Channel response to an extreme flood and sediment pulse in a mixed bedrock and gravel-bed river

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ABSTRACT: We exploit a natural experiment caused by an extreme flood (~500 year recurrence interval) and sediment pulse derived from more than 2500 concurrent landslides to explore the influence of valley-scale geomorphic controls on sediment slug evolution and the impact of sediment pulse passage and slug deposition and dispersion on channel stability and channel form. Sediment slug movement is a crucial process that shapes gravel-bed rivers and alluvial valleys and is an important mechanism of downstream bed material transport. Further, increased bed material transport rates during slug deposition can trigger channel responses including increases in lateral mobility, channel width, and alluvial bar dominance.

Pre- and post-flood LiDAR and aerial photographs bracketing the 2007 flood on the Chehalis River in south-western Washington State, USA, document the channel response with high spatial and temporal definition. The sediment slug behaved as a Gilbert Wave, with both channel aggradation and sequestration of large volumes of material in floodplains of headwaters and reaches where confined valleys enter into broad alluvial valleys. Differences between the valley form of two separate sub-basins impacted by the pulse highlight the important role channel and channel-floodplain connectivity play in governing downstream movement of sediment slug material.

Finally, channel response to the extreme flood and sediment pulse illustrate the connection between bed material transport and channel form. Specifically, the channel widened, lateral channel mobility increased, and the proportion of the active channel covered by bars increased in all reaches in the study area. The response scaled tightly with the relative amount of bed material sediment transport through individual reaches, indicating that the amount of morphological change caused by the flood was conditioned by the simultaneous introduction of a sediment pulse to the channel network. Copyright © 2015 John Wiley & Sons, Ltd.

KEYWORDS: sediment slug; extreme flood; channel response; Chehalis River; bed material transport

Introduction

In gravel-bed rivers, bed material transport is a primary process that shapes channel form (Church, 2006) and controls river channel lateral mobility (Wickert et al., 2013; Constantin et al., 2014). Sediment supply pulses (increase in sediment supply to a reach irrespective of source or channel response) and the formation and dispersal of bed material slugs (coherent bodies of sediment interacting with the alluvial channel) are particularly important instances of bed material transport, which can cause dynamic changes in river channel morphodynamics (e.g. Nicholas et al., 1995; James, 2010).

Past work to document the behavior of large sediment slugs has focused on retrospective studies of bed waves using terrain markers, stream surveys (including specific gauge analysis), and sediment budget analysis (e.g. Meade, 1982; James, 1989, 1991, 1993; Marron, 1992). In contrast, work resolving sediment slug evolution in multiple time steps has focused on small-scale sediment slugs such as deposits of individual channel-blocking landslides (Sutherland et al., 2002; Cui et al., 2003; Hoffman and Gabet, 2007; Short et al., 2015).

Retrospective studies of mega-scale and larger sediment slugs (sensu. Nicholas et al., 1995) have highlighted the importance of overbank sediment storage in creating asymmetrical bed-wave and sediment-wave dynamics (Major et al., 1996, 2000; Major, 2004; Gran and Montgomery, 2005) and long relaxation times before disturbed systems return to pre-disturbance or new equilibrium conditions (e.g. Pierson et al., 2011). This situation, where overbank storage provides a key role in mediating sediment flux, has been given the name ‘Gilbert Wave’ by James (2006, 2010), in reference to the classic case history of the impact of hydraulic mining debris in the Sacramento Basin. In such cases, overbank storage results in asymmetry between the early passage of a bed wave (as measured in changes in the local bed elevation through time) and a temporally skewed sediment wave (as measured in elevated sediment flux).

The behavior of sediment slugs in natural rivers is strongly governed by the virtual velocity of the slug material and its
connectivity, both laterally between channel and floodplain (e.g., Croke et al., 2013) and longitudinally between channel segments (e.g., Benda and Dunne, 1997; Hooke, 2003; Nelson and Church, 2012; Kuo and Brierley, 2014; Ferguson et al., 2015). It is important to recognize, however, that downstream sediment connectivity is grain size specific and that sediment pulses may be separated by transport processes and deposited simultaneously as discrete slugs in multiple river segments.

A natural experiment, allowing observation of the evolution and impact of a large mega-slug (sensu. Nicholas et al., 1995) in unprecedented detail, was set up when an extreme flood and over 2500 individual landslides occurred in the upper Chehalis Basin of south-western Washington State, USA in early December, 2007. Abundant remote sensing data, including multi-temporal LiDAR (light detection and ranging) data and frequent aerial photograph coverage, document the channel response to this event with high spatial and temporal definition. Here, we exploit this natural experiment to explore the influence of valley-scale geomorphic controls on sediment slug evolution and the impact of sediment pulse passage and slug deposition on channel stability and morphology.

The upper Chehalis Basin and December 2007 storm event

Basin context

The upper Chehalis Basin (Figure 1), includes the upper mainstem Chehalis, South Fork Chehalis, and Stillman Creek sub-basins. This portion of the Chehalis drains approximately 980 km² of the northern and eastern slopes of the Willapa Hills. The basin has moderate relief: 35% of the basin area has slopes greater than 30%, and the average and maximum elevations are 300 and 950 m, respectively. The hills are underlain by a combination of Eocene basalt, which contribute cobble, gravel, and fines to streams and Eocene-Pliocene clastic marine sedimentary rocks, which are generally friable and contribute sand to streams (Logan, 1987; Walsh, 1987). All three subject streams abruptly debouch from confined valleys in the Willapa Hills to broad alluvial valleys with substantially lower gradient (Figure 2). Land use in the hills is dominated by forestry for timber production, but these valley bottoms are used as farmland, typically separated from the river by a narrow riparian buffer of mature trees.

