Coalescing debris-fill complexes in headwater valleys of the Oregon Coast Range

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Submitted to Forest Science, February 1, 2006

Acknowledgements. This research was funded in part by the CLAMS Project, Pacific Northwest Research Station, USDA Forest Service. Shannon Hayes, John Green, Simon Mudd, Christine May, and Robin Beebee assisted with field work.

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Abstract

In headwater basins that are forested and where debris flows are common, debris-flow deposition forms valley-spanning dams of wood and boulders, and significant volumes of sediment are stored behind these dams. In three headwater streams (basin areas of approximately 2 sq. km) in the Oregon Coast Range, the following data were collected: (a) longitudinal channel profiles; (b) locations and heights of debris dams; (c) sediment storage volumes; and (d) exposed bank stratigraphy. In all three basins, these data indicate a zone of concentrated mainstem debris-flow deposition where stream gradients first decrease to approximately 10%. Deposits in this zone do not form fans per se but, rather, comprise coalescing debris-fill complexes. These complexes have large sediment storage volumes and represent a buffer between episodic sediment input from the hillslopes and continual sediment output to fish-bearing streams. A model of susceptibility to debris-flow inundation is based on hypothesized landslide susceptibility, an observed relationship between debris-flow runout length and elevation change between source and deposit, and tributary junction angles and corroborates the occurrence of coalescing debris-fill complexes where greater debris-flow susceptibility arises and is attributable to multiple sources, e.g., tributaries and main stem.

Keywords: Drainage patterns; Sediment storage; Natural dams; Network analysis; Debris flows
1. Introduction

Land-use practices (e.g., forest harvest and “stream cleaning”) have been blamed for the decline in numbers of salmonid fishes in the Pacific Northwest, but our understanding of the mechanisms linking land-use practices to fish abundance is still emerging. Specifically, sediment—both the right amount and the right kind, usually gravel—is one of the greatest “water quality” issues in the Pacific Northwest and other places (e.g., Jacobson 1995; Jacobson and Gran 1999), but the movement of gravel from headwater sources—in the Pacific Northwest, debris-flow deposits—to downstream, fish-bearing reaches is poorly understood. A better understanding of the transition between debris-flow and fluvial processes is necessary for understanding downstream effects of disturbance on aquatic habitat. Moreover, this transition is a key interface in the movement of sediment from sources in active orogens to sinks in sedimentary basins and has implications for the long-term evolution of longitudinal stream profiles and, therefore, relief in such orogens (Lancaster and Grant in press).

Restoration in mountain drainage basins of the Pacific Northwest often seeks to maintain moderate and steady supply, transport, and storage of gravel, which is required for salmonid spawning, e.g., by gravel-impounding wood emplacements of wood to improve gravel retention and landslide-preventing “leave areas” exempted from timber harvest (Ted Turner, Weyerhaeuser, Eugene, OR, pers. comm., 2003). Before most landslide-derived sediment can affect fish-bearing streams, it must be fluvially evacuated from headwater valleys, where the overwhelming majority of debris flows deposit—in large part because lengths of debris-flow-prone channels far surpass lengths of fluvially dominated channels (Stock and Dietrich 2003). Most debris flows initiated on hillslopes bounded by steep streams will likely be deposited in
those high-gradient reaches, and even debris flows that deposit in lower-gradient reaches must be removed by fluvial processes in order to contribute to the off-shore sediment budget.

The transition from debris-flow to fluvial processes represents a crucial linkage in the system of sediment delivery from hillslopes to larger streams (and off-shore) and must occur in numerous mountain landscapes (e.g., Taiwan, Hovius et al. 1997; Japan, Ishikawa 1990; and various mountain ranges in the USA, Eaton et al. 2003, Meyer et al. 1995), but the characteristics of this linkage are not well understood. In particular, while the existence of debris fans (analogous to alluvial fans but formed by debris-flow deposition) at junctions between debris-flow–dominated tributaries and fluvially dominated main channels is widely noted, such fans mark abrupt transitions in geomorphic process dominance. Where process transitions are smoother, styles of debris-flow deposition are different and have different implications for the morphology of those deposits and their valleys, both in longitudinal profile and in cross-section.

In forested headwater streams and valleys at the transition between mass-movement and fluvial processes, sediment delivery from the hillslopes to the channel and valley network is apparently dominated by mass-movement processes (Eaton et al., 2003), but sediment export is dominated by fluvial processes. Far from “chutes” that quickly transport all sediment supplied from upstream, these streams’ transport capacity is episodically overwhelmed by volumes of wood, boulders, gravel, and finer sediment that may linger for centuries or longer (Casebeer, 2004).

