Interpreting flash flood palaeoflow parameters from antidunes and gravel lenses: An example from Montserrat, West Indies

MELANIE J. FROUDE*†, JAN ALEXANDER*, JENNI BARCLAY* and PAUL COLE‡
*School of Environmental Sciences, University of East Anglia, Norwich Research Park, Norwich, Norfolk NR4 7TJ, UK (E-mail: m.froude@sheffield.ac.uk)
†Department of Geography, University of Sheffield, 9 Northumberland Road, Sheffield S10, UK
‡Earth Sciences, Plymouth University, Drake Circus, Plymouth, Devon PL4 8A, UK

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ABSTRACT

The wavelength of stationary water-surface waves and their associated antidune bedforms are related to the mean velocity and depth of formative flow. In past published sand-bed flume experiments, it was found that lens structures were preserved during antidune growth and change, and the dimension of the lenses was empirically related to antidune wavelength, and thus could be used to estimate flow velocity and depth. This study is the first to compare observations of formative flow conditions and resulting sedimentary structures in a natural setting, testing the previously published relationship at a field-scale. Trains of stationary and upstream migrating water-surface waves were prevalent during the flash flood in October 2012 in the Belham Valley, Montserrat, West Indies. Wave positions and wavelengths were assessed at 900 sec intervals through the daylight hours of the event within a monitored reach. The wave data indicate flow depths up to 1·3 m and velocity up to 3·6 m sec⁻¹. Sedimentary structures formed by antidune growth and change were preserved in the event deposit. These structures include lenses of clast-supported gravel and massive sand, with varying internal architecture. The lenses and associated low-angle strata are comparable to sand-bed structures formed from stationary and upstream migrating waves in flume experiments, confirming the diagnostic value of these structures. Using mean lens length in the event deposit underestimated peak flow conditions during the flood and implied that the lenses were preserved during waning flow.

Keywords Antidunes, flash flood, gravel lenses, lahar, palaeoflow.

INTRODUCTION

Sedimentary structures associated with supercritical flow conditions are very poorly documented in comparison with structures produced in subcritical flow conditions and, consequently, are still rarely recognized in the rock record. (Maejima et al., 2009). It is very difficult, both technically and because of inherent hazards, to observe sedimentary processes in natural supercritical flows above mobile beds directly because of the fast, turbulent, sediment-laden water. Sedimentary structures resulting from formation, migration and truncation of antidunes have been documented in sand-bed flumes with and without bed aggradation (e.g. Alexander et al., 2001; Yokokawa et al., 2010), but there are no published studies in natural settings that compare the formative flow conditions and the resulting sedimentary structures. Alexander et al. (2001) suggested that the dimensions of lenses formed in flumes under stationary water-surface waves are related to antidune wavelength and thus could be used to estimate flow velocity and
depth (Fig. 1). The Alexander et al. (2001) empirical relationship was used by Duller et al. (2008) to infer flow conditions in the 1918 Katla jökulhlaup from lenses preserved in sand-granule grade deposits. While this demonstrated the utility of the relationship to palaeoflow reconstruction, there were no direct measurements of the flow conditions which generated the documented deposit.

This article presents a unique data set that reinforces the field applicability of empirical relationships to the study of sedimentary structures generated by supercritical flow. Measurements of water-surface waves in a flash flood and lenses in the gravel deposits of the same flood within a study reach are compared to test the Alexander et al. (2001) relationship for lens size as an indicator of flow conditions, for the first time on a field-scale example. These measurements are compared to flow velocities derived from flow competency relationships using the dimensions of clasts observed in transport in the flood, and boulders in the event deposit.

Deposit grain size has been used to indicate palaeoflow velocity (e.g. Costa, 1983; Komar, 1987; Clarke, 1996; Mather & Hartley, 2005), but this is beset with problems associated with accounting for form drag, emergence, grain packing, sorting and exposure, suspended sediment concentration, slope and flow unsteadiness (Laronne & Carson, 1976; Church, 1978; Andrews, 1983; Church et al., 1998; Meirovich et al., 1998; Shvidchenko & Pender, 2000; Carling et al., 2002a,b; Petit et al., 2005; Billi, 2011; Parker et al., 2011; Alexander & Cooker, 2016). Importantly, the sizes of sediment available for transport impacts the reliability of grain size as a palaeoflow indicator. This study provides a test for the usefulness of antidune lens size as an alternative indicator of depositional flow conditions. The diagnostic value of lens size for flow conditions may be more reliable in conditions where flow is fully turbulent, rapidly varying, unsteady and Newtonian. This is often the case in flash floods: flows have the competence to transport larger clasts than are available for

![Fig. 1.](image-url)
transport. In these situations, maximum clast size would underestimate palaeoflow velocity. Antidune wavelength is independent of grain size, so lens length could generate more reliable estimates, and using both methods provides more information and thus confidence to interpretations.

### Water-surface waves and antidunes

Stationary water-surface waves are often observed in rivers and floods when flow velocity is greater than or equal to the celerity of water-surface waves, i.e. when the Froude number, \( Fr \), is equal to or greater than 1 (\( Fr = U/(gh)^{1/3} \), where \( U \) is mean velocity, \( h \) is flow depth, and \( g \) is acceleration due to gravity). In these flow conditions, over a non-cohesive mobile bed, feedback between the flow and the bed will form sediment waves (antidunes) that are in-phase with, or nearly in-phase with, the water-surface waves (Fig. 1). The wave crests of stationary water-surface waves remain in fixed positions or move slowly upstream or downstream (Núñez-González & Martin-Vide, 2011). It is important to differentiate these waves from a standing wave, which in the strict sense is a stationary wave which oscillates about a fixed point (node) at a frequency relative to the confining conditions; waves related to antidunes do not oscillate in this way.

Supercritical or near-critical flows that contain moving gravel make the collection of precise velocity data particularly difficult. Measuring depth precisely in fast flows over non-cohesive mobile beds is also challenging. However, it is possible to estimate flow conditions from water-surface wave dimensions (Tinkler, 1997a,b; Douxchamps et al., 2005) and this approach is used herein. Kennedy (1963) defined relationships between the wavelength, \( \lambda \), of stationary water-surface waves and mean flow velocity, \( U \), for two-dimensional waves (where the crest-parallel length is long compared to the wavelength):

\[
U = \sqrt{\frac{g\lambda}{2\pi}} \tag{1}
\]

and between \( \lambda \) and the mean flow depth, \( h_m \):

\[
h_m = \frac{\lambda}{2\pi} \tag{2}
\]

where two stationary water-surface wave trains interact at an angle (due to bed or bank topography), the resulting waves may have short crests and are thus called ‘three-dimensional waves’ (Yokokawa et al., 2010) or colloquially ‘rooster tails’ (Kennedy, 1961). Kennedy’s (1961) Eq. 1 modified for three-dimensional waves is:

\[
U = \sqrt{\frac{g\lambda}{2\pi}} \cdot \left[ 1 + \left( \frac{\lambda}{\lambda_t} \right)^2 \right]^{1/2} \tag{3}
\]

where \( \lambda_t \) is the transverse wavelength of the waveform.

Antidune bedforms develop in-phase or intermittently in-phase (with breaking waves) with the water-surface waves and have the same wavelength (Allen, 1982), which is controlled by the water depth and velocity and thus largely independent of the bed composition (grain size) (cf. Cartigny et al., 2014). Consequently, preserved antidunes can be used to estimate palaeoflow depth and velocity (e.g. Kennedy, 1963; Barwis & Hayes, 1985; Alexander & Fielding, 1997; Fralick, 1999; Carling et al., 2009). Antidune wavelength is considered a more reliable palaeoflow indicator than dune size (Vesipa et al., 2014).

Trains of stationary water-surface waves have been observed in many settings where fast shallow flow occurs, including, for example, glacier outburst floods (Magnússon et al., 2012; see video clip at youtube.com/watch?v=9sryall5700), rain-storm induced flash floods in arid and semi-arid settings (Reid & Frostick, 1987; Jones & Mulholland, 1999), rockbed and gravel-bed channels (Tinkler, 1997a,b), sand-bed channels on beaches (Grant, 1997) and in lahars (Pierson & Scott, 1985; Hayes et al., 2002). The prevalence of water-surface waves in certain environments highlights the potential of antidune deposits for palaeoflow reconstruction, if sedimentary structures associated with antidunes are preserved and identified in the sediment record.

