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Recycling of Pleistocene valley fills dominates 135 ka of sediment flux, upper Indus River

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Recycling of Pleistocene valley fills dominates 135 ka of sediment flux, upper Indus River

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Abstract

Rivers draining the semiarid Transhimalayan Ranges at the western Tibetan Plateau margin underwent alternating phases of massive valley infill and incision in Pleistocene times. The effects of these cut-and-fill cycles on millennial sediment fluxes have remained largely elusive. We investigate the timing and geomorphic consequences of headward incision of the Zaskar River, a tributary to the Indus, which taps the >250-m thick More Plains valley fill that currently plugs the endorheic high-altitude basins of Tso Kar and Tso Moriri. In situ ¹⁰Be exposure dating and topographic analyses show that a phase of valley infill gave way to net dissection and the NW Himalaya's first directly dated stream capture in late Marine Isotope Stage (MIS) 6, ~135 ka ago. Headwaters of the Indus are currently capturing headwaters of the Sutlej, and rivers have eroded >14.7 km³ of sediment from the Zaskar headwaters since, mobilising an equivalent of ~8% of the Indus' contemporary sediment storage volume from only 0.3% of its catchment area. The resulting specific sediment yields are among the rarely available rates averaged over the 10⁵-yr timescale, and surpass ¹⁰Be-derived

denudation rates from neighbouring catchments three- to tenfold. We conclude that recycling of Pleistocene valley fills has fed Transhimalayan headwaters with more sediment than liberated by catchment denudation, at least since the last glacial cycle began. This protracted release of sediment from thick Pleistocene valley fills might bias estimates of current sediment loads and long-term catchment denudation.

Keywords

Transhimalaya; Zaskar; Indus; valley fill; drainage capture; in-situ cosmogenic ^{10}Be

Highlights

- We directly date one of Himalaya's youngest drainage captures at the end of MIS 6
- Headwaters of the Indus are currently capturing headwaters of the Sutlej
- 10^5 -yr fluvial sediment flux of up to $\sim 270 \text{ t km}^{-2} \text{ yr}^{-1}$ from Pleistocene valley fills
- Up to 8% of Indus' sediment storage volume eroded from 0.3% of catchment area
- Recycled sediment yields are up to ten times higher than catchment denudation

1. Introduction

The Indus is amongst Earth's largest river systems in terms of discharge, catchment area, and sediment load (Milliman and Meade, 1983; Clift, 2002; Milliman and Farnsworth, 2011). It drains large parts of the western Tibetan Plateau margin, Transhimalayan ranges, Karakoram, and western Himalayan mountain ranges (Inam et al., 2008), and may be as old as the early uplift of the Tibetan Plateau (Rowley and Currie, 2006; Decelles et al., 2007; C.

Wang et al., 2008). Sedimentary analyses of the submarine Indus fan indicate that the river has not changed very much since ~45 Ma when it was already flowing westward through the Indus-Tsangpo Suture Zone (Clift et al. 2001; Clift, 2002) where the Eurasian and Indian continental plates meet. Others proposed that the Indus established its course much later (Sinclair and Jaffey, 2001). During its lifetime the Indus has tapped various sources of sediments, partly by headward incision into the Tibetan Plateau interior, partly by exploiting tectonic structures (Clift and Blusztajn, 2005; Garzanti et al., 2005).

Various mechanisms influence how rivers incise into the Tibetan Plateau over millennial to geological timescales (Bendick and Bilham, 2001; Lavé and Avouac, 2001; Sobel, 2003; Ouimet et al., 2007; Korup et al., 2010). Aridity can slow down discharge-driven fluvial incision, preserve plateau remnants, and create long-lived internally drained river systems (Sobel, 2003). Conversely, headward river incision aids plateau dissection through drainage capture. Yet systematic analysis and dating of drainage capture events are rare, especially along the western margin of the Tibetan Plateau and the Transhimalayan upper Indus basin. There, sedimentary basin-fills up to several hundred metres thick smother an alpine landscape, offering a stark contrast to stretches of deeply dissected bedrock gorges. We hypothesize that paired high-level terraces in the upper Zaskar River, a major tributary (~15,000 km²) of the Indus, mark the change from net valley infilling to dissection. Dating those terraces should allow comparing the resulting long-term fluvial sediment yields with catchment-wide denudation rates based on ¹⁰Be concentrations in river sands accumulated over tens of millennia. Our objective is thus to constrain when and how much of the large valley fills in upper Zaskar headwaters were eroded, and to clarify the role of this erosion for sediment-flux estimates in this major Transhimalayan river.

