ABSTRACT

SMITH, STEPHEN GERALD. Tectonic and Climatic Controls on Landscape Evolution in the Hangay Mountains, Mongolia and Olympic Mountains, USA. (Under the direction of Karl W. Wegmann).

Mountain belt evolution progresses via the interplay between forces that work to build topography and those that work to wear it down. Of these forces, tectonics and climate are among those central to developing a better understanding of both short and long-term erosion, especially with respect to how spatial and temporal patterns of erosion reflect variability in these parameters. A more robust knowledge of the feedbacks between tectonics, climate, and erosion across different regions of the globe is beneficial not only for modeling the geologic evolution of mountainous terrain, but also for better predicting the response of landscapes to future change.

Chapter 1 of this dissertation focuses on the Hangay Dome of central Mongolia, a midcontinental mountain range characterized by long wavelength high topography, basaltic eruptions spanning 30 million years, and an abundance of flat-topped summit plateaus. Although previous research has led to the development of several competing hypotheses regarding the formation of the range, the origin and timing of the Hangay Dome uplift remains unresolved. Using Landsat imagery, GIS, and cosmogenic \(^{10}\)Be, geomorphological investigations of paleo-topography, erosion rates, and the relative contribution of glacial activity to total erosion provide insight into the nature of Hangay Dome landscape evolution since the Mid-Miocene. Reconstruction of paleo-valleys with \(>600\) m of local relief, sluggish mean erosion rates (<45 m/Myr), and dominance of glacial erosion suggest that there has been no dramatic change in tectonic forcing of the Hangay Dome since \(~13\) Ma, and that high amplitude climate oscillations beginning in Pliocene time have led to an environment influenced primarily by the activity of glaciers.

Chapters 2 and 3 similarly explore the connections between tectonics and climate, but move from Mongolia to the Olympic Mountains of Washington and deal with much shorter timescales (late Holocene to present). More specifically, chapter 2 highlights the sedimentary record of Lake Quinault, a \(~70\) m deep moraine-dammed lake located on the western front of the Olympic Mountains. Analysis of several cores from the lake reveals a record of deposition over the past \(~4000\) years that is clastic-dominated and driven by large flood events. Using \(\mu\)XRF data as a proxy for individual flood layers, we assessed the
periodicity of overall hydrologic variability and the distribution of extreme events (paleofloods) throughout the late Holocene. Despite the lack of a long-term trend, notable peaks in event number occurred during ~2350 to ~2450 cal BP and ~1910 to ~2010 cal AD. The period ~2350-2450 cal BP is unremarkable within the context of existing Pacific Northwest paleoclimate records, but the spike in extreme events over the last century suggests that an increase in the frequency of large storms tracking over the Olympic Peninsula is connected to recent climate change.

Finally, Chapter 3 takes a broader look at recent landscape evolution in the Olympic Mountains by exploring the hypothesis that precipitation is linked to the volume of landslide erosion over a ~15 km-wide, 1250 km² swath of the range. Statistical analysis reveals a significant correlation between landslide volume and mean annual precipitation, and assessment of landsliding during the period 1990-2015 shows that 98% of recent landslide volume was produced in the windward, high-precipitation side of the range. A calculated erosion rate from landsliding of 0.28 ± 0.11 mm yr⁻¹ is also similar to other estimates of erosion throughout the region, which suggests that climate is a significant component of landscape evolution over both short and long timescales in the Olympic Mountains.
Tectonic and Climatic Controls on Landscape Evolution in the Hangay Mountains, Mongolia and Olympic Mountains, USA

by
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A dissertation submitted to the Graduate Faculty of North Carolina State University in partial fulfillment of the requirements for the degree of Doctor of Philosophy

Marine, Earth, and Atmospheric Sciences

Raleigh, North Carolina
2016

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DEDICATION

To Mom, Val, and Ilona. We are stronger together than we could ever be alone.

“Nothing worth doing is completed in our lifetime; therefore, we are saved by hope. Nothing true or beautiful or good makes complete sense in any immediate context of history; therefore, we are saved by faith. Nothing we do, however virtuous, can be accomplished alone; therefore, we are saved by love.” – Reinhold Niebuhr
BIOGRAPHY

Stephen G. Smith was born in Philadelphia, PA and raised in Bear, DE. He earned a Bachelor of Science degree in geology from Bucknell University in 2008 and a Master of Science degree in oceanography from the University of Rhode Island in 2010. He spent 18 months traveling and working in the environmental consulting industry before enrolling in the PhD program at NC State in 2012.
ACKNOWLEDGEMENTS

I have had a wonderful PhD experience at NC State, and owe gratitude to many organizations and individuals both associated with and separate from the University. My committee members, Karl Wegmann, Lonnie Leithold, Del Bohnenstiehl, and John Gosse deserve special thanks for all their valuable input and support. This dissertation has been made exponentially better as a result of their guidance and suggestions.

Regarding work in Mongolia, I would like to thank Naraa Mandah for his excellent driving and navigation throughout the rugged interior of the country, and Dashka Khorlooand, Gantulga Bayasgalan, Nathan Lyons, and Matthew Morriss for their camaraderie and support in the field. Mongolia research was supported by National Science Foundation Research Grants EAR-1009702 and EAR-1009680. For Washington, I owe special thanks to the Quinault Indian Nation for providing access to the natural gem that is Lake Quinault. Field logistics also greatly benefited from the advice and assistance of Bruce Wagner, Larry Gilbertson, and Bill Armstrong of the Quinault Indian Nation Department of Fisheries. I’d also like to thank Jubril Davies and Bruce Riddell for their help with obtaining sediment cores, as well as Robert Lane, Deanna Metevier, Corey Moore, Catherine Opalka, and Sharese Roberts for their assistance with computer and laboratory analyses. Washington research was supported by a Geological Society of America Graduate Student Research Grant and National Science Foundation Grant EAR-1226064.

I would also like to thank all of the faculty and staff at NC State who have contributed to my personal and professional development as both a scientist and educator. Whether formal or informal, I have cherished our interactions and consider myself extremely lucky to be a member of the Wolfpack. To the fellow graduate students (and several undergraduates) within the Department of Marine, Earth, and Atmospheric Sciences – thank you for your friendship, your camaraderie, and for lending a helping hand whenever necessary. Of course, this dissertation would not have been possible without my family and closest friends. This is especially true with respect to the three most important and beautiful women in my life (you know who you are!) who have made me the man I am today. I am forever grateful for your love and support.
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CHAPTER 1

Paleotopography and erosion rates in the central Hangay Dome, Mongolia: Landscape evolution since the Mid-Miocene

1.1 Abstract

Standing over 2 km above the surrounding topography and flanked by orogen-scale strike-slip faults, the Hangay Dome in central Mongolia is characterized by long wavelength high topography, basaltic eruptions spanning 30 million years, and an abundance of flat-topped summit plateaus. However, despite decades of research, the origin and timing of the intraplate Hangay Dome uplift continues to be debated. Using Landsat imagery, GIS, and cosmogenic beryllium-10, we employ geomorphic investigations of 1) paleotopography preserved beneath basalt flows of known age, 2) erosion rates at various temporal scales, and 3) the relative contribution of glacial activity to total erosion to provide insight into the nature of landscape evolution in the Egiin Davaa region of the central Hangay Dome since the middle Miocene. Reconstruction of paleo-valleys cut into Paleozoic basement rock that exhibit a degree of local relief (>600 m) similar to the modern landscape, sluggish mean erosion rates (<45 m Myr\(^{-1}\)), and dominance of glacial erosion suggest that there has been no dramatic change in tectonic forcing of the study area since ~13 Ma, and that high amplitude climate oscillations beginning in the Pliocene have led to an environment influenced primarily by the activity of glaciers. These results provide support for uplift onset during the Oligocene or early Miocene, quantify landscape evolution since the middle Miocene, and underscore the importance of considering geomorphic archives found on Earth’s surface when building models of intra-continental epeirogeny.
1.2 Introduction

The Hangay mountain range, or Hangay Dome, defines a 2 x 10^6 km² region of high, long-wavelength topography in central Mongolia with a maximum elevation of 4008 m at the summit of Otgon Tenger Uul (Figs. 1 & 2). Despite the range’s physiographic prominence in central Asia, both the origin and evolution of the Hangay Dome remain unclear, as prior studies have failed to reach a unifying theory for the Cenozoic geologic history of the region (e.g., Molnar and Tapponnier, 1975; Windley and Allen, 1993; Cunningham, 2001; Petit et al., 2002, 2008; Yanovskaya and Kozhevnikov, 2003; Bayasgalan et al., 2005; Tiberi et al., 2008; Savatenkov et al., 2010). Perhaps this lack of agreement is unsurprising – the origins of intraplate mountain belts are often enigmatic since their formation is left unexplained by conventional models of plate tectonics where vertical surface movements are driven largely by the horizontal motion of plates interacting at their boundaries. Instead, long wavelength cratonic uplift, or epeirogeny, is usually indicative of dynamic interactions between the asthenosphere and lithosphere combined with isostatic responses, and researchers have stressed the importance of considering this manner of surface deformation when interpreting the geologic record (Conrad and Husson, 2009; Braun, 2010). As a result, recent research has aimed at gaining a better understanding of this complex aspect of the Earth system and the mechanisms controlling the generation of dynamic topography (Lithgow-Bertelloni and Silver, 1998; Moucha et al., 2008; Braun et al., 2013; Flament et al, 2013; Guillaume et al., 2013).

Since Earth surface processes are often linked to the tectonic regime of a particular area, geomorphic observations of the Hangay Mountain range have the potential to yield valuable information on the nature of its evolution. Prior studies have suggested that the Hangay Mountains may be a product of relatively recent, rapid uplift since 5 Ma (Yarmolyuk et al., 2008), requiring that geomorphic data regarding erosion rates, topographic relief, and drainage basin geometry reflect this tectonic activity (Kirby and Whipple, 2001; Montgomery and Brandon, 2002; Wobus et al., 2006; Cyr et al., 2010; Whittaker, 2012; Gallen et al., 2013). Conversely, others have suggested that the majority of uplift of the Hangay mountains began and ended during the Oligocene epoch, in which case geomorphic data are likely to indicate that the mountains are evolving relatively slowly owing to a lack of recent tectonic forcing (Yanshin, 1975; Höck et al., 1999; Cunningham, 2001; Walker et al., 2007; West et al., 2013; Caves et al., 2014). With the goal of narrowing the ~25 Myr
uncertainty in the timing of uplift, this paper details both field and remotely-sensed geomorphic data that, coupled with $^{40}\text{Ar}/^{39}\text{Ar}$ ages of Cenozoic basalts, cosmogenic $^{10}\text{Be}$ erosion rates, and $^{10}\text{Be}$ surface exposure ages of moraine boulders, provides constraints on the topographic and erosional evolution of the central Hangay Dome.

A primary objective of this research, which was inspired by observations of significant topographic relief preserved in former paleo-valleys filled by Miocene basalt flows, is to elucidate the geomorphic signature of central Hangay Dome landscape evolution since the eruption of these lavas. The first component of this study focuses on expansive ridge-top basalts in the vicinity of the centrally located Egiin Davaa (Egiin Pass) (Fig. 2). Preserved on either side of the Hangay watershed divide that separates north-flowing tributaries of the Selenga River from internally drained basins of the Gobi Altai to the south, these Miocene basalts completely filled and concealed the underlying paleo-valleys. The preservation of this paleo-landscape provides a rare and valuable opportunity to document the relief in the Hangay Dome at a time prior to emplacement of the lowermost and presumably oldest lava flow. Age constraints on the succession of basalt flows not only permit a comparison of Paleogene and modern topographic relief across the central Hangay, but also allow for the derivation of erosion rates since the uppermost lava flow (e.g., Gani et al., 2007). Ridge-top bedrock terrestrial cosmogenic nuclide (TCN) erosion rate estimates are used to further establish the rate of relief generation since the end of volcanism in the study area.

Although some researchers posit cessation of Hangay uplift by the end of the Oligocene (e.g., Yanshin, 1975), the modern range exhibits significant topographic relief in the form of deeply incised U-shaped valleys. Therefore, this study also evaluates the role of glaciers in shaping the modern topography of the Egiin Davaa region and serving as the primary driver of late Cenozoic landscape evolution. There is compelling evidence for an increase in erosion rates in mountainous regions since the onset of Northern Hemisphere Glaciation ~3 – 2 Ma (e.g., Molnar and England, 1990; Peizhen et al., 2001; Molnar, 2004; Charreau et al., 2011; Herman et al., 2013; Herman and Champagnac, 2016), although the cause and global impact of this increase has been debated for decades (Syvitski and Milliman 2007; Willenbring and von Blanckenberg 2010; Hidy et al., 2014). As such, we explore the role of climate in sculpting the modern topography of the central Hangay Dome by coupling Last Glacial Maximum (LGM) end moraine volumes with TCN exposure ages in order to estimate the minimum glacial erosion of bedrock. This approach aims to test the hypothesis that
glacial erosion rates are significantly higher than those averaged over multiple glacial/interglacial cycles, which, if proven, would provide support for the notion that the modern central Hangay landscape reflects Quaternary climate-driven relief generation as opposed to a recent, rapid change in tectonic uplift.

Collectively, the analyses presented here were conducted to test three hypotheses: 1) that late Miocene paleo-relief is comparable to modern relief; 2) that erosion rates over multimillion-year timescales are low and analogous to modern erosion rates; and 3) that the modern landscape of the high Hangay has been shaped primarily by the activity of glaciers. The results of this study, combined with insights from a broad, multi-institutional collaboration (e.g., Meltzer et al., 2012) to study Hangay Dome rock and surface uplift histories and lithosphere-mantle dynamics, have implications for understanding the scale and pace of asthenosphere-lithosphere interactions that create long wavelength intraplate topography, for global climate dynamics, and for landscape response to both climatic and tectonic forcing.

1.3 Background and Geologic Setting

The Hangay Dome is situated at the kinematic transition between the Baikal rift province to the north and the Gobi-Altai transpressional range to the south (Figs. 1 & 2; Baljinnyam et al., 1993; Cunningham, 2001). Precambrian granites, gneisses, migmatites, and schists form the basement of the Hangay region (Kepezhinskas, 1986; Cunningham, 2001), and although only locally exposed, this basement complex appears to underlie the entirety of the Hangay Dome (Kovalenko et al., 1996; Cunningham, 2001). The cratonic block of central Mongolia served as a nucleus around which arcs and related subduction complexes accreted (Badarch et al., 2002). These younger terranes amalgamated throughout the early Paleozoic to early Mesozoic during the development of the Central Asian Orogenic Belt, which records the suture of the Siberian and North China cratons (e.g., Takahashi et al., 2000; Badarch et al., 2002; Jahn et al., 2004; Jolivet, 2015). Cenozoic deposition in the central Hangay Mountains is dominated by the accumulation of alluvium in shallow basins and widespread basaltic volcanism. Quaternary glaciations have left large moraine complexes and valleys filled with alluvium, while small Holocene cinder cones occur sporadically throughout the region (Cunningham, 2001; Tiberi et al., 2008).
Currently, central Mongolia experiences ~5 mm yr\(^{-1}\) of northward-directed shortening related to the India-Asia collision, and global positioning system (GPS) measurements indicate that the region is also moving eastward with respect to stable Eurasia (Calais et al., 2003). Shortening, as well as related E-W left-lateral shear is accommodated by a ~350 km-long system of sinistral strike-slip and normal faults flanking the north and south sides of the inherently intact block of the Hangay Dome (Baljinnyam et al., 1993; Calais et al., 2003; Bayasgalan et al., 2005; Walker et al., 2007; Jolivet et al., 2013). The long-term kinematics of the E-W fault system across central Mongolia appear to be consistent with the modern GPS strain field, suggesting that unlike in the Mongolian Altai to the west, there is no requirement for vertical axis rotation of the Hangay region (e.g., Bayasgalan et al., 2005). The Hangay Dome is situated between two prominent zones of E-W striking left-lateral faults, the Gobi-Altai to the south and the Bulnay fault to the north, both of which have generated great (M\(_{w}\) 8+) earthquakes in the twentieth century (e.g., Baljinnyam et al., 1993; Schlupp and Cisternas, 2007). In contrast to its surroundings, the interior of the Hangay Dome has had no significant instrumentally-recorded earthquakes (Walker et al., 2015).

Late Cenozoic slip associated with extensional faulting in the transfer zone between these two fault systems, near the center of the Hangay Dome, is also consistent with the regional modern strain field, and likely began after regional uplift (San’kov et al., 2000; Cunningham, 2001; Walker et al., 2008; Walker et al., 2015).

Running along the axis of the Hangay Dome is a NW-SE trending belt of flat-topped peaks that defines the catchment divide between waters draining northward into Lake Baikal and the Arctic Ocean and waters that flow west and south into the internally-drained basins of the Mongolian Valley of Lakes and Gobi Desert. These flat-topped peaks, or summit plateaus, have previously been interpreted as uplifted remnants of a Jurassic-aged peneplain (e.g., Devyatkin, 1975; Cunningham, 2001; Jolivet et al., 2007; West et al., 2013). Elevations of these low-relief, shallow-gradient surfaces are also coincident with estimates of the regional mean Quaternary glacial equilibrium line altitude (ELA), thereby suggesting that climate-controlled glacial and periglacial processes may also be responsible for their current morphology (e.g., Brozović et al., 1997; Lehmkühl et al., 2004; Vassallo et al., 2007; Egholm et al., 2009; Arzhannikov et al., 2012; Lyons et al., 2013; Margreth et al., 2016). In addition, the widespread distribution of glacial landforms such as cirques, steep U-shaped valleys, and moraines indicate that the range’s high relief has been recently generated at
least in part by extensive Quaternary glacial activity (Lehmkuhl & Lang 2001; Lehmkuhl, 2004; Rother et al., 2014).

Located ~2000 km from the nearest active plate tectonic boundary, the India-Eurasia collision zone, the precise origin of the epeirogenic Hangay Dome remains enigmatic. To date, there have been many hypotheses put forth in an attempt to explain the formation of the range’s high relief and peak heights. These hypotheses include, but are not limited to, far-field effects of India-Asia convergence, Pacific plate subduction, mantle plume/hot-spot activity, asthenospheric upwelling, magmatic underplating of the crust, and mantle flow diversion around a crustal keel (Molnar and Tapponnier, 1975; Windley and Allen, 1993; Cunningham, 2001; Petit et al., 2002, 2008; Yanovskaya and Kozhevnikov, 2003; Bayasgalan, 2005; Tiberi et al., 2008; Savatenkov et al., 2010). In addition to the breadth of proposed origins, the timing of uplift also remains uncertain. For instance, on the basis of tilted sediments, Devyatkin (1975) originally argued that uplift and doming began in the middle Oligocene. This hypothesis was supported by the initiation of significant Cenozoic alluvial sedimentation in southern flanking basins of the Hangay (northern margins of the Gobi Valley of Lakes) in mid-upper Oligocene time (Yanshin, 1975; Höck et al., 1999). Recent oxygen and carbon isotopic analyses from paleosols preserved in these same basin sediments and moisture transport pathway modeling by Caves et al. (2014) also suggest that an increase in the elevation of the Hangay Dome with respect to surrounding depositional basins occurred during the Oligocene. Caves et al. argue that the onset of aridity in basins south of the Hangay during the early Oligocene implies contemporaneous surface uplift of the Dome, and although aridification of central Asia has also been linked to a global cooling event at the Eocene-Oligocene transition (e.g., Dupont-Nivet et al., 2007; Kraatz and Geisler, 2010), the diachronous record of aridity in basins across the Hangay suggests an east to west growth of topography that likely accentuated the wider climate trend (Caves et al., 2014). However, contrary to this “old Hangay” hypothesis, other researchers have argued that uplift of the Hangay was coeval with development of the Altai mountain range circa 4 Ma (Yarmolyuk et al., 2008). Evidence cited in support of this “young Hangay” hypothesis includes a marked shift in the style of volcanism around this time (Yarmolyuk et al., 2008). This change in eruptive behavior is well documented, with recent $^{40}$Ar/$^{39}$Ar ages for basalts from the Hangay region indicating that multiple episodes of laterally extensive flows occurred between ~28 and ~4 Ma, followed by a later stage of less voluminous, valley-filling eruptions between ~4 Ma and the present. The locations of these
flows have no discernable spatial age progression, but do indicate that the greater Hangay region has experienced volcanism for the past 30 Myr (Ancuta et al., in review).

Regardless of constraints placed on the timing of an underlying process responsible for the development of the Hangay Dome as a topographic feature, the arid-to-semiarid climate and periglacial environment of the central Hangay likely preclude rapid landscape evolution, and recent research suggests that the discrepancy between the current topographic form of the Hangay Dome and the associated climatic and erosional gradient is indicative of a significant lag time for the geomorphic response to external forcing (West et al., 2013). In summary, a single hypothesis that explains these previous disparate observations has not been tested.

1.4 Methodology

This study is focused within and around the Egiin Davaa region of the central Hangay Mountains, a roughly 2,500 km$^2$ area uniquely characterized by flat-topped summits (3000+ m in elevation) that are capped by Cenozoic basalt flows (Fig. 3). These basalts unconformably overlie Paleozoic granites of the Hangay batholith (Takahashi et al., 2000; Jahn et al., 2004), and the undulating contact between the volcanic and plutonic rocks preserves the paleotopographic surface that existed at the onset of volcanism (Bayasgalan et al., 2007). The valleys exposing this contact are steep-walled and lead upstream to headward-retreating glacial cirques that supply boulders and coarse sediment to the channel bottoms.

Following our three hypotheses, the study methodology includes three primary components:

i. The reconstruction and subsequent analysis of paleotopography using Landsat imagery and ArcGIS software.

ii. A calculation of total erosion since emplacement of the uppermost, youngest basalt flow using a combination of basalt $^{40}$Ar/$^{39}$Ar ages and cosmogenic $^{10}$Be erosion rates.

iii. A comparison of total erosion to estimated glacial erosion during the late Pleistocene last glacial period (~40,000 to ~10,000 yr BP).
1.4.1 Quantifying Mid-Miocene paleotopography

Field observations of deep, basalt-filled paleo-valleys motivated our reconstruction of the relict topographic surface for the quantification of relief present at the time that Cenozoic lava flows were emplaced across the Egiin Davaa region of the central Hangay Mountains (Fig. 4). For this analysis, the study area was limited to the ~500 km² area outlined in Fig. 5, which includes all locations where the basalt/basement contact was visible, but excludes the area to the north of the NW-SW trending Egiin Davaa normal fault in order to avoid biases introduced by the ~225 m of slip that has occurred during and since emplacement of the basalt flows (Walker et al., 2015). The study area also includes the majority of flat-topped basaltic summits present in the central Hangay, and the known age range of these basalt flows provides time constraints for our geomorphic analyses. Ancuta et al. (in review) obtained \(^{40}\)Ar/\(^{39}\)Ar ages from ~230 basalt flows throughout the Hangay Dome, of which 24 are located within this study area (Fig. 5). Four of these samples, with ages between 12.9 and 9.2 Ma, were collected specifically from basalt flows in contact with the underlying basement rock. As a result, it is assumed that the paleotopographic surface preserved beneath the flows represents the variation in surface elevations across the central Hangay approximately 13 Myr before present.