### Sedimentary structures resulting from antidune formation, migration and truncation

Flume-generated antidunes are well-documented (Jopling & Richardson, 1966; Yagishita & Taira, 1989; Alexander et al., 2001; Yokokawa et al., 2010; Carling et al., 2014), and antidunes have been described from a wide range of modern settings including rivers (e.g. Harms & Fahnestock, 1965; Shaw & Kellerhals, 1977; Langford & Bracken, 1987; Cotter & Graham, 1991; Alexander & Fielding, 1997), alluvial fans (e.g. Blair & McPherson, 1999; Blair, 2000), washover fans
(Barwis & Hayes, 1985), glacial outwash streams (e.g. Fralick, 1999; Kjaer et al., 2004), glacial outburst floods (e.g. Burke et al., 2010; Lang & Winesmann, 2013), jökulhlaups triggered by volcanic activity (e.g. Duller et al., 2008; Marren et al., 2009), tsunami run-up deposits (e.g. Fujiwara & Tanigawa, 2014) and also in the rock record (e.g. Browne & Plint, 1994; Maejima et al., 2009). In most settings, the morphology of the original antidune is rarely preserved sufficiently well to measure wavelength (Alexander & Fielding, 1997; Fielding, 2006). The sedimentary structures resulting from formation, migration and truncation of antidunes are more likely to be preserved than the topographic form of antidunes on a bed. These structures are increasingly well-understood and recognizable (cf. Alexander et al., 2001 and Duller et al., 2008).

Combining the flume observations of Middleton (1965), McBride et al. (1975), Foley (1977), Yagishita & Taira (1989), Yokokawa et al. (1999), Alexander et al. (2001), Breakspear (2008), Yokokawa et al. (2010) and Cartigny et al. (2014) provides a scheme of reference for field interpretation of the sedimentary architecture of antidune deposits (Table 1; Fig. 2). Different modes of water-surface wave breaking have been associated with different sedimentary structures. Cartigny et al. (2014) demonstrated that ‘stable antidunes’ (non-breaking waves) produced upstream-dipping laminae sets; the thickness and internal architecture depending on factors including bed aggradation rate. ‘Unstable antidunes’ (formed under breaking waves) produce lenses with a variety of internal fabric and lamination depending on wave behaviour. Sand lenses with boundary-conformable laminae are characteristic of ‘unstable antidunes’ (Middleton, 1965; Hand, 1974; Alexander et al., 2001; Duller et al., 2008; Cartigny et al., 2014). Gravel lenses with comparable internal architectures to such sand lenses were observed by Breakspear (2008).

**THE BELHAM RIVER ON THE ISLAND OF MONTSESSAT, WEST INDIES**

The Soufrière Hills volcano on the island of Montserrat (16°72′N, 62°18′W, Fig. 3A and B) has been in episodic eruption since July 1995 (Wadge et al., 2014). Eruptive products (pyroclastic density currents, lahars, debris avalanches and tephra-fall) from the lava dome have been deposited across the southern half of the island. The copious volumes of unconsolidated pyroclastic debris are being remobilized, transported and redeposited by frequent flash floods (lahars; Barclay et al., 2007; Alexander et al., 2010). Flash floods result from high-intensity rainfall (more than 500 mm of rain often falls in a single event at intensities ≥50 mm h⁻¹) from localized and convective storms, and less frequent large-scale synoptic weather systems including hurricanes (Gray, 1979; Matthews et al., 2002). The magnitude of flash flooding has increased in response to catchment changes, resulting directly and indirectly from volcanic activity (Alexander et al., 2010; Froude, 2015).

The Belham River Valley is located on the north-western flanks of the Soufrière Hills volcano (Fig. 3C). Pre-1995, the whole catchment was densely vegetated with areas of dry (29%), mesic (48%) and wet forest (13%) (Young, 2008; Fig. 4A to C) and small-scale agricultural production. Since July 1995, volcanic activity has damaged or destroyed large areas of vegetation, resulting in bare earth surfaces and a predominance of dry forest species on revegetating slopes. Extensive volcanic deposits and the changes in vegetation have profoundly altered run-off dynamics (Alexander et al., 2010; Froude, 2015). Between July 1995 and March 2013, 5.1 × 10⁷ m³ (estimated from repeat topographic survey; Froude, 2015) of volcanic sediment was deposited on the Belham Catchment. Flash floods remobilized more than 10 × 10⁶ m³ of this primary volcanic material into the Belham Valley with an estimated 4.4 × 10⁶ m³ remaining in channel storage and the rest transported offshore (Froude, 2015).

Volcanic activity, increases in flash flooding magnitude and frequency, and increased sediment flux have transformed the morphology of the Belham River from its former narrow (2 to 3 m) gravel-bed channel, confined by steep valley sides (Fig. 4D and H). The valley floor has aggraded by an average of 9.45 m (maximum 31.11 m; Fig. 5) and steepened. From the Sappit River confluence to point E on Fig. 3C, the gradient of the Belham Valley has increased from 0.8° to 2.86°. Due to aggradation, the channel width has increased from ca 2 m to between 59 m and 182 m (Froude, 2015). Local aggradation rates vary between successive flash floods and local incision also occurs, where steep-sided channels are cut and filled in successive
<table>
<thead>
<tr>
<th>Study</th>
<th>Experimental conditions</th>
<th>Median grain size (D50) and SD in brackets</th>
<th>Froude number</th>
<th>Bedform wavelength (m)</th>
<th>Bedform amplitude (m)</th>
<th>Bedform migration</th>
<th>Bedform dimensionality</th>
<th>Sedimentary structures (reference Fig. 2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Middleton (1965)</td>
<td></td>
<td>0.19 mm (1.44)</td>
<td>1.1–1.7</td>
<td>1.0–2.0</td>
<td>0.05</td>
<td>Upstream</td>
<td>2D</td>
<td>1–6, 8, 16</td>
</tr>
<tr>
<td>McBride et al. (1975)</td>
<td></td>
<td>0.34 mm and 0.50 mm (0.71 and 1.26)</td>
<td>1.4 and 1.0</td>
<td>0.1–0.2</td>
<td>0.02–0.08</td>
<td>Downstream</td>
<td>2D</td>
<td>16</td>
</tr>
<tr>
<td>Runs 7/12 and 6/12</td>
<td></td>
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<td></td>
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<td></td>
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<td></td>
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<tr>
<td>Foley (1977)</td>
<td></td>
<td>μ = 0.28 mm (1.42)</td>
<td>1.37</td>
<td>0.4</td>
<td>0.06</td>
<td>Stationary</td>
<td>3D</td>
<td>8, 12</td>
</tr>
<tr>
<td>Yagishita &amp; Taira (1989)</td>
<td></td>
<td>μ = 0.5 mm (1.42)</td>
<td>1.66–1.76</td>
<td>0.6–0.7</td>
<td>0.03</td>
<td>Upstream</td>
<td>2D</td>
<td>8, 17</td>
</tr>
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<td>Yokokawa et al. (1999)</td>
<td></td>
<td>0.18 mm</td>
<td>1.3</td>
<td>0.4–1.0</td>
<td>0.01–0.02</td>
<td>Upstream</td>
<td>2D</td>
<td>5–6, 8–9, 11–13</td>
</tr>
<tr>
<td>Alexander et al. (2001)</td>
<td></td>
<td>0.42 mm (0.71)</td>
<td>1.5–1.71</td>
<td>0.76–1.14</td>
<td>0.035–0.066</td>
<td>Stationary – upstream</td>
<td>2D</td>
<td>3, 6–11, 13–16</td>
</tr>
<tr>
<td>Breakspear (2008) A, Run 1</td>
<td></td>
<td>1.21 mm (1.49)</td>
<td>1.59</td>
<td>0.4</td>
<td>0.02</td>
<td>Upstream</td>
<td>3D</td>
<td>1, 5, 8–9, 12–15, 18</td>
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<tr>
<td>Breakspear (2008) B, Run 2</td>
<td></td>
<td>1.06 mm (1.32)</td>
<td>1.35–1.75</td>
<td>0.4</td>
<td>0.018</td>
<td>Upstream</td>
<td>2D/3D</td>
<td>2, 5, 12–13, 19, 22</td>
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<td>Yokokawa et al. (2010)</td>
<td></td>
<td>0.18 mm (0.44)</td>
<td>1.26–2.62</td>
<td>0.4–0.5</td>
<td>0.015–0.02</td>
<td>Stationary – upstream</td>
<td>2D and 3D</td>
<td>2–3, 5–11, 14–16</td>
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<tr>
<td>Cartigny et al. (2014) A</td>
<td></td>
<td>0.35 mm and 0.265 mm</td>
<td>Fr50 = 1.0–1.41</td>
<td>–</td>
<td>–</td>
<td>Upstream</td>
<td>2D</td>
<td>7, 18–19</td>
</tr>
<tr>
<td>Stable antidunes Runs [7, 11], [18–19, 22]</td>
<td></td>
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</tr>
<tr>
<td>Cartigny et al. (2014) B</td>
<td></td>
<td>0.16 mm</td>
<td>Fr50 = 1.07</td>
<td>–</td>
<td>–</td>
<td>Upstream</td>
<td>2D</td>
<td>3, 8, 10, 14, 20–21</td>
</tr>
</tbody>
</table>
events (Fig. 4F and I). The Belham River is ephemeral, and for most of the year the valley floor is dry; flash flooding occurs for short periods after rain storms. Flow is highly unsteady, intensely turbulent and sediment-laden (Barclay et al., 2007; Alexander et al., 2010). Stationary waves are prevalent in the middle and lower reaches of the river during all observed flow events.

