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Deciphering topographic signals of glaciation and rock uplift in an active orogen: a case study from the Olympic Mountains, USA

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ABSTRACT

Estimating recent patterns of erosion and rock uplift within Cenozoic orogens has proven difficult as signals of these processes have been obfuscated by Plio-Pleistocene glaciation. The topography of many mountain ranges integrates the effects of long-lived rock uplift, Late-Cenozoic climate variation, and post-glacial landscape adjustment. In this study, we employ a suite of topographic analyses to study the relief of an active mountain range on a sub-catchment scale in an effort to separate the long-term signal of rock uplift from perturbations due to shorter-lived climate signals. We focus on the Olympic Mountains, USA, where patterns of exhumation and glaciation have been previously estimated; however, our methods and results are broadly applicable to other orogens.

Our analysis shows that Plio-Pleistocene alpine glaciers and the Cordilleran Ice Sheet have reduced the elevations of channel profiles and created anomalously low channel relief in the Olympic Mountains. Large low-gradient areas formed at lower elevations where ice sheets were present and alpine glaciers widened and deepened valleys. In the more rugged core of the range, near-threshold hillslopes along the margins of the oversteepened glacially-carved valleys, dominate the range. This implies a strong Plio-Pleistocene glacial climate control on the topography over

26 the more recent evolution of the Olympic Mountains. However, the broad relief
27 structure of the range appears to still record the regional rock uplift pattern and is
28 suggestive of an east-plunging antiform, consistent with folding of the subducting
29 plate or underplating of accreted rocks.

30

31 1. INTRODUCTION

32 Landscapes are a record of tectonic and surface processes. Given the means
33 to decipher the signatures of individual processes acting within these landscapes, we
34 can use them to understand past and present tectonic deformation. This has been
35 done with demonstrable success in fluvial-dominated landscapes (e.g. Kirby and
36 Whipple, 2001; Duvall et al., 2004; Whittaker et al., 2008; Adams et al., 2016).
37 However, many tectonically active landscapes do not record purely fluvial signals
38 because high rock uplift rates mean that many Cenozoic mountain ranges have risen
39 to high elevations where lower temperatures and increased precipitation have led to
40 the formation of alpine glaciers.

41 Recognizing patterns of tectonic rock uplift from modern topography within a
42 heavily glaciated landscape is difficult (e.g. Brozović et al., 1997; Brocklehurst and
43 Whipple, 2007) for two reasons. (1) Glaciers have imprinted their geometry on a
44 landscape initially formed by fluvial processes, which mixes topographic signatures
45 (Whipple et al., 1999; MacGregor et al., 2000; Brocklehurst and Whipple, 2002;
46 2004; 2006; Anderson et al., 2006), and (2) erosion by fast-moving, temperate
47 glaciers can outpace erosion by rivers and in some cases rock uplift (e.g. Hallet et
48 al., 1996; Brocklehurst and Whipple, 2002). Such efficient glacial incision has
49 created a mismatch between Quaternary erosion rates and longer-term rates of
50 erosion and/or rock uplift (e.g. Moon et al., 2011; Godard et al., 2012; Glotzbach et
51 al., 2013; Herman et al., 2013) before the initiation of Northern Hemisphere
52 glaciation ca. 3 Ma. It follows that glacial incision can alter relationships between
53 topography and rock uplift rate that may have previously existed (Brozović et al.,
54 1997; Brocklehurst and Whipple, 2007; Yanites and Ehlers, 2012). It is, therefore,

difficult to use common topographic metrics to estimate patterns of rock deformation in glacial landscapes (Brocklehurst and Whipple, 2007; Moon et al., 2011).

Previous work has made significant progress in understanding how glaciers can modify valley profiles and surrounding topography. It has been suggested that glaciers have the ability to increase the topographic relief of a mountain range by inducing isostatic uplift of mountain peaks via significant mass removal (e.g. Molnar and England, 1990; Montgomery and Greenberg, 2000; Yanites and Ehlers, 2016). In addition, numerical models and observations have shown that glaciers are able to increase local relief by widening valleys, lowering channel elevations to a greater extent than rivers, and by forming hanging tributary valleys (Small and Anderson, 1998; Whipple et al., 1999; MacGregor et al., 2000; Brocklehurst and Whipple, 2002; 2006; Anderson et al., 2006; Yanites and Ehlers, 2012). Conversely, recent work shows glaciers can also reduce relief, particularly when they preferentially erode at higher elevations within a landscape (Brozović et al., 1997; Whipple et al., 1999; Brocklehurst and Whipple, 2002; 2006; Naylor and Gabet, 2007), or protect topography by armoring it with cold based glaciers (Thomson et al., 2011; Yanites and Ehlers, 2012). Previous studies also show that some glacial systems can simultaneously create high elevation, low-relief landscapes in the upper reaches, and deeply incised valleys further downstream (Herman et al., 2011; Steer et al., 2012).

Thus, we are left with two seemingly incompatible hypotheses for how glacial processes modify topography in tectonically active orogens. Should we expect that glaciers increase or decrease topographic relief? How would this change the relationship between relief and rock uplift rate? The aim of this study is to test the efficacy of these competing hypotheses to alter preexisting topography across a

range that has experienced a high spatial variability in Plio-Pleistocene equilibrium line altitudes (ELA), and where long-term erosion and rock uplift rates have been estimated from low-temperature thermochronometry and river incision (Brandon et al., 1998; Pazzaglia and Brandon, 2001). The diverse landscapes and glacial history of the Olympic Mountain range (**often referred to simply as the “Olympics”**) (Figure 1) allow us to compare and contrast the ability of glaciers with different erosive capabilities (e.g. variable ELA values) to reorganize orogen topography and channel network geometry. Building on previous work, we utilize signals in the modern topography of the Olympics to assess the impact of glacial overprinting. We use a broad suite of topographic metrics from the scale of drainage basins to the finer scale of channel reaches with slope-area analysis. Such methods are common in purely fluvial settings, but here we demonstrate that slope-area analysis accurately describes the development of relief along channel profiles regardless of the exact incision mechanism (fluvial or glacial) (see Section 3.3). Using multiple metrics allows us to constrain the relative influence of glacial and fluvial relief development. From this we constrain the portions of the mountain range that most likely record information about patterns of rock uplift rate.