To perform a quantitative analysis of this paleotopography, the spatial extent of ridge-top basalts was mapped using 30 m-pixel resolution Landsat Thematic Mapper (TM) imagery. Several cloud-free images were available from http://earthexplorer.usgs.gov, and an image acquired on July 27, 2006 by the Landsat 5 satellite was chosen for its minimal snow and vegetative cover. After selection, the Landsat TM image was imported into the ArcMap 10.1 software environment and displayed using various band combinations. Basalts were mapped using a 1-5-6 band combination (1 = red, 5 = green, 6 = blue) as this combination maximized the color contrast between basalts and the underlying granitic basement rock (Figs. 3 & 6). The mapped contact was then verified in the field at several locations in June, 2013; excellent agreement between GPS coordinates and remotely-mapped basalt/basement contacts (within the uncertainty of the Landsat image resolution) did not necessitate any modifications to mapping conducted solely with Landsat imagery and ArcMap.
The ‘pre-basalt’ paleotopographic land surface was approximated using a 30 m ASTER Digital Elevation Model (DEM). ASTER data were retrieved from the online USGS EarthExplorer tool, courtesy of the NASA Land Processes Distributed Active Archive Center (LP DAAC), USGS/Earth Resources Observation and Science (EROS) Center, Sioux Falls, South Dakota, https://earthexplorer.usgs.gov/. The process of approximating this surface was initiated by constructing points along the basalt/basement contact at 100 m intervals and then deriving an elevation for each point from the DEM. Based on these point elevations, the paleo-surface (paleo DEM) was constructed using the kriging method contained in the Geostatistical Analyst Extension to ArcMap. For the interpolation, we implemented ordinary kriging with a K-Bessel model type and assumed a 15% mapping error for the basalt/basement contact. In practical terms, this approach resulted in the ‘removal’ of the ridge-top basalts from the modern topography in order to represent a 3-dimensional approximation of the basalt/basement contact (Fig. 7). Uncertainty associated with the interpolated paleotopographic surface is heavily dependent upon the proximity of the mapped basalt/basement contact, and ranges from <20 m along the contact itself to >100 m in portions of the study area that are >3 km from the nearest mapped contact location. After interpolation of the paleotopographic surface, ArcMap’s Spatial Analyst tools were used to extract topographic swath profiles from the paleo and modern DEMs in order to quantify and compare paleo and modern relief. A total of eleven swath profiles were extracted, of which five cross the study area in a SW-NE direction and are perpendicular to the additional six profile lines that trend NW-SE. Minimum, mean, and maximum elevation profiles were plotted for each swath using the minimum, mean, and maximum DEM cell (cell area = 90 m$^2$) elevations within circular focal windows of 1 km radii located every 500 m along the length of the swath. These series of overlapping circles defined 2 km-wide swaths, and elevations within each circular window were derived using the focal statistics tool contained in the Spatial Analyst Extension to ArcMap. Plotting the swath profiles provided visual aids for comparing topographic relief between the modern and relict landscapes (Fig. 8).

In addition to minimum, mean, and maximum values, the elevation range among the 90 m$^2$ DEM cells within each 3.14 km$^2$ circular window was calculated in order to quantify the paleo and modern topographic relief at the same 500 m intervals along each swath profile. Since the swath profiles vary in length between 7 and 30 km, relief statistics were aggregated into 5-km or smaller segments in order to yield comparable mean and maximum relief values.
across profiles. Uncertainties in the tabulated topographic data include a $2\sigma$ (95%) vertical accuracy of 20 m for ASTER-based modern elevations (LP DAAC) and spatially heterogeneous uncertainties for interpolated paleotopographic elevations, which can exceed 100 m in areas distant from the mapped basalt/basement contact. However, we do note that uncertainty of the interpolated surface is relatively low (<50 m) across the central portion of our study area where the contact is well-mapped and paleotopographic relief is highest, as supported by field observations. As such, we feel confident that that interpolated surface is an accurate reconstruction of paleotopography in much of the study area, and are able to quantitatively assess, and thus objectively compare, the degree of relief present in the modern and relict surfaces at various spatial scales and locations.

1.4.2 Estimating erosion since the end of volcanism

Ridge-top basalt flows were dated with step-heated $^{40}$Ar/$^{39}$Ar on groundmass in order to determine the timing of emplacement for the youngest flows in the study area, which encompasses all the basalt-capped high terrain near Egiin Davaa (Fig. 5). Nine ridge-top flows analyzed by Ancuta et al. (in review) span 9.6 – 4.1 Ma, and based on these ages it is estimated that between 9.6 and 4.1 Ma the study area was covered by a thick and continuous series of lava flows that filled and obscured all pre-existing valleys. Based on this model of valley filling, the vast majority of relief present in the study area today has been created as a result of erosion since the late Miocene to early Pliocene. However, the ages of ridge-top basalt flows do not provide any information on the amount of summit plateau erosion that has occurred since the cessation of volcanism. Published $^{10}$Be summit erosion rates of 12 – 20 m Myr$^{-1}$ for the Sayan mountain range located north of the Hangay Dome in southern Siberia (Jolivet et al., 2013) and 28 m Myr$^{-1}$ for the Ih Bogd massif in the Gobi Altai south of the Hangay Dome (Jolivet et al., 2007) suggest that summit plateau erosion in this region of Asia is not negligible when considered over several million years. As such, erosion rates were determined with $^{10}$Be measured in seven ridge-top TCN samples collected from various summit plateaus comprised of granitic bedrock located across the central Hangay (Fig. 9).

Owing to an approximately exponential decrease in cosmic ray flux through rock as a function of depth beneath the Earth’s surface, the concentration of TCN is inversely proportional to the erosion rate. If a rock has not been exposed for sufficient time to achieve
dynamic equilibrium, the concentration will underestimate the actual steady erosion rate. Thus site-specific $^{10}$Be erosion rates are interpreted to reflect minimum estimates of the long-term erosion rate of the summits unless we can validate the assumption that the concentration is at dynamic equilibrium. The exposure time necessary to obtain dynamic equilibrium shortens with increasing erosion rate, so surfaces experiencing rapid erosion rates can obtain dynamic equilibrium in centuries, whereas very slowly eroding surfaces take over 3 Myr. Previous studies have observed long, slow rates of denudation for low-relief summit plateau surfaces (e.g., Anderson, 2002; Margreth et al., 2016), and considering available knowledge of the tectonic and climatic regime of the Hangay Dome since the mid-Miocene (e.g., Caves et al., 2014), it is probable that the summit TCN concentrations have reached dynamic equilibrium and are controlled only by erosion. As such, the range of TCN erosion rates measured throughout the Dome establishes upper and lower bounds for the average pace of summit erosion since eruption of the uppermost basalt flow.

All TCN erosion rate samples were collected using a rock hammer and chisel from the upper 3 cm of well-exposed and intact (regolith-free) bedrock outcrops, including tors. No correction for topographic shielding was necessary owing to the high, flat landscape position of the sample sites, and no correction for snow cover was made since we assume seasonal snow depth is insignificant (<1 m) due to the arid climate of the region, especially during the cold winter months. Although it is possible this assumption is invalid and snow cover has indeed influenced TCN concentrations of the samples, this yields a small error relative to the age range (9.6 – 4.1 Myr) used to calculate total erosion. TCN samples were processed at the Georgia Institute of Technology Cosmogenic Nuclide Geochronology Laboratory according to procedures outlined in Kohl & Nishiizumi (1992) and Frankel et al. (2010) for beryllium extraction from quartz. Samples were analyzed by accelerator mass spectrometry at the Purdue Rare Isotope Measurement Laboratory (PRIME Lab), and erosion rates were calculated from $^{10}$Be concentrations using the CRONUS-Earth online calculator, version 2.2 (http://hess.ess.washington.edu/; Balco et al., 2008).

The determination of summit plateau $^{10}$Be erosion rates for the past few million years allowed for a more robust estimation of the total volume of rock that has been removed from the study area since emplacement of the uppermost basalt flow. Thus, the combined approach used to calculate total incision was to quantify the valley relief observed below the
study area's summit plateaus and add an additional thickness of basalt that would have been eroded over the past 4.1 – 9.6 Myr according to the mean rate of summit plateau erosion derived from TCN measurements (Fig. 10). We recognize that the minimum and maximum estimates of summit plateau erosion are possibly from more competent rocks than the often highly fractured basalts in the valley, but the variation in erosion among different rock types is difficult to predict. Fortunately, this contributes a small error when we consider that the incision of the post-lava topography is much greater than the slow erosion of the granitic portion of the summit plateaus from which the $^{10}$Be-derived erosion rates are calculated.

In addition to the 9.6 – 4.1 Myr range used to calculate summit denudation, uncertainty in the calculation of total rock volume removed from the study area includes our assumption of a flat, continuous basalt plateau that existed prior to the onset of erosion. Like the assessment of paleotopography, the reconstruction of this surface includes a 2σ (95%) vertical uncertainty of 20 m for ASTER-based modern elevations (LP DAAC) and spatially heterogeneous uncertainties for the interpolated pre-erosion surface, which are small (<15 m) in areas where basalt is present, but can exceed 100 m at points distant from modern basalt-capped summits. To deal with this uncertainty, we implement a conservative estimate of 60 m for area-wide vertical uncertainty, which, when coupled with the 9.6 – 4.1 Ma age range used for the onset of erosion, provides confidence that we are encompassing all possible erosion scenarios.

Isolated flat-topped summit plateaus in the study area were connected to form a continuous high-elevation plateau by generating an interpolated surface via the ordinary kriging method with a K-Bessel model type contained in the Geostatistical Analyst Extension to ArcMap (Fig. 11). This interpolated surface roughly approximates what we believe was a continuous basalt plateau at the time of the latest lava flow. The additional thickness of rock was added above this surface to accommodate for erosion of the summit that has occurred since dissection of the plateau began. We assume a constant erosion rate over this time to estimate the maximum and minimum thicknesses removed. The modern landscape was subtracted from the interpolated basalt plateau surface in order to calculate the total volume difference, which represents the amount of rock that has been eroded since the eruption of the uppermost basalt flow. Normalizing this volume to the total study area and dividing by
9.6 or 4.1 Ma establishes minimum and maximum erosion rate estimates in units of meters of vertical surface lowering per million years.

1.4.3 Comparing long-term erosion to Late Pleistocene glacial erosion

Several studies have used volumes of sediment stored in moraines to establish both glacial sediment budgets and the flux of sediment leaving a catchment over a specific period of time (e.g., Small, 1987; Delmas et al., 2009; Züst et al., 2014). This research provides a framework for estimating the rate of erosion during a glacial episode by calculating the volume of a well-preserved terminal moraine, estimating the time span of moraine formation, and considering various ratios of terminal moraine storage to overall glacial erosion.

During the 2011 and 2012 field seasons, rock samples were collected from multiple moraines throughout the central Hangay for TCN exposure dating, which assumes that a stable geologic surface will accumulate cosmogenic nuclides with exposure time according to a location-dependent production rate (Gosse and Phillips, 2001; Dunai, 2010). A total of twelve samples from granitic boulders at three terminal moraines located within ~100 km of Egiin Davaa (Gilgar Uul, Khaak Nuur, and Chuluut Gol) were utilized for this study (Fig. 12). Samples were processed at the Dalhousie University Geochronology Centre according to procedures outlined in Kohl & Nishiizumi (1992) and Hidy et al. (2013) for beryllium extraction from quartz. Processed samples were subsequently analyzed at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory (CAMS-LLNL) and exposure ages were calculated from $^{10}$Be concentrations using the CRONUS-Earth online calculator, version 2.2 (http://hess.ess.washington.edu/; Balco et al., 2008). For our calculations we assumed negligible boulder erosion since the time of deposition, no topographic or mass shielding, and no inheritance of cosmogenic nuclides from previous exposure.

Samples from these sites were selected not only because they are in close proximity to Egiin Davaa, but also because each terminal moraine is large (hundreds of meters wide), relatively well defined, and situated on stream reaches with uncomplicated valley floor topography, all of which minimize the total error in estimating moraine volume (Fig. 13). In ArcGIS, the volume of each moraine was calculated by interpolating a basal surface that approximates the elevation and topography of the valley floor that existed at each location.
prior to the glacial event that led to the deposition of the moraines. The basal surface was constructed using the ordinary kriging method with a K-Bessel model type contained in the Geostatistical Analyst Extension to ArcMap, and is based on elevations of the modern valley floor that are just upstream and downstream of each moraine. To determine total moraine volume, the basal surface was subtracted from the modern 30 m ASTER-based topography of each moraine. These moraine volumes are considered to be minimum estimates since we did not account for erosion of the moraines, and also because we believe that the elevation of the interpolated basal surface is likely higher than the elevation of the true valley floor that existed prior to moraine deposition.

The $^{10}$Be surface exposure ages from terminal moraines were used to make estimates for the time it took each terminal moraine to be deposited. Constraints on the timing of moraine abandonment, coupled with existing literature on the extent and period of OIS-2 and OIS-3 ice advances in the Hangay mountains (e.g., Lehmkuhl et al., 2004; Rother et al., 2014) as well as time spans of moraine deposition at similar latitudes in other regions of the world (e.g., Phillips et al., 1997; Ivy-Ochs et al., 2004; Briner et al., 2005; Licciardi & Pierce, 2008), allowed us to make minimum and maximum estimates for the duration of each glacial event that contributed to individual moraine deposition. Multiple ratios of moraine sediment storage to total glacial erosion were also considered since previous research suggests that moraines always underestimate the total sediment volume produced by a glacier (e.g., Larsen and Mangerud, 1981; Small, 1987; Sanders et al., 2013). For each ratio, minimum and maximum rates of glacial erosion for individual valleys were estimated by extrapolating moraine volume to yield a total volume of glacial erosion, dividing the total volume by the duration of deposition, and normalizing this value to the total upstream catchment area (Fig. 14). Rates of glacial erosion were compared to (1) the estimated rates of erosion since 9.6 – 4.1 Ma and (2) basin average erosion rates spanning $\sim 10^4$ years calculated from $^{10}$Be concentrations in river sands (Hopkins, 2012).
1.5  Results

1.5.1  Modern versus paleo topographic relief

Averaging across all 11 swath profiles, the modern and relict landscapes have mean relief values of 388 and 187 m, respectively (Table 1). These means represent averaged individual relief values calculated using a circular window with a 1 km radius (see Section 3.1). Dividing the profile lines into segments of 5 km (or less to include end segments of profiles not divisible by 5) reveals that, for the modern landscape, the average maximum relief along each swath profile segment is 626 m and the greatest relief along any single profile segment is 818 m (profile line 10). For the relict surface, which represents the landscape prior to the onset of volcanism roughly 13 Ma (Ancuta et al., in review), the average maximum relief along each swath profile segment is 344 m, and the greatest relief along any single profile segment is 606 m (Table 1). Considering the entire study area (340 km$^2$), there is a maximum relief of 1026 m in the modern landscape versus 903 m in the paleo-landscape.

1.5.2  Long and short-term erosion

The spatial analysis of erosion yields an estimated 118 ± 32 km$^3$ of rock that has been removed from the study area since 9.6 – 4.1 Ma. TCN results yield a range of summit erosion rates varying from 2.8 – 15.4 m Myr$^{-1}$, with an average rate of 8.1 m Myr$^{-1}$ (Table 2). Based on the normal distribution of rates, we assume 8.1 m Myr$^{-1}$ is a good approximation of late Miocene-to-present summit surface lowering throughout the study area, and accounting for the range of basalt summit plateau surface ages (9.6 – 4.1 Ma), utilization of this rate for calculation of summit surface lowering indicates the removal of an additional 17 – 41 km$^3$ of rock, which results in a total volume removal of 135 – 159 ± 32 km$^3$ in the region analyzed (Table 3). Normalizing to the study area, dividing by the various erosion duration scenarios, and including all tabulated uncertainties, this volume yields an erosion or surface lowering rate of 25.1 – 77.2 m Myr$^{-1}$, with an average rate of roughly 41 m Myr$^{-1}$ for the midpoint (6.9 Ma) of the 9.6 – 4.1 Ma age range of the uppermost basalts (Table 3).

Moraine sediment volumes, catchment areas, and estimates of total erosion for scenarios in which moraine volume represents 20%, 50%, or 80% of all material eroded out of the
catchment during the duration of moraine deposition are reported in Table 4. The 20%, 50%, and 80% scenarios were used in order to make conservative estimates of minimum and maximum glacial erosion rates since studies of moraine sediment budgets throughout the world have calculated or assumed that end moraines store anywhere from 20 – 80% of the total material eroded by glaciers from a valley during a period of glaciation (e.g., Larsen and Mangerud, 1981; Small, 1987; Sanders et al., 2013). This wide range covers a broad spectrum of moraine storage in glaciated landscapes across the world and incorporates such uncertainties as the percentage of meltwater sediment output, the degree of subglacial deposition, and the possible incorporation of material leftover from a previous glacial advance. For the 10Be moraine boulder samples collected as part of this study, the mean exposure ages were 30.6 ka (n = 3; 1σ = 15.2 ka), 21.8 ka (n = 4; 1σ = 2.0 ka), and 15.4 ka (n = 2; 1σ = 0.28 ka) for the Khaak Nuur, Chuluut Gol, and Gilgar Uul end moraines, respectively (Table 5). This distribution of exposure ages does not provide direct constraints on the duration of moraine deposition, but it does allow us to conclude that each moraine was deposited during the late Pleistocene last glacial maximum. In the context of the Quaternary glacial chronology that has been published for the Hangay and Sayan ranges, as well as estimates regarding the time span of moraine formation as a percentage of the total duration of a glacial interval (Gosse et al., 1995; Lehmkuhl et al., 2004; Delmas et al., 2009; Arzhannikov et al., 2012; Rother et al., 2014; Züst et al., 2014), this provides the basis for assuming a range of 1000 – 10,000 years for the duration of glacial erosion that led to the formation of each moraine. Although 10,000 years is a longer window of time than most estimates of last glacial maximum moraine deposition (e.g., e.g., Phillips et al., 1997; Ivy-Ochs et al., 2004; Briner et al., 2005; Licciardi & Pierce, 2008), this enables us to partly compensate for uncertainties in the history of moraine deposition at Gilgar Uul, where the 15.4 ka mean exposure age comes from boulders situated within a smaller recessional moraine located ~5 km upstream of the larger terminal moraine used for the erosion rate calculation. The terminal moraine was likely formed during the last maximum extent of Hangay glaciers that took place between 40 and 35 ka during OIS 3 (Rother et al., 2014), in which case it’s possible that later re-advances, while failing to reach this same extent, may have contributed some volume of sediment to the moraine. As such, the 1000 – 10,000 yr time window, in conjunction with the various ratios for moraine volume to total glacial erosion, represents the attempt to include this type of uncertainty into the calculated range of glacial erosion rates. Using these parameters, and assuming a porosity of 20% for moraine sediment based on measurements made by Burki et al. (2010), glacial erosion
rates derived from moraine volume range from 29 – 3700 m Myr\(^{-1}\) for the three glacial valleys utilized in the study (Table 6).

1.6 Discussion

Field and geochronologic evidence indicate that at approximately 13 Ma volcanic activity began producing a series of voluminous basalt flows that gradually accumulated and ultimately covered the pre-existing topography in the Egiin Davaa region of the central Hangay Mountains (Figs. 4, 5, & 10). The Landsat and GIS-based paleotopographic reconstruction presented here indicates that this pre-basalt landscape was characterized by a degree of topographic relief comparable to that of the present day, with >600 m elevation differences between valleys and ridgelines over horizontal distances of a few kilometers. Although analysis of the 13 Ma landscape yields an estimate of paleo-relief that is approximately 50% of the magnitude of relief in the modern landscape (Table 1), which seems to suggest a paleotopographic surface that is comparably subdued, the estimate of paleo-relief is considered an extreme minimum since our ability to thoroughly map the basalt/basement contact is limited by the available exposures, causing an underestimation of paleo-relief in the region (it is unlikely that the reach with the greatest relief was ever exposed). As a result, we can make minimum estimates for the depths of paleo-valleys that are exposed in along modern valley walls, but we cannot assess the total depth of a paleo-valley that extends below the elevation of the modern valley, and in these cases the true paleo-relief at this location may be much greater than our estimate. As a result of this limitation, the maximum relief observed in each topographic swath profile segment of the paleo-landscape more likely represents a value closer to the mean relief of the paleo-landscape within each segment since the interpolated surface is based only on known elevations of the basalt/basement contact, it is unlikely that the calculated paleo-relief is an overestimate. Under this assumption, the fact that the average maximum relief across all profile segments of the paleo-landscape is 107% of the mean relief present in the modern landscape suggests the amount of relief that was present during the mid-Miocene is more or less equivalent to that which exists today (Table 1). Even if this maximum relief versus mean relief assumption is invalid, the assessment of the paleo-landscape nevertheless indicates that locally, relief exceeded 600 m across short horizontal distances (<3 km) prior to 13 Ma, which we interpret as evidence suggesting (1) that by 13 Ma this region of the
central Hangay was sufficiently high relative to the regional surroundings that ≥600 m deep valleys could have been carved; and (2) that there has not been an abrupt change in tectonic forcing since that time.

Based on ages of ridge-top basalt flows, volcanism continued to cover the granitic paleo-landscape for several million years, with the final ridge-top flows erupting by ca. 4 Ma. At this point, erosive forces, as opposed to volcanic deposition, began to shape the landscape through the carving of the drainage network visible in the study area today. After correcting for summit erosion and using a 9.1 – 4.6 Ma window to estimate uncertainty surrounding the onset of this erosive regime, calculation of the long term erosion rate over the past several million years ranges from 25 – 77 m Myr\(^{-1}\). Considering that erosion rates measured in areas of active uplift are typically several hundred meters per million years or higher (Montgomery and Brandon, 2002 & references therein), this estimate is at least one order of magnitude lower than what might be expected if the Hangay range was experiencing a rapid rate of tectonic uplift in the past 4 Ma.

Bolstering this assertion is the general agreement of our results with modeled exhumation rates based on low temperature detrital thermochronology, which were calculated to be ≤30 m Myr\(^{-1}\) over the last 130 million years for both the Selenga and Orkhon River basins located just north of our study area (McDannell et al., 2014). Furthermore, using the modal size of vesicles trapped in basalt flows from the Egiin Davaa area as a paleo-elevation proxy, Sahagian et al. (2014) concluded that the Hangay Dome has experienced roughly 1 km of uplift since the eruption of 9 Ma basalt flows. Although the data from Sahagian et al. cannot provide additional constraints on the onset of uplift, such as whether it has been gradual or episodic in nature, the ~0.1 km Myr\(^{-1}\) rate is coherent with the model of a slowly evolving mountain range and the possibility that uplift commenced prior to 9 Ma and well exceeds ~1 km. Our findings are also in good agreement with basin averaged erosion rates measured using \(^{10}\)Be from catchments throughout the central Hangay. These rates, which integrate the signal of erosion over the past tens of thousands of years, range from 12 – 20 m Myr\(^{-1}\) (Hopkins, 2012) and provide further evidence that the Hangay Dome lacks an erosional signal of a recent, rapid change in tectonic forcing. Overall, the consistency of our rates with \(10^4\) yr erosion rates and \(10^7\) yr exhumation rates provides a compelling argument for a stable tectonic regime that has been in place for tens of millions of years. Rather than tectonic forcing, our estimated rates of glacial erosion from moraine volumes suggest that
climate is the primary driver of erosion throughout this region of the central Hangay, and that repeated glacial episodes have caused extensive erosion over the past 2.6 Ma, albeit at a slower rate during the shorter, relatively warm periods such as the present interglacial.

A simple conceptual model of landscape evolution can be formulated from the above interpretations. Beginning ~13 Ma the topography of the Egiin Davaa study area was essentially in-filled and smoothed by basaltic lava flows over a period of ~4 to ~10 Myr. However, assuming that the $^{10}$Be basin-average erosion rates of $12 - 20$ m Myr$^{-1}$ are representative of the pace of erosion throughout the central Hangay, which includes lower elevation areas that were not extensively glaciated, it is reasonable to infer that glacial activity has been a significant and effective agent of erosion at Egiin Davaa, thus yielding the $25 - 77$ m Myr$^{-1}$ average rate of erosion calculated since $9.6 - 4.1$ Ma. Although the uncertainties in moraine storage and duration of glacial activity result in variable rates of erosion, the conservative median estimate that each moraine stores 50% of the material eroded from the catchment and that the moraine was deposited over a span of 5000 years yields a glacial erosion rate of ~$190$ m Myr$^{-1}$ averaged across all three studied catchments. At this rate, the current topography of the Egiin Davaa region and other high elevation areas of the central Hangay could have been shaped predominantly by glaciers and thus reflect the product of climatic forcing. This notion of climate-controlled erosion is supported by the presence of an extensive granitic saprolite within the upper Orkhon Gol valley that is buried by a basalt flow with an eruptive age of $3.05 \pm 0.06$ Ma (Ancuta et al., in review). Since the modern climate regime of central Mongolia precludes saprolite formation, it is evident that a significant climate shift has occurred in the past 3 Myr, and this shift undoubtedly changed the nature of weathering and erosion across the Hangay Dome.

