Topographic Controls on Pyroclastic Density Current Dynamics: Insight from 18 May 1980 deposits at Mount St. Helens, Washington (USA)

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1.0 Introduction

Understanding the dynamics of pyroclastic density currents (PDCs), arguably the most deadly consequence of volcanism (e.g., Auker et al., 2013), is crucial for accurately predicting the hazard potential of future explosive eruptions. PDCs travel at velocities that can exceed 100 m s⁻¹, and can maintain high temperatures and dynamic pressures over much of the runout distance. In many cases PDCs have enhanced mobility relative to other types of geophysical mass flows due to PDC's ability to retain high internal pore pressure over much of the flow duration. This can occur where there is trapped gas in a matrix with abundant fine ash (Bursik and Woods, 1996; Roche et al., 2004; 2008; 2010; Brand et. al., 2014a), and/or with autofluidization, where air is engulfed or injected upwards into the current as a PDC propagates and deposits across rough substrates (e.g., Chédeville and Roche, 2014; 2015). Due to the highly irregular topography around volcanic centers, understanding the influence of surface roughness, topographic irregularities, and slope on flow dynamics and runout distance of PDCs is critical for hazard assessment. This paper addresses two poorly understood aspects of PDC behavior: (1) the influence of slope on current dynamics, and (2) the controls on bedform development and migration in deposits of density stratified currents.

The 18 May 1980 eruption of Mt St Helens (MSH) produced column collapse-derived PDCs that deposited material within the crater and crater breach (slopes of 5-8°), along the steep flanks (10°-30°), and throughout the shallow-dipping pumice plain (~5°), up to 8 km north of the volcano (Figure 1). Column collapse occurred intermittently during the afternoon of 18 May between 1215 and 1815 hrs local time (Table 1). The pumice plain deposits, which are comprised of four thick (5 – 12 m each), massive and poorly-sorted flow units (total thickness ~40 m), are interpreted to represent the deposits of concentrated portions of PDCs generated

during the afternoon of the eruption (*cf.* Figure 1c of Brand et al., 2014a). The flank deposits, hereafter referred to as the *Proximal Bedded Deposits* (PBD), differ in that they are dominantly stratified to cross-stratified (Rowley et al., 1985; Beeson, 1988).

The objective of this work is to compare the PDC deposits from the crater and along the steep flank of the volcano with deposits of the pumice plain to understand the overall evolution of the afternoon PDCs with regards to flow dynamics, density stratification, transport, and sedimentation processes as a function of slope and distance from the vent. MSH is an excellent place to assess the influence of topography on PDC flow dynamics given the wide range of slopes, topographic obstacles, and surface irregularities; excellent modern day exposures; and previous work on the 1980 PDC deposits (i.e., Rowley et al., 1981 and 1985; Criswell, 1987; Beeson, 1988; Brand et al., 2014a; Mackaman-Lofland et al., 2014). In addition, deposits along the steep flank form dune bedforms, allowing for new insights into the processes that lead to bedform development in PDC deposits.

2.0 Methods

The inner-crater and proximal flank deposits were mapped with distance from source into the eastern pumice plain. Slope along the steep flanks is determined from pre- 18 May 1980 topographic maps (Beeson, 1988). Since the pre- 18 May 1980 topography of the pumice plain was substantially altered by the debris avalanche, slope beyond the flanks is assumed based on the gradient along the top of the PDC deposit fill.

Present day inner-crater deposits are exposed but highly eroded and inaccessible due to down cutting within the crater and north flank. The flank deposits are only partially exposed relative to photographs in previous work (Rowley et al., 1985; Beeson, 1988). The lower flank

and eastern pumice plain deposits remain well-exposed. As such, data from an unpublished master's thesis (Beeson, 1988), collected when the inner crater and flank deposits were well-exposed and accessible, are combined with samples and observations collected by the authors for this study. Four locations are compared herein: inner crater deposits, flank deposits, flank to pumice plain transition deposits, and the pumice plain deposits (Figure 2).

Measurements include granulometry, componentry, bedform shape, and bedform size. For granulometry, Beeson (1988) dried and sieved his 78 samples using a Fritsch Analysette Microsplitter. Samples collected by the authors here were sieved down to 8 mm in the field and down to 2 mm in the lab (also oven dried prior to lab sieving). Ash size was analyzed using the Coulter counter LS-100R laser particle size analyzer at the University of Washington Tacoma. The fines ratio ($F2/F1$) for PDC deposits is used to determine the degree of fines depletion or fines enrichment (where $F2 \leq 0.0625$ mm; $F1 \leq 1$ mm). Brand et al. (2014a), which considered all PDC deposits from the afternoon of the 18 May 1980 eruption, found that PDC deposits with $F2/F1 \leq 0.1$ are fines-depleted, $F2/F1 \leq 0.15$ are fines-poor, $F2/F1$ between 0.16 – 0.31 are fines-normal, and $F2/F1 \geq 0.32$ are fines-enriched.

Componentry from select samples, which include pumice, lithics, glass, and crystals, was measured in the field (grains >8 mm) and in the lab (grains down to 0.125 mm). In addition, the statistical method of Pollock (2013) is used to determine if the PDCs entrained lithic blocks from the paleo-substrate along the steep flanks of the volcano. The method assumes that the lithic population in the coarsest fall deposit from the 18 May 1980 eruption (found along the flank) is an acceptable proxy for vent-derived clasts during the climactic, column-collapse PDC-generating phase of the eruption. Lithic componentry was completed for the fall deposits and PDC deposits along the flanks and throughout the pumice plain. For each sample, 100 lithic

blocks (>64 mm) were collected, identified, and classified into one of the four recent eruptive period units (3 ka–present): Goat Rocks, Kalama, Castle Creek, and Pine Creek (see Hausback 2000, for a description of the lithologies). The average of three sampled fall deposits is used to represent the vent-derived clast population.

Samples of PDC deposits are compared to fall deposits using the Chi-Square Test for Homogeneity (e.g., Eisenhart, 1935). This method determines the statistical likelihood of the PDC lithic population having the same source as the fall deposits (vent erosion), or the likelihood that lithics were entrained from a second source (volcano substrate). A P-value represents the probability of obtaining a higher Chi-square value than the one calculated if both samples are derived from the same parent population (see Pollock, 2013, for method details). A significance interval of 0.05 distinguishes between samples that are likely to contain locally entrained blocks from those that do not. P-values > 0.05 indicate that the distribution of lithic lithologies in the PDC sample is consistent with those derived only from vent erosion. P-values < 0.05 indicate that the population of lithics in the PDC sample consists of both vent-derived clasts and locally entrained clasts from the substrate.

3.0 Previous Work: Phases of the 18 May 1980 Eruption and Pumice Plain PDC Deposits

The crater and north side of MSH were reshaped and eroded down to bedrock by the initial debris avalanche and lateral blast (Christiansen and Peterson, 1981; Rowley et al., 1985). PDCs began traveling northward through the breach and along the northeast and northwest flanks ~ 4 hours into the eruption (Phase III of Criswell, 1987), and continued throughout the afternoon of 18 May during Phases IV and V (Table 1; Figure 1c). PDC deposits from this period are found within the crater (inner crater deposits), along the flanks (Proximal Bedded

Deposits; PBD), and throughout the pumice plain (Figures 1 and 2) to a maximum distance of 8 km north of the vent. Table 1 gives the sequence of phases and deposits at MSH on 18 May 1980, based on previous work.

The original extent of the PBD along the flank, as mapped by Rowley et al., (1985), extended 100 m above the post-eruption crater floor along Sugar Bowl dome and down along the eastern flank, where slopes range from 10-30° (Figures 1a, 2). PBD were < 6 m thick where found above the crater floor and up to 20 m thick along the eastern flank (west of and below Sugar Bowl dome; Figure 1a). The PBD transition into the pumice plain deposits along the middle to base of the flank (Figure 2d), where an increase in the size and proportion of lithic clasts was also noted. Small, ≤ 4 m thick patches of PBD were also found along the 26° slope west of the breach. The western slope has significantly less surface topographic irregularities (e.g., slot canyons) to the flank east of the breach, which Rowley et al. (1985) suggest may have influenced the location and rate of deposition. Descriptions of crater and flank facies are provided in Table 2 and discussed in detail below. Terminology was adopted from Branney and Kokelaar (2002).

The thickest deposits from the afternoon PDCs are found throughout the pumice plain (Figure 1a and 2d) and are comprised of four primary flow units. These units, found across the entire pumice plain, are briefly described here for comparison purposes later in the paper; a more detailed account can be found in Brand et al. (2014a).

Units I and II are correlated with the first voluminous PDCs generated during eruptive Phase III (Table 1), and are interpreted to have filled in and around the debris avalanche hummocks scattered across the pumice plain. In proximal regions, both units I and II are dominantly diffusely-stratified and/or diffusely cross-stratified lapilli tuff in proximal regions,

126 suggesting transport and deposition under highly unsteady conditions where basal conditions
127 varied from low to somewhat higher basal shear stress, resulting in periods of traction transport
128 and/or granular flow boundary conditions. Evidence for traction transport is interpreted as a
129 consequence of variability in surface roughness and topographic irregularities, and unsteadiness
130 within the currents. In contrast, the distal deposits of Units I and II are dominated by massive
131 lapilli tuff, suggesting rapid sedimentation with negligible boundary shear stress. The change in
132 conditions is interpreted as the result of current compaction due to increased sedimentation with
133 distance from source.

134 Units I and II thicken with distance from source, suggesting portions of the currents
135 bypassed more proximal locations to deposit the bulk of the sediment distally. Pumice lenses are
136 common in both units and increase in size and abundance distally, suggesting efficient
137 segregation processes within the aggrading current. Units I and II are largely fines-depleted up to
138 6 km from source and fines-normal to fines-enriched to their distal exposures up to 8 km from
139 source. This pattern suggests elutriation was a dominant process across much of the depositional
140 region, but fines became progressively trapped as the currents densified due to compaction of the
141 granular mixture across the terrain. Pumices within the deposits are completely rounded within a
142 kilometer of the base of the volcano; thus, ash added by comminution is ruled out for the
143 increase in fines with distance from source (c.f., Dufek et al., 2008).

144 PDC Units III and IV are correlated with the climactic, most voluminous phase of the
145 eruption (Phases IV and possibly V; Table 1). Both flow units are dominated by lithic tuff
146 breccias and lithic block-rich massive lapilli tuffs. Lithic blocks are primarily derived from crater
147 widening and entrainment of clasts from the debris avalanche hummocks across the pumice plain
148 (Pollock, 2013; Pollock and Brand, in review), suggesting the flows traveled up and over

topography as well as between debris avalanche hummocks. The general absence of fabric, dominant fines-normal to fines-enriched deposits, weak evidence for segregation of lithics, and lack of pumice lenses in Unit III suggests transport and deposition from a highly-concentrated, fluidized current with suppressed size-density segregation and negligible basal shear rates (c.f., Druitt, 1995; Druitt et al., 2007). In contrast, the massive nature, accumulation of lithic blocks at the basal flow boundary, fines-depletion up to medial locations (up to 5.3 km) and abundant pumice lenses in Unit IV suggest transport and deposition from a concentrated current with negligible basal shear stress, but less concentrated relative to Unit III such that density segregation and fines elutriation occurred.

4.0 Results

Table 2 details basic descriptions and interpretations for lithofacies found within the crater and along the steep flank.

