Pyroclastic dune bedforms: macroscale structures and lateral variations. Examples from the 2006 pyroclastic currents at Tungurahua (Ecuador)

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ABSTRACT

Pyroclastic currents are catastrophic flows of gas and particles triggered by explosive volcanic eruptions. For much of their dynamics, they behave as particulate density currents and share similarities with turbidity currents. Pyroclastic currents occasionally deposit dune bedforms with peculiar lamination patterns, from what is thought to represent the dilute low concentration and fluid-turbulence supported end member of the pyroclastic currents. This article presents a high resolution dataset of sediment plates (lacquer peels) with several closely spaced lateral profiles representing sections through single pyroclastic bedforms from the August 2006 eruption of Tungurahua (Ecuador). Most of the sedimentary features contain backset bedding and preferential stoss-face deposition. From the ripple scale (a few centimetres) to the largest dune bedform scale (several metres in length), similar patterns of erosive-based backset beds are evidenced. Recurrent trains of subvertical truncations on the stoss side of structures reshape and steepen the bedforms. In contrast, sporadic coarse-grained lenses and lensoidal layers flatten bedforms by filling troughs. The coarsest (clasts up to 10 cm), least sorted and massive structures still exhibit lineation patterns that follow the general backset bedding trend. The stratal architecture exhibits strong lateral variations within tens of centimetres, with very local truncations both in flow-perpendicular and flow-parallel directions. This study infers that the sedimentary patterns of bedforms result from four formation mechanisms: (i) differential draping; (ii) slope-influenced saltation; (iii) truncative bursts; and (iv) granular-based events. Whereas most of the literature makes a straightforward link between backset bedding and Froude-supercritical flows, this interpretation is reconsidered here. Indeed, features that would be diagnostic of subcritical dunes, antidunes and ‘chute and pools’ can be found on the same horizon and in a single bedform, only laterally separated by short distances (tens of centimetres). These data stress the influence of the pulsating and highly turbulent nature of the currents and the possible
role of coherent flow structures such as Görtler vortices. Backset bedding is interpreted here as a consequence of a very high sedimentation environment of weak and waning currents that interact with the pre-existing morphology. Quantification of near-bed flow velocities is made via comparison with wind tunnel experiments. It is estimated that shear velocities of ca 0.30 m.s$^{-1}$ (equivalent to pure wind velocity of 6 to 8 m.s$^{-1}$ at 10 cm above the bed) could emplace the constructive bedsets, whereas the truncative phases would result from bursts with impacting wind velocities of at least 30 to 40 m.s$^{-1}$.

**Keywords** Backset lamination, dune bedforms, pyroclastic currents, stoss-aggradation, Tungurahua.

**INTRODUCTION**

**Turbulent pyroclastic currents**

Pyroclastic currents are ground-hugging gas–pyroclast mixtures whose flow is generally driven by gravity and triggered by volcanic eruptions (Sulpizio et al., 2014; Dufek, 2016; Palladino, 2017). A range of flow processes, ranging from concentrated granular flows dominated by particle–particle interactions, to fluidized mixtures and up to fully turbulent, fluid-supported currents is interpreted from the study of deposits (e.g. Fisher, 1990; Cole & Scarpati, 1993; Branney & Kokelaar, 2002; Douillet et al., 2013a; Sulpizio et al., 2014). The flow-bed boundary layer is the coupling element between the flow structure and the bed morphology, being influenced by and influencing both the flow and the bed (Branney & Kokelaar, 1992; Branney & Kokelaar, 2002; Dellino et al., 2004a; Douillet et al., 2013a; Breard et al., 2016). The processes acting at this flow-bed boundary layer have a crucial influence on the grain-size distribution, the texture or fabric and depositional patterns of the sediment (Branney & Kokelaar, 2002). Deposits containing fine-scale laminasets forming dune bedforms have been interpreted as resulting from the low-concentration, turbulence-dominated, end member of pyroclastic currents (e.g. Sparks & Walker, 1973; Cole, 1991; Druitt, 1992; Dellino et al., 2004b – commented by Le Roux, 2005 – Douillet et al., 2013b). Such ‘dilute pyroclastic currents’ may result from decreased density resulting from thermal expansion of entrained air (Andrews, 2014), flow stripping at cliffs (Douillet et al., 2013a), or may be related to the initial eruptive dynamics, in particular for highly explosive maar volcanoes (e.g. Waters & Fisher, 1971; Jordan et al., 2013). Dilute pyroclastic currents are envisioned to behave as a particular type of particulate density current, i.e. a flow with gravity as the driving force, and where turbulently suspended particles, that ultimately sediment to form deposits, are the agent of excess density driving the flow (Simpson, 1982). As such, dilute pyroclastic currents are likely to be related to subaqueous turbidity currents, and a comparison of their dynamics and sedimentary signature is therefore highly relevant (Branney & Kokelaar, 1992; Kneller & Buc- kee, 2000). A smooth continuum between the granular and turbulent end-member flow behaviours is often modelled and expected in nature (Burgisser & Bergantz, 2002; Breard et al., 2016).

The stoss-aggrading nature of pyroclastic bedforms

Dune bedforms emplaced by pyroclastic currents are notorious for their stoss-aggrading nature producing a variety of backset laminations (e.g. Schmincke et al., 1973; Cole, 1991; Douillet et al., 2013b, and references therein). Early authors have suggested that these characteristics could link to the interpretation as Froude-supercritical bedforms (Fisher & Waters, 1969, 1970; Waters & Fisher, 1971; Crowe & Fisher, 1973; Mattson & Alvarez, 1973; Schmincke et al., 1973). Subsequently, this stoss-aggrading characteristic of the deposits has largely been taken as a straightforward argument for parental Froude-supercritical flows without further debate (Fisher, 1977; Yokoyama & Tokunaga, 1978; Wohletz & Sheridan, 1979; Walker et al., 1981; Fisher et al., 1983; Suthren, 1985; Charland & Lajoie, 1989; Sohn & Chough, 1989; Giannetti & Luongo, 1994; Brand & White, 2007; Gençalioğlu-Kuşcu et al., 2007; Brand & Clarke, 2009, 2012; Brand et al., 2009; Kelfoun et al., 2018).
Pyroclastic dune bedforms have been questioned in the pyroclastic backset laminations and Froude-supercritical Vellinga rents (Ponce & Carmona, 2011). Recently, following a similar interpretation for turbidity currents (Douillet et al., 2009). The direct link between backset laminations and Froude-supercritical bedforms has been questioned in the pyroclastic context as well as for turbidites (e.g. Kubo & Nakajima, 2002; Douillet et al., 2013b) and alternative interpretations have been suggested. In particular, the differential draping of a fallout load in a weak current was suggested to explain the processes of ‘regressive climbing dunes’ for pyroclastic bedforms (Douillet et al., 2013b), following a similar interpretation for turbidity currents (Ponce & Carmona, 2011). Recently, Vellinga et al. (2018) showed that, for analogue and numerical experimental currents, preferential stoss-aggradation occurs downstream of a hydraulic jump, in the subcritical regions dominated by slow, vertically-expanding current.

