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The mysterious grooves of Volcán Bárcena: a review of the role of streamwise counter-rotating vortices during erosion by dilute pyroclastic density currents

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Abstract

Although erosion during high-energy passage of a pyroclastic density current (PDC) causes great damage, analyses of the effects of such erosion are sparse in scientific literature compared to observations and interpretations of depositional processes. In this paper, we review observations of surfaces where PDCs have eroded sets of grooves that provide information on the erosion process. We postulate that in some cases, the grooves were carved by streamwise vortices in the boundary layer of the PDC and review possible fluid dynamic instabilities that can give rise to such vortices. For the prominent grooves at Volcán Bárcena, Mexico, we propose that a fluid dynamic instability, which we dub the “groovy instability,” occurred and caused formation of erosive counter-rotating vortices. This instability occurs when the particle concentration boundary layer thickness, δ_c , is larger than the velocity (shear) boundary layer thickness, δ_u , i.e., $L = \delta_c / \delta_u > 1$. In subaqueous turbidity currents, these vortices have a typical wavelength of $\sim 25 \delta_c$. If this relation is applied to the grooves formed on Volcán Bárcena, the inferred particle concentration boundary layer is estimated to have been < 1 m thick. We postulate that a transition between erosion of grooves and deposition of dunes at Volcán Bárcena occurred when hydraulically supercritical flow on the upper flanks changed to subcritical flow about halfway down the mountain. We call attention to boundary layer dynamics in erosive pyroclastic density currents at a dimension that is difficult to scale quantitatively in laboratory experiments and is usually not resolved computationally and to the need for incorporating such dynamics into models of PDC dynamics.

Keywords Pyroclastic density currents · PDC · Grooves · Erosion · Vortices · Volcán Bárcena

Introduction

Pyroclastic density currents (PDCs) have a complicated fluid dynamic structure that evolves in both space and time as a

PDC propagates from its source (see literature citations in Lube et al. 2020; Brosch and Lube 2020). Most of what is known about PDCs, and their hypothesized internal structure, comes from studies of the deposits left behind and sedimentation processes (e.g., Moore and Sisson 1981; Fisher et al. 1987; Valentine 1987). However, the destructive power of PDCs can be increased by erosion of the substrate and entrainment of particles into the body of the flow (e.g., Kieffer and Sturtevant 1988; Thouret 1990; Sparks et al. 1997; Branney and Kokelaar 2002; Scarpati and Perrotta 2012; Brand et al. 2014; Bernard et al. 2014; Dufek et al. 2015; Roche 2015; Dufek 2016; Fauria et al. 2016; Pollock et al. 2019). Sediment entrained by erosion of the surface affects the dynamics of other flows such as powder snow avalanches and turbidity currents (e.g., Parker et al. 1986; Rastello and Hopfinger 2004; Hall et al. 2008). However, less is known about the erosion and entrainment processes during propagation of PDCs, or their structure during the erosive phase, because observations are scarce, measurements within the high-

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energy multiphase flow are difficult, and exposures of erosion surfaces in the field are limited.

Erosion surfaces at a number of volcanoes contain distinctive sets of grooves, and we suggest herein that these groove sets contain information about the erosion process and the structure of the flow during erosion. Such sets were prominent enough to capture the attention of early explorers: Darwin noted that they were eroded into the flanks of Beagle Crater in the Galapagos Islands (reported in Grant and Estes 2009). Lacroix (1904) noted erosion on the slopes of Mount Pelée, Martinique, mentioning “deep parallel grooves, hollowed out...in ancient tuffs,” and cited others who had found them on Soufrière St. Vincent. Richards (1956) provided the first extensive documentation of sets of grooves with his observations of the 1952 eruption of Volcán Bárcena, Mexico, and attributed them to erosive “tephra avalanches.” Moore (1967) noted that the grooves at Volcán Bárcena changed to depositional dunes on the lower slopes and compared them with features observed at Taal volcano in the Philippines. Kieffer and Sturtevant (1988) described grooves eroded into substrates by the 1980 lateral blast of Mount St. Helens, USA. Pierson et al. (1990) illustrated a set of grooves in the ice cap of Nevado del Ruiz, Colombia, and noted their similarity to the grooves carved in erodible materials by pyroclastic density currents at other volcanoes, such as Mount St. Helens and Volcán Bárcena. Sparks et al. (1997) and Calder et al. (2000) documented grooves at Lascar Volcano, Chile, pointing out that some (but not all) grooves were downstream of boulders and that it was necessary to ascertain to what extent the grooves represent a “natural periodicity and circulation within the flow with a scale related to flow thickness and velocity or are a reflection of the characteristic roughness of the substrate.” Our results presented below confirm this suggestion of natural periodicity and circulation within the flow.

In this paper, we first review known occurrences of sets of erosional grooves carved by pyroclastic flows and compare and contrast suggested mechanisms for groove formation. We then suggest a new mechanism for groove formation based on descriptions of grooves at Volcán Bárcena. We demonstrate that sets of grooves on the erosion surface reveal characteristics of the erosion process and the structure of the boundary layer of the flow during an erosive stage. We suggest a model for the structure of the basal layer of an erosional PDC that forms grooves. Our model is developed for the properties of *dilute PDCs*, and we emphasize that our focus is on the structure of a PDC during *erosion*, a stage of flow for which few observations have been reported, rather than sedimentation. However, in the “Discussion” section, we speculate on how studies of the basal layer of sediments overlying an erosion surface may help with interpretation of the erosion process. We conclude by suggesting some profitable areas for future research.

We emphasize here that we are considering only grooves that occur in sets, in contrast to isolated individual channels or small-scale striae caused by impacting or dragging debris. All descriptions in this paper are from the literature as we did not visit any sites described other than the erosion surface of the 1980 lateral blast at Mount St. Helens (Kieffer and Sturtevant 1988).

A review of erosion surfaces with grooves

Sets of grooves have been eroded into the substrate during at least six volcanic eruptions, e.g., Mount St. Helens, USA (1980); Lascar, Chile (1993); Nevado del Ruiz, Colombia (1985); Soufrière Hills, Montserrat (1997); and Crater Lake, USA (~7700 years ago). Descriptions in the literature of grooves at Taal Volcano from the 1965 eruption (Moore et al. 1966), at Cerro Colorado, Mexico (Wohletz 1980), Latera and Baccano craters, Italy (Mattson and Alvarez 1973), and Beagle Crater, Galapagos [reported in Grant and Estes 2009], are not detailed enough to allow determination of whether these features are isolated channels or sets of grooves comparable to those described herein. The U-shaped channels on Koko Head (Oahu, Hawaii, USA) described by Fisher (1977) are often cited in the literature on grooves, but it is not possible to tell from the literature if these are groove sets or relatively isolated channels.