Further evidence for late Cenozoic climate change as the sculptor of modern relief is the lack of any discernable sedimentary veneer between the Paleozoic and older basement and the overlying Cenozoic basalt flows. This is true for paleo-valleys as well as paleo-summits, which suggests that the paleo, or pre-basaltic, landscape was dominated by a bare bedrock surfaces and limited valley fill. Although previous research has shown that lava flows are capable of significant erosion (e.g., Greeley et al., 1998; Williams et al., 2004; Ferlito and Siewert, 2006; Siewert and Ferlito, 2008; Kerr, 2009; Williams et al., 2011), we consider it highly unlikely that the basalt flows in the study area would have eroded the pre-existing landscape to such a degree that there would be such sparse field evidence of preserved
sediments anywhere along the basalt/basement contact. Furthermore, the preserved granitic saprolite in the Orkhon Gol valley suggests that weathered or eroded sediments would have been readily preserved had they indeed existed atop the Paleozoic bedrock. The interpretation of an environment characterized by limited weathering and erosion is in stark contrast to the modern central Hangay, where the majority of valleys have been back-filled with sediment during the Quaternary, and rivers are entirely alluvial and appear to exhibit minimal sediment transport capability at present (Fig. 15). This discrepancy in sediment production is hypothesized to have resulted in response to the Pliocene-to-Quaternary climate trend that gradually created a more erosive environment. In addition to a consistent trend of global cooling throughout the Cenozoic (e.g., Zachos et al., 2001), analyses of sedimentary records throughout central Asia indicate significant changes in regional climate beginning in the late Miocene. Specifically, studies from Lake Baikal and the Chinese Loess Plateau, which have correlative paleoclimate records (Song and An, 2010), reveal an intensification of the Asian winter monsoon ca. 8 Ma followed by the rapid onset of major, Milankovitch-like climate oscillations shortly after 4 Ma (Williams, et al., 1997; Kashiwaya et al., 2001; Kashiwaya et al., 2003; Takamatsu et al., 2003; Fan et al., 2006; Xu et al., 2012). The sedimentary signal of this change is recorded in basin stratigraphy as a sudden increase in mass accumulation rates, and has been identified in the internally-drained basins to the west and south of the Hangay Dome as well as in Lake Baikal to the north, into which sediments from the northern flank of the Hangay are deposited via the Selenga River system (Devyatkin, 1981; Kashiwaya et al., 2001). Amplified sediment accumulation in these basins suggests increased erosion of the surrounding high terrain (i.e. the Hangay Dome), which was likely spurred by the continuous growth and decay of glaciers in a cooler and more rapidly fluctuating climate regime. Thus, even in the absence of tectonic forcing, high latitude and high elevation regions such as the central Hangay were able to develop significant topographic relief as the result of quasi-periodic glacial/interglacial cycles.

Nevertheless, one alternate explanation for slow landscape evolution at Egiin Davaa is the possibility that significant tectonic uplift persisted throughout the mid-to-late Miocene and created an expansive high plateau (the Hangay Dome), but that this uplift is not expressed in the calculated erosion rates at Egiin Davaa due to the effects of valley-filling lava flows. In this scenario, the 9.6 – 4.1 Ma basalt flows at Egiin Davaa halted erosion by filling headwater valleys and causing fluvial incision to re-initiate further downstream. Without
valleys to focus erosion at the drainage divide, fluvial erosion propagated very slowly upstream from the margins of the lava flows, yielding an anomalously low erosion rate for the Egiin Davaa region. Headwater valleys finally reformed as icecaps grew atop summits in the study area, ultimately sculpting the modern relief and landscape. Thus, uplift may have been continuous and significant, but the geomorphic response to uplift was muted across the study area since the pre-existing peripheral valley network was filled by lava.

Although this scenario explains some of the observations made at Egiin Davaa, and our results do not preclude the possibility of minor tectonic uplift (< 1 km) since the late Miocene, multiple lines of evidence provide much greater support for the interpretation that tectonic forcing has played a negligible or very minor role in the evolution of the modern landscape, and that climate change best explains the observations and data from this study. This evidence includes the aforementioned modeled exhumation rates (McDannell et al., 2014) indicating slow, stable erosion over the past ~100 Myr as well as results from basalt vesicle paleo-altimetry (Sahagian et al., 2014) that show evidence for uplift of no more than 1 km over the past 9 Myr. Furthermore, the distribution of summit-capping basalt flows is limited to the Egiin Davaa region, and therefore this alternative scenario would not be applicable beyond the study area. However, $^{10}\text{Be}$ basin average erosion rates (Hopkins, 2012) are consistent (and low) across basins of varying size and hypsometry throughout the Hangay Dome (including Egiin Davaa), which suggests that much of the Hangay is evolving similarly. River systems are also integrated into the landscape as evidenced by equilibrium channel profiles both within and outside of the study area that provide further evidence against the persistence of a relict landscape that has yet to experience a propagating wave of erosion (cf. Gallen et al., 2011; Gallen et al., 2013).

Additional support for the role of climate in Hangay Dome landscape evolution is the significant correlation between mean annual precipitation and erosion for large drainage basins throughout the range (West et al., 2013). Referencing a weaker correlation for small headwater catchments, West et al. suggested that the relationship may partly be explained by the role of glacial erosion at high elevations, but subsequently failed to find a strong, positive relationship between catchment hypsometry and erosion, which typically indicates a dominant influence of glacial processes. However, the great extent of alluvial fill in valleys across the central Hangay may be obscuring this relationship by masking a significant percentage of the true relief between ridge and valley bedrock. It's also possible that the
dominance of glacial erosion is unique to Egiin Davaa owing to the history of valley-filling eruptions that essentially reset the headwater relief structure.

Overall, our assessment of paleo-relief, long-term erosion, and the role of glaciers in shaping the modern landscape indicates that there is no discernible geomorphic signal of a recent, rapid change in uplift in our study region of the Hangay Dome. Rather, our results from the Egiin Davaa area paint the picture of a mid-Miocene landscape with pre-existing topographic relief similar (>600 m) to the present-day central Hangay Dome, but with very limited sediment production due to an arid climate that precluded swift erosion. Circa 13 Ma, the eruptive output of a concentration of high elevation volcanic centers gradually filled adjacent river valleys and led to an eventual smoothing of the local landscape. Although others have suggested that basalts are present along ridgetops as a result of topographic inversion from recent graben uplift (c.f., Yarmolyuk et al., 2008; 2015), our observations and results agree with the interpretation of Cunningham (2001) that the multiple fault-bounded half-grabens distributed throughout the Hangay Dome are only minor sediment sinks. Furthermore, it is difficult to reconcile the accumulation of basalt flows (now preserved along the continental drainage divide) in a large depositional basin with the lack of observable sedimentation at the contact between basalt flows and Paleozoic basement. As such, our research suggests that the Egiin Davaa study area was a region of high topography before 13 Ma, and that sediment was effectively evacuated from, or did not exist in, valley bottoms prior to the eruptive phase.

Despite providing evidence against recent, rapid changes in uplift, our results are limited to the ages of the earliest basalt flows at Egiin Davaa (~13 Ma), so we are unable to provide additional constraints on the cause, onset, and/or progression of Hangay Dome uplift prior to 13 Ma. However, we point to recent research in the Hangay and Sayan mountain ranges with results that align well with the evaluation of significant paleo-relief and relatively slow rates of uplift and erosion that have persisted for longer than the past 13 Myr. By measuring carbon and oxygen stable isotopes in paleosol carbonates throughout Mongolia, Caves et al. concluded that surface uplift of the Hangay began by the early Oligocene (~30 Ma), blocking moisture from Siberia and thus creating more arid conditions in the northern Gobi desert (2014). This model of epeirogenic evolution agrees with earlier work showing that significant sedimentation, presumed to be a response to increased elevation of the Hangay Dome, began in the Mongolian Valley of Lakes in the middle Oligocene (Yanshin, 1975;
Höck et al., 1999), a scenario which agrees well with our surface interpretations. Perhaps more telling is the remarkable coherence between the results of this study and those of similar work by Jolivet et al. (2013) in the Sayan Mountain range located north of the Hangay Dome near Lake Baikal. Jolivet et al. also used basalt-capped summits to reconstruct Miocene paleotopography, and in combination with long-term exhumation rates derived from apatite fission track data and short-term erosion rates derived from $^{10}\text{Be}$ data, arrived at the conclusion that the Sayan Mountains were influenced by a long-wavelength uplift beginning in the Oligocene or early Miocene. Regional application of this deduction, coupled with mean exhumation and erosion rates ($12 – 20 \text{ m Myr}^{-1}$) from the Sayan that are quite similar to those for the Hangay ($12 – 77 \text{ m Myr}^{-1}$), lends considerable support for the hypothesis that uplift of the Hangay Dome began roughly 20 – 30 Myr ago.

### 1.7 Conclusions

Observations and quantitative reconstruction of paleotopography in the Egiin Davaa region of the central Hangay Dome reveal a degree of topographic relief present before 13 Myr ago that is similar and potentially equivalent to the topographic relief found at Egiin Davaa today. Relief at 13 Ma, which exceeds 600 m in places, suggests that volcanism occurring from 13 – 4 Ma in Egiin Davaa took place outside of a depositional basin and that the topographic distribution of basalts has not changed considerably since emplacement. The lack of a sedimentary veneer between basalt flows and underlying Paleozoic basement rock supports this interpretation.

A known age range of ridge-top basalt flows provides a reference surface from which to calculate the total volume and rate of erosion since emplacement of the basalts. Incorporating the results of $^{10}\text{Be}$ TCN measurements to correct for otherwise unknown summit denudation, erosion rates across the Egiin Davaa portion of the central Hangay dome range from 25 – 77 m Myr$^{-1}$ over the past 9.6 – 4.1 Myr. This range of rates compares well with long-term estimates of exhumation ($\leq 30 \text{ m Myr}^{-1}$; McDannell et al., 2014), and is similar and of the same order of magnitude to $^{10}\text{Be}$ basin-average erosion rates throughout the central Hangay ($12 – 20 \text{ m Myr}^{-1}$; Hopkins, 2012) that provide information on the pace of erosion over the past 30,000 – 50,000 years. Compared to studies of erosion in tectonically active mountain ranges, these slow rates of exhumation
and erosion provide no conspicuous evidence for recent and/or active geomorphic response to a recent or rapid change in uplift rates in the central Hangay.

Quantification of moraine volumes enables estimation of a range of Pleistocene glacial erosion rates. These rates, although widely variable, are concentrated above 100 m Myr⁻¹ and suggest that glacial erosion is responsible for much of the relief creation that has occurred in Egiin Davaa since emplacement of mid-to-late Miocene topography-blanketing basalt eruptions. Derivation of these rates, when viewed in the light of an abrupt shift to a cooler and more highly variable climate circa 3 Ma, suggest that the dramatic relief of the modern high Hangay may have been sculpted primarily as the result of late Cenozoic climatic forcing.

1.8 Acknowledgements

Constructive comments by M. Jolivet, J. West, and an anonymous reviewer were critical to improving an earlier version of the manuscript. We would like to thank Nrangerel “Naraa” Mandah for his excellent driving and navigation throughout central Mongolia, and Gantulga Bayasgalan, Dashchariv Khorloo, Molor Erdenebat, Nathan Lyons, Tara Forstner, and Matthew Morriss for their camaraderie and support in the field. This work is supported by U.S. National Science Foundation Research Grants EAR-1009702 and EAR-1009680.

1.9 References


Herman, F., Champagnac, J., 2015. Plio-Pleistocene Increase in Erosion Rates in Mountain Belts in response to Climate Change. Terra Nova. DOI: 10.1111/ter.12186


Table 1
Topographic relief statistics for modern and paleo Hangay Dome landscapes.

<table>
<thead>
<tr>
<th>Profile segment</th>
<th>modern mean relief (m)</th>
<th>modern max relief(^a) (m)</th>
<th>paleo mean relief (m)</th>
<th>paleo max relief (m)</th>
<th>paleo mean / modern mean(^b)</th>
<th>paleo max / modern max(^c)</th>
<th>paleo max / modern mean(^d)</th>
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<td>613</td>
<td>307</td>
<td>409</td>
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<td>5-10 km</td>
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<td>129</td>
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\(^a\)Max relief is defined as the difference (m) between the lowest and highest points along the entire profile segment.

\(^b\)Value obtained by dividing paleo mean relief (m) by modern mean relief (m).

\(^c\)Value obtained by dividing paleo max relief (m) by modern max relief (m).

\(^d\)Value obtained by dividing paleo mean relief (m) by modern max relief (m).
### Table 1 (continued)

<table>
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<th>modern max relief (m)</th>
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<tr>
<td>0-5 km</td>
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<td>598</td>
<td>223</td>
<td>413</td>
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<td>0.69</td>
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<td>5-7 km</td>
<td>487</td>
<td>627</td>
<td>119</td>
<td>283</td>
<td>0.24</td>
<td>0.45</td>
<td>0.58</td>
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<tr>
<td>Average</td>
<td>415</td>
<td>613</td>
<td>171</td>
<td>348</td>
<td>0.45</td>
<td>0.57</td>
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<td>End: 47.04167 N, 100.0333 E</td>
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<td>0-5 km</td>
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<td>655</td>
<td>145</td>
<td>369</td>
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<td>5-11 km</td>
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<td>556</td>
<td>134</td>
<td>312</td>
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<td>0.81</td>
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<td>Average</td>
<td>300</td>
<td>606</td>
<td>140</td>
<td>341</td>
<td>0.51</td>
<td>0.56</td>
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<td>Start: 47.16028 N, 99.96361° E</td>
<td>End: 47.07583 N, 100.0836 E</td>
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<td>222</td>
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<td>10-12.5 km</td>
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<td>524</td>
<td>70</td>
<td>164</td>
<td>0.23</td>
<td>0.31</td>
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<td>306</td>
<td>604</td>
<td>115</td>
<td>216</td>
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<td>0.70</td>
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<td>End: 47.11222 N, 100.1225 E</td>
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<tr>
<td>0-5 km</td>
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<td>0.45</td>
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<td>5-10 km</td>
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<td>594</td>
<td>312</td>
<td>610</td>
<td>0.99</td>
<td>1.03</td>
<td>1.93</td>
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<tr>
<td>10-12 km</td>
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<td>541</td>
<td>41</td>
<td>225</td>
<td>0.18</td>
<td>0.42</td>
<td>0.99</td>
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<tr>
<td>Average</td>
<td>257</td>
<td>604</td>
<td>152</td>
<td>410</td>
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<td>0.68</td>
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<td>End: 47.15333 N, 100.1633 E</td>
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<td>1.10</td>
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<tr>
<td>5-10.5 km</td>
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<td>0.59</td>
<td>0.87</td>
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<td>743</td>
<td>163</td>
<td>418</td>
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<td>0.56</td>
<td>0.98</td>
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<td>Start: 47.24917 N, 100.1153 E</td>
<td>End: 47.19806 N, 100.1883 E</td>
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<tr>
<td>0-5 km</td>
<td>392</td>
<td>648</td>
<td>331</td>
<td>523</td>
<td>1</td>
<td>0.81</td>
<td>1.33</td>
</tr>
<tr>
<td>5-7 km</td>
<td>408</td>
<td>677</td>
<td>44</td>
<td>227</td>
<td>0</td>
<td>0.34</td>
<td>0.56</td>
</tr>
<tr>
<td>Average</td>
<td>400</td>
<td>663</td>
<td>187</td>
<td>375</td>
<td>0</td>
<td>0.57</td>
<td>0.95</td>
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</tr>
<tr>
<td>Average</td>
<td>365</td>
<td>610</td>
<td>171</td>
<td>359</td>
<td>0.49</td>
<td>0.59</td>
<td>1.07</td>
</tr>
</tbody>
</table>

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*a* Max relief is defined as the difference (m) between the lowest and highest points along the entire profile segment.

*b* Value obtained by dividing paleo mean relief (m) by modern mean relief (m).

*c* Value obtained by dividing paleo max relief (m) by modern max relief (m).

*d* Value obtained by dividing paleo mean relief (m) by modern max relief (m).
Table 2

Cosmogenic $^{10}\text{Be}$ results for summit bedrock denudation rates.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation</th>
<th>Thickness $^a$</th>
<th>Production rate</th>
<th>Shielding</th>
<th>Quartz $^d$</th>
<th>$^{10}\text{Be}/^{9}\text{Be}$ $^e,f$</th>
<th>$^{10}\text{Be}$ conc. $^g,h$</th>
<th>Denudation rate $^f,i,j$</th>
</tr>
</thead>
<tbody>
<tr>
<td>MN0711-04</td>
<td>47.5210</td>
<td>100.5699</td>
<td>2311</td>
<td>3</td>
<td>29.01</td>
<td>0.383</td>
<td>1</td>
<td>70.3550</td>
<td>0.3095</td>
<td>33.10 ± 0.50</td>
</tr>
<tr>
<td>MN0711-06</td>
<td>48.0241</td>
<td>99.7999</td>
<td>2310</td>
<td>3</td>
<td>29.19</td>
<td>0.382</td>
<td>1</td>
<td>70.1286</td>
<td>0.3005</td>
<td>61.40 ± 0.90</td>
</tr>
<tr>
<td>MN0711-08</td>
<td>47.8040</td>
<td>99.1972</td>
<td>2375</td>
<td>3</td>
<td>30.44</td>
<td>0.39</td>
<td>1</td>
<td>70.1694</td>
<td>0.3022</td>
<td>69.60 ± 1.00</td>
</tr>
<tr>
<td>MN0711-11</td>
<td>47.5336</td>
<td>98.4102</td>
<td>3250</td>
<td>3</td>
<td>53.24</td>
<td>0.505</td>
<td>1</td>
<td>70.1304</td>
<td>0.3036</td>
<td>265.0 ± 3.00</td>
</tr>
<tr>
<td>MN0711-13</td>
<td>47.5438</td>
<td>98.3734</td>
<td>3506</td>
<td>3</td>
<td>61.95</td>
<td>0.543</td>
<td>1</td>
<td>70.5136</td>
<td>0.3150</td>
<td>236.8 ± 2.90</td>
</tr>
<tr>
<td>MN0811-29</td>
<td>46.7191</td>
<td>102.3287</td>
<td>2609</td>
<td>3</td>
<td>35.11</td>
<td>0.419</td>
<td>1</td>
<td>63.6738</td>
<td>0.3111</td>
<td>31.90 ± 0.50</td>
</tr>
<tr>
<td>MN0811-38</td>
<td>47.1045</td>
<td>99.9248</td>
<td>3322</td>
<td>3</td>
<td>55.07</td>
<td>0.515</td>
<td>1</td>
<td>70.1622</td>
<td>0.3002</td>
<td>114.6 ± 1.50</td>
</tr>
</tbody>
</table>

**Mean denudation rate = 8.11**

$^a$The tops of all samples were exposed at the surface, and samples ranged from 2-4 cm in thickness with a mean thickness of 3 cm.

$^b$Constant (time-invariant) local production rate based on Lal (1991) and Stone (2000). A sea level, high-latitude value of 4.8 at $^{10}\text{Be}/g$ quartz was used.

$^c$Isotope ratios were normalized to $^{10}\text{Be}$ standards prepared by Nishiizumi et al. (2007) with a value of $2.85 \times 10^{12}$ and using a $^{10}\text{Be}$ half-life of $1.36 \times 10^6$ years.

$^d$A density of 2.7 g cm$^{-3}$ was used based on the granitic composition of the samples.

$^e$A mean blank value of 192,021 ± 70,088 $^{10}\text{Be}$ atoms $(^{10}\text{Be}/^{9}\text{Be} = 7.4 \times 10^{-15} \pm 2.7 \times 10^{-15})$ was used to correct for background.

$^f$Uncertainties are reported at the 1σ confidence level.

$^g$A $^{10}\text{Be}$ model ages include a 6 percent uncertainty in the production rate of $^{10}\text{Be}$ and a 4 percent uncertainty in the $^{10}\text{Be}$ decay constant.

$^h$Beryllium-10 model ages were calculated with the CRONUS-Earth online calculator (Balco et al. 2008) version 2.2 (http://hess.ess.washington.edu/).
Table 3
Volumes and rates of erosion for multiple uppermost basalt flow age scenarios.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Study area (km$^2$)</th>
<th>Eroded rock volume (km$^3$)$^a$</th>
<th>Added rock volume from summit erosion (km$^3$)$^b$</th>
<th>Total eroded rock volume (km$^3$)</th>
<th>Surface lowering (erosion) rate (m Myr$^{-1}$)$^a,b$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_b = 4.1$ Ma</td>
<td>175</td>
<td>17.5</td>
<td>117.7 $\pm$ 31.6</td>
<td>135.2 $\pm$ 31.6</td>
<td>62.7 $\pm$ 14.5</td>
</tr>
<tr>
<td>$A_b = 6.9$ Ma</td>
<td>526.1</td>
<td>29.2</td>
<td>146.9 $\pm$ 31.6</td>
<td>176.1 $\pm$ 31.6</td>
<td>40.8 $\pm$ 8.7</td>
</tr>
<tr>
<td>$A_b = 9.6$ Ma</td>
<td>410</td>
<td>41.0</td>
<td>158.7 $\pm$ 31.6</td>
<td>200.7 $\pm$ 31.6</td>
<td>31.4 $\pm$ 6.3</td>
</tr>
</tbody>
</table>

$^a$ Uncertainty based on an estimated 60 m error in area-wide vertical accuracy.

$^b$ Calculation based on a mean summit lowering rate of 8.1 m Myr$^{-1}$ (Table 2).
Table 4
Magnitudes of total erosion for scenarios in which moraine volume accounts for 20%, 50%, or 80% of all material eroded from the catchment by glaciers.

