Rockwall slope erosion in the northwestern Himalaya

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Key Points:

- Rates of periglacial rockwall slope erosion are defined for the northwestern Himalaya using cosmogenic ^{10}Be concentrations in sediment from medial moraines.
- Tectonically driven uplift offers a first order control on patterns of rockwall slope erosion.
- Precipitation and temperature play secondary roles in this erosion.

Abstract

Rockwall slope erosion is an important component of alpine landscape evolution, yet the role of climate and tectonics in driving this erosion remains unclear. We define the distribution and magnitude of periglacial rockwall slope erosion across 12 catchments in Himachal Pradesh and Jammu and Kashmir in the Himalaya of northern India using cosmogenic ^{10}Be concentrations in sediment from medial moraines. Beryllium-10 concentrations range from $0.5\pm0.04x10^4$ to $260.0\pm12.5x10^4$ at/g SiO_2, which yield erosion rates between 7.6±1.0 and 0.02±0.04 mm/a. Between ~0.02 and ~8 m of rockwall slope erosion would be possible in this setting across a single millennium, and >2 km when extrapolated for the Quaternary period. This erosion affects catchment sediment flux and glacier dynamics, and helps to establish the pace of topographic change at the headwaters of catchments. We combine rockwall erosion records from the Himalaya of Himachal Pradesh, Jammu and Kashmir and Uttarakhand in India and Baltistan in Pakistan to create a regional erosion dataset. Rockwall slope erosion rates progressively decrease with distance north from the Main Central Thrust and into the interior of the orogen. The distribution and magnitude of this erosion is most closely associated with records of Himalayan denudation and rock uplift, where the highest rates of change are recorded in the Greater Himalaya sequences. This suggests that tectonically driven uplift, rather than climate, is a first order control on patterns of rockwall slope erosion in the northwestern Himalaya. Precipitation and temperature would therefore come as secondary controls.

Keywords: periglacial erosion; rock uplift; climate; cosmogenic isotopes; sediment flux

1. Introduction

A number of studies have underlined the importance of the erosion of bedrock-dominated slopes, referred to here as rockwall slopes (Fig. 1), in catchment sediment flux, relief production, topographic configuration and glacier dynamics of steep alpine environments (Heimsath and McGlynn, 2008; MacGregor et al., 2009; Seong et al., 2009; Ward and Anderson, 2011; Benn et al., 2012 Scherler and Egholm, 2017; Orr et al., 2019). The lateral erosion of slopes has been shown to exceed rates of vertical incision through glacial and fluvial processes, and therefore to a greater extent than previously thought, contribute to denudation budgets and landscape change on the catchment and mountain range scale (Brocklehurst and Whipple, 2006; Foster et al., 2008).

The erosion or destabilisation of rockwalls is commonly attributed to climate-modulated processes (Hales and Roering, 2005; Böhlert et al., 2008; Krautblatter and Moore, 2014). Changes to slope hydrology and weathering environments, permafrost degradation, and stress redistribution from changing glacier extents and the (un)loading of slopes through deposition and erosion can each decrease slope stability and lead to mass wasting (André, 2003; Cossart et al., 2008; McColl, 2012; Gallach et al., 2018). Periglacial weathering processes that are driven by moisture and temperature variability, and include freeze-thaw, frost cracking and ice wedging, are particularly critical for the detachment and disintegration of rock from rockwalls (Heimsath and McGlynn, 2008; McColl and Davies, 2013; Eppes and Keanini, 2017). This detachment is considered stochastic, and through rockfall and other mass wasting delivers debris to glacier surfaces (Sanders et al., 2012; Gibson et al., 2017; Sarr et al., 2019).

Rockwall erosion is also sensitive to the lithology and structure of the bedrock slope, glacial/fluvial erosion and seismicity (Sanchez et al., 2009; Leith et al., 2010). Disentangling climatic and non-climatic controls of slope failure and longer term rockwall slope evolution is challenging, with several studies arguing that a combination of factors instead dictates the stability of steep rockwalls in alpine regions (McColl, 2012; Krautblatter and Moore, 2014; Gallach et al., 2018).

Orr et al. (2019) were able to identify a tentative positive relationship between periglacial rockwall slope erosion and precipitation in the northwestern (NW) Himalaya by comparing erosion rates inferred from cosmogenic nuclide concentrations of medial moraine sediment across three glacier systems. Higher rates of erosion were determined for catchments with enhanced monsoon precipitation. Rather than identifying precipitation as the only control, their study instead suggests that rockwall slope erosion is more complex, and is likely dictated by the interaction between tectonics, climate, topography, and surface processes that are specific to each catchment. This challenges the argument that in the tectonically active ranges of the Himalaya, the rate of debris transfer from the hillslope to the glacier surface is largely controlled by rock uplift and topographic steepness (Scherler et al., 2011; Gibson et al., 2017).

In this study, we seek to better define the distribution and magnitude of periglacial rockwall slope erosion in the NW Himalaya by building upon the work of Orr et al. (2019) and quantifying erosion rates for a suite of 12 catchments. Rates of rockwall slope erosion are derived from cosmogenic nuclide ^{10}Be concentrations measured in sediment from medial moraines. Our new erosion dataset is combined with existing rockwall slope erosion records from Seong et al. (2009), Scherler and Egholm (2017) and Orr et al. (2019). This regional rockwall erosion dataset is compared to records of catchment-wide erosion and exhumation for the NW Himalaya to evaluate the extent to which rockwall slope erosion may differ from other records of landscape change, which have been averaged across various spatial and temporal scales. We compare patterns of rockwall slope erosion to variations in geology, tectonics, climate and topography throughout the region, to resolve the primary controls of rockwall slope erosion in the NW Himalaya. Steep north-south gradients in elevation, slope, relief, rock uplift and precipitation has made the Himalayan-Tibetan orogen an ideal location to evaluate these controls (Bookhagen and Burbank, 2006, 2010; Scherler et al., 2011). In line with existing assessments of the principle controls of rockwall slope erosion in alpine settings (Hales and Roering, 2005; Böhlert et al., 2008; Krautblatter and Moore, 2014), and landscape change more generally throughout the orogen (Thiede et al., 2004; Grujic et al., 2006; Clift et al., 2008; Gabet et al. 2008; Wulf et al., 2010; Deeken et al., 2011), our hypothesis is that climate-modulated processes will largely dictate the patterns of rockwall slope erosion in the NW Himalaya. Finally, we determine to what extent rockwall slope erosion and its controls, in this high-altitude and high relief setting, can contribute to the longstanding debate over the significance of climate versus tectonics in driving both short- and long-term landscape change. Until now, much of the research that has contributed to this debate is based in either unglaciated or deglaciated environments. This study will provide a unique insight into how erosional processes modulated by climate and/or tectonics operate within glaciated catchment headwaters.