The December 2007 storm: high magnitude flood and widespread landslides

During the period of December 1–3, 2007, heavy rains and rapid snow melt in the upper Chehalis Basin (Figure 1) caused extreme flooding on the Chehalis River and tributaries and triggered widespread landslides in the river’s watershed. Following up to 36 cm of snowfall on December 1, a tightly focused and warm atmospheric river (Figure 3) melted that snow and dumped nearly half a meter of rain at gauges in the Willapa Hills (Mote et al., 2008; Mass and Dotson, 2010; Neiman et al., 2011) generating an extreme flood (estimated to be on the order of a 500 year recurrence interval). The flood swamped stream gauging equipment and so the United States Geological Survey (USGS, 2008) used high water marks to estimate a peak instantaneous discharge of approximately 1800 m³ s⁻¹ at Doty (Figure 1), which is more than twice the previous flood of record at that gauge. Two significant

Figure 1. Upper Chehalis Basin overview. Elevation Data USGS 30 m DEM (Gesch et al., 2009). This figure is available in colour online at wileyonlinelibrary.com/journal/espl
 (>10-year recurrence interval) floods occurred following the 2007 event in 2009 and 2012 (Figure 4).

During and immediately after the storm, saturated soils gave way in over 2500 landslides in the upper Chehalis Basin (Figure 5) (Sarikhan et al., 2008; Department of Natural Resources [DNR], 2010; Turner et al., 2010; Whittaker and McShane, 2012; Lingley et al., 2013). These slides were mostly shallow rapid slides or debris flows, originating in areas of...
recent forest harvest and along forestry roads (Turner et al., 2010; Whittaker and McShane, 2012) and terminating in the channel or floodplain of the Chehalis River and tributaries.

Airborne LiDAR elevation data from prior to the event (Puget Sound LiDAR Consortium (PSLC), 2005) and several years after the event (PSLC, 2012), were available covering much of the area of the mainstem river and tributaries; additionally, aerial orthophotos with complete coverage of the study area were acquired in 1990 (USGS), 2006, 2009, and 2011, and 2013 (United States Department of Agriculture, USDA).

The availability of these data sources bracketing this extreme event provides the opportunity to examine channel response to an extreme flow and sedimentation event. In particular, this case study gives the opportunity to observe details of the early-stage evolution of a large mega-slug.

### Methods

#### Landslide volume and texture

Landslide volumes were determined using maps delineating slide area, compiled from oblique aerial photographs shortly after the 2007 event and subsequent orthophotos, and ground observations to determine typical depths. The Washington DNR conducted an aerial reconnaissance to map landslides in parts of the Chehalis River basin following the 2007 storms (Sarikhan et al., 2008) and distributed these data in a geographical information system (GIS) shapefile (DNR, 2010). This shapefile was compared to 2008 orthophotos (DNR, 2008) and edited to provide an updated coverage of landslides from the 2007 storm. This coverage was annotated with estimated percent delivery to a stream.
for each slide, based on slide deposit proximity to mapped streams, and percent vegetative cover in 2008 and 2013 to determine revegetation and stabilization of the slide scars through time.

Field observations to evaluate typical landslide depths and texture were collected in September 2011 along 20 miles of the Weyerhaeuser 1000 mainline road, which follows the mainstem Chehalis upstream of Pe-Ell (Figure 5). Key aspects of this work are summarized here, but additional details are described by Watershed GeoDynamics and Anchor (2012). Slides were classified into two categories: debris slides and debris flows. Slide depths and connectivity of the slide to the channel network were estimated for each of the 24 separate slide features that crossed the road or was readily visible. The results of this assessment were extrapolated through the study basin based on landslide area and type to estimate the volume of sediment contributed to the channel network. The procedure multiplied the area of each mapped landslide by the average depth and delivery ratio for that type of slide. Slide volumes were converted to weight using an average bulk density of 1.04 t m$^{-3}$ (USDA Natural Resources Conservation Service [NRCS], 2013).

Sediment inputs from the landslides were partitioned into bedload and suspended load based on grain size properties of the average grain size distribution of three grab samples from representative landslides that crossed the Weyerhaeuser 1000 mainline road and the texture of dominant soils in the basin (USDA NRCS, 2013).

Channel response

LiDAR differencing

LiDAR data from 2005 and 2012 (PSLC, 2005, 2012) allows evaluation of both local patterns of erosion and deposition and the total change in sediment storage within the study reaches. Detection of geomorphic change between raster surfaces is a somewhat complex process that depends on the confidence in elevation estimates for each individual raster (e.g. Wheaton et al., 2010). To help address this, the analysis used a simple and conservative method to filter out digital elevation model (DEM) uncertainty and find areas of ‘real change.’ First, all areas with absolute change in magnitude < 30 cm were filtered out based on the assumption that these changes are due to noise in the LiDAR data (the 2005 LiDAR dataset had errors relative to ground control points of up to 25 cm; PSLC, 2005) or due to local agricultural grading. This value was selected as a conservative value slightly higher than the typical errors for the LiDAR bare earth model relative to ground control points of ±12 cm and ±6 cm (±2$\sigma$) (PSLC, 2005, 2012 respectively). Second, the uncertainty ($u$) for each DEM cell was defined as:

$$u = x \tan \alpha \quad (1)$$

where $x$ is the DEM cell size and $\alpha$ is the local slope between adjacent grid cells. This approach gives an uncertainty value based on the range of elevations expected to lie within the DEM cell area. Cells were excluded from the analysis when the estimated change was less than the uncertainty for either compared raster. The resulting elevation change map shows areas of erosion and deposition and can be used to estimate the total change in storage component of the sediment budget.

Aerial photograph analysis

Wetted channel and bar positions were delineated from aerial orthophotos acquired in 1990 (1 m panchromatic, USGS DOQ, 2006, 2009, 2011, and 2013 (1 m color, USDA Farm Service Agency NAIP). Additionally, the position of the 2012 wetted channel was delineated from the LiDAR data. Channel conditions in 2006 and change from 1990 and 2006 – a period with typical hydrologic conditions (Figure 4) – were used as a baseline against which conditions and changes following the 2007 event could be compared.

Flows at the time of aerial photograph acquisition ranged from 0.9 to 2.6 m$^3$ s$^{-1}$ (Table I), which are all very low compared to the mean annual flow of 16.3 m$^3$ s$^{-1}$ and typical flood flows (two-year recurrence interval) of 264 m$^3$ s$^{-1}$. For each year prior to 2013, three zones were manually delineated along the channel: wetted channel, unvegetated bars, and areas outside of the active channel (hereafter ‘floodplain,’ although this may include some upland areas). This mapping directly yields some basic morphologic parameters, including active channel width, proportion bar area, and low-flow wetted channel width. The 2013 dataset became available after the channel delineation was complete, but local observations from this dataset supplement the analysis.