The role of topography

The possibility of stoss-aggradation being forced by the inherited bed topography was put forward for turbidity currents (Nakajima & Satoh, 2001; Kubo & Nakajima, 2002). In the pyroclastic context, it was suggested that the shape of a bedform alone could force saltating particles to deposit preferentially on stoss-faces in a feedback effect (Douillet et al., 2014). Indeed, the saltation threshold (the minimum shear required from a current to put particles in saltation) was measured for pyroclastic particles at various bed-slopes and found to be 50% higher on stoss-faces versus lee-faces (Douillet et al., 2014). The role of bed-slope has been further emphasized in the type of pyroclastic bedform produced, with evidence that dune bedforms are more likely to develop when the flow climbs up a ridge (Breard et al., 2015), that stoss-deposition is more frequent when flows climbed up-ridge than flowing down (Druitt, 1992), and that giant regressive bedforms resulted from a dual bypass upstream and downstream from a depositional zone of backset beds (Brown & Branney, 2004). Pyroclastic bedforms continuing their growth over several pulses or flows (e.g. Walker, 1984; Cole, 1991; Sulpizio et al., 2007); the interaction of the subsequent flow with the pre-existing topography can have a dominant role in sedimentation. Apart from Druitt (1992), no studies have yet addressed lateral variations within single pyroclastic bedforms, or the retro-controls of a developing bedform on its subsequent growth.

Erosional structures from turbulent pyroclastic currents

Since the recognition of the primary nature of pyroclastic dune bedforms, most documented outcrops included steep truncations of their stoss faces, generally called ‘chute and pools’ and interpreted as representing the signature of hydraulic jumps (here called Froude-jumps, i.e. a jump of the Froude regime from supercritical to subcritical) in some basal layer of the flow (e.g. Schmincke et al., 1973; Gençalioğlu-Kuşçu et al., 2007; Brand et al., 2009). Recently, most characteristics of such stoss-side truncations, in particular their abrupt and steep termination, could be reproduced using experimental short-lived air bursts unrelated to a jump in flow regime (Douillet et al., 2017).

Under water, scours caused by wall jets have been the focus of extensive work in the engineering context (e.g. Balachandar & Reddy, 2013). Interestingly, Dumas et al. (2005) noted that experimental bedforms created under combined flows had a ‘boxy profile’ when stoss faces were steepened by intense local scouring.

Besides these truncated bedforms, erosional, longitudinal U-shaped channels formed during the flow of pyroclastic currents have intrigued volcanologists for decades. Richards (1959) documented large-scale, parallel furrows that covered the upper flanks of Barcena volcano (Mexico) briefly after the 1952 eruption. Those furrows terminated abruptly at change of slope, and large boulders were sometimes present at their upstream initiation. Similar features were reported by Kieffer & Sturtevant (1988) from the 1980 Mount St. Helens eruption (Washington, USA); they were present in zones of flow reattachment downstream of sheltered regions and were attributed to scouring by longitudinal vortices resulting from flow instabilities induced by topography. Fisher (1977) also documented U-shaped (and V-shaped) channels in stratified pyroclastic deposits, interpreted as erosion from...
‘base surges’, and suggested that a flow front would develop a: “cleft and lobe pattern”, with “lobes being individual turbulent cells that splay outward from the source”. Thus most interpretations involve coherent structures related to a flow’s turbulence.

This article documents the sedimentary information found in pyroclastic dune bedforms at the lamina scale and through a survey of closely-spaced lateral transects. It is shown here that single bedforms vary drastically in their aggrading nature, and three processes of stoss-deposition unrelated to Froude-supercritical flows are suggested, which enable estimations of near-bed velocities of the parental currents.

THE TUNGURAHUA SET OF SEDIMENT PLATES

Context of the deposits of the 2006 eruption of Tungurahua

The dataset presented here was created from the unconsolidated deposits of the pyroclastic currents triggered during the 17 August 2006 explosive eruption of Tungurahua volcano (Ecuador). These pyroclastic currents deposited two end member sedimentary facies as a consequence of their interaction with the ravines on the steep flanks of the volcano. The valley-bottom facies is dominated by coarse-grained, (up to metric size boulders), unsorted, massive layers (1 to 5 m thick; Kelfoun et al., 2009; Douillet et al., 2013a; Hall et al., 2013). The marginal facies occurs outside curves of valleys, on shoulders and overbanks, and forms individual patches of a few hundreds of metres in extent consisting of fields of metre-size dune bedforms (Douillet et al., 2013b). This marginal facies is dominantly composed of ash organized in cross-laminated bedsets. The marginal facies patches were interpreted as resulting from dilute and turbulent currents that were created locally, where the main flow bodies would pass a cliff and incorporate large amounts of air (Douillet et al., 2013a).

These dune bedform fields represent unique deposits of this type worldwide because they can be investigated both in terms of surficial shape, extent and internal patterns. A downstream decrease in their dimensions (Douillet et al., 2013a) and evolution of outer shape has been recognized (Douillet et al., 2013b). Four different shape types were defined (Fig. 1), and a spatial transition was documented from ‘elongate’ (in proximal zones), ‘transverse’ (at the onset of deposition zones), sinusoidal and ‘lunate’ (intermediate in individual deposition zones), to ‘two-dimensional’ (2D) bedforms (in the most distal zones of spreading). This organization is very similar to that observed in the behaviour of experimental turbidity currents (Spinewine et al., 2009).

The sediment plate dataset

The dataset consists of ca 50 m² of outcrops from the 2006 deposits. The outcrop sections were hardened in their primary arrangement by impregnating the surface with epoxy, creating
sediment plates, a type of thick lacquer peels (method described in Douillet et al., 2018). The difference in resin uptake according to capillary forces (deriving from the grain-size distribution) for each lamina reveals sedimentary structures in the resulting plate, so that stratification is enhanced and underlined to a level of detail that would not otherwise be accessible.

Sets of plates were created for each of the four types of dune bedform previously recognized in Douillet et al. (2013b): ‘elongate’, ‘transverse’, ‘lunate’ and ‘2D’ (Fig. 1; Table 1). For each bedform, a transect consisted of a 3 m long outcrop cut perpendicular to the crest, from the superficial stoss-base to the lee, if possible over the entire thickness of the 2006 deposits (1-0 to 2-2 m).