2. Regional Setting

The Himalayan-Tibetan orogen is the result of the continued continental collision and partial subduction between the Indian and Eurasian lithospheric plates (Searle et al., 1997). The Indus-Tsangpo Suture Zone (ITSZ) defines the collision zone between these plates in the NW Himalaya and contains remnants of the Neo-Tethys Ocean (Fig. 2). The suture zone marks the northern boundary of the Tethyan Himalaya (Searle, 1986; Steck et al., 1998; Schlup et al., 2003). Between the early Miocene and Pleistocene, deformation driven crustal shortening initiated the development of a sequence of foreland propagating thrust systems that divide the lithotectonic units that lie south of the Tethyan Himalaya. The South-Tibetan Detachment (STD) and the Main Central Thrust (MCT) bound the Greater Himalaya Crystalline Core Zone to the north and south, respectively (Frank et al., 1973; Searle and Fryer, 1986; Walker et al., 1999; Miller et al., 2001; Vannay et al., 2004). This unit has been divided into two sub-units: southern Greater Himalaya sequence (GHS-S) and northern Greater Himalaya sequence (GHS-N; DeCelles et al., 2001; Thiede and Ehlers 2013). South of the Greater Himalaya and MCT lies the Lesser Himalaya sequence, which is bounded to the south by the Main Boundary Thrust (MBT). South of the MBT lies the Sub-Himalaya and Main Frontal Thrust (MFT; Upreti, 1999; Miller et al., 2000; Vannay et al., 2004). Continued crustal shortening and thrust and strike-slip faulting throughout the orogen means that the NW Himalaya remains tectonically active (Hodges et al., 2004; Vannay et al., 2004; Bojar et al., 2005), even though some regions in northern India, such as Ladakh, have undergone tectonic quiescence or dormancy since the early Miocene (Kristein et al., 2006, 2009). Hodges (2000), Yin and Harrison (2000) and Streule et al. (2009) provide further details of the Himalayan lithotectonic units and the timing of movement throughout the fault systems.

Two atmospheric systems primarily govern northwest Himalayan climate: the Indian summer monsoon that advects moisture from the Indian Ocean between late May and September, and the Northern Hemispheric mid-latitude westerlies, which bring moisture from the Mediterranean, Black and Caspian seas between December and March (Benn and Owen, 1998; Gadgil 2003; Lang and Barros 2004; Wulf et al., 2010; Mölg et al., 2013). A steep south-north precipitation gradient likely became established during the late Miocene, perpendicular to the strike of the mountain belt (~8 Ma; Qiang et al., 2001; Liu and Dong, 2013), due to the high elevation ranges of the Greater Himalaya inhibiting the northward migration of moisture to the interior of the orogen. Monsoon air masses are forced to ascend, condense and form clouds along the Himalayan front, which creates a rain shadow down the leeside of this orographic barrier (Bookhagen et al., 2005a, b; Wulf et al., 2010). During times of increased monsoon strength, moisture is thought to penetrate farther into the interior of the orogen (Finkel et al., 2003 Bookhagen et al., 2005a, b; Wulf et al., 2010). The northern hemispheric mid-latitude westerlies operate at higher tropospheric levels than the Indian summer monsoon. The orographic capture of moisture transported by this atmospheric system is therefore focused in high elevation ranges (> 4500 m asl) as winter snowfall (Weiers 1995; Lang and Barros 2004). Today, mean annual precipitation declines from ~1500−3000 mm in the Lesser and Greater Himalaya ranges, to <150 mm in the interior of the Tethyan Himalaya and Tibetan Plateau (Bookhagen and Burbank, 2006; Fig. 2).

The distribution and magnitude of precipitation has been shown to vary both temporally and spatially throughout the Himalayan-Tibetan orogen during the late Quaternary (Burbank et al., 2003; Bookhagen et al., 2005a, b). Fluctuations in monsoon strength driven by changes in orbital insolation, the migration of the intertropical convergence zone, convective localized monsoon storms and sporadic heavy rainfall are thought to cause some of this variability (Finkel et al., 2003; Owen et al., 2008; Thomas et al., 2016). On the local to regional scale (10^{2-4} km^2), topography and wind direction exert controls on the migration of moisture throughout the NW Himalaya (Bookhagen et al., 2005a, b), and create localized microclimates throughout individual mountain ranges (Benn and Owen, 1998; Bookhagen and Burbank, 2010; Wulf et al., 2010). Landscape change in the NW Himalaya is precipitation sensitive, where shifts in the availability and source of moisture is shown to initiate changes to sediment flux, hillslope processes (Bookhagen et al., 2005; Bookhagen and Burbank, 2006; Sharma et al., 2017; Kumar et al., 2018) and the timing of glaciation (Owen and Dortch, 2014; Saha et al., 2018).

Studies have shown that glaciation in the Himalayan-Tibetan orogen is largely influenced by climatic conditions including shifts in the strength or behaviour of regional and/or global atmospheric and oceanic systems (Owen and Sharma 1998; Watanabe et al., 1998; Solomina et al., 2015; 2016; Saha et al., 2018). The Himalayan Holocene stages (HHs; Saha et al., 2018), Himalayan-Tibetan Holocene glacial stages (HTHS; Saha et al., 2019), semi-arid western Himalayan-Tibetan orogen stages (SWHTs; Dortch et al., 2013) and monsoonal Himalayan-Tibetan stages (MOHITs; Murari et al., 2014) provide regional syntheses of the glacial records throughout the NW Himalaya (Table 1). Variability in the timing and forcing of glaciation across short distances (10^{1-2} km) throughout the NW Himalaya is commonly attributed to microclimatic variability and local geologic factors such as topography and glacier type (Barr and Lovell 2014; Anderson et al., 2014; Owen and Dortch 2014).

2.1. Study Areas

We selected 12 accessible catchments along the south-north precipitation gradient of the NW Himalaya (Figs. 2, 3, Supplementary Item 2). Each catchment supports either a cirque or small valley glacier with distinct and well-preserved medial moraines. The northern-most sites of this study are located in the Ladakh and Zanskar Ranges of the Ladakh region in Jammu and Kashmir of northern India and the Shigar region of Baltistan in Pakistan (Fig. 2). For this latter site, a pre-existing erosion dataset was reanalyzed. The Indian summer monsoon delivers two-thirds of the annual precipitation to Ladakh (87 mm/a; Table 1), whereas the mid-latitude westerlies provide the primary source of moisture to the Shigar region. Glaciers in the Ladakh region are small (1–10 km^2) cold-based sub-polar glaciers, which are precipitation sensitive and sublimation dominated (Benn and Owen, 2002).

The arid/semi-arid climatic setting of the Ladakh region is largely responsible for the preservation of very old landforms and sediment deposits (>400 ka; Owen et al., 2006; Hedrick et al., 2011; Orr et al., 2017, 2018) and slow rates of landscape change (<0.07±0.01 mm/a; Dortch et al., 2011a; Dietsch et al., 2015). The investigated Gopal, Stok and Amda catchments are three north-facing transverse catchments in the high-altitude desert landscapes of the northern Zanskar Range in Ladakh that retain small valley glaciers (Figs. 2, 3; Table 1). Karzok and Mentok are northeast-trending catchments that drain the Rupshu Massif in central Zanskar of the Ladakh region. Cirque glaciers occupy the upper reaches of these catchments.