The sediment deposited during 2009 to 2013 in the Belham Valley ranges from clay-grade (volcanic ash) to boulders (up to and greater than 2 m diameter). The valley floor varies in space and time from boulder fields (Fig. 4E) to nearly flat sand sheets with primary current lin- eation (Fig. 4H) to mud draped surfaces (Fig. 4I). The thickest mud drapes form where flood water is locally and temporarily dammed (Fig. 4J). The bed has been repeatedly surveyed and sampled over the past two decades, and deposits examined where they are exposed in natural channel banks or artificial trench walls (Fig. 4K).

METHODS

This study compares flow conditions during a sediment-laden flash flood with palaeoflow reconstructions using antidune lenses and grain size in the event deposit. To overcome the challenges inherent in observing sediment-laden flash floods, a camera was installed in a remote site overlooking a reach in the Belham River Valley. Imagery from this monitoring camera was used to measure flow properties (water-surface waves, boulders in transport and channel dimensions) during a flash flood event on 13 to 14 October 2012. Deposits of this event (before significant reworking) were examined on the dry bed and in trenches within 400 m of the camera site. Sedimentary structures developed in gravel that are comparable to the flume sand structures are documented, confirming their diagnostic application. Data on stationary waves in the flash flood and gravel lens architecture in the event’s deposits are used to test the flume-
derived empirical relationship between antidune lens length and flow conditions (Alexander et al., 2001).

**Seismic flow monitoring**

Flow is not directly gauged, but three-component seismometers, installed as part of the volcano monitoring network (Fig. 3C), continuously record the ground movements, and some of these movements are generated by flood events. The character of the seismic signal is used by the Montserrat Volcanic Observatory to monitor the occurrence and relative magnitude of flash floods, based on signal duration and number of peaks. Flash floods are identified as very long period (ca 10 min to several hours) seismic signals (Zobin et al., 2009) that have an elevated 2 to 5 Hz signal and peaks between 1 Hz and 30 Hz, and are accompanied by seismic noise in the lower frequencies (<1 Hz). Elevated seismicity corresponds to increased turbulence, flow volume or sediment load (Doyle et al., 2011). Seismic data thereby not only show the onset and duration of a flood, but also indicate peaks in discharge during an event. Signals interpreted as flash floods were verified by correlation with records of visual observation of the flood and

Fig. 4. Sites of interest in the Belham River Valley. Foreground scale of the photograph included, unless otherwise stated in figure caption. (A) Dry Forest. (B) Mesic Forest. (C) Wet Forest. (D) The lower Belham River Valley pre-eruption, looking towards the Soufrière Hills (photograph reproduced with permission, Richard Herd). Field of view (12·5° horizontal, 5° vertical). (E) Looking upstream from 100 m upstream of B1 (Fig. 3C), April 2000. The bed surface was covered with 0·5 to 1·0 m boulders that were resting (not buried) on a sand bed. Wood debris was also scattered across the valley floor. (F) Looking upstream from 300 m upstream of B1 (Fig. 3C). A large flash flood on 2 May 2006 incised 2 m into the valley floor (Sušnik, 2009). The two-storey orange house was gradually buried by lahars deposits between 2002 and 2010 (Froude, 2015). The incised channel width is 22 m. (G) Looking upstream towards the trench site (Fig. 3C), March 2012. The bed surface was ca 40% gravel: 60% sand with cobble-boulder bars in the centre of the valley and sand-pebble bars in the centre of sandy channels (ca 15 m wide) (Froude, 2015). Gravel bars remained unmodified in their central parts between November 2010 and March 2012. Vegetation growing within the channel indicates minimal disturbance. (H) Looking downstream and southward from site B2 (Fig. 3C) on 5 May 2011, immediately following a small flash flood. The half-buried house in the background (edge of valley floor) is ca 10 m wide. (I) Looking upstream and southward from 200 m downstream of site B2 (Fig. 3C) on 27 May 2011, immediately following a small flash flood. The half-buried house to the top right of the photograph is same as in (H). The flow had incised into a gravel bar in the centre of the valley floor, and deposited mud drapes over the bed surface down the entire valley floor. Some of the mud drapes have been removed by subsequent low-level flow. (J) Flow on 27 May 2011 pooled in commercial aggregate mining trenches (between site B1 and B2, Fig. 3C), leaving mud drapes. The trench width was ca 20 m, length ca 100 m and water depth <0·5 m. (K) Vertical exposure (the north side of the valley at B2, Fig. 3C) created by commercial aggregate extraction. Photograph from March 2011. Person is 1·57 m tall.

Fig. 5. Cross-section profile of a transect 100 m downstream of the trench site (Fig. 3C) showing elevation change due to primary volcanic and flash flood activity between July 1995 and March 2013 (Froude, 2015). Elevation data: pre-eruption DEM (Wadge, 2000); May 2003 (dGPS survey; Barclay et al., 2007); May 2005 (dGPS survey; Sušnik, 2009); November 2006 (dGPS survey; Darnell, 2010); November 2007 (dGPS survey; Darnell, 2010); June 2010 (Airborne LiDAR; SAC, 2011); March 2011 (dGPS survey; Froude, 2015); February 2012 (dGPS survey; Froude, 2015); March 2013 (dGPS survey; Froude, 2015). Locally elevations in this part of the valley varied by ±1 m between March 2011 and March 2013.

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rainfall records (Froude, 2015). Since 1995, the monitoring network has detected >585 rainfall-triggered flash floods of which >351 events have been verified visually: most of the unverified events occurred during the hours of darkness.

Remote monitoring camera: image acquisition and processing

In-person direct visual observation of valley-wide flows was limited to sites >5 km from the volcanic edifice, and these areas were frequently modified by commercial aggregate extraction. To monitor a section of the natural channel, unaffected by aggregate extraction, a telemetered remote camera was installed 3-8 km from the volcano in March 2012, upstream of any likely sand quarrying activity. The field of view of the camera encompassed a 50 to 69 m wide, 113 m long reach of the Belham River (herein referred to as the camera site).

The equipment comprised of a Stardot™ SC 5 megapixel IP camera (StarDot Technologies, Buena Park, CA, USA) with a wide-varifocal auto iris lens, powered by two 100 watt solar panels charging two 12-volt car batteries. Image capture was controlled by a custom built Raspberry Pi™ field computer running a Python script which captured five megapixel images at a rate of one frame per second during daylight hours, storing each image on a local hard drive. The camera was telemetered to the Montserrat Volcano Observatory, facilitating two-way data transfer.

Objects moving within the flow and surface flow features were measured within a sample cross-section, constrained by permanent poles and boulders of precise known locations (Fig. 6A). Photograph coordinates for the poles and boulders were obtained using DigitizeIt™ software. One pixel distance between P1 and R1 (vector A; Fig. 6A) equated to 0.026 m horizontal ground distance. One point between P2 and P3 equated to 0.047 m horizontal ground distance. Considering the distance between the vertices P1 and P3, a linear equation was derived to calculate the change in pixel:metre ratio between the foreground and background of the channel cross-section:

\[ y = 0.00035x + 0.026 \]  

where \( x \) is the pixel distance from vertex P1 along the vector P1 to P3, and \( y \) is a scaling coefficient to convert one pixel to metres. The equation was tested using objects of known size spread across the sample cross-section (Fig. 6A). Error ranged from 0.1 m in the background of the image to 0.05 m in the foreground of the image.