2. BACKGROUND

The Olympics are part of the forearc high of the Cascadia subduction zone which extends from northern California to British Columbia (Figure 1). This forearc high marks the topographic and structural apex of an accretionary wedge formed by the obduction of sediments onto the North American plate from the subducting Juan de Fuca plate (Tabor and Cady, 1978a). Though only weakly metamorphosed (up to greenschist facies), the primarily turbidite lithologies of the Olympics increase in

grade from the shore of the Pacific Ocean to the core of the range (Figure 2). This metamorphic pattern has been attributed to deeper exhumation caused by longer periods of sub-aerial erosion in the core of the range (Brandon and Calderwood, 1990; Brandon et al., 1998).

Analysis of zircon and apatite fission track (Brandon and Vance, 1992; Brandon et al., 1998), and apatite (U-Th)/He (Batt et al., 2001) cooling histories suggests that exhumation of the accretionary wedge has occurred at a near-constant rate of ~1 km/Myr since the emergence of the Olympics at ~14 Ma. Similar erosion rates have also been proposed on shorter timescales (10^4 - 10^5 yr) from river incision rates (Pazzaglia and Brandon, 2001). These studies suggest an asymmetric spatial pattern of exhumation where the highest rates are located to the west of the range divide, and decrease outward towards zero at the coast (Figure 3).

Since the start of the Pleistocene alpine glaciers were likely active in every valley of the Olympics, but the size of the glaciers was highly variable (e.g. Porter, 1964; Montgomery, 2002). While glaciers extended to the Pacific Ocean on the western side of the range (Thackray, 2001), on the eastern side the lengths were probably limited by to the abutting Cordilleran Ice Sheet (Figure 2b) (Porter, 1964). Stratigraphic evidence shows that the ice sheet advanced into western Washington at least six times during the Pleistocene, but marine isotope records suggest that ice sheets were likely present in the region as far back as 2.5 Ma (Booth et al., 2003). Basal shear stress values of the ice sheet are estimated to have been low compared to modern alpine glaciers, with ice sheet sliding velocities around hundreds of meters per year (Booth, 1986; Brown et al., 1987). This suggests a system dominated by rapid mass transport under low driving stresses, which led to extensive low-gradient outwash plains in front of the ice sheet (Booth et al., 2003).

Pleistocene equilibrium line altitudes (ELA) of alpine glaciers varied significantly across the range (Figure 3a), where the ELAs were ~1.5 km lower on the west side of the range than the east side (Porter, 1964). This ELA trend was likely driven by the orographic precipitation pattern that is still prominent today (Figure 3a). Modern precipitation rates can be an order of magnitude higher on the west side due to an ENE-prevailing wind direction, and a strong rain shadow effect. However, the Pleistocene climate may have been impacted by the local climate created by the Cordilleran Ice Sheet and lower mean annual temperatures (Porter, 1964). As noted by Montgomery (2002), the alpine glaciers of the Olympics have left variable topographic overprinting throughout the range. This overprinting suggests that the glaciers were likely more effective agents of erosion than the rivers occupying the range before them. There is ample topographic evidence to show that alpine glaciers have caused valley widening and deepening over the succession of past glacial periods (Montgomery, 2002; Montgomery and Greenberg, 2000). Montgomery and Greenberg (2000) even suggested that mass removal by glaciers changed the rock uplift rates by initiating isostatic feedbacks, which led to the uplift of Mount Olympus, the highest peak in the range at 2429 m.

3. METHODS

Since rock uplift is ultimately the engine driving relief production and erosion in active orogens, the topography of mountain ranges should provide some constraint on the spatial distribution of rock uplift rates, and the commensurate response of surface processes (hillslope diffusion, fluvial and glacial incision). Below we utilize several topographic metrics that operate on variable spatial scales to explore the patterns of relief in the Olympic Mountains. We determine local relief,

geophysical relief, and hypsometry to analyze the gross modern relief of the range, which is a product of a mixture of landscape processes, such as glaciers, rivers, and hillslope diffusion. For a more detailed analysis at the scale of an individual process, we calculate hillslope angles and channel relief. With these tools, we aim to understand the influence of various tectonic and surface processes on shaping the Olympic Mountains. Our geomorphic analyses were performed on a 10 m-resolution National Elevation Dataset provided by the United States Geological Survey (www.ned.usgs.gov). While the signals of ice sheet erosion and topographic modification are important for this region of the world, the focus of this paper is specifically on alpine glaciation. For completeness, we have extended our topographic calculations into zones affected by the Cordilleran Ice Sheet, but in general these observations are omitted from our analysis unless otherwise noted.

3.1. Local Relief, Hillslope Angle, and Hillslope Aspect

A local relief map was created by calculating the maximum difference in elevation found within a 5 km-diameter moving window (Figure 2c). This diameter was selected because 5 km is roughly the distance between most major valleys. In this way, we analyzed relief over a larger area than just hillslopes, but small enough to detect spatial gradients. Maps of hillslope angle (Figure 2d) and hillslope aspect (Figure 2f) were created with standard ArcGIS algorithms by calculating the maximum dip and dip direction, respectively, across a 30 m-square (3 x 3 pixel) moving window. We extracted swath profiles of elevation and local relief data at both the range (Figure 3) and ridge (Figure 4 and Figure S1) scale.

As a means of constraining the magnitude of valley depth, or missing material below ridges, we calculated “geophysical relief” using the technique of Brocklehurst

and Whipple (2004). To do this, we carried out the following process on individual basins: (1) extract elevation values from the basin boundaries (divide), (2) fit a cubic spline surface between these elevations, (3) identified ridges that protrude from the surface and repeated the interpolation to include these points, and (4) subtracted modern topography from the interpolated surface (Figure 5a). Geophysical relief is different from local relief in that it measures relief within a specific basin rather than random regions defined by a circle. Thus, geophysical relief is useful for estimating the volume of a given basin.