The Lahul-Spiti and Kullu district catchments are located in Pir Panjal and Greater Himalaya ranges of the Himachal Pradesh in northern India. Precipitation is primarily sourced from the Indian summer monsoon (950–1020 mm/a; Table 1). Glaciers are large, temperate and melt dominated, and fed by precipitation from the summer monsoon and mid-latitude westerlies (Benn and Owen, 2002; Su and Shi, 2002). The Urgos valley glacier extends throughout the upper reaches of a southeast trending tributary catchment of the Miyar basin in the Lahul-Spiti district (Fig. 3). Panchi is a north-facing catchment with a small valley glacier, located north of the Keylong and Darcha villages. Shitidar and Batal are north facing tributary catchments with two glaciers each. The Chhota Shigri and Hamtah catchments are also north facing and are each occupied by one glacier. Beas Kund is a southeast trending catchment located on the southern slopes of the Pir Panjal Range in the Kullu district. Two valley glaciers are contained in this catchment. The Indian summer monsoon also dominates annual precipitation in the Uttarkashi district of Uttarakhand, northern India, a region our study revisits and reanalyses the rockwall erosion data of Orr et al. (2019).

3. Methodology

3.1. Topographic analyses

Geomorphic maps of the 12 investigated catchments were prepared in the field and then refined using Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) global digital elevation models (GDEMs; 30-m-resolution), Landsat Enhanced Thematic Mapper Plus (ETM+) imagery and

Google Earth imagery. Topographic parameters including catchment area, 3-km-radius relief, mean slope, hypsometry and aspect were calculated using the Spatial Analyst Toolbox in ArcMap 10.1. These analyses were also conducted for Baltoro glacier system in the Shigar of Baltistan, Pakistan (Seong et al., 2009) and Bhagirathi glacier system in the Uttarkashi district of Uttarakhand, northern India (Orr et al., 2019) to enable comparisons between rockwall slope erosion and catchment parameters throughout the NW Himalaya.

3.2. Cosmogenic nuclide erosion rate theory for rockwall slopes

Production of cosmogenic ^{10}Be via neutron spallation occurs within the first few metres of the bedrock surface of the rockwall slopes and decreases approximately exponentially with depth. As material detaches from the rockwall slopes through erosion, new material moves into the zone of nuclide production (Lal, 1991; Uppala et al., 2005; Balco et al., 2008). The ^{10}Be inventory at the rockwall slope surface is the integrated cosmogenic nuclide production within quartz, during its exhumation through the zone of production (Eq.1).

$$N(z) = \left(\frac{P_0}{\lambda + \frac{E}{z_*}}\right) e^{-\frac{z}{z^*}} \tag{1}$$

N is the measured nuclide concentration at depth z, Po is the nuclide surface production rate, z^* is the e-folding length scale for falloff of production with depth between the surface ($z_* = \Lambda/\rho$, where Λ is the spallogenic mean free path and ρ is the target material density), E is the erosion rate and λ is the radiogenic decay.

Rates of rockwall slope erosion can be inferred by measuring cosmogenic ^{10}Be concentrations within medial moraine sediment. In our study areas, medial moraines form within the glacier ablation zones as a result of englacial sediment melt out. This sediment is sourced and transferred from accumulation zone rockwall slopes to the glacier surface via rockfall processes and avalanching, before being transported englacially to the equilibrium line of the glacier and exhumed to the surface (Matsuoka and Sakai, 1999; Goodsell et al., 2005; MacGregor et al., 2009; Mitchell and Montgomery, 2006; Dunning et al., 2015; Fig. 1). The ^{10}Be concentration of the medial moraine sediment reflects the averaged nuclide inventory of the source rockwall slopes. Due to the stochastic nature of rockwall slope erosion, the ^{10}Be concentrations across the rockwalls are unlikely to be spatially uniform. The mean concentrations of these rockwall slopes are instead considered steady in time and linked to the mean erosion rate (Ward and Anderson 2011).

$$E_{RS} = \frac{P_0 z_*}{N} \tag{2}$$

The medial moraine nuclide concentrations and rockwall slope production rates are used to calculate the rockwall slope erosion rate (E_{RS}; Eq. 2; Lal,1991; Granger et al., 1996; Balco et al., 2008). The longer the original sampled material has remained within the production zone of the rockwall slopes before being transferred to the glacier surface, the greater the ^{10}Be concentration measured in the medial moraine sediment, and therefore the slower the inferred rockwall slope erosion rate. This approach to quantifying rockwall slope erosion accounts only for the neutronic component of incoming cosmic rays and assumes a negligible loss of ^{10}Be due to radioactive decay and steady state erosion over time (von Blanckenburg, 2005). Further details of this methodology and its assumptions are provided by Supplementary Item 1 and in Ward and Anderson (2011) and Sarr et al. (2019).

3.3. Cosmogenic nuclide analyses

Orr et al. (2019) argue that the rates of rockwall slope erosion in the upper Bhagirathi catchment in the Uttarkashi district of Uttarakhand are best represented by the rates derived from the centermost medial moraine of Gangotri glacier. This is because the ^{10}Be concentrations of the medial moraine fall within uncertainty of each other, and that the other medial moraines are shown to receive input from the lateral moraines and hillslopes along the ablation zone of the glacier. The study recommends that multiple samples should be taken from each medial moraine and/or glacier to constrain and evaluate any variability in rockwall slope erosion throughout the catchment headwaters. Two or fewer samples are only appropriate when the medial moraine is well preserved with steep relief ridges, has no interaction with ablation zone slopes and where other sampling locations do not fit these criteria. With these recommendations in mind, we carefully collected between one and five samples from stable and well-defined medial moraine ridges for our 12 catchments studied (Figs. 3, 4). Each sample location was ≥

200 m^2 in area, to avoid sampling from a single source slope or rockfall event (see Supplementary Item 1). Approximately 3 kg of sediment with a grain size of <3 cm (clay-coarse gravels) was collected for each sample using bulk sediment sampling methods of Gale and Hoare (1991). Detrital samples of this grain size are shown to infer time-averaged erosion rates effectively, and for this study, are representative of the processes that contribute to rockwall slope denudation (Lal, 1991; Seong et al., 2009; Delunel et al., 2010; Puchol et al., 2014; Tofelde et al., 2018). Each sample was named using the initial term 'G' for 'glacier' followed by an abbreviated term for the catchment name. The samples were numbered in ascending order from the glacier snout, for glaciers with more than one sample. For example, the G_{Cht1} sample was located closest to the snout of Chhota Shigri in Lahul-Spiti, whilst G_{Cht5} was located farthest up-glacier.

Each sample was crushed and sieved in the Sedimentology Laboratories at the University of Cincinnati. A sample aliquot with equal input from all grain size fractions was created for each sample to avoid any one grain size from being overrepresented in the ^{10}Be analysis. The sample aliquot was crushed, sieved and the 250–500 μm fraction was retained for processing. The extraction of quartz and ^{10}Be isolation and purification was conducted at the Geochronology Laboratories at the University of Cincinnati, using the chemical procedures of Nishiizumi et al. (1989), von Blanckenburg et al. (2004) and Wittmann et al. (2016). The ^{10}Be/^{9}Be was measured using accelerator mass spectrometry at the Purdue Rare Isotope Measurement (PRIME) Laboratory at Purdue University (Sharma et al., 2000). Native ^{9}Be was measured via ICP–OES for each sample upon the recommendations of Portenga et al. (2015). The total ^{9}Be, including native ^{9}Be, rather than just the ^{9}Be carrier, was then used to calculate the ^{10}Be concentrations for the dataset.

