A new view of the dynamics, stability and longevity of volcanic clouds

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A B S T R A C T

Powerful explosive volcanic eruptions inject ash high into the atmosphere, which spreads to form umbrella clouds. Identifying key physical processes governing the dynamics, stability and longevity of umbrella clouds is central to assessing volcanic hazards as well as the nature of volcanic forcings on climate. Here we present a series of laboratory experiments producing turbulent particle-laden jets and subsequent axisymmetric intrusive gravity currents into a stratified environment. Our experiments reproduce many of the main dynamical regimes observed during the formation of an explosive volcanic column, and highlight new dynamics for the umbrella cloud. Theoretical predictions of column collapse from a simple model of a turbulent jet are in good agreement with experimental observations as well as previous studies. Depending on the flow intensity, the strength of the initial environmental density stratification and the particle concentration at the source, resulting umbrella clouds can, however, evolve through a series of new regimes as a result of the dynamics of particle sedimentation within these flows as well as from their bases. Using scaling theory we show that during cloud spreading, internal sedimentation drives the growth and intermittent overturn of thin, gravitationally unstable “particle boundary layers” (PBLs) as particle-rich plumes. This PBL-driven convection can have remarkable effects ranging from progressive dilution of clouds to their catastrophic overturn and collapse. In natural eruptions, whether the dynamics of PBLs play a major role in particle sedimentation depends on the grain size distribution inside the cloud and on eruption column height. In general, particles larger than ~60 μm–1 mm are expected to settle individually, whereas finer particles will accumulate PBLs resulting in the formation of armless mammatus clouds or dangerous gravity currents at much larger distances from the volcanic vent than ever considered before. Such dynamics is apparent in observations of numerous modern eruptions and is inferred from the deposits of historic and prehistoric eruptions for where there exist appropriate data. Consideration of the consequences of these phenomena for problems such as volcanic hazards to humans and climate change may, thus, be very important in the assessment of future eruptions.

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1. Introduction

Many explosive volcanic eruptions are characterized by turbulent jets that ultimately deliver dense mixtures of gas and ash into the atmosphere to an altitude of 20–50 km (Wilson, 1976). During a Plinian eruption the bulk density of these flows is rapidly reduced as a result of the mechanical entrainment and thermal expansion of cold atmosphere (Sparks and Wilson, 1976; Woods, 1988). A major effect of this process is that the jet can become less dense than the atmosphere and rise as a buoyant plume to a level of neutral buoyancy (LNB). At this altitude, the mixture spreads as a turbulent gravity current, forming an umbrella cloud the structure of which is further modified by stratospheric winds and turbulent diffusion (Carey and Sparks, 1986; Sparks et al., 1992).

Sedimentation of ash and pumice from these clouds presents a number of potentially severe hazards. Coarse ash and lapilli can cause roof collapses, or blockages of drinking water, waste disposal and power distribution systems, whereas fine and very fine ash can provoke health problems (Hornwell, 2007) or major air travel disruptions (Casadevall, 1994). The prediction of ash cloud behavior is therefore key to evaluating these classes of hazards as well as well-established climatic impacts linked to these eruptions (e.g., Robock, 2000).

Analyses of pyroclastic deposits and real-time observations suggest that most of the volcanic fragments discharged at the vent are still in suspension when the flow reaches the neutral buoyancy level (e.g., Bonadonna and Phillips, 2003). Sedimentation of the largest pumice and ash fragments is governed by their individual settling velocities, whereas particle aggregation is thought to enhance the fallout of fine ash (Rose and Durant, 2011). The residence time of the cloud mainly thus depends on the total mass and grain-size distribution of the particles injected into the atmosphere (Bursik et al., 1992).

Numerous tephra dispersal models have been developed in order to understand various aspects of the transport and deposition of volcanic ashes (Bonadonna and Phillips, 2003; Bonadonna et al., 1998; Bursik et al., 1992; Carey and Sparks, 1986; Ernst et al., 1996; Pfeiffer et al., 2005). Despite their different levels of complexity, all these models

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treat sedimentation by assuming that particle loss is mainly governed by the settling velocity of each particle, which is broadly supported by field studies and by laboratory experiments (Carey et al., 1988; Ernst et al., 1996; Sparks et al., 1991; Veitch and Woods, 2000).

Recent observations, however, suggest that processes governing sedimentation may be far more spatially complex and time-dependent than inferred from field and modeling studies. In particular, observations of the 1994 Reventador (Chakraborty et al., 2006) and 1991 Mount Pinatubo eruptions (Chakraborty et al., 2009) suggest that the dynamics of volcanic umbrella clouds can be strongly modified by large scale gravitational (Rayleigh–Taylor-type) instabilities driven by processes related to internal particle sedimentation, which will lead inevitably to time-dependent sedimentation. Similar buoyancy effects are also observed in a large number of historical eruptions (Fig. 1). These instabilities can take several forms such as relatively large rounded mammatus clouds (Fig. 1A, B and D), particle-rich fingers (Fig. 1C), or very thin ash veils (Hobbs et al., 1991). Several mechanisms have been proposed to explain these intriguing features (Bonadonna et al., 2002; Durant et al., 2009; Schultz et al., 2006) but their origin remains unclear in particular for mammatus clouds (e.g., Schultz et al., 2006). However, on the basis of Fig. 1 alone, one can expect that the construction of tephra fallout deposits should reflect the time- and space-dependent flux of material from the umbrella. However, interpreting the resulting local heterogeneities in the field in terms of dynamics for the cloud is not straightforward especially because numerical models cannot be used at the outcrop scale. To the best of our knowledge the only study to identify the role of episodic sedimentation from an umbrella cloud is the field-based work of Branney (1991) who argues that discontinuous “patches” of finely stratified pumice in the Whirrnesside tuff result from episodic expulsion of dense particle-rich material from umbrella cloud.

The goal of this paper is to investigate whether the enigmatic buoyancy effects summarized in Fig. 1 play a major role in the evolution of the umbrella cloud as a whole. For this, we present an extensive series of novel laboratory experiments simulating particle-laden jets with reversing buoyancy and capable of forming stable umbrella clouds. Dynamic scaling laws for dilute multiphase mixtures are used to show that our experiments are reliable analogs for natural explosive volcanic eruptions. In addition to reproduce buoyant plumes and collapsing fountains, our experiments show that resulting umbrella clouds may follow different unexpected dynamical regimes. The conditions leading to the formation of either a stable plume or a collapsing fountain in our experiments are understood with a simple theoretical model of turbulent jets. A scaling theory is then used to show that when a stable cloud forms, its long term behavior is strongly controlled by the internal sedimentation of fine particles into thin, gravitationally unstable “particle boundary layers” (PBLs), which intermittently overturn. Finally, we address the key issue of the extent to which the dynamics of these PBLs govern the stability and longevity of natural volcanic clouds.