For the transverse and lunate bedforms, several sets parallel to one another were impregnated, in order to document lateral variations within a single bedform (Figs 2, 3 and 4). Further, plates oriented perpendicular to the inferred flow direction were created to infer lateral continuities (Figs 12 and 14). Finally, one 6 m long set was created from the stoss side of the transverse bedform to the crest of the next one, in order to image the connection between

Table 1. Summary of all bedforms investigated.

| Plate name | Location (valley) | Latitude | Longitude | Profiles | Type* | GPR† |
|------------|-------------------|----------|-----------|----------|-------|------|
| Lunate     | Achupashal        | 1°25'59"N 64"W | 78°29'28"W | 3 x 6    | Ash strat | No   |
| Transverse | Achupashal        | 1°25'59"N 19"W | 78°29'28"W | 3 x 6 + 1 x 12 | Ash strat and block isol | Yes |
| Elongate   | Achupashal        | 1°26'36"N 96"W | 78°28'22"W | 1 x 6    | Ash and lapilli strat | Yes |
| 2D-1       | Achupashal        | 1°26'03"N 84"W | 78°29'48"W | 1 x 6    | Ash strat and lapilli mas | Yes |
| 2D-2       | Achupashal        | 1°26'03"N 48"W | 78°29'48"W | 1 x 4    | Ash strat and lapilli mas | Yes |
| Chontal    | Juive Grande      | 1°25'49"N 08"W | 78°27'14"W | 1 x 6    | Lapilli mas and strat | Yes |

*strat, stratified; mas, massive; isol, isolated.
†A ground penetrating radar (GPR) survey was carried on the bedforms prior to their dissection and is the focus of a forthcoming manuscript – see Dujardin (2014) for preliminary results.

Fig. 2. Surface shape (A) respective to (B), and the trenches realized to impregnate the transects (C) respective to (D) for the of the transverse respective to lunate bedform.

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Fig. 3. (A) and (B) Transects within the transverse bedform. All transects are formed by six individual plates 50 cm broad, forming a 3 m long profile. Transects T1a and T1b connect to form a 6 m long profile with flow direction from T1a towards T1b.
Fig. 3. (continued).
Fig. 4. Transects within the lunate bedform. All transects are formed by six individual plates 50 cm broad, forming a 3 m long profile.
successive structures (Fig. 3, sets T1a to T1b). One set of plates was made for an ‘elongate’ bedform (Fig. 15), a ‘lunate’ bedform in the Chontal area (later on ‘Chontal’ transect, Fig. 16), and two sets in two successive ‘2D’ bedforms (Fig. 17). A total of twelve sets of six plates, each 50 cm wide, were collected and exploitable after transport (Table 1).

SEDIMENTARY STRUCTURES AND FACIES

The dune bedforms are organized in laminasets and bedsets with coherent patterns separated by sharp unconformities (Figs 3 and 4). The general stratigraphic architecture is characterized by partially preserved lee-side (downstream dipping) bedsets, truncated by long erosive planes. These truncations are either sub-horizontal or very steep (see below) and subsequently covered by sets of backset laminae (upstream dipping). No conventional ripple beds were observed. Single bedforms are composed of several facies types and unconformities. Whereas the transverse and lunate bedforms are largely dominated by well-laminated sets of ash-size clasts (<2 mm), two ‘2D’ bedforms and a second lunate (‘Chontal’) bedform were found to have an almost massive fabric, consisting of an unsorted mixture of ash (0 to 2 mm diameter), up to blocks (>5-6 cm diameter). The ‘elongate’ bedform investigated here is transitional between these coarse-grained and fine-grained end members. Hereafter, the main stratification facies are described based on the observations of the ash dominated bedforms (transverse and lunate), and lateral variations are described between parallel transects in single bedforms (transverse and lunate). The coarse-grained features (Chontal, 2D1 and 2D2) are inventoried and discussed below in the Coarse-grained bedforms section (Figs 15 to 17).

Stoss features

Steep truncation trains
The most striking and ubiquitous features are steep truncations that cut into the body of the bedforms from upstream (Fig. 5). These erosive contacts are subplanar over at least 1 m distance in their upstream part. They evolve through a sudden break in angle to steep truncations between 35° and 90° to the horizontal (Fig. 5A and B). The downstream termination of these truncations reaches the limit of the palaeo-surface where the crest smoothens and vanishes into apparently concordant lee-side laminae. In several cases, the continuation of truncations evolves on the lee side into horizons containing shear structures which formed contemporaneously with the erosive cut (see Crest features section). Interestingly, these structures repeat in nearby patches (Figs 5A and 7B). Subsequent truncations generally cut through the filling of the former one (stacking upstream from one another). Truncations within a patch have relatively similar dimensions but tend to become steeper in successive occurrences.

Steep backset lineations
The holes that formed in response to the steep truncation events are always subsequently filled with a ‘steep backset lineation’ facies (Fig. 5). The material forming the infill can: (i) have a similar grain-size distribution as the undercut laminae (Fig. 5A); (ii) be richer in coarser clasts; (iii) be depleted in fines; or (iv) be coarser as well as fines-depleted together. The texture of the filling ranges from massive fabrics with clast-alignment patterns, to ‘lined’ or laminated. The term lineation is used here to stress that some coherent sets of outlines are present, but they cannot be confidently interpreted as depositional ‘laminae’, because they may also correspond to secondary truncations within a massive deposit. All of these stratigraphic indicators (truncation line, fabrics, lineations and laminations) render a coherent infill plane that spans angles from 90° to <30°, adopting the morphology of the cutting, and smoothing it over deposition of successive beds.

The basal (upstream) termination of these pseudo-strata are downlaps that approach the base of truncations tangentially, whereas toplaps either organize as converging bundles, evolve into concordant pure-aggradational crests (see Pure aggradation crests section), or form splay and fade structures. The resulting sets have roughly the shape of very steep sigmoids. The steep backset lineations evolve progressively away (upstream) from the erosion lines into smoother lenses or become re-incised by a subsequent truncation.

Erosive-based backsets
The combination of steep truncations covered with steep backset lineations are always found together, and form what is defined as ‘erosive-based backsets’ (Figs 5A to C and 7B). Those correspond to what is often described or interpreted as so-called ‘chute and pools’, a term that
implies deposition dominated by a Froude-jump (a jump in flow Froude-regime, like a hydraulic jump) and thus intrinsically contains a non-proven interpretation (see e.g. Schmincke et al., 1973; Gençalioğlu-Kuşcu et al., 2007; Brand & Clarke, 2009).

Interestingly, the erosive-based backsets are visible at multiple scales with invariant morphology, and span over two orders of magnitude with ‘trough to crest’ heights ranging from <5 cm (Fig. 5C) to >1 m (Fig. 5A). On occasions, the structural scale is correlated with the grain-size distribution of the subsequent infill (for example, Fig. 5A versus Fig. 5C). The smallest examples are, however, exclusively found at the base of the 2006 deposits, in a bedset formed of light grey silt-size ash (Figs 5C and 10). Many small-scale erosive-based backsets are found as lee-side features, with a continuous transition to the backset ripple facies visible on lee sides (see Lee-side features section, Fig. 7A).

**Overturning truncations**

The preserved laminasets immediately below truncation planes are occasionally recumbent and overturned in the downflow direction on a thickness of about 1 cm and over lengths up to 15 cm (Fig. 5D). These features could be reproduced by analogue experiments using air jets creating short-lived bursts and are beyond the scope of this paper (see Douillet et al., 2017). Briefly, they are interpreted as the signature of strong basal bursts or swirls that impact the stoss-face of bedforms. This locally produces an erosive airflow that is so strong that it infiltrates within the bed and coherently disturbs and overturns the particle bed.