Morphodynamic analysis

Combining the LiDAR differencing and morphologic mapping for each period between aerial photographs yields a picture of the morphodynamics in each reach. Overlying the channel zone polygons gives a view of surface state change through time, including areas of bank erosion (floodplain to wetted, floodplain to bar, and bar to wetted), accretion (wetted to bar, wetted to floodplain, and bar to floodplain), and no change. These changes are then combined with the LiDAR difference map to show the total volume of material eroded and deposited in the 2006 channel, in the area eroded between 2006 and 2011, and on the floodplain.

An additional set of useful descriptive metrics derived from the earlier mentioned analysis includes reach-average lateral bank migration rate (which may include erosion of both banks and does not necessarily imply a shift in the position of the centerline or thalweg), deposition, and vegetation colonization rates through time.

Bed material grainsize

The grain size distributions of 28 gravel/cobble bars along the Chehalis River and 5 bars along the South Fork Chehalis River were sampled during September 2010 to characterize the bed material through the study area at that time. Unfortunately, no grain size data are available from prior to the 2007 storm. Both armor and sub-armor layer samples were taken from bar head locations believed to be representative of mobile bed material (e.g. Klingeman and Emmett, 1982; Parker et al., 1982).

The armor layer sample followed the Wolman (1954) pebble count method ($n = 100$). The sub-armor layer sample was taken by scraping away the armor layer and excavating approximately 50 kg of underlying material by shovel. Particles larger than 32 mm from the sub-armor layer were sorted and weighed.

| Table I. Aerial photograph acquisition dates and river discharge |
|---------------------------------------------------------------|
| **Photograph acquisition date** | **Chehalis mainstem discharge at Doty (m$^3$ s$^{-1}$)** |
| July 10–15, 1990 | 2.0–2.6 |
| July 21, 2006 | 0.9 |
| August 1, 2009 | 1.3 |
| August 20, 2011 | 1.0 |
| August 14, 2013 | 1.8 |
in the field, and smaller fractions were retained for laboratory sieve analysis.

Bed elevation
Changes in river bed elevation at selected locations were evaluated at three sites in the upper Chehalis Basin. Specific gauge analysis at these sites tracked changes in water level through time for a particular flow (e.g., Jemberie et al., 2008) to indicate trends in bed elevation through time. These data were compiled for three USGS gauges in the study area, two on the mainstem at Doty and Adna (Figure 1), and one on the South Fork Chehalis River above the confluence with Stillman Creek (Wildwood). Stage records for the Chehalis River gauges were filtered for flows between 1.7 and 2.0 m$^3$ s$^{-1}$ at Doty (only stage is recorded at Adna) and for the South Fork for flows between 0.4 and 0.5 m$^3$ s$^{-1}$.

Study reach delineation

The results and discussion are organized at the scale of study reaches that were selected to characterize the channel response along the profile of the Chehalis and tributaries based on notable changes in valley morphology (Figure 6). Analysis sub-reaches were defined within some of these reaches at breaks in LiDAR coverage or more subtle changes in morphology.

One reach (South Fork Headwaters) was selected to illustrate the channel response in a sedimentation zone (sensu, Church, 1983) of a low-order tributary stream directly coupled to landslide inputs. It is located on Hanlan Creek, a third-order (13.6 km$^2$ basin area) stream tributary to the South Fork Chehalis River and illustrates a response typical of headwaters sedimentation reaches. The reach is below an abrupt break in gradient from 10% to 3% and is upstream of a constriction where the valley width decreases from approximately 100 m to approximately 10 m. No LiDAR data were available at this site and so interpretation is based only on aerial photographs. Another reach (The Mainstem Headwaters Reach) includes 13 miles of mainstem river through the confined valley upstream of Pe-Ell. It was selected to illustrate changes typical of higher-order streams above the mountain front.

Two reaches, Boistfort on Stillman Creek and Pe-Ell on the mainstem Chehalis, represent channel response at the mountain front where the streams exit the Willapa Hills. The bed material in both of these reaches is dominated by cobble-sized material. The Boistfort reach on Stillman Creek extends from its confluence with the South Fork Chehalis River to the mountain front. Upstream of the reach, Stillman Creek flows through a narrow valley and has a gradient of approximately 2%. The Boistfort Reach flows through a 150 m wide floodplain slightly incised into a large alluvial fan, which protrudes into the larger South Fork Chehalis River Valley. The reach’s slope gradually decreases from ~1% at the head of the fan to ~0.3% at the confluence of Stillman Creek and the South Fork Chehalis. Bridges and their abutments locally confine the flow, but there are no major bedrock controls on the channel gradient or alluvial valley width. In contrast, the Pe-Ell reach on the mainstem Chehalis occurs downstream of an abrupt break in slope where the river gradient decreases from approximately 0.5% to less than 0.1% (River Mile [RM] 107 in Figure 2). It includes a locally wide floodplain zone between the confined upstream valley and incised channel downstream (Figure 6).

Figure 6. Reach locations, surface grain size distributions (2010 data), and valley elevation relative to the low-flow water surface, created using the 2012 LiDAR dataset following the method of Jones (2006), except upstream of Pe-Ell where the 2008 LiDAR dataset was utilized. Note prominent constrictions at RM 101, 93.5, and 87 and relatively incised nature of the mainstem Chehalis above RM 90, with the exception of the small open area in the vicinity of RM 105. This figure is available in colour online at wileyonlinelibrary.com/journal/espl
Downstream of the mountain front, reaches were classified either as transport reaches or response reaches. Transport reaches are delineated where the channel boundary is mixed bedrock and alluvium, where the valley is confined and gradient controlled by bedrock, or where both conditions exist (Figure 6). Transport reaches occur along the mainstream between USGS River Mile (RM) 104.4 and 91.3 (Doty and Leudinghaus Reaches) and between RM 89 and 87 (Lower Valley Reach). Surficial bed material in the Doty Reach is dominated by cobble and gravel bars with substantial areas of bedrock and boulder exposure, the Leudinghaus Reach is gravel dominated with approximately 30% cobble sized material and local bedrock exposures, and the Lower Valley Reach is gravel dominated with approximately 12% cobble sized material.