This method for quantifying erosion assumes that sediment storage at the rockwall slopes of each catchment is limited and that the transport of sediment from the rockwall to the medial moraine is rapid (von Blanckenburg, 2005). Ward and Anderson (2011) developed an analytical expression to quantify the accumulation of cosmogenic nuclides during the transport of sediment from the source rockwall to the medial moraine. They found that ^{10}Be accumulation during the burial, englacial transport and exhumation of sediment to the glacier surface was negligible in landscapes with denudation rates ≤ 1 mm/a. This model was implemented in our study because some of the records of erosion local to our investigated catchments, particularly in Uttarakhand, exceed this threshold (0.13–5.37 mm/a; Vance et al., 2003; Lupker et al., 2013; Scherler et al., 2014). Moreover, the glaciers of this study share similar glacier geometries, surface velocities and debris cover characteristics as to those described in the study by Ward and Anderson (2011). The modelled ^{10}Be accumulation during this transport was then subtracted from the total ^{10}Be sample concentration for each sample, before deriving the rockwall slope erosion rates (Supplementary Item 1; Table S1).

Rockwall slope erosion rates were calculated using Equation 1 and 2, which are described in detail by Lal (1991), Granger et al. (1996), Balco et al. (2008) and Dortch et al. (2011). A 2σ uncertainty ascribed to the AMS results was propagated through the erosion rate calculations. Berylium-10 production rates were calculated and corrected for topographic shielding for each rockwall slope using a combination of Delunel et al. (2010) and Dortch et al. (2011) codes in MATLAB R2017.a, an ASTER 30-m GDEM (16-m vertical resolution), a calibrated sea-level high-latitude ^{10}Be spallogenic production rate from Martin et al., (2017; http://calibration.ice-d.org/) and a ^{10}Be half-life of 1.387 Ma (Korschinek 2010, Chmeleff 2010). Ward and Anderson (2011) and others have demonstrated that muonic production of ^{10}Be within amalgamated supraglacial debris sourced from the rockwall is largely negligible, and for the process timescales of rockwall slope erosion can be omitted from our calculation scheme (Braucher et al., 2003; Akçar et al., 2014; Sarr et al., 2018).

Widespread avalanching and minimal snow retention on the steep rockwall slopes within our study areas reduces our concern about the effects of snow shielding on the erosion dataset. Scherler et al. (2014) estimated the impact of snow shielding on nuclide concentrations using remote sensing derived observations of snow cover duration and field-based measurements of annual daily snow depth in Uttarakhand, northern India. This data is unavailable for our entire study area. However, we have applied a 5.3% correction to our erosion rates; the mean correction value made by Scherler et al. (2014) for ten catchments in Uttarakhand with similar topographic and climatic characteristics to our study area. This correction does not change any broad trends in the erosion dataset. However, due to the ambiguity attached to these correction estimates and the variability in mean annual precipitation across our study area, we prefer to use the uncorrected erosion rates herein.

3.4. Statistical analysis of erosion dataset

We calculated the Pearson Correlation Coefficient values (*p*) between the ^{10}Be rockwall slope erosion rates and climatic, topographic, tectonic and geologic parameters. A *p*-value of <0.01 (at >99% confidence level) was applied. Each considered parameter has proven to influence rockwall and/or slope stability in other alpine regions (McColl, 2012; Krautblatter and Moore, 2014). Principle Component Analysis (PCA) was then used to identify and evaluate the possible controls of rockwall slope erosion in the NW Himalaya (The R Core Team, 2018; Supplementary Items 3, 4). This approach has been successfully applied in other studies to identify and evaluate the nature and magnitude of the environmental and landscape response to changes in climate (Edwards and Richardson 2004; Sagredo and Lowell, 2012; Seaby and Henderson, 2014). The topographic parameters included catchment and glacier area, mean catchment, rockwall and glacier slope, catchment 3-km-radius relief, mean catchment and snowline elevation and glacier aspect. Climatic variables included mean annual precipitation (weather stations [as referenced in Table 1] and TRMM [1998−2009]) and temperature (weather stations and CRU2.0 [as referenced in Table 1]), mean rockwall slope temperature and minimum catchment temperature. Catchment specific temperatures were calculated using an adiabatic lapse rate of 7°C/km and methods outlined in Orr et al. (2019). Additional variables, such as sample grain size and mean apatite fission track (AFT) cooling age (as referenced in Table 5), were also included within these analyses. The latter enables us to identify correlations between modern erosion rates and regional denudation histories on the million-year timescale. The Uttarkashi dataset (Kirti, Bhagirathi) was not included in these analyses because the rockwall slope erosion rates characterize an extensive basin system with numerous tributary catchments, rather than a single catchment as the remaining dataset does. This dataset is examined in more detail in the discussion section below. *P*-values were also calculated between rockwall slope erosion rates and catchment parameters for Ladakh and Lahul-Spiti as discrete regions (Supplementary Item 4). The other studied districts were not subject this regional analysis because the datasets are restricted to only one or two catchments.

4. Results

Catchment relief is subdued in the Ladakh region study areas in Jammu and Kashmir (0.7−1.0 km), despite the imposing, high-altitude mountain peaks and rockwalls (>5500 m asl) that mark the headwater limits of each catchment (Table 2). The mean rockwall slopes range between 26.3±12.4 and 35.2±15.5°. The topography of the Lahul-Spiti region in Himachal Pradesh is more severe than Ladakh, even with lower mean elevations (<4500 m asl); the investigated catchments are larger (13.9−44.9km^2), and have greater relief (1.2±0.3−1.8±0.5 km) and mean rockwall slopes (32.8±12.8−47.2±11.9°).

The ablation zone of the Lahul-Spiti and Kullu glaciers are partially to completely covered by debris, whereas in Ladakh, coverage is <30% of the glacier surfaces (Fig. 4; Table 3). Beryllium-10 sample concentrations for the Ladakh and Lahul-Spiti/Kullu catchments range from 6.0±0.7x10^4 to 260.0±12.5x10^4 at/g SiO_2 and 0.5±0.04x10^4 to 30.6±1.0x10^4 at/g SiO_2, respectively (Fig. 3; Table 4). On average, ~1% of the total ^{10}Be concentration of each sample was the result of ^{10}Be accumulation during transport from the source rockwall slopes to the medial moraine. For our study, this necessary correction to the final ^{10}Be concentrations had a negligible impact on the derived erosion rates (Table 4; Supplementary Item 1). Rates of rockwall slope erosion for the Ladakh region ranged between 0.02±0.004 and 1.0±0.2 mm/a, while rates in Lahul-Spiti/Kullu ranged from 0.2±0.02 to 7.5±1.0 mm/a (Figs. 3, 5; Table 4)

The catchment parameters with the most statistically significant relationship with rockwall slope erosion include mean rockwall slope, mean catchment and snowline elevation, mean annual precipitation, mean annual temperature and mean AFT cooling age (Table 5). For the district-specific analysis, the same parameters are strongly correlated with rockwall slope erosion in Ladakh (p= <0.01;

Supplementary Item 4). None of the parameters have a strong statistical correlation with the inferred erosion rates for the Lahul-Spiti district.