4.1 Crater and Flank Deposits

Crater deposits are ≤ 3 m thick and found within the gradually sloping ($5 - 8^\circ$) crater breach, ~ 1.92 km from the crater center (Figure 1a). The crater deposits are similar to the deposits throughout pumice plain in that they are poorly-sorted, block-rich massive lapilli tuffs (Figure 2b). Components include pumice, lithics and whole or broken phenocrysts. Zones or lenses of lithic breccia are common; pumice lenses and fines-depleted deposits are not observed. The massive lapilli tuff and a lithic breccia lens were sampled (orange triangles in Figure 3). Both generally have a larger median phi than the pumice plain massive facies but are somewhat better sorted (Figure 3).

In contrast, the flank facies PBD lithofacies include massive (unstratified) beds, plane parallel stratification, large-scale dune bedforms and small-scale, subsidiary dune bedforms, as detailed in Table 2; stratified and cross-stratified deposits dominate the upper flanks (Figure 2c). The PBD grade laterally (downslope) into less-stratified, massive deposits along the mid- to lower flanks of the volcano (transitional facies; Figure 2a and 2d), ultimately grading into the main facies deposits of the eastern pumice plain (Figure 2a and 2e). Thus, a significant variety in bedding characteristics exists between the crater, the PBD, and the main pumice plain deposits.

The PBD are composed of three primary facies associations, hereafter referred to as bed sets (BS1, BS2, and BS3) following the classification of Beeson (1988; Figure 4). Each bed set is distinguished by changes in bedding characteristics and separated by sharp, often erosive contacts. The lithofacies that comprise each bed set are complex and highly variable with distance from vent, necessitating the more detailed lithofacies analysis provided in Table 2 and below. Broadly speaking, BS1 is dominantly cross-stratified with large and small-scale progressive and regressive dunes, BS2 is dominantly massive or diffusely stratified, and BS3 contains planar stratification, cross-stratification or massive bedding. These bed sets transition downstream into less stratified, more massive deposits with complex interrelationships in the mid- to lower-flank deposits.

4.1.1 Lower Bed Set (BS1)

BS1 deposits have planar-stratified to cross-stratified bedforms with abundant internal bed subsets bounded by erosional surfaces (Figure 4). BS1 is the thickest of all the bedding sets (around 8 meters maximum) and is in direct contact with the bedrock of the volcanic flank (either pre-1980 lithologies or early 18 May 1980 deposits), although the basal contact is rarely exposed. Strata are distinguished by abrupt changes in grain size (alternating between coarse-

grained stratification and fine-grained stratification; Table 2) and componentry (dominated by pumice lapilli, lithic lapilli, or a combination of pumice and lithics in a matrix of crystals and ash; Figure 3f). Pumice are subangular to subrounded; lithics are generally angular to subangular. Lithofacies within BS1 are dominated by regressive bedforms, progressive bedforms, and many subsidiary type I and II bedforms (Table 2). Medium to thick massive beds (> 1 m), plane parallel, and wavy massive lapilli tuff beds are present to a lesser degree. Pumice lenses and lithic lenses occur sporadically.

In comparison to the other bed sets, BS1 has the widest variability in median grain size and sorting with distance from source (Figure 3; Table 2). The deposits primarily plot within the field of dilute PDC deposits (Figure 3e; after Walker 1971; 1984) but also overlap with the main PDC-deposit facies within the pumice plain. Block-sized lithic concentrations occur as elongated, continuous trains or lenticular lenses. Lithic-block concentrations are most common in downslope, lower gradient locations near the transition from flank to pumice plain deposits. BS1 grain-size distributions range from fines-poor to strongly fines-enriched, but are on average fines-normal, and there is no pattern of increasing or decreasing fines either with distance from the vent or increasing depositional slope (Figure 3d).

4.1.2 Middle Bed Set (BS2)

The contact between BS1 and BS2 deposits is unconformable, and often has an angular relationship with the cross-strata of BS1 (Figure 4). BS2 is 1.0 m to 5.0 m thick and can be subdivided into two to three medium to thick-bedded, massive lapilli tuff subsets ranging from 10 cm to >100 cm. Subset contacts are generally sharp, although lenses of fine-grained type II subsidiary dune bedforms (Table 2) variably occur at subset boundaries. Additional localized and minor bedding features include 3 – 30 cm thick planar stratification and lenticular lithic

concentrations. Pumice concentrations are rare throughout BS2. Strata are distinguished by changes in grain size and componentry (Figure 3). Pumice are subangular to subrounded; lithics are generally angular to subangular.

BS2 grain-size distributions also primarily plot within the field of dilute PDCs with some overlap with the main facies pumice plain deposits (Figure 3e). While BS2 has wide variability in median grain size and sorting with distance from source, the spread is less than that found in BS1 (Figures 3a and 3b). Average fines data indicates that BS2 borders on fines-poor, and there is no pattern of increasing or decreasing fines either with distance from the vent or increasing depositional slope (Figures 3c and 3d).

4.1.3 Upper Bed Set (BS3)

The contact between BS3 and BS2 is sharp and often erosional (Figure 4). The maximum thickness of this bedding set is indeterminable due to erosion and reworking of surface strata. Common depositional features include regressive bedforms, subsidiary type II bedforms, and fine-grained, thin (3-10 cm) to medium (10 cm) plane parallel to wavy massive lapilli tuff beds. Pumice and lithic lenses rarely occur in the bed set. BS3 grain-size distribution is fines-poor to fines-normal and generally finer-grained and better sorted than the other bed sets (Figure 3). There is no pattern of increasing or decreasing fines with increasing depositional slope (Figure 3d). Pumice are subangular to subrounded; lithics are generally angular to subangular.

4.3 Nature of Bedforms along the Upper Flanks

All major, large-scale bedforms are found to the west and below Sugar Bowl dome (Figures 1a, 2c and 5) on slopes of 10 – 30°. Regarding terminology, form height refers to the thickest part of the bedform crest (often referred to as wave height). Wavelength is the distance

between successive crests; form length is the horizontal distance between the first stoss side to the last lee side expression. Form length was also recorded because consecutive bedforms were not always exposed to measure wavelength.

4.3.1. Progressive and Regressive Dune Bedforms

Progressive dune bedforms, low angle, broad bedforms with thick, well-developed strata on the lee-side (Table 2; Figure 6b and 7a), are found on depositional slopes of 12°- 16°. They occur most often in the upper portion of BS1. Isolated bedforms have form lengths from 2 – 14 m and form heights from 0.4 to 2.5 m (Figure 6b). Where found, consecutive progressive bedforms have wavelengths from 3 – 7.6 m.

Regressive dune bedforms, which are low-angle, broad bedforms with thick, well-developed strata on the stoss-side and generally thinner strata on the lee-side (Table 2; Figure 7b-d), are found on depositional slopes of 10°- 28°. Isolated bedforms have form lengths from 2 – 15 m and form heights from 0.5 – 5 m. Where found, consecutive bedforms have wavelengths from 2 to 8 m. Locally, regressive bedforms have well-developed, high-angle stoss-side bedding (Figure 7d and e). Cores of regressive dune bedforms with steeply dipping backset bedforms are truncated on the stoss-side by steeply up-flow dipping truncation surfaces (from 30 up to an extreme of 68°) with a sigmoidal shape. A series of well-developed, thick, upflow-stacking beds, also sigmoidal in shape, are found upstream with accumulation lengths of up to 15 m (Figures 7d and 7e). These steeply-dipping backset beds are only found on slopes of 15°-26°. They are better sorted and have lower F_2/F_2 and $Md\phi$ values relative to low-angle regressive bedforms (Table 2). The steeply-dipping backset beds are generally fines-normal.

Progressive and regressive bedforms show systematic changes in form length, wavelength and form height as a function of slope. From a decreasing slope of 16° to 12°

degrees, progressive bedform wavelengths decrease 4.5 m, form lengths decrease 10 m, and form heights decrease 2 m (Figure 6b). Similarly, the wavelength of regressive bedforms decreases by 2.5 m and the form height decreases by 1.25 m for every 5° decrease in basal slope (Figure 6c). In addition, the number of stoss-side bed sets and length of stoss-side accumulation, while somewhat random between slopes of 10-20°, increase significantly above slopes of 20° (Figure 6d).

4.3.2 Compound Bedforms

Compound bedforms involve all three bed sets (Figure 5c and 5d). These bedforms are *only* found within 3 km of the vent in the saddle west of Sugar Bowl dome and immediately below Sugar Bowl dome, on slopes ranging from 10°- 28°. They are large-scale, longitudinal dunes that internally consist of a core (BS1) and a sequence of draping beds (BS2 and BS3), separated by distinct erosional surfaces (cf., Beeson, 1988; Figure 5c and d). Bedform amplitudes range from 5 – 15 m, form lengths range from 10 – 50 m, and wavelengths range from 25 – 50 m. The average form length decreases by 6.25 to 25 m and amplitudes decrease by 3 to 5 m for every 5° decrease in slope (Figure 6a). The compound bedforms coincide with the thickest PBD deposits.

Cores include progressive, symmetric and/or regressive sandwave bedforms with alternating coarse-scale and fine-scale stratification and complicated inter-relationships due to multiple erosion surfaces (Figure 5d). Some cores consist of internally massive plane parallel beds with an upstream-dipping erosion surface that develop the entire feature into a dune-shape. In only one location a truncation surface was noticed on the lee-side of the bedform. All other compound bedforms have stoss-side erosional surfaces ranging from 5 to 47°. Strata on the upstream side of dunes alternate between coarse- and fine-grained strata, often have lenticular

concentrations of pumice, and are generally more than twice as thick as strata found on the lee-side of bedforms. The stoss-side strata thin toward and over the crest of the bedform, and generally maintain the same thin, planar nature that parallels the core in the down-flow direction.

BS2, which mantles BS1 in the compound bedforms, is often thinnest over the crest of the internal bedform and thicker in the troughs. On the source-ward side of compound bedforms, BS2 is often truncated by an erosional surface, followed by BS3. Like BS1, the BS3 strata are thicker on the upstream side of the bedforms (Figure 5d).

4.4 Flank to Pumice Plain Transition Deposits

The proximal bedded deposits gradually transition downslope into the thick, massive deposits of the pumice plain (Figure 2). Across this transition individual bed sets become thicker and less numerous (Figure 8). In all bed sets there is a distinct increase in lithic blocks, which are found scattered or as concentrations, and coarse pumice lenses.

Since continuous exposures were not available in the 1980's, Beeson (1988) used lithic concentration zones in the lower portion of BS1 to trace the transition to the proximal pumice plain deposits. Changes to BS1 include a gradual thickening and coarsening of individual beds and, eventually, a decrease or absence of large-scale dune bedforms. BS2 is dominated by internally massive, thick beds with an increase in thickness and occurrence of lithic and pumice concentration lenses across the transition. BS3 coarsens toward the pumice plain, primarily due to an increase in the proportion and size of pumice clasts, and maintains coarse-scale stratification across much of the transitional zone. Over the transition zone on low slopes the compound bedform large-scale morphology is subdued (less pronounced) due to thin beds over the crest of the bedform and thick beds in the troughs of the bedforms.

All bed sets continue to thicken, coarsen, and become more massive as the slope decreases; only massive lapilli tuff with pumice lenses and faint, diffuse stratification are found beyond the final break in slope and across the shallow-dipping eastern pumice plain (slope of 5°).

5.0 Discussion

PDCs are fundamentally morphodynamic; they are subject to rapid flow transformation due to nonlinear interaction with the substrate due to erosion, fluid turbulence, fluid-particle interactions, air entrainment, deposition and topographic effects (e.g., Branney and Kokelaar, 2002; Pittari et al., 2006; Brand et al., 2014a). These rapid flow transformations lead to considerable vertical and lateral heterogeneity in both the physical properties of the current and subsequent deposits, as demonstrated by the afternoon PDC deposits from the 1980 eruption of MSH. The two biggest challenges to researchers in the realm of PDC dynamics are (1) identifying relationships between depositional features and the currents that produced them, and (2) understanding how PDCs behave as a function of initial conditions (i.e., mass flux, bulk concentration, column collapse or dome collapse dynamics) and as a result of interaction with topography. Both are necessary to improve our ability to understand and mitigate associated hazards, such as the influence of rheological transformations on dynamic pressure and runout distance.