Response reaches are defined where the channel boundary is primarily alluvium and the channel is free to migrate laterally across the valley bottom, typically forming a meandering planform (Figure 6). Two distinct response reaches were studied downstream of the mountain front on the mainstem Chehalis. The Ceres Hill Reach is positioned upstream of the confluence with the South Fork between RM 91.3 and 87.7, and the Adna Reach (85.8 to RM 82.5) is located at the head of a continuous alluvial valley that extends downstream out of the study area to the vicinity of Centralia (RM 66, Figure 1). At Centralia, bedrock grade control reduces the upstream gradient, prohibiting downstream passage of gravel (inferred from a sand/silt bed for 10 km upstream). Surficial bed material in the Ceres Hill Reach at present consists of ~95% gravel with sand sheets over point bars, and in the Adna Reach consists of approximately 5% cobble material and 70-95% gravel, with the remainder sand. In addition, one response reach (Curtis) was studied on the South Fork Chehalis River downstream of the confluence with Stillman Creek; the bed material in this reach is dominated by gravel with approximately 20% cobble sized material.

Results

Landslide volume and texture

The average slide depth of the 24 features measured was 1.8 ± 1.1 m for debris slides and 2.0 ± 0.8 m for debris flows. The average amount of material from each slide that was estimated to be delivered to a stream was 57 ± 32% for debris slides and 90 ± 20% for debris flows (values ± 1 σ). Grain size samples of the landslide material indicate that 42% consisted of fines (sand and smaller), 46% consisted of gravel, and 12% consisted of cobble-sized material.

Application of this grain size distribution and these depths and delivery ratios results in the total estimated sediment yield to the channel network shown in Table II. Specific event yields in the study area range from a minimum of 2600 t km⁻² on the South Fork Chehalis River just above the confluence with Stillman Creek to a maximum of 22 000 t km⁻² where the mainstem Chehalis debouches from the mountains at Pe-Ell. These quantities are extremely high compared to typical specific sediment yields of 140 to 1300 t km⁻² yr⁻¹ for the basin area determined from suspended sediment measurements over the period 1962–1965 (Glancy, 1971) and a 38-year record of landslide activity (Weyerhaeuser, 1994a, 1994b), reflecting the extreme magnitude of the storm event.

Geomorphic change by reach

The following sections trace the impact of the landslide sediment, from headwaters reaches immediately below landslides, across zones of cobble and gravel deposition at the mountain front, and into distal reaches affected by finer grain size fractions. The focus of the discussion is on geomorphic change in sedimentation zones, both at the mountain front and further downstream, although changes in selected transport zones where repeat LiDAR data were available are also considered. Changes in individual reaches described later exist within an overall pattern, which is illustrated in Figure 7. Figure 7, which in many ways is the central exhibit of this paper, summarizes patterns of deposition and erosion and changes in lateral bank migration rate and channel width through time and provides context for the discussion of particular locations in the subsections that follow.

Figure 7A shows volumetric change between 2006 (pre-event) and 2012 (five years after event) for the study reaches, showing the along-channel patterns of sediment deposition. Average lateral migration rates for the flood (2006–2009) period were substantially higher than pre-flood (1990–2006) and post-flood periods (2009–2011) for reaches closest to the sediment source (Figure 7B) and suggests the slug moved downstream through time as the lateral migration rates and active channel widths (Figure 7C) increased in the 2009–2011 period farther downstream. These effects are discussed in further detail in the following sections.

Headwaters reaches channel response

The active channel width in the Mainstem Chehalis Headwaters Reach increased from 11 m in 2006 to 38 m in 2009 (Figure 7C). In the South Fork Chehalis Headwaters Reach, active channel width increased from approximately 15 m in 2006 (or less, as much of the channel margin is obscured by canopy cover in the available orthophoto) to nearly 60 m as established coniferous trees were ripped up and the entire valley bottom was inundated with sediment from debris flows and large wood rafts from both debris flow and local sources (Figure 7C). In most areas of the reach, the low-flow channel was confined to a single thread in 2009, but relief on the bar surface indicates that, during high flows, a braided channel would fill the entire valley bottom (Figure 8A). By 2011, the average active channel width had contracted to less than 30 m in the South Fork Chehalis Headwaters and 23 m in the mainstem headwaters, and By 2013 the South Fork channel had begun to downcut through the 2007 deposit, organizing the valley bottom into a

### Table II. Landslide volumes and sediment yields for the 2007 event compared with typical basin sediment yield

| Basin          | Total slide volume (m³) | Total slide weight (t) | Specific event yield (t km⁻²) | Typical sediment yield (t km⁻² yr⁻¹) |
|----------------|-------------------------|------------------------|-------------------------------|-------------------------------------|
| Mainstem Chehalis | 3 300 000               | 3 500 000              | 6400 at South Fork Confluence | 140 (1962–1965)                     |
| S. Fork Chehalis  | 370 000                 | 390 000                | 22 000 at Pe-Ell              | 1300 (1955–1993)                    |
| Stillman Creek   | 1 100 000               | 1 200 000              | 2600 at Stillman Confluence   | —                                    |
| Total           | 4 900 000               | 5 100 000              | 10 000 at South Fork Confluence | 200 (1955–1993)                    |

*Using bulk density of 1.04 t m⁻³ based on Natural Resources Conservation Service soil data for the study area (USDA NRCS, 2013).
Figure 7. Summary of along-channel morphologic changes in the study area. This figure is available in colour online at wileyonlinelibrary.com/journal/espl
distinct channel and floodplain where early seral vegetation rapidly established (Figure 8B), with the channel returning to a width similar to the 2006 channel (10–15 m). The Mainstem Headwaters Reach followed a generally similar pattern, increasing from 11 m width prior to the flood to 38 m in 2009 and then contracting back to 23 m by 2011. In between 2006 and 2009, alluviation locally smothered bed topography, with particularly notable effects where riffles composed of bedrock or colluvial boulders disappeared under thick gravelly deposits (cf. Zunka et al., 2015).