Mount St. Helens (1980)

The type locale for sets of grooves created by a pyroclastic flow is at Mount St. Helens where grooves were eroded into the substrate by the lateral blast on May 18, 1980 (Fig. 1; Kieffer and Sturtevant 1988). The grooves have wavelengths of ~7 m and lengths up to ~100 m. They occur at distances of 3.5 to 9 km from the crater in two settings: regions of flow reattachment downstream of sheltered regions (Fig. 1) and on ridges oriented at an angle to the flow direction (Fig. 6 in Kieffer and Sturtevant 1988). These grooves are no longer visible in satellite images or in the field because they are obscured by vegetation.

Kieffer and Sturtevant (1988) attributed the grooves in different settings to two flow instabilities that produced streamwise longitudinal vortices: the Görtler instability that can arise when flow follows a concave wall and the crossflow instability that can occur when flow is subparallel to an obstacle such as a ridge crest.

Volcán Bárcena, Mexico (1952)

With the possible exception of the grooves resulting from the 1980 lateral blast at Mount St. Helens, the most spectacular and best-described set of grooves formed by a pyroclastic

Fig. 1 Grooves formed during the lateral blast at Mount St. Helens in 1980 in relatively unconsolidated volcanic ash on the headwall of South Coldwater Creek. Flow was from right to left. For scale, note geologist

(lower left corner). Reprinted from Kieffer and Sturtevant (1988) with permission

density current (or currents) occurs on the flanks of Volcán Bárcena located (Fig. 2) on Isla San Benedicto, Mexico, approximately 480 km off the western shore of Mexico. Because these grooves were carved on a freshly created volcanic edifice that was newly emergent from the ocean and had not undergone significant weathering and hillslope/fluvial reworking, their properties, such as dimensions and wavelengths, provide a unique data set. We discuss these grooves in detail and propose a new mechanism for groove formation later after giving this overview of known grooved surfaces.

Lascar Volcano, Chile (1993)

The eruption of Lascar on April 19–20, 1993, produced pyroclastic density currents that eroded the landscape on many scales—from large-scale removal of vegetation and soil and sedimentary deposits to fine scale abrasion that left striae on individual boulders (Fig. 3, Sparks et al. 1997; Calder et al. 2000). The flow was so erosive that when it encountered coarse breccias containing minerals and rocks of differing hardness, it smoothed them uniformly (note the large clast eroded flush with the surface in the middle of Fig. 3). Scree and talus were stripped from channels and steep slopes of the volcano. Grooves were created on the heavily eroded surface in settings where velocities would have been high, e.g., on steep slopes or

where flow was accelerated through constrictions. The grooves had wavelengths of 0.5–2 m and depths of 0.1–0.3 m. The wavelength of some of these grooves may have been determined by bedding in older pyroclastic density current deposits. Some, but not all, grooves were associated with large boulders in a crag-and-tail geometry. Sparks et al. (1997) concluded from their observations of the distribution of pumice-rich flows versus lithic-rich flows that the flows were segregated into a lithic-rich basal part and a pumice-rich upper part and that the basal lithic-rich part of the flow was responsible for much of the erosion. They called these flows “high-concentration granular avalanches in which particle interactions dominated.” Particle sizes in the lithic-rich deposits ranged up to 1 m.

No flow instability mechanism has been proposed for the creation of these grooves.

Nevado del Ruiz, Colombia (1985)

On November 13, 1985, pyroclastic density currents poured down across snow and ice on the summit ice cap of Nevado del Ruiz generating catastrophic lahars that killed more than 23,000 people in the town of Armero. These pyroclastic density currents melted snow and then mixed with liquid water generated by the melting, this interaction enhancing the formation of the lahars (Pierson et al. 1990; Pierson 1995). Some

Fig. 2 (a) Grooves on the eastern flank of Volcán Bárcena, Mexico (Richards 1956). Crater has a diameter of 700–750 m, and grooves have typical widths of ~3–6 m. Light-colored areas are ash from later eruptions. (b) A pyroclastic density current moving down the flank of Volcán Bárcena on August 12, 1952. No scale was provided by Richards (1956), but we estimate that it is of O (10m) in thickness at the front. The prominent lobe-and-cleft structure on the front of the flow are not related to the mechanism proposed herein. Photo is an enlargement from a 16 mm duplicate Kodachrome motion picture frame published by Richards (1956) and is one of the very few photos in existence from the 1952 eruption. There are no observations that limit the timing of groove formation, and it is possible that an earlier larger flow determined the characteristics of the grooves with later flows, such as this one, deepening and widening them. Reproduced from Richards (1956) with permission from Springer Publishing

ice was incorporated into the pyroclastic density currents, including one 2-m ice block that was moved 3.2 km from the source near the summit crater. The flows carved radial grooves into the snow on the steep slopes of the summit ice cap (Fig. 4). Pierson et al. (1990) estimated the relief in the grooved areas to be less than 2 m except where the grooves were eroded along pre-existing fractures (see also Thouret 1990). These are the only grooves reported to date in snow and ice. The grooves were not significantly deflected by ice topography or crevasses. Pierson et al. (1990) noted the similarity to grooves at Mount St. Helens and Volcán Bárcena. No flow instability has been proposed for the creation of these grooves.

Soufrière Hills, Montserrat (1997)

Sparks et al. (2002) documented extensive erosion by the 1997 Boxing Day eruption at Soufrière Hills, Montserrat. The steep slopes of Fergus Mountain were stripped bare and scoured to a height of at least 300 m above the valley floor (Fig. 5). Pronounced grooves with a wavelength in the range of 1–10's of meters (estimated from Fig. 18 in Sparks et al. (2002)) were carved into the hillside, generally away from the volcano but with local deviations showing some topographic influence. A debris avalanche deposit emplaced prior to the pyroclastic density current was also scoured and striated.

Fig. 3 Grooves in relatively unconsolidated sediments at Lascar (F in upper third of the image) and striae on harder rock (middle third of photo). For scale, note geologist standing on a lag breccia (lower third of photo). Flow was from left to right. This image is downstream of a funnel that accelerated the flow. Image provided by Stephen Sparks, published in Sparks et al. 1997, reprinted with permission

Grooves formed in the surface were overlain by tar that was especially well preserved within the grooves. Sparks et al. (2002) estimated that minimum internal speeds of 80–120 m/s were attained. No flow instability has been proposed for the creation of these grooves.