Debris fans exist where steeper, narrower debris-flow–prone tributary valleys join lower-gradient, wider valleys down which debris flows do not generally travel. In contrast, many debris-flow–prone valleys are not terminated by abrupt confluences with larger valleys and instead become gradually wider, less steep, and therefore less prone to debris flows. Where
numerous smaller tributaries feed such gradually changing valleys, debris-flow deposits do not form distinct fans but, rather, valley-inundating fills. Debris-flow deposition in such valleys in the Oregon Coast Range typically forms valley-spanning debris dams that are composed primarily of wood and impound significant sediment volumes, although some dams result from riparian treefall. These impoundments typically fill with sediment to form steps and force nominally bedrock channels to be alluvial (Montgomery et al. 1996, 2003; Massong and Montgomery 2000; Lancaster and Grant, in press). Whereas deposition on debris fans is likely relatively insensitive to changes affecting debris-flow dynamics because the changes in such fixed, morphological conditions such as valley width and gradient are so large and abrupt, deposition in valleys with more gradual changes in those morphological conditions is probably more sensitive to changes in unfixed conditions including climate, fire regime, forest harvest practices, and anything else that affects the balance between forces driving and resisting debris-flow runout.

Several aspects of the latter valleys distinguish them from other parts of the hillslope-fluvial system. First, sediments stored on the valley bottom appear to be floodplains and are genetically similar to floodplains in that they are formed by over-bank deposition, but the processes building these sediment bodies are largely debris-flow deposition or backwater, fluvial deposition upstream of debris dams. Second, these valleys have underlying bedrock strath surfaces apparently formed by lateral migration, but that migration is driven by avulsions due to debris-flow and wood deposition. Third, many valleys are aggraded and are typically five times wider than their channels but are generally part of an actively incising network. Finally, wood and boulders cause avulsions, control grade, and store sediment in these valleys qualitatively as in
lower-gradient streams, but quantities are typically far greater at the debris-flow–fluvial transition.

This paper investigates patterns of valley-filling debris-flow deposition in the southern Oregon Coast Range (Tyee Formation). We introduce the term, coalescing debris-fill complex, to describe the deposits in parts of the valley network where valley-inundating debris-flow deposits, typically from numerous sources, are particularly concentrated. These coalescing debris-fill complexes are described, and their existence as distinct features is concluded, using multiple lines of evidence. Field data include surveyed longitudinal channel profiles, debris-dam heights and locations, mapped debris-flow runout tracks, valley-bottom sediment volumes, and deposit stratigraphy. An analysis of digital elevation models (DEMs) based on the runout data appears to reproduce important aspects of the observed patterns of valley-bottom storage.

2. Field sites in the Oregon Coast Range, USA

Field work was sited in three drainage basins in the Oregon Coast Range: (a) a tributary to Cedar Creek (“Cedar Creek”), (b) a tributary to Hoffman Creek (“Hoffman Creek”), and (c) Bear Creek (Figure 1, Table 1). These basins were chosen for their similar lithologies and drainage areas but different shapes and the fact that neither valley-bottom nor mid-slope roads have affected debris-flow runout and deposition in the mainstem valleys (Swanson et al. 1977). The basins are underlain by massive, gently dipping, rhythmic interbedded sandstone and siltstone of the Eocene Tyee Formation except for the southern-most part of the Cedar Creek basin, which is underlain by intrusive volcanic rocks of an unnamed formation (Peck 1961), which may have thermally altered and thus hardened nearby sandstone. Topography in the basins is steep—valley sideslopes are typically ~40°—and highly dissected. Soils are shallow and have low bulk
densities (e.g., Reneau and Dietrich 1991, measured values of ~1 kg/m³ at a similar site in the Oregon Coast Range). The climate is maritime with warm, dry summers and mild, wet winters. Diffusive hillslope transport processes and debris flows deliver sediment to the valley network (as in, e.g., Dietrich and Dunne 1978; and Benda 1990).

In this part of the Oregon Coast Range, forest biomass is typically less than but of the same order of magnitude as the mass of the soil layer, especially in mature stands (Grier and Logan 1977; Sidle 1992; Heimsath et al. 2001). The forest is primarily Douglas-fir (*Pseudotsuga menziesii*) and secondarily western hemlock (*Tsuga heterophylla*), western red cedar (*Thuja plicata*), red alder (*Alnus rubra*), and bigleaf maple (*Acer macrophyllum*). Douglas-fir, western hemlock, and western red cedar typically reach breast-height diameters (dbh) in excess of 2 m in “old-growth” (i.e., older than ~250 yrs) stands. Wood is therefore a significant part of the mass moved by landslides and debris flows and an important structural element in the beds of stream channels, where it can remain for decades or longer (Hyatt and Naiman 2001). After wood decays out of debris dams, a boulder lag, which may persist indefinitely, often remains. Parts of all of the sites have been harvested in the last half-century, but at some sites, many large, cut logs were left in low-order channels and are found in many debris dams.