**Planar truncations**

Planar truncations are common and seem to be parallel to the local slope over a decametre scale. They occur either as the upstream continuity of the steep truncation events, or reach a

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**Fig. 5.** Stoss-side features. (A) to (C) Erosive-based backset trains at different scales: (A) bedform scale (Trans-T4P1 to T4P4); (B) vertical truncation with vertical and overhanging infill of backset lineations (Trans-T3P1); (C) small-scale erosive-based backsets in silt-sized ash; (D) truncation with overturned lamination (Luna-T3P2).
Fig. 6. Crest features. (A) Pure-aggradation crest bedset building on a stoss-erosive palaeo-crest and subsequently cut by planar truncation (Trans-T4P4). (B) Pure-aggradation crest with upstream preferential deposition in the terminal sedimentation phase of growth (Trans-T2P3 and T2P4). (C) Regressive (stoss-depositional) beds containing prograding laminasets. Note that laminasets vanish as soon as the palaeo-crest is reached (plate from a previously investigated transverse bedform presented in Douillet et al., 2015).

Fig. 7. Backset ripples. (A) Patches of backset ripples and erosive-based backset trains (Trans-T4P1). (B) Propagation of a backset ripple structure through the stratigraphy, with evolving behaviour from preferential stoss or leede-deposition (regressive and progressive, Trans-T3P4). The pink line follows the successive position of the crest.
palaeo-surface where they vanish as apparently concordant beds. Occasionally, some planar truncations cut through the whole length of a bedform (transverse plate T1a, elongate, 2D1, 2D2 and Chontal). They are, however, not to be interpreted as a global feature linked with one particular flow event at the eruption scale (Fig. 6A).

Crest features

Pure aggradation crests
In several occurrences, the knickpoint of palaeo-crests at a given stratigraphic level exhibits continuous lamination that is preserved from the stoss to the lee side (net deposition on both sides; Fig. 6A and B). Further, the crest point is found to be shifted laterally between successive laminae and, in most cases, in the upstream direction (as described for the ‘regressive climbing dunes’ in Douillet et al., 2013b). This shift of the crestlines does, however, not represent a migration of the whole structure (since the root and body of the bedforms are not translated themselves) but results from preferential deposition on stoss faces.

Prograding laminasets in regressive bedsets
On the stoss-side of bedforms, individual beds seemingly massive and stoss-aggrading are in fact compounds of thin, oblique, prograding laminasets of higher order – i.e. of finer scale – (Fig. 6C). The geometry of the containing beds is clearly stoss-aggrading (which has often been interpreted as indicating Froude-supercritical flow conditions) and an upstream shift of the crest is visible. However, the individual laminae inside a bed are prograding, i.e. with accretion in the downstream direction. Upon passing the crest of the bedform, those beds vanish on the lee side.

Lee-side features

Planar lee laminasets
Lee sides are largely dominated by steeply dipping, planar laminasets (from 10° to >25°). These show strong variations in terms of lamination intensity, ranging from massive thin layers, to diffuse sub-planar beds, and up to very crude and well-developed laminasets. These variable planar facies mainly consist of ash and the fabrics range from well-sorted to unsorted at the lamina scale, involving any types of doublets from coarse to medium ash, sometimes also including lapilli horizons or anecdotic outsized clasts (up to 10 cm diameter, see Coarse-rich trough fillings section, Fig. 9). When preserved, a continuation with stoss-aggrading beds of any kind is observed. Planar lee-side laminasets are often truncated, and testify that only a small part of the sedimentary history on stoss faces is preserved.

Backset ripples
Numerous ripple-sized structures (albeit not necessarily genetically, but with similar dimensions, ca 10 to 30 cm length) with clear stoss aggradation and preferential upstream deposition are found as part of the lee of bedforms, intercalated within the planar laminasets (Fig. 7). These structures seem to be stacked on one another (in duplex), or occur with two or three periodic repetitions within a synchronous bed (Fig. 7A). Such backset ripples occur in close proximity to the smaller erosive-based backsets, sometimes even

Fig. 8. Train of three shark fin structures interpreted as representing shear instabilities at the base of the flow.

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in train on the same horizon/isochronous surface (Fig. 7A). In a few instances, these backset ripples can locally evolve into a preferential downstream aggradation trend when aggrading in the stratigraphy, before returning to a regressive trend (Fig. 7B). Some of the backset ripples tend to be longer and flattened (ca 30 cm long and 5 cm thick) as well as amalgamating together, resembling small-scale hummocky cross-stratification.

Overturned shark fin structures
Many horizons seem to be at the limit between erosive and concordant and occur superimposed on the planar lamination trend. They include peculiar soft sediment deformation features which are overturned in the flow direction and with a shark fin shape (ca 1 cm thick and 4 cm long, Fig. 8). Such ‘shark fins’ are found to be preceded upstream by a ‘ploughed zone’ suggesting downstream migration. They were thus interpreted as shear horizons related to traction carpets on lee sides (Douillet et al., 2015). More than 200 shark fins were analyzed and are the focus of a forthcoming manuscript. The shark fins further occur in periodic trains on an isochrone surface, and a wave mechanism was inferred to explain their formation. A linear stability analysis showed that waves can develop at a shear interface without being linked with surface tension or a density driven restoring force. Shark fins often occur downstream of erosive palaeo-crests, where the erosive line vanishes into concordant beds.

Coarse-rich trough fillings
The troughs lying between the base of a lee and stoss of the following bedform often exhibit horizons containing lags of coarser material (Fig. 9). Three main types of basal coarse lenses were observed:

![Fig. 9. Coarse-grained lags and lenses. (A) Superficial lag formed of light grey pumice (Trans-T1bP3 to T1bP6). (B) Horizon with oversized clasts that vanish laterally into finer grained particles and eventually disappears (Trans-T1bP1 to T1bP4). (C) Relatively coarse and massive lens that forms on the stoss side of a palaeo-crest and vanishes on the lee (Trans-T3P3 to T3P5).](image-url)
1 Superficial lags (Fig. 9A): the whole surface of the 2006 bedform fields was covered by a pluri-centimetric layer of centimetre diameter, low density pyroclasts landed by fallout. A lens of similar clasts is systematically accumulated at the base of the stoss faces. Its thickness varies around 3 to 10 cm.

2 Outsized clast horizon (Fig. 9B): A horizon in otherwise ash dominated bedsets contains several largely outsized clasts up to >20 cm diameter. Such sporadic and anomalous horizons are laterally fining into diffuse beds of centimetre diameter clasts on their upstream continuation, and vanish completely further downstream.

3 Massive lenses and lensoidal layers (Fig. 9C): a massive and relatively coarse-rich (including clasts up to 5 cm) layer thickens to >10 cm on the stoss of a palaeo-bedform. It can be followed over to the crest, but vanishes within tens of centimetres downstream on the lee side. In the elongate bedform, the basal part of such a layer also exhibits a short dyke and soft sediment deformation that was interpreted as a brief intraflow injected during flow.