<table>
<thead>
<tr>
<th>Basin</th>
<th>Moraine area (km²)</th>
<th>Moraine volume (km³)</th>
<th>Rock-equivalent volume (km³)</th>
<th>Total erosion (km³)</th>
<th>Total basin-wide surface lowering (m)²</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Gilgar Uul</strong></td>
<td></td>
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</tr>
<tr>
<td>Vₘ = 20%</td>
<td></td>
<td></td>
<td></td>
<td>0.13</td>
<td>1.1</td>
</tr>
<tr>
<td>Vₘ = 50%</td>
<td>116</td>
<td>0.033</td>
<td>0.026</td>
<td>0.05</td>
<td>0.5</td>
</tr>
<tr>
<td>Vₘ = 80%</td>
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<td></td>
<td></td>
<td>0.03</td>
<td>0.3</td>
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<tr>
<td><strong>Khaak Nuur</strong></td>
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<tr>
<td>Vₘ = 20%</td>
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<td></td>
<td></td>
<td>0.75</td>
<td>3.7</td>
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<tr>
<td>Vₘ = 50%</td>
<td>204</td>
<td>0.187</td>
<td>0.150</td>
<td>0.30</td>
<td>1.5</td>
</tr>
<tr>
<td>Vₘ = 80%</td>
<td></td>
<td></td>
<td></td>
<td>0.19</td>
<td>0.9</td>
</tr>
<tr>
<td><strong>Chuluut Gol</strong></td>
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<tr>
<td>Vₘ = 20%</td>
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<td>0.28</td>
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<tr>
<td>Vₘ = 50%</td>
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<td>0.069</td>
<td>0.055</td>
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<tr>
<td>Vₘ = 80%</td>
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²This number is the result of dividing the total volume of erosion by the basin area.
Table 5
Cosmogenic $^{10}$Be results for moraine boulder exposure ages.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Latitude ($^\circ$N)</th>
<th>Longitude ($^\circ$E)</th>
<th>Elevation (m)</th>
<th>Thickness (cm)</th>
<th>Production rate</th>
<th>Shielding factor</th>
<th>Erosion rate (cm yr$^{-1}$)</th>
<th>Quartz (g)</th>
<th>Be carrier (g)</th>
<th>$^{10}$Be/$^{9}$Be</th>
<th>$^{10}$Be conc. ($^{10}$Be Be atom at g$^{-1}$ SiO$_2$)</th>
<th>Age (ka)</th>
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<tbody>
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<td>Gilgar Uul (GU)</td>
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<tr>
<td>MN12-31</td>
<td>48.2094</td>
<td>98.79527</td>
<td>2531</td>
<td>3</td>
<td>33.27</td>
<td>0.409</td>
<td>0.00001</td>
<td>20.1408</td>
<td>0.8079</td>
<td>6.984 ± 0.123</td>
<td>5.033 ± 0.135</td>
<td>15.2 ± 1.4</td>
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<td>MN12-33</td>
<td>48.2103</td>
<td>98.75703</td>
<td>2519</td>
<td>3</td>
<td>33.01</td>
<td>0.408</td>
<td>0.00001</td>
<td>20.0533</td>
<td>0.8037</td>
<td>7.033 ± 0.163</td>
<td>5.132 ± 0.157</td>
<td>15.6 ± 1.4</td>
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<td>MN0711-15A</td>
<td>47.4621</td>
<td>98.56841</td>
<td>2676</td>
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<td>36.11</td>
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<td>0.00001</td>
<td>20.3840</td>
<td>0.8111</td>
<td>17.09 ± 0.316</td>
<td>12.52 ± 0.341</td>
<td>34.2 ± 3.1</td>
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<td>MN0711-15B</td>
<td>47.4626</td>
<td>98.56909</td>
<td>2677</td>
<td>3</td>
<td>36.14</td>
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<td>20.2411</td>
<td>0.8115</td>
<td>22.11 ± 0.406</td>
<td>16.05 ± 0.436</td>
<td>43.7 ± 3.9</td>
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<tr>
<td>MN0711-15C</td>
<td>47.4631</td>
<td>98.56913</td>
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<td>36.12</td>
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<td>0.00001</td>
<td>20.0644</td>
<td>0.8060</td>
<td>6.912 ± 0.128</td>
<td>5.010 ± 0.137</td>
<td>13.9 ± 1.2</td>
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<tr>
<td>Mean age =</td>
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<td>19.9984</td>
<td>0.8111</td>
<td>8.997 ± 0.248</td>
<td>6.589 ± 0.224</td>
<td>23.8 ± 2.2</td>
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<tr>
<td>MN08-11-32B</td>
<td>47.4115</td>
<td>100.2476</td>
<td>2274</td>
<td>3</td>
<td>27.49</td>
<td>0.378</td>
<td>0.00001</td>
<td>20.0789</td>
<td>0.8017</td>
<td>7.601 ± 0.241</td>
<td>5.475 ± 0.205</td>
<td>19.8 ± 1.8</td>
</tr>
<tr>
<td>MN08-11-32C</td>
<td>47.4156</td>
<td>100.2634</td>
<td>2219</td>
<td>3</td>
<td>26.45</td>
<td>0.372</td>
<td>0.00001</td>
<td>20.1028</td>
<td>0.8132</td>
<td>7.863 ± 0.210</td>
<td>5.739 ± 0.191</td>
<td>21.6 ± 2.0</td>
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<td>47.4483</td>
<td>100.2321</td>
<td>2088</td>
<td>3</td>
<td>24.11</td>
<td>0.357</td>
<td>0.00001</td>
<td>20.1026</td>
<td>0.8156</td>
<td>7.242 ± 0.128</td>
<td>5.299 ± 0.141</td>
<td>21.8 ± 1.9</td>
</tr>
<tr>
<td>Mean age =</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

2. The tops of all samples were exposed at the surface, and samples ranged from 2-4 cm in thickness with a mean thickness of 3 cm.

b. Constant (time-invariant) local production rate based on Lal (1991) and Stone (2000). A sea level, high-latitude value of 4.8 at $^{10}$Be/g quartz was used.

c. A density of 2.7 g cm$^{-3}$ was used based on the granitic composition of the samples.

d. Uncertainties are reported at the 1σ confidence level.

f. Reported values include boron correction factors of $4 \times 10^{-5}$ ± $1 \times 10^{-5}$ (Gilgar Uul and Chuluut Gol) and $1 \times 10^{-4}$ ± $0.2 \times 10^{-4}$ (Khaak Nuur).

e. Isotope ratios were normalized to $^{10}$Be standards prepared by Nishiizumi et al. (2007) with a value of $2.85 \times 10^{12}$ and using a $^{10}$Be half-life of $1.36 \times 10^6$ years.

f. Mean $^{10}$Be/$^{9}$Be blank values of $6.41 \times 10^{-15}$ ± $1.29 \times 10^{-15}$ (GU), $3.60 \times 10^{-15}$ ± $3.93 \times 10^{-16}$ (KN), and $4.11 \times 10^{-15}$ ± $3.42 \times 10^{-16}$ (CG) were used to correct for background.

h. Propagated uncertainties include error in the blank, carrier mass (1 percent) and counting statistics.

j. Beryllium-10 model ages were calculated with the CRONUS Earth online calculator (Balco et al. 2008) version 2.2 (http://hess.ess.washington.edu/).
### Table 6
Glacial erosion rates for various scenarios of erosion duration.

<table>
<thead>
<tr>
<th>Scenario (erosion duration)</th>
<th>Erosion rate [m Myr⁻¹]</th>
<th>Gilgarg Uul</th>
<th>Khaak Nuur</th>
<th>Chuluut Gol</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Kyr</td>
<td>290 - 1100</td>
<td>920 - 3700</td>
<td>820 - 3300</td>
<td></td>
</tr>
<tr>
<td>2 Kyr</td>
<td>140 - 570</td>
<td>460 - 1800</td>
<td>410 - 1600</td>
<td></td>
</tr>
<tr>
<td>5 Kyr</td>
<td>57 - 230</td>
<td>180 - 730</td>
<td>160 - 660</td>
<td></td>
</tr>
<tr>
<td>10 Kyr</td>
<td>29 - 110</td>
<td>92 - 370</td>
<td>83 - 330</td>
<td></td>
</tr>
</tbody>
</table>
Figure 1. Tectonic setting of the Mongolian Plateau and Hangay Dome in the context of the India-Asia collision. AT = Altyn Tagh; BR = Baikal Rift; DL = Mongolian Depression of Lakes; GA = Gobi Altai; HD = Hangay Dome; HR = Hövsgöl Rift; KL = Kunlun Shan; MA = Mongolian Altai; QS = Qilian Shan; TB = Tunka Basin; TS = Tien Shan. The area of Figure 2 is shown within the black box that includes the entire country of Mongolia.
Figure 2.  (Top) Shaded relief digital elevation model of Mongolia with the Hangay Mountains labeled in the central part of the country and the study area outlined with a black rectangle.  (Bottom) Shaded relief digital elevation model of the study area with major rivers depicted using black lines; dashed-line segments signify areas of ephemeral flow.  Note the deeply incised valleys with up to 1000 meters of relief.  ED = Egiin Davaa (Egiin Pass); CG = Chuluut Gol (Chuluut River); OG = Orkhon Gol; TG = Tuin Gol.
Figure 3. Landsat Thematic Mapper (TM) images of the study area created using various band combinations. (A) Band 5-4-3 (5 = red, 4 = green, 3 = blue) “natural color” image. Basalt summits in the center of the image appear black to dark purple, whereas underlying granitic basement rock is light purple to pink in color. (B) Band 1-5-6 image. The basalts are deep purple and the granitic basement rock is yellow. Due to the sharp contrast, this was the image used to map the basalt/basement contact. (Imagery acquired by the Landsat 5 satellite on July 27, 2006). The upper portion of the Chuluut Gol and the location of Egiin Davaa are depicted on both images for reference to their locations on Figure 2 and Egiin Davaa (ED) is identified.
Figure 4. Field photos of the modern Hangay Mountain landscape. In (A), a paleo-valley with ~600 m can be traced by the contact (white dashed line) between Cenozoic basalt flows and the underlying granitic basement. (B) depicts the contact between granitic basement (light-colored rocks in foreground) and basalt (darker-colored rocks in the distance) as seen while walking along a summit in the study area. In (C), stacked Cenozoic basalt flows comprise most of the image, but two instances of the sharp contact with the underlying basement rock are visible, one in the background and another in the lower right side of the image. (D) a basalt-filled paleo valley is observed in cross section.
Figure 5. Landsat TM “natural color” image showing 12 of the $^{40}\text{Ar}/^{39}\text{Ar}$ basalt ages referenced from Ancuta et al. (in review) for use in this study. Note that there are 24 total $^{40}\text{Ar}/^{39}\text{Ar}$ basalt ages within the study area, but several are from the same location and have nearly identical ages. In these cases only a single sample location and a corresponding mean age are shown. Black dots mark the location of basalt flows in contact with the underlying Paleozoic-to-Mesozoic basement rocks, whereas white dots mark the location of ridge-top basalt flows. Numbers are the mean age in millions of years. See Ancuta et al. (in review) for full details of each sample, including precise location and age uncertainty. The extent of the study area for paleo-topographic and erosion rate analyses is outlined by the black dashed line. (Base imagery acquired by the Landsat 5 satellite on July 27, 2006).
Figure 6. Oblique Landsat TM band 1-5-6 image of the study site with the mapped basalt/basement contact shown in black. The white line indicates the trace of the Egiin Davaa normal fault (e.g. Walker et al., 2015), beyond which the contact was not mapped so that any recent fault slip would not bias results of the paleo-topography and erosion rate calculations.
Figure 7. Visual representation of the digital elevation model creation process for modern and paleo topography in the Hangay Mountain study area. Shown in (A) is an oblique view of the July 27, 2006 Landsat TM “natural color” image; below the Landsat image is a digital elevation model of the area enclosed by the black oval. In (B), the summit-capping basalts are shaded in gray. These basalts were “removed” from the landscape in order to create the digital elevation model of paleo-topography.
Figure 8. Topographic swath profile locations and example profiles for both modern and paleo topography within the study area. The locations of profiles 1-11 are shown as black dashed lines on the DEMs, with example profiles 2 and 9 indicated in bold. The profile graphs below these DEMs show the minimum, mean, and maximum elevations along profiles 2 and 9, which were constructed using a circular window with a 1 km radius.
Figure 9. Landsat TM image from July 27, 2006 showing summit plateau TCN sampling locations (with samples IDs in brackets) and corresponding denudation rates rounded to the nearest whole number. Inset figure is a DEM of Mongolia with a black box outlining area of larger figure. Sample IDs are as follows: 04 = MN0711-04; 06 = MN0711-06; 08 = MN0711-08; 12 = MN0711-12; 13 = MN0711-13; 29 = MN0711-29; 38 = MN0811-38. The study area for the paleotopographic and erosion rate analyses is enclosed in a black rectangle. See Tables 1 and 2 for full sample descriptions and detailed results of TCN denudation rate calculations.
Figure 10. Diagram depicting the process of “filling in” the topography in order to calculate the volume of rock removed from the study area via denudation.
Figure 11. The process of calculating eroded rock volume from the study area. In step 1 (top), an area for analysis is delineated. In step 2, an “uneroded” landscape is created by interpolating across mountain summits within the study area and adding different volumes of rock to compensate for the range of estimated summit erosion since emplacement of the ridge-top basalts. Step 3 is the subtraction of the modern topography from the uneroded landscape (DEM) of modern landscape shown in panel 3) to produce a DEM of the rock volume that has been removed. Note that, as expected, the largest areas of erosion correspond to the deepest valleys in the modern landscape. Note: Scale bar and north arrow at top are approximately correct for all images.
Figure 12. Landsat TM image from July 27, 2006 with moraine boulder TCN sampling locations (with samples IDs in brackets) and corresponding exposure ages rounded to the nearest thousand years. Inset figure is a DEM of Mongolia with a black box outlining area of larger figure. GU = Gilgar Uul; KN = Khaak Nuur; CG = Chuluut Gol. The study area for the paleotopographic and erosion rate analyses is enclosed in a black rectangle. See Tables 5 and 6 for full sample descriptions and detailed results of TCN exposure age calculations.
Figure 13. Google Earth images of each study area used in the glacial erosion rate analysis. Interpreted LGM moraines are partly outlined by black dashed lines. CG = Chuluut Gol; GU = Gilgar Uul; KN = Khaak Nuur. Coordinates of the centers of each image (indicated by white crosshairs) are as follows: CG = 47° 23’ 21” N, 100° 22’ 32” E; GU = 48° 08’ 06” N, 98° 48’ 23” E; KN = 47° 29’ 16” N, 98° 32’ 07” E.
Figure 14. Visual illustration of the moraine volume analysis, using the Chuluut Gol study area as an example. (A) First, moraine and drainage basin area are defined. (B, #1) A DEM is created for the moraine, and (B, #2) a basal surface is interpolated for the moraine based on low points in the topography surrounding the moraine as well as the channel elevation of the stream that bisects the moraine. (B, #3) The basal surface is subtracted from the moraine DEM to obtain moraine thickness, which is then used to calculate volume.
Figure 15. Photograph showing a wide, alluvial river valley (Chuluut Gol) with several ephemeral channels. This type of transport-limited alluvial system, which exhibits significant sediment aggradation, is typical of the central Hangay Mountain range. Dark specks in center of valley are cattle for scale. Photo location: 47.3353° N, 100.1056° E, oriented 215° (SW).
CHAPTER 2

A ~4000 year record of hydrologic variability in the Olympic Mountains, Washington

2.1 Abstract

Sedimentological and geochemical analyses of gravity and piston cores retrieved from Lake Quinault, a 15 km² and 70 m deep moraine-dammed lake located on the western front of the Olympic Mountains in Washington, USA, reveal a record of deposition over the past ~4000 years that is detrital-dominated and driven by large flood events. Individual historic flood layers preserved in sediment cores contain a coarse-grained basal deposit that is enriched in terrigenous organic matter. These same flood layers display peaks in the ratio of incoherent to coherently scattered X-ray radiation, or inc/coh, from µXRF core scans, a parameter which traces the concentration of coarse-grained, organic-rich sediment throughout each core. Given this relationship, we utilize the inc/coh ratio as a proxy for individual flood event layers that allows for an assessment of the periodicity of overall hydrologic variability and the distribution of extreme events (paleofloods) throughout the late Holocene. Despite the lack of a long-term trend, notable peaks in event number occurred during ~2350 to ~2450 cal BP and ~1910 to ~2010 CE. The period ~2350-2450 cal BP is unremarkable within the context of existing Pacific Northwest paleoclimate records, but the spike in extreme events over the last century suggests that an increase in the frequency of large storms tracking over the Olympic Peninsula is connected to recent climate change. This connection, coupled with high spectral power of the inc/coh time series at multi-decadal and multi-centennial periodicities, leads us to hypothesize that the record of hydrologic variability as preserved in in Lake Quinault sediments reflects trends in climate conditions favorable for the formation of north Pacific atmospheric rivers. Since atmospheric rivers are modulated by fluctuations in large scale atmospheric and oceanic teleconnections such as the Pacific Decadal Oscillation, and are predicted to increase in frequency and severity during the course of the 21st century, understanding past hydrologic variability has important implications for the landscape and ecosystem response of Olympic Mountain catchments to future climate warming.
2.2 Introduction

The Olympic Mountains, which constitute a large portion of the Olympic Peninsula in western Washington, are unique in the United States for the combination of rugged, glacially-sculpted topography and large swaths of temperate rain forest (Fig. 1). Rainforest conditions exist predominantly on the western, windward side of the range owing to orographic forcing as moisture-laden, landfalling Pacific storms encounter the steep, mountainous terrain (e.g. Minder et al., 2008; Neiman et al., 2011; Gavin & Brubaker, 2015). The bulk of precipitation, which can exceed 5 m annually throughout the range’s highlands, falls during the winter season as the Aleutian low pressure system intensifies and drifts south from the Gulf of Alaska, subsequently steering moisture and winds toward the Pacific Northwest (PRISM Climate Group, 2012; Gavin and Brubaker, 2015). These typically cool, rainy winters transition to warm and dry summers once the Aleutian low weakens and the North Pacific high pressure system strengthens and brings northwesterly air flow, low humidity, and sunny skies to the region (Ware & Thomson, 2000; Patterson et al., 2013).

Although the seasonal climate of the Pacific Northwest is largely governed by the relative strength and position of the Aleutian low and North Pacific high pressure systems, at sub- and inter-decadal timescales the El Niño-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) ocean-atmosphere teleconnections drive much of the long-term variability in precipitation (e.g. Mantua et al., 1997; McCabe & Dettinger, 1999; Mantua & Hare, 2002; Miller & Goodrich, 2007; Ersek et al., 2012). With respect to western Washington, winters are generally cooler and average winter precipitation generally greater when ENSO and PDO indices are both negative. These effects extend to seasonal snowpack and subsequent melt-driven discharge on the Olympic Peninsula, which also increase when ENSO and the PDO are in a negative state (Redmond and Koch, 1991; Gavin and Brubaker, 2015). Adding further complexity to trends in precipitation, snowpack, and river discharge in the Olympic Peninsula is the role of atmospheric rivers (ARs). ARs are defined as long, narrow atmospheric plumes characterized by high water vapor content and strong, low-level winds responsible for transporting large amounts of water from lower to higher latitudes (Neiman et al., 2011; Gimeno et al., 2014). The AR that connects tropical moisture with the Pacific Northwest is known colloquially as the ‘Pineapple Express’ (Lackmann & Gyakum, 1999; Dettinger, 2004; Gimeno et al., 2014), and this AR has been responsible for the majority of floods in western Washington since 1980 as a result of abundant precipitation.
coupled with seasonally-high temperatures and melting levels that favor rain rather than snow at higher elevations (Neiman et al., 2011). However, despite the connection between ARs, precipitation, and river discharge, there is less certainty concerning the relationship between longer-term climate variability and conditions favorable to ARs. Limited prior research suggests Pineapple Express storms are more pronounced during positive PDO and neutral or near-neutral ENSO states (Dettinger, 2004), which is an opposite relationship with respect to the aforementioned general trend of greater winter precipitation during times of negative PDO.

Thus far, neither the precise nature nor the long-term variations of these complex interacting processes are completely understood, and ocean-atmosphere climate dynamics and their influence on landscape evolution remain the subject of ongoing research with respect to both the Pacific Northwest and other regions of the globe (e.g., Redmond & Koch, 1991; Peng et al., 2005; Conroy et al., 2010; Nelson et al., 2011; Kirby et al., 2012; Ersek et al., 2012; Wilhelm et al., 2013; Rodrigo-Gámiz et al., 2014). This underscores the importance of obtaining paleoenvironmental data in an effort to build a more robust knowledge of past patterns so as to better predict future response to a warming climate. If past variability in parameters such as river discharge and sediment transport are better understood and placed in context of known ocean-atmosphere dynamics influencing Pacific Northwest climate, then future change can be addressed and proper preparations can be made. These types of assessments are especially important given the concern surrounding projected global climate change and the potential for more extreme and unpredictable weather in the region (Mote & Salathé, Jr., 2010; Weller et al., 2012). Preparing for future change is also crucial for regional stakeholders such as the Quinault Indian Nation who have jurisdiction over Lake Quinault. In a joint effort with the U.S. Fish & Wildlife Service, the Quinault Indian Nation maintains the Quinault National Fish Hatchery, which was established in 1968 to buffer declines in local salmon and steelhead fisheries (U.S. Fish & Wildlife Service, 2009). In the Pacific Northwest, proposed links between salmon productivity and climate variables such as sea surface temperature, phases of the PDO, and flood frequency (e.g., Mantua et al., 1997; Stockner et al., 2003; Finney et al., 2010) suggest that studying the sedimentary record of Lake Quinault may also have implications for future trends in lake and catchment ecology.
In general, previous research addressing Holocene environmental variability within the Olympic Mountains has focused primarily on forest paleoecology, fire history, and vegetation change (e.g. McLachlan et al., 1995; Brubaker and McLachlan, 1996; Gavin and Brubaker, 1999; Heusser et al., 1999; Gavin et al., 2001; Gavin et al., 2013; Gavin and Brubaker, 2015), and thus there is a need for additional proxy records detailing Holocene variability in catchment-scale surface processes and hydrology. A commonly utilized methodology for reconstructing the history of Earth surface processes is through the careful study and analysis of lacustrine sedimentary records. Depending on local geographic, geologic, and hydrologic variables, such as lake geometry, local climate, and sedimentation rate, these records can span tens of thousands of years and provide researchers with valuable insight into the nature and drivers of deposition through time for various regions across the globe (e.g. Koinig et al., 2003; Guyard et al., 2007; Chapron et al., 2007; Debret et al., 2010; Kirby et al., 2010; Nelson et al., 2011; Jouve et al., 2013; Finkenbinder et al., 2014, Schillereff et al., 2014). Furthermore, in the case of stream-fed lakes, if there is a valid relationship between discharge and the deposition of detritus, then lake sediments can serve as an archive of past hydrologic variability, which can provide valuable data on general patterns of flooding and/or the frequency of singular extreme flood events (e.g. Bøe et al., 2006; Moreno et al., 2008; Parris et al., 2009; Page et al., 2010; Jones et al., 2012; Gilli et al., 2013; Schillereff et al., 2014; Schlolaut et al., 2014).

Our research focuses on the sedimentary record of Lake Quinault, a ~15 km2 moraine-dammed lake that acts as a sediment trap and defines base level for the upper Quinault River. The upper Quinault River catchment (UQRC), which is the source of the vast majority of sediment and water to the lake, drains a mountainous area almost entirely under the designation of Olympic National Park or Olympic National Forest, and has experienced such minimal disturbance as a proportion of drainage basin area that it should be considered broadly representative of a montane-to-alpine drainage basin free of human interference to forest and hydrologic systems (Fig. 2). Averaged across the UQRC, yearly precipitation totals nearly 5 m, and the typical annual hydrograph is marked by large winter floods, sustained above-average flow during snowmelt in June and July, and minima in August or September (Gavin and Brubaker, 2015; U.S. Geological Survey, 2016). Despite seasonal snowpack in the upper elevations, times of exceptionally high discharge are much more common in winter due to the combination of intense rainfall and soils at or near saturation (Bountry et al., 2005). These high discharge events drive detrital-dominated sedimentation
into the deeper, distal areas of the lake that otherwise receive only sparse detritus during low flow conditions and have little to no autochthonous sediment contribution (Stockner et al., 2003; L. Workman, personal communication, 2013). Overall, the physical characteristics of Lake Quinault and the UQRC make it an ideal setting to assess the record and patterns of late Holocene hydrologic variability and changes in the frequency of extreme discharge events.

In this paper we present a ~4000 year, high-resolution (<3 yr) continuous record of sedimentation in Lake Quinault and use core-scanning X-ray fluorescence (µXRF) elemental data as a paleohydrology and paleoflood proxy. We also implement Lomb-Scargle periodograms and continuous wavelet transforms to characterize the dominant periodicities and temporal variability of the proxy record. Finally, we compare 100+ years of Quinault River streamflow data to modern atmospheric and climate indices in an attempt to decipher potential controls on recent river discharge and how this may relate to past variability. The significance of the findings is discussed relative to Pacific Northwest climate and catchment-scale dynamics of the Pacific Ocean-facing Olympic Mountains.