3. **Methods**

3.1 **Field Data**

The main channels of the study basins were surveyed with hand level and tape measure during summer, 2000. Steps in the profile were surveyed in detail, and notes were taken regarding their characteristics, whether in the bedrock itself or made of debris dams composed of wood, boulders, or a mixture of the two (i.e., wood-and-boulder jam). Dam height was defined as
the difference between the minimum of elevations upstream of the dam and the maximum of elevations downstream of the dam, i.e., plunge-pool depths and wood and boulder heights above channel beds were excluded. Profile surveys continued upstream, as feasible, to include all significant valley-floor deposits.

Stream gradients were derived from the surveyed profiles. In order to reduce the variability introduced by debris dams and impounded sediments, gradients were calculated between points at regular (10-meter) elevation intervals, proceeding from upstream to downstream. In practice, points were chosen as close to these elevation intervals as possible because surveyed points did not in general exist at these precise intervals.

Surveyed longitudinal channel profiles were matched to those extracted from digital elevation models (DEM) by transforming the extracted profiles to match known locations of tributary junctions and basin outlets. Contributing areas at every surveyed point were interpolated from DEM-derived values associated with points along the transformed, extracted profiles.

At intervals of approximately ten channel widths, valley cross-sections were surveyed, and the depths of sediment under the surveyed surfaces were found and integrated to calculate cross-sectional areas of sediment storage on the valley floor. Extensions of bedrock valley sides to the valley-bottom bedrock were inferred by extrapolating from the valley sideslope surfaces and assuming that the valley-bottom bedrock cross-section is horizontal (although that “assumption” is based on morphologic evidence on the ground, e.g., of exposed bedrock at the interface between sideslope and valley bottom) and valley sideslopes are 40°. At tributary mouths, the valley edge is the imaginary line connecting the upstream and downstream valley walls. Valley sideslope angles most often measured in the field were 40°, but angles as high as 60° were
measured for short distances upslope. This method therefore results in some under-estimation of
the storage volume due to using the sideslope surface for sideslopes with non-zero colluvial
depths and where bedrock sideslopes are over-steepened.

Where the channel bed was not bedrock, depth to bedrock beneath the channel bed was
estimated by linear interpolation between bedrock points up- and downstream of the alluviated
(or colluviated) section. For alluviated reaches that are too long for linear interpolation (e.g.,
>100 m), only discrete “wedges” (defined by a downstream debris dam and an upstream break in
slope, often at the base of another debris dam) were interpolated. Where significant “stacking” of
deposits occurs (common especially in the Cedar and Hoffman Creek sites), this method results
in underestimation of total storage. Unit storage is then defined as the cross-sectional area of
valley-bottom storage divided by the upstream basin (contributing) area and is, therefore,
dimensionless.

Descriptions and scale photographs of bank stratigraphy were taken at sites of valley cross-
sectional surveys. The characteristics of individual sedimentary facies in stream-bank exposures
reflect the processes responsible for deposition. Three different classes of deposit facies were
identified in the field as indicative of particular depositional processes: (1) Unsorted, angular,
matrix-supported clasts indicate deposition by debris flow (Collinson, 1978). (2) Sorted,
rounded, clast-supported gravels indicate deposition by high-energy fluvial processes. (3) Fine-
sediment layers interbedded with organics indicate deposition by low-energy fluvial processes.
The facies relationships, their relative arrangements in the valley, and the valley and deposit
morphology of these three simple types of strata indicate the mechanisms responsible for their
deposition.
3.2 Analysis of Digital Elevation Models

Network structure is a powerful control on sediment dynamics (e.g., Jacobson 1995; Jacobson and Gran 1999) and is currently called upon to explain observed differences in tributary influences on main streams, particularly the effects of debris flows (Benda et al. 2003, 2004a,b), but “network structure”, though it deserves quantitative treatment, is still often treated qualitatively (Richards 2002). Here, we employ a new DEM-based measure of susceptibility to inundation by debris flow.

The Channel-Hillslope Integrated Landscape Development model (CHILD), was modified to perform the DEM analyses. CHILD employs a triangulated irregular network (TIN) of elevations, obtained here through bilinear interpolation of a 10-m DEM derived from USGS 7.5’ quadrangle maps with 40-ft. (12.2-m) contour intervals. Locations of nodes in the TIN are random and uniformly distributed but conditioned on proximity to neighbors such that the distance between nearest neighbors is at least two-thirds the original grid spacing (i.e., nodes are at least 6.67 m apart).