Our paper is organized in the following way. Section 2 presents the experimental device and the scaling laws used to check that the main processes acting on the dynamics of volcanic plumes are reproduced in laboratory. Qualitative observations of flows spanning a large range of conditions are presented in Section 3. In Section 4 we develop a theoretical model and a scaling theory to understand the stability of our particle-laden jets and clouds. Implications of this work for natural eruptions and limitations linked to our approach are discussed in Section 5. Conclusions and future directions are presented in Section 6.

2. Laboratory experiments and scaling analysis

2.1. Experimental device

The experiments are isothermal and conducted in a 0.8 m high tank with a 1 m × 1 m cross-section (Fig. 2A). This tank is first filled with an aqueous NaCl solution with either a stepwise or linear density stratification. In most experiments the tank is filled with a 0.2 m-thick basal layer of salt water and an overlying 0.4 m-thick layer of fresh water that we introduce carefully to avoid mixing across the density interface. We conduct additional and dynamically equivalent experiments with a continuous linear density stratification constructed using the double bucket technique (see Section 2.3). At the start of each
experiment, we inject a well-stirred mixture of fresh water and well-sorted Custer feldspar particles (Fig. 2B) at a fixed rate using a long syringe. This setup enables us to vary the main source and environmental conditions that govern flow dynamics: (i) the particle mass, (ii) the volumetric flow rate, (iii) the nozzle size, and (iv) the stratification in the tank. To cover the full range of conditions appropriate for natural conditions we vary the particle concentration at the source and the input flow rate. Table 1 lists the experiments and the conditions imposed at the source. A typical injection lasts between 1 and 10 min and is characterized quantitatively with an array of high-resolution video cameras. Following the injection, a long time exposure camera records the complete evolution of the suspended particle-laden mixture.

2.2. Scaling laws of dilute multiphase mixtures

The complexity of the dynamics of explosive volcanic plumes raises the question of the reliability of our experiment to reproduce natural phenomena. Here, we present a scaling analysis to ensure that our laboratory experiments are a suitable analog for natural eruption plumes.

A number of parameters related to the distinct dynamics of the particles and the particle–fluid mixture in the jet and the umbrella are required to scale our experiments. All the variables used are specified in the Notation section. The subscripts $\rho$ and $\mu$ are used to distinguish properties at the vent (i.e., source of the jet) and at the LNB directly above the vent (i.e., source of the umbrella), where the mixture spreads laterally. We note that since the height of the LNB and related properties are controlled by environmental and source conditions for the jet (Morton et al., 1956), scales for the cloud are ultimately imposed by the source conditions for the jet.

The Reynolds number ($Re$) quantifies the importance of inertial and turbulent stirring in the flow of the particle–fluid mixture,

$$Re = \frac{VL}{\nu},$$

where $V$ and $L$ are the characteristic velocity and length scales for the flow and $\nu$ is the kinematic viscosity of the mixture. For the vertical jet, the characteristic velocity and length scales are the exit velocity $U_0 = Q_0/\pi R_0^2$ and the vent radius $R_0$, respectively, with $Q_0$ the input flow rate. For the umbrella cloud, the characteristic length and velocity scales are the overshoot height $h_m = z_m - z_0$ and the lateral spreading rate $W_m = \sqrt{h_m g_m^*}$, where $z_m$ and $z_0$ are the maximum and neutral buoyancy heights, respectively. Here $g_m^* = g(\rho - \rho_w)/\rho_w$ is the reduced gravity of the jet at the maximum height $z_m$.

### Table 1

Experimental conditions imposed at the source. Regime: FG: fingering; LY: layering; LC: late collapse; PC: partial collapse; TC: total collapse. *: linear stratification experiment (see Fig. 3D).

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<th>Exp.</th>
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<th>$Q_0$ (m$^3$ s$^{-1}$)</th>
<th>$M_0$ (m$^4$ s$^{-2}$)</th>
<th>$F_0$ (m$^4$ s$^{-2}$)</th>
<th>$F_u$ (m$^4$ s$^{-2}$)</th>
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<td>$4.3 \times 10^{-3}$</td>
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with \( g \) the acceleration of gravity, \( \rho_a \) the density of the ambient fluid and \( \rho \) the density of the particle-fluid mixture, which is given at any height by

\[
\rho = \epsilon_p \rho_p + \left(1 - \epsilon_p \right) \rho_f,
\]

(2)

where \( \epsilon_p \) is the particle volume fraction, and \( \rho_p \) and \( \rho_f \) are the particle and fluid densities, respectively.

The evolution of a jet depends strongly on its density, which is governed by turbulent entrainment across the stabilizing density interface that defines its sides. To characterize the balance between the stabilizing buoyancy and driving inertial forces governing the dynamics of entrainment and mixing, we use a Richardson number (\( Ri \)):

\[
Ri = -\frac{g \tau_f}{\nu g}.
\]

(3)

For the jet, \( L = R_b \), \( V = U_0 \) and \( g^* = g[(\rho - \rho_a)/\rho_a] \). Entrainment and mixing are enhanced where \( -Ri < 1 \), the jet will rise as a plume where \( Ri > 0 \) and can potentially collapse as a fountain where \( Ri < 0 \).

To characterize the balance between the buoyancy and additional viscos forces governing the convection driven by the dynamics of thin particle boundary layers at the cloud base we define a Grashof number (\( Gr \)):

\[
Gr = \frac{g^* L^3}{\nu^2} = \frac{Ri}{Re^2}.
\]

(4)

where \( L \) is the bubble size and \( g^* = g \) for the umbrella cloud.

In dilute mixtures, particle-particle interactions are negligible but the mechanical coupling between particles and the fluid phase is complex. The two-way stress coupling between the phases can be characterized by two dimensionless numbers. The Stokes number (\( St \)) is a ratio of the time scale for the inertial response of particle \( \tau_p \) to the time scale for flow of the continuous phase \( \tau_f \) and is defined as

\[
St = \frac{\tau_p}{\tau_f} = \frac{1}{18} \frac{\rho_p d_p^2 V}{\mu L}.
\]

(5)

This parameter measures the tendency of particles to stay coupled to turbulent motions. Here, \( f \) is a drag factor, \( d_p \) is the particle diameter (or effective particle diameter for non-spherical particles (cf., Pfeiffer et al., 2005)), and \( \mu = \nu \) is the dynamic viscosity of the fluid phase. \( f \) can be related to the particle Reynolds number \( Re_p \) (Burgisser et al., 2005),

\[
\frac{f}{1 + 0.15 Re_p^{0.687}} = 0.0175 \frac{1}{1 + 42, 500 Re_p^{-1.16}}.
\]

(6)

where

\[
Re_p = \frac{V_t d_p}{\mu},
\]

(7)

with \( V_t = \tau_p g \) the particle terminal fall velocity. Where \( St > 1 \) particles decouple from the fluid phase and enhance turbulence by momentum transfer during their motion and fallout, whereas for \( St < 1 \) particles remain coupled to eddies and reduce turbulence intensity (Elghobashi, 1994).