Crater Lake, Oregon USA (~7700 years ago)

Robinson et al. (2017) described two groups of linear subparallel ridges (“groove-ridge sets”) on pumice deposits from the ~7700-year-old deposits at Crater Lake, Oregon (Fig. 6). Analysis of the geometry of these ridges suggests that they

were carved by a scouring flow that originated from the north rim of Crater Lake. Each group originated as “a single large ridge on the lee side of a local topographic high, and the ridges bifurcate, increase in number, and decrease in height downslope from the highs” (Robinson et al. 2017). In one case, a single ridge that is 20 m high, 400 m wide at its base, and 500 m long branches into individual ridges that are 1–2 m high and 20 m wide. Grooves between the ridges are erosional and deep enough to cut through the pumice and expose andesitic lava. Robinson et al. (2017) pointed out a resemblance to grooves seen at Volcán Bárcena, Mount St. Helens, Lascar, Montserrat, and Koko Head.

Fig. 4 Summit of Nevado del Ruiz showing grooves in (mostly) snow on the flanks of the ice and snowfield. For scale, the crater at the top is about 0.8 km diameter wide, and the distance from the near rim of the crater to

the ice terminus in the lower right is about 1.8 km. Image courtesy of Thomas Pierson. From Pierson et al. (1990) reprinted with permission

They proposed that the groove-ridge sets were carved by vortices downstream of topographic obstacles, as had been proposed by Kieffer and Sturtevant (1988) for grooves at Mount St. Helens, and that these vortices split, increasing in number but decreasing in size. Wedge-shaped mounds between the two sets of groove-ridge sets may be remnants from the same process. No flow instability has been proposed for the creation of these grooves.

Case study: grooves on Volcán Bárcena as indicators of fluid-dynamic instabilities

With the possible exception of the grooves resulting from the 1980 lateral blast at Mount St. Helens, the most spectacular and best-described set of regularly spaced grooves formed by a pyroclastic density current (or currents) occurs on the flanks of Volcán Bárcena (Fig. 2) located on Isla San Benedicto, approximately 480 km off the western shore of Mexico (Fig. 7). These grooves were carved on a freshly created volcanic edifice that was newly emergent from the ocean and had not undergone significant weathering and hillslope/fluvial reworking when observed by Richards (1956).

The eruption of Volcán Bárcena on August 1, 1952, was only intermittently observed (e.g., Fig. 2b), but Richards

(1956) reconstructed the eruption sequence from stories of tuna fishermen in the vicinity at various times and from a few short scientific expeditions. The volcano was born and grew to maturity during a brief period of a few months in the summer of 1952: “Few volcanoes have grown as rapidly: few have had such a short life” (Williams 1952). Volcán Bárcena formed on the north flank of an eroded pyroclastic cone, Montículo Cinerítico, which is now present only as the prominent ridge that extends toward the south from the rim of Volcán Bárcena (visible in Fig. 7a). Richards (1956) pointed out the absence of “straight-type grooves” on Montículo Cinerítico and attributed the gullies there to rain erosion. In contrast, the grooves present on Volcán Bárcena are straight. We thus do not consider Montículo Cinerítico further, nor do we consider channels formed by rain within the linear grooves on Volcán Bárcena.

Richards (1956) reported that the rim of Volcán Bárcena was at ~335 m elevation when the eruption occurred, and the volcano had steep 33° slopes above ~152m. On the southeast side, there was no change of slope above the sea cliffs, but on other sides the slope decreased at about 152m elevation. In the decades since the eruption, erosion has reduced the summit to ~280 m elevation, a loss of 55m. We discuss the likely effect of the slope change on the pyroclastic density current at the end of the “Results” section.

Fig. 5 Erosional grooves (indicated by the orange arrow) on the shoulder of Fergus Mountain, Montserrat. The grooves appear as the bands particularly visible on the upper right shoulder of the central ridge slanting downslope toward center right. However, they occur everywhere, including across the concentric bands that are outcrops of

pyroclastic layers in the bedrock toward the center and bottom of the image. We estimate from Fig. 7 in Sparks et al. (2002) that the bright colored valley fill behind Fergus Mountain is 200 m wide. Photo is Fig. 18a in Sparks et al. (2002), reproduced with permission. Original image provided by Stephen Sparks

The substrate on the flanks of Volcán Bárcena was covered with tephra because Volcán Bárcena grew by tephra fallout from water-magma interactions. The only particles of boulder size were either near the rim positions or at sea-level and associated with the emission of lava there. Richards (1956, p. 112) speculated that the near-rim boulders “may have been ejected from the crater independently from the emission of tephra avalanches, or perhaps they were deposited as the tephra avalanches dilated when flowing over the crater rim.” In either case, it does not appear that large particles were a major part of the substrate into which the grooves were eroded.

Richards (1956) summarized the groove properties at that time as follows: Grooves began ~12–21 m below the rim and extended to the eroded sea cliffs on the southeast side. However, grooves on the east and west flanks generally, but not always, terminated at the change of slope. They were best developed below the low areas of the crater rim, that is, where pyroclastic density currents generated inside the crater were most likely to have spilled over the rim. The grooves were typically 3–6 m wide, ~1–3 m deep, and averaged ~240–270 m in length. The longest groove was ~424 m. In cross-section, the grooves were U-shaped, and crests between them were wider than the troughs. Individual grooves did not coalesce. The upper ends of the grooves may have been covered with

tephra, but they appeared to develop gradually from the small non-grooved area immediately adjacent to the rim. The groove width showed little variability or a slight divergence downslope. Boulders were present near the upper end in some, but not all, grooves. However, Richards (1956) provided no description of material erupted from the vent, nor the grain size or sorting of the substrate, nor of any deposits of material eroded from the grooves. The linear grooves typically contained smaller sinuous channels that appeared to resemble runoff channels. Although observations were very limited, Richards (1956) established that these drainage patterns had been formed within 6 weeks of the eruption.

At or below the change of slope, grooves were wider and shallower (see profile in Fig. 7b). Their lengths were variable, but they were typically only *c.* 60 m long. If the change of slope was abrupt, an individual linear groove appeared to terminate, but if it was gradual, the groove was generally recognizable below the change of slope. Where the slope change was gradual, grooves displayed a poorly developed herringbone pattern (strictly, the pattern is chevron rather than herringbone; however, we use the more common “herringbone” for consistency with older literature). In this region, some grooves coalesced, but the lower ends of grooves diverged downslope. Debris sometimes occurred at the lower ends of

Fig. 6 Two groups of linear ridges bounded by dashed lines on north flank of Timber Crater at Crater Lake, OR. Arrows point to andesite outcrops. The square box in A shows the location of topographic profiles across ridges shown in B. Reproduced from Fig. 3 in Robinson et al. (2017), with permission

the grooves when the change of slope was moderate (observed between Volcán Bárcena and Crater Herrera) (Fig. 8). On the southwest side of the cone, a field of transverse dunes with wavelength ~ 10 m was present at the break of slope (Moore 1967).