Elevations on the original rectangular DEM were based on an inverse-distance-weighting algorithm (e.g., Weber and Englund 1992, 1994) which introduced terrace-like artifacts, where the flat parts of the “terraces” correspond to the contours of the original topographic map. For purposes of the calculations presented here, gradients were calculated along flow paths using nodes “at” (i.e., closest in elevation to) the original contours: for each node, our gradient calculation algorithm stepped downstream and upstream (along the path of greatest contributing area), as necessary, to find the appropriate nodes for gradient calculation; the elevations used were those of these down- and upstream nodes, and the distance used was that along the edges
(i.e., line segments connecting nodes in the triangulation) forming the flow path between those nodes (Dan Miller, Earth Systems Institute, personal communication, 2003).

Areas topographically susceptible to landsliding are found on the basis of gradient and contributing area, as in Montgomery and Dietrich (1994) and Dietrich et al. (1995), but unknown parameters such as soil depth and hydraulic conductivity are removed, such that we have

\[ I^* = \frac{b \sin \theta \cos \theta}{A} \left( 1 - \frac{\tan \theta}{\tan \phi'} \right), \]

where \( A \) is contributing area, \( b \) is contour width (node or pixel width in this context), \( \theta \) is slope angle, \( \phi' \) is effective friction angle, and lower values of \( I^* \) (m\(^{-1}\)) correspond to greater susceptibilities to landsliding. Contributing area is calculated according to single flow directions along the steepest edge connecting a node to its downstream neighbors, similar to the D8 algorithm of O’Callaghan and Mark (1984) and identical to that used in previous applications of CHILD (Tucker et al. 2001a,b; Lancaster et al. 2001, 2003). A failure probability is then assigned to each node:

\[ P_f = e^{-I^*/\lambda}, \quad S_{us} \leq S \leq \tan \phi' \]

\[ = 1, \quad S > \tan \phi' \]

\[ = 0, \quad S < S_{us} \]

where \( \lambda (= 0.0011) \) is a decay scale chosen such that \( P_f = 1/(60,000) \) when \( A/b = 16 \text{ m} \) and \( S = S_{us}, S = \tan \theta \); and \( S_{us} \) is the value of \( \tan \theta \) at the maximum in equation (1) \( (S_{us} = 0.380 \) for \( \tan \phi' = 0.90, \) or \( \theta = 21^\circ \) for \( \phi' = 42^\circ \)). Slopes less than \( S_{us} \) are assumed to be “unconditionally stable,” while slopes greater than \( \tan \phi' \) are assumed to be “unconditionally unstable” (Mont-
gomerly and Dietrich 1994). The values chosen for (1) \( \lambda \) and (2) \( \phi' \) are somewhat arbitrary but reasonable: (1) Typical landslide intervals for failure-prone hollows are 6000 yrs (Reneau and Dietrich 1991; Montgomery et al. 2000), and if \( P_f \) is the probability of landsliding at a site in a given year, then the probability of landsliding at slope angles corresponding to \( S_u \) is one-tenth that of “slide-prone” sites. (2) Side-slopes are typically 40° in the study areas, and the slope angle for unconditional instability should be higher than angles observed for soil-mantled slopes in the field.

Values of \( P_f \) are then “sent” downstream and a weighted amount “added” to nodes according to the weighting function, 

\[
f_w' = \left(1 - \frac{|X - \mu_X|}{\sigma_X}\right) \left(\prod_{i} \cos \alpha_i \right) \times L, \mu_X - \sigma_X \leq X \leq \mu_X + \sigma_X.
\]

where \( X = Z - L/R_{LZ} \), \( L \) is horizontal distance along the network and \( Z \) is the elevation difference between the potential initiation site and a downstream point, both in meters; \( \mu_X \) and \( \sigma_X \) are the mean and standard deviation of \( X \), respectively, calculated from the best linear fit between \( L \) and \( Z \) for debris flows mapped in the field (Figure 2); \( R_{LZ} \) is the slope of that fit (Figure 2); and \( \alpha_i \) are the direction angle changes at major tributary junctions (defined as points where contributing area at least doubles). The first term in parentheses is a triangle function centered on \( \mu_X \) and dropping to zero at \( \mu_X \pm \sigma_X \). It is multiplied by the cosines of angle changes at major tributary junctions, as in Lancaster et al. (2003), and the distance to the potential initiation point to account for entrainment or “bulking up” by debris flows along their runout paths. A measure of unit (i.e., per unit contributing area) debris-flow “impact” is then
\[ M_{df} = \frac{\sum_{i=1}^{N} f_{ui} P_f a_i}{A}, \]  

(4)

where \( N \) is the number of upstream nodes and \( a_i \) is the area of the \( i \)-th upstream node (on a TIN, as employed herein, the Voronoi area, defined as the locus of points closer to a given point than to any other).