5. Discussion

Rockwall erosion rates vary by up to two orders of magnitude throughout the NW Himalaya (0.02±0.04–7.6±1.0 mm/a; Fig. 5). Considering the inherent complexities of periglacial-glacial environments, the application of cosmogenic nuclide analysis in these settings, and the range and variability in denudation recorded for this region (e.g., Vance et al., 2003; Scherler et al., 2014; Thiede and Ehlers, 2013), this is perhaps unsurprising. No relationship is apparent between ^{10}Be concentration and proximity of sample location to either a glacier margin or snout. Variability in ^{10}Be within the catchments is likely because the medial moraine sediment is poorly mixed and/or has a non-proportional sediment supply from the rockwall that is dominated by stochastic rockfall events (Small et al., 1997; Muzikar, 2008; Ward and Anderson, 2011).

The strong variability in physical settings of the catchments prevent any meaningful interpretations or comparisons between specific erosion rates. Moreover, time-averaged nuclide derived erosion rates come with large uncertainties when characterizing local areas (≤10^1 km^2), which has been shown to underestimate the true rates (Yanites et al., 2009; Willenbring et al., 2013; Sadler and Jerolmack, 2014). Instead, we focus on the broad trends of this rockwall slope erosion dataset for the NW Himalaya. Rockwall slope erosion decreases with distance north from the MCT; up to two orders of magnitude difference in erosion exist between Uttarakhand, Himachal Pradesh, Jammu and Kashmir and Baltistan (Fig. 5). The Urgos catchment in northern Lahul-Spiti slightly deviates from this trend with erosion rates of 3.2±0.5 and 7.6±1.0 mm/a, which are equivalent to those records in Kullu and southern Lahul-Spiti. The elevated rates may be attributed to increased annual precipitation in Miyar, which exceeds much of Lahul-Spiti (snowfall: 120−400 cm/a; Patel et al., 2018) and allows for more rapid erosion. Alternatively, the low ^{10}Be concentrations could be due to the input of fresh debris from the large, steep relief lateral moraines along Urgos glacier (Fig. 4e, f).

The applicable timescales of this time-averaged dataset, although varied (~0.1−24.6 ka), means that the erosion rates encompass recognized shifts in climate, sediment flux, glacier mass balance and seismicity, which themselves operate across various timescales (10^{1-6} years; Barnard et al., 2001; Finkel et al., 2008; Owen and Dortch, 2014; Scherler et al., 2015). Between ~0.02 and ~8 m of lateral rockwall slope erosion is possible for a single millennium in the NW Himalaya. When these rates are extrapolated for the whole Quaternary period, an estimated ~2 km of rockwall retreat is accomplished in the NW Himalaya, which are similar estimates to the Sierra Nevada in the Western USA (Brocklehurst and Whipple, 2002). The magnitude of rockwall slope erosion evident in the NW Himalaya not only demonstrates the importance of slope erosion through periglacial processes, specifically frost cracking in high-altitude alpine settings, but also the significance that localized erosion has for understanding wider landscape change (Small and Anderson 1998; Hales and Roering, 2005, 2007; Moore et al., 2009; Sanders et al., 2012, 2013). The rates of rockwall slope erosion reflect, in part, the pace of topographic change at the catchment headwaters.

The magnitude of erosion, particularly in the GHS-S, is sufficient to affect the strength of hillslope-glacier coupling, catchment sediment flux and contribute to topographic change such as the production of relief, the migration of catchment divides, and the reconfiguration of drainage basins (Oskin and Burbank, 2005; Naylor and Gabet, 2007; Heimsath and McGlynn, 2008; MacGregor et al., 2009 *ibid*). The rockwall slope erosion rates share a significant association with mean rockwall slope: the greater the mean rockwall slope, the more rapid the erosion (Fig. 6a, Table 5). This points to important feedbacks between these variables, where the rockwall slope angle and erosion rate limit one another. A tentative relationship can also be recognized between relief and rockwall slope erosion; where catchments with the high-altitude peaks (>5800 m asl), narrow ridgelines and high relief (>1.2±0.2 km), record the highest rates of erosion. Part of this is because catchments with rockwall slope erosion rates >1 mm/a have mean rockwall slopes that exceed the 35° threshold, above which slopes are unable to retain regolith, snow or ice (Gruber and Haerberli, 2007; Nagai et al., 2013). This means that rockfall and avalanching is pervasive. More extensive glacier debris cover in these catchments compared to those with slower erosion demonstrate that coupling between rockwall and glacier is enhanced in catchments with steep accumulation areas, and that slope is important in moderating hillslope debris flux (Regmi and Watanabe, 2009; Scherler et al., 2011; Table 3). Other studies also recognize the importance of slope in landscape change, some of which argue that slope gradients can be used to infer rates of background denudation (Portenga and Bierman, 2001; Finlayson et al., 2002; Burbank et al., 2003; Ouimet et al., 2009; Scherler et al., 2011, 2014).

Rates of rockwall slope erosion in Uttarkashi and Ladakh districts are either equivalent to, or exceed by up to one order of magnitude, the local catchment-wide erosion and exhumation rates (Fig. 6). Quaternary exhumation rates range between ~0.1 and 3 mm/a in the study areas (Thiede et al., 2004; Theide and Ehlers, 2013). Catchment-wide rates for the Lahul-Spiti and Kullu districts are unavailable because much of the region remains glaciated (Owen and Dortch, 2014). Orr et al. (2019) caution that comparing these erosion datasets can be problematic as they refer to landscape change through a variety of erosional processes and across various spatial and temporal scales. Nevertheless, the order of magnitude difference in these rates shows that erosion at catchment headwaters in the NW Himalaya largely outpace the entire drainage basins (Oskin and Burbank, 2005; Naylor and Gabet, 2007), and that erosion can vary significantly across short distances downstream (Scherler et al., 2014). Time-averaged rates for small areas such as catchment headwaters and rockwall slopes are sensitive to short-term local change, including single mass wasting events, and are therefore expected to record more rapid rates of erosion than a catchment-wide perspective (Yanites et al., 2009; Willenbring et al., 2013). The Karzok catchment in central Zanskar of Ladakh deviates from this trend as the rockwall slope erosion either equals or is slower than the catchment-wide erosion and exhumation rates (Fig. 6). The preservation and gradual reworking of landforms and sediment deposits that date to > 400 ka is likely affected by the low background denudation recorded in this region (Hedrick et al., 2011). A possible explanation is that sediment residence times exert a stronger control on the catchment-wide erosion signal in these ancient landscapes than the scale and various surface processes operating in the catchment area.

5.2. Controls of slope erosion

Considerable efforts have been made in recent years to define the parameters that control hillslope stability, and therefore determine the frequency and magnitude of mass wasting events (Matsuoka, 2001; Ballantyne, 2002; Hales and Roering, 2005; Regmi and Watanabe, 2009; Fischer et al., 2006, 2012; Sanders et al., 2012, 2013). The interactions between topography, climate, hydrology, geologic setting and cryosphere dynamics are shown to control rockfall activity. Of the catchment parameters that can be defined in the NW Himalaya, mean rockwall slope as already discussed, mean catchment and snowline elevation, mean annual precipitation, mean annual temperature, and mean AFT cooling ages show the strongest correlation with rockwall slope erosion rates (Figs. 6, 8; Table 5).