2.3 Background and Geologic Setting

2.3.1 Regional Geology

The Olympic Mountains represent the sub-aerial portion of the accretionary wedge formed as a consequence of oblique subduction of the Juan de Fuca plate beneath the North American plate across the Cascadia margin. Around 15 Ma, a decrease in accommodation space led to uplift and exposure of Olympic Mountain segment of the subduction wedge above sea level (Brandon et al., 1998). At the surface, the Olympic Mountains consist primarily of the Olympic subduction complex (OSC), which is characterized by highly deformed, stratigraphically discontinuous, and partially metamorphosed sedimentary rocks (Fig. 1; Tabor & Cady, 1978; Brandon et al., 1998). Juxtaposed against the rocks of the OSC is a horseshoe-shaped surface exposure of older oceanic basalts situated on the footwall of the eastward-dipping Hurricane Ridge thrust fault (Tabor & Cady, 1978; Brandon & Calderwood, 1990; Babcock et al., 1994). These basalts, which make up the Eocene
Crescent Formation (ECF), form a border along the northern, eastern, and southern sides of the range.

As the Juan de Fuca plate continues to converge and subduct beneath the North American plate at 36 mm yr\(^{-1}\), rates of uplift increase landward from \(\sim0.3\) km Ma\(^{-1}\) near the Pacific Coast of Washington to \(\sim1\) km Ma\(^{-1}\) near the summit of Mount Olympus, which, at an elevation of 2430 m (7980 ft), is the present-day high point of the range (Brandon et al., 1998; DeMets & Dixon, 1999). Previous research has shown that net uplift, or rock influx into the range, is balanced by the processes of denudation, or eroded sediment flux out of the range, and thus the Olympic Mountains are believed to maintain a flux steady state in which there is no net addition or loss of mass over geologic timescales (Brandon et al., 1998; Pazzaglia & Brandon, 2001; Willett and Brandon, 2002; Stewart and Brandon, 2004).

### 2.3.2 Study Site

The upper Quinault River catchment encompasses a \(\sim600\) km\(^2\) area along the southwestern flank of the Olympic Mountains and is underlain primarily (\(\sim90\%\)) by the Olympic subduction complex with a small area of ECF basalts present near the upper Quinault River’s junction with Lake Quinault (Figs. 2, 3, & 4; Tabor & Cady, 1978). The catchment has a history of multiple glaciations, and the most recent Oxygen Isotope Stage 2 (OIS 2) glaciation likely reached its maximum extent around 30 ka. During the retreat of the glacier that occupied the Quinault valley during MIS 2, a period of re-advance or ice stagnation built the terminal moraine that now impounds Lake Quinault (Fig. 4; Moore, 1965; Thackray, 1996; Thackray, 2001; Bountry et al., 2005). Staley (2015) reports a luminescence age of 18.6 ± 5.1 ka from glacial outwash located 4 km downstream from and correlated to the Lake Quinault moraine, and this serves as the best estimate regarding the timing of lake formation. As warmer conditions developed and persisted throughout the Holocene, ice continued to retreat up the Quinault valley, and only small (<0.25 km\(^2\)) cirque glaciers, such as the glacier on Mt. Anderson that marks the headwaters of the East Fork of the Quinault River, are present in the catchment today (Quinault Indian Nation, 1999).

As the Quinault River has incised into the terminal moraine, the surface water elevation, areal extent, and volume of water within Lake Quinault have likely decreased (Bountry et al., 2005). Today, Lake Quinault has a surface water elevation of 57 m, covers \(\sim15\) km\(^2\), and
has average and maximum depths of 40 m and 70 m, respectively. Sedimentation in the lake is concentrated at the delta formed by the upper Quinault River, of which there are two main tributaries, the North and East Forks, which join approximately 30 river kilometers upstream of the lake (Bountry et al., 2005).

Anthropogenic disturbance within the catchment is mostly limited to the late 1800s and early 1900s when the area was open to logging and mature forests and woody debris were cleared from within the floodplain and main channel of the alluvial upper Quinault River. The fluvial response to this period of activity was an acceleration of lateral channel migration that led to subsequent increases in erosion and sediment supply. This transition occurred relatively abruptly and was associated with minimal changes in channel geometry and fluvial processes during the first half of the 20th century (Bountry et al., 2005). Today, the upper Quinault River just upstream of Lake Quinault has been described as having an irregular meandering morphology with shorter reaches of braided and anastomosing channel forms (O’Connor et al., 2003). Field observations of the UQRC indicate that the dominant mode of sediment transport down hillslopes is via landsliding. Lending support for these observations is a GIS assessment of catchment topography indicating that approximately half the hillslopes in the UQRC are at angles greater than 30°, as well as a 59-year landslide inventory that catalogued 668 landslides throughout the catchment using aerial photos spanning the years between 1939 and 1998 (Quinault Indian Nation, 1999). In addition to transporting material downslope, landslides and other mass wasting processes supply the majority of sediment to the upper Quinault River; bank erosion contributes only a small fraction of the sediment ultimately delivered to Lake Quinault by its river (Bountry et al., 2005).

Overall, owing to the combination of steep slopes, high rainfall, a singular dominant fluvial input, high trapping efficiency, simple bathymetry, and event-driven deposition of fine-grained sediments, Lake Quinault provides an excellent opportunity to unravel a sedimentary record of late Holocene hydrologic variability in a nearly pristine catchment.
2.4 Methods

2.4.1 Coring and core-logging

In May 2013, piston and gravity cores were recovered from eight sites spanning the long axis of Lake Quinault at water depths between 35 and 70 m (Fig. 5; Table 1). Cores were obtained using a Kullenberg coring system (Kelts et al., 1986) deployed from a specially constructed platform (R/V KRKII) with a tower and moon pool. The recovered piston cores range from 3.7 to 6.9 m in length and were collected in 7 cm outer diameter polycarbonate tubes. The accompanying gravity cores (denoted with a -G suffix) were recovered simultaneously at each coring site to provide a relatively undisturbed sample of the sediment-water interface, and range in length from about 0.3 to 0.6 m. The piston cores were cut into ≤1.5 m segments at the Lake Quinault field site, and all core segments were capped, sealed, and transported to the LacCore Facility at the University of Minnesota in Minneapolis where they were placed in cold storage before further processing and sampling that occurred approximately ten days after recovery.

During processing, cores were split and sediment surfaces were cleaned with glass slides in preparation for imaging and description. Core images were generated on a Geotek Geoscan-IV linescan camera mounted on a dedicated MSCL-CIS track, and core halves were scanned on a Geotek XYZ Multisensor Core Logger for high-resolution point sensor magnetic susceptibility (MS) at 5 mm resolution. Core lithologies were then described and sampled, which included collection of plant macrofossil material to be used for AMS-\(^{14}\)C analyses in the construction of sediment age versus depth models. Photos and initial descriptions were later used to visually measure event layer thicknesses and correlate layers between cores.

2.4.2 Age modeling

Sediment chronology was established using twelve \(^{14}\)C AMS dates of assorted plant debris sampled from cores 2 and 3. Each organic sample was cleaned with 1N HCl and NaOH (ABA method; Olsson, 1986) to remove any possible contamination from carbonate coatings or humic acids during burial in soils or at the lake bottom. Pre-treated samples were sent to Direct AMS in Seattle, Washington for analysis (Table 2). Resultant \(^{14}\)C AMS ages were
calibrated and converted to calendar years using OxCal 4.2 software (Ramsey, 2009) and the IntCal13 calibration curve (Reimer et al., 2013) with the northern Hemisphere Zone 1 extension (Hua et al., 2013). To provide age constraints for core 1, calibrated ages from core 2 were applied to core 1 by visually correlating layers and depths between cores, which allowed us to assign a specific depth in core 1 to each age obtained from core 2. Correlation between cores was performed using Corelyzer software made available through the CoreWall project (www.corewall.org). Age versus depth models for cores 1 and 3 were then built with a Bayesian approach using Bacon software (Blaauw and Christen, 2011).

2.4.3 µXRF geochemistry, particle size, and isotopic analysis

Relative down-core elemental abundances for cores from two locations, sites 1 and 3, were generated using an ITRAX µXRF scanner at the Large Lakes Observatory in Duluth, Minnesota. Both the piston and gravity cores from each site were scanned by a molybdenum X-ray tube using a 2 mm sampling step and a 30 second dwell time. During data reduction, artifacts introduced by core section breaks and sediment gaps were deleted from the dataset. Based on methodology outlined by Ohlendorf et al. (2015), elemental abundances were normalized to both inelastic Compton scattering (inc; incoherent scattering of radiation at a lower energy than the incoming X-rays) and elastic Rayleigh scattering (coh; coherent scattering of radiation at the same energy as the incoming X-rays) to correct for potential biases introduced by sediment matrix-related effects such as density, porosity, and water content, as well as changes in X-ray tube power during the measurement period. Results of inc versus coh normalization were nearly perfectly correlated (R > 0.99), indicating little to no difference in the data output with respect to normalization technique. We ultimately chose to normalize using the coh parameter and thus all elemental data presented in this paper reflects the division of element raw intensities by the corresponding intensities of coherent radiation. Although 34 elements were included in the µXRF analysis (Al, Si, P, S, Cl, Ar, K, Ca, Ti, V, Cr, Mn, Fe, Ni, Cu, Zn, Ga, Ge, Se, Rb, Sr, Y, Zr, Sn, Ba, Ce, Pr, Ta, W, Re, Pt, Au, Pb, Bi), we considered only common elements used in previous µXRF studies of lacustrine sediments that were measured in high abundance (>500 counts/measurement) by the ITRAX scanner and thus also exhibited the highest signal to noise ratios (Ca, Fe, K, Mn, Rb, Si, Sr, Ti, and Zr). From the µXRF results, we also calculated and plotted the inc/coh ratio since it has been shown to record changes in sediment density and/or organic content (Guyard et al., 2007; Fortin et al., 2013). This
relationship is explained by the theory that the magnitude of incoherent scattering between electrons is higher for elements with a low atomic mass, and therefore the $inc/coh$ ratio is linked to the mean atomic number and could be partly related to carbon content (Crouda et al., 2006; Guyard et al., 2007). Furthermore, since incoherent scattering results from collisions between photons with particles larger than the wavelength of the incoming X-ray radiation (e.g., Harris & Bertolucci, 1978), the $inc/coh$ ratio may also theoretically correlate positively with particle size if other properties such as chemical composition, water content, and density exhibit minimal variation throughout a given core interval.

In order to test for a correlation between core particle size and core-scanning µXRF data, the gravity core from location 3 was sub-sampled at 5 mm intervals along its entire length for particle size analysis. Core 3 was chosen since the sedimentation rate is higher than core 1, which thus made it more straightforward to compare particle size and µXRF data to layers visible in the core image. The particle size distribution of each bulk untreated sample was characterized using a Beckman Coulter LS13-320 Laser Diffraction Particle Size Analyzer equipped with a Universal Liquid Module that measures particle size between 0.04 and 2000 microns.

To assess the organic carbon content and composition of Lake Quinault sediment, five bulk samples from the uppermost 30 cm of the location 5 gravity core (5-G) were prepared for elemental and isotopic analysis. To remove inorganic carbon, samples were treated with aqueous 4N HCl for 48 hours, then vacuum dried at room temperature without rinsing. Once dried, the samples were transferred to tin boats for quantification of nitrogen and carbon content using a Thermo-Electron EA 1112 elemental analyzer. Measurements were corrected for the formation of salts during the acidification process, and the relative precision for both C and N measurements was 2%. Following combustion in the elemental analyzer, CO$_2$ was transported via helium to a Conflo III interfaced Thermo Delta V Isotope Ratio Mass Spectrometer (IRMS) for stable carbon isotopic measurements. $^{13}$C/$^{12}$C ratios are expressed as per mil relative to the VPDB standard using conventional δ$^{13}$C notation. Samples were run against an acetonilide working standard calibrated against several in-lab and NBS/NIST reference materials, and two sets of duplicate samples were run to ensure data precision. Absolute measurement precision was 0.2‰.
2.4.4 Time series analysis

Although μXRF measurements were made at a constant sampling interval, a gradual increase in sedimentation rate at the core sites results in data series that are unevenly spaced in time. Accordingly, spectral analysis of μXRF and MS data was completed using the SLOMBS program developed by Pardo-Igúzquiza and Rodríguez-Tovar (2012) that combines the Lomb-Scargle periodogram with a Monte Carlo evaluation of the statistical significance of the estimated power spectrum. The Lomb-Scargle approach is especially useful for unevenly spaced data since it directly utilizes the irregular time series and avoids interpolation that would otherwise decrease data resolution (Pardo-Igúzquiza and Rodríguez-Tovar, 2006, Pardo-Igúzquiza and Rodríguez-Tovar, 2011, Pardo-Igúzquiza and Rodríguez-Tovar, 2012 and Rodrigo-Gámiz et al., 2014). Using the SLOMBS program, power spectrums for cores 1 and 3 were estimated for 500 frequencies over the frequency range $[0, 0.2]$. Prior to analysis, linear trends were removed from the datasets to remove the effect of increased sedimentation rate on the signal. For each time series, $10^N$ (N = number of data points) permutations were performed, and linear smoothing with 7 terms was used. Following the approach of Rodrigo-Gámiz et al. (2014) and Jiménez-Espejo et al. (2014), we limited our interpretation of results to cycles well above the maximum frequency resolution of the approach (as defined by the Nyquist frequency, or 0.5 multiplied by the lowest sampling frequency of the core record) and also avoided interpreting very low frequencies that constitute a significant portion of the total time represented by the record (i.e. periodicities > 1000 yr in a core record spanning ~4000 yrs). Of the frequencies considered, we focused on those with the highest spectral power that achieved a confidence level of 95% or greater.

Following SLOMBS analysis, μXRF data were interpolated (resampled) at time steps of 2 and 0.25 years for cores 1 and 3, respectively, so that continuous wavelet transforms (CWT) could be used to evaluate how the frequency components of a signal vary with time. This interpolation was necessary because CWT requires the input of an evenly-spaced time series. Once resampled, analyses were carried out using the wavelet coherence toolbox developed by Grinsted et al. (2004) for use in MATLAB.
2.4.5 Extreme event (paleoflood) characterization

In order to define an extreme hydrologic flood event from the upper Quinault river basin, a 100-point moving window of the mean and standard deviation were calculated for the detrended and resampled core 1 *inc/coh* dataset (motivation for choosing this parameter is covered in the results and discussion sections). Detrended and resampled measurements were utilized in order to minimize sampling biases introduced by the gradually increasing sedimentation rate and resultant uneven sampling of µXRF in time. After applying the moving window, extreme events were then identified as *inc/coh* values that exceeded the sum of the running mean and standard deviation at that particular point. In order to be considered as a separate event, rather than part of a single large event, we required that event values be separated by at least one data point with an *inc/coh* value that did not exceed the sum of the running mean and standard deviation. Selected events were also compared to core 1 photos as an additional check on the method to ensure that an obvious single deposit was not misidentified as being multiple events. Overall, this characterization allowed us to limit the biasing effects of changes in the location of sediment source (e.g., river mouth position on the upper Quinault River delta relative to the core location) and/or the nature of sediment delivery (e.g., changes in sediment focusing) by identifying only those proxy signatures that significantly exceeded (>1σ) the background values of the associated time interval. Using the average µXRF sampling step of 2 years and the 100-point window, mean values and standard deviation are averaged over a moving period of 200 years.

2.4.6 Historical discharge and climate data

Historical discharge data for the Quinault River were obtained from the United States Geological Survey (USGS) National Water Information System website at http://waterdata.usgs.gov/nwis. The Quinault River gaging station is located roughly 500 m downstream of Lake Quinault’s outlet (egress channel) to the lower Quinault River (47.45778 N, 123.88806 W) and thus instantaneous discharge values are buffered by water storage within the lake. Daily, monthly, and annual discharge statistics covering the period from October 1, 1911 through October 1, 2015 were acquired for comparison to monthly-tabulated oceanic and atmospheric data obtained from the NOAA Climate Prediction Center.
2.5 Results

2.5.1 Core lithologies and age models

Lake Quinault cores consist predominately of physically stratified detrital sediment ranging in size from medium-coarse sand at the lake delta (cores 4, 6, & 7) to clayey silt at the lake center (cores 3, 5, & 8) to silty clay at the distal end of the lake (cores 1 & 2) (Fig. 6). This grain size pattern reflects proximity to the source of water and sediment into Lake Quinault (i.e. the upper Quinault River) and coarsening-upward trends exhibited in each core are attributed to gradual lake delta progradation that has increased this proximity through time at a given core location. As much of the research in this paper is focused on µXRF data obtained for cores 1 & 3, we limit further discussion to the sedimentary records at these locations.

2.5.1.a Core 1

Core 1 was recovered from a flat sub-basin on the distal slopes of the lake, roughly 700 m from the lake’s outlet (Fig. 5). Core 1 sediment, which spans ~5.3 m, generally consists of tan to grey silt and clay interstratified with distinct organic-rich graded silt layers interpreted to represent deposition during high-discharge events (Fig. 7). These layers range in thickness from <1 cm to 8 cm and, based on visual inspection and physical properties, appear to be characterized by a basal unit enriched in organic matter and containing some sand-sized particles that grades upward to silt and clay-sized particles containing much less or no organic material. In some cases, these deposits are also capped by very fine, blue-gray clay laminae. Based on the calculated age model, core 1 sediments span a period of 3855 years with the sediment surface representing the year 2013 CE (Fig. 8). Average sediment accumulation rate increases from less than 1 mm yr\(^{-1}\) near the base of the core to several mm yr\(^{-1}\) at the top, but true rates of sedimentation have likely been highly variable as a result of episodic, event-driven deposition.
2.5.1.b Core 3

Core 3, recovered from the lake’s deep central basin, is ~6.5 m in length and mainly consists of tan to grey very fine sand and silt interstratified with distinct organic-rich graded sandy silt layers (Fig. 7). Compared to core 1, the layers in core 3 are generally thicker (>15 cm in some cases) and have a larger mean particle size, which is expected since the location of core 3 is significantly closer to the lake delta and likely receives a greater and coarser-grained sediment input during high discharge events. Owing to core 3’s location in the lake’s deep and flat central basin, sediment focusing may also contribute to the higher rate of sedimentation. The core 3 age model indicates that the core record spans 642 years, or the period 1371-2013 CE, with a relatively constant mean sedimentation rate of approximately 1 cm yr\(^{-1}\) (Fig. 8). Although deposition from year to year is likely highly variable due to the event-driven nature of sediment delivery to the lake basin, the higher sedimentation rate suggests the possibility that, for any given time interval, core 3 records deposition from a greater number of high discharge events as compared to core 1.

2.5.2 Geochemistry, magnetic susceptibility, and particle size

Correlation matrices for core 1 and 3 µXRF elemental abundance are provided as Tables 3A & 3B. For both core 1 and core 3, Fe, K, Si, and Ti all have strong positive correlations with one another (R ≥ 0.65) and a strong negative correlation with inc/coh. Ca, Rb, and Zr also share significant positive correlations (R ≥ 0.50) with one or more elements and a negative correlation with inc/coh, but in general these correlations are weaker and less consistent across elements. Mn is not strongly correlated to any other element, which is consistent with previous research indicating that Mn is not influenced by detrital input and more often may be a tracer of lake-bottom redox conditions (Croudace et al., 2006). With the exception of Mn and inc/coh, all elemental profiles show a decrease toward the sediment surface with periodic high and low frequency fluctuations superimposed on this long-term trend (Figs. 9 & 10). The inc/coh profile shows similar fluctuations, but the overall trend is reversed and inc/coh instead gradually increases toward the sediment surface. Mn shows no apparent trend with depth, but is punctuated by a series of large peaks in abundance that occur roughly every 50 cm. Superimposed on these low frequency peaks are smaller, higher frequency fluctuations that are continuous throughout the record.
Magnetic susceptibility, though sampled at a lower resolution than the µXRF data, parallels the general trends of Fe, K, Si, and Ti and shows the same upward decrease as well as high and low frequency fluctuations throughout the record (Figs. 9 & 10). Particle size, which is available only for the gravity core from site 3, has a strong positive correlation with $inc/coh$ ($R = 0.77$) and a strong negative correlation with MS ($R = -0.77$) and elements other than Mn, especially Si ($R = 0.73$) (Fig. 11). The particle size distribution is dominated by clay and silt, with up to 25% fine sand comprising the coarsest intervals. Overall, visual observations indicate that particle size is greatest in darker layers of the core and lowest in the lightest layers. This likely reflects the nature of deposition in Lake Quinault, in which high discharge events lead to the transport of both larger particles as well as a greater abundance of terrestrial organic material derived from surface erosion. This interpretation is supported by $\delta^{13}C$ and C/N values of samples taken from the thick dark brown layer near the top of core 5C-1G that indicate this deposit is enriched in terrestrial soil and plant debris (Fig. 12; Table 4; Prahl et al., 1994; Kaushal & Binford, 1999; Leithold & Blair, 2001; Leithold et al., 2013). This layer in core 5C-1G, a sample from which contains 5% organic carbon by weight, is correlative to the thick, dark brown, and coarse-grained layer in core 3C-1G, indicating that it is also enriched in terrestrial soil and plant debris.

### 2.5.3 Time series analysis

Considering the similar trends among µXRF time series, spectral analysis focused on $inc/coh$ scattering ratios since $inc/coh$ is most strongly correlated to particle size. Results of SLOMBS analysis for core 1 shows several broad power peaks at multi-centennial and multi-decadal frequencies (Fig. 13). In the low frequency bands, $inc/coh$ exhibits high power at various periodicities between 100 and 500 yrs, whereas shared periodicities in higher frequency bands cluster between 25-35 yrs and 50-90 yrs. There is also power at 7-8, 12, and 18-yr periodicities, but the amplitude of these signals is lower and less distinguishable from the red noise of the dataset. Core 1 spectral power is largely echoed by the core 3 results, which are generally noisier, but show common peaks at periodicities of 7, 12, 18, 25-35, 50-55, 80, and 100-150 yrs (Fig. 13).

Results of the CWT reveal that the multi-centennial periodicities are relatively constant through time, especially those centered around 250 and 500 yrs that have consistent 95% confidence across the wavelet spectrum (Fig. 14). Owing to the shorter time series of core
3, such low frequency signals are unresolvable, but the core 3 wavelet spectrum does show consistently strong power at a periodicity centered around 128 yrs that is also reflected to some extent in the spectrum for core 1. As for multi-decadal periodicities, power at 60-90 yrs is relatively consistent for core 1 despite a lack of continuity at the 95% significance level. A similarly consistent signal is apparent at periodicities in 40-70 yr range for core 3. In both spectrums, the power at <20 yr and 25-40 yr periodicities is more sporadic, having a much lower degree of continuity throughout the full length of the time series.

### 2.5.4 Extreme event (paleoflood) characterization

In total, the analysis of the inc/coh time series of core 1 resulted in the identification of 122 extreme events over a period of 3800 years, with extreme events defined as inc/coh values that exceed the sum of the running mean and standard deviation at a particular point in time. This number of events suggests we have isolated those with an average recurrence interval of approximately 30 years or greater (Fig. 15). Assessing the distribution of these events through the past 3800 years (Fig. 16), no outstanding trend in the frequency of extreme events is apparent. The majority of intervals are characterized by 2-5 extreme events, with notable exceptions being 0 events from ~1150 to ~1250 cal BP in addition to 7 and 8 events each during the periods ~2350 to ~2450 cal BP and ~1910 to ~2010 CE, respectively.