3.2. Basin Hypsometry

Hypsometry has been used as a metric for analyzing the geomorphic form of large tracts of land. Previous research has suggested that the hypsometry of landforms is dependent on fluvial, glacial and tectonic processes (Kirkbride and Matthews, 1997; Brozovic et al., 1997; Montgomery et al., 2001; Brocklehurst and Whipple, 2004). Brocklehurst and Whipple (2004) showed that the degree of glaciation of a drainage basin is heavily dependent on the position of the ELA, as this position largely controls the areas and elevations of the landscape where glacial incision is focused. We generated hypsometric (cumulative frequency) curves for individual basins, which record the area above a given elevation plotted against that elevation. To allow for comparison across the range, we have normalized the elevation and cumulative area to the maximum and minimum values in each basin (Figure 5b). Because of this normalization, the area under this curve, the hypsometric integral (HI), lies between 0 and 1. We estimated this integral using:

$$HI = \frac{Z_{mean} - Z_{min}}{Z_{max} - Z_{min}} \quad (1)$$

where Z_{max} , Z_{min} , and Z_{mean} are the maximum, minimum and mean elevations within each basin, respectively. To isolate the relief structure of fluvial and alpine glaciated landscapes, we have omitted portions of basins that were likely affected by the Cordilleran Ice Sheet (see basins in Figure 2d).

3.3. Channel Relief and Topology

The change in elevation of a channel over a given horizontal distance (i.e. the reach-scale channel relief) is simply a function of the local channel slope. However, it is well known that natural channels have lower gradient sections downstream (though this reduction need not be monotonic) as drainage area increases downstream in both glacial (MacGregor et al., 2000; Anderson et al., 2006) and fluvial (Flint, 1974) landscapes. In order to account for this covariation in slope and area, we normalize local slope values by the upstream drainage area. This concept was initially proposed by Flint (1974) who observed that concave-up (henceforth concave and convex will be discussed with an upward reference) channel profiles can be reasonably described by a relationship between the local-slope (S) (m/m) and the upstream drainage area (A) (m^2) to a negative power, referred to as the concavity, θ

$$S = k_s A^{-\theta} \quad (2)$$

where k_s is the area-normalized slope, referred to as the channel steepness (dimensions depend on value of θ). In order to normalize channel steepness values across the Olympics we used a single reference concavity for all analyses. Therefore, all discussions of slope-area analysis are in reference to a normalized channel steepness (k_{sn}), though we omit the word “normalized” for brevity.

Previous authors have suggested that former glacial incision has left modern fluvial channels with concavities that are far from what is typically expected for persistent fluvial conditions. Brocklehurst and Whipple (2007) demonstrated that glaciers have a tendency to lower apparent concavities upstream of the former glacier terminus. In our analysis, we set θ to a commonly-used fluvial value of 0.45, and then analyzed how the channel steepness/relief (k_{sn}) changed along the length of the basin. In the following sections, we show that this value is a reasonable choice for the Olympics, as it produces relatively uniform channel steepness values in the Clearwater Basin, which has been suggested to be the least glacially impacted in the range (e.g. Pazzaglia and Brandon, 2001; Montgomery, 2001).

Slope-area analyses from previous work suggest that the channel steepness of *graded* rivers is positively correlated with rock uplift rates and inversely correlated with erosional efficiency. Erosional efficiency is related to climatic characteristics and channel bed resistance to erosion (e.g. Wobus et al., 2006). Because of this relationship, a map of channel steepness values can record spatial distribution of channel incision rates as influenced by spatial and temporal variations in rock uplift and erosional efficiency within a mountain range. In part, our aim here is to examine the extent to which these correlations are modified in a landscape where the channels are not graded. Abrupt changes in the channel profile, referred to as knickpoints, can be readily recognized by punctuated changes in channel steepness values downstream (Figure 2c and Figure 6). In the Olympics, knickpoints can have two general geometric forms; an increase in downstream channel steepness creates a convex knickpoint, while a downstream reduction forms a concave knickpoint (e.g. Royden and Perron, 2013).

While modern river channels, including those that experienced glacial incision have been studied using slope-area analysis (e.g. Montgomery, 2001; Brardinoni & Hassan 2006; 2007; Brocklehurst and Whipple, 2007; Robl et al., 2008; Hobley et al., 2010), many assumptions generally adopted in purely fluvial landscapes do not apply in mixed glacial-fluvial landscapes. For instance, in our study we do not require that the Olympic Mountains are in topographic steady-state (where elevations remain constant over time), nor do we imply that our slope area analysis relates directly to the processes of glacial incision, or that rock uplift rates need to be spatially uniform. We emphasize that this technique provides a robust, purely geometric construct for understanding the importance of spatial changes in channel relief without demanding an understanding of all parameters within a specific incision law (fluvial or glacial).

We quantified channel network topology and the vertical structure of channel steepness using χ analysis. The result of this calculation is a new distance term, χ (m), which is a scaled version of the distance from the outlet of the river of the channel, x (m). The integration of Eq. (2) shows that χ is the integral of the upstream accumulation area (Perron and Royden, 2013):

$$z(x) = z(x_b) + k_s \chi \quad (3a)$$

with

$$\chi = \int_{x_b}^x \left(\frac{A_0}{A(x)} \right)^\theta dx \quad (3b)$$

where x_b is the position of the outlet of the river (Olympic Peninsula shoreline in this study), and A_0 is a reference area used to preserve the length dimensionality of χ ($A_0 = 1 \text{ m}^2$ in this study). As indicated in Eq. (3a), the channel steepness determines the slope of elevation- χ relationships. These plots are referred to as linearized profiles or χ plots, and they provide a graphical representation of the change in channel steepness as a function of elevation (Figure 7). Therefore, in landscapes where the

relief is mostly dictated by the relief of channels, these plots can illuminate how relief scales with elevation. Where rock uplift rates, lithology, and climate parameters are spatially uniform, and have been temporally constant, χ plots should be nearly linear and values of χ should be consistent across drainage divides (Willett et al., 2014). In essence, this theory suggests that ranges in steady topographic state and uniform conditions, should have symmetric divides, and drainage networks sharing a divide should be similarly sized.