2. Study area

The Indus River flows through deep bedrock gorges and broad braidplains upstream of the Nanga Parbat syntaxis, an area of rapid exhumation and bedrock incision (Burbank et al., 1996; Zeitler et al., 2001; Garzanti et al., 2005). Denudation rates along the Indus drop from $>1000 \text{ mm ka}^{-1}$ near the syntaxis (Burbank et al., 1996) to $\sim 10 \text{ mm ka}^{-1}$ at the western Tibetan Plateau margin (Munack et al., 2014). Denudation rates estimated from cosmogenic ^{10}Be inventories are low despite the steep alpine topography (Dortch et al., 2011; Dietsch et al., 2014; Munack et al., 2014), so that the Transhimalayan ranges of Ladakh and Zaskar host some of the oldest glacial landforms in the Himalaya-Tibet orogen (Owen et al., 2006; Hedrick et al., 2011). Massive staircases of river-derived fill terraces (Fort et al., 1989; Clift and Giosan, 2014) located up to 400 m above present river levels, together with stacks of lake sediments and local landslide and fan deposits (Hewitt, 2002; Phartiyal et al., 2005; 2013), testify to major alternating cut-and-fill cycles in the upper Indus catchment (Blöthe et al., 2014). Some of these prominent sediment bodies are between 100 and 530 ka old (Owen et al., 2006; Blöthe et al., 2014; Scherler et al., 2014), and demonstrate the longevity of valley fills in the rain shadow of the Higher Himalayas. How these large valley fills affect sediment flux and drainage patterns along the western Tibetan Plateau margin (Blöthe and Korup, 2013; Clift and Giosan, 2014) remains to be resolved in more detail, though.

To elucidate the response of major rivers to major cut-and-fill cycles we investigate a large valley fill of the Rupshu Plateau (Fig. 1) where the soil mantled, gently sloping, and slowly denuding western Tibetan Plateau margin grades into the steep and rugged Transhimalayan ranges (Munack et al., 2014). We focus on the dissected and $>250\text{-m}$ thick valley fill of the More Plains, which is part of an extensive aggradation surface in the headwaters of the Zaskar River (Fig. 1). Terraced remnants of this valley fill (Fig. 2) occur

throughout the Sumka (298 km²) and Tozai catchments (475 km², Fig. 1) that line the topographically inconspicuous drainage divide between the Indus and the Sutlej Rivers (Fig. 3) that drain the mountain belt to the northwest and southeast, respectively. At about 4765 m asl, the gently sloping surface of the More Plains features abandoned river channels pointing north into the Zara River.

The geology of our study area (Fig. 3) is dominated by sedimentary shales, sandstones, and siltstones of the pre-collisional Kurgiakh Formation (Garzanti et al., 1986; Steck, 2003, Fig. 3, #28), marls and limestones of the Lilang group (Fig. 3, #22), and intrusive crustal granites (Fig. 3, #26) and local gabbros (Fuchs and Linner, 1996; Steck, 2003; de Sigoyer et al., 2004), forming much of the Tso Moriri and Tso Kar lake basins. Siltstones of the Lamayuru Formation (Fig. 3, #14) and Lamayuru Flysch of the Neotethyan slope underlie the southern Tso Moriri basin (Frank et al., 1977; Steck, 2003). Flowing along the southern slopes of the pre-collisional granodioritic Ladakh Batholith and cutting through some ophiolitic mélangé (Steck, 2003, Fig. 3, #9), the Indus River bounds the study area in the north. Dolomitic limestones of the Kioto Formation dominate the western parts of the Tozai catchment (Hayden, 1904; Steck, 2003; Fig. 3, #20).

3. Methods

To assess late Quaternary valley filling and dissection in the upper Zaskar near the More Plains, we combine in-situ produced cosmogenic ¹⁰Be analyses in river sediment, amalgamated surface and depth-profile samples, with topographic analyses and field observations.