Image coordinates for the crests of water-surface waves in wave trains were obtained using DigitizeIt™ software. The distance between image coordinates was calculated, and the location of the mid-point coordinate (equidistant between the two wave crests) was located on the P1 to P3 (vector, Fig. 6A), providing a value of \( x \) for Eq. 4, and thereby the scaling coefficient \( y \) to convert each measurement of wavelength into
metres. The average crest to crest distance was calculated for each wave train by dividing the total wave train length by the number of waves in the train. Boulder dimensions were also obtained using Eq. 4.

**Bed topography and topographic change**

To define changes in the valley and channel geometries in response to flash flood events, differential GPS surveys (dGPS) of the Belham Valley floor between the coast and point E (Fig. 3C), were undertaken in February 2012 and March 2013, and compared with earlier surveys (Barclay et al., 2007; Sušnik, 2009; Alexander et al., 2010; Darnell, 2010; Darnell et al., 2010).

A Leica AT302 GPS antenna (vertical accuracy ±10 to 20 mm, horizontal accuracy 10 to 20 mm; Leica Geosystems Inc., Norcross, GA, USA) was used in kinematic mode on-foot. A quasi-systematic method was adopted, in which the perimeter of individual topographic features, such as bars and banks (top-edge and base), was surveyed, in addition to a series of transects across the valley. Leica Geo-Office Software™ was used to post-process the data with reference to the base station (MVO1), orbit information, precise ephemeris and antenna calibration. Resolved points were assessed for quality, and points with a root-mean-square error >20 mm were removed from the data set. Points for each survey were interpolated in ESRI ArcMap™ software using a series of hard breaklines (along abrupt changes in slope) and a hard clip, which restricted the interpolation to the point cloud. Constrained Delaunay triangulation was used to connect the points in a series of triangles. Triangulated Irregular Network (TIN) surfaces retain information on steep breaks in slope if these are delineated during the data collection (Brasington et al., 2000; Darnell, 2010), providing useful markers of terrace erosion and channel migration between surveys. The TINs were converted directly to an elevation raster (digital elevation model – DEM) with a regular 10 × 10 m grid using a linear interpolation. Vertical accuracy is ±30 to 40 mm. Elevation change and volumetric change related to flash floods were calculated by differencing the DEMs.

Between the dGPS surveys in February 2012 and March 2013, only one large flash flood (lasting over 24 h with multiple peaks in discharge) occurred, on 13 to 14 October 2012. In addition, there were 25 small to moderate events that remained within sub-channels, did not modify bed elevation by more than the DEM error and left little or no trace in the sediment record.

**13 to 14 October 2012 flash flood measurements**

Images from the permanent monitoring camera were sampled systematically and analysed at 900-second intervals between 16:15:00 Coordinate Universal Time (UTC) 13 October 2012 and 22:00:00 UTC 14 October 2012. Local seismicity was recorded on three seismometers (MBLY, MBGH and MBGB; Fig. 3C). Direct rainfall measurement was by a tipping-bucket rain gauge on a Vantage Pro 2 weather station (station SGH2011 – Davis Instruments Corporation, Hayward, CA, USA; Fig. 3C), in the Belham Catchment at 325 m above sea-level. The rain gauge ceased functioning, just after 4:00:00 UTC on 14 October 2012. Rainfall intensity was estimated from Météo-France Guadeloupe rainfall radar images at 15-minute intervals. The radar time series correlates well with the timing of rainfall peaks registered on the SGH2011 weather station on 13 October, but underestimates measured peak rainfall intensity by 46%. This underestimate is based on the comparison of radar-derived estimates with rain gauge data when it was operational. Therefore, radar data intensity estimates on 14 October 2012 were adjusted accordingly.

Combined with seismic data and rainfall measurements, images at the camera site were used to monitor changing stage (estimated from proportion of the valley floor occupied by flow), timing of peak flow stages, the occurrence, geometry and size of stationary water-surface waves, the size of moving boulders, apparent flow turbulence and the timing of rainfall. Valley width occupied by flow served as a proxy for discharge, given the shallow sub-channel depth and high width:depth ratio of 112.

Hand-held cameras (still and video) provided supplementary images of conditions downstream (sites B1 and B2; Fig. 3C). This section of valley was highly modified by aggregate mining, but remained accessible during flood events. Flow velocity was estimated using a float distance transit at sites B1 and B2 (Fig. 3C), during daylight. This method provides an estimate of surface flow velocity along the travel vector of the floating object. These were the only direct measurements of velocity acquired during the flood event. Flow velocity and depth at the camera site were estimated using water-surface wave geometry, and Eqs 1 and 2.
sediment was sampled using a ‘bucket’ method with a 500 ml container at mid-water depth at the edge of the main channel at site B2 (Fig. 3C), during waxing flow (16:16:00 UTC 13/10/12, 16:21:00 UTC, 13/10/12) and during waning flow (21:00:00 UTC, 14/10/12). The dried mass was used to estimate the suspended sediment concentration and the grain-size distribution from laser-diffraction with a Malvern Mastersizer 2000 (Malvern Instruments Limited, Malvern, UK; Table 2).

Palaeoflow velocity estimates from grain size

Two methods were used to estimate flow velocity using the dimensions of large clasts observed to move in the flash flood and in the event deposit (Table 3). The Clarke (1996) method estimates the minimum critical force needed to initiate the movement of a submerged boulder in a steady flow. The critical flow velocity is calculated from the drag force \( C_D \) acting on the boulder, fluid density \( \rho \), and the cross-sectional area of the boulder. Both cubic and spherical shapes are considered and results averaged to estimate critical stream velocity. The Alexander & Cooker (2016) method considers forces acting on a boulder in unsteady flow, including an impulsive force generated by change in flow conditions.

13 to 14 October 2012 flash flood study sites

Two key sites within a study reach are discussed in this article: the camera site and the trench site (Fig. 3C). A sample transect across the camera site was surveyed before the flash flood on 12 October 2012 (Fig. 7A) and after the flood on 15 October 2012 (Fig. 7C). Grain size was estimated at 0.1 m intervals along this transect, and the presence or absence of vegetation recorded. Silt and sand grade (<2 mm) were classed as a single category, and medium, coarse and very coarse pebbles, cobbles and boulders were measured at each sample point. Geomorphological sketch maps drawn pre-flood and post-flood were used to record changes in channel structure at the camera site. Pre-event surveys showed that the dry valley floor had a gradient of 2.8° and was braided (Fig. 7B and E). The number of sub-channels, their depth and boundary characteristics changed during the flood. Both before and after the flood, the sub-channel depths were <0.2 m and channel floors were sandier than the gravel bar surfaces.

Table 2. Flow conditions at monitoring sites B1 and B2. Reynolds number is calculated as a range to incorporate the error (±0.3 m) for water depth.

<table>
<thead>
<tr>
<th>Site</th>
<th>Date/Time</th>
<th>Floating object</th>
<th>Surface velocity (m sec(^{-1}))</th>
<th>Estimated mid-depth velocity (m sec(^{-1}))</th>
<th>Water depth ( h ) (m)</th>
<th>Channel width ( W ) (m)</th>
<th>Suspended sediment concentration (g m(^{-3}))</th>
<th>Estimated Reynolds number</th>
</tr>
</thead>
<tbody>
<tr>
<td>B2</td>
<td>14/10/2012</td>
<td>Wooden log</td>
<td>2.17</td>
<td>1.86</td>
<td>0.5</td>
<td>0.5</td>
<td>10.5</td>
<td>359658 - 1518631</td>
</tr>
<tr>
<td></td>
<td>13/10/12</td>
<td>Coconut</td>
<td>2.55</td>
<td>2.45</td>
<td>0.5</td>
<td>0.5</td>
<td>13.29</td>
<td>116490/3</td>
</tr>
</tbody>
</table>

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Between the October 2012 flash flood and the subsequent field campaign in March 2013, deposits in the accessible (lower) reaches of the valley were incised a small amount (channel depths < 0·3 m) by small run-off events (Fig. 3C). Commercial aggregate excavation by backhoe, 400 m downstream of the camera site (Fig. 3C), created a large trench (herein called the trench site) on the north side of the valley creating temporary vertical exposures typically < 0·5 m high, extending laterally up to 150 m north-west (Fig. 8E). These provided the best vertical exposure of the 2012 flash flood event deposit, and the only exposure at a site that was previously unmodified by aggregate excavation. The trench was photographed (for example, Fig. 12A), sketched, logged and sampled.