### 2.6 Discussion

#### 2.6.1 Flood deposition in Lake Quinault

Detrital-dominated, event-driven sedimentation within Lake Quinault is evidenced by the continuous sequence of graded layers characterized by basal units of relatively coarse sediment and a small, yet significant contribution of organic material that gives the sediment a light or dark brown hue. Above this basal unit, particle size of the layer becomes finer and organic matter decreases, giving the sediment a tan or light gray color. Where bioturbation has been limited, a lamina of very fine, bluish-grey clay caps the layer. Based on a knowledge of local catchment and lake processes, in which coarse-grained detritus enriched in terrestrial organic material is delivered to the lake only during times of high discharge, as well as similarity to deposits described from other lakes around the world (e.g., Arnaud et
al., 2005; Chapron et al., 2007; Lauterbach et al., 2012; Gilli et al., 2013; Schillereff et al., 2014), these layers are interpreted as flood deposits sourced from the upper Quinault River. We believe that during high discharge events (typically associated with winter storms), elevated runoff, discharge, erosive power, and transport capacity conspire to carry coarse detritus and terrestrial organic matter to deep and distal (relative to the delta) areas of Lake Quinault. Conversely, when discharge is low or only slightly elevated, such as during fairweather conditions or the protracted period of spring snowmelt, only fine particles are transported to deep water regions owing to reduced runoff, erosive power, and transport capacity (Kashiwaya et al., 1987, Chen et al., 2004, Peng et al., 2005; Conroy et al., 2008; Cuven et al., 2010; Warrier et al., 2014).

With respect to the signature of each individual flood layer, the coarse brown basal units are likely deposited when high river discharge leads to a suspended sediment concentration that exceeds the necessary density threshold to generate a hyperpycnal flow (underflow). Sediment not immediately deposited then gradually settles out of suspension according to its size and density, which explains the bluish-gray clay cap observed in some layers. Owing to the seasonal nature of rainfall at Lake Quinault, it is likely that the distribution of flood layers present in the lake’s central and distal areas is heavily skewed to the winter season, as 199 of the 200 highest daily discharge measurements from the years 1911-2015 took place between the months of October and March (U.S. Geological Survey, 2015).

2.6.2 A proxy record of flooding

Careful analysis of the uppermost thick event layer present between 7-25 cm depth in core 3-G, which can be correlated across all cores from the central and distal areas of the Lake Quinault basin, serves to highlight the physical and geochemical signature of an individual flood deposit (Fig. 11). Radiocarbon dating of plant debris present within this layer revealed a post-modern age, indicating it was deposited after the year 1950. This result, when calibrated to the post-bomb $^{14}$C curve (Hua et al. 2013; Table 2) and coupled with constraints from age-model interpolation and average sediment accumulation rates, suggests that this layer was deposited during the flood of March 19, 1997. This Quinault River flood produced a mean daily discharge of 1206 m$^3$ s$^{-1}$ at the Lake Quinault egress channel (lower Quinault River gaging station) and left the lake turbid for nearly 5 months (L. Workman, personal communication, 2013). This flood, which was facilitated by the
presence of a strong atmospheric river (Dettinger, 2004), is the highest recorded discharge since gaging began in 1911 and thus may be considered the 100-year flood of record for the catchment (U.S. Geological Survey, 2015).

Grain size, µXRF, and MS data all track the brown-colored and coarse-grained lower half of the layer via large excursions from mean values (Fig. 11). Mean grain size and inc/coh increase abruptly at the base of the layer, reach a maximum, slightly decrease above the maximum, and then increase to a second peak before decreasing to mean values where the deposit transitions back to a tan/grey color. With respect to particle size, such oscillatory patterns have been interpreted to reflect the rising and falling limbs of the flood hydrograph (Mulder et al., 2003, St.-Onge et al., 2004; Gilli et al., 2013; Simonneau et al., 2013), and we posit that this may also be the case for this particular layer in Lake Quinault. The inc/coh ratio, which has been used previously as a tracer of organic matter within lake sediments (Guyard et al., 2007; Burnett et al., 2011; Jouve et al., 2013), appears to hold the same relationship for Lake Quinault, with highest inc/coh values in the coarsest and darkest portions of the 1997 layer. Trends in MS, Si, and other elements are less straightforward, being opposite in sign to those in mean grain size and inc/coh and having more muted oscillations within the coarse brown layer. This result is contrary to some previous research that has shown detrital flood layers are marked by increases in µXRF abundance of elements such as Fe, K, Si, and Ti (e.g. Czymzik et al., 2013), as well as additional research that has shown a similar positive correlation with MS (Foster et al., 2003; Osleger et al., 2013; Li et al., 2013). However, a growing body of µXRF research has demonstrated that these correlations are particularly site-specific (Schillereff et al., 2014), and in the case of Lake Quinault we point to previous research on flood stratigraphies in lake sediments that indicates Fe, K, Si, and Ti can also be positively correlated with the finer-grained portion of a given flood layer, perhaps as a result of an increase in the proportion of clay minerals (Cuven et al., 2010; Wilhelm et al., 2013; Wirth et al., 2013). As such, the signature of Lake Quinault flood layers is interpreted in the manner described below.

Increased mean particle size reflects the relationship between high discharge into the lake and the ability to transport coarser particles more effectively from the delta to greater distances across the lake. As discharge wanes, the flood layer becomes normally graded as finer particles slowly fall out of suspension from the water column. Increased inc/coh abundance throughout the brown, coarse portion of the layer tracks the input of coarse
detritus enriched in terrestrial organic matter to the lake bottom, which is likely deposited via underflows when discharge is at or near its peak. The fact that the suite of positively correlated µXRF elements (Fe, K, Si, Ti, etc.) and MS show a marked decrease at the base of each flood layer, followed by an increase to mean values as particle size and organic matter decrease, is explained in one of two ways. First, these elements may truly exist in higher abundance in clay and silt-sized sediment in Lake Quinault, as suggested by prior research outlined above. In this scenario, clay and silt-sized particles would have a higher percentage of iron-bearing minerals that would cause the concordant increase of MS in these intervals (Dearing, 1999). Alternatively, it is possible that there is such a degree of chemical and magnetic homogeneity across all sizes of Lake Quinault sediment that decreases in elemental and MS abundance reflect dilution by organic material. This effect is difficult to quantify since µXRF data yields only relative abundances and the carbon content of Lake Quinault sediments is so low (~5% by weight even at the base of the 1997 flood layer), but remains a plausible explanation given that lake deposition is dominated by detrital input. Within this framework, we reason that the inc/coh parameter of the core logs serves as a viable paleohydrology and paleoflood proxy for Lake Quinault, with downcore peaks in inc/coh interpreted to reflect deposition of coarse sediment enriched in organic matter during periods of elevated river discharge into the lake.

2.6.3 Sources of uncertainty

Given Lake Quinault’s proximity to the Cascadia subduction zone as well as a network of active, shallow crustal faults in western Washington, one limitation in reconstructing a record of climate-controlled hydrologic variability, especially with respect to sedimentary event layers, is the possibility that some of these event layers may have been formed as the result of subaqueous flows triggered by seismic shaking. A companion study of Lake Quinault (Leithold et al., in revision) explores the lake’s paleoseismic record using sediment cores and seismic profiling, and the conclusions drawn from that study are used to inform the interpretations made here. The last known rupture of the Cascadia Subduction Zone occurred on January 26, 1700 CE and produced an estimated M_w 9.0 earthquake (Satake et al., 1996; Yamaguchi et al., 1997). Previous work also suggests a minimum of eight additional megathrust earthquakes that have occurred along the northern Cascadia margin during the past 3500 years (e.g. Atwater, 1987; Atwater, 1992; Atwater & Hemphill-Haley, 1997; Atwater et al., 2003; Kelsey et al., 2005; Nelson et al., 2006; Goldfinger et al., 2011;
Goldfinger et al., 2012; Graehl et al., 2014). Furthermore, sedimentary records from inland areas contain evidence for at least seven upper plate earthquake events sourced from the nearby Seattle, Canyon River, and Saddle Mountain faults during this same time period (Karlin & Abella, 1992; Karlin & Abella, 1996; Karlin et al., 2004; Walsh & Logan, 2007; Witter et al., 2008; Barnett et al., 2015), indicating multiple sources of seismicity with the potential to impact Lake Quinault.

Based on sedimentological and seismic evidence from Lake Quinault, Leithold et al. (in revision) identified markers of prodelta erosion and deposition indicative of delta-front collapse possibly caused by the 1700 CE megathrust earthquake. However, significant basin-wide disruption of Lake Quinault sediments, including wide-spread sediment degassing is limited to the interval 1650 to 1450 cal BP as evidenced by an abrupt shift in seismic facies correlated to a zone of disrupted core laminae. Although no similar seismic facies shifts are identified since this time interval, more recent earthquakes may nevertheless have had significant impacts on the nature of sediment delivery to the lake.

Based on globally observed effects of seismic shaking in high-relief mountainous terrain, it is assumed that nearby earthquakes of a sufficient magnitude (M_w 6.0 or above) have a measurable impact on the geomorphology of the upper Quinault River catchment (e.g. Dadson et al., 2004; Yanites et al., 2010; Hovius et al., 2011; Parker et al., 2011; Tang et al., 2011), an inference supported by predictions of strong to severe shaking throughout the Quinault catchment for M_w 7.4 and M_w 9.0 earthquakes along the Canyon River-Saddle Mountain fault and Cascadia subduction zones, respectively (USGS, 2009; USGS, 2016).

This impact would likely be characterized by widespread slope failures that would release vegetation, regolith, and bedrock from hillslopes and lead to increased sediment delivery to valley bottoms of the upper Quinault River catchment and ultimately to Lake Quinault, with the potential effect of temporarily impounding water and sediment behind landslide deposits, which could reduce river discharge and downstream sediment transport. The breaching of landslide dams, if occurring catastrophically, could also mimic levels of discharge and sediment transport associated with flood events resulting from excessive precipitation. Alternatively, even if landslide dams do not form, landslide activity may liberate a vast quantity of sediment that becomes immediately available for transport, leading to a protracted period of increased sediment input to the lake.
In addition to the aforementioned possibility of delta slope failure, change in delta morphology is also a likely consequence of seismic shaking, and has the potential to alter the nature of sediment delivery and distribution within the lake. Overall, all of these earthquake-related effects could bias the paleoflood record in Lake Quinault by creating hiatuses or abnormal increases in sediment deposition at a single sediment core location, and observations made in the aftermath of recent earthquake events in mountainous terrain indicate that fluvial systems may take years–to-decades to return to equilibrium (e.g. Dadson et al., 2004; Hovius et al., 2011; Howarth et al., 2012; Wang et al., 2015). Although adequate discharge is necessary to transport coarse particles to the distal areas of the lake, the *inc/coh* proxy, which is believed to trace the input of coarse-grained detritus enriched in terrestrial organic matter, may be biased by the overall volume of sediment and organic material that is available for transport at the time of flooding. By liberating vast quantities of sediment and vegetation from hillslopes, earthquake-driven landsliding may make it more likely for a given flood event to achieve the suspended sediment concentration necessary to generate a hyperpycnal flow. In this scenario, the deposition of coarse-grained, organic-rich detritus at the distal areas of Lake Quinault would be weighted toward those floods that occurred at times of greater sediment availability. In addition to coseismic landsliding, greater sediment availability could be the result of landsliding as a result of high-magnitude precipitation, or perhaps as the result of a prolonged period of below-average discharge during which time sediment continues to accumulate owing to the absence of large flood events.

Further sources of uncertainty include lateral delta migration that may have changed flow patterns and sediment supply to specific areas of the lake, erosion of existing deposits by underflows, and changes in sediment supply due to aseismic landsliding. There is also the possibility of human influences on sediment mobilization, although a prior study of microfossil abundance in the uppermost sediments of Lake Quinault found no evidence of changes regarding in-lake biological productivity as a result of human settlement in the Quinault valley (Stockner et al., 2003). Finally, as the Lake Quinault delta has gradually prograded to a position closer to the core 1 location, it is likely that the average deposit thickness for a flood event of a given magnitude has increased. However, visual inspection of the core record reveals that flood layers are >2 mm in thickness throughout all depths of the core 1 record, indicating that the 2 mm µXRF sampling resolution is adequate for capturing the *inc/coh* signature of all such flood layers during the past ~4000 yrs.
Altogether, it is acknowledged that the nature of both coseismic and aseismic processes may have resulted in “missing layers” from the Lake Quinault record or a bias toward flood events that occurred during time of greater sediment availability. However, in spite of these uncertainties, we believe that the vast majority of extreme flood events over the past ~4000 years ultimately left a discernible sedimentary signature in the deepest areas of the Lake Quinault basin, and that the lake serves as a valuable recorder of past hydrologic variability for the Pacific Ocean-facing western flank of the Olympic Mountains.

2.6.4 The paleoflood record of Lake Quinault

Owing to the sources of uncertainty discussed above, as well as the overall trend of increasing grain size attributed to delta progradation, we neither attempt to constrain the magnitude of discharge events associated with individual flood deposits nor make any inferences on changes in flood magnitude over the late Holocene. Instead, we focus on assessing changes in the frequency of extreme flood events as well as broader trends in the periodicity of flooding, which we believe are well constrained by the inc/coh proxy record of paleohydrology from Lake Quinault sediments.

The lack of a long-term trend in paleoflood frequency is mostly unsurprising since previous paleoclimate research on the Olympic Peninsula has suggested relatively stable climate conditions throughout the late Holocene (e.g. Brubaker and McLachlan, 1996; Gavin et al., 2001; Gavin and Brubaker, 2015). Irrespective of the lack of a long-term trend, previous lake studies of past hydroclimatic variability in Washington have linked precipitation variability to the El Niño Southern Oscillation and the Pacific Decadal Oscillation (Nelson et al., 2011; Steinman et al., 2012). Considering the results of wavelet analysis for cores 1 and 3, which reveal high power at multidecadal and multi-centennial periodicities, it is possible that Lake Quinault sediments are also preserving a signal of changing phases of the PDO, as the periodicities identified from both cores 1 and 3 are broadly consistent with those identified for the PDO over the past millennium (Fig. 14; MacDonald & Case, 2005). However, the historical discharge record of the Quinault River shows little to no correlation with the PDO index since 1950, and there is no relationship between the largest 50 recorded daily discharge measurements and positive or negative PDO (or ENSO) values. Nevertheless, the lack of robust late Holocene records of the PDO in the Pacific Northwest makes this association a target of future work, and we acknowledge the possibility that the
weak relationship between discharge and the PDO index since 1950 underrepresents a potentially stronger connection prevalent throughout much of the late Holocene.

Particularly noteworthy is the high number of extreme events recorded between ~2350 to ~2450 cal BP and over the past 100 years, the latter of which may relate to the observed warming that has persisted since the turn of the twentieth century (IPCC, 2014). As a result of this warming, it is possible that conditions favorable to the formation of atmospheric rivers became more common, just as models suggest that continued warming could increase the frequency and intensity of AR storms in the future (Dettinger et al., 2011). In particular, AR-driven Pineapple Express storms capable of bringing tremendous moisture to the Pacific NW may have increased in number over the past century, a conjecture partly supported by a ~50% increase between 1950 and 1990 in the number of days on which large-scale atmospheric circulation patterns were symptomatic of the Pineapple Express condition (Dettinger, 2004). The period ~2350 to ~2450 cal BP may have been characterized by similarly favorable conditions, although prior late Holocene paleoclimate records from the Olympic Peninsula (e.g., Brubaker & McLachlan, 1996; Gavin & Brubaker, 1999; Greenwald & Brubaker, 2001; Gavin et al., 2013) and other areas of the Pacific Northwest (e.g., Nelson et al., 2011; Ersek et al., 2012) provide no evidence for drastic changes during this time. Nevertheless, just as Kirby et al. (2012) interpreted a history of Holocene pluvial episodes from Lower Bear Lake as resulting from increased frequency of atmospheric river storms tracking across southern California, we hypothesize that flood deposition in Lake Quinault reflects the frequency of large storms tracking over this particular portion of the Olympic Peninsula.

The precise physical mechanism by which Pineapple Express storms are (or are not) steered over the Quinault River catchment is beyond the scope of this study, but it may partly reflect a connection with phases of the PDO since the majority of the most vigorous Pineapple Express moisture transport events have occurred during positive phases of the PDO (Dettinger, 2004). As a result, it is reasonable to suggest that the record of extreme events in Lake Quinault largely reflects the occurrence of exceptional atmospheric river storms focused over the slopes of the southwestern Olympic Mountains, which may be controlled at decadal-to-centennial frequencies by large-scale ocean and atmospheric teleconnections such as the PDO.
2.7 Conclusions

Sediment cores from Lake Quinault, WA yield a ~4000 year record of detrital-dominated, flood-driven deposition sourced from the upper Quinault River. Precipitation, which can exceed 5 m yr$^{-1}$ in the highlands of the upper Quinault River catchment, accumulates predominantly during winter as the Aleutian low intensifies and steers large storms toward the mountainous terrain of coastal northern Washington. During large winter storms, the combination of elevated runoff, river discharge, stream power, and sediment transport capacity throughout the UQRC lead to the deposition of graded flood layers in the deep and distal areas of Lake Quinault. Ranging from a few mm to several cm in thickness, these flood layers are characterized by a coarse basal unit enriched in river-borne organic matter likely deposited via hyperpycnal flows. The close association between these basal units, mean particle size, and the $inc/coh$ parameter derived from core-scanning $\mu$XRF allows for a high resolution (2 mm) proxy-based reconstruction of late Holocene hydrologic variability. Spectral results reveal high power at multi-decadal and multi-centennial periodicities, and statistical analysis suggests a significant increase in event frequency over the past 100 years that may reflect the occurrence of storms associated with the north Pacific atmospheric river known colloquially as the Pineapple Express. The most intense of these storms, which generally occur during positive phases of the PDO, are capable of producing tremendous precipitation and subsequent flooding in the Lake Quinault catchment. Persistent lake turbidity and high rates of detrital input, both of which are a consequence of large floods, have been tied to decreased productivity within Lake Quinault that correlates to declines in salmon escapement and run size (Stockner et al., 2003). Consequently, if such extreme events have indeed become more frequent over the past century, our results have significant implications for the future ecological, hydrologic, and geomorphic responses of Lake Quinault and the UQRC in light of model predictions indicating that Pineapple Express storms will increase in number and severity under a warming climate regime (e.g. Dettinger, 2004).

2.8 Acknowledgements

We give special thanks to the Quinault Indian Nation for providing us access to the natural gem that is Lake Quinault. Furthermore, the logistics of our study particularly benefited from
the advice and assistance of Bruce Wagner, Larry Gilbertson, and Bill Armstrong of the Quinault Indian Nation Department of Fisheries. We’d also like to thank Chris and Tom Iversen at Locharie Resort for providing pleasant and relaxing accommodations throughout the duration of fieldwork. We thank Jubril Davies and Bruce Riddell for their help with obtaining sediment cores, as well as Corey Moore and Deanna Metevier for their assistance with laboratory analyses. Discussions with Walt Robinson, Gary Lackmann, and Sandra Yuter strengthened our understanding of atmospheric processes for the Pacific Northwest. This work was supported by the National Science Foundation Geomorphology and Land Use Dynamics Program, EAR-1226064.

2.9 References


Guyard, H., Chapron, E., St-Onge, G., Anselmetti, F.S., Arnaud, F., Magand, O., Francus, P., Mélières, M., 2007. High-altitude varve records of abrupt environmental changes and mining activity over the last 4000 years in the Western French Alps (Lake Bramant, Grandes Rousses Massif). Quaternary Science Reviews 26, 2644-2660.


Holocene climate variability, Lake Tutira, North-Eastern New Zealand. Marine Geology 270, 30-44.


Table 1
Coring sites at Lake Quinault.

<table>
<thead>
<tr>
<th>Coring Site</th>
<th>Latitude (*N)</th>
<th>Longitude (*E)</th>
<th>Water Depth (m)</th>
<th>Core Length(^a) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>47.46845</td>
<td>-123.88138</td>
<td>47</td>
<td>5.3</td>
</tr>
<tr>
<td>2</td>
<td>47.46877</td>
<td>-123.90732</td>
<td>38</td>
<td>5.9</td>
</tr>
<tr>
<td>3</td>
<td>47.47669</td>
<td>-123.86113</td>
<td>70</td>
<td>6.5</td>
</tr>
<tr>
<td>4</td>
<td>47.48769</td>
<td>-123.84807</td>
<td>37</td>
<td>3.7</td>
</tr>
<tr>
<td>5</td>
<td>47.48039</td>
<td>-123.85674</td>
<td>67</td>
<td>5.6</td>
</tr>
<tr>
<td>6</td>
<td>47.48497</td>
<td>-123.85215</td>
<td>58</td>
<td>5.9</td>
</tr>
<tr>
<td>7</td>
<td>47.48664</td>
<td>-123.85018</td>
<td>50</td>
<td>6.3</td>
</tr>
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<td>8</td>
<td>47.4754</td>
<td>-123.8712</td>
<td>58</td>
<td>5.7</td>
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</table>

\(^a\)Core lengths are for piston cores only. Gravity core lengths are reported on Figure 5.
Table 2

$^{14}$C results for core sediments

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (cm)</th>
<th>Material dated (frag. = fragments)</th>
<th>$^{14}$C age (yr BP)</th>
<th>Calibrated age range $^{a,b}$ (95%) (yr BP)</th>
<th>Modeled age range $^{a,c}$ (95%) (yr BP)</th>
<th>Modeled age $^{a,c}$ - weighted mean (yr BP)</th>
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<tbody>
<tr>
<td>2</td>
<td>146</td>
<td>leaf frag.</td>
<td>102±28</td>
<td>15-268</td>
<td>221-274</td>
<td>248</td>
</tr>
<tr>
<td>2</td>
<td>286</td>
<td>leaf/woody plant frag.</td>
<td>1625±28</td>
<td>1414-1595</td>
<td>1365-1551</td>
<td>1450</td>
</tr>
<tr>
<td>2</td>
<td>328</td>
<td>leaf/woody plant frag.</td>
<td>1683±25</td>
<td>1534-1691</td>
<td>1557-1774</td>
<td>1650</td>
</tr>
<tr>
<td>2</td>
<td>369</td>
<td>large twig</td>
<td>2075 ± 25</td>
<td>1953-2124</td>
<td>1942-2124</td>
<td>2037</td>
</tr>
<tr>
<td>2</td>
<td>456</td>
<td>small twig</td>
<td>2725 ± 32</td>
<td>2759-2876</td>
<td>2762-2913</td>
<td>2824</td>
</tr>
<tr>
<td>2</td>
<td>600</td>
<td>woody plant frag.</td>
<td>4483 ± 27</td>
<td>4985-5290</td>
<td>4974-5285</td>
<td>5155</td>
</tr>
<tr>
<td>2</td>
<td>619</td>
<td>woody plant frag.</td>
<td>4724 ± 29</td>
<td>5326-5583</td>
<td>5319-5562</td>
<td>5408</td>
</tr>
<tr>
<td>3</td>
<td>30</td>
<td>woody plant frag. modern</td>
<td>-60 to -6.6</td>
<td>-48 to 4</td>
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</tr>
<tr>
<td>3</td>
<td>150</td>
<td>leaf</td>
<td>139 ± 23</td>
<td>7-279</td>
<td>78-205</td>
<td>138</td>
</tr>
<tr>
<td>3</td>
<td>274</td>
<td>woody plant frag.</td>
<td>319 ± 22</td>
<td>307-459</td>
<td>222-341</td>
<td>311</td>
</tr>
<tr>
<td>3</td>
<td>437</td>
<td>woody plant frag.</td>
<td>300 ± 22</td>
<td>300-452</td>
<td>405-494</td>
<td>451</td>
</tr>
<tr>
<td>3</td>
<td>555</td>
<td>woody plant frag.</td>
<td>661 ± 28</td>
<td>559-672</td>
<td>547-651</td>
<td>581</td>
</tr>
</tbody>
</table>

