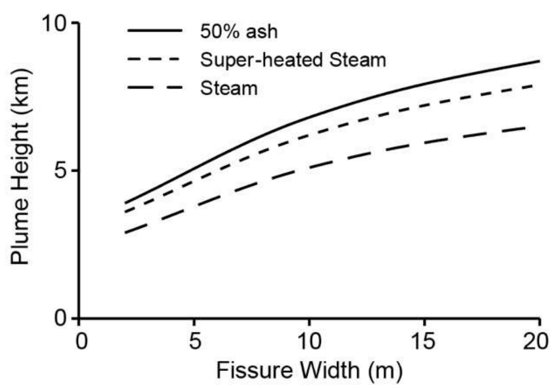

Figure 2.

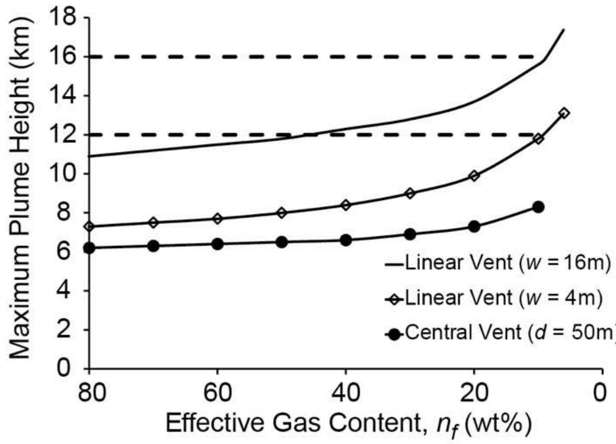

Figure 3.

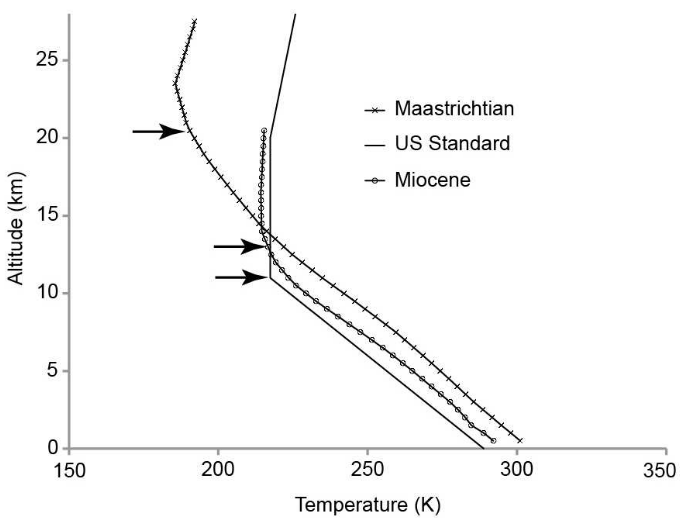

Figure 4.

Appendix A.
Supplementary Material for: Assessing eruption column height in ancient flood basalt eruptions

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Appendix A

Here we provide direct comparisons of the *4C* and *Plumeria* buyouant plume rise models. *4C* is based on the approach contained in the series of Glaze et al. papers (Glaze and Baloga, 1996; Glaze et al., 1997; Glaze, 1999; Glaze and Baloga, 2002; Glaze et al., 2011). A complete definition of *4C* for circular vents is given in Glaze et al. (1997). *Plumeria* (Mastin, 2007)is a user-friendly program that is freely available through the USGS Cascade Volcano Observatory (http://vulcan.wr.usgs.gov/Projects/Mastin/Publications/G3*Plumeria*/framework.html, last accessed 15 July 2013). The plume rise physics contained in *Plumeria* are based on Woods (1988; 1993).

Both *Plumeria* and *4C* begin with mass, momentum, and thermal energy conservation and both approaches model the buoyant rise of solid ash, water vapor (magmatic and entrained), ambient air, and liquid water. Both models also include the release of latent heat during water vapor condensation. The fundamental assumption of both models is that the effects of convection and turbulence on mass conservation can be estimated through an entrainment parameter (Morton et al., 1956). Although recent work by Carazzo et al. (2008) have indicated that the entrainment parameter may be variable, a constant value has been shown by numerous studies to be adequate to first order and is assumed here. Because both modeling approaches are based on Morton et al. (1956), the basic physics are similar, and maximum plume heights predicted by both approaches are also similar.

Table A1 details predicted plume heights from *Plumeria* and *4C*. The criterion used by *Plumeria* to stop plume rise is the point at which the upward velocity drops below 1 m/s. Because *4C* typically uses a value of 10 m/s to estimate the maximum plume height, the height

Table A1. Predicted maximum plume heights from *Plumeria* and *4C*; radius=vent radius, u=bulk plume velocity, H=maximum plume height (see text for details).

radius (diameter) (m)	*Plumeria* H (km) (u < 1 m/s)	*Plumeria* H (km) (u < 10 m/s)	*4C* H (km) (u < 10 m/s)	Delta (km)	% diff
20 (40)	11.4	11.3	10.5	0.8	7.6%
50 (100)	15.4	15.3	14.6	0.7	4.8%
100 (200)	19.6	19.6	18.7	0.9	4.8%
150 (300)	23.1	23	22	1	4.5%
200 (400)	26	26	24.9	1.1	4.4%
250 (500)	28.6	28.5	27.5	1	3.6%

predicted by *Plumeria* when velocity drops below 10 m/s is given in column three. The fourth column shows the predicted plume heights with a cutoff value of 10 m/s using *4C*.