**ARCHITECTURE**

**Basal contact**

The base of the 2006 deposits is visible in the transverse and lunate sediment plates (Fig. 10). When exposed, the basal unit is a thick cross-laminated bed (ca 10 cm) of fine ash overlying coarse and weathered blocks. This zone contains numerous plant remnants, a few of them carbonized, and the lamination exhibits a variety of bedding patterns. Small (a few centimetres thick) backset ripples and erosive backsets are present. During digging out of the outcrops, non-carbonized plastic ropes and farming remnants were excavated. The basal unit grades abruptly (within a few centimetres) into massive to well-laminated, unsorted fine to coarse-grained ash. This latter content forms the majority of the 2006 deposits up to the surface for the transverse and lunate outcrops.

**Lateral variations**

Correlations between transects are not apparent, even when separated by <1 m within a single bedform. Overall, a dune bedform outcrop can in no case be taken as a 2D structure, and interpretations may vary drastically within tens of centimetres. An example is suggested on the transverse (Figs 11 and 12) and lunate (Figs 13 and 14) bedforms based on the main truncation features and subsequent sedimentary facies as well as cross-profiles.

**Transverse**

The transverse set consists of four flow-parallel transects (T1 to T4, Fig. 11) and one perpendicular cross-profile (CP, Fig. 12) linking the T3 and T4 plates in the crest area. The base of the 2006 eruption is visible in all transects except for T2.

**Transect T1** is 6 m long (T1a and T1b). Its upstream limit is aligned with the three parallel transects but it extends further down to the downstream following dune crest. Transect T1a is largely composed of downstream dipping beds lying over the basal contact of the 2006 deposits. These beds are, however, the theatre for numerous trains of backset ripples and erosive-based backsets. Some shark fin structures are found on relatively isochrone beds on the lee side.
lower part of T1b presents the character of an aggrading low angle bedform, which is subsequently truncated, with the final structure developing downstream from the palaeo-bedform. An outsized clast horizon is present in the trough between both bedforms in the lower third infill part of the 2006 deposits. The stoss side of both successive crests consists of a large-scale erosive backset zone terminating as unconformable topsets at the surface. These two stoss erosive backsets are not synchronous, and the genesis of the downstream bedform (T1b) pre-dated the final shaping of the upstream one (T1a). A long planar truncation that cuts through the upper part of the whole T1a bedform is lost into conformable beds over the lower lee side.

Transect T2 contrasts in its structural patterns with those of T1. Whereas the base is not exposed, the planar truncation can still be recognized. From T2 on to the following profiles, several series of steep truncations are visible, and may have been confounded as a single truncation in T1. The four stoss-truncations identified in T2 are steepening the bedform, and are followed by episodes of stoss deposition and crest aggradation.

Transect T3 marks the onset of subvertical truncation trains. It is the steepest part of the bedform, and up to seven trains of steep subvertical truncations followed by stoss aggrading beds are preserved. They can be correlated to some of the T2 truncation events through their stratigraphic position, the degree of truncation and the nature of the infill that contains faint lineations, some centimetric outsized clasts including light grey pumice lapilli clasts.

Transect T4 is located below the crest edge of the bedform. Five distinct packages of truncation trains are exposed. The main train of backset laminae is translating in the upstream direction over more than 1.5 m.

A cross-profile connecting T3 and T4 was made in the crest area (Fig. 12). It exhibits as much variability as the flow-parallel profiles. Truncation events are sometimes recognized, and seen to evolve with steep angles in the flow-perpendicular direction. They also cross-cut one another, which explains the complications upon the correlation of the individual profiles.

Lunate
The three lunate transects show more similarities with one another and the correlations are more confidently proposed (Figs 13 and 14). This bedform is located just 10 m upstream from the transverse one, so that the same colour-code is used, that may relate to the same flow events as for the transverse bedform.

The first major truncation event visible in transects T2 and TN lies ca 25 to 50 cm above the basal fine-grained unit. This truncation surface is planar to downstream dipping, and overlain by a coarse-rich horizon, a combination similar to the transects T1 and T2 from the transverse bedform. Two additional truncative surfaces occur 20 to 40 cm higher in the stratigraphy, and are planar downstream to slightly
upstream dipping. The stoss side of the lunate bedform is, as for the transverse one, cut by three steep truncation planes that can be correlated through the different profiles. The trough upstream from the stoss face is finally partially filled with a lens of light pyroclasts similar to the final fallout covering the surface. A distinct discordance on the upper part of the lee of transects T2 and TN-1 has been interpreted as a very short-throw slumping event that produced a horizon with convolute and disturbed soft sediment deformation features.

Three successive cross-profile plates were created between TN-1 and TN and confirm the correlations of truncation surfaces between transects (Fig. 14). Interestingly, the same richness of structures and amount of variability is visible in the cross-profiles thought to be perpendicular to the flow direction. In particular, numerous shark fin structures, backset ripples and small erosive-based backset beds are visible. Whether these small-scale structures are related to varying flow direction or to local slope-related transport processes on the outer tail of the lunate bedform is unclear.

**COARSE-GRAINED BEDFORMS**

**Elongate**

The elongate transect is transitional between the fine-grained bedforms and the courser-grained ones (Fig. 15). It contains ash to lapilli clasts,
organized in diffusely laminated bedsets to massive layers up to 15 cm thick. Overall the bedform has a low angle structure and little in common with the other ones in terms of stratal architecture. The base of the 2006 deposits could not be observed in nearby rain-gullies cutting several metres down into the surface. Sub-planar, downstream dipping laminasets of low angle cross-laminations cut into one another in the lower part. A low angle proto-bedform developed in the middle part of the plate, and consists of diffuse and low angle backset beds. This structure is then covered by several massive to faintly stratified layers with a grading contact. The latter layers are stacking up on the stoss of one another, resulting in a regressive lensoidal structure.

Chontal

Fig. 13. A possible correlation of the lunate transects. The colour coding is based on the same events as for the transverse bedform (these two structures are separated by ca 10 m in the field).

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The Chontal bedform has the outer shape of a relatively flat lunate bedform, yet its internal content is completely distinct from the lunate bedform described earlier (Fig. 16). It lies on the shoulder of two valleys that directed pyroclastic currents and come closer together before a sharp curve, so that the influence of two pyroclastic currents with diverging flow directions was hypothesized (see the ‘Chontal’ area in Douillet et al., 2013a). The base of the 2006 deposits could not be reached, although a height of >2 m was exposed. Three main sedimentary phases can be identified:

1 In the lower part, a unit of ash-dominated bedsets (>50 cm thick) containing small erosive-based backsets and planar truncations is exposed. Its content is relatively coarse, with centimetric clasts that are concentrated in the backsets’ trough. This lower part contrasts with the rest of the bedform, and its upper contact is sharp, although without clear sign of any truncation. This unit might be related to the July 2006 pyroclastic currents rather than those from August.

2 A very coarse and unsorted unit (ca 1-7 m thick) containing clasts from the ash size to blocks up to ca 10 cm diameter covers the lower ash bedsets. Whereas it appears as fairly massive at the outcrop, the corresponding sediment plate
exhibits several structures highlighted through an undefined but perceived lineation trend. On the stoss side, signs of large and very steep (70° to 90°) backset lineations are visible on the upper part, forming diffuse splay and fade structures; they evolve into less steep backset lines over a thickness of ca 1 m. On the lee side, planar lineations are perceived; they are superimposed with a clear fining-up gradient. This sedimentary facies is referred to as ‘coarse-lineated’ below (Four-fold formation mechanisms section).