The tendency for dense particles to remain in suspension in a turbulent flow is expressed with an additional sedimentation number (\( \Sigma \)), which is a ratio of the inertial response time of a particle \( \tau_p \) to the time scale for its sedimentation \( \tau_s \).

\[
\Sigma = \frac{\tau_p}{\tau_s} = \frac{1}{18} \frac{\rho_p d_p^2 g}{\nu V} = \frac{V_t}{V}.
\]

(8)

Thus, particle settling is enhanced for \( \Sigma > 1 \) and inhibited for \( \Sigma < 1 \). In the framework of a turbulent jet, \( V = U_0 \) and \( \Sigma \) indicates whether particles settle (\( \Sigma > 1 \)), rise (\( \Sigma < 1 \)) or are neutrally buoyant (\( \Sigma = 1 \)). In the framework of an umbrella cloud, \( V = W_0 \) and \( \Sigma \) indicates the extent to which particles settle (\( \Sigma > 1 \)) or are carried laterally (\( \Sigma < 1 \)).

The Stokes and sedimentation numbers completely characterize the behavior of particles within the jet and umbrella. For \( St < 1 \) and \( \Sigma < 1 \) particles are coupled with the fluid phase and efficiently mixed by larger eddies. This regime favors the transport and dispersion of the solid fraction into the environment. For \( St > 1 \) and \( \Sigma < 1 \) particles are decoupled from the continuous phase and either settle or are expelled from the flow. Intermediate configurations where \( St < 1 \) and \( \Sigma < 1 \), or \( St > 1 \) and \( \Sigma > 1 \), or \( St > 1 \) and \( \Sigma < 1 \) correspond to a third class where a strong two-way coupling between fluid and particles exists. In this regime, transient segregations of particles lead to particle gathering in zones of least velocity gradient and a possible formation of mesoscale structures (Agrawal et al., 2001; Burgisser et al., 2005).

2.3. Comparison between experiments and volcanic clouds

The values of the dimensionless numbers discussed above are calculated in our experiments and for natural volcanic clouds (Fig. 3A to C and Table 2). We note that the Reynolds numbers for explosive eruptions vary in the range \( 10^2 \leq Re \leq 10^7 \) for the vertical jet and \( 10^3 \leq Re \leq 10^{10} \) for the umbrella cloud. Such extreme Re are unapproachable at the laboratory scale, although our flows are at high-Re, fully turbulent and conducted under Re conditions comparable to many published studies (e.g., Burgisser et al., 2005).

Fig. 3A shows that our experimental range of \( \Sigma \) is consistent with natural flows, which implies that the degree of dilution at the source is well reproduced. In addition, our laboratory experiments yield values of Richardson number partially within the natural range at the base of volcanic jets. This agreement suggests that the balance between the stabilizing buoyancy and driving inertial forces is also well reproduced in our experiments.

Fig. 3B shows that our experimental range of particle volume fraction at LNB is significantly different to that for the natural case. This difference is a consequence of the limitation imposed by the height of the tank, which controls the maximum possible dilution by turbulent entrainment of salt water. Whereas explosive volcanic jets entrain atmospheric air over a few tens kilometers reducing particle concentration by several orders of magnitude, our small scale jets entrain salt water over a few tens centimeters reducing \( \rho_a \) by only one order of magnitude. An additional remark is the significant difference between the values of the Grashof number yielded in our experiments to that for the natural case. We note, however, that in both cases \( Gr > 10^2 \).

Fig. 3C shows that our experimental values of the Stokes and sedimentation numbers fall within the wide range covered by the natural systems. \( St \) and \( \Sigma \) change by several orders of magnitude from the jet to the umbrella cloud (Fig. 3C). Above the nozzle, particles have \( St < 1 \) and \( \Sigma < 1 \), which suggests that particles have no effect on the development of turbulence and that no sedimentation should occur from the jet. On the other hand, particles have \( St < 1 \) and \( \Sigma > 1 \) in the umbrella cloud. This suggests that particles are well mixed into the umbrella cloud, although they tend to settle rather than being dispersed laterally.

Another key parameter controlling the flow dynamics is the strength of the environmental stratification. The quasi-linear density gradient observed in nature is reproduced in our experiment with a single density interface for the sake of simplicity. This technique has been applied successfully in a number of studies (Fan, 1967; Hart, 1961; Kotsovinos, 2000) and does not affect the plume dynamics as long as the scaling for the strength of the stratification is the same as the strength of a linear density stratification (Morton et al., 1996).
This condition is verified by using the Brunt–Väisälä frequency ($N$) defined as

$$N^2 = -\frac{g}{\rho_s} \frac{d\rho_s}{dz}$$  \hspace{1cm} (9)$$

where $\rho_s$ is a reference density. Fig. 3D shows a direct comparison of column height between an experiment with two layers and another with a linear density gradient. Our comparison confirms that (i) both types of experiment are quantitatively identical, and (ii) the neutral buoyancy height is not imposed at the interface between the two layers. In the terrestrial atmosphere, the Brunt–Väisälä frequency is in the order of 0.01 s$^{-1}$, whereas for all our experiments $N$ is in the order of 0.1 s$^{-1}$. This difference is imposed for technical reasons because $N$ partially controls the maximum column height. In turn, the stratification imposed allows to reproduce geometrical aspect ratios observed in natural eruptions (i.e. $z_b/z_m \approx 0.7$ and $z_m/R_0 \approx 10^3$).

Finally, we note that contrary to natural explosive volcanic eruptions, only isolated plumes are produced in our experiments. Although variations in initial temperature may affect the plume dynamics, its effect on the flow development is relatively weak compared to other dynamical quantities such as the initial gas content, exit velocity and vent radius (Woods, 1988). At the LNB, the plume temperature is imposed by surrounding environmental conditions as a result of intense mixing with the atmosphere during the column rise (Holasek et al., 1996). Therefore, the effect of source temperature on conditions across the LNB is negligible and our isothermal assumption does not affect the quantitative conclusions presented in this study.

