Earth and Planetary Science Letters 377–378 (2013) 218–226

Contents lists available at SciVerse ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Impact of wind on the condition for column collapse of volcanic plumes

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ARTICLE INFO

Article history:
Received 29 January 2013
Received in revised form 5 June 2013
Accepted 28 June 2013
Available online 16 July 2013
Editor: T. Elliot

Keywords:
plume
collapse
buoyancy
wind
entrainment
Richardson number

ABSTRACT

The collapse of volcanic plumes has significant implications for eruption dynamics and associated hazards. We show how eruptive columns can collapse and generate pyroclastic density currents as a result of not only the source conditions, but also of the atmospheric environment. The ratio of the potential energy and the kinetic energy at the source quantified by the Richardson number, and the entrainment efficiency quantified by the radial entrainment coefficient have already been identified as key parameters in controlling the transition between a buoyant and collapsing plume. Here we quantify how this transition is affected by wind using scaling arguments in combination with a one-dimensional plume model. Air entrainment due to wind causes a volcanic plume to lower its density at a faster rate and therefore to favor buoyancy. We identify the conditions when wind entrainment becomes dominant over radial entrainment and quantify the effect of wind on column collapse. These findings are framed into a generalized regime diagram that also describes previous regime diagrams for the specific case of choked flows. Many observations confirm how bent-over plumes typically do not generate significant collapses. A quantitative comparison with the 1996 Ruapehu and the 2010 Eyjafjallajökull eruptions shows that the likelihood of collapse is reduced even at moderate wind speeds relative to the exit velocity at the vent.

1. Introduction

Whether a plume generated by an explosive volcanic eruption rises buoyantly into the atmosphere or collapses depends on the source conditions and the entrainment of air near the vent. The mixture of ash and gas ejected from the vent is negatively buoyant. However, provided the eruption columns initial momentum is high enough and the entrainment of air is efficient it can form a positively buoyant plume. On the other hand, if either initial momentum or air entrainment are insufficient to overcome the negative buoyancy, the eruption column will collapse either partially or completely and form pyroclastic density currents (PDCs). Quantifying the criterion that separates these distinct eruption styles has been the subject of intense study in physical volcanology as it has direct impact on hazard assessment (Sparks and Wilson, 1976; Wilson, 1976; Sparks et al., 1978; Wilson et al., 1980; Wilson and Walker, 1987; Carey and Sigurdsson, 1989; Valentine and Wohletz, 1989; Bursik and Woods, 1991; Woods and Caulfield, 1992; Neri and Dobran, 1994; Koyaguchi and Woods, 1996; Woods and Bower, 1995; Kaminski and Jaupart, 2001; Neri et al., 2003; Ogden et al., 2008; Carazzo et al., 2008; Kobs, 2009; Andrews and Gardner, 2009; Koyaguchi et al., 2010; Ogden, 2011; Shea et al., 2011; Suzuki and Koyaguchi, 2012; Carazzo and Jellinek, 2012; Saffaraval et al., 2012).

Previous studies have made great progress in identifying the critical parameters that influence column collapse. A major control on the condition for column collapse is the ratio of the potential energy and the kinetic energy evaluated at the source, which is quantified by

$$\Gamma_1 = \frac{-\mathrm{Ri}_0}{\alpha} = -g\left(\frac{\rho_{a0}-\rho_0}{\rho_{a0}}\right)\frac{r_0}{\alpha u_0^2} \qquad (1)$$

with α the radial entrainment coefficient, Ri_0 the Richardson number, r_0 the plume radius, u_0 the plume velocity, g the gravitational acceleration, ρ_0 the plume density, and ρ_{a0} the density of the ambient air. The 0-index indicates conditions at the source. The entrainment coefficient α was introduced by Morton et al. (1956), who assumed the radial entrainment of the plume is linearly proportional to the velocity of the plume, which is one of the key assumptions to reduce the complexity of the physics involved. Morton (1959) then showed how the non-dimensional number Γ_1 enables to distinguish between different types of forced plumes, i.e. plumes with a finite initial mass and momentum flow rate, of which a jet and a purely buoyant plume are the end members. Analogue experiments (Woods and Caulfield, 1992; Carazzo and