4. Results

4.1 Field Study

The surveyed profiles are visibly stepped, and although steps in bedrock and indurated sediments do exist, debris dams account for most visible steps (Figure 3). While the elevation profiles are generally concave, the profiles showing only the cumulative rise of natural dams have apparent convexities, where zones of closely spaced dams contribute significantly to the relief of the elevation profile (Lancaster and Grant, in press).

Measured unit storage vs. distance downstream (Figure 4c) is highly variable for all three study basins, but trends are apparent in the envelopes. All three unit storage distributions have maxima near the upstream extents of measurement. Unit storage appears to taper off to some lower stationary level for Hoffman Creek at 1200–1600 m from the divide and for Bear Creek at either 500 m or 1700–2000 m from the divide. Unit storage in Cedar Creek does not appear to reach a stationary level.

Locations and extents of coalescing debris-fill complexes were estimated based primarily on the locations of convexities in and the extents of the associated steep parts of the cumulative dam rise profiles, and the estimated extents of coalescing debris-fill complexes were adjusted with reference to the measured unit storages (Figure 3, Figure 4). The unit storages were especially
useful for determining the upstream extents of the complexes because, although the locations of debris dams often correspond to maxima in unit storage, values of the latter are often large for some distance upstream of these dams. Longitudinal locations, gradients, and contributing areas of these hypothesized complexes are listed in Table 2, and their gradients and contributing areas, along with power-law fits, are highlighted in Figure 5. Stream gradients in these reaches are relatively large, as high as 19.1% (10.8°)—consider that Benda and Cundy’s (1990) empirical model requires debris flows stop at gradients of 6.1% (3.5°). These reaches have lower concavities than those of the entire profiles (although the difference is small for Cedar Creek), and gradients just downstream decline more steeply, i.e., those downstream reaches have greater concavities than the reaches hypothesized to contain coalescing debris-fill complexes (Figure 5, Table 1). For these three streams, the gradient range of the complexes appears to be positively correlated with contributing area such that, for steeper streams, coalescing debris-fill complexes occur at greater drainage areas (Figure 5, Table 2).

On the one hand, the apparent correlation between gradient and contributing area in reaches with coalescing debris-fill complexes inferred from the debris-dam–relief profiles seems to confirm common sense: If the channel is steeper, debris flows will reach further downstream. On the other hand, the coalescing debris-fill complexes do not span a uniform range of gradient. If they did, then the complex in, e.g., Cedar Creek would extend downstream to even greater contributing areas.

The facies profiles show the intermingling of debris-flow and fluvial facies (Figure 6). Fluvial deposits, both coarse (gravel) and finer were found in steep reaches far upstream in all three study sites; debris-flow deposits and their associated debris dams can impound otherwise steep reaches and lead to deposition of even fine sediments. Debris-flow facies are prevalent in
bank exposures along most of the lengths of all three main channels (Figure 6). Although the relative proportions and distributions of debris-flow and other facies vary among the three sites, the reaches corresponding to inferred coalescing debris-fill complexes have greater fractions of debris-flow facies in bank exposures. The fact of greater proportion of debris-flow facies in reaches with many large debris dams (Figure 3, Table 2) is not surprising since those debris dams are typically formed by debris flows. Still, the facies relationships do confirm that debris-flow deposits comprise much of the fill in these reaches. The composite facies containing coarse and, more rarely, fine fluvial sediments reveal the interplay of processes forming the complexes.

4.2 Digital Elevation Model Analysis

Shaded relief maps with inferred channel networks show the visually apparent differences in network structure among the study basins, maps shaded according to $P_f$ from equation (2) show quantitative differences in distributions of hillslope areas susceptible to landsliding, and maps shaded according to equation (4) illustrate the quantitative implications of those different network structures and distributions of landslide susceptibility, $P_f$, for the inferred susceptibility to debris-flow deposition, $M_{df}$ (Figure 7).

These inferred debris-flow susceptibilities are shown graphically for the main channels of each basin in Figure 4 along with DEM-derived longitudinal channel profiles, contributing areas, and unit storage volumes measured in the field. Calculated peaks in unit debris-flow susceptibility, $M_{df}$, appear to correspond to tributary inputs, and the locations of those peaks correspond to measured peaks in storage, although the relative magnitudes of measured unit storage peaks are not always predicted by the calculation of $M_{df}$. The latter calculation does appear to capture some of the shape of apparent envelopes in measured unit storage. There also
appears to be some correspondence between zones of greater dam relief and larger values of $M_{df}$ with the notable exception of Bear Creek. Field observations indicate that, where $M_{df}$ is greatest but measured unit storage is zero and dam relief is absent (Figures 3 and 4), the valley has been recently scoured by a debris flow, the deposit of which is evident as the peak in unit storage at 1000 m from the divide (2000 m from the outlet), and the valley downstream of this point has been partially scoured even more recently by a debris flow (Figure 8a), the deposit of which is evident as the peak in unit storage at 1300 m from the divide (1700 m from the outlet). Similarly, Hoffman Creek has been recently scoured by a debris flow upstream of 400 m from the divide (2400 m from the outlet) (Figure 8b).