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<th>Variable [Units]</th>
<th>Experiments</th>
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<tr>
<td>$h_m$ [m]</td>
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<td>$10^{-8}-10^{3}$</td>
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<td>$\phi$</td>
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<td>$7.5 \times 10^{-10} - 10^{-1}$</td>
</tr>
<tr>
<td>$\phi_{\text{surf}}$</td>
<td>$10^{-6} - 10^{-1}$</td>
<td>$10^{-6} - 10^{-4}$</td>
</tr>
<tr>
<td>$\mu$ [kg m$^{-1}$ s$^{-1}$]</td>
<td>$7 \times 10^{-4} - 7 \times 10^{-3}$</td>
<td>$10^{-6} - 10^{-5}$</td>
</tr>
<tr>
<td>$\rho_s$ [kg m$^{-3}$]</td>
<td>$10^{-3}$</td>
<td>$3 \times 10^{-5}$</td>
</tr>
<tr>
<td>$\rho_m$ [kg m$^{-3}$]</td>
<td>998–1008</td>
<td>$10^{-3}$</td>
</tr>
<tr>
<td>$\rho_s$ [kg m$^{-3}$]</td>
<td>998</td>
<td>0.1–1</td>
</tr>
<tr>
<td>$G_r$</td>
<td>$106-3 \times 10^7$</td>
<td>750–2500</td>
</tr>
<tr>
<td>$Re$</td>
<td>950–4100</td>
<td>$10^{15} - 10^{20}$</td>
</tr>
<tr>
<td>$Re_p$</td>
<td>20</td>
<td>$10^{-4} - 10^{5}$</td>
</tr>
</tbody>
</table>
3. Qualitative observations

Our experiments investigate the different phenomena that occur as we vary both the initial volumetric flow rate and the particle volume fraction. For low particle concentrations and relatively high flow rates, the mixture of solid particles and fresh water mixes rapidly with the ambient salt water layer (Fig. 4). Although the particle-laden jet is denser than the surrounding fluid at the source, its density decreases as a result of turbulent entrainment and dilution to values lower than the ambient density. Resultant buoyancy forces augment the momentum flux imparted at the source to drive the plume to a LNB, which it initially overshoots to a height that depends on its momentum and the strength of the stratification, before collapsing back to the LNB as a fountain to drive a gravity current laterally, forming an umbrella. This behavior is shown in Fig. 4A (“stable plume”) and is a popular analog for the formation of a Plinian column (Wilson, 1976).

For intermediate particle concentrations and moderate flow rates, there is less entrainment and mixing of the ambient fluid. The flow separates into a buoyant plume with a low particle concentration and a dense fountain with a high particle concentration. In general, whereas the buoyant plume impinges and spreads along the density interface in the tank as an intrusion, the collapsing part generates a radially symmetric turbulent gravity current at the tank base. This behavior is shown in Fig. 4A (“partial collapse”) and corresponds to explosive volcanic eruptions in the transitional regime (Di Muro et al., 2004).

For high particle concentrations and relatively low flow rates, there is little entrainment and mixing and the mixture injected remains too dense to undergo a buoyancy inversion. The flow collapses on itself when its initial momentum is exhausted and forms a turbulent fountain feeding a radial gravity current at the bottom of the tank. This behavior is shown in Fig. 4A (“total collapse”) and is an analog to a volcanic collapsing fountain with associated pyroclastic density currents (Sparks and Wilson, 1976).

These three dynamical regimes have already been observed in previous laboratory studies of turbulent particle-laden jets (Carey et al., 1988; Sparks et al., 1991; Veitch and Woods, 2000) and jets with reversing buoyancy (Kaminski et al., 2005; Woods and Caulfield, 1992). These behaviors have also been inferred from field observations of past eruptions (Sparks and Walker, 1977), and are also reproduced in numerical simulations of explosive volcanic eruptions (Di Muro et al., 2004; Suzuki et al., 2005; Valentine and Wohletz, 1989). A remarkable new finding, however, is the dynamic and highly time-dependent nature of the evolution of umbrella clouds in the stable plume regime (Fig. 4B).

For source conditions close to the partial collapse regime, the stable plume forms an umbrella cloud but gravitational instabilities develop in the near field of the cloud where particles are more concentrated. Ultimately, these instabilities lead to a collapse from the neutral buoyancy height to the tank base with an associated gravity current, from which phoenix clouds may emerge (Fig. 4B and Movie 1). To our knowledge, this “late collapse” regime has not been observed in previous laboratory studies involving turbulent particle-laden jets.

At lower particle concentrations, we observe two different regimes, depending on the injection rate. At relatively low injection rates, the particle-laden plume overshoots the density interface in the tank and falls back rapidly forming an umbrella cloud. The cloud propagates slowly along the neutral buoyancy height producing little mixing across the density interface. A few minutes after spreading begins, a dense particle-rich boundary layer (PBL) forms by steady accumulation of particles. The growth of this particle boundary layer at cloud base is accompanied by a vertical oscillation that grows in amplitude until the PBL becomes sufficiently thick to detach and descend.

---

Fig. 4. Photographs of the laboratory experiments showing the different types of behavior observed as the volume flux and particle content were changed. (A) The three regimes observed for the jet: “stable plume” (top—exp. 24), “partial collapse” (middle—exp. 17), and “total collapse” (bottom—exp. 22). Symbols at the top left corner correspond to the symbols used in Fig. 5. (B) Evolution of the umbrella cloud in the stable plume regime: (i) the “layering” regime is characterized by the formation of layers (right—exp. 8) and thin, long, and slow fingers at the cloud base (left—exp. 25); (ii) the “settling-driven fingering” regime is characterized by a dense basal particle boundary layer (left—exp. 6), which may become sufficiently thick to detach and descend as large dense particle-rich fingers (right—exp. 3); (iii) the “late collapse” regime is characterized by the growth of a gravitational instability in the near field of the cloud (left—exp. 20) leading to a final collapse with an associated gravity current. “Phoenix clouds” may form if the current mixes efficiently with salt water (right—exp. 19). Note the different scales.
as large dense particle-rich fingers (Fig. 4B and Movie 2). Where particle-laden downwellings reach the bottom of the tank entrained fresh water is expelled upward, driving large-scale convective flow, in turn. This ‘settling-driven fingering’ regime has similarities with particle-rich fingers observed in the experiments of Cardoso and Zarrebini (2001).

At larger flow injection rates, the plume significantly overshoots the density interface and falls back, resulting in extensive mixing across the density interface. A resulting thick umbrella spreads and no oscillation is observed. Gradually, however, the umbrella breaks up into discrete (light brown) particle-laden layers separated by relatively thin (dark) water-rich layers a few tens of minutes after its emplacement. A particle boundary layer forms at the base of the umbrella and the loss of particles is driven by slow settling of individual particles within long fingers descending from the particle boundary layer (Fig. 4B and Movie 3). In common with the fingering regime, large-scale convection starts when particle-rich fingers reach the bottom of the tank. However, complete particle sedimentation in the layering regime takes one to two orders of magnitude longer than in the fingering regime. This “layering” regime has not been observed in previous laboratory experiments simulating a volcanic ash-cloud.