Mountain front channel response

The most pronounced sedimentation of all reaches in the study area occurred on the mainstem Chehalis at Pe-Ell (Figure 9). A total volume of 260 000 m$^3$ of material accumulated in this area, with a length-normalized net volumetric change of nearly 100 m$^3$ m$^{-1}$ in the reach between RM 106 and 107 (Figure 7A). Gravel was deposited in the channel, leading to bar formation and net bed aggradation (Figure 9A). Reach-average lateral migration rates in the period 2006–2009 (Figure 7B) ranged from 1 to 10 m per year and were expressed mostly as point-bar growth and reciprocal bank erosion with limited channel widening. Lateral migration rates decreased by 50 to 80% in the period from 2009 to 2011 following the pulse of sediment deposition. A large volume of sediment was deposited outside of the channel migration area and accumulated across the whole floodplain. The importance of bars in the morphologic structure of the river increased dramatically at Pe-Ell; bar growth accounted for nearly all of 50 to 60% increases in channel width (Figure 7C).

In contrast to the mainstem Chehalis at Pe-Ell, changes in slope and confinement are gradual along the Boistfort Reach of Stillman Creek. Additionally, there is no major valley constriction limiting the downstream extent of the sedimentation zone. These factors have resulted in less locally concentrated sediment accumulation over a much larger area along Stillman...
Creek (Figure 10). Length-normalized net volumetric change along Stillman Creek ranged from 20 to 65 m$^3$ m$^{-1}$, with peak deposition occurring 3–6 km downstream of the alluvial fan head, and a downstream-skewed distribution (Figure 7A). From the location of peak deposition downstream, the volume of sediment deposited on the floodplain dwarfed the volume of sediment deposited in the area where the channel migrated, while upstream lateral migration was more pronounced (6–10 m per year as opposed to 1.5–5 m per year downstream) and the volume of sediment deposited on point bars is comparable with the volume of sediment deposited on the floodplain. As with mountain front reaches at Pe-Ell on the mainstem Chehalis, the influence of bars and active channel width both increased dramatically along Stillman Creek (Figure 7C). The wetted width at low flow decreased slightly. In areas upstream of the peak of sediment deposition, the width of the active channel began to contract in the period from 2009 to 2011, while downstream of the peak, active channel width continued to increase (Figure 7C) and local lateral bank migration rates accelerated (Figure 7B).

Contraction of channel width and reduction of migration rates between 2009 and 2011 on the mountain-proximal reaches of both the mainstem Chehalis and Stillman Creek was accompanied by establishment of vegetation. This vegetation may ultimately stabilize a large proportion of the sediment pulse material in the floodplain.

Downstream mainstem reaches channel response

Important channel changes have also occurred downstream of the areas of maximum gravel and cobble deposition that are described earlier. In these areas, mechanisms forcing geomorphic change have included recent high water discharge, and deposition of both finer components of the landslide sediment (which may have been flushed rapidly through upstream reaches as suspended load), and locally derived sediment from bank erosion (which includes both gravel and fines). There is a distinct difference in the channel change in transport and response reaches. Channel alignment in transport reaches has been stable with net erosion caused by channel widening, while response reaches have migrated laterally, eroding floodplain and terrace surfaces and depositing material in growing point bars.

Transport reaches. Downstream transport reaches all show very similar morphodynamic changes (Figure 11). Both reaches with repeat LiDAR coverage experienced net erosion caused by channel widening and slight erosion from floodplain surfaces (Figure 7). Notably, much of the channel expansion in these reaches occurred through erosion of point bars, where shear stress may have been relatively low. This is interpreted to indicate that the erosion was a channel expansion to accommodate high flood discharge. The channel boundary expanded where the margin was most erodible rather than where shear stress was most concentrated on the outside of bends but bedrock and revetments prevented bank erosion.

In the Doty Reach repeat LiDAR coverage is not available. Planimetric channel changes, however, are very similar to those in the Leudinghaus Reach just downstream (Figures 7B and 7C). The USGS gauge at Doty shows modest change accompanying the 2007 flood, with specific gauge analysis indicating approximately 0.2 m of aggradation (Figure 12). In the Leudinghaus Reach, repeat LiDAR coverage shows very little growth of channel bars and floodplain erosion (in volumetric terms), relatively low lateral bank migration rates, and a slight (17%) increase in channel width, which was caused largely through erosion of the floodplain at the inside of point bars (Figure 11A). The influence of bars in the reach grew from 15% to 20% of the active channel area.

Response reaches. Downstream response reaches on the mainstem Chehalis (Ceres Hill and Adna) saw net length normalized aggradation of about 10 m$^3$ m$^{-1}$. In these reaches erosion was concentrated at the outside of meander bends and in the floodplain along the edges of terraces. Deposition was concentrated on point bars opposite bank erosion sites and also occurred in broad swaths across low-lying portions

Figure 10. Channel change along the Stillman Creek alluvial fan. This figure is available in colour online at wileyonlinelibrary.com/journal/esp
of the floodplain that had been inundated by flood waters (Figures 13 and 14).

Though bars in these reaches currently have a substantial fraction of surface sand (~5–10%) and are dominated by gravel, qualitative observations (Patricia Olsen, personal communication, 2013) indicate that the texture of bars in these reaches changed from cobble- and gravel-dominated prior to the 2007 flood to sand and gravel by 2013.

The average lateral bank migration rate in the Adna Reach increased from a baseline of 1.5 m per year (1990–2006) to 3 m per year over the period 2006–2009. In contrast to most other reaches in the study area, the lateral bank migration rate was faster between 2009 and 2011 (to 4 m per year) than it had been between 2006 and 2009. The width of the active channel and dominance of bars both increased (Figure 7C). Eroding banks in downstream sedimentation reaches are typically about 3 m higher than point bars growing across the river (Figure 6), and so the process of lateral migration erodes more volume from the floodplain than it deposits in bars within the active channel (Figure 7A). Thus, net mobilization of sediment from the floodplain by lateral migration can occur even in the absence of channel widening. This imbalance was, however, more than compensated by accumulation of fine sediment in low-lying floodplain areas, where up to 1.2 m of sediment was deposited, so that the overall change in storage in these reaches was positive.

Sediment slug dynamics

By integrating direct observations of morphologic change at the reach scale and observing along-channel patterns of morphologic change (Figure 7), it becomes possible to make inferences about the large-scale dynamics of sediment slug behavior associated with the 2007 event.