5. Discussion

The data and analyses presented herein corroborate the existence of zones within headwater valley networks that are characterized by infilling by debris flows, zones we term coalescing debris-fill complexes. While the data and analyses also reveal the locations of discrete debris fans at tributary junctions, coalescing debris-fill complexes are characterized by debris flows that travel along the mainstem valley and have disparate sources. While these complexes may lead to unit storage maxima at or near tributary junctions, some of the debris flows contributing to those maxima reach those junctions from upstream along the main stem rather than from the tributaries themselves. A typical example is the debris-flow–runout track mapped in Bear Creek and shown in Figure 7.

The data indicate the existence of a local maximum in sediment storage that begins where stream gradients first decrease to approximately 10%. Although storage per unit contributing area may have other local maxima downstream, these downstream maxima are generally the
result of tributary input and often correspond to discrete fans. The more upstream maxima correspond to debris flows that traverse significant reaches of mainstem valleys.

These upstream loci of debris-flow deposition are essentially debris fans that fill the valley bottom, i.e., coalescing debris-fill complexes. While these complexes often have fluvial components corresponding to deposition in backwaters formed by debris-flow deposits, these upstream complexes are more predominantly composed of debris-flow deposits than sediment storage complexes further downstream.

Evidence for a distinct phenomenon includes the following. First, the sediment storage data for all three basins imply zones of high unit storage with distinct beginnings and endings (especially in the cases of Bear and Hoffman Creeks; Cedar Creek is so “compressed” that the downstream end is somewhat ambiguous). Second, these storage zones correspond to zones of steepening in the dam relief profiles (i.e., where dams are relatively high and close together). Third, these storage zones, in which debris-flow facies predominantly comprise the bank stratigraphy, are sedimentologically distinct from downstream deposits, in which fluvial facies have relatively greater representation.

Calculated debris-flow susceptibilities, $M_{df}$, appear to predict locations and extents of the coalescing debris-fill complexes reasonably well. While observed longitudinal profiles, dams, and storage volumes are temporal snapshots of an inherently variable system, $M_{df}$ represents temporally averaged conditions. As noted above, the locations of greatest disagreement between measured unit storage and calculated debris-flow susceptibility are in locations of recent debris-flow scour. While some debris-flow–scoured channels are like curved-bottomed troughs and show little evidence of past debris-flow deposition, the parts of the Hoffman and Bear Creek valleys that have low unit storage but large calculated debris-flow susceptibilities do show
evidence of past deposition. The aforementioned scoured reach of Hoffman Creek contains hard, indurated debris-flow deposits. It is likely that this reach contained other, more erodible debris-flow deposits prior to the last scouring event. The reach of Bear Creek from ~500–1000 m from the divide has been scoured to bedrock, but the bedrock valley bottom is flat rather than curved. It is likely that debris-flow deposition promotes such flat valley bottoms by forcing the channel to move laterally and, thereby, to carve a flat (strath) surface.

If coalescing debris-fill complexes persist over times that are greater than the times required for tectonic uplift to create the relief represented by the surveyed reaches, then we could infer that the relatively low concavities in reaches with coalescing debris-fill complexes are effectively responses to the larger rates of deposition by debris flows in those reaches. That is, higher gradients are maintained because they are required to both remove debris-flow deposits and erode bedrock beneath those deposits (Lancaster and Grant, in press).