The exposures through the 18 May 1980 afternoon PDC deposits at MSH provide one of the few case studies where exposures are sufficient to correlate individual flow units from proximal to distal from source over a range of topographic conditions (Brand et al., 2014a; Mackaman-Lofland et al., 2014; Pollock and Brand, in review). The deposits at MSH allow us to infer density stratification and influence of topography (slope) on erosional, transport and

depositional changes within the current. The findings presented here will provide a series of hypotheses that can be tested experimentally, and also provides a data-rich case study that will be useful for partial-validation of computational models.

5.1 Interpretation of Bed Set Characteristics and Correlation with Pumice Plain Flow Units

The predominantly regressive nature and common presence of steeply-dipping backset bedforms within BS1 (Figure 7b-e), combined with multiple truncation surfaces, suggests strong influence of fluid turbulence on transport and deposition (e.g., Walker, 1984; Schmincke et al., 1973; Brand and Clarke, 2009; Douillet et al., 2013). These characteristics, combined with the often reverse graded and diffusely-stratified (due to grain alignment) nature of individual strata within bedforms and plane-parallel beds, suggests deposition from a predominantly turbulent current with a well-developed, high-concentration basal region. The basal region must be thin enough that motion is driven by shear stress exerted from the overriding, turbulent portion of the flow, promoting traction transport (Sohn, 1997) and grain segregation via grain-dispersive processes (Bagnold, 1954).

Thinning and shearing would initially increase as the currents accelerate and thin downslope. Shear force is also expected to be high at the base of the current given the high surface roughness of the volcanic flank (e.g., Figure 5a). Cross-stratification would occur when turbulent sweeps cross through and disrupt the basal region of the current; strong sweeps would erode into previously deposited strata, resulting in the bed truncations and cross-stratification common to BS1.

Medium to thick massive beds and plane parallel bedding, found interbedded in more proximal locations but becoming the dominant deposit characteristic with distance from source, are interpreted to represent conditions where the concentration and thickness of the concentrated

basal region increases due to high rates of sedimentation from the overlying turbulent suspension. The thickness would increase to the point that the effect of the overriding turbulence was largely suppressed, and sediment was deposited from a laminar flow regime (e.g., Lowe, 1982). The concentration and thickness of the basal region is expected to increase as current energy decreases away from source due to decreasing slope (Sulpizio and Dellino, 2008). Deposits and interpretations such as these were first established in the subaqueous sediment gravity current literature (e.g., Middleton, 1970; Lowe, 1982; Postma, 1986), and subsequently in the PDC literature (e.g., Sigurdsson et al., 1987; Fisher, 1990; Cole and Scarpati, 1993; Sulpizio et al., 2008).

While there is significant overlap of statistical parameters with all units found within the pumice plain, the BS1 stratigraphic position, $Md\phi$, sorting, and the spread of F2/F1 are most consistent with the pumice plain Units I and II (Figure 3). While primarily massive with diffuse stratification, Units I and II are also locally stratified to cross-stratified in the pumice plain, further supporting the correlation with BS1 and suggesting that the current(s) were relatively less concentrated in comparison with the later (climactic) afternoon PDCs (Table 1; Brand et al., 2014a). The multiple erosional surfaces suggest unsteady pulses of erosion and deposition from a single current, or from multiple, closely-spaced currents, the latter consistent with the correlation with two flow units in the pumice plain.

The thicker, massive layers within of BS2 and BS3, which also commonly display diffuse stratification and alternating coarsening and fining-upward sequences, are interpreted to have developed under sustained traction-carpet sedimentation where the grain size of supplied sediment varies with time due to unsteadiness in the current(s) (e.g., Lowe, 1982; Sohn, 1997). Sustained deposition under traction carpet transport mechanisms would result in diffusely

378 stratified beds; whereas, unsteadiness and intermittent sweeps of turbulent eddies from the upper
379 portion of the current would promote temporary deposition of the planar- and cross-stratification
380 observed throughout both bed sets (e.g., Wilson and Walker, 1982; Beeson, 1988; Cole and
381 Scarpati, 1993; Branney and Kokelaar, 1997). Thus intermittent stratification and reverse
382 grading is likely the consequence of traction carpet processes, granular shearing, and dispersive
383 forces within the basal region (Lower, 1982; Sohn, 1997), as described for BS1. The more
384 common massive bedding in BS2 and BS3 suggests that the responsible currents had a higher
385 concentration and/or thicker concentrated basal region relative to the current(s) that formed BS1.
386 The relatively higher concentration conditions could be a consequence of increasing mass flux at
387 the vent, which may lead to higher particle to gas ratios. This interpretation is supported by
388 observations that an increase in mass flux occurred throughout much of the PDC-producing
389 eruptive phases (e.g., Carey et al., 1990). The BS1 deposits also likely smoothed the landscape
390 to some extent, except for the low-amplitude bedforms, such that the effect of basal shear stress
391 may also have been diminished relative to what was experienced by the BS1 current(s).

392 The BS2 stratigraphic position, $Md\phi$, sorting, and the spread of F2/F1 values are most
393 consistent with pumice plain Unit III. Unit III also displays the least degree of grain segregation
394 and bedding features, consistent with the commonly massive and depression-filling nature of
395 BS2. Brand et al., (2014a) correlate Unit III with the climactic phase of the eruption and thus the
396 highest mass flux, further supporting the reason for a high concentration current relative to BS1,
397 as discussed above. Finally, the BS3 stratigraphic position, $Md\phi$, sorting, and especially the low
398 values of F2/F1 are most consistent with the pumice plain Unit IV. Weak stratification and cross-
399 stratification in Unit IV are also consistent with the alternating massive to stratified deposits of

BS3, suggesting the responsible current was less concentrated relative to that of the BS2-Unit III current.

5.2 PDC Regime Changes due to Changes in Slope

The massive deposits within the crater of MSH, < 2 km from the vent, contain lithic breccia lags and lenses within a massive lapilli tuff. Granulometry characteristics plot within the field of concentrated PDCs (Figure 3; after Walker, 1984) and the development of lithic lenses indicate deposition from a laterally-moving flow rather than deposition from ballistic or mass fall out from the column. Thus density stratification with the development of a thick, concentrated basal region and some degree of grain segregation developed within 2 km of the vent. Such massive deposits are interpreted to represent thick, concentrated currents where shear stress at the base is low to negligible due deceleration of settling clasts, a consequence of hindered settling, and high sedimentation rates (e.g., Lowe, 1982; Branney and Kokelaar, 2002; pages 66-71; Roche et al., 2004).

It is unclear exactly how the crater deposits correlate with the timing of deposition along the flank and throughout the pumice plain. They could have been deposited at any time on the afternoon of 18 May. If the currents were depositional before they reached the steep flanks, it would suggest that a collapsing column can transition into a density-stratified, sediment gravity current within two kilometers from source. However, it is equally possible that the crater accumulations were deposited during a waning phase of PDC deposition, which became depositional within the crater as energy decreased. Unfortunately, the data collected in the 1980s combined with the modern-day exposures are not sufficient to resolve this interpretation. This ambiguity motivates the question of the spatial and temporal scales over which PDCs derived from collapsing columns transition from momentum driven flows into sediment gravity currents.

The southern Sugar Bowl dome saddle is >100 m above the crater floor. Thus the deposits on the southern saddle likely result from the upper, dilute portions of the PDCs surmounting and depositing along topographic highs (e.g., Figure 1a and 2e), similar to the deposits found at Tungurahua volcano, Ecuador (Douillet et al., 2013). However, the majority of PBD are found along the west side and below Sugar Bowl dome in an area that is topographically *lower* than where currents were depositing in the crater breach (Figure 2a and 2f). The terrain here becomes steep (up to 30°) and highly irregular with many ridges and small drainages (Figure 5a and 5b).

We interpret that portions of the concentrated afternoon PDCs were funneled between Sugar Bowl dome and a topographic high to the east, then accelerated and thinned as they descended and spread across the steep slope. The concentrated basal region of the current(s) thinned to the point where the overriding turbulence strongly influenced transport and deposition (detailed interpretations in section 5.1), and the basal region instead behaved as a bedload at the base of a dominantly turbulent flow, marking a major change in current dynamics. The BS1 currents developed large-scale bedforms to create an undulating surface of regularly spaced, low-angle bedforms (e.g., Rowley et al., 1985). The generally fines-normal PBD suggest that elutriation was either unimportant in the concentrated basal region of the current or, more likely, that comminution of pumice and the addition of fine ash kept up with elutriation. Comminution is further supported by the subangular to subrounded nature of the pumice, which suggests that pumice were in a state of active but incomplete rounding upon deposition along the steep flanks. Comminution can result in a significant increase in the fines proportion (Dufek and Manga, 2008; Manga et al., 2011 and references therein), and is the most likely explanation for the fines-normal nature of the PBD.

The currents responsible for BS2 and BS3 are interpreted similarly; concentrated PDCs from later in the afternoon were funneled into this somewhat (initially) confined region and experience the same acceleration, spreading and thinning as they descended the steep slopes. The difference is that the more massive and diffuse stratified nature of the BS2 and BS3 deposits are consistent with higher concentration currents. BS2 and BS3 deposits largely mantle the BS1 bedforms < 3 km from source, suggesting currents with a thin-enough bedload and high-enough energy that topographic effects did not result in significantly thicker deposition in troughs and thinner deposition on crests. However, beyond 3 km from source BS2 and BS3 tend to be thicker in depressions and thinner on topographic highs, while BS1 bedforms persist. This observation suggests that BS2 and BS3 experienced a decrease in flow energy as slope decreases and provides further evidence that they were more concentrated relative to BS1 current(s).

Recent work by Chédeville and Roche (2014; 2015) highlight the importance of autofluidization, where an upward air flux results from substrate interstices as flow particles settle between grains. Experimental currents ($d = 0.08$ m) traveling across a rough, nonerodible substrate with a single sheet of glued beads ($d = 1.5 - 3$ mm) have up to twice the runout distances than currents with the same initial conditions traveling across a smooth substrate. Chédeville and Roche (2015) found that autofluidization occurred for slopes from $0 - 30^\circ$. However, for slopes $>16^\circ$, a strong deceleration phase, interpreted as the result of higher shearing and energy dissipation due increased collisions as the currents thinned across higher slopes, resulted in shorter runout distances.

Bareschino et al. (2008) demonstrate a second mechanism for autofluidization – “plunging breakers”. Their experiments, which involve granular flow within a large, rotating drum, simulate dense granular flows traveling across an incline. Increased velocities resulted in

flow front instabilities, where the upper portion of the current head extends out beyond the basal region (overhanging) and collapses as an avalanche, resulting in substantial entrainment of ambient air.

Given the high surface roughness and irregular topography along the flanks, it is likely that mechanisms for autofluidization described by Chédeville and Roche (2014; 2015) and Bareschino et al. (2008) were important for the 18 May 1980 PDCs descending the steep flanks. Plunging breakers would arise both from the increase in velocity and thinning of the current, as well as interaction with topographic obstacles and irregularities. Evidence for supercritical flow conditions (high flow regime bedforms) also suggests that entrainment of ambient air would be important along the steep flanks, further influencing flow dynamics (Bursik and Woods, 1996). Thus, in addition to simple thinning of the basal region of the current, expansion of the currents due to autofluidization and ambient air entrainment also likely served to change flow behavior from that of concentrated PDC conditions to relatively dilute, turbulent PDC conditions, as described above.