3. The final unit (10 to 35 cm thick) consists of a bedset that becomes diffusely laminated, with a greater thickness on the stoss than on the lee side. Ash laminations are intercalated with more massive and fining-up layers.

**Two-dimensional bedforms**

The two 2D bedforms are located in an area on the counter-slope of the volcano flanks, past the base of the edifice. The parental flows must have crossed the valley of the Chambo River, and were interpreted as spreading against the local slope at the downstream limit of inundation of the pyroclastic currents. The base of the deposits was not encountered with the sediment plates, yet erosive rain-gullies testify to a local thickness >4 m.
The lower part of the transect is massive, unsorted, including blocks up to 15 cm diameter (Fig. 17). It is overlain by a diffusely laminated, finer grained bedset, ca 30 cm thick.

Both units are truncated on their stoss face by a layer with evidence for: (i) the occurrence of deposition on the stoss side; (ii) backset lineations visible in the ‘2D1’ bedform; and (iii) a locally very varying nature from massive to diffuse lamination. Finally, a low angle, second truncation of the stoss face is covered by another bedset of diffuse to massive layers consisting of ash to lapilli clasts.

**INTERPRETATION**

The numerous sedimentary features exposed in the dataset bring a range of questions. There are two ways to interpret their formation. In the ‘traditional’ scheme, the flow regime would dictate the sedimentation processes. Here, the occurrence of stoss-side deposition would thus be interpreted as indicating Froude-supercritical or transcritical antidunes and chute and pools. In this frame, relatively stable flows and the oscillation of a free upper surface would be needed.

However, this interpretation is not favoured here, since experimental and observational evidence about pyroclastic currents point towards pulsating and unsteady flows (e.g. Hoblitt, 1986; Brown et al., 2007; Andrews & Manga, 2012; Breard & Lube, 2017) where no stable conditions could develop. In this context, a discussion is opened to suggest revision of the generally accepted distinctive features to interpret sedimentary structures related to Froude-supercritical flows. It is shown that the bedforms’ structural patterns can be well explained by mechanisms driven by the basal topography and bed shapes that would interact with flow bursts and pulses in highly turbulent flows at very high sedimentation rates.

**Steepness and lateral variations: the invalidation of flow regime interpretations**

The transverse bedform drastically evolves in terms of stratal architecture over the four lateral profiles (Figs 3, 11 and 12). If put in a flow-regime interpretation, T1 would probably be interpreted as a subcritical dune and T2 as an antidune, whereas T3 and T4 would represent chute and pool structures. It has been suggested that antidunes (indicating Froude-supercritical flows) would occur in more proximal areas than chute and pools (indicating hydraulic jumps/ Froude jumps) and that distal deposits would in turn be dominated by subcritical bedforms (e.g. Schmincke et al., 1973). The data here...
invalidate this model, since the three types of stratal architectures coexist in the same structure and on the same horizon. It is considered very unlikely that three distinct flow regimes are recorded within few metres of one another, within a single bedform and at the same time. Indeed, it is not possible to explain how a flow could be radically distorted in a Froude-regime so that a single structure would coexist within three different parts of a bedforms’ phase diagram.

Additionally, the extreme steepness of the backset beds contrasts with the experimental evidence of bedding related to Froude-

**Fig. 16.** (A) Interpreted transect of the bedform from the Chontal area. (B) Zoom of the zone highlighted in (A) reveals the coarse-lineated facies and contact with lowermost unit.
supercritical flows, which are generally at low angle and cross-cutting one another (e.g. Alexander et al., 2001; Cartigny et al., 2014; Vellinga et al., 2018). As already highlighted from their surface expression, the geometrical relationships of the dimensions and steepness of the bedforms are not compatible with Froude-supercritical flows (Douillet et al., 2013b), an argument further exacerbated by the internal patterns evidenced here. Finally, lee-side erosion is virtually absent from the outcrops, an observation that excludes their interpretation as antidunes. Indeed, antidunes form as a mould of a gravity wave and contain backset bedding when the wave migrates upstream, a feature that should thus be associated with concurrent lee-side erosion due to this same upstream migration of the wave.

In this context, it is shown that all sedimentary features evidenced here can be interpreted without involving a flow in a Froude-supercritical regime (see next sections). It is suggested that the sedimentary architecture of the structures is dominated by the flow unsteadiness and turbulence (related to the Reynolds dimensionless number) rather than by the flow regime.
(defined through the Froude dimensionless number). In particular, backset bedding is related to highly depositional and slow currents with depletive and waning dynamics.

**Turbulent bursts**

*Scale invariant features and turbulent vortices*

The erosive-based backset structures show a striking scale-invariance of the patterns over two orders of magnitude (Figs 5 and 10). The largest documented examples in the literature (at Laacher See, Germany) reach a scale of several metres and are formed by very coarse-grained lapilli (pebble-size) pumice clasts with a similar stratal architecture (see fig. 8 in Douillet et al., 2018; Schmincke et al., 1973). If a dimensional scaling could simply be translated between the size of the structures to their parental flow process, then this would mean that the same flow processes occur at different scales. A scale-invariant flow process that could form steep truncations can be related to turbulent eddies. Indeed, the eddy downscale-cascading inherent to turbulent flows (see Kolmogorov scale/enstrophy cascade rate, e.g. Vallis, 2006) predicts such eddies from the flow scale to the centimetre scale.

Recent large-scale pyroclastic current experiments further showed that mesoscale turbulence clusters would form and concentrate particles outside turbulent eddies (Breard et al., 2016). It is suggested here that such turbulent clusters inherent to pyroclastic currents could impact on the stoss face of bedforms as erosive bursts, followed by highly depositional moments and form erosive-based backsets. This possibility is to be verified through Reynolds-scaled analogue or numerical modelling, as well as further field investigations on the genetic and timing relations between successive truncations.

*Narrow timescales and high sedimentation*

The sub-vertical truncations and subsequent sub-vertical backset beds as fillings suggest a highly and fast varying phenomenon (Fig. 5A). Their occurrence as subsequent repetitive trains in a single nest further suggests a pulsating behaviour of flushing away and deposition within a narrow time window. In the absence of cohesive forces, the organization of the filling as subvertical aggrading lamination inside a truncation nest is not likely to be preserved without extremely fast burial to avoid gravitational collapse. Thus, very high sedimentation rates with deposition of laminasets several tens of centimetres thick must have occurred within seconds within a pressing/plastering current. Such sharp transitions in sedimentary behaviour testify to the highly turbulent nature of the flows with sudden evolution between extremes of erosive and depositional behaviours.