Richards (1956) listed seven features that must be addressed by a hypothesis to explain the origin of the grooves. Above the change of slope: (1) their origin close to the crater rim; (2) U-shape; (3) straightness; and (4) non-coalescing arrangement. Below the change of slope: (5) increasing width downslope, (6) “turbulent patterns” on the cone (which we

interpret to be the herringbone pattern shown in Fig. 3), and (7) local deposition at the lower ends.

Richards (1956) ruled out three possibilities: (1) rain erosion, because of the large size of the grooves and the absence of dendritic drainage; (2) erosion by lava bombs rolling downhill, by comparison with properties observed at Parícutin Volcano; and (3) ash landslides, by comparison of groove properties with landslide scars on the interior of the crater. By comparison with observations at Mount Pelée, Richards (1956) concluded that the grooves were eroded by “tephra avalanches,” avoiding the term “pyroclastic” because he did not see the incandescence implied by the term “pyro.” He reported the velocity of one avalanche (December 12, 1952) as “30 knots (15.4 m/s) or more.”

It is not known for sure how many density currents may have occurred during the active phase of Volcán Bárcena’s 1952 eruption. Richards (1956) stated it was likely that the eruption sequence began just when a tuna fishing boat was nearby on August 1, 1952. That activity was still in progress when the tuna boat left on August 4–5. Williams (1952) argued that ejecta were hurled to great heights by strong gas pressure, and that only later “as the vapor pressure diminished, did the weaker Vulcanian explosions, accompanied by explosions that lifted the ejecta only a short distance above the crater rim, form glowing avalanches that swept down the outer flanks of the cone.” When Volcán Bárcena was visited on August 12, 1952, it was in continuous eruption, and the photograph of the density current shown in Fig. 2b was obtained. According to Richards (1956), the eruptions “diminished in intensity in late August or early September.” He also states that the grooves were formed in August or September 1952, and that they “were nearly obliterated by superposed dendritic drainage by March 1953”. We assume that although there may have been multiple density currents, the characteristic wavelength of the grooves was set during an early eruption.

Flow Instabilities

We propose that the grooves were formed by erosion from counter-rotating streamwise vortices in a narrow boundary layer of the pyroclastic density current, a proposition that bears some similarities to past models of erosion by longitudinal vortices (Fig. 9). In this model, the grooves were eroded where the down-going fluid in adjacent vortices eroded into the substrate. The eroded material was then cycled through the vortices, either to be piled onto adjacent ridges or lofted into the main body of the flow. The existence of a preferred groove wavelength of O (1–10’s m) implies a strong fluid-mechanical selection process. In this section, we briefly outline other fluid dynamic instabilities that should be examined for such an analysis and consider which of these instabilities might be applicable to the grooves on Volcán Bárcena. We then discuss

Fig. 7 (a) Isle San Benedicto, with inset showing Richards' image from 1952. Main image from Planet.com Cubesat taken in 2019. Image courtesy of Planet Labs Inc. Montículo Cinerítico mentioned in the text is the ridge that extends toward the southwest from the central crater. Rain

gullies drain off of Montículo Cinerítico toward the W-SW and differ significantly in morphology from those on Volcán Bárcena. (b and c) Elevation profiles on west and east slopes respectively. Note shallowing of west slope at X between 100 and 150 m elevation

Fig. 8 Herringbone pattern on Volcán Bárcena grooves on the east side of the crater (Richards 1956, Plate XXII). White rectangle shows area where the herringbone pattern is particularly prominent. Grooves

estimated to be 3–6 m wide at the top of the image. Reproduced with permission from Springer Publishing

our model for vortex generation and apply the model to Volcán Bárcena. We discuss our assumptions and the likely mechanisms that terminated the grooves, before suggesting directions for future observations and research in the final section.

A number of flow instabilities can produce vortices at a geologic scale, but many of these can be negated for Volcán Bárcena as summarized in Table 1. This situation motivates a search for other mechanisms of groove formation.

When considered from the aspect of particle loading, PDCs have two end members—concentrated and dilute. Although dilute PDCs are one of the most frequently occurring PDCs (Lube et al. 2014, 2020), they have received less research attention than their concentrated end member. We assume that the PDC shown in Fig. 2b is typical of the PDC that formed the grooves at Volcán Bárcena. The lobe-and-cleft structure on the front of the flow indicates that the density, at least of the basal part of the flow, is greater than atmospheric density. On the other hand, since the PDC has just emerged from the crater rim region, we assume that there has been insufficient time for significant stratification to occur and that the billowy appearance of the flow, therefore, suggests that it is a dilute turbulent

PDC. Competing processes of sedimentation, erosion, and entrainment of both substrate material and ambient air as the flow travels down the slopes will determine the evolution of the flow as it progresses. We also invoke an assumption that the processes of erosion, entrainment, and mixing during the groove-carving are such that even though the flow may become stratified, the basal layer remains dilute. As we point out in the “Discussion” section, these assumptions provide a rich opportunity for future research in computational, laboratory, and field experiments.

The lobe and cleft structure is commonly seen on the front of pyroclastic density currents (e.g., Fig. 2b), but it is unlikely to be a mechanism by which long straight grooves can be eroded. Lobes and clefts develop via a fluid-mechanical instability within the foremost section of density current fronts, as a result of the unstable density stratification on the underside of such fronts (Hartel et al. 2000a, b; Cantero et al. 2007). This instability results in the formation of localized regions of counter-rotating streamwise vorticity (Simpson 1997; Nasr-Azadani and Meiburg 2014). The lobe-and-cleft instability exists only in the neighborhood of the current front, so that it can affect a given substrate location only during

Fig. 9 Helical flow structure produced by counter-rotating vortices and inferred bed erosion. From Karcz (1967). Reprinted with permission

the brief, transient phase when it passes this location and would thus be unlikely to exist for periods long enough to erode grooves to depths of meters during passage. In addition, the lobes and clefts meander, split and merge as they propagate (cf. Fig. 19 in Nasr-Azadani and Meiburg 2014), so that they would not be expected to give rise to the very straight, non-coalescing grooves observed on the slopes of Volcán Bárcena. The multiphase flow instability to be discussed below, on the other hand, can result in counter-rotating streamwise vortex pairs that extend over the entire length of the current.

Large-scaled grooves of dimensions comparable to those on Mount St. Helens and on Volcán Bárcena have been reported for grooves formed by some subaqueous turbidity currents, such as on the Mississippi delta region of the USA (Coleman 1981) and Cap de Creus, Spain (Lastras et al. 2007; Puig et al. 2008). Although there are differences between subaerial and subaqueous settings (Doronzo et al. 2016), the common features in both the scale and linearity of the grooves suggest that the mechanism of groove formation is independent of the phase (gas or liquid) of the ambient

medium into which the eroding flow intrudes. We propose that an instability known to occur under some conditions in multiphase flows, such as turbidity currents (Hall et al. 2008), could be responsible for the volcanic grooves, and for convenience, we herein refer to this as the “groovy instability.”

