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The timing and intensity of column collapse during explosive volcanic eruptions

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ABSTRACT

Volcanic columns produced by explosive eruptions commonly reach, at some stage, a collapse regime with associated pyroclastic density currents propagating on the ground. The threshold conditions for the entrance into this regime are mainly controlled by the mass flux and exsolved gas content at the source. However, column collapse is often partial and the controls on the fraction of total mass flux that feeds the pyroclastic density currents, defined here as the intensity of collapse, are unknown. To better understand this regime, we use a new experimental apparatus reproducing at laboratory scale the convecting and collapsing behavior of hot particle-laden air jets. We validate the predictions of a 1D theoretical model for the entrance into the regime of partial collapse. Furthermore, we show that where a buoyant plume and a collapsing fountain coexist, the intensity of collapse can be predicted by a universal scaling relationship. We find that the intensity of collapse in the partial collapse regime is controlled by magma gas content and temperature, and always exceeds 40%, independent of peak mass flux and total erupted volume. The comparison between our theoretical predictions and a set of geological data on historic and pre-historic explosive eruptions shows that the model can be used to predict both the onset and intensity of column collapse, hence it can be used for rapid assessment of volcanic hazards notably ash dispersal during eruptive crises.

1. Introduction

The greatest volcanic hazard to lives and property on the ground are pyroclastic density currents (PDC), the largest of these being produced by collapse of "Plinian" columns during explosive eruptions. The collapse of a Plinian column is controlled by the evolution of buoyancy in the column. In a volcanic plume, the bulk density of the mixture of hot gas and pyroclasts is rapidly reduced by turbulent entrainment and thermal expansion of relatively cold atmospheric air (Sparks and Wilson, 1976; Woods, 1988). A stable Plinian column forms when the bulk density of the plume attains values below atmospheric density before its initial upward momentum has been exhausted, whereas collapse occurs when the initial momentum is not sufficient to carry the flow up to the point of buoyancy inversion. The transition between these two different dynamic behaviors is of crucial importance for the assessment of volcanic hazards to humans and climate change and it has been extensively studied. Laboratory experiments (Woods and Caulfield, 1992; Kaminski et al., 2005; Dellino et al., 2007; Carazzo and Jellinek, 2012), one dimensional steady eruption column models (Wilson et al., 1980; Kaminski and Jaupart, 2001; Carazzo et al., 2008; Koyaguchi et al., 2010), and 2-D and 3-D numerical models (Valentine and Wohletz, 1989; Neri and Dobran, 1994; Di Muro et al., 2004; Suzuki et al., 2005; Ogden et al., 2008; Suzuki and Koyaguchi, 2009, 2012) all agree on the conditions of collapse and show that it is mainly controlled by the mass discharge rate feeding the column and the gas content at the vent.

A number of geologic field studies and real-time observations have shown that column collapse is rarely total. Where the conditions of collapse are met, the jet commonly separates into a less dense part that forms a buoyant plume and a denser part that collapses to produce PDC. This phenomenon occurs in explosive eruptions ranging in size across the whole known volumetric range from small to large (Di Muro et al., 2004), and it has been directly observed and/or inferred from the structure of the deposits for numerous eruptions including the Bishop Tuff (Wilson and Hildreth, 1997), Taupo (Wilson and Walker, 1985), 79 Vesuvius (Sigurdsson et al., 1985; Carey and Sigurdsson, 1987), 1150 Quilotoa (Di Muro et al., 2008), 1300 Mt Pelée (Villemant et al., 1996; Carazzo et al., 2012), 1815 Tambora (Self et al., 2004), 1912 Novarupta–Katmai (Fierstein and Hildreth, 1992), 1937 Rabaul (Mckee et al., 1985), 1963 Agung (Self and Rampino, 2012), 1977 Ukinrek

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(Kienle et al., 1980), 1979 Soufrière St Vincent (Carey et al., 1988), 1980 Mt St Helens (Rowley et al., 1981), 1984 Mayon (see Fig. 1 of Kaminski et al., 2005), 1991 Mt Pinatubo (Scott et al., 1996), 2009 Sarychev Peak eruptions (Rybin et al., 2012), and 2010 Eyjafjallajökull eruptions (Bursik et al., 2012).

The key mechanisms that control the mass partitioning between the buoyant rising column and the dense collapsing fountain, which we refer thereafter as the "intensity of collapse", are poorly known. The intensity of collapse controls the PDC run-out distance on the ground while the remaining mass flux feeds the atmospheric ash plume, which may impact aviation safety and global climate through volcanic ashes and aerosols. It is thus a critical parameter for the rapid assessment of the predominant risk, which may pass from one of ash injection into the atmosphere to one of ground-level hazard as eruptive source conditions vary.

The goal of this paper is to better understand the controls on the mass partitioning during partial column collapse. To this aim, we first present an exhaustive review of field data from historic and prehistoric eruptions that underwent partial column collapse. We then give the results of new laboratory experiments reproducing hot particle-laden jets, and a 1D model to predict the conditions of entrance into the regime of partial collapse. Then, we use a simple theory to predict the intensity of collapse in our experiments, which is found to be in good agreement with field data. Our results emphasize the role of magma temperature and gas content on the intensity of collapse during volcanic eruptions.