3.1. Cosmogenic ¹⁰Be analyses

We collected sediment samples from five perennial rivers draining quartz-bearing lithologies (catchment areas ranging between 4 and 270 km²) to derive catchment-wide denudation

rates from concentrations of cosmogenic ^{10}Be in these samples (Figs. 1, 3, Table 1). We also collected two amalgamated grab samples for cosmogenic ^{10}Be exposure dating of terrace treads flanking the deeply incised Sumka River (PTOP and DP_Surf; Figs. 1, 2, 4, Table 2), each consisting of 35 well-rounded quartz-bearing pebbles (*b*-axes of 2-6 cm). At the location of DP_Surf, we further collected samples from a gravel pit from depths of 20 to 160 cm at 20-cm intervals (DP_020 to DP_160; Fig. 4; Table 2). For both the PTOP and DP_Surf sampling locations, which are ~3 km apart and at similar elevations, we recorded the topographic shielding and coordinates with a handheld GPS receiver with <10 m horizontal, and <20 m vertical, accuracy.

We isolated the 125-500 μm -sized quartz-grain fraction from the samples and the crushed and sieved pebble aggregate (surfaces and depth profile) samples following standard procedures for in-situ cosmogenic ^{10}Be analysis (Kohl and Nishiizumi, 1992). We extracted beryllium using ion chromatography at the Australian Nuclear Science and Technology Organisation (ANSTO) following procedures described in Child et al. (2000). All samples were spiked with a ^{10}Be -free solution of ~370 μg ^9Be prepared from a beryl crystal solution at 1080 ($\pm$ 0.7%) ppm. Full procedural chemistry blanks were prepared from the same beryl and gave a final mean $^{10}\text{Be}/^9\text{Be}$ ratio of $5.15 \pm 1.34 \times 10^{-15}$ ($n=4$, mean of two procedural blanks). The $^{10}\text{Be}/^9\text{Be}$ ratios were measured at the ANSTO ANTARES accelerator (Fink and Smith, 2007) and were normalised to the 2007 KNSTD standard KN-5-2 with a nominal $^{10}\text{Be}/^9\text{Be}$ ratio of $8,558 \times 10^{-15}$ (Nishiizumi et al., 2007). Blank corrected $^{10}\text{Be}/^9\text{Be}$ ratios ranged between 16.8 and 229.4×10^{-13} with analytical errors ranging from 1 to 2% (See Tables 1 and 2). Errors for the final ^{10}Be concentrations (atoms/g) were calculated by summing in quadrature the statistical error for the AMS measurement, 2% for reproducibility, and 1% for uncertainty in the Be spike concentration.

We calculate terrace surface ages from the depth-profile samples collected at DP-Surf with a Monte Carlo simulation based on the cosmogenic nuclide ingrowth equation of (Braucher et al., 2011):

$$C(x, t, \varepsilon) = \frac{nP e^{-x/\Lambda_n}}{\frac{\varepsilon}{\Lambda_n} + \lambda} \left[1 - e^{-t\left(\frac{\varepsilon}{\Lambda_n} + \lambda\right)} \right] + \frac{m_1 P e^{-x/\Lambda_{m_1}}}{\frac{\varepsilon}{\Lambda_{m_1}} + \lambda} \left[1 - e^{-t\left(\frac{\varepsilon}{\Lambda_{m_1}} + \lambda\right)} \right] + \frac{m_2 P e^{-x/\Lambda_{m_2}}}{\frac{\varepsilon}{\Lambda_{m_2}} + \lambda} \left[1 - e^{-t\left(\frac{\varepsilon}{\Lambda_{m_2}} + \lambda\right)} \right] \quad (1)$$