Inferred adjustment of stream gradients over geologic time to coalescing debris-fill complexes combine with the fact of their formation in valley reaches with gradually varying gradients to have profound implications for the effects changes in the factors controlling debris-flow runout lengths, especially those changes that do not generally affect sediment transport capacities in the stream network. The most likely of such changes is decreasing wood volume in the initiation and runout zones of debris flows typically depositing in coalescing debris-fill complexes. Numerical modeling results of Lancaster et al. (2003) indicate that removing wood altogether from the Hoffman Creek site might increase average runout lengths by more than 400% and maximum runout length by more than 100%. Runout lengths of debris flows depositing in the steep valleys typically occupied by coalescing debris-fill complexes would likely increase far more than runout lengths of debris flows depositing on fans: Once a debris
flow reaches the distal end of a fan, valley width and gradient typically increase and decrease, respectively, abruptly and by a large amount, but valley widths and gradients do not change as abruptly or greatly downstream of current coalescing debris-fill complexes. As shown in Figure 5, however, stream gradients do decrease more “quickly” with increasing contributing area, i.e., the streams have higher concavity, downstream of the coalescing debris-fill complexes, and this higher concavity may be an adjustment of the fluvial system over geologic time to lower sediment supply by debris flows. If changes in, say, wood volume were to effectively send debris flows well into these more concave reaches, the fate and residence time of debris-flow deposits in such reaches would probably also change, but even the direction of that change is unclear. On the one hand, we might suppose that the greater discharges downstream would evacuate the deposits more quickly. On the other hand, we think it is more likely that the stream power available to evacuate such deposits would actually be lower in these downstream reaches because of their higher concavity. That is, stream powers adjusted to the evacuation of fewer debris-flow deposits might become overwhelmed by more debris-flow deposition. Stream reaches even further downstream might therefore actually experience a decreased sediment supply due to increased debris-flow runout lengths, though such an inference is speculative.

It is important to note that, while runout lengths of debris flows typically depositing on fans might be less sensitive to changes in the factors influencing runout length (e.g., wood loading), even small changes might be important because debris fans are effectively buffers between debris-flow–prone tributaries and larger streams. Small increases in runout lengths might make the bypassing of those buffers more frequent. Clearly, the sensitivity of sediment supply in streams of different sizes and contributing network structures to changes in factors influencing debris-flow runout lengths bears further study.
6. Conclusion

Multiple lines of evidence point to the existence of coalescing debris-fill complexes, which are valley-filling deposits composed of predominantly debris-flow deposits and also fluvial gravels and fine sediments. Heights and spacings of debris dams in surveyed longitudinal channel profiles, profiles of valley-bottom storage per upstream contributing area, relationships between stream gradients and contributing areas, and bank stratigraphy all support the existence of these complexes.

Such complexes are different from debris fans: While the latter are typically isolated, fan-shaped deposits at the mouths of debris-flow–prone tributaries to main streams not prone to debris flows, coalescing debris-fill complexes fill main-stem valleys of debris-flow–prone valley networks and typically form where deposition zones associated with several debris-flow–prone tributaries converge on a reach of the main-stem valley, although these complexes can also extend and branch upstream into tributaries. Their existence and extent are determined by local valley gradient, upstream network structure, and distributions of landslide-prone hillslopes. A new predictor of debris-flow impact on any given point in the channel network is based on theoretical and empirical relationships and appears to be a useful indicator of the likelihood of debris-flow deposition but does not differentiate between areas prone to both scour and deposition and those prone to deposition only. While scour may sometimes be associated with coalescing debris-fill complexes, they are predominantly depositional zones. The new predictor of debris-flow impact is generally inclusive of coalescing debris-fill complexes but does not predict their location exclusively. Rather, maxima of debris-flow impact may at any given time be—and for two of the field sites are—associated with valleys with low storage volumes.
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Figure 1. Top: map of the Oregon Coast Range with locations of Cedar Creek, Hoffman Creek, and Bear Creek sites (outlined with solid lines). Bottom: enlarged shaded-relief maps of the (a) Cedar Creek, (b) Hoffman Creek, and (c) Bear Creek sites. See Table 1 for elevation ranges.

Figure 2. Debris-flow runout length (horizontal distance) vs. elevation difference between initiation site and downstream end of debris-flow deposit for debris flows mapped in Hoffman (all except longest runout; Lancaster et al., 2003) and Bear Creek sites in summers of 1999 and 2000.

Figure 3. Longitudinal profiles and cumulative relief of dams vs. distance from the outlets for (a) Cedar Creek, (b) Hoffman Creek, and (c) Bear Creek sites surveyed in summer, 2000. Vertical exaggeration is 10 times (after Lancaster and Grant, in press) Locations of observed coalescing debris-fill complexes are shaded in gray.

Figure 4. Comparisons of DEM analyses and sediment storage data for Cedar (left), Hoffman (middle), and Bear (right) Creek sites. (a) Longitudinal profile (elevation vs. distance downstream from drainage divide) of main channel and contributing area at points along that profile from the TINs of Figure 7. (b) Unit influence of debris-flow deposition at points along the main channel, i.e., $M_{df}$ vs. distance downstream from the basin divide. (c) Cross-sectional areas of valley-bottom storage calculated from longitudinal channel and cross-sectional valley surveys. Locations of observed coalescing debris-fill complexes are shaded in gray.

Figure 5. Stream gradient vs. contributing area for main channels in study basins. Points identified as falling within reaches of observed coalescing debris-fill complexes (Figure 3) are
black (vs. gray for points outside of these reaches), and power-law fits to these black points are shown.