Around 3.3 km from source, along slopes of 22-28°, the deposits begin to grade into a complicated intermixture of bedforms, stratified to diffusely-stratified deposits with splay-and-fade features (e.g., pg. 106 of Branney and Kokelaar, 2002), and massive block-rich deposits (Figure 8). Truncation surfaces are abundant, suggesting highly unsteady currents with periods of erosion, while the massive block-rich deposits and splay-and-fade features suggest rapid deposition and flow transformations over short distances (< 100 m). The deposits continue to thicken and become more massive with fewer truncation surfaces with distance from source, suggesting rapid sedimentation and low shear stress at the base of a thicker concentrated basal region current relative to the thinner basal region and more dilute conditions on the upper slopes.

The influence of slope on density current evolution and depositional patterns is explored for PDCs using field relationships (Gurioli et al, 2002; Cole et al., 2005), a combination of field observations and laboratory experiments (Macias et al., 1998; Roche et al., 2004; 2008; 2010; 2014; Roche, 2015), and a combination of field observations and theory (Sulpizio et al., 2007; Sulpizio and Dellino, 2008; Bernard et al., 2014). Similar studies have been completed for turbidity currents (e.g., Mulder and Alexander, 2001; Gray et al., 2005). These studies show that, as flow energy decreases with distance and decreasing slope, the sediment supply from the upper portions of the current would increase, resulting in a thickening of the concentrated basal underflow. Thickening of the concentrated basal region due to increased sedimentation from the suspended load would suppress or eliminate the influence of turbulence from the overriding current. If the current is fines-rich, due to the presence of fines produced via fragmentation or comminution, the current may transition to a fluidized PDC with high internal pore pressure (e.g., Roche et al, 2010); if the current is fines-poor, the flow may behave closer to a dry granular flow (e.g., Bernard et al., 2014). In addition, the ratio of the resisting stress to driving stress will increase as the slope decreases. Thickening of the underflow region and a decrease in slope would also result in a partial redirection of momentum flux per unit area (Iverson and Denlinger, 2001), effectively decreasing basal shear stress and increasing normal bed frictional forces at the flow-deposit contact (Sulpizio and Dellino, 2008; Sulpizio et al., 2008). These conditions would promote deposition, and offer an additional explanation for the thickening of deposits and lack of bedding features along shallow slopes.

Sediment gravity currents crossing a break in slope often undergo a hydraulic jump, resulting in increased deposition, increased turbulence and, as a result, an increase in ash elutriation (e.g., Macias et al., 1998; Gray et al., 2005; Sulpizio and Dellino, 2008). If the upper

portion of the current(s) remains more dense than the atmosphere it will override the depositional base and continue downstream as a density current. If the upper dilute portion is less dense than the atmosphere, the fines will loft into a co-ignimbrite cloud.

While the presence of coarse lithic clasts increases with distance from source within the PDC deposits, a co-ignimbrite ash deposit is not apparent between flow contacts across the lower flank and eastern pumice plain. In addition, the eastern pumice plain deposits are thick and fines-normal, supporting the former interpretation of continued lateral flow of the afternoon currents. It is possible that co-ignimbrite plume(s) also developed due to increased turbulence at the break in slope, but were either non-depositional in the study area (transported elsewhere by wind) or subsequently eroded by later PDCs.

Exposures of massive lapilli tuff persist more than 7 km to the north, reaching Spirit Lake. The lack of clast fabric, fines-normal to fines-enriched nature of the eastern pumice plain deposits and increasing thickness with distance from source suggest retrogradational aggradation from currents that likely maintained high internal pore pressure and fluid-like behavior across much (if not all) of their runout distance. The influence of changing flow dynamics on the runout distance of the current is not clear since the presence of the lake prevents determination of the full runout distance.

The transition from massive crater deposits, to the PBD along the steep flanks, to the transitional deposits along the mid-flanks, and finally back into the massive deposits in the eastern pumice plain, suggests that slope and uneven topography strongly influence the dynamics and depositional pattern of a given current. One important issue to recognize is that, similar to the massive crater accumulations, it is not possible to know for certain whether each unit within the PBD were deposited over the duration of a given, semi-sustained, waxing and waning

current, or during a waning phase of PDC generation. The difference in thickness between the flank and pumice plain units, combined with a lack of lithic blocks on the upper flanks, suggests portions of the current were non-depositional and bypassed the upper flanks to deposit in the shallow-dipping pumice plain. This retrogradational pattern is further supported by the general thickening of PDC deposits distal from source, and thinning toward source throughout the pumice plain (Brand et al., 2014a). It is also possible that the mid- and upper-flank deposits were deposited during the waning phase of flow.

5.3 Evidence for Erosion along the Steep Flanks

Experiments and numerical modeling of dry granular flow demonstrate that erosion most affects flow behavior above angles of roughly half the repose angle of the erodible material (~15° for most granular substances), and that erosion into a granular substrate along slopes near the angle of repose can increase runout by up to 40% (e.g., Mangeney et al., 2010; Farin et al., 2014; Bernard et al., 2014). Erosion along the steep flanks of volcanoes is noted for PDCs at Tungurahua volcano, Ecuador (Bernard et al., 2014) and in pumice flows in column collapse-derived PDCs at Lascar volcano, Chile (Calder et al., 2000). Componentry analysis of deposits exposed at Tungurahua suggest erosion primarily occurred on slopes > 25°; whereas exposures at Lascar suggest erosion primarily where the PDCs encountered topographic obstacles or moved through constrictions. Bernard et al. (2014) numerically investigate the mechanisms and consequences of substrate erosion on current runout. They find that thickness of the erodible layer and fluctuations in basal shear stress due to variations in current thickness are the primary controls on a current's erosive capacity. Their model suggests a 5 – 20% increase in runout due to current bulking. However, the experiments of Farin et al. (2014) suggest that bulking is not

the cause of runout increase. Thus the influence of erosion by granular flows along steep flanks remains poorly understood. Given that the currents at MSH were dilute, turbulent, and fluidized PDCs, not dry granular flows, further complicates our ability to determine the influence of substrate erosion on ultimate runout distance.

Along the flanks of MSH, the prominent truncation surfaces between bed sets indicate that the initial passing of the currents were strongly erosive. Pervasive truncation surfaces within bed sets, often with stoss side aggradation or aggradation along steeply-dipping backset beds, indicate intermittent erosion and rapid deposition, likely a function of turbulent sweeps across the depositional boundary followed by periods of rapid deposition from the concentrated basal region. Features such as these emphasize the highly unsteady nature of currents descending the steep flank.

The primary lithology along the eastern upper flank is Castle Creek basaltic and basaltic andesite lava flows, which are exposed within steep slot canyons. Deposits are found both within and on top canyon walls, indicating the currents that deposited here experienced high surface roughness and partial channelization (Figures 5b and 5c). Two samples from BS2 were collected from exposures along the mid-to-upper flank (transitional zone, slope $\sim 20^\circ$; Figure 2a) and analyzed for componentry. Both samples are enriched in Castle Creek Andesite relative to the fall deposits, consistent with an increase in the upstream lithology (Figure 8). In addition, both samples have P-values $\ll 0.05$, providing statistical evidence for upstream substrate erosion in both bed sets (BS1 and BS3 were not sampled due to lack of lithics in this location and lack of modern day exposure downstream). The statistical results indicate that erosion occurred upstream within the slot canyon where Castle Creek lavas and rubble are exposed (Figure 5b). These findings are consistent with the results of Bernard et al. (2014) and Calder et

al. (2000) who find that substrate erosion is promoted at slopes of $\sim 25^\circ$ (former) and in the presence of topographic obstacles or constrictions (latter).

The largest blocks found in each sample range from 8 – 29 cm in their longest direction (five largest clasts for each sample presented in Figure 8d). The theoretical approach of Roche (2015) demonstrates that, for a dilute, turbulent PDC traveling across a 10° slope, the largest dense lithic that can be entrained is 18 cm (for densities of $2000 - 3000 \text{ kg/m}^3$). However, the simplifying assumptions in Roche (2015) break down for the PBD because the surface roughness is significant (greater than a single clast diameter) and the current is not fully dilute. Regardless, the larger entrained blocks found along the flanks of MSH suggest that dilute PDCs with concentrated basal regions are capable of entraining dense lithics up to 29 cm on slopes of $> 25^\circ$.

5.4 Relationships between bedform morphology and PDC dynamics

Despite decades of advancements, our ability to interpret PDC dynamics from deposit characteristics, especially those deposited by turbulent PDCs, remains limited. Turbulent PDCs span the spectrum from fully turbulent, low particle-concentration currents where the sediment load is dominantly transported in suspension, to turbulent PDCs where density-stratification and high sediment concentration in the basal region of the current largely suppresses turbulence in the depositional region (Sigursson et al., 1987; Branney and Kokelaar, 1992; Druitt, 1992; Cole and Scarpati, 1993). Fully dilute-turbulent to higher concentration, stratified, turbulent PDCs form in a range of eruptive conditions including (but not restricted to) lateral blasts (e.g., 1980 eruption of Mt St Helens; Kieffer, 1981), Plinian column collapse (Rowley et al., 1985; Cole and Scarpati, 1993), ash-cloud surges that can detach from concentrated PDCs (Fujii and Nakada, 1999), and phreatomagmatic eruptions (e.g., Moore, 1967; Belousov and Belousova, 2001;

Brand and Clarke, 2012; Brand et al., 2014b). The low particle concentration allows dilute PDCs to surmount large obstacles, increasing their hazard potential.

Due to their dynamic similarity, much of the existing theory for PDC transport and deposition mechanisms is derived from our understanding of turbidity currents (e.g. Lowe, 1982; Postma, 1986; Kneller and Buckee, 2000), despite the fact that gas-particle flows have the potential to behave as compressible fluids (Kieffer, 1981; Wohletz, 1998; Esposti Ongaro et al., 2008). There is clear evidence in the deposits of both dilute PDCs and turbidity currents that current velocity is a controlling factor for bedform morphology and migration. For example, dune wavelengths are observed to increase with increasing slope in dilute PDC deposits, suggesting that bedform wavelength is a function of the flow velocity (Crowe and Fisher, 1973; Fisher and Schmincke, 1984). Similarly, dunes found in turbidity current deposits associated with higher flow velocities, such as in channels, tend to have longer wavelengths and commonly appear to migrate upstream (e.g., Stromboli Canyon, Kidd et al., 1998; and the El Julan and Icod wave fields, Canary Islands, Wynn et al., 2000; Wynn et al., 2002). Morris et al (1998) note that wavelengths of regressive bedforms formed in turbidity currents decrease with distance from source, which they associate with an increase in sediment concentration, decrease in velocity, decrease in flow thickness, or combination of the three. The same trend is noted in the dune deposits of dilute PDCs, both along the flanks of Mt St Helens (this study) and in dilute PDC deposits commonly found in tuff cones and tuff rings (e.g., Wohletz and Sheridan, 1979; Brand and White, 2007). The proximal bedded deposits at MSH demonstrate that these same controlling parameters influence bedform formation and migration in turbulent, density stratified currents.

Unfortunately, the hydraulic and emplacement interpretations of PDCs responsible for dune bedform are dominantly based on relationships derived for open channel flows (*cf.* Fedele et al., 2007; Spinewine, et al., 2009), with few studies actually testing these relationships for deep ambient density underflows (with the exception of Hand, 1974). Thus sediment transport and deposition within PDCs, and the controls on bedform morphology and migration, remain poorly understood and debated (e.g., Valentine, 1987; Spinewine, et al., 2009; Douillet et al., 2013). Identifying the relationships between parameters such as bedform wavelength, amplitude, accretion, and migration of dune bedforms to current velocity, thickness, and density concentration profile would not only give us a better fundamental understanding of the dynamics of these complex, density stratified flows, but will also allow us to better appreciate and predict the hazards associated with them. Here we extract relationships that link the PBD to flow regime, and highlight deposit features that likely record important information about the current(s) that formed them, but as of yet cannot be directly linked to PDC conditions at the time of emplacement.