where C is the ^{10}Be concentration [at. g^{-1}], x is depth below surface [cm], t is the exposure time [yr], ε is the surface denudation rate [$\text{g cm}^{-2} \text{ yr}^{-1}$], P is the total surface production rate of ^{10}Be [at $\text{g}^{-1} \text{ yr}^{-1}$], λ is the radioactive decay constant ($\lambda = 4.998 \pm 0.043 \times 10^{-7}$), n , m_1 , m_2 are the relative contributions of neutrons, slow muons, and fast muons, respectively to the total ^{10}Be production ($n = 98.87\%$, $m_1 = 0.27\%$, $m_2 = 0.86\%$; Braucher et al., 2011), and Λ_n , Λ_{m_1} , Λ_{m_2} are their respective attenuation lengths ($\Lambda_n = 160 \text{ g cm}^{-2}$, $\Lambda_{m_1} = 1500 \text{ g cm}^{-2}$, $\Lambda_{m_2} = 4320 \text{ g cm}^{-2}$; Braucher et al., 2011). Note that in Eq. (1) ε reflects a surface denudation rate expressed in units of length per time [cm yr^{-1}] multiplied by density [g cm^{-3}]. We calculate the total ^{10}Be surface production rate (P) using the time-dependent altitude and latitude scaling scheme of Stone (2000), a ^{10}Be sea-level high-latitude (SLHL) spallation production rate of $3.94 \pm 0.20 \text{ at g}^{-1} \text{ yr}^{-1}$ (Heyman, 2014), and the muogenic production rates reported in (Braucher et al., 2011). Our ^{10}Be SLHL spallogenic production rate is statistically identical to those reported in (Borchers et al., 2016). We use the updated ^{10}Be half-life of $1.387 \pm 0.012 \text{ Ma}$ (Chmeleff et al., 2010; Korschinek et al., 2010) in all our calculations.

The imperceptibly sloping surfaces ($\sim 0.25^\circ$) from where we collected DP_Surf and PTOP have distinct frost polygons, and a thin veneer of aeolian silt, such that we assumed negligible erosion of the abandoned More Plains surface. Field observations and satellite

images confirm that the DP_Surf and PTOP sampling sites did not experience any colluvial deposition. Hence, we run our exposure model with a zero denudation rate. We allow density to vary as a free parameter between 1.9 to 2.3 g cm⁻³, and obtain a best-fit terrace surface age by minimising χ^2 (Bevington and Robinson, 2003):

$$\chi^2 = \sum_{i=1}^N \left[\frac{C_i - C(x, t, \varepsilon, \rho)}{2\sigma_i} \right]^2, \quad (2)$$

where C is the calculated ¹⁰Be concentration [at. g⁻¹] using Eq. (1), and C_i and σ_i are random samples from the normally distributed concentration measurements and their standard deviation. We estimate the 68% confidence intervals of our best-fit terrace surface age as $\chi_{min}^2 + 1$ (Bevington and Robinson, 2003). Our calculations neglect possible shielding effects on ¹⁰Be production rates by ice and snow. We believe that this simplification is justified, given that local glacial chronologies – particularly since MIS 5e – show only minor glacier advances that could have influenced ¹⁰Be concentrations (Taylor and Mitchell, 2000; Owen et al., 2006; Hedrick et al., 2011); the high-altitude desert of Zaskar also rarely favours thick snow cover (Burbank and Fort, 1985).

We derived basin-wide production rates from a 30m SRTM DEM (available at <http://earthexplorer.usgs.gov>) of the study area using the same ¹⁰Be SLHL production rates and scaling scheme as above. We correct for topographic shielding following Codilean (2006) and use Eq. (1) to calculate catchment-wide denudation rates.

3.2. Eroded sediment volume calculations

To estimate the volume of sediment eroded from the valley fills in the upper Zaskar, we used two different methods. First, we emplaced a hypothetical dam on the 30-m SRTM DEM of the study area and calculated the backfill volume using standard GIS fill-routines (method #1), resulting in a sediment wedge with a horizontal surface (Fig. 1, 5). Second, we used the

elevation of terrace remnants that flank the modern rivers to reconstruct the former valley floor of the Sumka and Tozai valleys (method #2). Based on our 30-m DEM we measured 2.5-km-wide sections spaced 500 m apart and normal to the mean former drainage direction of the valleys, which we estimated from satellite images (Fig. 5). For each cross-section, we extracted the raster cell values from the SRTM DEM with local slopes $<15^\circ$, which we took as an upper threshold for terrace-surface slopes. We used a nonlinear median quantile regression (Koenker, 2005) to reconstruct the concave-upward former valley bottom by connecting the trends of the terrace treads:

$$E_t = a * D^2 + b * D + c \quad (3)$$

where E_t is the elevation of terraces [m], D is distance upstream of an arbitrary starting point [m], a [m^{-2}] and b [1] are estimated coefficients and c [m] is the intercept. We used Eq. 3 to estimate the former elevation for each cell upstream of the dissected fill, and subtracted the SRTM DEM to calculate the eroded volume and resulting average sediment yields within 2σ uncertainty.