$^{a}$ yr BP refers to years before 1950 C.E.
$^{b}$ Dates calibrated using Oxcal 4.2 software, IntCal13 curve, and N.H. Zone 1 ext. (Ramsey, 2009; Hua et al., 2013; Reimer et al., 2013).
$^{c}$ Ages were modeled using Bacon software (Blauuw and Christen, 2011).
### Table 3A
Correlation matrix for core 1 μXRF elemental abundances

<table>
<thead>
<tr>
<th>element</th>
<th>Ca</th>
<th>Fe</th>
<th>K</th>
<th>Mn</th>
<th>Rb</th>
<th>Si</th>
<th>Sr</th>
<th>Ti</th>
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<tbody>
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<td>Fe</td>
<td></td>
<td>0.58</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>K</td>
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<td></td>
<td></td>
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<tr>
<td>Mn</td>
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<td>0.18</td>
<td>-0.09</td>
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<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Rb</td>
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<td>0.44</td>
<td>0.75</td>
<td>-0.04</td>
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<td></td>
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<tr>
<td>Si</td>
<td>0.62</td>
<td>0.62</td>
<td>0.77</td>
<td>-0.19</td>
<td>0.44</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Sr</td>
<td>0.44</td>
<td>0.10</td>
<td>0.12</td>
<td>-0.05</td>
<td>0.10</td>
<td>0.19</td>
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<tr>
<td>Ti</td>
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<tr>
<td>Zr</td>
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<td>0.44</td>
<td>0.52</td>
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<td>0.37</td>
<td>0.53</td>
<td>0.12</td>
<td>0.54</td>
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### Table 3B
Correlation matrix for core 3 μXRF elemental abundances

<table>
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<th>element</th>
<th>Ca</th>
<th>Fe</th>
<th>K</th>
<th>Mn</th>
<th>Rb</th>
<th>Si</th>
<th>Sr</th>
<th>Ti</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td></td>
<td>0.51</td>
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<td></td>
<td></td>
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<tr>
<td>K</td>
<td>0.42</td>
<td></td>
<td>0.84</td>
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</tr>
<tr>
<td>Mn</td>
<td>0.18</td>
<td>0.58</td>
<td>0.39</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rb</td>
<td>0.28</td>
<td>0.73</td>
<td>0.83</td>
<td>0.34</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si</td>
<td>0.52</td>
<td>0.70</td>
<td>0.82</td>
<td>0.30</td>
<td>0.64</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sr</td>
<td>0.61</td>
<td>0.36</td>
<td>0.32</td>
<td>0.06</td>
<td>0.28</td>
<td>0.45</td>
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<tr>
<td>Ti</td>
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<td>0.77</td>
<td>0.82</td>
<td>0.31</td>
<td>0.66</td>
<td>0.73</td>
<td>0.54</td>
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<tr>
<td>Zr</td>
<td>0.41</td>
<td>0.06</td>
<td>0.12</td>
<td>-0.06</td>
<td>0.09</td>
<td>0.14</td>
<td>0.31</td>
<td>0.36</td>
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Table 4
Concentration and isotopic composition (elemental and stable) of organic carbon in core S-G sediment

<table>
<thead>
<tr>
<th>Depth (^a) [cm]</th>
<th>%C</th>
<th>(\delta^{13}C)</th>
<th>C/N</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>1.25</td>
<td>-27.1</td>
<td>9.80</td>
</tr>
<tr>
<td>10</td>
<td>0.74</td>
<td>-25.2</td>
<td>7.50</td>
</tr>
<tr>
<td>11</td>
<td>0.78</td>
<td>-25.4</td>
<td>7.21</td>
</tr>
<tr>
<td>15</td>
<td>5.50</td>
<td>-26.6</td>
<td>24.8</td>
</tr>
<tr>
<td>22</td>
<td>1.62</td>
<td>-26.7</td>
<td>12.0</td>
</tr>
</tbody>
</table>

\(^a\) 0 cm = sediment-water interface
Figure 1. Satellite image of Washington’s Olympic Peninsula. Lake Quinault is colored in light blue and outlined in white. The locations of Seattle, Mount Olympus, the Strait of Juan de Fuca, and the Cascadia subduction zone are provided for reference. Inset map shows area of larger image with respect to the Pacific Northwest region of the United States and Canada.
Figure 2. Satellite image of Lake Quinault and the upper Quinault River catchment (UQRC). The UQRC is outlined in white; the confluence of the north and east forks of the UQRC can be seen approximately 10 km upstream of the northeast end of Lake Quinault. Inset map show location of larger image within the Olympic Peninsula.
Figure 3. Inset: Simplified tectonic map of the Cascadia Subduction zone, showing 36 mm yr$^{-1}$ convergence between the Juan de Fuca and North American plates (Base image obtained from serc.carleton.edu). Larger image: Simplified geologic map of the Olympic Peninsula and surrounding area. Rocks of the Olympic Subduction Complex (OSC) are colored in green, and basalts of the Eocene Crescent Formation (ECF) are colored in purple. Faults are indicated by solid or dashed black lines. The Hurricane Ridge Fault (HRF) marks the contact between the OSC and peripheral rocks such as the ECF. Nearby active fault abbreviations are as follows: Seattle Fault zone (SFZ); Saddle Mountain Fault (SMF); and Canyon River Fault (CRF); Lake Creek – Boundary Creek Fault (LCBCF).
Figure 4. Photos of Lake Quinault and the Upper Quinault River catchment. A. View of Lake Quinault looking northeast from the Late Pleistocene moraine that impounds the lake. A two-lane road runs along the moraine and provides scale for the foreground. Photo by Larry Workman, Quinault Indian Nation. B. Aerial view of the Lake Quinault delta and the downstream end of the Upper Quinault River. White specks in the lower right of the image are single family homes. Photo by Sam Beebe, Ecotrust.
Figure 5. Satellite image of Lake Quinault with core locations marked as yellow circles. Location of Lake Quinault on the Olympic Peninsula is marked by a white star on the inset satellite image. Lake depth contours shown at 10 m intervals; the 70 m, 60 m, and 50 m contours are labeled. Bathymetric data obtained from the Academic Seismic Portal (ASP) at the University of Texas Institute for Geophysics (UTIG), doi:10.1594/IEDA/500068.
Figure 6. Photographs of gravity cores obtained from each coring location. Core images are aligned so that those nearest to the delta are displayed to the right side of the figure and those furthest from the delta are to the left. Please refer to Figure 4 for precise coring locations. The black dashed line marks the base of the flood layer believed to have been deposited during March 1997. This layer can be traced easily across cores 2-6, with the base of the layer less certain in cores 4 and 7. At sites 2-6 the layer is normally graded, enriched in fine, brown-colored woody plant debris near its base, and capped by fine gray silty clay. Note that each gravity core exhibits stratigraphy generally representative of that present in the longer piston cores from each location. For better display, the width-to-length ratio, brightness level, and contrast of each image were increased using Adobe Photoshop CS6.
Figure 7. A. Photograph of core 1. Composite photograph consists of 5 separate images from each core section (length ≤ 150 cm). B. Mapped flood layers within core 1. Mapped layers include only those with readily distinguishable upper and lower boundaries. C. Photograph of core 3. Composite photograph consists of 7 separate images from each core section (length ≤ 150 cm). D. Mapped flood layers within core 3. As for core 1, mapped layers include only those with readily distinguishable upper and lower boundaries. Note that some layers may appear black in color – this is a result of multiple thin and stacked flood layers that may be unresolvable in print.
Figure 8. Age-depth models for cores 1 and 3 constructed using the IntCal13 calibration curve (Reimer et al., 2013) and the Bayesian modelling software Bacon (Blauuw and Christen, 2011). The calibrated $^{14}$C dates are shown in blue (before 1950 A.D.) and green (post-1950 A.D.). The red dashed lines denote the weighted mean age models; black-to-gray shaded regions depict the 95% confidence window for the age models, with darker shading indicating more likely ages.
Figure 9. Core 1 magnetic susceptibility (MS) and select µXRF elemental abundances. The composite photo of the core has been shrunk to fit the image and processed to increase brightness and contrast. MS and µXRF data have been smoothed using a 10-point window.
Figure 10. Core 3 magnetic susceptibility (MS) and select μXRF elemental abundances. The composite photo of the core has been shrunk to fit the image and processed to increase brightness and contrast. MS and μXRF data have been smoothed using a 10-point window.
Figure 11. Location 3 gravity core photograph, mean particle size, \textit{inc/coh}, Si, and MS. The flood layer interpreted to have been deposited during the March 1997 storm is highlighted in blue.
Figure 12. Location 5 gravity core with $\delta^{13}C$, C/N, and %C sampling locations and values indicated. The MS profile of core 5-G is also included to show the similarity of the deposit signature to that of core 3-G (Figure 11).
Figure 13. SLOMBS power spectra for core 1 (A & B) and core 3 (C & D) inc/coh time series. A. Lomb-Scargle periodogram of spectral power associated with low frequencies [0 0.02] (top) and a plot of the achieved significance level for each peak shown (bottom). B. Same as in A, except for higher frequencies [0.02 2.0]. C and D are the same as A and B, but for core 3. The confidence peak for the [0 .005] frequencies in C is greyed out since these frequencies are considered too low to interpret (> 200 yr periodicities in a < 700 yr core record). For all plots, frequencies with high power are labeled with their associated period in years, rounded to the nearest whole number.
Figure 14. A. Detrended and resampled time series of core 1 inc/coh (bottom) and its continuous wavelet power spectrum (top). Thick black contours designate regions of the wavelet power spectrum that are above the 95% confidence level against red noise. Areas outside the cone of influence (COI) have been left blank since edge effects may distort the signal in these areas. B. Same as A, but for core 3.
Figure 15. Time series of core 1 inc/coh parameter with running mean denoted by the solid black line. The sum (and difference between) the running mean and the running standard deviation are plotted as navy blue lines. Extreme events, defined as those inc/coh measurements that exceed the sum of the running mean and running standard deviation, are highlighted in a bold blue color, whereas the remainder of the time series is colored a lighter blue. The 1997 flood layer is the uppermost event identified and is marked with an arrow.
Figure 16. Histogram of extreme events identified using the core 1 \textit{inc/coh} time series and binned into 100-year intervals beginning in 2013 CE.
Figure 17. Squared wavelet coherence (WTC) between the 993-1997 CE reconstructed PDO (MacDonald and Case, 2006) and core 1 (top) and core 3 (bottom) inc/coh time series. Areas of significant signal coherence (95% confidence) between the time series are shown with thick black contours, and areas outside the cone of influence (COI) have been left blank since edge effects may distort the signal in these areas. Arrows pointing left indicate that the signals are out-of-phase, which is expected given that precipitation on the Olympic Peninsula is greater when the PDO is negative (e.g. Miller and Goodrich, 2007). The WTC can be considered the local correlation between time series in time frequency space; for complete details of the method, see Grinsted et al., 2004.
CHAPTER 3

Precipitation, landsliding, and erosion across the Olympic Mountains, Washington

3.1 Abstract

In the Olympic Mountains of Washington State, landsliding is the primary surface process by which material is eroded from hillslopes and delivered to river networks. However, the volumetric percentage of landslide material transported to valley bottoms as a result of triggering by either large earthquakes or high-magnitude atmospheric precipitation events remains unknown. To test the hypothesis that precipitation is linked to erosion, nearly 1000 landslides were mapped across a ~15 km wide x ~85 km long (1250 km$^2$) swath of the Olympic Mountains and the volume of hillslope material moved by each slide was calculated using previously published area-volume scaling relationships. Divided among a grid of 40 blocks of roughly equal area, landslide volume was compared to mean annual precipitation data acquired from the PRISM climate group for the period 1981-2010. Statistical analysis reveals a significant correlation ($r = 0.5; p < .001$) between landslide volume and mean annual precipitation, and assessment of the period 1990-2015 shows that 98% of landslide volume was produced in the windward, high-precipitation side of the range during this interval. Normalizing to area, this volume yields an erosion rate of $0.28 \pm 0.11$ mm yr$^{-1}$, which is similar to other estimates of erosion throughout the Olympic Mountains, including those from river sediment yield, cosmogenic $^{10}$Be, fluvial terrace incision, and thermochronometry. The lack of large historic earthquakes makes it difficult to assess the precise relative contributions of precipitation and seismic shaking to total erosion, but our results suggest that climate, and more specifically a sharp precipitation gradient, plays a significant role in driving erosion and landscape evolution over both short and long timescales in the Olympic Mountains.
3.2 Introduction

Situated along an active plate boundary in a region that receives abundant precipitation, previous research has explored the roles of both tectonics and climate in the topographic and structural evolution of Washington’s Olympic Mountains (Willett, 1999; Montgomery & Greenberg, 2000; Montgomery, 2001; Montgomery & Greenberg, 2000; Brandon, 2002; Stolar et al., 2007), but the relative contribution of each mechanism to the surface response of the landscape remains unknown. The interplay between climate, tectonics, and mountain belt evolution is a research topic that has garnered much attention and debate in the 21st century (e.g. England & Molnar, 1990; Burbank et al., 1996; Montgomery et al., 2001; Riebe et al., 2001; Bonnet and Crave, 2003; Whipple, 2009; Binnie et al., 2010; Korup et al., 2010; DiBiase and Whipple, 2011; Ferrier et al., 2013; Gasparini & Whipple, 2014), and although it is accepted that precipitation facilitates erosion in mountainous regions, deciphering the relative contribution of precipitation to overall rates and spatial patterns of erosion has proven difficult. Much of this difficulty likely arises due to complexity inherent in erosional processes at various spatial and temporal scales as well as inconsistencies in feedback cycles between climate, tectonics, and erosion across different regions of the globe. In addition to studies addressing mountain belt evolution over long (> 10^6 yr) timescales (e.g. Willett, 1999; Montgomery et al., 2001a; Bishop, 2007; Whipple, 2009), recent research has also sought to better understand the relationship between climate and tectonics over much shorter timescales by attempting to constrain the feedbacks between rainfall, earthquakes, and erosion during and after singular meteorological or seismic events (e.g. Dadson et al., 2004; Guzzetti et al., 2004; Yanites et al., 2010; Hovius et al., 2011; Parker et al., 2011; Tang et al., 2011; Lira et al., 2013; Marc et al., 2015). These studies offer valuable information on the erosional response of particular areas to individual geologic phenomena, but nevertheless no empirical consensus has been reached regarding the net effect of climate on erosion rates.

With respect to the Olympic Mountains, previous research has detailed various facets of the region’s geologic evolution and led to the conclusion that the range maintains a long-term (> 10^6 yr) flux steady-state (Willett & Brandon, 2002) with exhumation, surface uplift, and erosion rates that increase gradually from the Pacific Ocean toward the central massif (Brandon et al., 1998; Pazzaglia & Brandon, 2001). Over shorter timescales, less is known about patterns of erosion, with research limited to river incision and 10Be-derived catchment-
averaged erosion rates from the Clearwater River basin (Wegmann and Pazzaglia, 2002; Belmont et al., 2007). Furthermore, although much research has focused on the paleoseismic record of the nearby Cascadia Subduction Zone, the most recent rupture of the CSZ occurred more than 300 years ago on January 26, 1700 CE (Satake et al., 1996; Yamaguchi et al., 1997) and thus the impact of a megathrust earthquake on the surface processes of individual catchments within the Olympic Mountains remains unclear. Based on globally observed effects of seismic shaking in high-relief mountainous terrain, it is assumed that the 1700 CE event resulted in widespread slope failures that may have imparted a decadal-scale impact on fluvial networks and associated ecosystems (e.g. Dadson et al., 2004; Yanites et al., 2010; Hovius et al., 2011; Parker et al., 2011; Tang et al., 2011), but this assumption is unproven and it is unknown how the sediment flux from such a singular event would compare to the flux derived from, for instance, the 668 landslides inventoried in the Quinault River catchment of the Olympic Mountains during the seismically quiescent period 1939 to 1998 (Quinault Indian Nation, 1999). As a result, the relative role of seismicity versus precipitation in driving the long-term pace of erosion is a fundamental, yet unanswered question crucial not only for understanding the broader evolution of this and other mountain ranges in tectonically-active temperate climatic regions, but also for developing a more complete knowledge of catchment sediment flux and how it may vary over time and space in response to multiple external forcings.

In this paper, we use a ~1250 km² swath across the Olympic Mountains that is oriented parallel to both tectonic (exhumation and uplift) and precipitation gradients of the range as a natural laboratory to test the following two hypotheses: 1) That the volume of sediment moved by landslides mimics patterns of precipitation across the range; and 2) That erosion via aseismic landsliding is a significant contributor to the long-term rate of erosion. We discuss our results relative to previous research conducted within the Olympic Mountains as well as field evidence for possible upland response to past climatic and/or seismic events.

### 3.3 Geology and Climate of the Olympic Mountains

The Olympic Mountains represent the sub-aerial portion of the accretionary wedge formed as a consequence of oblique subduction of the Juan de Fuca plate beneath the North American plate across the Cascadia margin. Around 15 Ma, a decrease in accommodation
space led to uplift and exposure of Olympic Mountain segment of the subduction wedge above sea level (Brandon et al., 1998). At the surface, the Olympic Mountains consist primarily of the Olympic subduction complex (OSC), which is characterized by highly deformed, stratigraphically discontinuous, and partially metamorphosed sedimentary rocks (Fig. 1; Tabor & Cady, 1978; Brandon et al., 1998). Juxtaposed against the rocks of the OSC is a horseshoe-shaped surface exposure of older oceanic basalts situated on the footwall of the eastward-dipping Hurricane Ridge thrust fault (Tabor & Cady, 1978; Brandon & Calderwood, 1990; Babcock et al., 1994). These basalts, which make up the Eocene Crescent Formation (ECF), form a border along the northern, eastern, and southern sides of the range.

As the Juan de Fuca plate continues to converge and subduct beneath the North American plate at 36 mm yr\(^{-1}\), rates of uplift increase landward from \(\sim 0.3 \text{ km Ma}^{-1}\) near the Pacific Coast of Washington to \(\sim 1 \text{ km Ma}^{-1}\) near the summit of Mount Olympus, which, at an elevation of 2430 m (7980 ft), is the present-day high point of the range (Brandon et al., 1998; DeMets & Dixon, 1999). Previous research has shown that net uplift, or rock influx into the range, is balanced by the processes of denudation, or eroded sediment flux out of the range, and thus the Olympic Mountains are believed to maintain a flux steady state in which there is no net addition or loss of mass over geologic timescales (Brandon et al., 1998; Pazzaglia & Brandon, 2001; Willett and Brandon, 2002; Stewart and Brandon, 2004).

Geographically, the Olympic Mountains constitute a large portion of the Olympic Peninsula and are unique in the United States for the combination of rugged, glacially-sculpted topography and large swaths of temperate rain forest (Fig. 2). Rainforest conditions exist predominantly on the western, windward side of the range due to orographic forcing as moisture-laden, landfalling Pacific storms encounter the steep, mountainous terrain (e.g. Minder et al., 2008; Neiman et al., 2011; Gavin & Brubaker, 2015). The bulk of precipitation, which can exceed 5 m annually in the higher-elevation western side of the range (Fig. 3), falls during the winter season as the Aleutian low pressure system intensifies and drifts south from the Gulf of Alaska, subsequently steering moisture and winds toward the Pacific Northwest (PRISM Climate Group, 2012; Gavin and Brubaker, 2015). While snow in lowland areas is very rare, during cold storms a significant amount of precipitation may fall as snow atop high ridges (Minder et al., 2008). Modeling of orographic precipitation patterns across the range show that deposition of snow at high elevations has a minor role
in rainfall distribution, but that cold storms with low freezing levels can lead to increased leeward precipitation produced via downwind advection of frozen hydrometers generated in the orographic cloud (Zängl, 2007; Minder et al., 2008). The typically cool, rainy winters transition to warm and dry summers once the Aleutian low weakens and the North Pacific high pressure system strengthens and brings northwesterly air flow, low humidity, and sunny skies to the region (Ware & Thomson, 2000; Patterson et al., 2013).

3.4 Methods

The chosen study area encompasses a ~1250 km$^2$ NW-SE swath across the entirety of the Olympic Mountains (Fig. 4). For comparison between landsliding and precipitation, the swath was subdivided into a grid of 40 blocks of roughly equal area (~30 km$^2$). Block demarcations were based on Washington State township and range subdivisions (6 mi x 6 mi grid), but note that the numbering scheme used for this study is our own and does not correlate to the public land survey numbering system. Within each block, individual landslides were mapped manually using all available satellite imagery in Google Earth (individual images provided by DigitalGlobe, Landsat, NASA, USDA Farm Service Agency, and U.S. Geological Survey) for the years 1990-2015. Landslide borders were carefully drawn to avoid grouping multiple scars as a single unit; flow paths, chutes, and other features outside of the main scar were not mapped as part of the total landslide area (Fig. 5). Due to difficulties in identifying landslide boundaries in high elevation areas with little-to-no vegetation, possible landslides in these areas were not included in the assessment. Any landslides believed to be associated with anthropogenic activity, i.e. landslides associated with logging roads of the commercial forest lands surrounding Olympic National Park, were also excluded from the study.

Mapped landslides were imported into the ArcMap 10.3 software environment in order to tabulate areas for individual polygons, and landslide volumes were calculated using area-volume scaling parameters reported in Larsen et al. (2010). Based on field observations and satellite imagery, landslides with areas ≥ 5000 m$^2$ were considered deep-seated bedrock landslides, whereas those with areas < 5000 m$^2$ were labeled shallow failures consisting predominately of soil and vegetation. Landslide volumes were summed to yield a total landslide volume for each block within the mapping grid, and volumes were
subsequently normalized to block area to correct for minor differences in block size. Uncertainties in the normalized volumes include an estimated 15% mapping error as well as reported standard deviations ($\sigma$) for $\alpha$ and $\Upsilon$ in the formula, $V = \alpha A^\Upsilon$ ($V = $ landslide volume, $A = $ landslide area, $\alpha$ & $\Upsilon = $ empirically-derived constants) used to estimate the volume of a landslide based on its surface area. For bedrock slides, $\sigma_\alpha = 0.03$ and $\sigma_\Upsilon = 0.02$, and for shallow soil slides $\sigma_\alpha = \sigma_\Upsilon = 0.005$ (Larsen et al., 2010). Propagating these uncertainties into the volume calculation for each grid block means that total volume uncertainty is heavily dependent upon landslide size distribution. When total volume is spread among many similarly-sized landslides, uncertainty is small since all individual landslides are unlikely to be biased in the same direction. However, in the cases where the total landslide volume is dominated by a single or several very large landslides, errors are less likely to be balanced and thus uncertainty is high (e.g., Emberson et al., 2016).

Modeled 30-year mean annual precipitation (1981-2010) was obtained from the PRISM Climate Group at Oregon State University and compared to total landslide volume for each ~30 km$^2$ grid cell of the study area swath, which extends from the western foothills to the northeast corner of the range. The spatial distribution of precipitation is a community model output of the PRISM Climate Group, and data covering the entire Olympic Peninsula were obtained in ASCII format and imported to ArcMap, where mean annual precipitation was averaged for each grid block (Fig. 3). The Pearson correlation coefficient, $r$, between landslide volume and precipitation was calculated using MATLAB; a permutation test with $10^6$ iterations was utilized to establish the significance of the correlation. Mean elevation and mean slope were also calculated for each grid block using USGS 10 m Digital Elevation Model (DEM) data available via the National Elevation Dataset (http://nationalmap.gov/elevation.html). In the same manner described for precipitation data, correlations between these values and landslide volume were computed and tested for significance using MATLAB.

During the mapping procedure, each landslide was labeled according to the date of the imagery on which the landslide first appeared. Landslides present in the earliest (1990) imagery have unknown timing, but those that occurred during the years 1990-2015 were isolated in order to calculate an erosion rate for the study area based on the total volume of material released by landslides over the 25-year period (Fig. 6). Total landslide volume and associated erosion rates were also derived for “high precipitation” and “low precipitation,”
regions of the study area in which high precipitation regions correspond to grid blocks having mean annual precipitation $\geq 2.5 \text{ m yr}^{-1}$ (blocks 1-18, 20-21) and low precipitation regions correspond to grid blocks having mean annual precipitation $< 2.5 \text{ m yr}^{-1}$ (blocks 19, 22-40).