5.4.1 Regressive and Progressive Bedforms – Evidence for Flow Criticality

Flow criticality is governed by the ratio of the inertia forces over gravitational forces, which is represented by the Froude numbers (Fr), $Fr = \frac{U}{\sqrt{g'h}}$, where U is velocity and h is the flow depth. Reduced gravity is accounted for with $g' = g \frac{(\beta - \alpha)}{\alpha}$, where g is the gravitational acceleration, β is the bulk density of the current, and α is the density of the atmosphere. Studies in open channel flow show that progressive bedforms develop under subcritical conditions, where $Fr < 1$, and regressive bedforms develop under supercritical conditions, where $Fr > 1$ (Blatt et al., 1980). While it is well-established that flow criticality controls bedform shape and

accretion in open-channel flow, it is unclear if the same relationships hold for density stratified, sediment gravity currents, such as dilute PDCs.

The PDC deposits along the flank show no evidence for liquid water at the time of deposition, such as accretionary lapilli, vesiculated tuff, bedding plane sags or soft sediment deformation. Thus bedforms developed within dry, two phase (gas and particles) PDCs. Recall that regressive bedforms dominate, especially on steep slopes ($>15^\circ$) and in proximal locations. Progressive bedforms are only found on slopes of $<15^\circ$ and at distances >2.5 km from source. The location of regressive bedforms on steeper slopes and most proximal to source is consistent with where we might expect high energy, supercritical flow conditions. Further, the increase in dune bedform wavelength and amplitude with increasing slope, and decreasing wavelength and amplitude with decreasing slope and with distance from vent, indicate the flow energy and velocity are the primary controls on bedform form length and form height. However, even though regressive and progressive bedforms were never found within the same bedding horizon, the overlap in regressive and progressive bedform locations on slopes $<15^\circ$ suggests that velocity is not the only controlling factor on bedform development.

The steeply-dipping backset beds along the steepest regions of the flank are consistent with what is interpreted elsewhere as chute-and-pool features, which occur across a hydraulic jump (when a current abruptly transitions from supercritical to subcritical; e.g., Jopling and Richardson, 1966; Schmincke et al., 1973). While rarely preserved in the fluvial record, chute-and-pool features are common in the deposits of dilute PDCs (e.g., Schmincke et al., 1973; Brand and White, 2007; Brand et al., 2009; Brand and Clarke, 2009). The internally massive nature of the PBD steeply-dipping backset beds, and increasing number of upstream dipping bed sets with increasing slope (Figure 6d), suggest that high rates of sedimentation on the upstream

side of bedforms was also important. Thus while the erosive pulse may have elutriated fines, the concentrated portion that followed likely had high rates of deposition, preventing further fines elutriation.

While the deposits record information regarding the conditions at the time of deposition, they do not record the reason behind their initial development, continued aggradation, or consistent relationship between increasing wavelength and amplitude with increasing slope and decreasing wavelength and amplitude with decreasing slope beyond qualitative interpretations. These unresolved issues motivate further exploration of the conditions under which bedforms develop and evolve in density stratified, turbulent flow.

5.4.2 Formation of Compound Bedforms

The compound bedforms described here record deposition and differential erosion within individual currents and between different currents. The surficial expression of the compound bedforms is a softly undulating dune field where dune wavelengths and amplitudes decrease with decreasing slope and with distance from source (Figure 6a). Compound bedforms are only found within 3 km from source and along the steepest slopes, suggesting formation under the highest energy conditions. Their location, combined with the regressive nature of bedforms in BS1, suggest supercritical flow conditions, as discussed above. However, while the large-scale nature and proximal distribution suggest deposition under high flow regimes, a clear interpretation for these bedforms is difficult because they consist of all three bedding sets.

Initiation of the BS1 regressive bedforms, which are at the core of the compound bedform, is interpreted to result from (1) internal waves at the interface between the concentrated basal region and dilute overriding cloud (e.g., Yih, 1980; Valentine, 1987), (2) the influence of surface roughness, which may have initiated deposition (e.g., Douillet et al., 2014), or (more

likely) a combination of the two (e.g., Hand, 1974). Since the deposits in BS2 and BS3 largely mantle the initial bedforms of BS1 proximal to source and along the steepest portions of the flank, the responsible currents also likely experienced supercritical flow conditions. It is possible that the initial dune expressions of BS1 triggered the development of internal waves (formed at the interface between the concentrated basal region and upper dilute, turbulent region of the current) of the same wavelength in the subsequent PDCs responsible for BS2 and BS3 (e.g., Valentine, 1987).

6.0 Conclusions

Numerous, column collapse driven PDCs were generated during the afternoon of the 18 May 1980 eruption. The PDCs were depositional within the crater breach, leaving massive deposits ~ 2 km north of the vent along the shallow slope (5-8°). These PDCs are interpreted as thick, concentrated and mobile currents with high internal pore pressure and negligible shear stress at the base due to rapid sedimentation. The majority of the currents were funneled down the stair steps (north portion of the breach). The lack of May 18 PDC deposits along the stairsteps suggests the currents generated during the afternoon of the eruption were largely non-depositional until they reached the shallow-dipping pumice plain.

The PDCs funneled down the eastern flank accelerated and thinned as they descended and spread across the steep slope. A shift toward supercritical conditions likely resulted in a high degree of ambient air entrainment (e.g., Bursik and Woods, 1996), while upward fluxes of air from substrate interstices (e.g., Chédeville, and Roche, 2014; 2015) and plunging breakers across rugged, irregular topography (Bareschino et al., 2008) further inflated the currents. This combination of processes allowed the currents to transition to a dominantly turbulent, dilute PDC with a thin concentrated basal region that moved via traction transport, as recorded by the

strongly stratified to cross-stratified proximal bedded deposits. The multiple erosive surfaces found within the flank deposits suggest that strong, turbulent sweeps intermittently crossed the depositional boundary, resulting in temporary erosive conditions. An increase in lithics consistent with the upstream substrate geology indicate that substrate erosion occurred in areas of high surface roughness, constrictions and on slopes $> 25^\circ$. Unfortunately, the influence of substrate erosion on fluidized PDC dynamics and runout distance remains unclear for our field case study.

Regressive bedforms are found in the 18 May 1980 PDC deposits along the entire flank, but are dominant along the most proximal and steepest portions of the volcanic slope ($>15^\circ$). Progressive bedforms have smaller wavelengths and are only found >2.5 km from source on slopes $<15^\circ$. Bedform wavelength and amplitude for both regressive and progressive beds increase with increasing slope and decrease with decreasing slope. The combination of these observations suggests that flow conditions were supercritical along the upper region of the slopes, and that flow velocity has a strong control on bedform wavelength.

Compound bedforms, comprised of multiple bed sets and PDCs, are only found on the steepest slopes and within 3 km from source. The core of the compound bedforms, which are in contact with the high surface roughness substrate, are dominated by regressive bedforms. Generation of the bedforms is thus interpreted as a combination of sedimentation initiated by surface roughness and upstream-migrating bedforms in supercritical currents. Later currents mantled the initial bedforms to develop the compound nature < 3 km from source, but downstream begin to fill in the troughs and thin over the crests. This depositional pattern suggests that the later currents were also initially supercritical, and that development of internal

waves, known to occur in stratified flow (e.g., Yih, 1980), was influenced by the undulating nature of the surface.

The stratified and cross-stratified upper flank deposits gradually transition to massive deposits away from source as the slope shallows toward and across the pumice plain, suggesting a decrease in flow energy resulted in an increase in thickness of the basal region to the point where the influence of overriding turbulence was suppressed. Thus the flow behavior shifted again to one where high internal pore pressure, low basal shear, hindered settling and laminar flow conditions dominated, as recorded by the massive nature of the eastern pumice plain deposits.

Our work highlights the problem that the behavior of fluidized granular flows along steep slopes, or as a function of changing slopes, remains poorly constrained. Evidence for changes in flow dynamics and criticality as a function of changing slope suggests that models assuming one flow regime or depth averaged techniques, while sufficient for capturing large scale dynamics and associated risk, may not capture important changes in PDC behavior due to variations in topography. Finally, no quantitative relationship exists that links dune bedform characteristics (wavelength, amplitude, and direction of accretion) to PDC emplacement conditions (velocity, depth, particle concentration, and vertical grain-size distribution), limiting our ability to interpret pyroclastic bedforms. Questions to pursue with future work include:

- What impact does a change in flow dynamics and flow criticality have on the overall runout distance of a given current?
- How do bedforms develop in density stratified flows, and how are bedform morphologies related to flow criticality?

To examine these questions, the authors suggest scaled experiments and numerical models to explore conditions that promote transitions in flow criticality (subcritical to

769 supercritical, or vice versa), and the consequences of these transitions on general flow
770 dynamics and runout distance. We also recommend developing experiments and numerical
771 models for dilute PDCs where density segregation and the development of internal waves
772 along the interface of the traction-dominated basal region and the overriding dilute, turbulent
773 portion of the current can be explored relative to the development of bedforms along the
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8.0 References

- Andrews, B. J., & Gardner, J. E., 2009. Turbulent dynamics of the 18 May 1980 Mount St. Helens eruption column. *Geology*, 37(10), 895-898.
- Auker, M.R., Sparks, R.S.J., Siebert, L., Crosweller, H.S., Ewert, J., 2013. A statistical analysis of the global historical volcanic fatalities record. *Journal of Applied Volcanology* 2(1):1-24
- Bagnold, R. A., 1954. Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear. In *Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences* . Vol. 225, No. 1160, p. 49-63.
- Bareschino, P., Lirer, L., Marzocchella, A., Petrosino, P., & Salatino, P., 2008. Self-fluidization of subaerial rapid granular flows. *Powder Technology*, 182(3), 323-333.
- Beeson, D.L., 1988. Proximal Flank Facies of the May 18, 1980 Ignimbrite: Mount St. Helens, Washington. MS Thesis, The University of Texas at Arlington, p. 1-206.
- Belousov, A. Belousova, M., 2001. Eruptive process, effects and deposits of the 1996 and the ancient basaltic phreatomagmatic eruptions in Karymskoye Lake, Kamchatka, Russia. *Spec. Pub. Intern. Assoc. Sediment.* 30:35-60.
- Bendaña, S., Brand, B.D., Self, S., Dufek, J., 2012. Effects of slope on the dynamics of pyroclastic density currents from May 18th, 1980 Mt. St. Helens eruption. Abstract V41B-2778 presented at 2010 Fall Meeting, AGU, San Francisco, Calif., 3-7 Dec.
- Bernard, J., Kelfoun, K., Le Pennec, J. L., Vargas, S. V., 2014. Pyroclastic flow erosion and bulking processes: comparing field-based vs. modeling results at Tungurahua volcano, Ecuador. *Bulletin of Volcanology*, 76(9), 1-16.
- Blatt, H., Middleton, G., Murray, R., 1980. *Origin of sedimentary rocks*: Prince-Hall. Inc., New Jersey, 782.
- Brand, B. D., Mackaman-Lofland, C., Pollock, N. M., Bendaña, S., Dawson, B., Wichgers, P., 2014a. Dynamics of pyroclastic density currents: Conditions that promote substrate erosion and self-channelization—Mount St Helens, Washington (USA). *Journal of Volcanology and Geothermal Research*, 276, 189-214.
- Brand B.D., Gravely, D., Clarke, A.B., Nemeth, K., 2014b. A combined field and numerical approach to understanding dilute pyroclastic density current dynamics and hazard potential: Auckland Volcanic Field, New Zealand. *Journal of Volcanology and Geothermal Research*, 276, 215-232.
- Brand, B.D., Clarke, A.B., 2009. The architecture, eruptive history, and evolution of the Table Rock Complex, Oregon (USA): from a Surtseyan to an energetic maar eruption. *Journal of Volcanology and Geothermal Research* 180: 203-224. <http://dx.doi.org/10.1016>