4. Results

Several km-scale risers of the More Plains along the Sumka River near the village of Pang offer $>250\text{-m}$ thick outcrops dotted by badland-like sedimentary pillars (Fig. 2C). North-facing risers are degraded by permafrost flow failures and slumps (Fig. 2D). The exposed sediments are surprisingly homogenous in terms of bedding, composition, clast size, and shape. Sub-rounded, very coarse pebbles in a yellow ochre, silt-cemented matrix dominate this valley fill (Fig. 2 E–H). Beds in the eastern part of the section are between $<0.5\text{ m}$ to $>5\text{ m}$ thick, show fluvial imbrication, dip between 1° and 3° towards the north to northeast, and alternate from limestone- to granite-dominated. A 3-m deep and 10-m long trench that we

dug into the western More Plains surface (star on Fig. 5) was devoid of granitic clasts, consistent with the non-granitic lithologies of the Tozai catchment (Steck, 2003), as opposed to the upper Sumka catchment, which features intrusive rocks (Fig. 3).

A Monte Carlo simulation yields a best-fit exposure age for the More Plains surface (DP_SURF in Figs. 1 and 2) of 135 ka (Fig. 4). The 1- σ confidence limits calculated as $\chi^2_{min} + 1$ (Bevington and Robinson, 2003) are 122.5 ka and 143 ka. The amalgamated surface grab sample at PTOP (Figs. 1 and 2) has a ^{10}Be concentration equal to that of DP_SURF within 1 σ uncertainty (Fig. 4). Considering the identical elevation, surface slope (Fig. 2), material properties, and bedding of the sedimentary units underlying the DP_SURF and PTOP samples, we infer similar sources, pathways, and transport times, and hence similar exposure histories with comparable amounts of inherited ^{10}Be . Thus, we assign the same surface age for PTOP as the one obtained for DP_SURF, proposing $135^{+8}_{-12.5}$ ka as the earliest time when net aggradation of the More Plains switched to net incision.

Depending on the approach to reconstruct the former surface of the More Plains upstream of Pang (Fig. 1), the volumes of sediment that we estimate were lost to fluvial erosion vary by nearly a factor of three. For the Sumka valley, the removed sediment volume is at least $1.23 \pm 0.2 \text{ km}^3$, if assuming a formerly flat valley floor (method #1, Fig. 5A). In contrast, the concave-up sediment wedge estimated from a polynomial fit to the terrace surfaces (method #2) has a volume of $3.59^{+1.09}_{-0.9} \text{ km}^3$ (Fig. 5A). For the Tozai valley, the volumes obtained for both methods are 3.97 ± 0.7 and $11.1^{+2.96}_{-2.49} \text{ km}^3$, respectively (Fig. 5B). Keeping in mind that method #1 consistently underpredicts valley-fill volumes because it ignores the slope of the former valley floor, we estimate that rivers removed $\sim 14.7 \text{ km}^3$ of sediment since they began incising into the More Plains. The corresponding average sediment yields are $170^{+51}_{-42} \text{ t km}^{-2} \text{ yr}^{-1}$ for the Sumka valley (Fig. 1, red polygon), and

$329^{+88}_{-74} \text{ t km}^{-2} \text{ yr}^{-1}$ for the Tozai valley (Fig. 1, yellow polygon); the average yield for both valleys is $267^{+74}_{-62} \text{ t km}^{-2} \text{ yr}^{-1}$ (Table 3).

The ^{10}Be concentrations of the river-sand samples (Table 1, Fig. 1) translate to catchment-averaged denudation rates between 13.1 ± 0.5 and $26.4 \pm 0.7 \text{ mm ka}^{-1}$ (Table 1). The associated averaging timescales (von Blanckenburg, 2005) are 45.9 to 22.7 ka (Table 1), and thus postdate major glacial periods judging from the regional chronology of Pleistocene glaciations (Burbank and Fort, 1985; Taylor and Mitchell, 2000; Owen et al., 2006; Hedrick et al., 2011). These denudation rates correspond to average sediment yields of $36\text{--}50 \text{ t km}^{-2} \text{ yr}^{-1}$ (Table 1), assuming that sediment storage on hillslopes is negligible.