Along Stillman Creek (and the South Fork downstream of the confluence with Stillman), which flows in a relatively unconfined and gently concave channel crossing an alluvial fan, the disturbance pattern is coherent: deposition volume, lateral bank migration rates, and changes in active channel width all follow downstream-skewed distributions. Peak planimetric disturbance measured by three metrics – reach average lateral bank migration rate (10 m per year), change in active width (91% increase), and proportion of the channel occupied by bars (38% increase) – occurred between RM 96 and 98 in the period 2006 to 2009. In these areas between 2009 and 2011, vegetation establishment on bars outpaced erosion of new floodplain, leading to a reduction in channel width and proportion of the channel occupied by bars. The peak lateral bank migration rate between RM 93.4 and 94 occurred between 2009 and 2011; this is interpreted to indicate advancement of the gravel portion of the slug. If that interpretation is correct, then the virtual velocity of gravel composing the wave front was approximately 1.4 km per year.

Maximum aggradation along Stillman Creek between 2006 and 2012 occurred between RM 94 and 95.8, downstream of the maximum planimetric instability (RM 95.8–97), indicating a virtual velocity for the peak of the sediment wave of approximately 1.0 km per year. In the region of maximum aggradation, the lateral position of the active channel has remained relatively stable, but the channel has aggraded 0.6–1 m and 1.5–3 m thick sheets of gravel have been discharged onto the floodplain.

In contrast, the pattern of disturbance is relatively disordered along the mainstem Chehalis, where the long profile and valley width are influenced by bedrock. Transport reaches, where the channel is confined, were stable or showed net erosion due to channel widening. Sedimentation reaches, with broad alluvial floodplains, all showed substantial aggradation.
Observations of the bed material in the channel and channel’s morphodynamic response allow inference of the current (as of 2013 aerial photograph) position of the sediment slug in the system. The leading front of the gravel portion of the sediment slug has progressed approximately 25 km from the mountain front in the mainstem Chehalis River and 10 km from the mountain front in Stillman Creek; while the peak of disturbance has moved 8 km through the Chehalis River downstream of the mountain front and 3 km down Stillman Creek. Sand mobilized by the landslides has reached much further downstream, with the leading edge of disturbance moving out of the study area (>50 km from the mountain front) and the peak of disturbance occurring ~30–40 km from the mountain front along the mainstem Chehalis River. On Stillman Creek, the sand component of the wave has progressed into the South Fork Chehalis River approximately 13 km from the mountain front and the peak of disturbance is approximately 10 km from the mountain front. Comparison of these slug propagation rates for the mainstem Chehalis River and Stillman Creek with theoretical predictions confirms the important control channel scale exerts...
over sediment slug propagation rate. Beechie (2001) observed that sediment waves in gravel-bed rivers typically move at approximately a rate of 20 times the bankfull width of the channel annually, which on the mainstem Chehalis would translate to a rate of approximately 0.75 km per year and on Stillman Creek would translate to a rate of 0.45 km per year. The observed rates of the peak of the slugs’ disturbance are roughly similar to these theoretical predictions (20 to 75% higher) while the leading edge of the slugs have propagated much further downstream.

It is possible to construct rough sediment budgets considering the total estimated landslide volume, connectivity between landslides and the channel network, grain size distribution of the source sediment, and volume of aggradation downstream of the mountain front, as outlined in Table III. Most of the sediment moved by landslides in the Stillman Creek basin was remobilized and moved to the mountain front (75%), while a relatively small component of the material in the mainstem Chehalis River basin was transported downstream to the mountain front (19%). If it is assumed that the fraction of landslide volume delivered to streams and subsequently moved to the mountain front on Stillman Creek provides a minimum estimate of the fraction of landslide volume initially entrained by streams in the mainstem Chehalis basin, then it is possible to estimate that at least 30% of the connected landslide volume is coarse material that remains in storage along the mainstem Chehalis above Pe-El. This is consistent with field observations of sediment storage along and within the channel in the headwaters reaches upstream of Pe-El.

Discussion

Sediment slug evolution

Here we have focused on quantifying the geomorphic response of channel units impacted by the flood and sediment pulse, using coupled aerial photograph-based channel mapping and multiple LiDAR datasets bracketing the event. While most case histories of comparable events have focused on reconstruction from landscape and archival evidence, here remote sensing data are available showing the evolution and impact of large sediment slugs in fine detail. The subsequent discussion is structured to parallel the sediment pulse as it moves down valley and interacts with peculiarities of the channel network and considering impacts in various affected environments.

First we consider the importance of medium- to long-term storage of the pulse sediment in headwaters reaches, as illustrated by channel dynamics and sediment budget in the South Fork Headwaters Reach. Slug deposition caused valley wide aggradation, and the channel subsequently incised through this deposit, returning toward pre-disturbance channel dimensions and presumably grade over a period of about four years. In areas of relatively wide valley bottoms (such as the South Fork Headwaters Reach) this resulted in storage of large volumes of sediment, while in canyons and confined valleys only channel grade adjustments occurred. Much previous work has shown that sediment yield after landscape-scale disturbances typically decays exponentially over a period of decades to centuries, with half-lives for the volume of material temporarily stored in headwaters areas ranging from approximately 5 to 50 years in established examples (Adams, 1980; Pearce and Watson, 1986; Major et al., 2000; Dadson et al., 2004; Koi et al., 2008; Hovius et al., 2011; Huang and Montgomery, 2012), where half-life ($t_{1/2}$) is computed as:

$$t_{1/2} = \frac{\log 2}{\log \frac{V_i}{V_f}}$$

The variable $t$ represents the elapsed time of observation and $V_i$ and $V_f$ represent the initial landslide volume and landslide volume remaining in the upstream basin at the time of observation, respectively.

The pattern of exponential decay occurs because material mobilized by the disturbance becomes progressively more difficult for alluvial streams to reach as the most accessible deposits are carried away and as all deposits become stabilized through the formation of lag-armour deposits and establishment of vegetation. The sediment budgets constructed here allow calculation of the volume of sediment remaining in headwaters reaches of both Stillman Creek and the mainstem Chehalis, and therefore, the half-lives for alluvial export of landslide sediment from the headwaters. In the smaller and steeper Stillman Creek basin, where there is very little area like that illustrated by the South Fork Headwaters Reach, the half-life for storage of connected landslide sediment in the headwaters is extremely short – calculated to be 2.4 years using the values of connected landslide volume and remaining storage in Table III. In the large mainstem Chehalis basin, where the channel gradient is lower and the valley bottom has substantially more area for alluviation, the half-life is longer (17 years), but still relatively short compared to the examples cited earlier. The comparatively high rate of sediment export and short half-lives observed in the present study are likely results of the extremely high flow that accompanied initial sediment production, the high mobility and delivery ratio of debris flows that comprised many of the landslide events, and the relatively high flow event just two years after the initial 2007 flood. Sediment supply did not only increase (as, for example, the case of widespread seismic induced landslides), but flow energy available to transport sediment was also very high, allowing efficient evacuation of material from landslide deposits prior to possible stabilization by vegetation colonization or formation or armouring lag deposits by a sequence of moderate flow events.