Results

Groove formation at Volcán Bárcena

Hall et al. (2008) investigated the mechanism by which the interaction of a subaqueous particle-laden turbidity current with an erodible sediment bed can result in the formation of streamwise grooves along the surface of the bed. Because the postulated instabilities occur at the base of the flow and away from its lateral or top boundaries with the ambient fluid, the nature of the fluid into which the density current intrudes is only secondarily relevant to the model. Thus, the same physical mechanisms should qualitatively apply to both subaqueous and incompressible subaerial density currents,

Table 1 Instabilities considered and eliminated for Volcán Bárcena

Instability	Reason eliminated
Lobe and cleft	Operates only at flow head; see text
Görtler	Curvature too small on the affected slopes; see text
Cross-flow	No appropriate geometry that produces topography at an angle to the flow
Rayleigh-Bénard	No denser fluid above lighter fluid in our model
Kelvin-Helmholtz	No obvious shear layer between two fluids; Produces transverse, not longitudinal, structures
Taylor-Couette	No flow in a narrow gap with rotating inner wall; both concave and convex slopes are present as shown in Fig. 7b,c.
Granular flow (Rayleigh-Bénard or Taylor-Couette analog)	Applicable to dense granular flows, not to our assumed dilute density current; see text
Inflection point	No inflection point in the velocity profile within the boundary layer

although quantitative differences may exist regarding such features as growth rates and dominant wavelengths. These quantitative differences can be attributed primarily to the fact that subaqueous flows generally are closer to the Boussinesq regime assumed by Hall et al. (2008) than are pyroclastic density currents, especially if the flow is stratified (Valentine 1987). In this case, a concentrated basal flow layer may develop in the pyroclastic density current (e.g., Fig. 3 in Kelfoun 2017; Breard et al. 2016). However, we note that within the narrow velocity boundary layer where the instability mechanism of Hall et al. (2008) is active, local density differences may be sufficiently small so that the Boussinesq approximation can be expected to provide insight into the flow. Our model is intended to apply primarily to density currents in which all parts are dilute enough that particle-particle interactions do not dominate the momentum transfer (particle concentrations $< \sim 10^{-2}$ according to Lube et al. 2020).

During the erosive stage, particles are eroded and entrained into the flow rather than being deposited, and therefore the flow structure can be, and we propose is, different than during sedimentation. We emphasize that our assumptions concerning the structure and dynamics of the basal layer apply to the *erosive* stage where particle entrainment from the substrate into the PDC dominates, and not to the waning stage where particle settling from the PDC occurs.

The “groovy instability” mechanism that we propose as being responsible for the generation of the grooves involves a feedback loop as follows: Erosion and resuspension of deposits modify the effective density of the flow; i.e., fluid regions that have higher particle concentrations have a greater effective density. Hence, random spanwise variations in erosion result in spanwise density variations. Due to gravity, these spanwise density variations drive a secondary flow in the form of counter-rotating vortices. This secondary flow increases the shear stress in regions of high particle concentrations, which further enhances local erosion and resuspension, and thereby further amplifies the initial random spanwise variation. Due to the existence of the turbulent eddy cascade, initially the bed will be perturbed on many scales. The feedback mechanism described above is most efficient at a certain wavelength, but not necessarily the largest. This feedback leads to the preferred amplification of that wavelength, which sets the scale of the grooves.

Hall et al. (2008) applied a linear stability analysis to examine the mechanisms responsible for the formation of longitudinal grooves by multiphase flows. Using the three-dimensional Navier-Stokes equations in the Boussinesq approximation, their analysis accounted for the coupled interaction of the fluid and particle motion and the interaction of the density current with the erodible bed. The key features of this instability mechanism can be summarized as follows. A unidirectional base

flow has a velocity boundary layer of thickness δ_u . The magnitude of the streamwise velocity increases from zero at the sediment bed surface to a maximum value over a distance $O(\delta_u)$, while the shear stress decreases away from the sediment bed. Analogously, the sediment concentration gradient at the bed defines a concentration boundary layer of thickness δ_c . The velocity boundary layer thickness δ_u is a function of the Reynolds number Re ($Re = u_\infty \delta_u / \nu$ is an effective Reynolds number formed with the eddy viscosity ν , and u_∞ is a typical streamwise velocity). In contrast, δ_c is the thickness of the concentration boundary layer and is determined by the competition between the settling motion of the particles and their effective diffusivity, as expressed by the Péclet number, $Pe = u_\infty / w_s$, where w_s is the particle settling velocity. In such a unidirectional flow in its base state, the rate at which sediment is eroded due to shear stress is in equilibrium with the rate at which it settles out, so that the location of the sediment bed surface does not change with time. However, a small, wavy spanwise perturbation to the initially horizontal sediment bed surface will affect evolution of the coupled flow/sediment bed system. In the context of linear stability analysis, it is assumed that initially small random perturbations of all wavelengths are present in the flow. Linear stability theory then associates a growth rate with each of these wavelengths, and only those wavelengths with a positive growth rate will be amplified by the linear stability mechanism. Generally, the wavelength with the largest growth rate can be expected to dominate the flow and to generate the observed regular flow pattern (Fig. 10). Such instabilities, and the resulting coherent streamwise vortices, can appear in both laminar and turbulent flows.

Figure 10a shows an initial wavy perturbation of the surface of the sediment bed, with peaks and valleys. The semicircular open contours near the bottom in Fig. 10b and c show excess sediment concentration (solid lines) and sediment concentration deficits (dashed lines) relative to the initial base flow. These contours indicate that above the peaks, the sediment concentration is reduced so that the local density decreases and the fluid rises, whereas the sediment concentration grows above the valleys so that the local density increases and the fluid sinks. In Fig. 10b, the gray shading depicts the shear stress, with lighter (darker) areas indicating positive (negative) values. In Fig. 10c, the gray shading depicts the perturbation of the streamwise (out-of-plane) velocity, with lighter (darker) areas indicating positive (negative) values.

The downward fluid motion above the valleys brings fast fluid from far above the bed closer to the sediment bed, so that the shear stress increases locally, which leads to further erosion of the valleys. Conversely, the upward fluid motion above the peaks thickens the velocity boundary layer and locally reduces the shear, so that erosion at the peaks is further weakened. In this way, the initial wavy perturbation of the sediment bed is amplified. As a result of these conditions,

Fig. 10 Physical instability mechanism that creates counter-rotating streamwise vortices in the bottom boundary layer of a turbidity current, as derived from linear stability analysis. The main flow direction of the current is out of the plane towards the reader, so that the figure illustrates various features of the flow in a cross-plane. Adapted from Hall et al. (2008). Reprinted with permission

counter-rotating vortices develop that are superimposed on the streamwise velocity boundary layer, as indicated by the closed lines with arrows.