Experimental turbidity currents also exhibit a pulsating behaviour with short time recurrences (Cartigny et al., 2013), whereas experimental turbulent pyroclastic currents form: “series of sedimentation–erosion couplets that propagate” across the flow bed interface (Andrews & Manga, 2012). Such erosion–sedimentation coupletst would well explain the subsequent backset incisions. Considering the formation of sedimentary steep backsets and backset ripples, the interpretation here is that their growth is the sole consequence of high sedimentation rates against basal topographic obstacles, and their position is dictated by the local topography. Indeed, if sedimentation is high enough that any bed irregularity would trigger the formation of backset beds as a flow reaction, a self-sustaining process can occur, as seems to be visible. Further, scale would not be involved in such a process and this would explain why the same patterns are obtained at several sizes.

**Coherent turbulent structures**

In three-dimensions, the lateral variations are found to be abruptly evolving, with continuity of the order of less than a metre and crest-perpendicular truncation angles up to \(>30^\circ\) (Fig. 12). This emphasizes that the transient phenomenon cutting bedforms are also very localized. In addition, the truncations that probably remobilized large amounts of pre-deposited material are also evolving within tens of centimetres in the downstream direction into concordant aggrading planes. This indicates locally waning conditions, and that processes of recycling and cannibalism of the sedimentary structures over short distances dominate the dynamics.

Most erosional furrows and sharp scouring of bedforms in the literature, regarded independently of a specific environment, have been perceived as resulting from coherent turbulent cells (e.g. Richards, 1959; Fisher, 1977; Kieffer & Sturtevant, 1988; Dumas et al., 2005). Numerical simulations of turbulent structures over ripple beds have shown that coherent structures identified as ‘Görtler vortices’ could form and be a main agent of sediment entrainment (e.g. Zedler & Street, 2001). If occurring at the metre scale in
pyroclastic currents, such Görtler vortices would be very appropriate flow-structures to explain the formation of steep truncation events. In subaqueous experiments, the scouring of stoss-faces described by Dumas et al. (2005) could be attributed to a similar effect. At high Reynolds numbers, numerical simulations succeed in reproducing low-speed and high-speed streaks that could impact the stoss faces of bedforms (e.g. Cantero et al., 2008). The authors speculate that such streaks or Görtler vortices could be the dominant mechanism that shapes the stratal architecture of erosive backset events, in a similar manner as for smaller scale bedforms and the scouring described by Dumas et al. (2005). Alternatively, pulsating Froude-jumps might produce similar features, yet natural observations of these processes are almost exclusively linked with lee-side erosion (e.g. Dietrich et al., 2016; Hage et al., 2018) and no argument allows a solution to this question.

The steep truncations and backset lineations are similar to what is generally interpreted as chute and pools (for example, types I, II and IX of Schmincke et al., 1973; and types d and e of Cole, 1991). The revised interpretation here may be applicable for all of those structures.

**Coarse bedforms: the transitional limit between granular and turbulent flows**

**Coarse lineated facies: locally turbulent flows**

The imaging power of the impregnation method enabled the identification of subtle lineation/lamination patterns even in the coarsest and most unsorted bedforms (for example, Chontal and 2D outcrops). The parental currents were likely to be dominated by particle–particle interactions during their transportational phase in order to support the coarse content of the deposits, a hypothesis reinforced by the bad sorting of this facies (Figs 16 and 17). However, the faint occurrence of backset and aggrading crest lineations points towards tractional transport, probably related to mechanisms of support locally driven by the fluid’s turbulence. These bedforms may thus represent deposits recording the hypothesized transitional flow behaviour between ‘concentrated’ (granular) and ‘dilute’ (turbulent) end members of pyroclastic currents (e.g. Burgisser & Bergantz, 2002). This strengthens a previous interpretation that the flows responsible for the marginal overbank deposits locally emanate from valley-confined granular flows when a portion of these became turbulent.
as a result of air entrainment upon passing upstream cliffs (Douillet et al., 2013a). The closest processes that might explain this coarse-lineated sedimentary facies are found in the dam break analogue experiments conducted by Rowley et al. (2014), where fluidized flows deposited backset, lineated beds. Leclair & Arnott (2005) showed that turbidity currents with up to 36% particle concentration in the bedload layer could deposit as laminated beds, and this ratio is suggested as an order of concentration here. The finer-grained nature of the lenses and lensoidal layers are considered to result from parent flows with dominant particle–particle support (granular flows or bedload rich) over a thickness at least equaling the size of the largest transported clasts rather than fluid turbulence support down to the flow-bed boundary (Fig. 9C). The lenses vanish outside the troughs formed by the upstream toe of stoss-faces, and so they flatten topography by filling the troughs; they can be understood as the signature of a simple damming triggered in the pools at the toe of bedforms. No particular flow conditions are required and this is understood as a pure topography-triggered jamming or frictional freezing, due to the parental ‘granular’ part of the flows tripping against the obstacle formed by bedforms.

Four-fold formation mechanisms

Altogether, the outcrops from Tungurahua enable construction of a depositional scheme for pyroclastic bedforms compound of four formalional bricks (Fig. 18).

Differential draping fallout

The purely aggrading crests resemble climbing structures (ripples or dunes), apart from their upstream preferential deposition (Figs 6A, 6B and 7). Climbing structures are generally interpreted as resulting from sedimentation with higher depositional rates than translational (e.g. Allen, 1971; Ashley et al., 1982). In proglacial settings, climbing-dune cross-stratifications are related to high rates of transfer of sands from suspension to the bed and net deposition on a bedform’s stoss sides (Ghienne et al., 2010), a result likely to be transferable here. As already suggested for turbidites (Ponce & Carmona, 2011) and pyroclastic currents (Douillet et al., 2013b), the interpretation here is that the stoss-depositional crests result from a process of differential draping, whereby fallout-dominated deposition is enhanced on stoss-faces, due to the simple combination of the bed topography and trajectory of particles (Fig. 18B). This requires the bedsets to be sedimented in a gentle current with shear velocities below the saltation threshold and high fallout input. The fallout load would originate from upper parts of the pyroclastic currents, because of spatial and/or temporal changes in sediment transport rate.

The median diameter (Md) previously measured for the Tungurahua 2006 pyroclastic bedforms is around Md = 2 Phi (Douillet et al., 2013a) and can be used with the corresponding shear velocity ($u^*$) at the saltation threshold
measured for a flat bed in a wind tunnel (Douillet et al., 2014). This would imply that purely aggrading crests were emplaced by weak currents with shear velocities below \( u_* = 0.29 \, \text{m.s}^{-1} \), corresponding to near-bed velocities below 6 m.s\(^{-1}\) at a height of 10 cm above the bed for a pure wind.

**Slope-influenced saltation**

The regressive laminasets found on stoss faces, which internally consist of prograding individual laminae, that vanish downstream a crest (Fig. 6C), are understood here as representing slightly higher shear velocities than for differential draping bedsets. For these beds, the saltation threshold is partially reached, so that particles are transported near the bed and produce prograding laminae. Whereas the saltation transport is sufficient to transport away all particles down-slope and thus no deposition occurs on a lee side, it has a net loss of carriage on stoss faces (Fig. 18C). This is supported by wind tunnel measurements of the saltation threshold for pyroclasts at various bed slopes, where it was measured that the threshold is reached at up to 50% more shear velocity on a \(+25^\circ\) slope than on a downstream dipping bed with a \(-25^\circ\) slope (Douillet et al., 2014). Following the same comparison to wind tunnel measurements as before for the prograding stoss-laminasets, the shear velocities needed, considering that the saltation threshold is reached on stoss-faces (\( u_* \sim 0.39 \, \text{m.s}^{-1} \)) and surpassed on lee faces (\( u_* > 0.27 \, \text{m.s}^{-1} \)), means that velocities of ca 8 m.s\(^{-1}\) (for a pure wind) would be needed at 10 cm above the bed, slightly above that for pure aggrading crests.