- Brand, B.D., Semken, S., Clarke, A.B., 2009. Eruptive Conditions and Depositional Processes of Narbona Pass Maar Volcano, Navajo Volcanic Field, Navajo Nation, New Mexico (USA). *Bulletin of Volcanology* 71:49-77, doi 10.1007/s00445-008-0209-y.
- Brand, B.D., White, C.M., 2007. Origin and stratigraphy of phreatomagmatic deposits at the Pleistocene Sinker Butte Volcano, Western Snake River Plain, Idaho. *Journal of Volcanology and Geothermal Research* 160: 319-339.
- Brand B.D., Clarke, A.B., 2012. An unusually energetic basaltic phreatomagmatic eruption: Using deposit characteristics to constrain dilute pyroclastic density current dynamics. *Journal of Volcanology and Geothermal Research*. 243–244: 81–90.
- Branney, M. J., Kokelaar, P., 1997. Giant bed from a sustained catastrophic density current flowing over topography: Acatlan ignimbrite, Mexico. *Geology*, 25(2): 115-118.
- Branney, M., Kokelaar, B. P., 2002. Sedimentation of ignimbrites from pyroclastic density currents. *Memoir of the Geological Society of London*, 27, 150.
- Bursik, M.I., Woods, A.W., 1996. The dynamics and thermodynamics of large ash flows. *Bull. Volcanol.* 58:175-193
- Calder, E. S., Sparks, R. S. J., Gardeweg, M. C., 2000. Erosion, transport and segregation of pumice and lithic clasts in pyroclastic flows inferred from ignimbrite at Lascar Volcano, Chile. *Journal of Volcanology and Geothermal Research*, 104(1), 201-235.
- Carey, S., Sigurdsson, H., Gardner, J.E., Criswell, W., 1990. Variations in column height and magma discharge during the May 18, 1980 eruption of Mount St. Helens. *J. Volcanol. Geotherm. Res.* 43, 99-112.
- Chédeville, C., Roche, O., 2014. Autofluidization of pyroclastic flows propagating on rough substrates as shown by laboratory experiments. *Journal of Geophysical Research: Solid Earth*, 119(3), 1764-1776.
- Chédeville, C., Roche, O., 2015. Influence of slope angle on pore pressure generation and kinematics of pyroclastic flows: insights from laboratory experiments. *Bulletin of Volcanology*, 77(11), 1-13.
- Christiansen, R. L., Peterson, D. W., 1981. Chronology of the 1980 eruptive activity. *US Geol. Surv. Prof. Pap*, 1250, 17-30.
- Cole, P. D., Scarpati, C., 1993. A facies interpretation of the eruption and emplacement mechanisms of the upper part of the Neapolitan Yellow Tuff, Campi Flegrei, southern Italy. *Bulletin of volcanology*, 55(5), 311-326.

- Cole, P. D., Fernandez, E., Duarte, E., Duncan, A. M., 2005. Explosive activity and generation mechanisms of pyroclastic flows at Arenal volcano, Costa Rica between 1987 and 2001. *Bulletin of Volcanology*, 67(8), 695-716.
- Criswell, C. W., 1987. Chronology and pyroclastic stratigraphy of the May 18, 1980, eruption of Mount St. Helens, Washington. *Journal of Geophysical Research: Solid Earth* (1978–2012), 92(B10), 10237-10266.
- Crowe, B.M., Fisher, R.V., 1973. Sedimentary structures in base-surge deposits with special reference to cross-bedding, Ubehebe Craters, Death Valley, California. *Geol. Soc. Am. Bull.* 84, 663-682.
- Denlinger, R. P., Iverson, R. M., 2001. Flow of variably fluidized granular masses across three-dimensional terrain: 2. Numerical predictions and experimental tests. *Journal of Geophysical Research: Solid Earth* (1978–2012), 106(B1), 553-566.
- Douillet, G. A., Pacheco, D. A., Kueppers, U., Letort, J., Tsang-Hin-Sun, È., Bustillos, J., Hall, M., Ramón, P., Dingwell, D. B., 2013. Dune bedforms produced by dilute pyroclastic density currents from the August 2006 eruption of Tungurahua volcano, Ecuador. *Bulletin of Volcanology*, 75(11), 1-20.
- Douillet, G. A., Rasmussen, K. R., Kueppers, U., Castro, D. L., Merrison, J. P., Iversen, J. J., & Dingwell, D. B., 2014. Saltation threshold for pyroclasts at various bed slopes: Wind tunnel measurements. *Journal of Volcanology and Geothermal Research*, 278, 14-24.
- Druitt, T.H., 1995. Settling behaviour of concentrated, poorly sorted dispersions and some volcanological applications. *Journal of Volcanology and Geothermal Research* 65: 27-39.
- Druitt, T.H., 1992. Emplacement of the 18 May, 1980 lateral blast deposit east-northeast of Mount St. Helens, Washington. *Bulletin of Volcanology*, 54: 554-572.
- Druitt, T.H., Avard, G., Bruni, G., Lettieri, P. and Maez, F., 2007. Gas retention in fine-grained pyroclastic flow materials at high temperatures. *Bulletin of Volcanology* DOI 10.1007/s00445-007-0116-7.
- Dufek, J., Manga, M., 2008. In situ production of ash in pyroclastic flows. *Journal of Geophysical Research: Solid Earth* (1978–2012), 113(B9).
- Eisenhart, C., 1935. A test for the significance of lithological variations. *Journal of Sedimentary Petrology* 5:137-145.
- Farin, M., Mangeney, A., Roche, O., 2014. Fundamental changes of granular flow dynamics, deposition, and erosion processes at high slope angles: Insights from laboratory experiments. *Journal of Geophysical Research: Earth Surface*, 119(3), 504-532.

- Fedele, J.J., Hoyal, D.C.J.D., and Sheets, B.A., 2007. Experiments on bedforms created by density currents: *Eos Trans. AGU*, v. 88, p. Fall meeting supplement, abstract H41F-0839.
- Félix, G., Thomas, N., 2004. Relation between dry granular flow regimes and morphology of deposits: formation of levées in pyroclastic deposits. *Earth and Planetary Science Letters*, 221(1), 197-213.
- Fisher, R. V., 1990. Transport and deposition of a pyroclastic surge across an area of high relief: the 18 May 1980 eruption of Mount St. Helens, Washington. *Geological Society of America Bulletin*, 102(8), 1038-1054.
- Fisher, R.V., Schmincke, H-U., 1984. *Pyroclastic Rocks*. Springer-Verlag, Berlin.
- Fujii, T., Nakada, S., 1999. The 15 September 1991 pyroclastic flows at Unzen Volcano (Japan): a flow model for associated ash-cloud surges. *Journal of Volcanology and Geothermal Research*, 89(1), 159-172.
- Gray, T. E., Alexander, J., Leeder, M. R., 2005. Quantifying velocity and turbulence structure in depositing sustained turbidity currents across breaks in slope. *Sedimentology*, 52(3), 467-488.
- Gurioli, L., Cioni, R., Sbrana, A., & Zanella, E., 2002. Transport and deposition of pyroclastic density currents over an inhabited area: the deposits of the AD 79 eruption of Vesuvius at Herculaneum, Italy. *Sedimentology*, 49(5), 929-953.
- Hand, B.M., 1974. Supercritical flow in density currents. *J Sediment Petrol* 44: 637-648.
- Iverson, R. M., Denlinger, R. P., 2001. Flow of variably fluidized granular masses across three-dimensional terrain: 1. Coulomb mixture theory. *J. Geophys. Res*, 106(B1), 537-552.
- Jopling, A.V, Richardson, E.V., 1966. Backset Bedding Developed in Shooting Flow in Laboratory Experiments: NOTES. *SEPM Journal of Sedimentary Research*. Vol. 36
- Kidd, R.B., Lucchi, R.G., Gee, M., Woodside, J.M., 1998. Sedimentary processes in the Stromboli Canyon and Marsili Basin, SE Tyrrhenian Sea: results from sidescan sonar surveys. *Geo-Mar. Lett.* 18, 146-154.
- Kieffer, S. W. (1981) Fluid dynamics of the May 18 blast at Mount St. Helens. *US Geol. Surv. Prof. Pap.* 1250, 379-400.
- Kneller, B., Buckee, C., 2000. The structure and fluid mechanics of turbidity currents; a review of some recent studies and their geological implications *Sedimentology*, 47: 62-94.
- Lowe, D. R., 1982. Sediment gravity flows: II Depositional models with special reference to the deposits of high-density turbidity currents. *Journal of Sedimentary Research*, 52(1).

- Macias, J. L., Espindola, J. M., Bursik, M., Sheridan, M. F., 1998. Development of lithic-breccias in the 1982 pyroclastic flow deposits of El Chichon volcano, Mexico. *Journal of volcanology and geothermal research*, 83(3), 173-196.
- Mackaman-Lofland, C., Brand, B. D., Taddeucci, J., & Wohletz, K., 2014. Sequential fragmentation/transport theory, pyroclast size–density relationships, and the emplacement dynamics of pyroclastic density currents – A case study on the Mt. St. Helens (USA) 1980 eruption. *Journal of Volcanology and Geothermal Research*, 275, 1-13.
- Manga, M., Patel, A., Dufek, J., 2011. Rounding of pumice clasts during transport: field measurements and laboratory studies. *Bulletin of Volcanology*, 73(3), 321-333.
- Mangeney, A., Roche, O., Hungr, O., Mangold, N., Faccanoni, G., Lucas, A., 2010. Erosion and mobility in granular collapse over sloping beds. *Journal of Geophysical Research: Earth Surface* (2003–2012), 115(F3).
- Moore, J.G., 1967. Base surge in recent volcanic eruptions, *Bull. Volcanol.* 30, 337-363.
- Morris, S. A., Kenyon, N. H., Limonov, A. F., Alexander, J., 1998. Downstream changes of large-scale bedforms in turbidites around the Valencia channel mouth, northwest Mediterranean: implications for palaeoflow reconstruction. *Sedimentology*, 45(2), 365-378.
- Mulder, T., Alexander, J., 2001. The physical character of subaqueous sedimentary density flows and their deposits. *Sedimentology*, 48(2), 269-299.
- Pittari, A., Cas, R. A. F., Edgar, C. J., Nichols, H. J., Wolff, J. A., Marti, J., 2006. The influence of palaeotopography on facies architecture and pyroclastic flow processes of a lithic-rich ignimbrite in a high gradient setting: the Abrigo Ignimbrite, Tenerife, Canary Islands. *Journal of Volcanology and Geothermal Research*, 152(3), 273-315.
- Pollock, N., 2013. Field evidence for substrate entrainment by pyroclastic density currents and its effect on downstream dynamics at Mount St Helens, Washington (USA). University of Washington Master's Thesis
- Pollock, N., Brand, B.D., 2015. Field evidence for substrate entrainment by pyroclastic density currents and its effect on downstream dynamics at Mount St Helens (USA). International Union for Geodesy and Geophysics; Abstract VS06p-384 in Section VS06 – Benchmarking Pyroclastic Density Current Models: Code Inter-Comparison and Field Validation. Prague, Czech Republic. June 28, 2015
- Postma, G., 1986. Classification for sediment gravity-flow deposits based on flow conditions during sedimentation. *Geology*, 14(4), 291-294.
- Roche, O., Gilbertson, M.A., Phillips, J.C., Sparks, R.S.J., 2004. Experimental study of gas-fluidized granular flows with implications for pyroclastic flow emplacement, *J. Geophys. Res.*, 109, B10201, doi:10.1029/2003JB002916