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0012-821X/$ – see front matter
http://dx.doi.org/10.1016/j.epsl.2013.06.041

Jellinek, 2012) and numerical simulations (Valentine and Wohletz, 1989; Suzuki and Koyaguchi, 2012) applied to the problem of volcanic plumes have used this quantity to distinguish between buoyant and collapsing columns, whereby we use *buoyant* for a negatively buoyant jet that develops a positively buoyant plume and *collapse* for the case where only a jet is formed. The critical number for Γ_1, say Γ_{1c}, above which collapse occurs has been found to be of order 10, written as $O(10)$ (Woods and Caulfield, 1992; Suzuki and Koyaguchi, 2012).

Volcanic plumes rising in a quiescent atmosphere are rare (Volentik et al., 2010; Fontijn et al., 2011) and, therefore, there is a need for quantifying the effect of wind on plume dynamics. Wind can significantly affect the trajectory of the plume through bending over (Bursik, 2001; Bonadonna et al., 2005) that sometimes also results in distal bifurcation (Ernst et al., 1994) and the rise height of the plume through a more efficient entrainment of air (Graf et al., 1999; Bursik, 2001; Tupper et al., 2009; Degruyter and Bonadonna, 2012; Woodhouse et al., 2013; Devenish, 2013). The effect of wind on plume rise can be quantified by

$$\Pi = 0.552\frac{\bar{N}H}{\bar{v}}\left(\frac{\alpha}{\beta}\right)^2, \tag{2}$$

with H the plume height, $\bar{N}$ the buoyancy frequency averaged over the plume height, $\bar{v}$ the wind speed averaged over the plume height, and β the wind entrainment coefficient (see Degruyter and Bonadonna, 2012 for derivation). Finally, a strong wind advection often results in unique sedimentation patterns, as bent-over plumes are typically characterized by (i) no up-wind sedimentation, (ii) narrow deposits, and (iii) rapid proximal-medial thinning associated with enhanced sedimentation of block- to lapilli-size particles from the inclined section of the column (e.g., Bonadonna et al., 2005).

Although wind has been recognized as having a major control on plume dynamics, a quantification of its role in the generation of PDCs has yet to be determined. Here we present a theoretical framework for this using a one-dimensional plume model and compare it to observations. Such integral models cannot capture as much of the physics of a volcanic plume as three-dimensional models (Valentine and Wohletz, 1989; Neri and Dobran, 1994; Graf et al., 1999; Neri et al., 2003; Ogden et al., 2008; Kobs, 2009; Tupper et al., 2009; Ogden, 2011; Suzuki and Koyaguchi, 2010, 2012), but we gain quantitative first-order insights by a fast assessment of a very large parameter space through repeated random sampling using for example the Monte Carlo technique (Degruyter and Bonadonna, 2012) or the polynomial chaos quadrature weighted estimate technique (Bursik et al., 2012). We then use our results to make a quantitative comparison with observations of the 1996 Ruapehu and the 2010 Eyjafjallajökull eruptions.

2. Methods

We simulate a steady turbulent plume exposed to a cross flow using the model described and applied to volcanic plumes in Degruyter and Bonadonna (2012). A schematic of the problem investigated is shown in Fig. 1. We thereby assume that

1. pyroclasts in the plume are well mixed and both mechanically and thermally coupled with the gas phase such that the plume can be described as a single phase (Woods, 1988; Hort and Gardner, 2000; Burgisser et al., 2005),
2. velocity and buoyancy variations through a cross section of the plume follow a top-hat profile that is self-similar along the trajectory of the plume (Morton et al., 1956),