In all cases, the following input parameters are assumed: Ambient air temperature at the vent =15.86 C, Tropospheric lapse rate =-6.5 K/km, vent elevation=0 km, relative humidity =0%, tropopause elevation =11 km, eruption velocity=300 m/s, eruption temperature =1000 K, magmatic volatile content (assumed to be water) =3wt%. For consistency with values stated in Mastin (2007), a specific heat for the solid particles of 1100 J/K/kg (Sparks, 1986) is used in *4C*. *4C* uses the US Standard Atmosphere to define ambient temperature and pressure profiles, which has a stratospheric lapse rate of 1.053 K/km between 11–20 km, and a lapse rate of 2.5 between 20-32 km. The standard *Plumeria* input only allows use of a single lapse rate in the stratosphere. As a result, plumes that rise higher than 20 km may have additional discrepancies introduced by atmospheric temperature.

Maximum plume heights predicted by *Plumeria* and *4C* differ by only a few percent (Table A1, column 6), with that difference decreasing as the plume size increases. There are two significant ways in which the models differ. First, as discussed in Glaze and Baloga (1996), the thermal energy conservation of Woods (1988) is inconsistent with the rest of the system of conservation equations. Glaze et al. (1997) define the basic thermal energy, without latent heat, as

$$\frac{d}{dz}\left(\rho_B u r^2 C_B \theta\right) = 2\alpha\rho_a u r C_a \theta_a - \rho_a u r^2 g$$

where z is the vertical variable, ρ_a and ρ_B are bulk densities of the atmosphere and plume, u is upward plume velocity, r is plume radius, C_a and C_B are specific heats of the bulk atmosphere and the plume, θ and θ_a are plume and atmosphere temperatures, and g is gravitational acceleration. For comparison, Woods (1988) defines this same thermal energy balance as

$$\frac{d}{dz}\left(\rho_B u r^2 C_B \theta\right) = 2\alpha\rho_a u r C_a \theta_a + \frac{u^2}{2} g\left(\rho_a - \rho_B\right) r^2 - \rho_a u r^2 g$$

The additional term on the right hand side of the Woods (1988) expression is not consistent with the momentum conversation equation (Glaze and Baloga, 1996). Figure A1 illustrates the impact of this difference on the bulk plume temperature as a function of height. Although, the temperature differences are small, correcting this inconsistency in *4C* results in maximum plume heights 4-7% lower than predicted by the Woods (1988) model (Glaze, 1999). Note that this one discrepancy describes all of the observed maximum plume height differences in Table A1.

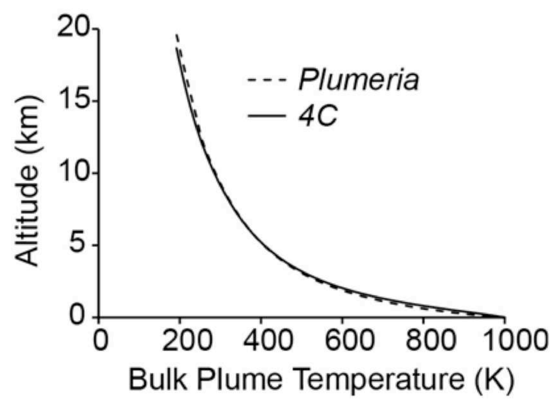

Figure A1: Comparison of bulk plume temperature as a function of altitude as estimated by the *Plumeria* and *4C* models. Both models assume a circular vent of radius 100 m, initial velocity of 300 m/s, initial temperature of 1000 K, and bulk volatile content of 3wt%. Inconsistency in the Woods (1988) thermal energy conservation equation results in small differences in the bulk plume temperature.

Another difference between *Plumeria* and *4C* is that Woods (1988) defines the transition between the jet and buoyancy regions as the point where bulk plume density drops below ambient density. However, as noted by Glaze (1999), this transition definition results in discontinuous solutions for all variables.

Alternatively, continuous solutions across the jet-buoyancy boundary (that converge at the same height above the vent) can easily be found for all variables. Figure A2 shows the transition height above the vent as a function of vent radius for both the continuous and discontinuous solutions. Requiring continuous solutions for all variables across this boundary (as dictated by physical systems) results in transition heights substantially lower than those determined using the discontinuous solution. Further, continuous solution transition heights do not vary much as a function of mass flux at the vent.

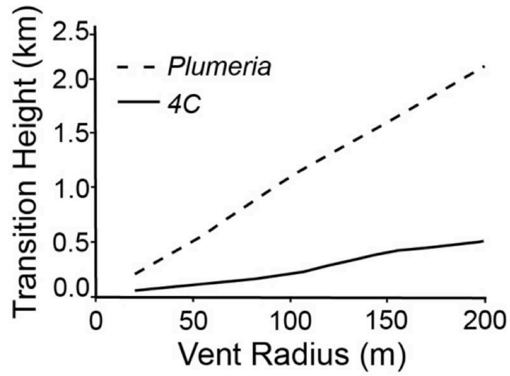

Figure A2: Figure after Glaze (1999). Comparison of heights where transition occurs between the jet- and buoyancy-driven regions for both the *Plumeria* and *4C* models. Both cases assume initial velocity of 300 m/s, initial bulk temperature of 1000 K, initial bulk volatile content of 3wt%, vent altitude of 0 km, and initial circular vent radius as shown on the x axis. The discontinuous transition employed by *Plumeria* results in an overestimate of the maximum plume height (Table A1).

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