**Figure 6.** Stratigraphy at locations of valley cross-section surveys for Cedar, Hoffman, and Bear Creek sites. Heights of stratigraphic sections are normalized such that relative proportions of debris-flow (black), fluvial-gravel (dark gray), and fluvial-fine (light gray) facies are shown vs. distance from the outlet. White spaces indicate cross sections with no bank exposures. Arrows indicate observed coalescing debris-fill complexes. Note that each of the horizontal axes of have different scales.

**Figure 7.** Analyses of digital elevation models (DEMs) for Cedar (top), Hoffman (middle), and Bear (bottom) Creek sites. (a) Shaded relief map, with drainage network for points with contributing area greater than 1 ha, of triangulated irregular network (TIN) used in analysis. TIN was formed by bilinear interpolation of the original grid at points chosen at random (but conditioned not to be too close to neighboring nodes; see Lancaster et al., 2003). (b) Grayscale map of $P_f$ (equation 2), the relative probability of failure based on calculation of $I^*$ from equation (1). (c) Grayscale map of $M_{df}$ (equation 4), the unit influence of debris-flow deposition on valley storage for points along the channel network, and outlines of reaches with observed coalescing debris-fill complexes. Note that the scales of failure probability shading in (b) are the same for the three basins, but the scales of debris-flow influence shading in (c) are different for each basin.

**Figure 8.** (a) Debris-flow–scoured reach of Bear Creek at 1800 m from the outlet (1200 m from divide). This photo shows the reach downstream of the originating tributary’s confluence with
the main stem and upstream of its terminus (deposit). This is the debris flow mapped in Figure 7.

(b) Debris-flow–scoured reach of Hoffman Creek >2400 m from the outlet (<400 m from the divide).
### Table 1. Drainage basin characteristics for the study sites.

<table>
<thead>
<tr>
<th>Site</th>
<th>Description</th>
<th>Drainage area, km²</th>
<th>Elevation range, m</th>
<th>Circularity $\alpha$</th>
<th>$\beta$, $\theta$, $R^2$ from slope-area fits $\beta$, $\theta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cedar Creek</td>
<td>Right-bank tributary of Cedar Creek, 1.8 km upstream of Sweet Creek confluence</td>
<td>1.84</td>
<td>79–539</td>
<td>0.796</td>
<td>43.5, 0.445, 0.685</td>
</tr>
<tr>
<td>Hoffman Creek</td>
<td>Left-bank tributary of Hoffman Creek, 1.7 km upstream of Siuslaw River confluence</td>
<td>2.03</td>
<td>10–265</td>
<td>0.596</td>
<td>208.0, 0.631, 0.965</td>
</tr>
<tr>
<td>Bear Creek</td>
<td>Left-bank tributary of Knowles Creek</td>
<td>2.19</td>
<td>97–480</td>
<td>0.514</td>
<td>283.0, 0.612, 0.828</td>
</tr>
</tbody>
</table>

$\alpha = \frac{4\pi A}{P^2}$, where $A$ is drainage basin area and $P$ is drainage basin perimeter (Mayer, 1990).

$S = \beta A^{-\theta}$, where $S$ is stream gradient; $A$ is contributing area in m²; $\beta$ is the steepness coefficient with units, m²; and $\theta$ is the concavity index.
Table 2. Distances, gradients, and contributing areas in reaches of coalescing debris-fill complexes inferred from longitudinal channel profiles

<table>
<thead>
<tr>
<th>Site</th>
<th>Distance (m)</th>
<th>Gradient (%)</th>
<th>Contributing area (km^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>from outlet</td>
<td>from divide</td>
<td>upstream end range</td>
</tr>
<tr>
<td>Cedar</td>
<td>420–940</td>
<td>1720–1200</td>
<td>11.4 8.4–19.1</td>
</tr>
<tr>
<td>Hoffman</td>
<td>1560–2300</td>
<td>1210–470</td>
<td>8.7  5.6–11.9</td>
</tr>
<tr>
<td>Bear</td>
<td>1428–2070</td>
<td>1580–940</td>
<td>11.0  7.4–13.2</td>
</tr>
</tbody>
</table>
FIGURE 2

\[ L = (3.95)Z - 53.7, \ R^2 = 0.863 \]
FIGURE 4

(a) Cedar
(b) Hoffman
(c) Bear

contributing area, m²
debris flow susceptibility, m
unit sediment storage

distance from divide, m

elevation, m

x 10⁴

FIGURE 5

Cedar fit, $y = 25x^{-0.40}$, $R^2 = 0.48$

Hoffman fit, $y = 19x^{-0.43}$, $R^2 = 0.64$

Bear fit, $y = 12x^{-0.36}$, $R^2 = 0.41$