5. Discussion

5.1. Valley infill, river capture, and dissection

Preserved channel patterns on the More Plains section consistently indicate that the Sumka and Tozai catchments formerly drained northward across the More Plains into the Zara River (Fig. 3), possibly into the Tso Kar catchment via what is today a major wind gap (Figs. 5, 7). Elevation profiles across the Sumka and Tozai valleys reveal a former aggradation surface graded to the N-NE dipping More Plains surface (Figs. 5, 7B; blue dots in profiles #2 and #3). Remnants of this formerly more expansive, but now dissected, valley floor rest >200 m above the lower Tozai and Sumka rivers (Fig. 2E–H). Mostly very coarse sub-rounded gravels in this fill point at fluvial and debris-flow sedimentation that operated over short distances and rapidly infilled a valley network. Although the Tozai and Sumka catchments share indistinct transfluence passes with the neighbouring Phirtse drainage in the Sutlej headwaters (Fig. 3; Thelakung La and southward), most of the valley fill would have been locally derived. We infer that the Tozai River tapped this formerly northward draining river system by incising into the More Plains valley fill shortly after 135 ka, essentially cutting

short the 80-km long horseshoe-shaped valley of the Zara River (Profile 'p-p₂' in Figs. 5, 7B-C, respectively). To our knowledge, this is the first directly dated river capture in the NW Himalayas near the drainage divide between the Indus and Sutlej Rivers.

The full extent of the former More Plains valley fill is vague, but would have reached the Sumka-Tozai confluence at least. A much larger valley fill extending to the present Zara-Tozai confluence is also conceivable (upstream of red dots in profile #3, Fig. 7B). If the Tozai valley drained north across the More Plains prior to the MIS 6-5e transition ('horseshoe' in Fig. 5), headward incision commencing here at ~135 ka would have forced the river to follow a steeper gradient towards the west ('shortcut' in Fig. 5). In any case, undercutting both down- and upstream isolated the More Plains between the Sumka ('p' in profile #8, Fig. 7) and the present upper Zara ('p₁' profile #1, Fig. 7) rivers; the latter is largely graded to the modern confluence with the Tozai River, whereas the former valley fill of the More Plains is not.

Hillslopes in the area around the Tozai-Sumka confluence (Fig. 6A) might be remains of the pre-capture bedrock topography. They appear to be adjusted to the More Plains surface, and rivers could have plausibly migrated headward here. Such a retreat towards the Pangyo basin east of the Tozai headwaters (Fig. 3), and along the lowest point of the 5090-m transfluence pass, is a possible scenario for future drainage changes. The longitudinal profiles of the Zaskar headwaters are much steeper than those of the Sutlej (Fig. 7A), and likely incising faster. Hence the Indus-Sutlej drainage is bound to migrate eastward here with the Indus headwaters gaining ground.

5.2. Catchment denudation and recycling of Pleistocene sediments

The ¹⁰Be-derived denudation rates of catchments fringing the endorheic lakes of Tso Kar and Tso Moriri (Fig. 1, Table 1) fit the picture of very slow landscape lowering in the upper Indus basin (Dortch et al., 2011; Munack et al., 2014). Denudation rates along the western shore of

Tso Moriri ($19\text{--}26 \text{ mm ka}^{-1}$, Fig. 1, Table 1) are statistically indistinguishable from those in the nearby Ladakh Batholith, but higher than in the Numah catchment (Fig. 1), which we expected to denude faster following river capture $\sim 135 \text{ ka}$ ago. Rock type might play a subordinate role, as the faster denuding catchments drain more competent rocks. The low denudation rates in Numah catchment might reflect incomplete headward incision of the Sumka River into its tributaries, whereas the higher rates in the Tsomo, Korzog, and Yan catchments might respond to active normal faulting along the western shoulder of the Tso Moriri half-graben (Hintersberger et al., 2010).