4. Theory

The results presented above suggest our 300 μm-particles do not settle individually from the cloud, but instead behave collectively and form PBLs leading to either settling-driven fingering, layering or late collapse. To characterize the mechanisms controlling the formation of these PBLs and subsequent fingers, we compare our experimental results with theory. Identifying the conditions in the cloud prior to PBLs formation is challenging in our experimental configuration since the conditions imposed at the nozzle are strongly modified during the jet rise. To build understanding we characterize the jet dynamics and the mechanics of PBL and fingering formation, in turn. Next, we address the combined effects of these processes on the residence times of suspended particles and, thus, the residence time of the cloud.

4.1. Formation of a stable plume

The formation of a stable plume as opposed to a collapsing fountain has been extensively studied in the context of explosive volcanic eruptions as it is of crucial importance for hazard assessment (Carazzo et al., 2008a; Suzuki et al., 2005; Valentine and Wohletz, 1989; Wilson et al., 1980). Here, we use our experimental data on the onset of column collapse to construct a simple 1D model based on a “top-hat” formalism (see Appendix A).

Our model predicts column collapse in a way consistent with our experimental results (Fig. 5). We note that when partial collapse occurs, the top-hat formalism does not apply because the jet is not in steady state. Thus, we put partial collapse in the collapsing regime. An additional observation from Fig. 5 is that the transition from the layering to the settling-driven fingering and late collapse regimes is observed for a critical Richardson number at the vent of −4 × 10−4. Because low-RiL jets tend to reach greater heights in the tank than high-RiL jets and thus promote turbulent mixing at the salt/fresh water interface when falling back from the maximum to the neutral buoyancy height, this result suggests that the evolution of the cloud depends strongly on the density gradient produced as a result of mixing across the initial density interface.

The good agreement between theoretical predictions and our experimental results suggests that the simple top-hat formalism captures the essential dynamics of our multi-phase jets. Consequently, this model can be used to calculate the key dynamical quantities at the neutral buoyancy height, which impose the conditions in the cloud.

4.2. Formation of particle boundary layers

The major qualitative features of the settling-driven fingering and the layering regimes are analogous to those observed in sedimentation-driven convection (Hoyal et al., 1999b; Kerr, 1991) and in diffusive convection in particle-heat systems (Green, 1987; Hoyal et al., 1999a). The periodic structure of the layered clouds issuing from low-RiL jets has also been observed in numerous past volcanic clouds including the 2010 Eyjafjallajökull eruption (Dacre et al., 2011; Folch et al., 2012). Whereas wind shear effects have been suggested to explain the formation of these thin ash layers, our experiments show that layering may even occur in the absence of wind, which suggests that an important mechanism is missing in our understanding of this phenomenon. The full characterization of our layering regime and its implications for the interpretation of ash concentration measurements will be addressed in greater detail in a future study. Here, we focus on the formation of PBLs in clouds issuing from high-RiL jets.

To quantify the tendency of the cloud to form a particle-rich layer at its base, we introduce the dimensionless number B such as (Marsh, 1988).

\[ B = \frac{V_i}{V_g} \]

(10)

where \(V_i\) is the growth rate of the gravitational instability given by (Whitehead and Luther, 1975).

\[ V_i = \frac{\Delta \rho_i g \Delta \rho_{col}}{3 \mu} \]

(11)

where \(\Delta \rho_{col}\) is the thickness of the PBL, and \(\Delta \rho_{i} = \rho_{bot} - \rho_{ov}\) with \(\rho_{bot}\) and \(\rho_{ov}\) the PBL and salt water densities, respectively. The number \(B\) characterizes the preferential mode of sedimentation. Where \(B < 1\) individual particle settling dominates, whereas where \(B > 1\) PBL formation is enhanced. The effective density of the particle boundary layer (\(\rho_{bot}\)) is not straightforward to estimate a priori for two reasons: (i) the time-averaged particle concentration within the PBL (\(c_{bot}\)) is not a simple function of the source conditions, and depends also on the dynamics of umbrella spreading; and (ii) the concentration of particles increases with depth across the PBL. However, the effective amount of particles in the PBL can be approximated using the particle concentration in the entire cloud (\(c_{bot}\)) calculated using our model of turbulent jets (Table 1). Since PBL is much thinner than the jet
wide at the LNB, conservation of volume requires that particle concentration in the PBL can be as much as 10 times bigger on average (i.e., \( \epsilon_{pbl} = 1 - 6 \times 10^{-2} \)). Replacing these values in Eq. (10) together with \( V_s = 0.07 \text{ m s}^{-1} \) and \( \delta_{pbl} = 0.5 - 1 \text{ cm} \) as observed in our experiments, we calculate \( B = 10 - 400 \). In our experimental configuration, PBLs are therefore likely to form before particles settle individually.

4.3. Finger formation

Particle-laden fingers observed in our experiment are likely to result from the growth and intermittent detachment of the PBL at the cloud base. To confirm this hypothesis, we use the simple conceptual model developed by Hoyal et al. (1999b) on the basis of well-established theory for turbulent thermal convection (Howard, 1964; Turner, 1973).

Assuming that mass transfer is independent of the layer depth (Howard, 1964; Turner, 1973) and thus governed by processes local to the boundary layer, the critical conditions for finger formation are defined by a local Grashof number,

\[
Gr_{pbl} = \frac{g'_{pbl} \delta_{pbl}^3}{\nu_{pbl}^2} > 10^3, \tag{12}
\]

where \( \nu_{pbl} \) is the effective kinematic viscosity of the particle boundary layer, and \( g'_{pbl} = g(\rho_{pbl} - \rho_s)/\rho_a \) is the reduced gravity of the PBL. The effective kinematic viscosity of the PBL can be approximated by using the Einstein–Roscoe formula,

\[
\nu_{pbl} = \nu_a (1 - \Gamma)^{-2/5}, \tag{13}
\]

where \( \nu_a \) is the kinematic viscosity of the salt water and \( \Gamma = \epsilon_{p}/\epsilon_{pmax} \) is an effective particle concentration, with \( \epsilon_{p} \) the concentration at the source and \( \epsilon_{pmax} \) the effective maximum concentration determined from our experiments (~0.1). Applying this approximation, the viscosity ratio \( \nu_{pbl}/\nu_a \) is in the range of 1.3–16 for our experiments. We note that no turbulent motions are observed within the PBL consistent with earlier analyses of turbulent Rayleigh–Taylor instabilities (Linden and Redondo, 1991).