Aerial imagery and a post-event survey transect at the trench site (prior to excavation) showed that the trench crossed a 23 m wide channel with 2·6° bed slope and gravelly surface on the north side of the valley (Fig. 8F). The channel boundaries were relatively unchanged by the 13 to 14 October 2012 flash flood. In November 2010, the bed surface at the trench site was near planar and covered by a thin laterally continuous massive mud layer (Fig. 8A). Small to moderate flash floods between November 2010 and March 2012 modified the bed forming shallow braided channels, and the surface sediment was sand within the channels and sandy-gravel on the bars (Fig. 8C).

### The 13 to 14 October 2012 Sediment-laden Flash Flood

The 13 to 14 October flash flood occurred during the passage of Tropical Storm Rafael which deposited 170 mm of rain in 28 h, with a maximum recorded intensity of 61·5 mm h⁻¹ and multiple peaks in rainfall (Fig. 9A). The first rain bands from Tropical Storm Rafael caused localized surface run-off and ponding water in trenches on the valley floor, observed at site B2 on 12 October 2012. Rainfall was initially sporadic (Fig. 9A), becoming near continuous from 14:34:00 UTC and flow was sustained by 16:10:00 UTC in a single 1 to 2 m wide channel on the north side of the valley at site B1. Local

---

**Table 3.** Estimates of flow velocity using water-surface wavelengths (using Eq. 1), clasts within the flow (using the methods of Clarke, 1996 and Alexander & Cooker, 2016), deposit lenses (assuming preservation of 38% and 56%) and deposit clasts (using the methods of Clarke, 1996 and Alexander & Cooker, 2016). Equations for the Clarke (1996) method are reviewed in Stokes et al. (2012).

<table>
<thead>
<tr>
<th>Date</th>
<th>Mean (m sec⁻¹)</th>
<th>max (m sec⁻¹)</th>
<th>Clarke (1996) method* (m sec⁻¹)</th>
<th>Alexander &amp; Cooker (2016) method† (m sec⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>13/10/2012</td>
<td>2·18</td>
<td>3·61</td>
<td>3·37</td>
<td>3·85</td>
</tr>
<tr>
<td>14/10/2012</td>
<td>1·82</td>
<td>2·65</td>
<td>2·36</td>
<td>2·69</td>
</tr>
</tbody>
</table>

*Assuming: acceleration of gravity (g) = 9·81 m sec⁻², density of clast (ρ) = 2650 kg m⁻³, density of fluid (ρ) = 999·1 kg m⁻³, bed slope angle (β) = 0·049 radians (camera site), 0·045 radians (trench site), lift coefficient (Cₗ) = 0·178 (cubic boulder), 0·2 (round boulder), drag coefficient (Cₜ) = 1·18 (cubic boulder), 0·2 (round boulder). †Assuming the boulder is spherical with a diameter equal to the mean axial length and the friction coefficient (Cₕ) = 1. The acceleration of flow (a) = 0·5 m sec⁻².
seismicity records confirm flood onset and indicate seven flood peaks (Fig. 9B); these correlate with visual estimates of flow stage, bedload size and quantity from the camera imagery.

The flash flood was highly unsteady, turbulent and sediment-laden. Flow velocity and depth at site B1 at the start of the event were 0·71 m sec$^{-1}$ and 0·5 m. Assuming a water temperature of 15°C, sediment grain density of quartz (2650 kg m$^{-3}$), with the flow conditions (Table 2), gives Reynolds numbers ($Re = \rho_lU_lh/\mu_m$, where $\rho_l$ is the bulk density of the flow and $\mu_m$ is the dynamic viscosity) 124 000 to 495 000 on the rising limb of the first flow peak at Site B1 and 380 000 to 1 520 000 on the falling limb of the last flow peak at Site B2. This result is consistent with previous inferences (Barclay et al., 2007; Alexander et al., 2010) that flash floods in this valley are Newtonian and fully turbulent.

Waves and surges in the 13 to 14 October 2012 flash flood – observations

Trains of stationary and upstream migrating waves were prevalent. At the camera site, each train consisted of 3 to 11 waves with wavelengths of 0·32 to 8·33 m (Fig. 9D). Assuming
two-dimensional waves, in the study reach the maximum mean flow velocity, $U$, was $3.6 \text{ m sec}^{-1}$ (Eq. 1), and maximum flow depth, $h_m$, was $1.3 \text{ m}$ (Eq. 2). Although most waves appear to be two-dimensional, the perspective and image resolution of photographs could mean that there were some three-dimensional waves that were not identified. Assuming a water temperature of 15°C and sediment grain density of quartz (2650 kg m$^{-3}$), the flow conditions give $Re$ values of 1 620 000 to 2 620 000, based on suspended sediment concentrations measured at sites B1 and B2 (Table 2). Table 4 contains estimates of maximum mean flow velocity and maximum flow depth for each sub-channel, showing the variation in flow conditions across the valley floor at this location.

At the camera site, waves extended partially or fully across sub-channels, varying with time. The first wave train formed shortly after occupation of the north sub-channel and persisted for ca 10 min, reducing in amplitude and disappearing as flow depth increased, but a new wave train formed at the same place in the channel.
Fig. 9. (A) Rainfall intensity calculated over 900-second intervals from the SGH2011 rain gauge (Fig. 3C) and the maximum pixel value for the Belham Catchment from the Météo-France Guadeloupe rainfall radar. (B) One-minute real-time seismic amplitude (RSAM) from MBLY, MBGH and MBGB. (C) Percentage of valley-floor width occupied by flow (red), passage of bores at the camera site (grey bars) and camera functionality. (D) The average wavelengths of water-surface wave trains present in the sample area (Fig. 6A) of camera images between 16:15:00 UTC 13/10/2012 and 22:00:00 UTC 14/10/2012, sampled every 15 min. Each wave train is located within a sub-channel on the valley floor (see Fig. 6B).

Table 4. Estimates of flow velocity (using Eq. 1) and flow depth (using Eq. 2), based on the maximum mean wavelength from wave trains observed at the camera site (Fig. 3C) between 16:15:00 UTC 13/12/2012 and 22:00:00 UTC 14/10/2012 (see Fig. 6D).

<table>
<thead>
<tr>
<th>Date</th>
<th>Sub-channel</th>
<th>Wavelength, $\lambda$ (m)</th>
<th>Mean flow velocity, $U$ (m sec$^{-1}$)</th>
<th>Flow depth, $h_m$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Min</td>
<td>Mean</td>
<td>Max</td>
</tr>
<tr>
<td>13/10/2012</td>
<td>North</td>
<td>0.42</td>
<td>0.94</td>
<td>1.38</td>
</tr>
<tr>
<td>13/10/2012</td>
<td>North-middle</td>
<td>1.62</td>
<td>2.44</td>
<td>3.34</td>
</tr>
<tr>
<td>13/10/2012</td>
<td>South-middle</td>
<td>1.85</td>
<td>3.81</td>
<td>8.31</td>
</tr>
<tr>
<td>13/10/2012</td>
<td>South</td>
<td>1.48</td>
<td>3.96</td>
<td>8.33</td>
</tr>
<tr>
<td>14/10/2012</td>
<td>North</td>
<td>1.75</td>
<td>3.10</td>
<td>4.48</td>
</tr>
<tr>
<td>14/10/2012</td>
<td>North-middle</td>
<td>1.58</td>
<td>2.54</td>
<td>3.94</td>
</tr>
<tr>
<td>14/10/2012</td>
<td>South-middle</td>
<td>1.63</td>
<td>1.81</td>
<td>2.00</td>
</tr>
<tr>
<td>14/10/2012</td>
<td>South</td>
<td>0.29</td>
<td>0.54</td>
<td>0.82</td>
</tr>
</tbody>
</table>

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within 30 sec. Waves were prevalent in all sub-channels; they grew and remained stationary for up to 9 min, or migrated upstream at rates <1 m sec\(^{-1}\). Migration upstream was not always continuous, and many wave trains paused for short periods (<10 sec) before continuing to migrate; this led to slower rates of migration, ca 0-1 m sec\(^{-1}\). Waves which migrated more rapidly tended to break en masse (i.e. more than one adjacent wave breaking at the same time), while slower or stationary waves broke individually (Video Clip S1); or sequentially, the breaking of one wave triggering the breaking of adjacent waves (Video Clip S2). Sometimes waves did not break, but dissipated as local flow conditions changed. Waves tended to reduce in height or disappear, when rain fell directly into the flow because flow depth locally (temporarily) increased (and Froude number decreased). One wave train in the south sub-channel persisted for over 60 min before disappearing, and a new wave train formed in its place shortly afterwards (~2 min). Some wave trains grew in length, with the addition of new waves at either end of the wave train. Entire wave trains were observed to change position across the valley floor in response to changes in channel boundaries, as well as migrating upstream.