Catchments with the most rapid rockwall erosion have a greater proportion of the rockwall slope above the snowline than below, and larger glacier accumulation areas, than those with lower erosion rates. Aided by high gradient slopes that are set in part by erosion, field assessments and satellite imagery suggest that snow and ice entrained debris is either removed from the rapidly eroding rockwalls via avalanching or is largely absent. Evidence of avalanching underlines the importance of snow processes and cover, whether set by climatic conditions or surface uplift, in the transfer of debris from the rockwall to the glacier system (Scherler et al., 2011, 2014).

Estimated surface temperatures of the rockwalls are similar to those considered optimal for mechanical weathering processes (-8 to -3°C), e.g., freeze-thaw, frost cracking and frost wedging (Brozović et al., 1997; Matsuoka and Sakai, 1999; Matsuoka, 2001; Hewitt, 2002; Hales and Roering, 2005; MacGregor et al., 2009; Table 1). The medial moraine sediment characteristics are consistent with sediment from the supraglacial realm, which have detached from source slopes by periglacial weathering processes (Benn and Lehmkuhl, 2000; Schroder et al., 2000; Benn and Owen, 2002; Hambrey et al., 2008; Lukas et al., 2012; Orr et al., 2019; Table 3; Supplementary Item 5). Rates of periglacial erosion are likely further enhanced by seasonal and/or diurnal thermal variability in exposed bedrock surfaces of our investigated catchments, which is determined in part by the topographic steepness (Gruber and Haerberli, 2007; Fischer et al., 2012; Nagai et al., 2013; Haeberli et al., 2017). However, for high elevation catchments (> 4000 m asl) and/or rockwall slopes of our study area that lack an insulating layer of snow due to threshold slopes, bedrock surfaces can reach temperatures below -8 °C, which inhibit further mass wasting (Ward and Anderson, 2011). This is tentatively reflected in the relationship between temperature and rockwall slope erosion; the catchments with lower regional temperatures record slower erosion rates (Fig. 6c). The rockwall debris flux of each catchment is therefore likely influenced by the feedbacks between elevation, temperature and slope.

A strong positive relationship between ^{10}Be-derived rockwall slope erosion and mean annual precipitation supports the view that the distribution and magnitude of Himalayan erosion and denudation is partly a function of orographically focused monsoon rainfall (Bookhagen et al., 2005a; Theide et al., 2004; Bookhagen and Burbank, 2006; Gabet et al., 2006; Wulf et al., 2010; Dey et al., 2016; Figs. 6c, 7). The argument that precipitation provides a first-order control on the frequency and magnitude of mass wasting events in alpine settings is common (Hovius et al., 2000; Iverson, 2000; Dortch et al., 2009). Eppes and Keanini (2017) argue that the proficiency of mechanical weathering processes such as sub-critical cracking is climate-dependent, and specifically limited by moisture. Sources of moisture in alpine environments include snowfall, rainfall and melt water. Although rockwall slope erosion is certainly influenced by the availability of moisture and is sensitive to the microclimatic conditions of each catchment, its distribution throughout the NW Himalaya cannot be fully explained by precipitation. A five-fold decline in precipitation occurs between the first topographic high of the Lesser Himalaya (900±400 m asl) and the interior ranges of the orogen (Bookhagen et al., 2005a, b; Bookhagen and Burbank, 2006; Fig. 7). If precipitation were the primary control of rockwall slope erosion as we hypothesised, then we would expect to find that our maximum erosion rates coincide with maximum rainfall, and that a notable decline in these rates would be observed with distance north into the Greater Himalayan interior. However, our results show that this is not the case. Scherler et al (2014) make a similar observation, where the highest catchment-wide rates in Uttarakhand are also located north of the precipitation maxima. To further emphasize this point, there is an order of magnitude difference in the rockwall slope erosion rates between the GHS-N and the Tethyan Himalayan, yet a small decline in annual precipitation of < 300 mm.

Since the Late Miocene the steep orographic barrier of the Himalaya has restricted the northward advancement of moisture (Bookhagen et al., 2005a; Wulf et al., 2010), therefore preventing any subsequent major shift in the overall intensity or distribution of precipitation (Bookhagen et al., 2005a; Bookhagen and Burbank, 2010; Boos and Kuang, 2010; Thiede and Ehlers, 2013). The overall pattern in rockwall slope erosion throughout the NW Himalaya is therefore unlikely to be an artifact of a previous climatic regime, despite short-term fluctuations in monsoon strength during the Quaternary potentially affecting rockfall activity on the catchment scale (Thompson et al., 1997; Gupta et al., 2003; Fleitmann et al., 2003; Demske et al., 2009). One major concern in evaluating the role of climate in long-term landscape change is that the denudation records are averaged across million-year timescales and are therefore unable to account for the importance or variations in the Indian summer monsoon (Bookhagen et al., 2005a; Thiede and Ehlers, 2013). This study is able to show that erosion records that reflect landscape change on timescales that would be sensitive to fluctuations in monsoon strength (10^{2-5} years), i.e., rockwall slope and catchment-wide erosion, are not unilaterally controlled by precipitation.

The patterns in rockwall slope erosion rates are most closely associated with regional AFT cooling ages (Figs. 6d, 7; Table 5). Much attention has been paid to understanding the patterns of cooling ages and exhumations rates in the Himalaya, and the feedbacks between tectonics and climate that are responsible for the distribution and intensity of Himalayan denudation across million-year timescales (Schelling and Arita, 1991; Srivastava and Mitra, 1994; Thiede and Ehlers, 2013). Many studies have argued that denudation is primarily governed by climate; orographic precipitation causes rapid erosion and exhumation along the Himalayan front and Lesser Himalaya (Zeitler et al., 2001; Thiede et al., 2004; Grujic et al., 2006; Biswas et al., 2007; Sharma et al., 2017; Kumar et al., 2018). However, young AFT ages (<10 Ma) and rapid rates of exhumation throughout the Lesser Himalaya and GHS-S instead reflect a close interaction between tectonics, denudation and monsoon-enhanced erosion, rather than just the latter (e.g., Wobus et al., 2003; Thiede et al., 2004; Vannay et al., 2004). Coupling between climate and tectonics becomes less evident farther into the Greater Himalayan interior; while the GHS-N becomes progressively more arid, the AFT ages remain <17 Ma and exhumation rates > 5mm/a (Thiede and

Ehlers, 2013; Schlup et al., 2003; Fig. 7). The pattern in AFT ages and inferred exhumation histories for the NW Himalaya, like our rockwall slope erosion dataset, cannot therefore be fully explained by precipitation. Instead, there is the argument that the patterns of Himalayan denudation are instead a function of tectonically controlled rock uplift; the result of crustal wedge deformation from the Indo-Eurasian collision and the flat-ramp-flat geometry of the Main Himalayan Thrust (e.g. Burbank et al., 2003; Bollinger et al., 2006; Herman et al., 2010; Robert et al., 2011; Godard et al., 2014). The lateral and vertical transport of rock over the ramp since the late Miocene has resulted in rapid and continuous exhumation, and the generation of steep topographic relief (Cattin and Avouac, 2000; Godard et al., 2004; Lavé and Avouac, 2000, 2001). Young AFT cooling ages and rapid rates of exhumation are therefore focused throughout the Lesser Himalaya and GHS-S (Fig. 7). This is consistent with our patterns in rockwall slope erosion, therefore indicating that tectonically driven rock uplift throughout the NW Himalaya is likely to provide a major control on patterns of denudation since the late Paleogene, and also influence late Quaternary records of erosion (Scherler et al., 2011, 2014). The climatic parameters of precipitation and temperature are therefore likely secondary controls. Moreover, this confirms that similar casual relationships between rock uplift and erosion operate throughout the glaciated and non-glaciated regions of the NW Himalaya.