Hence, the spanwise perturbation affects evolution of the coupled flow/sediment bed system in two ways. First, it places the peaks of the interface into regions of the current where the base flow profiles have lower shear stress (hence less erosion) and lower sediment concentration (hence less settling). If the velocity boundary layer is thinner than the concentration boundary layer, $L = \delta_u/\delta_c < 1$, and the reduction in shear stress and erosion outweighs the reduction in sedimentation at the peaks, so that the peaks keep growing. Similarly, at the valley locations, the increase in shear stress outweighs the increase in sedimentation, so that the valleys are further eroded.

Second, at the same time, the small, wavy spanwise perturbation sets up a secondary flow so that longitudinal counter-rotating vortices form. In the downflow regions, high-velocity, freestream fluid is brought closer to the sediment bed, which results in a further increase of the local bed shear stress (and hence the erosion rate) in the valley regions. Conversely, the upflow regions lead to a locally thicker velocity boundary layer above the peaks, so that the bed shear stress and erosion rate are further reduced there. In general, the effect described in the preceding paragraph dominates (Hall et al. 2008). An approximate criterion for the onset of instability is $L = \delta_u/\delta_c < 1$, that is, where the velocity gradient near the boundary is steeper than the concentration gradient.

Lube et al. (2020) present conceptual models for both concentrated and dilute PDC's including both velocity and concentration boundary layer profiles. Their profiles suggest that the condition, $L = \delta_u/\delta_c < 1$, for the onset of the groovy instability is characteristic of PDCs dominated by dilute transport behavior (right side panels of Fig. 1a in Lube et al. 2020). In this case, grooves would be expected in the substrate. The situation is more complex in their model for a concentrated PDC. In this case, a dense bottom layer that has nearly constant velocity and particle concentration is formed. No grooves would be expected on the substrate in this case. However, in the more dilute layer flowing along above this dense basal layer, the velocity gradient is again steeper than the concentration gradient, and hence, the groovy criterion is satisfied. If the basal layer is erodible and the dilute layer is sufficiently energetic to be erosive, then grooves might form on top of the basal layer.

For subaqueous flows, Hall et al. (2008) found the most amplified wavelength to be $O(25)$ times the concentration boundary layer thickness. Therefore, if the wavelength, λ , of grooves eroded by the vortices can be measured, the thickness of the concentration boundary layer can be estimated as $\sim \lambda/25$. However, we note that this factor of $O(25)$ is subject to some uncertainty in the present context, since factors such as particle/particle collisions, non-Boussinesq effects, or non-Newtonian rheology are not modeled in any detail.

A groove-ridge pair hosts two vortices with opposite rotations (Fig. 10a). At Volcán Bárcena, the grooves were 3–6 m wide, and ridges between them were slightly larger (Richards 1956). We take a typical combined groove and ridge width to be ~ 10 m. The inferred concentration boundary layer thickness is then ~ 0.4 m, i.e., less than $O(1)$ m.

The question then arises as to how this conclusion regarding boundary layer thickness can be tested in the absence of direct observations of an eroding flow? We speculate that observations and interpretations of the basal sediments that overlay an erosion surface may provide some information. For example, Lé and Pitman (2009) proposed that when granular flows are erosive, a mixing layer forms between the flowing material and the basal surface where erosion occurs. We envision a similar layer in an eroding particle-laden gas and propose that such a mixing layer could be equivalent to a collapsed version of our particle concentration boundary layer. We speculate that vortices arising from the groovy instability would reside in this mixing layer. The basal deposits containing eroded material represent the *collapsed* concentration boundary layer and therefore give a *minimum* size for the vortices. We also note that basal deposits are derived from upstream sources and that eroded material is either deposited downstream or lofted into the flow and that a mixing layer includes material from different places and times during propagation of the PDC.

Unfortunately, such data are not available for Volcán Bárcena. At Mount St. Helens, Kieffer and Sturtevant (1988) found that grooves eroded by the 1980 lateral blast ~ 8 km from the volcano had an average transverse wavelength of 7 m. Moore and Sisson (1981) found a coarse basal unit containing abundant eroded material at distances corresponding to the grooves. This unit was 0.25–0.75 m thick, a dimension in agreement with the conclusion for our analysis of Volcán Bárcena that the mixing layer is thin. At other sites at Mount St. Helens, Fisher et al. (1987) found a similar layer, A0, and Pollock et al. (2019) found bed material worked into undulose and recumbent flame structures. These ranged from 0.04 to 1.8 m in height and were interpreted as the record of instabilities similar to Kelvin-Helmholtz shear instabilities. The data suggest a mixing layer $O(\sim 1\text{m})$ thick. Dellino et al. (2008) described a “basal coarse layer” from the subplinian Pollena (472 AD) eruption of Vesuvius. From their Fig. 2, it appears that the basal coarse layer is <0.1 m thick. These comparisons are encouraging, and future detailed research documenting erosion surfaces and their overlying mixing layers, both vertically and in the downstream flow direction, will hopefully elucidate the relation between the erosion surface and overlying sediments in detail.

Groove termination at Volcán Bárcena

Grooves on the upper half of the cone of Volcán Bárcena have consistent characteristics on all flanks, but the structures on the lower half of the cone vary. On the east flank, the declivity is uniform from the crater rim down to the top of the sea cliffs. The grooves extend “slightly more than half-way to the base of the cone” (Richards 1956). We suggest that in this location, the vortices dissipated due to increasing turbulence on the lower slopes of the volcano, a process that has been demonstrated to occur for counter-rotating streamwise Görtler vortices in flow along plates in the laboratory (Fig. 11a). In some places, not well documented, Richards (1956, Table 4) reported that “groove debris may occur at lower ends when the change in declivity is moderate.”

On the west flank of Volcán Bárcena, the grooves are either buried or replaced by a new unit of material at an upstream-facing scarp at an elevation of $\sim 150\text{m}$ (Fig. 12). The scarp diminishes in height from south to north (right to left in Fig. 12a). The new unit displays periodic features with a wavelength of ~ 10 m (Fig. 12b). Moore (1967, p. 357) interpreted these features to be dunes and suggested that they formed where the change in slope impeded the flows, resulting in the beginning of deposition.