- 1009
1010 Roche, O., Montserrat, S., Niño, Y., Tamburrino, A., 2008. Experimental observations of water-
1011 like behavior of initially fluidized, dam break granular flows and their relevance for the
1012 propagation of ash-rich pyroclastic flows. *J. Geophys. Res.* 113, B12203.
1013 doi:10.1029/2008JB005664
1014
1015 Roche, O., Montserrat, S., Niño, Y., Tamburrino, A., 2010. Pore fluid pressure and internal
1016 kinematics of gravitational laboratory air-particle flows: insights into the emplacement dynamics
1017 of pyroclastic flows. *J. Geophys. Res.* 115, B09206. doi:10.1029/2009JB007133.
1018
1019 Roche, R., 2015. Nature and velocity of pyroclastic density currents inferred from models of
1020 entrainment of substrate lithic clasts. *Earth and Planetary Science Letters* 418: 115-125.
1021
1022 Roche, O., Niño, Y., Mangeney, A., Brand, B., Pollock, N., & Valentine, G. A., 2013. Dynamic
1023 pore-pressure variations induce substrate erosion by pyroclastic flows. *Geology* 41(10), 1107-
1024 1110.
1025
1026 Rowley, P. D., MacLeod, N. S., Kuntz, M. A., & Kaplan, A. M., 1985. Proximal bedded deposits
1027 related to pyroclastic flows of May 18, 1980, Mount St. Helens, Washington. *Geological Society*
1028 *of America Bulletin* (0016-7606), 96 (11), p. 1373.
1029
1030 Schmincke, H., Fisher, R.V., Waters, A.C., 1973. Antidune and chute and pool structures in the
1031 base surge deposits of the Laacher See area, Germany. *Sedimentology*, 20:553-574.
1032
1033 Sigurdsson, H., Carey, S. N., Fisher, R. V., 1987. The 1982 eruptions of El Chichon volcano,
1034 Mexico (3): Physical properties of pyroclastic surges. *Bulletin of volcanology*, 49(2), 467-488.
1035
1036 Sohn, Y. K., 1997. On traction-carpet sedimentation. *Journal of Sedimentary Research*, 67(3).
1037
1038 Sparks, R. S. J., Moore, J. G., & Rice, C. J., 1986. The initial giant umbrella cloud of the May
1039 18th, 1980, explosive eruption of Mount St. Helens. *Journal of volcanology and geothermal*
1040 *research*, 28(3), 257-274.
1041
1042 Spinewine, B., Sequeiros, O. E., Garcia, M. H., Beaubouef, R. T., Sun, T., Savoye, B., & Parker,
1043 G. (2009). Experiments on wedge-shaped deep sea sedimentary deposits in minibasins and/or on
1044 channel levees emplaced by turbidity currents. Part II. Morphodynamic evolution of the wedge
1045 and of the associated bedforms. *Journal of Sedimentary Research*, 79(8), 608-628.
1046
1047 Sulpizio, R., Dellino, P., 2008. Sedimentology, depositional mechanisms and pulsating
1048 behaviour of pyroclastic density currents. *Developments in Volcanology*, 10, 57-96.
1049
1050 Sulpizio, R., De Rosa, R., Donato, P., 2008. The influence of variable topography on the
1051 depositional behaviour of pyroclastic density currents: the examples of the Upper Pollara
1052 eruption (Salina Island, southern Italy). *Journal of Volcanology and Geothermal Research*,
1053 175(3), 367-385.
1054

- Walker, G.P.L., 1971. Grain-size characteristics of pyroclastic deposits. *Journal of Geology* 79: 696-714.
- Walker, G. P., 1984. Characteristics of dune-bedded pyroclastic surge bedsets. *Journal of volcanology and geothermal research*, 20(3), 281-296.
- Wohletz, K.H., 1998. Pyroclastic Surges and Compressible Two-Phase Flow. In *From Magma to Tephra*, edited by A. Freundt and M. Rosi, pp25-51, Elsevier, Amsterdam.
- Wohletz, K.H., Sheridan, M.F., 1979. Model of Pyroclastic Surge. *Geological Society of America Paper* 180
- Wilson, C. J. N., Walker, G. P. L., 1982. Ignimbrite depositional facies: the anatomy of a pyroclastic flow. *Journal of the Geological Society*, 139(5), 581-592.
- Wynn, R.B., Masson, D.G., Stow, D.A.V., Weaver, P.P.E., 2000. Turbidity current sediment waves on the submarine slopes of the western Canary Islands. *Mar. Geol.* 163, 185-198.
- Wynn, R.B., Piper, D.J.W., Gee, M.J.R., 2002. Generation and migration of coarse-grained sediment waves in turbidity current channels and channel-lobe transition zones. *Marine Geology* 192: 59-78.
- Yih C-S., 1980. Stratified flows. *Academic Press New York*, pp 103 – 141.

Figures Captions:

Figure 1: (a) Digital Elevation Map of Mt. St. Helens showing distribution of 18 May 1980, afternoon PDC deposits and debris avalanche hummocks (modified from Rowley et al., 1985). The legend is used in both Figures 1 and 2. The dashed lines are only present in Figure 2; (b) Sequence of events on 18 May 1980, highlighting the afternoon PDCs as the focus of the work presented here. ‘Ash flow’, synonymous with PDC, is terminology used to describe the phases of the eruption in Rowley et al. (1985); (c) Photograph of Plinian eruption column collapsing to the north around 15:50 Pacific Standard Time (photo credit: Joseph G. Rosenbaum, provided by the Cascade Volcano Observatory, United States Geologic Survey).

Figure 2: (a) Vertically exaggerated profile of eastern flank demonstrating the increase in resisting force relative to driving force with decreasing slope (assuming mass stays the same); m = mass; g = gravitational acceleration; θ = slope (after Sulpizio and Dellino, 2008). Point 0 meters on the horizontal axis represents the vent. (b) Massive PDC deposits exposed in the crater of MSH (photo taken by Steve Self in early 1980s; thickness ~ 3 m). (c) Proximal bedded deposits found high on the steep flanks of $\sim 25^\circ$; (d) transitional deposits found on moderate slopes of $\sim 12^\circ$. (e) Massive pumice plain deposits found beyond the break in slope (slopes $\sim 5^\circ$). (f) Google Earth image with modified Rowley et al. (1985) map overlay. The dashed lines mark the modern day exposures (as of 2015).

Figure 3: (a) – (e) Granulometry data comparing pumice plain units to proximal bedded deposits bed sets: Only the eastern pumice plain deposits are included in plot (d) to highlight the downstream fines content. Solid lines in (e) represent the Walker et al. (1984) pyroclastic flow field, whereas the dotted lines represent their pyroclastic surge field; (f) componentry data for pumice plain units and proximal bedded deposits bed sets. P = pumice, C = crystals and L = lithics.

Figure 4: Examples of the proximal bedded deposit bed-sets (BS). Note the clear truncation surface between each. Paleo-topography was rarely exposed. (a) Flow direction right to left, 1 m shovel for scale; (b) flow direction left to right; BS1 is ~ 8 m thick for scale; (c) flow direction right to left, person for scale.

Figure 5: (a) Photo taken from the beginning of proximal bedded deposits, looking downslope toward the north-northwest (Spirit Lake in the background). Most of the deposits in this area have been eroded, except for the deposits on right side of photo. Notice high degree of surface roughness and availability of granular material on the surface. (b) Photo taken standing on the PBD looking upslope into a slot canyon comprised of Castle Creek basalt and basaltic andesite lava flows. PBD are found within and overtop the slot canyon, suggesting the canyon was present during the 18 May 1980 eruption. Notice the high degree of surface roughness. (c) Oblique aerial photo of undulating compound bedforms of the PBD. Scale unknown (photo taken in early 1980s), but the wavelength should be between 20-50 m. (d) Cross-section through a compound bedform showing the BS1 core and draping nature of BS2 and BS3, although erosional truncations are common between bed sets.

Figure 6: (a) - (c) Plots of depositional slope, amplitude and wavelength with distance from vent for (a) compound bedforms, (b) progressive bedforms, and (c) regressive bedforms; (d) plot of

stoss side accumulations and number of stoss side packages within regressive bedforms as a function of slope. Form length (horizontal distance between first stoss side to last lee side expression) was also recorded because consecutive bedforms were not always exposed to measure wavelength.

Figure 7: (a) Progressive bedform with thicker strata on the lee side (flow direction right to left), with 1 m scale bar (1980s photo provided by S. Self); (b) – (e) regressive bedforms with prominent stoss side accretion and a series of steeply-dipping backset beds (flow direction right to left). Scale bar in (e) is 60 cm.

Figure 8: (a) Photograph of PDC deposits on mid-to-upper flank of Mount St. Helens. Flow direction from right to left; (b) Sketch of outcrop outlining bedding features and blocks. Black boxes represent sample locations for data in (c) and (d); (c) pie charts comparing lithic componentry and p-values of PDC deposits to fall deposits; (d) diameters of five largest lithic clasts per sample.

Table Captions:

Table 1: Summary of phases during the 18 May 1980 eruption with relative timing and duration of the PDCs that flowed north through the breach (modified from Criswell, 1987). Co-ignimbrite ash cloud height from Sparks et al. (1986). Eruption column height and mass flux estimates from Carey et al. (1990) and Andrews and Gardner (2009). Pumice plain PDC units and flank deposits are correlated with the various phases of eruption (see Brand et al., 2014a for unit correlation details).

Table 2: Terminology of Bedforms. Bedform classification modified from Beeson (1988)

Table 3: Granulometry data for crater, flank and eastern pumice plain (Avg. = Average).

Table 4: Range of dimensions for PBD bedforms (modified from Beeson, 1988)

Table 1: Summary of phases during the May 18, 1980 eruption with relative timing and duration the PDCs that flowed north through the breach (modified from Criswell, 1987). Co-ignimbrite ash cloud height from Sparks et al. (1986). Eruption column height and mass flux estimates from Carey et al. (1990) and Andrews and Gardner (2009). Pumice plain PDC units and flank deposits are correlated with the various phases of eruption (see Brand et al., 2014 for unit correlation details).