4. RESULTS

4.1. Local Relief, Hillslope Angle, and Hillslope Aspect

A swath across the entirety of the Olympic Mountain range in a direction parallel to tectonic convergence (54°) reveals that patterns of mean and maximum elevation have an asymmetric shape whereby the western flank of the range rises gently from the Pacific Ocean, and then decreases sharply on the eastern flank (Figure 3a). The local relief pattern across the range closely mimics the deviation between maximum and mean elevations (Figure 3), providing confidence that a 5 km-window for calculating local relief is accurately capturing the broad scale relief. However, the range is more radial in shape, as opposed to a simpler two-sided range. The most significant deviation from a simple two-sided range is created by a large valley flowing orthogonal to the profile – the Elwha valley (Figure 2d). This north-trending valley extends well into the topographic core of the range. Because of this, the highest elevations are actually found at the edges of the topographic core. While the Elwha valley topography creates a more complex orogen scale relief pattern, it does not obfuscate the broad regional pattern.

The spatial distribution of geophysical relief throughout the range is variable

(Table 1 and Figure 5b). However, a few notable trends can be observed. (1) The position of maximum geophysical relief occurs at lower elevations in large west side basins rather than small east side basins (c.f. Dungeness to Hoh in Figure 5a). (2) When comparing the total volume of each basin (V) we find that the basins on the east side of the range represent greater missing volumes per unit area, in other words the mean geophysical relief (GPR_{mean}) is higher on the east side ($GPR_{mean} = 0.45 \pm 0.9 \text{ } 1\sigma$, $N = 9$) than the west side ($GPR_{mean} = 0.30 \pm 0.9 \text{ } 1\sigma$, $N = 7$) (Table 1).

Even over small areas the distribution of hillslope angle values is highly variable (Figure 2d). As noted by Montgomery (2001), the basins of the western flank have a bimodal distribution of hillslope angles where there is a low-angle mode ($\sim 4^\circ$) and a high-angle mode near a threshold hillslope value ($\sim 30^\circ$) (see Figure S2). On the other side of the range most basins have a single mode where the mean is near a threshold hillslope value ($\sim 30^\circ$). While some east side basins could possibly contain a low-angle mode if areas affected by ice sheet were included, these modes would be much less significant than those exhibited by west side basins. The prominent low angle regions are highly correlated with wide valley floors, active and older Quaternary alluvium deposits, extant glaciers, glacially scoured lakes, glacial tills, and regions which were under the Cordilleran Ice Sheet (Figure 2b).

There are many examples of valleys and interfluvies that exhibit a high-degree of north–south asymmetry where the north flowing portions dominate the aerial extent of the basin or ridge (Figure 2f, Figure 4, and Figure S1). Figure 4 shows an example profile of one of these asymmetric ridges. Swath data of ridge elevations show that the north flowing tributaries supported significant glaciers as seen by valley parallel cirque basin topography. However, the south flowing portions of the ridge, representing the steep valley wall of the adjacent glacial valley, are nearly

linear and the minimum, maximum and mean elevation trends are all near threshold hillslopes ($\sim 30^\circ$) (Figure 4a). A swath of the northern portion of the ridge parallel to the strike of the ridge shows that glacial incision has created steep valley walls similar to those exhibited on the south side of the ridge (Figure 4b).

4.2. Basin Hypsometry

Normalized hypsometry curves show a continuum of basin elevation distributions throughout the Olympics (Figure 5b). Basins which flow from the range divide to the Pacific Ocean have larger areas at lower elevations, and thus a low hypsometric integral (HI mean = 0.28 ± 0.4 1σ , $N = 7$) (Table 1). Conversely, basins which lie to the east of the range divide have a larger area at higher elevations, yielding a higher hypsometric integral (HI mean = 0.45 ± 0.6 1σ , $N = 9$).

4.3. Channel Relief and Topology

Channel profiles on the east side of the Olympic range divide are on average steeper than on the west side (Figure 2c, Figure 6, and Figure 7). This pattern records the same range asymmetry exhibited in the swath data (Figure 3). Channel profiles from the west side exhibit a pattern where lower elevation portions of the landscape have lower channel steepness values and higher elevations are steeper. There are two common channel forms on the east side, those with nearly uniform and relatively high channel steepness values, and those that include large convexities where channel steepness values transition from higher to lower upstream (Figure 6 and Figure 7). This excludes portions of the landscape likely affected by continental ice. In general, these changes in channel steepness, which occur on both sides of the range, are discrete (i.e. they occur at knickpoints), which

suggests that they represent punctuated changes in erosional processes. These steepness transitions are not coincident with fault zones or lithology type (Figure 2 and Figure S3). High channel steepness values are often located in regions with high local relief, in the core of the range. High channel steepness channels are found in V-shaped valleys, while low channel steepness channels are often found in more U-shaped or flat floored valleys (Figure 6).

The abundance and length of river channels appears to be dependent on hillslope aspect in portions of the range. The ridge and valley asymmetry noted in the hillslope angle and aspect data, is also recorded in the river network topology (Figure 2e, Figure 2f and Figure S4). Few fluvial systems exist within portions of the range that are near threshold slope values, and south facing.

5. DISCUSSION

5.1. Glacial Relief Production and Reduction

In the following section, we look to the modern topography of the Olympics for signals of glacial impact, well aware that some amount of post-glacial landscape relaxation has likely already occurred. Indeed, this is a limitation for any post-glacial study. However, topographic relaxation is likely to only add noise to, or smooth out observations of glacial topography. As such, we move forward with our strongest and clearest results.

Local relief values are as high as 86% of the total range relief (2429 m), demonstrating that valley floor elevations approaching sea level are found in close proximity to the core of the range (Figure 6). These low elevation reaches and their low channel steepness values were likely created by glaciers lowering channels relative to previous fluvial conditions (e.g. Brocklehurst and Whipple, 2002; 2006;

377 Montgomery, 2002). As many authors before us (e.g. Sugden and John, 1976; van
378 der Beek and Bourbon, 2008; Valla et al., 2011), **we broadly use Penck's (1905)**
379 observation of overdeepening to be a reduction of valley floor elevation brought on
380 by enhanced glacial incision. Penck identified many accompanying features of these
381 overdeepenings – **“the trough, the trough's end, and the lake lying with in it, with its**
382 **shoulders and the hanging mouths of the side valleys.”** The Olympics contain many
383 excellent examples of these troughs, trough ends, hanging tributaries and lakes (e.g.
384 Figure 6). Though there are a few natural glacial valley lakes in the Olympics (Figure
385 2b), other depressions within valleys may have been filled in with glacial and fluvial
386 sediments.