In any case, the net sediment flux from the recycled More Plain valley fill exceeds headwater denudation inferred from ^{10}Be by a factor of three to ten. This reworking distorts the long-term sediment flux in the upper Zaskar River over the past two glacial-interglacial cycles. Differing timescales are not an issue here, as the lower ^{10}Be -derived denudation rates concern roughly a third of the time since incision into the More Plains began (Table 1). If anything, we would expect higher rates for the shorter integration time instead, given less time to accommodate phases of negligible geomorphic activity.

5.3. Regional context and relevance

We provide an estimate of sediment yields averaged over $>100 \text{ ka}$, building a rare link between geomorphic and geological timescales for the NW Himalayas. Our estimate of 14.7 km^3 of eroded valley fills in the upper Zaskar suggests that an equivalent of as much as 8% of the contemporary sediment volume stored in the Indus catchment (Blöthe and Korup, 2013) may have come from as little as 0.3% of its area. Clift and Giosan (2014) estimated that $7 \text{ to } 22 \text{ km}^3$ of valley fills were lost from the entire Zaskar catchment since 10 ka . The removal of stored sediments would have fed average yields of $90\text{--}280 \text{ t km}^{-2} \text{ yr}^{-1}$ over that shorter period, but strikingly similar to our estimated yields from the Zaskar headwaters

since 135 ka. Although obtained over different timescales, both estimates underline the relevance of (Trans-)Himalayan valley fills, and especially those in headwater settings, for understanding major catchment dynamics (Table 3).

Our findings from the More Plains support previous work in the region on multiple 10^4 -yr cycles of infilling of alpine bedrock landscapes, and subsequent removal of this sediment, leaving former valley floors stranded as terraces several hundred metres high. At the Zaskar-Indus confluence, Blöthe et al. (2014) identified at least two postglacial phases of valley infilling before ~200 ka, and 50-20 ka, preserved in >150-m and 30- to 40-m high terraces, respectively (Fig. 8). Near the More Plains the geomorphic legacy of these major sedimentary cycles is particularly striking, smothering a substantial fraction of the alpine relief below hundreds of metres of debris. Few comparably extensive former valley floors line the rivers in Ladakh and Zaskar. The More Plains are much younger, for example, than a prominent 233-ka (MIS 7) terrace on the Indus near Nimu (Fig. 8D #4). Another large terrace at Agham in the Shyok River similarly records a turn from net aggradation to net incision at ~124 ka (MIS 6-5e; Fig. 8C #2; also recalculated). Yet only the paired terraces of the More Plains allow estimating the sediment volumes lost to fluvial erosion with confidence. Most of the former valley floors in the region were abandoned at the transition from glacials to interglacials, although major valleys such as the Shyok-Nubra and the Zaskar also aggraded during interglacials. River capture and subsequent sediment release from the More Plains section, i.e. the present Sumka (and Tozai) valleys, commenced during late MIS 6, but pinpointing climatic drivers remains problematic. Hedrick et al. (2011) proposed five glacial advances for the Korzog catchment, based on ^{10}Be exposure dates of moraine boulders (Figs. 1, 8). The oldest and youngest of these advances were within only ~15 km and ~1 km from modern glacier snouts at 310 ± 4.1 ka and 3.6 ± 1.1 ka, respectively

(Fig. 8E, 'K'). Other advances occurred around 80 ka (Fig. 8E, 'Z'; Taylor and Mitchell, 2000) and ~24 ka (Hedrick et al., 2011). Deposition on the More Plains ceased around ~135 ka at the latest. Lacking evidence of major glaciations in the surrounding Zaskar ranges between ~250 ka and ~110 ka (Fig. 8E) so far rules out major glacial sediment sources for the More Plains, such that the search for alternative sources remains open.

6. Conclusions

Pleistocene cycles of widespread aggradation and dissection shaped the mountain valleys of the western Tibetan plateau and the Transhimalayan Ranges. We report the first directly dated river capture in the headwaters of the Zaskar River close to the drainage divide of the Indus and the Sutlej, and propose an onset of river incision at the MIS 6-5e transition. This particular capture event might be part of a gradual eastward extension of the Indus at the expense of the Sutlej. Rivers lowered valley floors by >250 m, bypassed some 80 km of drainage route, and released up to 8% of the sediment storage of the Indus River from <0.3% of its area. The resulting average net sediment yields from these eroding basin fill deposits offer rare estimates on the 10^5 -yr scale, and exceed yields derived from ^{10}Be -derived basin-wide denudation. We infer that reworking of stored sediment in the semi-arid Transhimalayan Ranges may distort average sediment flux estimates over periods spanning two glacial-interglacial cycles, and thus complicate straightforward comparison between processes of erosion, sediment transfer, and deposition in this area dominated by low rates of long-term denudation.