Phase	Time (hours)	Eruption Style	Column Height; Mass Flux (kg/s)	Timing and Duration of North Breach Funneled PDCs	Stratigraphic PDC Deposit Unit
I	0832-0900	Massive landslide followed by lateral blast	Co-ignimbrite plume reached 30 km in height	Lateral blast reached maximum distance in 5-7 minutes	Blast deposits (only found beyond 8.5 km from source, below Johnson Ridge)
II	0900-1215	Early Plinian	15-18 km; 6.3×10^6 - 1.3×10^7 kg/s	N/A	N/A
III	1215-1500	Early Ash Flow	13-16 km; 3.9×10^6 kg/s	1217 – PDC “poured through the breach”, generating 4-7 km elongated ash cloud; 1330-1400 – episodic PDCs emplaced in the upper Toutle Valley (unclear if these reached the pumice plain); 1430-1500 size of flows increased	Units I and II; BS1
IV	1500-1715	Climactic phase	13-16 km; 4.0×10^7 kg/s	1500 – “numerous, large” PDCs were observed to spill over the crater rim, engulfing the volcano, reaching Spirit Lake. PDC flux to the north abated around 1635. PDC activity that began to wane at 1605, ceasing at 1635.	Unit III and possibly Unit IV; BS2 and possibly BS3
V	1715-1815	Late Ash Flow	Not reported	Short-lived increase in activity during the waning phase produced a PDC pulse that flowed into the upper Toutle valley at 1745, reaching Spirit Lake at 1747. A second PDC pulse to the north occurred at 1810.	Possibly Unit IV and BS3
VI	1815-May 19 th , 1980	Waning and Post eruption phase	N/A	N/A	N/A

Table 2: Terminology of Bedforms. Bedform classification modified from Beeson (1988)

<i>Plane Parallel Stratification</i>				
Depositional Feature and Description	Median (ϕ)	Sorting (ϕ)	F2/F1	F2/F1 Average
Coarse-grained stratification Thin to medium bedded, plane parallel strata that alternate from coarse- and fine-grained due to the proportion of coarse pumice present. Reverse grading is common. Fine-grained beds are pumice-rich and moderately well-sorted. The coarser beds are fines poor, poorly-sorted, and clast-supported. Coarse-grained strata are most commonly found underlying dune bedforms or comprising the lee-side and cores of dune bedforms. Interpretation: Deposited under traction-transport dominated flow boundary, as evidenced by stratification, sorting and reverse grading.	<i>Fine-grained beds:</i> 1.3 to 2.7	<i>Fine-grained beds:</i> 1.52 to 1.78	<i>Fine-grained beds:</i> 0.219 to 0.398	<i>Fine-grained beds:</i> 0.308
	<i>Coarse-grained beds:</i> 1.13 to 0.74	<i>Coarse-grained beds:</i> 2.3 to 2.93	<i>Coarse-grained beds:</i> 0.106 to 0.226	<i>Coarse-grained beds:</i> 0.155
Fine-grained stratification Similar to coarse-grained stratification except thinner-bedded and dominated by finer-grained, plane parallel to wavy strata with minor grain size fluctuations between layers. Contacts between strata are sharp to erosional. Small-scale cross-stratification is also present (see Subsidiary Type I Bedforms described below). These are also most commonly found underlying dune bedforms or comprising the lee-side and cores of dune bedforms. Interpretation: Deposited under traction-transport dominated flow boundary, as evidenced by stratification, sorting and reverse grading.	0.95 to 2.6	1.69	0.203 to 0.238	0.227
Massive (unstratified) Beds				
Depositional Feature and Description	Median (ϕ)	Sorting (ϕ)	F2/F1	F2/F1 Average
Massive Beds Medium (10 – 30 cm) to thick (>100 cm), generally poorly-sorted beds that display little to no internal structures (massive) except for faint stratification due to discontinuous planar trains of pumice or lithic lapilli. Grading, if present, is normal with respect to lithic clasts, and coarse-grained and inverse with regards to pumice clasts. No pumice accumulations are found. The basal contact is sharp to erosional, as indicated by the common angular truncation of underlying strata. Unstratified beds are most common in deposits found near the base of the flank and into the pumice plain, but are also found in the upper flank deposits (slopes from 6°- 28°) within or between bedforms.	-0.39 to 1.97	1.46 to 2.90	0.130 to 0.416	0.189

Interpretation: Deposited under rapid sedimentation conditions where traction-transport was suppressed. Faint stratification and trains of clasts indicate unsteadiness in the current and fluctuations between negligible basal shear stress to relatively higher basal and internal shear stress, likely transitioning toward traction-transport conditions.				
Pumice-rich beds Clast-supported beds of rounded pumice lapilli and some lithic lapilli. Pumice-rich beds are lenticular in nature when found within larger bedforms or interdune regions and generally more continuous when found at the basal portions of bed sets. They are also common on the upstream side of steeply dipping (up to 30°) truncation surfaces and backset beds (e.g. Figure 6e). Reverse grading and faint stratification are common. Beds are typically poorly-sorted, somewhat fines-poor and polymodal to unimodal with the coarsest mode representing the increased pumice content. Interpretation: Pumice roundness, reverse grading and stratification suggest pumice-rich zones transported as a traction-carpet. Pumice lenses are typically thought to develop as pumice segregates from the main flow to be transported along a region of neutral buoyancy. The commonality of these lenses along the steep slopes as part of large bedforms suggests low density conditions near the depositional boundary, further supporting the semi-dilute nature of the current.	0.25 to -3.94	2.07 to 3.57	0.127 to 0.146	0.137
Large-scale Dune Bedforms				
Depositional Feature and Description	Median (ϕ)	Sorting (ϕ)	F2/F1	F2/F1 Average
Progressive Dune Bedforms: Low aspect ratio, somewhat symmetric bedforms with well-developed lee-side bedding and less extensive, shallow-dipping stoss-side bedding. Lee-side strata dip $< 10^\circ$ relative to the depositional slope and are comprised of thin to medium bedded, parallel, sub-parallel or curved strata that level to become parallel to underlying planar beds in the downstream direction. Stoss-side beds are poorly-developed and often coarser and less well-sorted relative to the lee-side strata. Interpretation: Deposition under turbulent, traction-transport conditions. The progressive nature suggests downstream	-0.92 to 1.65	1.66 -3.14	0.157 to 0.270	0.214

migration under low flow regime (Froude #<1) conditions.				
Regressive Dune Bedforms: Symmetrical bedforms with well-developed, low-angle stoss-side bedding. The core of regressive bedforms consist of planar to wavy bedding that dips parallel or up to 10° to the depositional slope. The cores are truncated on the stoss-side and draped with thicker stoss-side bedding relative to lee-side bedding. A series of well-developed concordant beds drape the upstream truncation surface with accumulation lengths from 2 to 15 m. The beds are heterogenous and contain fine-grained strata, coarse-grained strata and pumice-rich beds. Unconformities are common in the stoss-side accumulations. Interpretation: Deposition under turbulent, traction-transport conditions. The symmetric to regressive natures suggests upstream migration and deposition influenced by standing waves under high flow regime (Froude #>1) conditions.	-1.57 to 3.26	1.43 to 3.01	0.073 to 0.558	0.199
Steeply-dipping Backset Beds: Regressive bedforms with well-developed, high-angle stoss-side bedding. Cores of regressive dune bedforms with steeply dipping backset bedforms are truncated on the stoss-side by steeply up-flow dipping truncation surface (from 30 up to 68°) and have a sigmoidal shape. Upflow-stacking beds with a set of sigmoidal-shape truncations are often found upstream from the first truncation surface with accumulation lengths up to 15 m. Interpretation: These backset beds are interpreted to be the depositional representation of a chute and pool features, where a current abruptly transitions from supercritical to subcritical conditions (i.e., hydraulic jump; Jopling and Richardson, 1966). While rarely preserved in the fluvial record, chute and pool features are common in the deposits of dilute PDCs (e.g., Schmincke, 1973).	0.5 to 1.6	1.09 to 1.8	0.0279 to 0.393	0.185
Subsidiary Dune Bedforms – Smaller-scale features than PDB and RDB with form heights ~0.4 m and wavelengths of 0.3 – 3.5 m				
Depositional Feature and Description	Median (ϕ)	Sorting (ϕ)	F2/F1	F2/F1 Average
Subsidiary Type 1 Dune Bedforms (SUB1) SUB1 are most often found on the lee side of larger scale dune bedforms within sequences of thin to medium bedded, plane parallel to wavy beds. SUB1 are found on depositional slopes of 16°- 20°. Dunes consist of low-angle, fine-	0.95 to 1.97	1.42 to 2.13	0.104 to 0.282	0.212

grained strata (described above details) that most often dip in the downflow direction before leveling with the lower bounding surface. Bedforms can be progressive, regressive or stationary. They have form heights from <0.2 – 0.4 m and most often occur on slopes from 15-22° as isolated or multiple bedforms. <u>Interpretation:</u> Type I are described as secondary features that form at the “limbs” of large-scale bedforms or within sequences of parallel bedding and can be result in fluctuations in flow velocity to a lower speed (Beeson, 1988).				
Subsidiary Type 2 Dune Bedforms (SUB2): Bedforms that occur as thinly laminated (1-5 cm) to unstratified dune-like features or discontinuous packages of inclined lamina. SUB2 are found on depositional slopes of 6°- 23° and occur as solitary and consecutive sets and sometimes in between or capped by medium to thick-bedded deposits. Solitary form heights range from 0.1 to 1 m. Consecutive bedforms has wavelengths of <3.5 m and wave heights from 0.2 - 0.3 m. SUB2 have bimodal to unimodal grain sizes and is finer-grained and better-sorted than SUB1 and as such contains a high fines content than any other flank bedding type. <u>Interpretation:</u> Unlike Type I subsidiary bedforms, Type II bedforms are not always subordinate features because the dimensions (amplitude and wavelength) approach those of regressive and progressive bedforms. The finer grain size and thinner laminae suggests deposition from even lower particle concentration dilute PDCs.	1.00 to 3.13	1.37 to 2.29	0.113 to 0.532	0.276

Table 3: Granulometry data for crater, flank and eastern pumice plain (Avg. = Average).

	Md ϕ	Sorting (σ_ϕ)	F2/F1
Crater (n=2)	0.63 to 2.33 ϕ Avg. 1.83 ϕ	1.76 to 3.07 ϕ Avg. 2.11 ϕ	0.208 to 0.222 Avg. 0.215
Flank (n=85)	-3.94 to 3.13 ϕ Avg. 1.12 ϕ	1.09 to 3.57 ϕ Avg. 1.98 ϕ	0.0279 to 0.532 Avg. 0.193
BS1 (n=53)	-3.94 to 3.13 ϕ Avg. 1.09 ϕ	1.09 to 3.57 ϕ Avg. 1.99 ϕ	0.0279 to 0.532 Avg. 0.211
BS2 (n=25)	-0.39 to 2.07 ϕ Avg. 0.90 ϕ	1.44 to 2.90 ϕ Avg. 2.04 ϕ	0.104 to 0.273 Avg. 0.164
BS3 (n=7)	0.79 to 3.26 ϕ Avg. 1.87 ϕ	1.43 to 2.13 ϕ Avg. 1.69 ϕ	0.111 to 0.348 Avg. 0.149
Eastern Pumice Plain (n=12)	-4.20 to 2.12 ϕ Avg. 0.46 ϕ	1.88 to 3.46 ϕ Avg. 2.86 ϕ	0.0665 to 0.173 Avg. 0.137

Table 4: Range of dimensions for PBD bedforms (modified from Beeson, 1988)

Bedform Type	Compound Bedforms	Regressive Bedforms	Progressive Bedforms
Form Height (meters)	2 to 13 m	0.7 to 5 m	0.4 to 2.5 m
Wavelength (meters)	25 to 50 m	1.9 to 8 m	3.0 to 7.6 m
Form Length (meters)	10 to 40 m	2 to 13.5 m	2 to 13.75 m
Distance from Vent (kilometers)	2.23 to 2.97 km	2.29 to 3.27 km	2.53 to 3.27 km
Depositional Slope	12° to 28°	10° to 28°	12° to 16°

Figure 1

	Time (hours)	Eruption Phase/Style
I	0832-0900	Massive landslide followed by lateral blast
II	0900-1215	Early Plinian
III	1215-1500	Early Ash Flow
IV	1500-1715	Middle Ash Flow (Climactic phase)
V	1715-1815	Late Ash Flow
VI	1815-May 19 th , 1980	Post eruption phase

(b)

Figure 2

Figure 3

Figure 4

Figure 5:

Figure 6

Figure 7

Figure 8

(d) Five Largest Clasts (cm)

BS2 Sample 2	BS2 Sample 1
29 x 16 x 12.5	16 x 10 x 6.5
16.5 x 11 x 8	11 x 8 x 6
16 x 9 x 7	10 x 6.2 x 4
13 x 9 x 8	6.5 x 5.5 x 5
11 x 11 x 6	8 x 4.5 x 2