Applying the critical Grashof number definition (Eq. 12), the critically unstable PBL will have a thickness that scales as

\[
\delta_{pbl} \sim 10 \left( \frac{\nu_{pbl}^2}{\nu_{pbl}} \right)^{1/3}, \tag{14}
\]

which has been quantitatively tested in previous laboratory experiments (Hoyal et al., 1999b). This relationship suggests that settling-driven fingering can always form if the particle concentration in the PBL (\( \epsilon_{pbl} \)) remains large enough while the PBL grows to the critical thickness. This condition implies that most particles do not settle individually through the cloud/salt water interface (i.e., \( B > 1 \)), which is the case in our experiments.

To determine whether or not the growth and destabilization of PBLs lead to fingers formation, we analyze oscillations at the base of the cloud. Indeed, in settling-driven fingering systems, the vertical extent of the fingers depends on the slow diffusion of salt, which enhances their buoyancy, and the sedimentation of particles from their ends, which reduces their buoyancy relatively more rapidly. The loss of buoyancy causes a descending finger to slow and then rise back towards the base of the cloud, where it overshoots the LNB and excites an oscillation at the base of the entire umbrella. Since the thickening of the PBL is controlled by the flux of particles entering into the layer, we expect that,

\[
\delta_{pbl}^3 \sim T_{max} Q_{pbl}. \tag{15}
\]

where \( T_{max} \) is the longest measured period of umbrella oscillation, and \( Q_{pbl} \) is the flux of particles. Here, \( Q_{pbl} \) is taken to be proportional to the flux of material cascading from the top of the jet when the flow falls back from its maximum height to the neutral buoyancy height. Therefore, \( Q_{pbl} \) scales as \( A_{ml}(g' \Delta z_{ml})^{1/2} \), with \( A_{ml} \) and \( g' \Delta z_{ml} \) the surface area and reduced gravity of the jet at the maximum height, which can be calculated from the source conditions using our model of jets (Appendix A). The longest period of cloud oscillation \( (T_{max}) \) is calculated using measurements of the displacement of the cloud/salt water interface (Fig. 6A). These measurements are performed twice for each experiment at different locations along the distance from the source providing a 95\% confidence level. The longest period of cloud oscillations is then determined from spectral analysis using the method of Welch, 1967 and a Thompson multitaper algorithm.

Fig. 6B shows that Eq. (15) is verified suggesting that oscillations and fingering instabilities result from the growth and destabilization of PBLs at the cloud bases. An additional remark is that quasi-periodic oscillatory behavior driven by convective PBL dynamics is not ubiquitous. In particular, for higher values of \( T_{max} Q_{pbl} \) initial oscillations grew rapidly in amplitude (Fig. 6A), leading, in turn, to a late collapse of the entire umbrella cloud and the formation of a turbulent gravity current. This observation suggests that a further analysis of the stability of the PBL could lead to a prediction for this catastrophic collapse.

![Fig. 6](image-url) (A) Measurements of interface displacement as a function of time for two experiments in the layering regime (dotted curve) and in the late collapse regime (solid curve). (B) Comparison between the estimated critical thickness of the particle boundary layer (\( \delta_{pbl} \) defined on the sketch) and the longest period of the oscillations observed. Each symbol corresponds to one experiment reported in Fig. 4.
4.4. Impact on the residence times

The new PBL dynamics identified in our experiments raise the question of the residence time of suspended material in the cloud. To quantify this effect, we introduce the dimensionless number \( T \) such as

\[
T = \frac{T_f}{T_s},
\]

where \( T_s = H/V_s \) and \( T_f = H/V_f \) are the residence times when individual particle settling and when PBL dynamics dominates, respectively. Here, \( V_s = g \delta_p^{3/5} \) is the characteristic velocity of flow into the resulting fingers with \( Q_f = \pi V_s \delta_p^2 / 4 \) the corresponding volume flux into the fingers. Replacing variables in Eq. (16) together with Eq. (14), one can re-write \( T \),

\[
T \approx \frac{0.42 V_s^{4/5}}{\left( g \delta_p V_s \right)^{4/15}}
\]

Replacing the parameters in this equation with consistent values, we estimate \( T \approx 0.5 \) to 2 suggesting that particle-rich fingers at the base of the cloud may either promote or impede particle sedimentation in our experiments depending on the effective particle content in the PBL. We note that in spite of the uncertainties on \( \epsilon_{pbl} \), this result corresponds to finger velocities \( V_s \approx 3 \sim 14 \) cm s\(^{-1}\) consistent with our observations (Movies 1 & 2).

5. Implications for volcanic ash-clouds

We have shown that large-scale gravitational instabilities at the cloud bases can arise in our experiments as a result of the growth and intermittent overturn of thin PBLs formation. Our experimental results are well understood with a theoretical model for the jet and a scaling theory for the cloud. The scaling analysis presented in Section 2.2 suggests that our experimental multiphase flows are reliable analog for the transport and sedimentation of fine (low \( S_t \)) volcanic ash in the atmosphere. Thus, we now investigate whether or not PBLs formation influence the behavior of volcanic clouds and can explain the intriguing features shown in Fig. 1.

5.1. Particle boundary layers in volcanic clouds

To quantify the tendency of volcanic clouds to form PBLs at their base, we use the \( B \) number introduced in Section 4.2 (Eq. 10). In volcanic ash-clouds, however, particle settling velocities vary from large particles falling at high Reynolds number to small particles falling at low Reynolds number. Using the formulae of Bonadonna et al., 1998 for \( V_s \) valid for a range of particle Reynolds number leads to three formulations for \( B \):

\[
B_{\text{low}} \approx \frac{\Delta \rho \rho_f}{\rho_p d_p^2} \text{ for } \text{Re}_p < 0.4,
\]

\[
B_{\text{int}} \approx \frac{75 \Delta \rho \rho_p^{1/3} \rho_f^{1/3} V_s^{2/3} \delta_p^{1/2}}{\mu d_p^{1/2}} \text{ for } 0.4 \leq \text{Re}_p < 500,
\]

\[
B_{\text{high}} \approx \frac{1}{3} \frac{\Delta \rho \rho_f^{1/2} \rho_p^{1/2} V_s^{1/2} \delta_p^{1/2}}{\mu d_p^{1/2}} \text{ for } \text{Re}_p > 500.
\]

Replacing \( \delta_p \) with Eq. (14) and assuming that \( \rho_f \sim \rho_p \) leads to a generic form for \( B \) appropriate over the full range of \( \text{Re}_p \) conditions:

\[
B \approx k \xi_p^{1/3} \left[ \frac{\rho_p d_p^2}{\rho_f d_p^2} \right]^{n/3},
\]

where \( \xi_p = \epsilon_{pbl} / (1 - \Gamma)^3 \), \( k \) is a geometric constant and \( n \) is a constant that depends on the form of the drag law. For high-\( \text{Re}_p \) particles, \( k = 33 \) and \( n = 1/2 \), whereas \( k = 128 \) and \( n = 1 \) for intermediate-\( \text{Re}_p \) particles, and \( k = 600 \) and \( n = 2 \) for low-\( \text{Re}_p \) particles.