Wave trains occupied most sub-channels at sites B1 and B2 when visited at 16:16:00, 16:56:00, 17:50:00 and 23:00:00 UTC on 13 October 2012, and at 12:00:00, 13:00:00 and 21:00:00 UTC on 14 October 2012. Visual observations of wave trains at sites B1 and B2 were in-person and with higher resolution image capture, but for significantly shorter duration than at the camera site. At site B1 when waves within trains grew and diminished in height, they also changed between undular and breaking form. Most waves observed were two-dimensional, with wavelengths between 0.3 m and 1.2 m; they persisted for <10 min at locations both in the centre of the channel and near the channel edge. The largest wave train observed (nine waves), with a mean wavelength of 5-4 m (centre of channel at site B1 17:15 to 18:15 on 14 October 2012), contained three-dimensional waves (with a mean transverse wavelength of 3.9 m) which changed between undular and breaking form (Video Clips S3 and S4), mostly remaining stationary with short periods of upstream migration. Some individual waves in the train broke and reformed, but the train as a whole persisted for longer than the one-hour observation period. Flow velocity was estimated as 1.6 m sec\(^{-1}\) using Eq. 3; this is lower than flow velocity estimated by float distance transit (Table 2), but is thought to better estimate mean conditions compared with the float measurement, because the water-surface waves persisted for a long period (more than an hour) and covered a larger part of the channel (190 m\(^2\)), and hence local variations in velocity with depth or position within the wave train were captured.

Three hundred and forty-nine turbulent surges (bores) were recorded during the daylight hours of the flood, occurring at 6 to 6002 second intervals (Fig. 9C); on average, 108 bores were recorded per hour during the event (minimum, 2 per hour and maximum, 190 per hour). These were downstream propagating and channel wide with distinct breaking fronts (Video Clip S5). The bores travelled at an average velocity of ca 6 m sec\(^{-1}\) and induced temporary increases in flow depth of up to half a metre; during waxing, flow bores gradually increased the flow stage; however, during waning conditions, depth change was associated with the front and did not increase flow stage. When a bore approached a wave train (within ca 10 m), the waves tended to break concurrently generating a localized short-duration upstream surge. Most waves broke during the passage of a bore, and those which did not reduced in height. The bores were not observed at downstream Sites B1 and B2.

At the camera site, moving cobbles and boulders were observed episodically when they were temporarily exposed by flow, or when they were partially submerged. The biggest clast moving had a maximum observed diameter of 0.89 m, at 17:00 13/10/2012 in the north channel. The biggest clast observed in the middle channels and south channel was 0.62 m at 20:30 on 13/10/2012 and 0.4 m at 19:00 on 13/12/2012, respectively.

**DEPOSIT SEDIMENTOLOGY**

Three vertical sections from the trench are presented herein, located: (i) parallel to flow within the channel; (ii) near-parallel to flow and parallel to the channel edge; and (iii) perpendicular to flow, cross-cutting the centre of the channel (Fig. 8E). Facies boundaries, sedimentary structures and lamination were delineated (Fig. 10), and architectural element diagrams were drawn (Figs 12 to 14). Grain-size analyses of samples are presented in Fig. 11.

The deposits exposed in the trenches consist of: (i) matrix-supported gravel lenses (Gmw); (ii) clast-supported gravel (Gcl), sometimes
Fig. 10. Sedimentary logs of two flow-parallel and one flow-perpendicular trench sections (Fig. 8E). Table 5 contains the facies code. Sediment sample locations are shown by blue arrows with numeric ID.

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organized in a lenticular pattern; (iii) laminated sand (Sh); (iv) massive sand (Sm); (v) silt and fine sand (Fm); (vi) pebbly sand with low-angle cross-beds (GSI); (vii) laminated gravel (Gh); and (viii) beds of massive gravel (Gmm) (Fig. 10; Table 5).

**Sedimentary structures formed in the 13 to 14 October 2012 flood event**

A distinct sedimentary contact was recognized over >1000 m laterally around the trench faces, at depths varying from 0.45 to 1.4 m below the natural deposit surface (see Logs 1 and 3: Fig. 10). This is interpreted as the base of the 2012 flash-flood deposits. Repeat dGPS surveys in March 2012 and March 2013 record ca 0.3 m of aggradation at the trench site, indicating that although there was net aggradation, the pre-event bed was remobilized during the flood (i.e. some bed erosion during the flood followed deposition, scour and fill).

The 13 to 14 October 2012 event deposit includes horizontally laminated sand (Sh), dipping sand and pebble beds (GSl), clast-supported gravel (Gcl) and massive sands (Sm) (Table 5). The event deposit recorded in Log 1 (Fig. 10) predominately consists of clast-supported, open framework to partly open-framework gravel with lenticular structures, interbedded with massive sand (Fig. 12). Lens bases cut into older lenses or massive sand beds. The lowermost lenses cut into pre-event thick massive sand beds (sample S3, grain size shown in Fig. 11). There is a thin layer (<0.02 m) of medium to very coarse sand at the base of most of the gravel lenses (S2, grain size shown in Fig. 11). Where the tops of the lenses are not truncated by other lenses, they are overlain by massive sand. Sand also forms the matrix in the gravel at the top of the lenses. In larger lenses (>0.84 m), the long axis of gravel clasts is inclined towards the lens centre, forming a concave-up pattern. The smallest lenses contain massive finer gravel (relative to the whole deposit). At the site of Log 1, there was one large clast in the event deposit (maximum diameter of 0.37 m), and all others had a maximum diameter <0.13 m.

In the trench face recorded in Log 2 (Fig. 10), the 2012 flash flood deposit is the upper 0.36 m, but in this flow-parallel section, it was partly obscured by collapse of the trench face. A flow-perpendicular section excavated at this location (Fig. 13) exposed inter-bedded planar laminated pebbly sand (Sh, beds <0.02 m thick, Fig. 10) and clast-supported pebble–cobble gravel lenses with erosional bases. Gravel clast fabric forms concave-up structures within most lenses, although some contained convex-upward clast patterns (Fig. 13). The lens basal boundaries are erosional, but the sand matrix within some lenses appears to grade into the surrounding Sh beds. There are also discontinuous beds of clast-supported gravel of large pebble–cobble grade (Gh). The 0.2 m thick massive mud bed (S5 grain size shown in Fig. 11 and structures in Fig. 13), by comparison with dGPS survey data, corresponds to the extensive mud layer deposited on the valley floor at this location in November 2010 (Fig. 8A). It is not possible to distinguish deposits from small–moderate flash floods that occurred between November 2010 and October 2012, but ground observations on

**Fig. 11.** Grain size of deposit at sampled units (see Fig. 10). The 64–0 to 0.5 mm fraction was determined using dry sieving. Laser-diffraction analysis (using the Malvern Mastersizer 2000) was used to ascertain the grain-size distribution of particles between 0.1 μm and 2.0 mm. The sample number locations are indicated on the logs in Fig. 10.

**Table 5.** Facies code descriptions.

<table>
<thead>
<tr>
<th>Lithofacies code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fm</td>
<td>Massive mud</td>
</tr>
<tr>
<td>Sh</td>
<td>Horizontally laminated sand</td>
</tr>
<tr>
<td>Sm</td>
<td>Massive sand</td>
</tr>
<tr>
<td>GSI</td>
<td>Dipping, laminated pebbly sand</td>
</tr>
<tr>
<td>Gh</td>
<td>Laminated gravel</td>
</tr>
<tr>
<td>Gmm</td>
<td>Massive matrix-supported gravel</td>
</tr>
<tr>
<td>Gmw</td>
<td>Matrix-supported gravel with grading and containing weak wavy lamination</td>
</tr>
<tr>
<td>Gcl</td>
<td>Clast-supported gravel from open framework (no matrix) to clast-supported</td>
</tr>
<tr>
<td></td>
<td>with matrix, graded and laminated</td>
</tr>
</tbody>
</table>
15 October 2012 show that this valley-edge location had been occupied by flow. The largest clast in the event deposit at the site of Log 2 was 0.12 m (maximum diameter).