**Truncative bursts**

The steep truncations represent anomalies that disturb the previous weak-current sedimentary processes (Fig. 5). Short-lived, highly erosive basal pulses related to coherent turbulent structures at the flow-bed boundary are the best candidate to explain these features, and they represent the high-energy moments of the pyroclastic currents (Fig. 18D). These are followed directly by moments of very high deposition, yet lateral velocities must be present to ensure that subvertical lamination is plastered against those truncations. Those turbulent high velocity clusters must be advected in the downstream direction as well as close to the bed, in order to impact mainly on stoss faces, yet smaller ones seem to brush lee faces as well. In order to produce overturning at truncations, as observed in the deposits, burst jets with velocities ranging from 28 to 40 m.s\(^{-1}\) were needed in small-scale experiments (Douillet et al., 2017), which is taken as a lower range value for the real deposits of Tungurahua, largely above the values for aggrading phases.

**Granular-based events**

The coarse and massive lensoidal layers that are occasionally interbedded in bedform patterns represent the granular-flow part of the bedform-forming pyroclastic currents (Figs 9B, 9C and 18E). These sporadic events may be more common than their sedimentary signature, since the lenses often vanish as sedimentary bypasses, with no information on a granular-based passage. The flows were probably at least as thick as the coarsest deposited clasts (see *Coarse lags: topographic pools* section). Interestingly, experimental granular flows passing over a bedform experience a jump similar to a Froude-jump (granular jump) and could leave a similar sedimentary signature (Viroulet et al., 2017).

**Four mechanisms and no equilibrium conditions**

The whole variety of Tungurahua’s bedforms can be reconstructed with a combination of the four formation mechanisms and, more generally, can be applied to most pyroclastic bedform deposits elsewhere. Notably, such pyroclastic structures do not exhibit any kind of equilibrium bedform (for example, sustained progradation/translation). This is likely to be a result of rapidly changing conditions and the absence of any stable flow over intervals comparable to the timescales of growth of a sedimentary structure. These bedforms are a simple stack of the four depositional processes interacting with the pre-deposited structures. This disequilibrium is further supported by the systematic spatial stability of pyroclastic bedforms, at Tungurahua and elsewhere: Once a bed morphology is initiated, no migration is observed, over several metres thickness of deposition and through the variety of depositional mechanisms. Hence, no stable conditions were reached during the flow of the parent pyroclastic current.

**Flow energy**

When interpreting sediments, we probably never look at the moments of the flow where it has its
highest energy, but only at the ones where deposition occur, likely always during vanishing periods. As such, sediments are thus unlikely to reflect any high energy events. The most erosive and, indirectly, energetic events, reported here, are the truncative bursts. All stoss-aggrading features are interpreted here as low energy events. This is further supported by the fact that the final depositional bedsets almost always contain strong stoss-aggrading patterns. Thus stoss-aggradation belongs to the waning, rather than strongly stoss-aggrading patterns. This is further supported by the fact that the final depositional bedsets almost always contain strong stoss-aggrading patterns. Thus stoss-aggradation belongs to the waning, rather than being highly energetic periods of a pyroclastic current, when the flows vanish.

Even in an interpretation as ‘Froude-supercritical bedforms’, the sedimentary beds should not be related to high energy flows. Indeed, although many interpretations refer to Froude-supercritical flows as highly energetic, this is a confusing statement. The flow regime, defined as sub-critical or super-critical (or lower and upper flow regime, respectively), corresponds to a flow state where the Froude number (Fr) is Fr < 1 respectively Fr > 1. This ratio informs on the kinetic over potential energy of a flow, but in no way on the amount of total or kinetic energy of the flow. In nature, Froude-supercritical flows may in most cases represent waning, low energy conditions, rather than high energy events. Compared to a subcritical flow, supercritical conditions may well be attained through a decrease in flow depth rather than an increase of velocity, and would as such correspond to a decrease in energy. Thus, only the truncative events are considered here as possibly related to highly energetic events, if looked at over the temporal sedimentation phases of a flow.

CONCLUSION

The dataset presented here represents an extensive and fine scale investigation of pyroclastic bedforms and their lateral variations. The sedimentary architecture of the structures largely consists of stoss-aggrading features such as backset ripples, erosive-based backset bedset trains, draping crests with preferential stoss deposition, or lensoidal layers. These features are punctuated by important truncation events that attack the stoss-face of bedforms with angles up to the vertical.

The presented structures can be explained through the combination of four formational mechanisms, namely: ‘differential draping’, ‘stoss-influenced saltation’, ‘truncative bursts’, and ‘granular-based pulses’. The kinetic energy related to the constructive phases of the sedimentary history is likely to be very low and involves high rates of sedimentation and weak currents.

Coarse-grained, unsorted and apparently massive deposits contain similar truncations and backset beds as laminated bedforms, yet are less pronounced. These deposits represent the signature of the depositional portion of a current at the transition between turbulent and granular: although granular during transport, the sedimentary dynamics were turbulent enough to produce faint lineations and stratification. The parental flows are thus understood as granular currents that have locally become turbulent in the vicinity of their deposition zone and testify to the gradual continuum between dense pyroclastic flows (granular-based) and dilute pyroclastic currents (turbulent supported).

The location of a pyroclastic bedform is spatially very stable. This contrasts with the high variability in stratal architecture within a single dune. Within tens of centimetres, truncations can reach angles of >30° in the flow-perpendicular direction, and pass from sub-vertical truncations to concordant lamination in the flow-parallel direction. Altogether, the features evidence that no equilibrium is reached, and that these dune bedforms are transient structures. All features point towards very pulsating behaviour, and the dominant role of turbulence and local coherent turbulence structures in the vicinity of the bed (possibly Göttler vortices), interacting with the topographic expression of the previously deposited bedform.

Whereas pyroclastic bedforms have long been interpreted as antidunes and chute and pools related to Froude-supercritical flow regime, this interpretation is not favoured here. The data show that the features usually taken as diagnostic for Froude-supercritical bedforms are evolving drastically within a single structure. The interpretation thus puts emphasis on the combined role of pre-existing morphology and local scouring. Accent is put on the high turbulence and extreme sedimentation, hence related to high Reynolds numbers within unsteady, weak and waning currents. Quantitative estimates of nearby flow-velocity are made with comparison to wind tunnel experiments on saltating pyroclasts. These estimates suggest that a pure wind velocity of 6 to 8 m.s⁻¹ could emplace the constructive bedsets, whereas the truncative phases would result from bursts at 30 to 40 m.s⁻¹.
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