Inspection of Eq. (21) indicates that \( B \) is governed primarily by particle size \( (d_p) \) and is relatively insensitive to the particle volume fraction \( (\epsilon_{pbl}) \) over the range in \( \text{Re}_p \). The latter quantity is not straightforward to evaluate without measurements of oscillation frequency at volcanic cloud bases. Here, we assume that particle concentrations in the PBL lie within the maximum range of values for the entire cloud \( (\epsilon_{pbl}) \) and up to ten times more concentrated, consistent with our experiments. Particle concentrations in the cloud are found to be \( \epsilon_{pbl} \approx 10^{-6} \sim 10^{-4} \) using the model Carazzo et al. (2008b) for various source and environmental conditions including gas contents at the vent ranging from 7.5 \( \times \) \( 10^{-4} \) to \( 10^{-2} \) vol\% (Table 2). This range of values for \( \epsilon_{pbl} \) is found to be fully consistent with previous theoretical calculations (Sparks et al., 1994; Woods, 1995) and direct measurements in ash clouds (Bursik et al., 1994). Thus, we retain \( \epsilon_{pbl} = 10^{-6} \sim 10^{-4} \) as a reasonable range for particle concentration in the PBL.

Fig. 7 shows the expected transitional conditions \( (B = 1) \) over a wide range of particle sizes, densities, cloud altitudes, and for the full range of observed \( \text{Re}_p \) conditions. Estimations of \( B \) for high-\( \text{Re}_p \) particles show that not surprisingly large particles (typically \( > 1 \sim 6 \) mm) always settle individually (Fig. 7A). On the other hand, the sedimentation of low-\( \text{Re}_p \) particles is expected to be governed by PBL dynamics (Fig. 7C). Lastly, intermediate-\( \text{Re}_p \) particles are likely to either settle individually or to form PBLs depending on the effective particle concentration (Fig. 7B).

5.2. Comparison with natural data

The analysis presented above shows that the mode of particle sedimentation in volcanic ash-clouds is mainly governed by the grain size distribution and to a lesser extent by the altitude reached into the atmosphere. With an aim of testing the theoretical predictions with field observations, we reviewed data available on 16 eruption phases including phreatomagmatic, Vulcanian and Plinian events (Fig. 8). We use total grain size distributions to determine a mean value and a sorting coefficient for error estimations. The cloud altitude is estimated by using the empirical relationship \( z_m \sim 0.7z_{m0} \), where the maximum column height \( z_{m0} \) is inferred from satellite measurements (Carey et al., 1990; Holasek et al., 1996) or from the distribution of lithic material in the deposits around the vent (Carey and Sparks, 1986).

Fig. 8 compares the collected field observations with the theoretical transition \( B = 1 \). Because low-\( \text{Re}_p \) particles always tend to form PBLs (Fig. 7C), whereas high-\( \text{Re}_p \) particles always tend to settle individually (Fig. 7A), the only relevant curves for Fig. 8 are those given in Fig. 7B for intermediate \( \text{Re}_p \) particles. We note, however, that for the lowest \( \epsilon_{pbl} \) considered, all intermediate-\( \text{Re}_p \) particles should settle individually (Fig. 7B). For this specific condition \( (\epsilon_{pbl} = 10^{-6} \) ), the transition \( B = 1 \) is simply given by the low-/intermediate-\( \text{Re}_p \) transition (Fig. 8).

The comparison between theoretical predictions and field data shows that depending on the effective amount of particles in the PBL, particles larger than \( 63 \sim 125 \) \( \mu \)m at sea level and \( 250 \) \( \mu \)m \( \sim 1 \) mm in high atmosphere should settle individually, whereas finer particles are more likely to settle collectively and to form PBLs. In
the majority of the Vulcanian and Plinian eruptions shown in Fig. 8 settlement is expected to be governed by individual particle settling, as assumed in turbulent suspension models (Bursik et al., 1992). A number of Plinian events, however, including the well known 1980 Mt. St. Helens and 1982 El Chichon eruptions are expected to be in the PBL regime.

An important remark is that whereas the population of large pumice and lapilli can be accurately reconstructed from deposits in the field, the population of the finest ash is more difficult to assess because small particles remain in suspension for weeks and fall out far from their point of emission. Therefore, the median grain-sizes reported in Fig. 8 represent maximum values and it is possible that more Vulcanian and Plinian eruptions could be in the PBL regime. This assumption is supported by the thorough re-analyses of the 1980 Mt. St. Helens and 1982 El Chichon deposits, which show that the proportion of fine ash has been underestimated in previous studies (Durant et al., 2009; Rose and Durant, 2008). Lastly, the collected field observations suggest that PBL formation may have been important in highly-explosive phreatomagmatic eruptions producing fine volcanic ash including the recent Eyjafjallajökull eruption (Fig. 8).

The origin of mammati in atmospheric and volcanic clouds remains poorly understood (Schultz et al., 2006). Our results support the idea that mammati clouds observed during the 1980 Mount St. Helens eruption (Fig. 1A) formed as a result of the growth of a PBL at the base of the umbrella cloud (Durant et al., 2009). Overall, our results suggest that internal sedimentation of the finest fraction of particles may lead to the formation of PBLs at the base of volcanic clouds. Whether these PBLs will result in the formation of rounded mammati clouds, particle-rich fingers or thin ash veils depends on the characteristics of the PBLs (Section 4.3). Consequently, more work is needed to characterize the dynamics of these PBLs in the natural case. In particular, real-time measurements of particle sizes and concentrations at the base of volcanic clouds would be valuable.

3.5. Impact on cloud stability and longevity

In spite of the lack of field data required to interrogate the stability of PBLs in nature, it is interesting to quantify the effect of finger formation on the residence time of suspended particles. For this, we use the \( T \) number defined in Section 4.4 considering only low-\( R_p \) particles which always form PBLs,

\[
T \approx 0.42 \left( \frac{g}{\nu^2} \right)^{1/4} \left( \frac{1 - \Gamma^{5/4}}{\rho_p^{1/2}} \right)^{3/5}, \tag{22}
\]
where \( g' = g(\rho_f - \rho_a) / \rho_a \). Applying typical values for volcanic ash clouds (see Section 2.3) leads to \( T \approx 10^{-1} \) for 10 \( \mu \)m-particles and \( T \approx 10^{-2} - 10^{-3} \) for 1 \( \mu \)m-particles. This result suggests that depending on the grain-size distribution in the PBL, finger formation reduces the residence time of fine particles in volcanic ash-clouds by at least one and up to three orders of magnitude compared to the process of individual particle settling. These estimates should be considered carefully since other mechanisms affecting the dynamics of fingers may play an important role.