Table III: Sediment budget for mountain front reaches and upstream basins

| Basin              | Total connected landslide volume (m$^3$) | Downstream aggradation (m$^3$) | Estimated washload (42%)$^b$ | Upstream connected storage remaining (m$^3$)$^c$ |
|--------------------|-----------------------------------------|-------------------------------|------------------------------|-----------------------------------------------|
| Mainstem Chehalis  | 2 000 000                               | 260 000                       | 110 000                      | 1 630 000                                     |
| Stillman Creek     | 690 000                                 | 370 000                       | 150 000                      | 170 000                                       |

$^a$Volumes presented here are less than those in Table II because these are reduced to show only the estimated volume connected to stream channels and available for fluvial transport.

$^b$This represents the volume of sand and finer material that must have been eroded from landslides to mobilize the deposited volume of gravel.

$^c$This represents a maximum estimate because the volume of gravel transported downstream away from the mountain front reaches is unknown (though presumed small).
The pulse in the Elwha system evolved as a typical dispersive times the typical volume of annual alluvial sediment transport.

ers, sediment pulses abruptly introduced on the order of 10 erial, a single discrete pulse may influence multiple zones along pathways for understanding sediment pulse impacts. In particular for sorting mechanisms and grainsize specific transport this case study highlights the importance of considering the pulse material into long-term storage in channel floodplains in mountain front sedimentation zone reaches. The material has been moved into storage by several factors. Channel aggradation occurred due to overwhelming sediment supply and a pattern of downstream-reduction in sediment transport capacity. This, in turn, increased the depth and sediment transport competence of overbank flows, and caused a large volume of bed material to spill into overbank areas. On Stillman Creek this pattern was dominated by gradual downstream reduction in sediment transport capacity, while along the mainstem Chehalis, backwater effects of local constrictions (including bedrock, temporary large wood jams (Entrix, 2009), and bridges) resulted in both more pronounced local aggradation and different responses in transport and sedimentation reaches.

Both mountain front reaches show that even in the midst of slug passage, the volume of sediment aggraded in the channel is dwarfed by the volume of overbank storage (Figure 7A). Such overbank storage is a factor differentiating river response to macro-scale sediment slugs (e.g. Sutherland et al., 2002; Hoffman and Gabet, 2007) from mega- and super-scale slugs, where major channel planform and valley floor adjustment occur. Further, overbank storage is a key mechanism in James’ (2006, 2010) Gilbert Wave model. This storage will regulate the long-term supply of sediment from the slug to downstream reaches. Gradual reworking of sediment stored in the floodplain will occur as the channels migrate, metering out sediment from the 2007 slug for a long period (decades to centuries) into the future during individual high flow events.

On the mainstem Chehalis, some volume of smaller grainsize fractions moved through the Pe-Ell Reach at the mountain front and were flushed rapidly through transport reaches great distances downstream. This highly mobile material (some fraction potentially moving as washload through upstream reaches) passed to downstream sedimentation zone reaches where it accumulated in a second discrete slug and caused major change in channel form and migration rate. Thus, this case study highlights the importance of considering the potential for sorting mechanisms and grainsize specific transport pathways for understanding sediment pulse impacts. In particular, a single discrete pulse may influence multiple zones along a river profile concentrated in regions where transport mechanisms change.

It may be particularly instructive to contrast the pattern of storage in the Chehalis River study area with that occurring after the historic dam removals on the Elwha River. In both rivers, sediment pulses abruptly introduced on the order of 10 times the typical volume of annual alluvial sediment transport. The pulse in the Elwha system evolved as a typical dispersive sediment wave (e.g. Lisle et al., 2001), with a limited volume of storage occurring only within the channel (East et al., 2015; Warrick et al., 2015), while in the present case a large fraction of the total pulse volume moved into long-term storage on the floodplains. The wave on the Elwha matched typical wave evolution as it has been modeled in many flume studies (e.g. Cui et al., 2003; Madej et al., 2009; Humphries et al., 2012; Podolak and Wilcock, 2013). Peak flows on the Elwha during the study period reported by East et al. (2015) and Warrick et al. (2015) were approximately half of a two-year recurrence interval flood, and so the river was not hydraulically connected to the floodplain to facilitate lateral sediment movement.

In the present case, the behavior of the observed sediment slugs is not well-described by the framework developed by analysis of flume studies, which typically explicitly neglect consideration of along-stream channel and valley topography and stream-floodplain sediment exchanges (e.g. Lisle et al., 2001). It does not conflict with this framework, but the primary factors controlling evolution of this sediment slug are those that, by necessity, are not typically considered in a flume environment. Pronounced deposition occurred in several discrete zones along the channel profile where decreased slope, confinement, or both factors reduced shear stress in the downstream direction. In-channel sedimentation at such locations combined with effects of an extreme flood and large wood jams to create strong channel-floodplain connectivity and resulted in Gilbert Wave-like sediment slug evolution.

Major interactions between channel and floodplain are, by definition (Nicholas et al., 1995), the factor differentiating mega- and super-scale sediment slugs from the more commonly observed and modeled macro-scale slugs. In addition to providing a major storage reservoir, floodplain connection also changes the hydraulic implications of slug deposition. Macro-scale slugs steepen the slope on the downstream limb of the bed wave increasing flow energy; this acts as a negative feedback mechanism leading to erasure of the bed wave. In the case of mega-scale slugs connected to floodplains, such as the one presently described, sedimentation in the channel results in reduction in channel depth, increase in channel width, and an increase in flow connection to the floodplain, and potentially increased meander amplitude. These are all factors that counteract the impact of increased relief between the bed wave crest and downstream reaches that may lead to reduced flow energy in reaches impacted by slug deposition, potentially leading to persistent sedimentation in an impacted reach. Only when the scale of a slug becomes large enough to force channel straightening or increase the gradient of the valley floor does it again clearly increase flow energy in the impacted reach.