387 Figure 8 shows a conceptual model for how glacial overdeepening affects
388 channel relief when it occurs at different elevations within a landscape. When high
389 elevation portions of the landscape are overdeepened, local channel slopes are
390 reduced. When overdeepening is focused at lower elevations within the landscape,
391 channel slopes are reduced at lower elevations, but channel slopes may increase at
392 higher elevations unless the drainage divide migrates dramatically. Many studies
393 have observed these changes in natural and modeled settings (e.g. MacGregor et
394 al., 2000; Anderson et al., 2006; Brocklehurst and Whipple, 2002; 2006; van der
395 Beek and Bourbon, 2008). The trough end represents a portion of the valley that is
396 markedly steeper than the trough itself. Therefore, glacial overdeepening, especially
397 at lower elevations, can be readily observed as the marked downstream reduction in
398 channel steepness coincident with concave knickpoints (see Figure 6, and Figure 7).

399 Increasing upstream channel steepness patterns in fluvial landscapes have
400 been attributed to an increase in rock uplift rate in other mountain ranges (e.g.
401 Wobus et al., 2003). Gradually increasing patterns of rock uplift like those shown in

Figure 3 do create upstream increases in channel steepness, however, this effect does not create discrete concave knickpoints (Whipple and Kirby, 2001; Cooper et al., 2016) like those seen in the Olympics. Neither concave or convex knickpoints are co-linear in map view as might be expected if they were fault controlled (Figure 2 and Figure S3). Furthermore, linearized profiles show that convex knickpoints are not found at the same elevation, as might be expected for a change in relative base level fall (e.g. Wobus et al., 2006). In fact, most knickpoint locations occur at river confluences (Figure 6 and Figures S5-S6), suggesting a downstream change in erosion capacity as a function of ice or water discharge (e.g. Whipple et al., 1999; MacGregor et al., 2000).

The relief of channels, fluvial and glacial, represent 80-90% of the total relief in most of the Olympics, and other mountain ranges around the world (e.g. Whipple and Tucker, 1999). The remaining 10-20% of the relief is formed by the hillslopes. This implies that the distributions of elevations throughout a range is mostly dependent on the geometry of the channels (e.g. Willgoose and Hancock, 1997). Because overdeepening tends to concentrate elevations near the ELA, this effect is recorded in the hypsometry of a basin (Brocklehurst and Whipple, 2004). While care needs to be taken when comparing normalized elevations from hypsometric curves and non-normalized ELAs, the basins represented in Figure 6 span a similar range of elevations, and therefore, the relative positions in the landscape are preserved. The hypsometric integrals of basins are quite low on the west side of the range, suggesting large areas of low elevation are present where ELAs were lower during the Pleistocene (Porter, 1964). Conversely, hypsometric integrals of the basins on the east side of the range are considerably higher, as was the Pleistocene ELA. These relationships are consistent with the work of Yanites and Ehlers (2012) who

demonstrated that the hypsometric integral was lowered considerably once a large fraction of the landscape was higher than the ELA within a landscape evolution model.

There is a remarkable similarity of the form of the hypsometric curve (Figure 5b) and linearized channel profiles (Figure 7) of the same basin. This is because the channel relief controls the topographic structure of a mountain range, be they fluvial or glacial channels, and therefore marked downstream reductions in channel steepness should be expected in any basin with low hypsometric integrals, as should convex knickpoints within basins with high hypsometric integrals. By definition, low hypsometric integrals record more area at lower elevations within a landscape. However, the only way to create such a condition is to create a significant area of low relief at low elevations. High hypsometric integrals require more area at higher elevations, which records larger areas of low relief at high elevations. These same relief trends are accurately recorded in linearized channel profiles, and in this form the spatial distribution of elevations is better preserved than in a hypsometric curve. This finding is important because slope-area and χ analysis are commonly used to investigate fluvial landscapes and their applicability to glacial settings is under appreciated.

Overdeepened portions of the valley are also coincident with broad, flat floors. Local valley cross section profiles are trapezoidal (Figure 6), where hillslopes dip toward the flat valley floor at consistently high angles ($\sim 30^\circ$). As the most significant valley floor flattening and widening is in the southwestern portions of the range where ELAs were the lowest, we attribute this topographic signature to heavy Pleistocene glacial incision. Montgomery (2002) came to a similar conclusion based on trends in valley width throughout the Olympics.

Channel topology and drainage divide asymmetries provide ample evidence of lateral incision throughout the range. The patterns of channel topology in the Olympics are consistent with a conceptual model where glacial incision may have effectively erased nearly all south flowing tributary systems within the Queets, Hoh, east fork of the Quinault, Dosewallips, Duckabush, and South Fork Skokomish basins. Topological disequilibrium is evident throughout the range as seen by drastic change in χ across drainage divides (Figure 2e and Figure S4) (Willett et al., 2014). These χ disequilibria imply that at some time during the evolution of the Olympics, north draining basins have grown/lengthened at the expense of the adjacent southern basins. Swath profiles across asymmetric ridges throughout the range (Figure 4 and Figure S1) also provide evidence of this divide migration, similar to what has been observed in other glaciated mountain ranges around the world (Oskin and Burbank, 2005; Naylor and Gabet, 2007). Efficient headward erosion of north flowing valleys, and valley floor widening of adjacent large trunk glaciers created a southward migration of drainage divides within east-west trending ridges.

Over time, these divides would have been pushed to a topographic state where continued southward divide migration would have led to divide lowering as the threshold landscapes were no longer able to steepen, because of limitations related to the cohesive strength of the local rocks, and the base of the southern ridge side would not be able to migrate because of the large abutting trunk glaciers. Such a topographic condition would be extremely difficult to create in a fluvial landscape because opposing fluvial systems with different erosion rates would simply adjust the slopes of their channels to stabilize the shared divide (e.g. Willett, 1999). The north flowing tributary basins are also dominated by threshold hillslopes where the topographic relief is limited by the hillslope length (Figure 4b), which is in turn

controlled by the spacing of tributary valleys (e.g. Burbank et al., 1996). Though we cannot quantify the magnitude of channel or interfluvial lowering, there is clear evidence recorded in the topography and channel network geometry that the production of relief due to glacial activity was limited on the western flank of the range.