To explain the sudden transition from erosion to deposition, we propose a shallow-water analogy. In such a model, the flow underwent a transition in this area from hydraulically supercritical (and erosional) to subcritical (and depositional) conditions through a hydraulic jump triggered by the change in declivity. Hydraulic jumps are well known to cause major changes of flow regime in rivers (e.g., Kieffer 1985; Gioia et al. 2011; Wilkinson et al. 2018). While they are lesser known to influence volcanic flows, they have been postulated for a number of volcanic settings: (1) for a small pyroclastic density current in the August 7, 1980, eruption at Mount St. Helens (Levine and Kieffer 1991); (2) for pyroclastic density currents in the 110,000 B.P. eruption at Laacher See, Germany (Freundt and Schmincke 1985); (3) for large ash flows resulting from Plinian eruptions (Woods and Bursik 1994; Bursik and Woods 1996); (4) for small pyroclastic density currents, such as that from the 1998–1999 Volcan de Colima eruption (Saucedo et al. 2002); (5) for lithic-breccia deposits in the 1982 eruption of El Chichon, Mexico (Macias et al. 1998); and (6) for dunes in deposits from the August 2006 eruption at Tungurahua Volcano, Ecuador (Douillet et al. 2013a, b). Further discussions of this topic are given in Branney and Kokelaar (2002) and Sulpizio et al. (2014).

We assume that the flow front velocity was 15.4 m/s, as estimated from Richards (1956) report of a flow front velocity of 30 knots or more for the December 12, 1952, event pictured in Fig. 2, and use our calculated thickness of $d=0.4$ m for the basal layer. This yields a Froude number, $Fr=v/(gd)^{0.5}=7.8$ (note: $d=0.4$ is not the thickness of the basal layer, but it is the

Fig. 11 (a) Laboratory smoke-wire visualization of Görtler vortex breakdown in flow of air along a plate. Flow is from top to bottom and grid spacing is 10 cm. Initial smoke filaments were ~1.5 cm apart. The vortices, manifested as straight streaks at the top of the image, develop oscillations around $x=90$ cm, leading quickly to the breakdown into turbulence. From Fig. 3 in Swearingen and Blackwelder (1987). Reproduced with permission. (b) The east flank of Volcán Bárcena showing features with an irregular texture that may reflect the breakdown of vortices on the lower flank of the volcano (arrowed). The diameter of the crater is 700–750 m, and the typical distance between grooves is 3–6 m at their proximal ends near the crater. Reproduced with permission

thickness associated with the concentration gradient at the wall. These two may be different.) The flow would thus be supercritical on the upper slopes of the volcano. Velocities decrease downslope of a hydraulic jump, and rapid

sedimentation occurs because of this deceleration. In addition, local turbulence increases when a flow goes through a hydraulic jump at a slope change because vorticity is generated by the slope curvature (Macias et al. 1998). At this point, entrainment

Fig. 12 (a) Transition from grooves to dunes on the western flank of Volcán Bárcena. The grooves are heavily modified by drainage channels in this view. The dunes begin downstream of the dark cloud shadow. (b) Detail of the groove-dune interface showing the dunes and grooves with superimposed drainage channels due to later fluvial erosion.

L and S indicate lee and stoss sides of the dunes. The arrow indicating flow direction points to the scarp separating the grooves and dunes (dashed line). Wavelength of the dunes is ~10 m. Both images from Richards (1956) reproduced with permission

of air might also reduce the density of the flow, thus contributing to increased sedimentation of entrained particles downslope.

The shallowing of slope creates a geometry where the Görtler instability (Table 1) also becomes possible. This instability is characterized by the Görtler number, $G \sim 43(\theta/R)^{1/2}$, where θ is a boundary layer thickness and R is a radius of curvature (Tani 1962; note that different analyses give slightly different coefficients than the 43 used in this approximation, but they do not change our results significantly). We fitted a radius of curvature to a number of cross-sections and found $G=0.7\text{--}1.5$. Any Görtler effect would thus be weak and likely to reinforce the groovy instability proposed herein and this could explain, for example, why the grooves are so long—the combined groovy and Görtler instabilities would be a robust mechanism.

Discussion

In our analysis, we have used the Boussinesq approximation that neglects density differences, except in the buoyancy term in the momentum equation. We note that the Boussinesq approximation generally assumes density differences of $O(10\%)$ or less. In the current context of pyroclastic density currents, global density differences can certainly far exceed that range. However, within the narrow velocity boundary layer where the instability mechanism of Hall et al. (2008) is active, local density differences may be of an order such that the Boussinesq approximation can be expected to provide insight into the flow. Referring, for example, to the corresponding turbidity current flow investigation by Sequeiros et al. (2012), their Fig. 4 suggests that within the narrow velocity boundary layer, the concentration varies by no more than $O(10\%)$. Nevertheless, in order to minimize non-Boussinesq effects, we suggest that the application to pyroclastic density currents is best for the surge-like end member of “dilute pyroclastic density currents” in which particle loading is light (concentrations $<10^{-2}$ as defined by Lube et al. 2020) and the main particle transport mechanism is turbulent suspension (e.g., Burgisser and Bergantz 2002; Dellino et al. 2008; Breard and Lube 2017). Qualitatively, our results probably apply even to flows with up to $\sim 10^{-1}$ loading in which fluid-mediated particle collisions occur, but do not dominate the transport of momentum (Fig. 3d, e in Lube et al. 2020).

Increased particle concentration at the base of the flow enhances the stratification. Pyroclastic density currents, even small ones such as that pictured in Fig. 2b, typically have heights of 10's–100 m and thus are 1–2 orders of magnitude thicker than the concentration boundary layer thickness. For turbidity currents, the approximation of δ_c equaling the current height is reasonable because density variations due to particle loading can be modest. In contrast, particle settling through

the gaseous component in concentrated pyroclastic density currents plays a major role, perhaps producing a highly stratified flow, and thus violating the Boussinesq approximation that we employ. Kelfoun (2017) suggests a two-layer model for such flows that consists of a concentrated basal current and an overlying dilute current. We suggest that in the two-layer model of Kelfoun (2017), the overlying dilute current can be ignored and the only part of the flow relevant to the assumed Boussinesq approximation is the concentrated basal layer. Our model is restricted to cases where even this “concentrated” basal layer is relatively dilute, i.e., particle concentrations $<10^{-1}$ as discussed above.

We have also assumed that the flow is monodisperse, but it is evident that pyroclastic density and turbidity currents contain particles of many sizes and that they are thus inevitably stratified. We suggest that a representative “current height” is the height of the layer of the dominant particles contributing to the erosion. Smaller particles will be in suspension over a much larger flow thickness, giving a visual impression of a current much thicker than the boundary layer (Fig. 2b). If, as we propose, sedimentary deposits overlying an erosion surface freeze-in the characteristics of the (upstream) mixing layer, then these deposits provide information about the particle sizes in the erosive flow.