2. Geological data on past eruptions

A first key step in understanding partial column collapse is to quantify it. In this section, we estimate the intensity of collapse, F, from the ratio of the mass of fall and flow deposits produced during the phase of partial column collapse for well-documented eruptions. Although partial column collapse has been often observed during historical eruptions, the interpretation of Plinian fall and PDC deposits from past eruptions is difficult because several mechanisms may affect the time relationship between the different units (e.g., a co-ignimbrite plume rising off a PDC). In addition, the proportions of fall vs flow material are difficult to estimate because of the variable erosion and wide range of densities between these components. Here, we selected eruptions for which either direct visual observations (1980 Mt St Helens and 1963–1964 Agung eruptions), or careful analyses based for example on the nature, composition and content of pumice, crystal and lithic fragments contained in the deposits (Bishop Tuff, Taupo, and 1912 Novarupta–Katmai eruptions) led the authors who conducted these studies to conclude that a partial column collapse did occur (Carey and Sigurdsson, 1985; Wilson and Walker, 1985; Fierstein and Hildreth, 1992; Wilson and Hildreth, 1997; Self and Rampino, 2012). Eruptions that formed a stable Plinian column without PDC (e.g., 1902 Santa Maria), a sustained fountain without fallout deposit (e.g., 1883 Krakatau), or which underwent a final column collapse (e.g., 1982 El Chichon) were discarded from the analysis.

2.1. The 1912 Novarupta–Katmai eruption

The Novarupta–Katmai eruption produced numerous fall layers interbedded with intraplinian PDC (Fierstein and Hildreth, 1992; Hildreth, 1991; Houghton et al., 2004). Fierstein and Hildreth (1992) separate this event into three main intervals of eruptive activity (Episodes I, II and III). Episode I began with a sustained Plinian column (layer A) with subordinate associated PDC ('all-rhyolite ignimbrite'), and was followed by dominantly valley-filling PDC (B1-ignimbrite, B2-mixed ignimbrite) with subordinate contemporaneous Plinian fallout (layers B1, B2, B3) (Fierstein and Hildreth, 1992; Hildreth, 1991). The intercalation of fall and flow units during this first episode is interpreted to result from an unstable column in the partial collapse regime (Fierstein and Hildreth, 1992; Houghton et al., 2004). Episode II was dominated by a sustained Plinian column (layers C and D) punctuated with the formation of proximal intraplinian flows and surges whose deposits are intercalated with the Plinian fallout C and D (Fierstein and Hildreth, 1992). Houghton et al. (2004) interpret this synchronicity as the result of a partial column collapse (PDC 5 in their Table 2). Episode III is also characterized by Plinian fall deposits (layers F and G) intercalated with thin intraplinian ignimbrite and several fall/flow layers (Fierstein and Hildreth, 1992). Houghton et al. (2004) also interpret this synchronity as the result of a partial column collapse (PDC 3, PDC 2 and PDC 1 in their Table 2).

Estimating F for each episode or unit is not possible because some important information such as the volume of each PDC deposit is not available. However, the careful fieldwork of Fierstein and Hildreth (1992) gives enough information to estimate F for the first stage of Episode I (layer A + all-rhyolite ignimbrite). During this initial phase, the volume of the "co-plinian" ash deposit (layer A) is found to be 9.7×10^{11} kg (Fierstein and Hildreth, 1992), whereas the volume of the contemporaneous PDC" (all-rhyolite ignimbrite) is estimated to be 2.5×10^{12} kg (Hildreth, 1991) (Table 1).

2.2. The 1980 Mt St Helens eruption

The May 18, 1980 eruption of Mt St Helens began with a violent lateral blast triggered by a large flank collapse. After this opening phase, a large anvil-shaped cloud that lifted off the surge area rose to 24 km by 0900 PDT (Carey and Sigurdsson, 1985, 1989). By 0925, a sustained eruption column developed (Carey and Sigurdsson, 1989) and rose to about 15 km. This sustained Plinian column oscillated by about 2 km during 3 h and deposited two fall units (B1 and B2). Numerous pyroclastic flows were generated by partial column collapse (Rowley et al., 1981; Carey and Sigurdsson, 1985) between 1215 and 1630 (B3 unit). Visual and radar observations indicate that a 14-km-high convective plume was still rising above the vent (Carey and Sigurdsson, 1989) and that a co-ignimbrite plume rose off the pyroclastic flows (Criswell, 1987) during B3. For the next 1.5 h, the column returned to a pure Plinian regime (B4 unit) (Carey and Sigurdsson, 1989).

Field observations show that the Plinian column was in the partial collapse regime during the B3 phase. The mass of pyroclastic flow deposits is estimated to be 2.9×10^{11} kg (Carey and Sigurdsson, 1989), whereas the mass of fall deposits, which includes material emplaced from the Plinian column and from the co-ignimbrite plume, is estimated to be 3.9×10^{11} kg (Carey and Sigurdsson, 1989). Field data and theoretical models suggest that when a co-ignimbrite plume forms, typically ∼35% of the total erupted material is elutriated from the pyroclastic flow and rise in the plume (Sigurdsson and Carey, 1989; Woods and Wohletz, 1991). Based on this work, we calculate a mass of 2.9×10^{11} kg for the Plinian fallout deposits and of 1×10^{11} kg for the co-ignimbrite deposits, the latter of which we add to the mass of pyroclastic flow deposits (Table 1).

2.3. The Bishop Tuff eruption

The Bishop Tuff was released at 0.76 Ma during the collapse of the Long Valley caldera (Wilson and Hildreth, 1997). The eruption produced nine fall layers (F1 to F9) interbedded with intraplinian ignimbrite (Wilson and Hildreth, 1997; Hildreth and Wilson, 2007). Wilson and Hildreth (1997) show unequivocally that the ignimbrite deposition did not simply follow after Plinian fall activity but is intraplinian in nature. Therefore, the whole eruption was in the