There is also an interesting asymmetry of the Elwha drainage divide, which may indicate recent peak lowering. The portions of the Elwha drainage divide that also represent the range divide are lower than the remaining portion of the basin divide (ignoring the lower portions that have been effected by the Cordilleran Ice Sheet) (Figure 9). This portion of the range crest has many prominent wind gaps suggesting a change in drainage area of the basins that share the range divide. Taken together, the low divide elevations and wind gaps suggest that the range crest has migrated, and that this migration occurred at the expense of peak elevations. It is possible that the headward erosion of the large west flowing glacial systems may have impinged on basins to the east and thus shifted the range divide.

5.2. Signals of Pre-Glacial Fluvial Relief

Every channel profile from the Olympic Mountains contains a low channel steepness reach at the highest elevations (e.g. Figure 6). This portion of the profile contains small cirque basins, colluvial reaches, hillslopes, and in some cases, extant glaciers. The lowest elevations in each profile also have similarly low channel steepness values due to alpine glacier overdeepening on the west side and continental ice erosion on the east side. In between the upper and lower low-channel steepness sections on the west side, channels are steeper and contain many minor convex and concave knickpoints. The very steep reaches on the west side may be

502 the product of downstream overdeepening because without commensurate channel
503 lengthening, downstream overdeepening would lead to upstream steepening. Only
504 some of the west side tributaries whose headwaters are lower than ~500 m exhibit a
505 relatively constant channel steepness value along their length (Figure 2c and Figure
506 7). Because of these observations we cannot be certain that large west side basins
507 still record an unmodified form of fluvial relief from a time before the onset of
508 glaciation. We include the Clearwater in this interpretation, which exhibits similar
509 patterns of channel steepness (Figure S5 and Figure S6) and hypsometry (Figure
510 5b) as the other west side glaciated basins. This does not directly call into question
511 previous research constraining the recent fluvial record from the deposits of the
512 Clearwater (Pazzaglia and Brandon, 2001; Wegmann and Pazzaglia, 2002),
513 because we cannot constrain the age of this topographic signal. Therefore, we
514 cannot rule out that the bedrock channel profile geometry could be much older than
515 the deposits currently found there.

516   The channel segments which remained above the ice sheet on the east side
517 exhibit a higher channel steepness. However, larger tributaries often contain a major
518 convex knickpoint where channel steepness decreases upstream. These major
519 convex knickpoints likely highlight a boundary in the landscape between regions that
520 where glaciers had little influence (downstream), and regions formerly dominated by
521 glaciers (upstream) - similar to what has been suggested in other mountain ranges
522 (Brocklehurst and Whipple, 2006; Hobley et al., 2010). Smaller tributaries that grade
523 into trunk streams downstream of these knickpoints exhibit a nearly uniform channel
524 steepness value (Figure 7). This suggests that they experienced relatively minor
525 glacial modification, as we would expect glacial erosion to create knickpoints and
526 that these knickpoints would still be observable. Due to the rain shadow across the

527 Olympics, and its effect on the Pleistocene ELA, the basins on the east side of the
528 range were spared from the heavy glacial erosion experienced by the west side
529 (Figure 3). This has preserved a stronger signal of the pre-glacial fluvial relief,
530 though we do not suggest that these channels perfectly reflect the initial fluvial
531 geometry.

532

533 5.3. Pattern of rock uplift recorded in topography

534 We have shown that alpine glaciers and continental ice created a pervasive
535 signal of channels with low steepness values, and an abundance of convex and
536 concave knickpoints. Therefore, most channel segments are the product of
537 Pleistocene glaciers and not necessarily rock uplift rates. Low hillslope angle zones
538 are found at lower elevations created by glacial valleys infilled with Quaternary
539 deposits, or by continental ice sheets (c.f. Figure 2b and Figure 2d). Near threshold
540 hillslopes, likely oversteepened from glacial erosion, dominate the rugged core of the
541 range. As such, we also find hillslope angle to be a weak metric for predicting spatial
542 patterns in rock uplift.

543 Due to the intensity of Pleistocene topographic modification, it is difficult to
544 constrain how much spatial patterns in climate influenced the mean elevation and
545 relief of the range before glaciation. Indeed, some amount of range asymmetry could
546 also be expected due to the intense rain shadow across the range (e.g. Willett,
547 1999). However, the topography of the eastern flanks of the Olympics is not a simple
548 function of fluvial systems which have steepened to compete with more aggressive
549 adjacent systems receiving considerably more precipitation. Many of these shorter,
550 steeper fluvial systems on the eastern flank are buffered from the larger systems on

551 the western flank by the large Elwha basin. Instead, we suggest these steep
552 channels are better records of an asymmetric rock uplift pattern.

553 Perhaps the most robust topographic indicator of rock uplift rates is the spatial
554 variation in local relief across the orogen, which records the magnitude of relief on a
555 longer wave length than the channel or hillslope scale, and thus may not be as easily
556 over printed by relatively short-lived climate signals. The mean and maximum local
557 relief patterns are nearly identical to the mean and maximum elevation patterns
558 across the range (Figure 2 and Figure 3). Brocklehurst and Whipple (2007) found the
559 same pattern in the heavily glaciated Southern Alps, New Zealand where they
560 concluded that these topographic signatures also scaled with rock uplift rates. In
561 addition, the mean geophysical relief is higher in the east side basins suggesting that
562 these basins may have formed under the influences of higher erosion and rock uplift
563 rates (Table 1).