3.5 Results

A total of 937 landslides were mapped throughout the study area. Summed together, the mapped landslides encompass an area of $4.5 \times 10^6 \text{ m}^2$ and a volume of $2.0 \times 10^7 \text{ m}^3$ (Table 1). Mean annual precipitation, $P_A$, and landslide volume, $V_L$, exhibit a strong positive correlation ($r = 0.5; p < .001$) with the data trend best described by the power function, $V_L = \chi P_A^\omega$, where $\chi$ and $\omega$ are empirical constants with values $\chi = 6.0 \times 10^{-5}$ and $\omega = 2.77$ (Fig. 7). Landslide volume is also positively correlated to mean slope, but visual inspection of the data suggests this correlation is mainly controlled by grid blocks 37-40 that have anomalously low values for mean slope and little to no corresponding landslide volume. When these grid blocks are removed and a correlation coefficient is calculated for grid blocks 1-36, there is no correlation between landslide volume and mean slope, but landslide volume and mean annual precipitation maintain the strong positive correlation. Landslide volume and mean elevation are neither correlated when including all grid blocks nor when excluding the outliers.

For the years 1990-2015, total estimated landslide volume for the study area is $4.7 \times 10^6 \text{ m}^3$ (0.0047 km$^3$). Dividing by a total study area of $1.26 \times 10^6 \text{ m}^2$ and a time span of 25 years yields an erosion rate of $0.15 \pm 0.06 \text{ mm yr}^{-1}$ (Table 2). Note that the $\pm 0.06 \text{ mm yr}^{-1}$ error is the result of incorporating uncertainty associated with the calculation of total landslide volume.

Partitioned between grid blocks of high versus low precipitation, $4.6 \times 10^6 \text{ m}^3$ of landslide volume comes from the region of high precipitation, or the windward side of the range. After normalization to area to account for a slight size difference between the two partitions, this value accounts for 98% of all landslide volume during the years 1990-2015 across the entire study swath of the Olympic Mountains (Fig. 8; Table 2). Calculating erosion for each of these partitioned volumes via the equation $R_e = V_L/A/t$, where $R_e$ is the erosion rate, $V_L$ is
landslide volume, \( A \) is the area considered, and \( t \) is the corresponding time interval, results in an area-wide surface lowering rate of \( 0.28 \pm 0.11 \text{ mm yr}^{-1} \) for the windward high precipitation western flank of the range and \( 0.005 \pm 0.002 \text{ mm yr}^{-1} \) for the leeward “low precipitation” region on the eastern flank of the orographic rain shadow created by the zone of high topography centered around Mount Olympus.

### 3.6 Discussion

Based on prior studies of post-landslide vegetation recovery (e.g. Guariguata, 1990; Smale et al., 1997; Francescato et al., 2001; Chou et al., 2009), the temperate maritime climate of the Olympic Peninsula, and observations made from satellite imagery of the study area spanning 1990-2015, we conclude that the majority of mapped landslides likely occurred during the past century, and that landslides possibly triggered by the 1700 CE CSZ earthquake have long been revegetated and rendered unidentifiable via imagery-based remote sensing techniques. As a result, we consider the volume of material moved by the mapped landslides to reflect aseismic climate-driven erosion in the Olympic Mountains, as there have been no large earthquakes beneath the Olympic Peninsula during the historic period (Rogers et al., 2016; Pacific Northwest Seismic Network, 2016). In the absence of seismic shaking, regional landslides are the probable result of intense precipitation or rapid snow melt contributing additional moisture to wet-season near-saturated regolith on hillslopes, thereby decreasing the effective normal stress and leading to slope instability (e.g. Swanson et al., 1986; Gerstel, 1999; Quinault Indian Nation, 1999; Benda et al., 2003; Wegmann, 2004). The significant correlation between landslide volume and mean annual precipitation supports this conclusion, as more and larger landslides are found in portions of the Olympic Mountains that receive greater amounts of rain and snowfall. Given that landsliding is the dominant mode of sediment transport from hillslopes to fluvial networks in mountainous environments (Hovius et al., 1997; Hovius et al., 2000; Korup et al., 2004; Korup et al., 2010; Wenske et al., 2012; Bennett et al., 2013), this suggests that the windward, wetter side of the Olympic Mountains erodes more quickly than the leeward, drier side during periods of seismic quiescence. This result is consistent with rates of long-term exhumation and river incision from the range (e.g, Brandon et al., 1998; Pazzaglia and Brandon, 2002).
Calculating the 1990-2015 erosion rate from landslides allows for an assessment of climate-driven erosion with respect to long-term estimates of Olympic Mountain landscape evolution. A rate of $0.28 \pm 0.11 \text{ mm yr}^{-1}$ for areas of high precipitation is similar to those calculated for the Clearwater River basin using cosmogenic $^{10}\text{Be}$ that integrate the catchment-averaged signal of erosion over the past 1500-4000 years depending on the rate associated with each individual sample (Belmont et al., 2007). This timespan includes several great earthquake events identified in paleoseismic studies focused on the CSZ (e.g. Atwater, 1987; Atwater, 1992; Atwater & Hemphill-Haley, 1997; Atwater et al., 2003; Kelsey et al., 2005; Nelson et al., 2006; Goldfinger et al., 2011; Goldfinger et al., 2012; Graehl et al., 2014), and therefore the comparable rate of modern aseismic erosion may indicate that atmospheric precipitation has played a significant role in setting the pace of erosion throughout the late Holocene. This inference is supported by a ~4000 year record of deposition in Lake Quinault that is dominated by climate-modulated trends of river discharge and sediment transport into the lake, with relatively limited evidence of disruption from seismic shaking (Leithold et al., in review; Smith et al., in prep). The $0.28 \pm 0.11 \text{ mm yr}^{-1}$ rate is also comparable to short term (annual-to-decadal) estimates of erosion based on Hoh River sediment yield (Nelson, 1986) as well as studies of longer-term Olympic Mountain evolution, including a ~140 ka record of river incision derived from Clearwater River strath terraces, the spatial pattern of exhumation derived from Zircon and Apatite fission track and U-Th/He thermochronology across the range, and estimates of erosion based on slope and mean local relief for a 10-km diameter swath of the range parallel to the vector of Cenozoic tectonic convergence (Brandon and Vance, 1992; Brandon et al., 1998; Pazzaglia and Brandon, 2001; Montgomery and Brandon, 2002).

The similarity of our calculated erosion rate with those from previous research conducted over various spatial and temporal scales suggests that it is an accurate assessment of recent landscape evolution in high precipitation areas of the Olympic Mountains. Furthermore, the relationship between our rate of $0.28 \pm 0.11 \text{ mm yr}^{-1}$ and the mean long-term exhumation rate of $0.28 \text{ mm yr}^{-1}$ for the entire Olympic Peninsula (Brandon et al., 1998) provides compelling evidence that climate exerts a major control on the pace of erosion. Even if we compare the lower bound of our calculated rate, $0.17 \text{ mm yr}^{-1}$, to the estimated exhumation rate for the central massif, $0.75 \text{ mm yr}^{-1}$ (Fig. 9; Brandon et al., 1998), it is still the case that nearly 25% of erosion (assuming flux steady state, in which long-term accretion of material at the front and base of the subduction wedge is equivalent

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to the volume of eroded material leaving via surface processes) may be the result of precipitation-induced landsliding. Nevertheless, this conclusion also raises questions about the nature of erosion on the leeward side of the range. Although every grid block assessed in this “low precipitation” portion of the range has a value of mean annual precipitation that exceeds the U.S. national average (PRISM Climate Group, 2012; NOAA, 2016), aseismic landslide volume is limited and appears to have a negligible contribution to overall erosion. One possible explanation for this negligible contribution is that since slope failures are not frequently triggered by local climatology and hydrology, more slopes may instead remain close to the critical threshold of stability over longer periods of time and thus fail more rapidly in the event of an earthquake. In this hypothesized scenario, relatively fewer landslides would occur throughout the windward side of the range since a lower percentage of slopes would be at or near the critical threshold, and thus the contribution of coseismic landslide volume to the long-term rate of erosion would be comparatively greater for the leeward side of the range.

However, confounding the above scenario is a ~0.17 mm yr\(^{-1}\) erosion rate we derived using sediment volume \((1.56 \times 10^7 \text{ m}^3)\) impounded behind the former Glines Canyon dam that was located on the Elwha River from 1927-2012 (Bountry et al., 2010). This rate was calculated by assuming a bulk reservoir sediment density of 1.5 g cm\(^{-3}\) (e.g. Avnimelech et al., 2001; Audry et al., 2004; Snyder et al., 2004; Lazzari et al., 2015), then converting to rock volume (assuming an average rock density of 2.7 g cm\(^{-3}\)) and dividing by the total upstream drainage area (632 km\(^2\)) and the time elapsed between measurements of reservoir volume (83 yrs). Although this rate is ~40% lower than our calculated rate for the windward side of the range, our hypothesis would predict an even lower rate of erosion for the Elwha River basin since it is located on the leeward side of the range. One explanation may be that climate is still an important driver in erosion for this side of the range, but that high-elevation mass wasting processes such as debris avalanches and/or flows are the primary means of sediment release rather than landslides. Another possible explanation is anthropogenic influence, but we reject this notion since Olympic National Park protects a vast majority of the upstream drainage area, and thus this unexplained discrepancy is a target of future work.

If our results are accurate, and precipitation contributes quite significantly to the long-term rate of erosion across the windward side of the range, a large question still remains as to the
potential impact that large earthquakes may have on surface processes of upland areas, and subsequently how this manner of erosion compares to that driven by climate. Although it is impossible to quantify coseismic landsliding associated with the most recent CSZ rupture in 1700 CE, field data and observations in the upper Quinault River valley do provide evidence for possible impacts of this event. Roughly 20 km upstream of Lake Quinault, a cut-bank along the East Fork of the Quinault River exposes a massive 15 m-thick, poorly-sorted unit consisting of large, angular boulders within a gravel and sand matrix. This massive unit both overlies and is capped by ~1 m-thick beds of well-sorted fluvial gravels, a stratigraphy suggestive of disruption in fluvial deposition caused by a large landslide event. The location of this deposit also corresponds with a prominent convexity in the Quinault River profile, which provides additional support for the possibility of a large, valley-blocking landslide (Fig. 10; Leithold et al., in revision). Immediately downstream of the landslide deposit, and coincident with the confluence of Graves Creek, is a 5 m-high alluvial terrace that is continuous along both banks of the Quinault River for hundreds of meters and includes multiple buried and rooted conifer tree stumps in growth position from which limited 

In addition to the landslide and terrace deposits at Graves Creek, the U.S. Bureau of Reclamation mapped and dated a series of well-pronounced fluvial terraces located along a ~15 km span of the Quinault River immediately upstream of Lake Quinault, with the lowest terrace yielding an age of 1610 ± 90 CE from detrital Populus charcoal and the intermediate terrace yielding an age of 1460 ± 50 cal BP from detrital conifer charcoal (Bountry et al. 2005). The age of the lower terrace, like the deposit at Graves Creek, brackets the 1700 CE CSZ earthquake and in this case suggests the possibility of channel aggradation as the result of increased sediment delivery initiated by widespread coseismic landsliding. The formation of the intermediate terrace may be similarly described, as the 1460 ± 50 cal BP age is consistent with evidence of paleoseismic sediment disruption in Lake Quinault that may have been the result of an earlier CSZ rupture or a large earthquake triggered by slip along a nearby upper crustal fault (Leithold et al., 2016). Collectively, these terrace deposits indicate that seismic shaking most likely does have an appreciable impact on erosion within
Olympic Mountain catchments, but that the degree of impact may depend on both the geomorphic conditions at the time of the event (e.g., antecedent soil moisture, duration since previous event; e.g., Wegmann & Pazzaglia, 2002) as well as the degree and spatial extent of ground shaking produced by each earthquake, which are related to the magnitude and location of the earthquake, but are also influenced by the surrounding topography (e.g. Meunier et al., 2008; Buech et al., 2010; Gallen et al., 2015). This is highlighted by particularly strong evidence for significant Lake Quinault sediment disruption ~1500 cal BP, but limited signs of lake-wide disturbance resulting from any known large subsequent earthquakes (Leithold et al., 2016).

Coupling our results with indirect evidence of seismically-induced erosion, it appears that the long-term erosional evolution of the Olympic Mountains is governed in part by a balance between gradual, consistent climate forcing and abrupt, infrequent tectonic forcing. During times of seismic quiescence, areas of the range that receive abundant precipitation continue to erode at a rate comparable to the long-term average, whereas the leeward portion of the range erodes much more slowly. When large earthquakes do occur and subsequently induce powerful ground shaking, it is probable that landsliding is widespread throughout all portions of the range, but it may also be the case that landsliding is of a greater density, area, and volume in those areas that are sheltered from receiving excessive orographic precipitation. Furthermore, even the season in which a large earthquake happens may matter in terms of hillslope sediment production. For instance, if a large earthquake takes place during the winter to early spring months when slopes are saturated, the shaking may generate more hillslope failures than if the same magnitude earthquake occurred when slopes are drier, such as in August or September. We acknowledge that this hypothesis is difficult to test in the absence of seismic activity, but future work will focus on developing a more complete understanding of short and long-term erosion rates for individual catchments on either side of the orographic precipitation gradient.

3.7 Conclusions

Sediment volumes calculated for mapped landslides from a ~15 km-wide swath of the Olympic Mountains reveal a significant positive correlation with mean annual precipitation. The windward side of the range, which receives abundant orographic precipitation, is
characterized by a greater density of landslides and associated volume of transported material as compared to the drier leeward side of the range. For the period 1990-2015, the total volume of landslide sediment from the windward, high-precipitation portion of the study area yields a mean erosion rate of $0.28 \pm 0.11 \text{ mm yr}^{-1}$. This rate is similar to other estimates of erosion throughout the Olympic Mountains, including those from river sediment yield, cosmogenic $^{10}\text{Be}$, fluvial terrace incision, and thermochronometry (Nelson, 1986; Brandon et al., 1998; Pazzaglia and Brandon, 2001; Belmont et al., 2007). The coherence between our rates and those previously published suggests that climate is an important driver of both short and long-term landscape evolution in regions of the Olympic Mountains receiving heavy annual precipitation. Additional work is necessary to determine the precise relative contributions of climate (i.e. precipitation) and tectonics (i.e. earthquakes) to overall erosion, but our results provide compelling evidence that climate plays an important and measurable role in delivering sediment from hillslopes to the fluvial network. This has significant implications for understanding mountain belt evolution in areas with similarly sharp precipitation gradients and for comprehending the mechanisms controlling source to sink sediment fluxes in such environments.

### 3.8 Acknowledgements

Thanks to NC State students Robert Lane, Deanna Metevier, Catherine Opalka, and Sharese Roberts for their assistance with landslide mapping, and to the staff of Olympic National Park for their assistance with fieldwork logistics. This work was supported in part by a Geological Society of America Graduate Student Research Grant and by the National Science Foundation Geomorphology and Land Use Dynamics Program, EAR-1226064.

### 3.9 References


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<td>3.71 x 10⁵ ± 1.41 x 10⁵</td>
<td>4.68 x 10⁵ ± 1.78 x 10³</td>
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ᵃA 15% error is assumed for the mapping procedure.
ᵇUncertainty includes propagated errors from the determination of landslide area and st. dev. for $\alpha$ and $\gamma$ in the formula, $V = \alpha Y$, used to calculate landslide volume ($V$).
A 15% error is assumed for the mapping procedure.

Uncertainty includes propagated errors from the determination of landslide area and st. dev. for $\alpha$ and $\Upsilon$ in the formula, $V = \alpha A \Upsilon$, used to calculate landslide volume ($V$).
Table 2
Landslide volumes and erosion rates for high and low precipitation areas of the study region.

<table>
<thead>
<tr>
<th>Study region</th>
<th>Area (km$^2$)</th>
<th>Mean elevation (m)</th>
<th>Mean slope (degrees)</th>
<th>Mean precip (m yr$^{-1}$)</th>
<th>landslide volume (m$^3$)$^a$ (1990-2015)</th>
<th>erosion rate (mm yr$^{-1}$)$^a$</th>
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<tr>
<td>High Precipitation (windward)</td>
<td>657</td>
<td>848</td>
<td>27.0</td>
<td>3.93</td>
<td>$4.60 \times 10^6 \pm 1.75 \times 10^6$</td>
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</tr>
<tr>
<td>Low Precipitation (leeward)</td>
<td>601</td>
<td>1172</td>
<td>24.7</td>
<td>1.63</td>
<td>$7.45 \times 10^4 \pm 2.83 \times 10^4$</td>
<td>$0.005 \pm 0.002$</td>
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<tr>
<td>All</td>
<td>1258</td>
<td>1010</td>
<td>25.7</td>
<td>2.78</td>
<td>$4.67 \times 10^6 \pm 1.77 \times 10^6$</td>
<td>$0.149 \pm 0.056$</td>
</tr>
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$^a$Uncertainty includes propagated error from landslide area and 1 σ st.dev. for α and γ in the formula, $V = \alpha A \gamma$, for volume (V).
Figure 1. Inset: Simplified tectonic map of the Cascadia Subduction zone, showing 36 mm yr\(^{-1}\) convergence between the Juan de Fuca and North American plates (Base image obtained from serc.carleton.edu). Larger image: Simplified geologic map of the Olympic Peninsula and surrounding area. Rocks of the Olympic Subduction Complex (OSC) are colored in green, and basalts of the Eocene Crescent Formation (ECF) are colored in purple. Faults are indicated by solid or dashed black lines. The Hurricane Ridge Fault (HRF) marks the contact between the OSC and peripheral rocks such as the ECF. Nearby active fault abbreviations are as follows: Seattle Fault zone (SFZ); Saddle Mountain Fault (SMF); and Canyon River Fault (CRF); Lake Creek – Boundary Creek Fault (LCBCF).
Figure 2. Satellite image of Washington's Olympic Peninsula. The locations of Seattle, Mount Olympus, the Strait of Juan de Fuca, and the Cascadia subduction zone are provided for reference, and the downstream stretches of major rivers are marked with white lines. Inset map shows area of larger image with respect to the Pacific Northwest region of the United States and Canada.
Figure 3. Left: Color-coded map of mean annual precipitation across the Olympic Peninsula for the period 1981-2010. The study area is outlined in black, and downstream reaches of major rivers are marked with black lines. Precipitation data obtained from PRISM Climate Group (2012). Right: Precipitation (blue) and elevation (black) profiles across the Olympic Mountains. Profile location is marked by the red line labeled A-A’ in the left image. Note the abrupt decrease in precipitation moving east across the range crest.
Figure 4. Satellite image of the Olympic Peninsula showing the 40-block mapping grid used for the study. Boundaries of individual grid blocks were based on Public Land Survey Township and Range subdivisions for Washington State; average block area is 31 km$^2$ (Table 1). The East Fork of the Quinault River is labeled for reference to comments made in the discussion, and the arrow points to the East Fork’s junction with Graves Creek, which is the location of images presented in Figure 9.
Figure 5. Left: Perspective-view example of delineated landslides within the Upper Quinault River catchment using Landsat imagery available in Google Earth (grid block 13). Right: Area calculations derived by importing the landslide polygons originally determined in Google Earth into ArcMap 10.3. In the lowest-elevation polygon, landslide volume is written in red font below the surface area determination. Volumes were calculated using power-law scaling parameters (e.g., Larsen et al., 2010).
Figure 6. Perspective-view example of a post-1990 landslide utilized for erosion rate calculation. Left: Imagery from 1990 with scar of future landslide outlined by the white dashed line. Right: Post-landslide imagery (from 2013) with same white dashed line. The depicted landslide occurred along Rustler Creek in the North Fork Quinault River catchment (grid block 13; Figure 4).
Figure 7. Plot of landslide volume (m$^3$) versus mean annual precipitation (m yr$^{-1}$) for each of the 40 blocks within the study grid. Vertical error bars represent propagated uncertainty in the volume calculation, which includes estimated 15% surface mapping errors as well as reported standard deviations for $\alpha$ and $\Upsilon$ in the formula, $V = \alpha A^\Upsilon$, used to calculate volume based on landslide area ($\sigma_\alpha = 0.03$ and $\sigma_\Upsilon = 0.02$ for bedrock slides and $\sigma_\alpha = \sigma_\Upsilon = 0.005$ for shallow soil slides; Larsen et al., 2010). The Pearson $r$ value for the correlation is 0.5 with a $p$ value of $< 0.001$ derived using a permutation test with $10^6$ iterations. For any given value of mean annual precipitation within a grid block, the resulting landslide volume is best predicted by the power function, $V_L = \chi P_A^\omega$, where $\chi$ and $\omega$ are empirical constants with values $\chi = 6.0 \times 10^{-5}$ and $\omega = 2.77$. 
Figure 8. Relative contribution of high-precipitation (grid blocks 1-18, 20-21) and low-precipitation (grid blocks 19, 22-40) areas to total landslide volume for the period 1990-2015. The study area is outlined in black, and downstream reaches of major rivers are marked with black lines. Over the 25 year span, 98% of landslide volume has been produced in the high precipitation, windward portion of the study grid, yielding an erosion rate of $0.28 \pm 0.11$ mm yr$^{-1}$ for this area. The line B to B' corresponds to figure 9.
Figure 9. Smoothed curve of integrated fluvial incision and long-term exhumation rates along the profile B-B’ that is shown in figure 8. The smoothed curve was obtained from Pazzaglia and Brandon (2002) and is based on reset apatite fission-track ages from Brandon et al. (1998) and terrace incision rates from Pazzaglia and Brandon (2002). For reference, the approximate location of Mt. Olympus along the profile is indicated by a black arrow.
**Figure 10.** Field and geochronological evidence for the impact of the 1700 CE Cascadia subduction zone earthquake on the upper Quinault River catchment (see arrow on Figure 4 for location). A. Contour map of the area surrounding the confluence of Graves Creek and the East Fork of the Quinault River. Hypothesized valley-blocking landslide is mapped in yellow and downstream event terrace is mapped in orange; consult discussion for additional details. Black star corresponds to the location of panel C; white star corresponds to the location of panel D. B. Panoramic photograph of the northern cut-bank of the Quinault River, just upstream of the confluence with Graves Creek. The 5 m-thick event terrace is visible on the downstream (left) side of the image and a portion of the 15 m-thick landslide deposit is visible on the right, upstream side of the photo. C. Photograph of the 15 m-thick landslide deposit; the boulder outlined in white has a long axis of approximately 1 m. D. Photograph of the event terrace from the opposite bank of the Quinault River, just downstream of Graves Creek. The red star indicates the location of the $^{14}$C sample acquired from the outermost growth rings of one of several large trees buried in place by the aggradation of sediment now recorded as the event terrace upon which a mature conifer forest is now growing. E. Diagrammatic view of OxCal $^{14}$C calibration results for the sample shown in panel D. The radiocarbon age of 219 ± 27 BP correlates to a calibrated age that potentially brackets the 1700 CE earthquake, the timing of which is marked on the graph by a red arrow. Percentages correspond to the probability that the $^{14}$C age represents the calendar age window indicated by the intersection of the blue $^{14}$C curve and the 219 ± 27 BP uncalibrated age (Ex. 46% probability that the sample dates to sometime within the span 1725-1815 CE). Note that percentages are based on reported results spanning 3σ (99.7%) certainty in the age determination. Base diagram obtained and calibration performed using Oxcal 4.2 software with the IntCal13 curve and N.H. Zone 1 ext. (Ramsey, 2009; Hua et al., 2013; Reimer et al., 2013).