PCA indicate that ~68% of the variance observed in rockwall slope erosion rates in the NW Himalaya can be explained by the six parameters discussed above (mean rockwall slope, mean catchment and snowline elevation, mean annual precipitation, mean annual temperature and mean AFT age; Fig. 8). To explain the remaining variance, other parameters must be considered. Rockwall lithology, rock strength and mass quality, and jointing and structure for example, affect the thresholds for mass wasting and have been shown to govern hillslope debris flux and rates of erosion (Hallet et al., 1991; Augustinus, 1995; Anderson, 1998; Hales and Roering, 2005; MacGregor et al., 2009; Fischer et al., 2010). Rockfall activity in the investigated catchments is therefore likely affected by the erodibility of the rockwall and the periglacial processes acting upon it (Heimsath and McGlynn, 2008; Eppes and Keanini, 2017; Moon et al., 2017). The significance of this parameter in the patterns of rockwall slope erosion on the regional scale is however less clear. Previous work has argued that the difference in rock strength between the crystalline sequences of the Lesser and Greater Himalaya is negligible, and has little influence upon the denudation histories of the orogen (Burbank et al., 2003; Scherler et al., 2011, 2014).

Studies throughout High Asia have shown that geomorphic change, specifically mass wasting events, are closely associated with neotectonism including earthquakes and/or persistent microseismicity (Hovius et al., 2000; Menunier et al., 2008; Dortch et al., 2009; Lupker et al., 2012). For example, earthquakes in Uttarakhand such as the 1991 Uttarkashi (M 6.1; Valdiya, 1991; Bali et al., 2003) and 1999 Chamoli (M 6.6; Rajendran et al., 2000) events are found to trigger mass redistribution on a scale that affects short term erosion rates (Bali et al., 2003; Scherler et al., 2014). The frequency of rockfall events and therefore rates of rockwall slope erosion in our catchments is therefore likely to be influenced in part by local tectonic activity.

A further candidate for rockwall slope erosion control is glaciation and glacial erosion; vertical incision and the debuttressing of slopes can lead to enhanced slope instability and failure (Naylor and Gabet, 2007; Heimsath and McGlynn, 2008; MacGregor et al., 2009; Fischer et al., 2010). Large, erosive temperate glaciers occupy catchments with rapid rates of rockwall slope erosion, while slower rates are from catchments with less erosive, sub-polar glaciers (Owen and Dortch, 2014). Past retreat and expansion of glacier ice may also have contributed to the evolution of the rockwalls; the downwasting of ice may encourage the unloading of slope debris, while a greater glacier volume may see an increase in glacial erosion processes acting upon the slope (Fischer et al., 2006; 2010, 2012; Herman et al., 2017). For example, the Hamtah glacier in Lahul-Spiti has retreated ~90 m in the last ~200 years, during which time rockwall slope erosion rates have exceeded 3 mm/a (Tables 1, 4; Fig. 5; Saha et al., 2018). Rockwall slope erosion is therefore likely a critical component of the catchment headwater's response to shifts in glaciation, where the redistribution of stress from changing ice extents likely decreases slope stability (McColl, 2012; Gallach et al., 2018). These processes are part of a complex feedback; supraglacial debris cover above a critical thickness can act to insulate glacier ice, while a thinner layer can enhance melt by decreasing the albedo of the glacier surface (Ostrem, 1959). The erosion and delivery of debris from the rockwall slopes to the glacier can therefore affect surface melt rates, the mass balance of the glacier, and more broadly the glacier's sensitivity to environmental change (Anderson et al., 2011; Immerzeel et al., 2014; Gibson et al., 2017). To complete the feedback; glaciation and glacier dynamics regulate glacial erosion processes and local climatic conditions, which are both parameters that are shown to affect rockwall slope stability (Heimsath and McGlynn, 2008; McColl, 2012; Anderson et al., 2018).

Rather than a single control, we have confirmed the initial conclusions of Orr et al., (2019) by finding that rockwall slope erosion is instead more likely the result of longstanding feedbacks between climate, tectonics, topography and surface processes within each catchment. The relative importance of these various parameters in driving rockwall slope erosion will likely vary across space and time. For example, the recognised relationship between rockwall erosion and slope for the NW Himalaya does not extend to the Lahul-Spiti district, if it were considered a discrete region. Only in Ladakh does the steepest catchment and rockwall slopes record the most rapid rates of erosion. This does not mean that rockwall erosion is unaffected by slope in Lahul-Spiti, however it does suggest that other parameters are also necessary to explain the patterns of erosion. No parameters discussed show a strong correlation with rockwall slope erosion in Lahul-Spiti (Supplementary Item 4). Explanations for this might be that the erosion of rockwalls is influenced by a combination of parameters which together affect rockwall slope stability, or that this erosion is sensitive to undefined parameters such as glaciation and glacial processes (e.g., glacier type and dynamics, glacial erosion). Deciphering erosion controls may not be possible due to the inherent complexities of glaciated catchments in this district (e.g., rapid cycles of glacier retreat/advance, shifts in fluvial/meltwater discharge and erosion, variability in glacial/non-glacial sediment source-sink sedimentation; Bookhagen et al., 2006; Adams et al., 2009; Bookhagen and Burbank, 2010; Saha et al., 2018;).

Controls of rockwall slope erosion may also be difficult to constrain across various temporal and spatial scales because for some catchments, once a threshold for a particular parameter has been met (e.g., moisture availability), rockwall slope erosion becomes predominantly limited by it. During a period of enhanced rainfall or monsoon along the Himalayan front for example (Bookhagen et al., 2005b; Clift et al., 2008), catchments with strongly contrasting geology and/or topography may display similar rockfall activity. In this case, the magnitude of precipitation is sufficient to govern rockwall slope stability and override any resistance to mass wasting (e.g., strong, non-erosive rock type or shallow, low gradient slopes). When averaged over time, these physically contrasting catchments will share a similar record of rockwall slope erosion. This may offer an explanation for why single high-magnitude events such as these, are viewed to be responsible for a significant proportion of the total landscape change in mountain environments (Hasnain 1996; Kirchner et al., 2001; Craddock et al., 2007; Wulf et al., 2010). To tackle some of these outstanding questions, rockwall slope erosion controls should be evaluated for glaciated catchments with comprehensive geologic and climatic data and well constrained records of glacial history, topographic change and mass wasting.

We suggest that rockwall slope erosion is largely influenced by catchment-specific conditions that vary over temporal and spatial scales. However, our study is able to demonstrate that the broad spatial patterns in rockwall erosion follow long-term trends in denudation throughout the NW Himalaya, and is therefore broadly controlled by tectonically driven rock uplift. The climatic parameters of precipitation and temperature are therefore likely secondary controls. This suggests that periglacial rockfall processes are part of the erosional response to structural change throughout the Himalayan-Tibetan orogen, and play a significant role within topographic change at catchment headwaters and the mass balance of the orogen. Identifying a more significant tectonic control to landscape change than climate is common; work in the wider Himalaya and the northern Bolivian Andes suggest that denudation patterns do not follow gradients in precipitation (Burbank et al., 2003; Gasparini and Whipple, 2014; Godard et al., 2014; Scherler et al., 2014).