Additionally, our analysis employed the Reynolds and Péclet numbers. Although the turbulent viscosity and density could be quantitatively different in air-particle and liquid-particle currents, Hall et al. (2008) found that the dominant wavelength depends only weakly on the Reynolds and Péclet numbers, and thus we ignore them to first order.

We have also assumed that the Volcán Bárcena density current was incompressible and can be characterized by a Froude number. However, gas-particle mixtures can exhibit features of both incompressible and compressible gas flow simultaneously (e.g., snow avalanches). Two types of abrupt changes in flow conditions can occur in compressible density currents (Eglit et al. 2007): those that result in a change in the height of the flow, as in the end member characterized by a Froude number, and those that compress the flow and change the flow density, like the end member of a compressible gas characterized by a Mach number. The velocity of sound in particle-laden gaseous density currents can be much less than 100 m/s, even as low as 1–10 m/s (Kieffer 1981), and it is therefore possible that some density currents are even internally supersonic. However, such flows possess a free upper surface and a hydrostatic pressure distribution and therefore do not support gas dynamic shocks even if the flow velocity is greater than the sound speed of the mixture (Timmermans et al. 2001). The behavior of pyroclastic density currents may thus not simplify to either end member. A well-studied mixture with such characteristics is the powder snow avalanche [e.g., Ancy 2004; Hakonardottir 2004; Eglit et al. 2007]. If the flowing fluid becomes sufficiently laden with

particles, it is possible that low sound speeds could lead to important compressibility effects, especially in the more heavily particle-laden boundary layer. We have assumed that, to first order, the effects of compressibility on vortex behavior are small, as has been suggested for Görtler vortices (Wadey 1991).

Conclusions and recommendations

As discussed earlier in this work, Richards (1956) listed seven characteristics of the grooves present at Volcán Bárcena that needed to be addressed by any hypothesis for their origin. The model proposed herein for erosion of the grooves by vortices addresses these characteristics as follows: First, the *U-shape* of the grooves reflects the erosional downcutting of the substrate by cylindrical vortices. Second, the *herringbone (chevron) pattern* is consistent with the particle paths of fluid in pairs of vortices, as illustrated in Fig. 9. Third, the observation that the *grooves do not coalesce* is consistent with the fact that the vortices diverge in a downslope direction as they travel down the slopes of the approximately conical volcano. Fourth, the *straightness* of the grooves reflects the fact that the driving force for the flow is the downslope component of gravity and that lateral driving forces are not significant. Fifth, their *origin below the crater rim* reflects the fact that it is only here that the boundary layer started to develop. Sixth, the *increasing width of the grooves downslope* arises both because of nonlinear growth of the instability and breakdown into turbulence. Seventh, *local deposition at the lower ends of the grooves* arises either from loss of transport capacity as turbulence develops (east flank) or as energy is dissipated in a hydraulic jump (west flank), or both.

There are insufficient data to assign specific groove-forming instabilities for the examples at Lascar, Ruiz, Soufrière Hills, and Crater Lake at this time. We suggest that, amongst the processes listed in Table 1, three should be considered in future investigations of sets of streamwise grooves in the erosion surfaces of pyroclastic density currents: the Görtler, the crossflow, and the groovy instabilities. In determining which, if any, instabilities may have occurred, our findings point to several questions that must be addressed in field observations: (1) “Was there curvature in the flow either from the geometry of the landscape or within the evolving flow field itself, so that the Görtler instability might have been important?”; (2) “Is there a geometry that is conducive to the crossflow instability, such as a ridge oriented obliquely to the flow?”; and (3) “In the absence of these effects, is there any evidence either on the erosion surface or in the overlying sediments that could point toward the groovy instability?”

Our findings point to some profitable areas for future research:

- (1) *Research on non-Boussinesq particle-driven density currents applied to pyroclastic density currents*, similar to that of Ancey (2004) for snow avalanches, to test the possible relation of this mechanism to vortices in granular flows.
- (2) *Studies on the formation and nature of vortices in compressible multiphase flows* to clarify role of particles in the vortices and the similarity, or difference, in vortex behavior between incompressible and compressible flows.
- (3) *Documentation of the detailed three-dimensional morphology of erosion surfaces* must be conducted as soon as possible after eruptions if an erosion surface is left exposed, because normal weathering and geomorphic processes will quickly alter the surface by degrading high points and infilling low points. Techniques such as airborne LIDAR (Cashman et al. 2013) and drone-based structure-from-motion (SFM) photogrammetry (Westoby et al. 2012; Carrivick et al. 2016) now make this feasible.
- (4) *Computational and laboratory/field-scale models* that (a) model compressible gravity flows, e.g., such as those of Timmermans et al. (2001) and Eglit et al. (2007), and (b) focus on erosive flow-substrate interactions, rather than solely on sedimentation-substrate interactions. Such models should yield velocity and concentration profiles under various conditions, such as those provided by Pittaluga and Imran (2014) for turbidity currents. Laboratory simulations, such as those of Brosch and Lube (2020), would provide valuable data if performed with high resolution of the velocity and concentration profiles near the wall and with erosive boundary conditions.
- (5) *A comparison of flows that entrain substrate material and that occur in a number of different geological settings*, such as debris flows, rock avalanches, lahars, pyroclastic density currents, snow avalanches, and dam break floods (Iverson and Ouyang 2015). A key question in understanding and modeling such flows is how the flow and substrate respond to the downward propagation of an erosive surface (“front”) as material is removed from the substrate and incorporated into the flow. Laboratory simulations can qualitatively elucidate some processes but scaling the boundary layer of sediment-laden flows to laboratory conditions is, in general, problematic. Some of the newer large-scale field simulations may be of immense help if the experiments can be made energetic, or directed, enough to cause erosion by outflowing ejecta.

Most current models of substrate entrainment processes use depth-integrated techniques in which turbulence and variations in shear are averaged to yield a single velocity in the downstream direction (e.g., Kelfoun 2017; de’ Michieli

Vitturi et al. 2019; Shimizu et al. 2019). In such models, the evolving boundary between the flow and the static bed is idealized as a sharp interface. In fact, as pointed out by Iverson and Ouyang (2015), the boundary—even in idealized two-dimensional flow—is indistinct because grains shift around before they become fully entrained. In addition to this grain-scale blurring of the erosive surface, we propose here that the counter-rotating vortices discussed above move material from the substrate to the main flow at a scale much greater than individual grain diameters.

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Code availability Not applicable

Author contribution SWK, EM, and JB were involved in all aspects of the work. JA was involved in the supercritical flow/hydraulic jump theory.

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Declarations

Competing interests The authors declare no competing interests.

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