564 A rock uplift pattern corresponding to the local relief pattern in the Olympics
565 would be in agreement with the deep exhumation pattern recorded in the reset zircon
566 fission track ages (Figure 3b) and the metamorphic grade of the core rocks (Figure
567 2a). The correlation of the relief and long-term exhumation pattern implies that the
568 pattern of relief is long lived. However, the asymmetric local relief pattern does not
569 match the asymmetric pattern of the erosion rates estimated from apatite fission
570 track data or the pattern of more recent river incision rates (Figure 3). It could be that
571 this cross-range erosion pattern suffers from the complications of partially reset
572 apatite fission track data, and the interpretation of a fluvial system whose recent
573 incision rates may not match long-term rock uplift rates (i.e. the Clearwater). The
574 work of Stolar et al. (2007) shows that if the broad topographic form of the range is
575 considered to be in consistent over long time scales, a predicted erosion rate pattern

would also scale with mean elevation and local relief based on the results from a coupled critical wedge-landscape evolution model (Figure 3b).

A rock uplift pattern that mimics the mean elevation and local relief could be created by an east-plunging antiform which deforms the rocks of the accretionary wedge and North American plate. Such a deformation pattern is likely linked to the east-plunging arch in the subducting slab of the Juan de Fuca plate observed from the analysis of earthquake hypocenters (Figure 1) (Crosson and Owens, 1988; Weaver and Baker, 1989; Brandon and Calderwood, 1990). While subducting ridged indenters have been shown to create rock uplift patterns similar to our observations in the Olympics (e.g. Bendick and Ehlers, 2014), there is also strong evidence to suggest that sedimentary underplating is important. Geophysical imaging (e.g. Clowes et al., 1987; Calvert et al., 2011) and geologic mapping (Tabor and Cady, 1978a, b) provide ample evidence for the accretion and imbrication of 10s of kilometers of sediments in the hanging wall of the Hurricane Ridge Fault. Three-dimensional seismic tomography shows that these sediments (Calvert et al., 2011) also have an east-plunging antiformal shape, but with sharper bend in the planform arc than the arc in the Juan de Fuca plate (Figure 1). The structural culminations associated with the underplating of these sediments could create a rock uplift pattern similar to that exhibited by the topographic relief. The early work of Tabor and Cady (1978a, b) also provides structural evidence of east-plunging antiformal deformation of the Coast Range Terrain rocks, and the erosional half-window exhibited by the Hurricane Ridge fault.

6. CONCLUSIONS

We suggest that the most robust record of recent rock uplift in the Olympic

601 Mountains is the modern medium-long wavelength topography. Here, as in many
602 other ranges around the world, rock uplift follows the trends in local relief (5 km-
603 relief), or more accurately, rock uplift drives the development of local relief. The
604 broad trend in local relief throughout the range is correlated with patterns of deep
605 exhumation on longer timescales as recorded in the metamorphic grade and the
606 oldest available cooling histories of bedrock (i.e. zircon fission track data). The rock
607 uplift pattern recorded in the topography is suggestive of a broad east-plunging
608 antiform, which could have been created by folding of the subducting plate (e.g.
609 Crosson and Owens, 1998; Brandon and Calderwood, 1990; Bendick and Ehlers,
610 2014), or duplex growth.

611   Today the finer-scale topography of the Olympics is controlled by hillslope and
612 fluvial processes. However, these processes are not in equilibrium with rock uplift
613 rates because the topography of the range is still dominated by the geometry set by
614 the now extinct glacial systems. In this paper, we have demonstrated that glaciers
615 have dramatically reorganized the topography and channel network topology of the
616 Olympics. In doing so, we have shown the utility of analytical methods that may be
617 **considered “traditionally fluvial”** (e.g. channel steepness and χ) in understanding the
618 evolution of channel topology and relief in mountain ranges which have been
619 influenced by other processes.

620   These changes have most certainly created a condition where post-glacial
621 erosion rates are influenced by hillslope and fluvial processes attempting to reach a
622 new equilibrium topographic state, and therefore, are not equivalent to rock uplift
623 rates. This condition is not unique to the Olympics. Indeed, many Cenozoic mountain
624 ranges around the world which experienced significant glaciation are likely in a
625 similar state. Ranges where the Pleistocene ELA reached low elevations will likely

be the most effected including: the European Alps (e.g. van der Beek and Bourbon, 2008), the North American Rocky Mountains, the Southern Alps of New Zealand (e.g. Brocklehurst and Whipple, 2004), the southern Andes (e.g. Thomson et al., 2010), and Alaska (e.g. Meigs and Sauber, 2000).

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841

842 FIGURE CAPTIONS

843 Figure 1. Topography and geologic features of the Cascadia convergent margin
 844 offshore of the of the Olympic Peninsula, Washington State, USA. Red lines are
 845 depth contours of the subducting slab (red numbers denote depth) (from Crosson
 846 and Owens, 1987). Black dashed lines are depth contours of the accretionary wedge
 847 sediments (black numbers denote depth) which is the plane of the Hurricane Ridge
 848 Fault (from Calvert et al., 2011). Note that the trace of the Hurricane Ridge Fault
 849 marks the top of these sediments (i.e. 0 km depth). The relative velocity of the Juan

850 de Fuca plate toward the North American plate is ~36 mm/yr with a bearing of ~54
851 (DeMets and Dixon, 1999).
852
853 Figure 2. Geology and geomorphology of the Olympic Mountains. a)
854 Simplified geologic map. Tectonostratigraphic units from Brandon et al. (1998).
855 Coast Range (CR) terrain – pillowed and massive basalts, diabase dikes, and rare
856 pelagic limestone and reddish mudstone. Upper Olympic Subduction Complex
857 (OSC) – mainly turbidite sandstone, and subordinate mudstone. Lower OSC – clastic
858 sedimentary rocks, mainly turbidites. Coastal OSC – turbidites, mudstones, and
859 minor pillow basalt. Faults and metamorphic grade from Tabor and Cady (1978b). L
860 – laumontite; Pr + Pu – prehnite plus pumpellyite; Pu – pumpellyite; Cl + Ep – chlorite
861 plus epidote. b) Simplified surface cover map (adapted from Washington Division of
862 Geology, 2010). Ice including extant alpine glaciers is masked in light blue. Natural
863 lakes formed by glacier overdeepening are shown in dark blue. Undifferentiated
864 Quaternary alpine glacial till deposits are marked in brown. Undifferentiated
865 Quaternary alluvial deposits are marked in yellow. The extent of the Cordilleran Ice
866 Sheet is delineated with a black dashed line. The summit of Mount Olympus is
867 marked with a red dot. c) Local relief (5 km-relief) overlain by a channel steepness
868 (k_{sn}) map of drainage areas > 2 km² with a 1 km smoothing window. d) Hillslope
869 angle map overlain by outlines of river basins excluding areas influenced by the ice
870 sheet: (1) Quinault, (2) Queets, (3) Clearwater, (4) Hoh, (5) Bogachiel, (6) Calawha,
871 (7) Sol Duc, (8) Elwha, (9) Morse, (10) Dungeness, (11) Quilcene, (12) Dosewallips,
872 (13) Duckabush, (14) Hamma Hamma, (15) North Fork Skokomish, (16) South Fork
873 Skokomish. e) Elevation data overlain by χ values of drainage areas > 2 km². The
874 range divide is shown in magenta. f) Shaded relief overlain by hillslope dip direction