Channel response to elevated bed material transport and extreme flood

Finally, it is possible to consider what lessons may be learned from this case study considering the linkage between sediment and water supply and channel form and morphodynamics. Generally for sand- and gravel-bedded rivers, increased sediment supply increases channel width and dominance of alluvial bars, often resulting in formation of braided fluvial systems (Church, 2006; Mueller and Pitlick, 2014). Further, while it has generally been accepted that increased sediment supply results in decreased channel stability, recent research has shown quantitatively that lateral channel mobility (that is, the rate of lateral channel migration causing reworking of the floodplain surface, including erosion and formation of islands and floodplain areas) is positively correlated with increasing bedload flux over a broad range of channel forms (Wickert et al., 2013; Constantine et al., 2014).

Extreme floods alone, however, can trigger substantial geomorphic change (Schumm and Lichy, 1965; Everitt, 1968; Burkhart, 1972; Costa, 1974; Gupta and Fox, 1974; Stevens et al., 1975; Wolman and Gerson, 1978; Osterkamp and Costa, 1987; Pizzuto, 1994; Costa and O’Connor, 1995; Hooke, 1996; Lane and Richards, 1997; Friedman and Lee, 2002). But in the present case, increased sediment supply and the occurrence of an extreme flow must have acted together to trigger the observed responses. The extreme flow provided the energy
needed to effectively move and redistribute sediment, probably increasing the quantity of sediment mobilized into the main branches from low-order tributaries, the quantity of sediment conveyed downstream, and rate at which the slugs advanced downstream above that which could have occurred given only typical flood flows. Conversely, elevated sediment supply altered the impact of the extreme flood, especially in areas close to the mountain front. Debris delivered from upstream choked the channel during floods, and increased flow depths outside the channel to the point that these flows were competent to transport a large volume of the bedload mixture onto floodplain surfaces.

Observations of channel changes in the project area are consistent with the existing theoretical framework connecting sediment transport rate and channel morphodynamics. Channel migration rates increased by a factor of 10 in the mountain front gravel sedimentation zones, and by a factor of one to five in downstream areas, including both sedimentation and transport reaches (Figure 15A). Notably, although absolute rates were relatively modest, lateral bank migration rates in transport reaches increased proportionally more than in sedimentation zones. At the peak of slug impact, the active channel widened by 50 to >100% in mountain front reaches and by 10 to 20% in downstream sedimentation reaches (Figure 15B). Nearly all of this widening was manifested as an increase in the dominance of active bars; in sedimentation zones, bars typically grew into the wetted channel and caused reciprocal bank erosion, while in transport reaches the floodplain along the inside of point bars eroded, allowing channel widening without shifting the thalweg. On average across all studied reaches, bar dominance increased from approximately 20 to 40% of the active channel, with the greatest increase (from 5 and 20% to 30 and 50%, respectively) in reaches closest to the mountain front, through which the greatest amount of bed-material was conveyed (Figure 15C).

By calculating the volume of downstream deposition, it is possible to estimate the minimum bed material flux into and through a particular reach in the study area. By doing this, the geographic patterns observable in Figure 15 can be generalized. Figure 16 shows correlations between bed material transport into a reach (normalized to contributing drainage area) and change in three dependent geomorphic parameters: lateral bank migration rate, active channel width, and proportion of the active channel occupied by bars. All three dependent parameters show a strong correlation to the bed material flux through the reach. Though correlation does not necessarily imply causation, the pattern of increasing lateral migration rates through time with passage of the sediment wave front, even in the absence of major floods, suggests a causal relationship; and this agrees with recent observations from both flume and other field environments. The relation between lateral bank migration rate and bed material flux supports the result of Wickert et al. (2013) in a river environment very different than that considered by Constantine et al. (2014) and further strengthens the case that bed material sediment flux is one of the fundamental driving forces behind channel migration. The relationships between change in active channel width and proportion of the active channel occupied by bars also correlate tightly to bed material flux through the reach, providing additional empirical support for the notion that increased non-cohesive sediment supply will increase channel width and the dominance of alluvial bars (e.g. Church, 2006).

Additionally, the control of channel planform and morphology by bed material flux may help reconcile the common perception of bed wave translation with the reality of bed wave dispersion. Cui et al. (2003) note that the disturbance in sediment transport rate may translate along the leading edge of a bed wave, even when topographic evolution of the wave is dominantly dispersive. It bed material flux (and not bed elevation) predominantly controls channel morphodynamic response to a sediment slug and bed material flux disturbance translates downstream, then easily observable planform disturbance would be expected to translate downstream with the

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**Figure 15.** (A) Percent increase in erosion rate during slug passage over 1990–2006 baseline, (B) percent increase in active channel width during slug passage compared to 2006 condition, (C) change in percent of active channel occupied by bars during slug passage compared to 2006 condition. This figure is available in colour online at wileyonlinelibrary.com/journal/espl
leading edge of the dispersing slug, potentially leading observers to fallaciously interpret topographic translation of the bed wave itself.

Conclusion

Two broad themes have been considered in this paper: the dynamic evolution of sediment slugs acting as Gilbert Waves moving through one relatively simple (Stillman Creek) and one complex (mainstem Chehalis) channel and the impact of that elevated bed material transport on channel forms and lateral mobility. Downstream attenuation in the wave volumes occurred because large volumes of material were exported from the channel to the floodplain, providing stores of material that may be reworked for decades or centuries to come. The particular patterns of this slug deposition responded to the peculiarities of the flood hydrology and channel and valley form; a coherent pattern of disturbance is evident along Stillman Creek, while the mainstem Chehalis is divided into highly responsive sedimentation zones and transport reaches that appear to have rapidly flushed the pulse sediment downstream. Dramatic increases in channel lateral mobility, active channel width, and the relative dominance of alluvial bars all occurred in response to the event. Though high flows were likely important in providing the competence to move large volumes of sediment, close correlations between bed material sediment flux through reaches and their geomorphic response highlight the important role bed material flux plays in governing channel form and rates and patterns of morphodynamic processes.

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