The face recorded in Log 3 (Fig. 10) was perpendicular to flow direction, cross-cutting the sub-channel on the north side of the valley floor (Fig. 8E). In the upper 1.33 m, the 2012 flash flood deposit (Fig. 14) contains clast-supported gravel (Gcl) organized in lenses that truncate and are superimposed on one another. The lens boundaries are delineated by gradational grain-size changes: fine pebbles–boulders within most lenses, to massive and weakly laminated medium to very coarse sand between the lenses (S6, grain size shown in Fig. 11). The horizontal sand beds (for example, at 0.77 m depth) are up to 0.2 m thick, and the planar laminae are discontinuous. The lenses predominantly contain concave-up laminae (finer gravel) and clast fabric patterns (cobble–boulder gravel) that are concordant with lens base, or downlap asymptotically onto the basal surface. Some smaller lenses (<0.2 m) are contained within tabular beds of clast-supported pebble-cobble gravel (Gh). The largest clast in the event deposit at the site of Log 3 had a maximum diameter of 0.18 m.

**DISCUSSION**

**Lens size in relation to antidune wavelength**

To infer antidune wavelength from lens size, the lens length ($L_4$) parallel to flow direction is required (Fig. 1). Lenses are elongate across stream because they are related to waves that extended partly or fully across the channel (rather than three-dimensional antidunes, cf. Yokokawa et al., 2010). Lenses were excavated to measure their dimensions in the trench face oriented approximately parallel to the channel axis. In other settings, where flow direction is not easily obtained, it may be possible to use grain fabric to estimate the angle of exposure relative to flow
direction to assess flow-parallel lens length. Although this was not assessed directly in this study, strong grain fabrics have been found in experimental and fluvial antidunes (Fig. 15; Yagishita & Taira, 1989; Cheel, 1990; Alexander & Fielding, 1997; Breakspear, 2008) and may be used as a palaeocurrent indicator.

The geometry of an antidune trough is determined by the wavelength of the formative water-surface wave; however, the geometry of the trough preserved as a lens depends not only on water-surface wavelength (antidune wavelength), but on the violence of wave breaking (Blair, 2000), local sediment flux and subsequent bed remobilization. Local sediment flux and the violence of wave breaking vary in both time and space during a flooding event so not all of an antidune is preserved in a sedimentary deposit, and preservation is not uniform. Consequently, individual lens size is unlikely to be an accurate record of an individual stationary wavelength. Mean lens size in one bed, however, may be related to the mean wavelength (temporal and spatial averaging). For example, Alexander et al. (2001) observed lenses in flow-parallel sections with an average length of 0.43 m (range 0.19 to 0.75 m, standard deviation 0.13 m) following a flume run with 0.07 m flow depth, 0.76 to 1.14 m wavelength stationary waves and sand addition giving a bed aggradation rate of 0.01 mm sec$^{-1}$. The ratio of mean lens length $L_L$ to measured water-surface wavelength, $\lambda$, in the flume was between 0.38 and 0.56. The range of lens lengths was very comparable to those from a run with no net bed aggradation. In a similar flume experiment with water depth 0.04 m, mean flow velocity 0.85 ms$^{-1}$ and sand-bed aggradation 0.01 mm sec$^{-1}$, Yokokawa et al. (2010) recorded

Fig. 13. (A) The 13 to 14 October 2012 flood event deposit in the uppermost 0.36 m of exposure at Log 2 (Figs 8E and 10); the upper 0.11 m of deposit was modified in places at this site during trench excavation. The section contains clast-support gravel lenses (Gcl) and planar laminated sand (Sh) and gravel (Gh, delineated by the line drawing in (B)). The massive mud (Fm) at the base of the exposed section corresponds to the bed surface at this location in November 2010 (Fig. 8A).
Fig. 14. (A) The 13 to 14 October 2012 event deposit in the uppermost 1.33 m of exposure at Log 3 (Figs 8E and 10). The section contains clast-supported gravel lenses (Gcl) and massive (Sm) to weakly planar laminated sand (Sh) and gravel [Gh, delineated by the line drawing in (B)].

Fig. 15. Imbrication of grains/clasts dipping upstream, on the stoss and lee antidune face formed under non-breaking wave conditions (Yagishita & Taira, 1989). The experimental conditions of each study are provided in Table 1. Yokokawa et al. (1999) showed that mean imbrication angle varied cyclically through a vertical antidune section formed by breaking water-surface waves, while Breakspear (2008) showed that clasts were consistently imbricated upstream in non-breaking and breaking wave experiments in gravels.
lenses from 3D antidunes with a mean lens length to water-surface wavelength ratio of 0.4 to 0.5.

Deposits from the 2012 flash flood were only recorded in one flow-parallel section (Log 1: Fig. 10). Gravel lenses had an average length of 0.75 m (range 0.23 to 2.92 m and standard deviation, 0.75 m). Assuming a 38 to 56% preservation (based on preservation in flume experiments), the mean antidune length at the gravel trench site was 1.34 to 1.98 m. This lies within the 1.48 to 8.33 m water-surface wavelengths recorded at the camera site. Using the antidune wavelength range inferred from lenses in the deposit gives estimated flow velocity of 1.43 to 1.76 m sec\(^{-1}\) and depths of 0.21 to 0.32 m at the trench site. These estimates are within the range (minimum to maximum) for flow parameters from water-surface waves at the camera site (Table 3).

**Lens preservation**

The flow conditions estimated from the mean length of lenses associated with the 13 to 14 October 2012 flash flood measured at the trench site are less than the average flow conditions and the maximum peak flow observed at the camera site. There are three possible reasons for this: (i) flow conditions in this part of the channel (400 m from the camera site) were different and water-surface waves that formed had shorter wavelengths compared with those at the camera site, even though the gradient and channel size were comparable; (ii) preservation of lens length was a lesser proportion of the antidune length than preservation observed in flume studies; or (iii) the lenses formed on 14 October, not during peak discharge on 13 October. The camera site and trench site were similar in terms of valley-floor gradient, width, channel pattern and bed surface texture and were situated in close proximity (400 m apart). Average flow conditions at the trench site are thought to have been similar to average conditions at the camera site. This leaves the possibilities that most of preserved lenses formed on the second day of the event and also that the percentage of lens preserved was less than observed in flume experiments.

In the Belham River deposits, there is a strong correlation between lens length and thickness (Fig. 16). As Bridge (1997) and Leclair et al. (1997) observed with dune-cross-set preservation, Alexander et al. (2001) found that the thickness of antidune lenses related to the depth of scour associated with larger amplitude antidunes and not with bed aggradation rate; indeed, Alexander et al. (2001) found it impossible to distinguish between antidune structures from aggrading and non-aggrading beds with the same flow conditions. This is unsurprising when the rate of scour

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**Fig. 16.** Lens length versus lens thickness from structures in the 13 to 14 October 2012 flood deposit. The linear trendline and \( R^2 \) value indicate a strong correlation between lens length and lens thickness. However, the relationship between lens length and lens thickness is best described by a power law equation. The slight variation around the trend may relate to small variations in the angle between wave axis and channel axis (i.e. not all waves are perpendicular to channel axis), variation in water-surface wave behaviour (breaking, migration) influencing lens base geometry and lens preservation.

\[
y = 0.1727x^{0.858} \\
R^2 = 0.983
\]

\[
y = 0.122x + 0.035 \\
R^2 = 0.927
\]
and antidune growth following wave breaking in the flume experiments is orders of magnitude faster than mean bed aggradation rate. The same is very likely to be true in most natural settings, where scour and fill will cause near-instantaneous local bed level change rate on the order of 1 to 10 mm sec\(^{-1}\), whereas mean bed aggradation rate, even in the rapidly aggrading Belham system, is 0-003 mm sec\(^{-1}\).