6. Conclusion

Rates of rockwall slope erosion are defined for 12 catchments in northern India, NW Himalaya and range between 0.02±0.04 and 7.6±1.0 mm/a. Rockwall slope erosion largely outpaces local catchment-wide erosion and exhumation, and is sufficient to affect catchment sediment flux, glacier dynamics and topographic change, such as the production of relief, the migration of catchment divides and the reconfiguration of drainage basins.

Erosion rates become progressively lower with distance north from the MCT; up to two orders of magnitude difference in erosion rates are observed between Uttarkashi, Kullu, Lahul-Spiti, and Ladakh and Shigar. Rather than a single control, rockwall slope erosion on a catchment-by-catchment basis is largely influenced by longstanding feedbacks between climate, tectonics, topography and surface processes. The relative roles of these parameters are likely to vary over various spatial and temporal scales.

Our study demonstrates that like records of denudation in the NW Himalaya, the broad trend in rockwall slope erosion cannot be fully explained by the distribution of precipitation. Instead rockwall slope erosion can be considered part of the erosional response to tectonically driven uplift, the product of Indo-Eurasian convergence and structural geology. The distribution and magnitude of erosion applicable to geomorphic (10^{2-5} years) and geologic (10^6 years) timescales in the NW Himalaya therefore suggests that tectonics, rather than climate, provide a first-order control on landscape evolution. Our study also demonstrates the importance of lateral rockwall slope erosion via periglacial processes in helping set the pace of topographic change at catchment headwaters of high altitude and high relief mountain ranges, and the significance that localized erosion has for understanding wider landscape change.

Acknowledgments

Data supporting the conclusions is in the process of being archived with the GFZ Data Services repository (http://dataservices.gfz-potsdam.de/portal/). In the interim and for review purposes, this data can be accessed in Supplementary Items 1–5. The authors acknowledge that this manuscript will not be published until the data is completely archived and publicly available.

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Tables

Table 1. Details of the investigated catchments.

Table 2. Catchment and glacier characteristics of the investigated catchments (uncertainties are expressed to 2σ).

Table 3. Medial moraine morphology and sediment descriptions.

Table 4. Medial moraine sample details, ^{10}Be concentrations and inferred rockwall slope erosion rates for the investigated catchments.

Table. 5. Pearson's Correlation Coefficient values (*p*) between ^{10}Be rockwall slope erosion rates and catchment parameters.

Figures

Fig. 1. Schematic diagram of glaciated catchment with primary debris transport pathways (ELA: equilibrium-line altitude). The rockwall is defined for this study as the headwater bedrock slopes above the ELA of each glacier. This study focuses on erosion via periglacial processes only. The medial moraines and supraglacial debris are revealed below the ELA (gray shading).

Fig. 2. Overview of the study area in the NW Himalaya. a) Study area location (black polygon) is outlined on a 5-km-radius relief map with swath polygons [bold polygons S1, 2, 4, 5 are referred to in Fig. 7] modified from Bookhagen and Burbank (2006). 3-km-radius relief dataset used in following analyses. b) ASTER GDEM of study area (see a) with investigated regions and districts outlined. c) Hillshade map of the study area is overlain by mean annual precipitation (TRMM 2B31; Bookhagen and Burbank, 2006). White circles: location of catchments of this study. Gray circles: location of published rockwall slope erosion rate studies: Baltoro glacier system (Seong et al., 2009), Chhota Shigri (Scherler and Egholm 2017), Bhagirathi glacier system (Orr et al., 2019). Major faults from Hodges (2000) and Schlup et al. (2003). KF- Karakoram Fault, SSZ- Shyok Suture Zone, ITSZ - Indus-Tsangpo

Suture Zone STD- South Tibetan Detachment, MCT- Main Central Thrust, MBT- Main Boundary Thrust, MFT- Main Frontal Thrust, MHT- Main Himalayan Thrust. Inset: simplified structure of the NW Himalaya, modified from Searle et al. (2011) and Schlup et al. (2011).

Fig. 3. Geomorphic maps of the study areas including sample ^{10}Be concentrations and rockwall slope erosion rates. 1: Catchment ridgeline, 2: 100-m-contour lines.

Fig. 4. Views of medial moraines and sampling locations for three investigated catchments (white and black dashed lines outline medial moraine ridges). a) Beas Kund medial moraine, b) Sampling of G_{Bea1} in Beas Kund, c) Chhota Shigri medial moraine, d) Sampling of G_{Cht5} of Chhota Shigri, e) Urgos medial moraine, f) Sampling of G_{Urg2}.

Fig. 5. Sample ^{10}Be concentrations (a) and rockwall slope erosion rates (b) for the NW Himalaya. Uttarkashi (Bhagirathi glacier system) and Shigar (Baltoro glacier system) datasets are from Orr et al. (2019) and Seong et al. (2009), respectively.

Fig. 6. Rockwall slope erosion rates and catchment parameters. a) Mean rockwall slope (black points) and 3km-radius relief (red triangles). b) Mean elevation (black points) and snowline elevation (blue triangles). c) Mean annual precipitation (black points) and mean annual temperature (green triangles). d) Mean AFT cooling ages (black points). PL: Power Law function.

Fig. 7. Erosion, relief and precipitation of the NW Himalaya with distance from the MFT (datasets from Bookhagen and Burbank 2010). Swath locations outlined in Fig. 1a (LH- Lesser Himalaya; GHS-S/N- Greater Himalayan sequence South/North; TH- Tethyan Himalaya; THD- Tethyan Himalaya Dome). Exhumation[1]: Exhumation rates (use erosion rate y-axis) are inferred from AFT cooling ages as referenced below, an AFT cooling temperature of 120°C, and a geothermal gradient of 25°C/km. a) Swath 1 (S1). Rockwall slope erosion: *this study*; catchment-wide erosion: Dortch et al. (2011a), Dietsch et al. (2015); AFT cooling ages: Kristein et al. (2006, 2009). b) Swath 2. Rockwall slope erosion: *this study*, Scherler and Egholm (2017); Kristein et al. (2006, 2009). b) Swath 2 (S2). Rockwall slope erosion: *this study*, Scherler and Egholm (2017); AFT cooling ages: Schlup et al. (2003, 2011), Thiede et al. (2006), Walia et al. (2008). c) Swath 4 (S4). Catchment-wide erosion: Scherler et al. (2014); AFT cooling ages: Jain et al. (2000), Thiede et al. (2004, 2005, 2009), Vannay et al. (2004). d) Swath 5 (S5). Rockwall slope erosion: Orr et al. (2019); catchment-wide erosion: Vance et al. (2003), Lupker et al. (2012); AFT cooling ages: Sorkhabi et al. (1996), Searle et al. (1999), Thiede et al. (2009).

Fig. 8. PC1/PC2 plot for the catchment parameters that contribute to the distribution and magnitude of the rockwall slope erosion. Parameters with strongest correlation with erosion are labelled. Proportion of variance: PC1 (0.68), PC2 (0.17), PC3 (0.07), PC4 (0.04).

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