875 (aspect) map. Yellow outlines mark the extent of swath data in Figure 4 and Figure
876 S1.

877

878 Figure 3. a) Elevation and climate data across the Olympic Mountain range parallel
879 to the direction of tectonic convergence ($\sim 54^\circ$). Maximum and mean elevations are
880 shown in thin and thick brown lines, respectively. Mean annual precipitation data
881 (PRISM data; Daly et al., 1994) are shown in blue. Equilibrium line altitude data from
882 Porter (1964) are represented by a red trend line. b) Comparison between patterns
883 of estimated erosion and relief across the Olympic Mountain range. Maximum and
884 mean local relief values are shown in thin and thick brown lines, respectively. The
885 blue and red polygons mark the erosion patterns estimated from Pazzaglia and
886 Brandon (2001) based on low-temperature thermochronometry and river incision,
887 respectively. The black envelope is the erosion pattern from Stolar et al. (2007)
888 based on a coupled critical wedge-landscape evolution model. The grey bar denotes
889 the zone of reset zircon fission track samples from Brandon and Vance (1992).

890

891 Figure 4. Elevation profiles from a ridge within the Queets basin. See Figure 2f for
892 location. Grey envelopes and black line show maximum, minimum, and mean
893 elevations. a) Values calculated perpendicular to the ridge. b) Values calculated
894 parallel to the ridge within the landscape bound by the dashed lines in (a). Note that
895 the slopes in these ridges appear to be near threshold values ($\sim 30^\circ$).

896

897 Figure 5. a) Geophysical relief as a function of elevation of example basins from the
898 Olympic Mountains. b) Normalized basin hypsometry curves throughout the range.
899 Westside basins (blue) have a much lower hypsometric integral (number in brackets)

900 than eastside basins (red). The Clearwater curve is dashed blue for easy
901 identification. The black curve represents a landscape with a Gaussian distribution of
902 elevations (hypsometric integral = 0.5) for reference. The marked difference between
903 high and low hypsometric integral basins is caused by the position of the ELA within
904 the basin.

905

906 Figure 6. Topography of example basins from Figure 5a. See Figure 2d for
907 locations. Channel profiles are colored by local channel steepness values (k_{sn})
908 calculated with a 0.5 km smoothing window. The grey envelope represents the
909 elevations between the channel and a interpolated cubic spline surface above the
910 channel. Major convex and concave knickpoints are shown as white and black dots,
911 respectively. Solid black lines show the change in drainage area as a function of
912 distance. Inset plots show example cross sections from a position along each profile.
913 a) The Queets basin. b) The Hoh basin. c) The Elwha basin. d) The Dungeness
914 basin. Note: we have not removed the manmade Elwha dams from this dataset
915 (these are the two lowest convex knickpoints).

916

917 Figure 7. Linearized channel profiles from the basins in Figure 5a. a) Hoh River (4).
918 b) Elwha River (8). c) Queets River (2). d) Dungeness River (10). χ values are shown
919 for river reaches with accumulation areas $>1 \text{ km}^2$ to reduce the effects of hillslopes
920 and extant glaciers. The slopes of these plots are the local channel steepness
921 values. Note the highly linear (grey) profiles that were not as heavily modified by
922 glacial erosion (i.e. lack major knickpoints).

923

924 Figure 8. Conceptual model showing changes in river profiles under the influence of
925 glacial overdeepening. The black line is the initial river profile with a single channel
926 steepness. The red line represents a profile where more glacial erosion has occurred
927 higher in the landscape due to a higher glacier ELA. The blue line represents a
928 profile where more glacial erosion has occurred at lower elevations due to a lower
929 glacier ELA. Note: in both cases the ELA has been placed at the midpoint of the
930 profile. a) Traditional channel profiles (long profiles). b) Linearized channel profiles.
931

932 Figure 9. Topography of the Elwha drainage divide. Note the lower mean elevation
933 and sizable wind gaps associated with the portion that represents the range divide.

Table 1. Characteristics of Olympic Mountain drainage basins.

Basin Name	Number*	Range Side	Area (km ²)	Volume (km ³)	GPR _{mean} (km)	Hypsometric Integral
Quinault	1	West	1103	435	0.39	0.24
Queets	2	West	738	261	0.35	0.23
Clearwater	3	West	402	75	0.19	0.28
Hoh	4	West	764	314	0.41	0.27
Bogachiel	5	West	303	79	0.26	0.27
Calawah	6	West	311	70	0.23	0.35
Sol Duc	7	West	504	139	0.28	0.29
Mean					0.30	0.28
Elwha	8	East	691	399	0.58	0.48
Morse	9	East	95	39	0.41	0.49
Dungeness	10	East	383	150	0.39	0.52
Quilcene	11	East	131	49	0.37	0.41
Dosewallips	12	East	249	147	0.59	0.51
Duckabush	13	East	175	82	0.47	0.49
Hamma Hamma	14	East	197	77	0.39	0.41
N. Fork Skokmish	15	East	178	91	0.51	0.41
S. Fork Skokmish	16	East	148	51	0.34	0.36
Mean					0.45	0.45

*See Figure 2d for basin locations. The standard deviation on the west side mean GPR and HI values are 0.09 km and 0.04, respectively. The standard deviation on the east side mean GPR and HI values are 0.09 km and 0.06, respectively.