The preserved lens length should increase with parent water-surface wavelength if other conditions are held constant; however, flow was rapidly varying and sediment-laden. Sediment transport during the breaking of hydraulic jumps (bores and water-surface waves) was observed at the camera site during the 13 to 14 October 2012 event. The most visible change was an increase in cobble-boulder transport corresponding with increased discharge, turbulence and bore incidence. In a positive surge moving downstream, if the depth of the bore increases, the ratio of conjugated depths (difference between depth of initial flow and bore depth) increases, and the potential energy loss at the hydraulic jump increases (Cartigny et al., 2014). To satisfy the conservation of mass and momentum, the velocity of the bore must increase with increased bore depth, if initial flow conditions do not change, and this leads to an increase in bore Froude number (Chanson, 2004). As the conjugate depth ratio increases and bore Froude number increases, lens preservation will decrease because higher \(Fr\) corresponds with increased hydraulic jump size, strength and competence (Khezri & Chanson, 2012b), so that breaking bores and water-surface waves with higher \(Fr\) are more destructive of the bed surface. Alexander & Cooker (2016) discuss the reason for this greater entrainment of large particles in rapidly changing flow.

Bores observed in this study had Froude numbers ranging from 1.11 to 1.47, corresponding with the undular (\(Fr < 1.3\)) and breaking, turbulent (\(Fr > 1.4\)) bores observed by Khezri & Chanson (2012a,b); whereby turbulent breaking bores have been documented to set particles in motion (Khezri & Chanson, 2012b). The bores during the 13 to 14 October 2012 flood appeared most turbulent and more frequently during rising and peak flow stages, and it is thought that these bores exerted more energy to remobilize parts of the bed surface. Given the scale of the bores relative to the water-surface waves, it is likely that they were a major control on lens preservation and the maximum clast size deposited during rising and peak flow stages. These periods are concurrent with the largest water-surface waves measured at the camera site; waves during these periods were quicker to break after formation and broke more violently than waves during waning flow stages.

The largest lens preserved in the deposit (2.92 m) relates to flow velocities between 2.85 m sec\(^{-1}\) to 3.46 m sec\(^{-1}\) (assuming 38 to 56% preservation). This exceeds the maximum observed flow velocity on 14 October, but just under that estimated for 13 October of 3.61 m sec\(^{-1}\) (Table 3). The largest lens was two standard deviations larger than the mean lens length, while the occurrence of water-surface waves on 13 October equating to velocities of 2.85 to 3.46 m sec\(^{-1}\) was between one and two standard deviations greater than the mean flow velocity on that day. Flow velocities on 13 October, consistent with the formation of the largest lens size, occurred much more frequently than their preservation would imply. Mean lens length (1.31 to 1.98 m) corresponded with mean wavelenght (±1 standard deviation) for times shown in Table 6. The percentile values show that mean lens length is representative of mean conditions measured at the camera site at the onset of flow (16:30:00 to 16:45:00 UTC 13/10/2012) and in waning stages (from 17:15:00 UTC 14/10/2012). The mean lens length underestimates mean wavelength between these times, but is still within range of observed water-surface wavelength. The occurrence of bores was below the average number per hour (108) for all records in Table 6 after 18:30:00 UTC 13/10/2012, and very few bores (<10) were observed after 18:00:00 UTC 14/10/2012, corresponding with declining discharge and mean water-surface wavelength (mean flow velocity).

Estimates of velocity from the largest boulder in the deposit (Table 3) corresponded well with average flow conditions on 13 October and were closer to maximum flow conditions on 14 October. However, the largest clast was nearly three times the size of the next largest in the event deposits (of 0.18 m), equating to flow velocities of 1.78 m sec\(^{-1}\) and 1.86 m sec\(^{-1}\) (the Alexander & Cooker, 2016; and Clarke, 1996; methods, respectively). The Alexander & Cooker (2016) method is considered more applicable than the Clarke (1996) method in these settings, because the method considers forces in unsteady flow, even though acceleration could not be measured. This figure underestimates the mean flow velocity on 13 October but is close to the mean flow velocity on 14 October. In this case, the Alexander & Cooker (2016) method applied to
Table 6. Water-surface wave measurements at the camera site during 13 to 14 October 2012 flood event, when mean wavelength (±1 standard deviation) corresponded with mean lens length (1.31 to 1.98 m).

<table>
<thead>
<tr>
<th>Date Time</th>
<th>Number of wave trains measured</th>
<th>Mean wavelength (m)</th>
<th>Standard Deviation</th>
<th>Percentile score (x = 1.31 m)</th>
<th>Percentile score (x = 1.98 m)</th>
<th>Discharge (% of valley floor occupied by flow)</th>
<th>Bore frequency (number per hour)</th>
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the largest boulder in deposit gives a fair estimate of velocity, but the other boulders in the deposits were much smaller and would have misrepresented the velocity in the flow. The boulders also generate velocity estimates less than the observed peak flow velocities, supporting the idea that most if not all of the event deposit was formed after the peak flow and during waning conditions. The lens size does not rely on the size of sediment available and therefore may be a more universally useful means to estimate flow conditions than maximum clast size.

This study confirms that lens architecture can be related to antidunes not only for sand beds (cf. Alexander et al., 2001; Duller et al., 2008) but also for gravel. This demonstrates the potential to use the ratios of water-surface wavelength (and antidune wavelength) to lens size as a record of time-averaged conditions during a flood event. Instantaneous flow metrics can otherwise be difficult to measure, even within a flume environment, in flows in which there is spontaneous wave breaking and rapid feedback between the flow and bed, fast, localized variation in depth, velocity and suspended sediment concentration. The application of Alexander et al. (2001) ratios in this study demonstrates that mean lens length may underestimate mean flow conditions in rapidly varying natural channels. More data are required from different natural settings to improve application of the relationship between antidune length and preserved lens length. It is not known how lens preservation varies with patterns of wave breaking, sediment mobility or Froude number, because the architecture of few antidune beds has been documented.

CONCLUSIONS

Numerous trains of water-surface wave were observed in the October 2012 flash flood in the Belham Valley on Montserrat and these played an important role in sedimentary processes. Water-surface wave trains are prevalent in other similar floods in the Belham Valley and in a multitude of other environments.

The October 2012 event deposits are dominated by lenses that are interpreted as structures formed by antidunes developing and changing during the growth, breaking and dissipation of stationary water-surface wave trains. The characteristics of the gravel lenses are similar to the sand lenses formed by antidunes in flume experiments (cf. Alexander et al., 2001). This confirms that even where the antidune bedforms are not preserved, diagnostic structures may be recognized and lens architecture can be related to antidunes in both sand and gravel deposits.

Mean lens size in a bed is a function of both antidune wavelength (and thus the water-surface wavelength) and the degree of antidune preservation. Thus, lens length, measured parallel to the flow direction, as indicated, for example, by grain fabric, can be used to estimate palaeoflow velocity and depth, through the relationship between antidune wavelength to water-surface wavelength. Using the preservation fractions of 0.38 to 0.56 calculated by Alexander et al. (2001) for sand lenses in a flume experiment applied to the gravel lenses observed in Belham Valley event deposit underestimates average flow conditions over an entire event and is nearer to that on the afternoon of 14 October 2012. Flow velocity estimated from maximum clast size in the same deposit gives velocities close to the mean on 13 October and the maximum on 14 October, confirming that the deposit mostly or entirely formed after the peak flow. It is likely that most if not all of the lenses formed during waning flow (with peak flow lenses having been reworked), although it is also possible that the preservation of gravel lenses was less than sand lens preservation in the Alexander et al. (2001) flume experiments. This demonstrates that antidune lenses are a good indicator of flow conditions (although they may underestimate peak flow) and, because they do not rely on grain-size availability, they may be more universally applicable than competence methods in some settings.

ACKNOWLEDGEMENTS

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NOMENCLATURE

$\alpha$ Acceleration of flow (cf. Alexander & Cooker, 2016)
$\beta$ Bed slope angle (radians, cf. Clarke, 1996; Stokes et al., 2012)
REFERENCES


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Supporting Information

Additional Supporting Information may be found in the online version of this article:

**Video Clip S1.** Water-surface waves migrating upstream and individual wave breaking (Fig. 3C) from 18:06:40 to 18:07:27 UTC on 13/10/2012.

**Video Clip S2.** Water-surface waves breaking sequentially, the breaking of one individual wave in the train triggering the breaking of adjacent waves (Fig. 3C) from 18:06:40 to 18:07:27 UTC on 13/10/2012.

**Video Clip S3.** Water-surface wave train observed at site B1 (Fig. 3C) at 17:15 UTC on 14/12/2012.

**Video Clip S4.** Water-surface wave train observed at site B1 (Fig. 3C) at 17:30 UTC on 14/12/2012.

**Video Clip S5.** Downstream propagating, channel-wide turbulent surge (bore) (Fig. 3C) from 16:53:30 to 16:54:56 UTC on 13/10/2012.