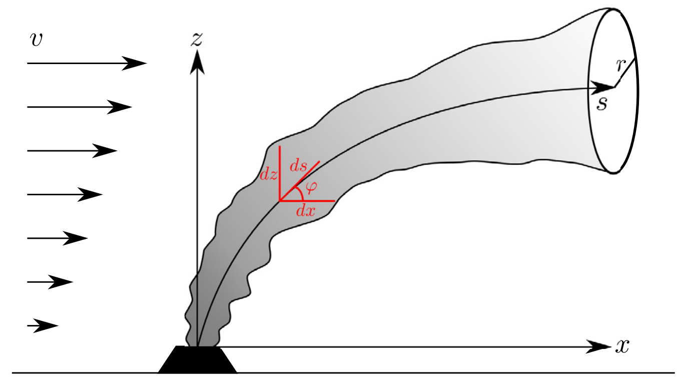

Fig. 1. Schematic representation of the coordinate system used by the one-dimensional plume model (after Hoult et al., 1969, and Degruyter and Bonadonna, 2012). See Table 1 for the definition of the symbols.

3. the rate of entrainment consists of two mechanisms: one is due to the difference between the plume (centerline) velocity and the wind velocity component parallel to the plume and the other is due the wind component perpendicular to the plume (Hoult et al., 1969; Hewett et al., 1971; Hoult and Weil, 1972; Bursik, 2001).

We do not incorporate the effects of jet overpressure on the plume (Saffaraval et al., 2012) nor injection of external water in the plume due to surface water or atmospheric humidity (Woods, 1993; Koyaguchi and Woods, 1996; Glaze and Baloga, 1996; Glaze et al., 1997). We also do not include the pressure-drag term that may result from the wind–plume interaction, which is used in some models (e.g., Ooms, 1972; Schatzmann, 1979), but that is considered not to be very important (Coelho and Hunt, 1989). These assumptions allow us to write the governing equations as

$$\frac{d}{ds}\left(\rho_d u r^2 \phi_d\right) = 2u_\varepsilon r \rho_a, \tag{3}$$

$$\frac{d}{ds}\left(\rho_v u r^2 \phi_v\right) = 0, \tag{4}$$

$$\frac{d}{ds}\left(\rho_s u r^2 \phi_s\right) = 0, \tag{5}$$

$$\frac{d}{ds}\left(\rho u^2 r^2\right) = g(\rho_a - \rho)r^2 \sin\varphi + v\cos\varphi \frac{d}{ds}\left(\rho u r^2\right), \tag{6}$$

$$\rho u^2 r^2 \frac{d\varphi}{ds} = g(\rho_a - \rho)r^2 \cos\varphi - v\sin\varphi \frac{d}{ds}\left(\rho u r^2\right), \tag{7}$$

$$\frac{d}{ds}\left(g\frac{\rho u r^2(c\theta - c_a\theta_a)}{\rho_{a0}c_a\theta_{a0}}\right) = -\frac{\rho}{\rho_{a0}}ur^2N^2\sin\varphi \tag{8}$$

where we refer to Table 1 for the definition of the symbols. Eqs. (3), (4) and (5) are the conservation of mass flow rate for dry air, exsolved volatiles, and pyroclasts, respectively. Eq. (6) is the conservation of momentum flow rate in the direction of the plume centerline s, Eq. (7) is the transverse momentum equation, and Eq. (8) gives the conservation of heat flow rate, where the buoyancy frequency N is defined by

$$N^2 = \frac{g^2}{c_a\theta_{a0}}\left(1 + \frac{c_a}{g}\frac{d\theta_a}{dz}\right). \tag{9}$$

This equation corresponds to the conservation of heat flow rate defined in Glaze and Baloga (1996) and Glaze et al. (1997). To close the set of equations we use the volume fraction relationship and the ideal gas law for the gasses in the plume (Glaze et al., 1997),

$$\phi_d + \phi_v + \phi_s = 1, \tag{10}$$

$$(\rho_d\phi_d R_d + \rho_v\phi_v R_v)\theta = P. \tag{11}$$