Particle aggregation is a particularly important phenomenon in volcanic ash clouds (Rose and Durant, 2011). Effects related to this process are not observed in our experiments, either in the cloud or in the resulting deposit, most likely because conditions for aggregation in water are not met for our 300 \( \mu \)m-particles. In natural eruption clouds, however, particle aggregation will cause the lowest-\( \Re_b \) particles to enter into higher-\( \Re_b \) regimes. Whether this phenomenon will enhance individual particle settling or PBLs formation is not straightforward to evaluate. Indeed, aggregation processes increase the mean particle sizes, but also reduce the effective particle densities and to a lesser extent particle concentrations in the PBL. On the base of Eq. (21) alone, one can infer that an increase of particle sizes should promote individual particle settling, whereas a reduction of particle densities should promote PBL formation. Because both effects alter the effective value of \( B \) in opposite manner, and also because their relative importance varies according to the \( \Re_b \) regime, an exact treatment of this problem would require a more sophisticated model that would not be supported by our experiments. Nevertheless, we argue that a significant fraction of the original particle size distribution should remain in a \( \Re_b \) regime that favors PBL formation, as is apparent in the observations summarized in Figs. 1 and 8.

An additional important effect controlling the behavior of volcanic umbrella clouds is the presence of atmospheric winds. High velocity winds can advect the cloud, enhance turbulent mixing and may impede the formation of fingers at the base of the cloud (Linden, 1974). This effect can be quantified by comparing the time scale for turbulent shear \( W_o / H \) and for flow into a finger \( h_f / V_f \)

\[
T^* = \frac{W_o h_f}{V_f H}
\]  

(23)

where \( W_o \) is the wind speed and \( h_f \) is the finger length. Fingers will form if \( T^* > 1 \) and will be impeded if \( T^* < 1 \). The maximum finger length can thus be estimated by replacing \( T^* = 1 \). For the 1980 Mt. St. Helens eruption, we estimate \( W_o = 30 \text{ ms}^{-1} \) (Carey and Sparks, 1986; Durant et al., 2009), \( H = 10 \text{ km} \) (Sarna-Wojcik et al., 1981) and \( V_f = 0.55 \text{ ms}^{-1} \) consistent with a mean grain size of 4.4\( \phi \) (Fig. 8). Replacing these values in Eq. (23) yields \( h_f \approx 185 \text{ m} \), which seems to be consistent with observations in Fig. 1A. We note that in the presence of very high-speed winds (\( W_o = 60 \text{ ms}^{-1} \), \( h_f \approx 91 \text{ m} \) and fingers are still expected to form. An additional remark is that in the near absence of wind (\( W_o < 0.5 \text{ ms}^{-1} \)), \( h_f \approx 11 \text{ km} \) and fingers are expected to reach the ground and possibly to spread as gravity currents. No-wind conditions during an eruption are rare, but possible as inferred from Plinian fallout deposits with a radially symmetric distribution (Papale and Rosi, 1993). We argue that in the near absence of wind, the 1980 Mt. St. Helens eruption cloud could have formed possibly dangerous gravity currents at much larger distances from the volcanic vent than ever considered before.

6. Conclusions

We have presented a new set of laboratory experiments understood with theory in order to understand the origin of the enigmatic buoyancy effects observed at the base of volcanic clouds. We show that particle-laden umbrella clouds may follow unexpected dynamical regimes depending on the flow intensity, the strength of the initial environmental density stratification and the particle concentration. Internal sedimentation of the fine fraction of particles during cloud spreading leads to the growth of thin PBLs at cloud bases. For natural eruptions, whether the dynamics of PBLs play a major role in volcanic sedimentation depends on the grain size distribution, particle concentration and to a lesser extent plume height. Although large and heavy particles settle individually, the remaining fine fraction of particles is likely to form PBLs and subsequent fingers. An exhaustive review of available field data from historic and prehistoric eruptions reveals that PBLs formation affected sedimentation in varied types of sub-aqueous and subaerial explosive events. The impact of these new dynamics is mainly to reduce the residence time of fine particles by one to three orders of magnitude compared to the process of individual particle settling.

Our work explains the origin of the enigmatic buoyancy effects observed in some recent volcanic ash-clouds (mammatus, particle-rich fingers, ash veils, destabilization of the whole umbrella cloud) and provides insights into understanding the formation of finely stratified pumice deposits (Branney, 1991). Catastrophic overturn and collapse of the umbrella could present a potentially significant and previously unrecognized volcanic hazard. Understanding whether our late collapse regime could occur in nature and lead to the occurrence of late pyroclastic flows is therefore an exciting future direction of investigation.

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\( \alpha_c \) entainment coefficient
\( \delta_{p,0l} \) critical particle boundary layer thickness (m)
\( \epsilon_p \) particle volume fraction
\( \epsilon_{p,\text{av}} \) effective maximum particle concentration
\( \mu \) fluid dynamic viscosity (Pa s)
\( \nu \) fluid kinematic viscosity (m\(^2\) s\(^{-1}\))
\( \nu_a \) kinematic viscosity of the salt water (m\(^2\) s\(^{-1}\))
\( \rho \) mixture density (kg m\(^{-3}\))
\( \rho_a \) ambient density (kg m\(^{-3}\))
\( \rho_f \) fluid density (kg m\(^{-3}\))
\( \rho_p \) particle density (kg m\(^{-3}\))
\( \rho_r \) reference density (kg m\(^{-3}\))
\( \tau_\text{f} \) time scale for flow (s)
\( \tau_\text{s} \) time scale for sedimentation (s)
\( \xi_{\text{pbl}} \) particle response time (s)
\( \Sigma \) sedimentation number

**Subscripts**
- \( o \) values at the source of the jet
- \( f \) values in the finger
- \( m \) values at the maximum jet height
- \( \text{pbl} \) values in the PBL
- \( u \) values at the source of the cloud

where \( \text{Ri}=g R U^2 \) is the local Richardson number, and \( A \) is a dimensionless parameter that depends on the flow structure (Carazzo et al., 2006). The latter parameter has been constrained using laboratory measurements on various jets and plumes, and can be calculated using Eqs. (13) to (17) of Carazzo et al., 2008b. Eqs. (24) to (27) can be used to describe the self-similar evolution of the jet and thus to predict the conditions leading to the formation of a stable plume (Fig. 5).

**References**