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Tables

Table 1. Results of ^{10}Be analyses in river sediment samples and denudation rate calculations.

Sample ID	AMS ID	Drainage System	Lat ^(a) [°N]	Long ^(a) [°E]	Outlet ^(a) [m a.s.l.]	DEM Lat ^(b) [°N]	DEM Long ^(b) [°E]	Basin Area ^(c) [km ²]	Basin Slope ^(c,d) [degrees]	$^{10}\text{Be}/^9\text{Be}$ ^(e,f,g) [10 ⁻¹⁵]	Sample Mass [g quartz]	Carrier Mass [mg]	^{10}Be Conc. ^(g,h) [10 ³ at.g ⁻¹]	Denudation Rate ^(g,i) [mm ka ⁻¹]	Averaging Timescale ⁽ⁱ⁾ [ka]	SSY ^(k) [t.km ⁻² .yr ⁻¹]
<i>Internally drained</i>																
Korzog	B5605	Tso Moriri	32.96459	78.25183	4567	32.96754	78.25183	104	17.06 ± 9.66	3151 ± 37	26.459	0.373	2966 ± 75	20.80 ± 0.53	28.8	39.52 ± 1.02
Yan	B5607	Tso Moriri	33.09564	78.27039	4728	33.09564	78.27039	270	16.61 ± 8.67	3221 ± 43	30.450	0.376	2658 ± 69	25.46 ± 0.67	23.6	48.38 ± 1.28
Spang	B5609	Tso Kar	33.22702	77.98815	4700	33.22702	77.98815	133	18.60 ± 8.21	3073 ± 38	22.281	0.375	3453 ± 88	18.67 ± 0.49	32.1	35.48 ± 0.93
Tsomo	B5610	Tso Moriri	32.93547	78.26654	4606	32.93547	78.26654	4	18.55 ± 8.35	1676 ± 24	18.637	0.377	2265 ± 60	26.43 ± 0.71	22.7	50.22 ± 1.35
<i>Externally drained</i>																
Numah	B5608	Tozai River	33.12427	77.87433	4802	33.12427	77.87433	19	18.75 ± 6.18	3473 ± 85	18.308	0.376	4761 ± 158	13.06 ± 0.45	45.9	24.82 ± 0.85

a) Coordinates determined using hand held GPS and referenced to WGS84 Datum.

b) Coordinates refer to 30-m SRTM DEM pixel in channel nearest to sampling point in the field.

c) Calculated using 30-m SRTM DEM.

d) Uncertainties represent one standard deviation of the mean catchment slope.

e) $^{10}\text{Be}/^9\text{Be}$ ratios were normalised by SRM KN-5-2 (same as 07KNSTD) with a nominal ratio of $8,558 \times 10^{-15}$ (Nishiizumi et al., 2007).

f) Corrected for a mean (n=2) procedural blank of $5.15 \pm 1.34 \times 10^{-15}$.

g) Uncertainties expressed at 1 σ level.

h) Final uncertainties include: i) AMS analytical errors (the larger of the counting statistics errors and the one standard deviation of the repeated measurements), ii) standard reproducibility (2%) and iii) errors in 9Be-carrier concentration and quartz mass (1%), in quadrature.

i) Calculated using altitude/latitude scaling based on Stone (2000), topographic shielding based on Codilean (2006), and production mechanisms based on Braucher et al. (2011); See text for complete details and constants used.

j) Calculated using a depth of 60 cm (cf. von Blanckenburg, 2005).

k) Specific sediment yield (SSY) calculated with a bulk density of 1.9 g.cm^{-3} .

Table 2. Results of ^{10}Be analyses in amalgamated pebble surface and depth-profile samples.

Sample ID	AMS ID	Lat ^(a) [°N]	Long ^(a) [°E]	Elevation ^(a) [m a.s.l.]	Depth Below Surface ^(b) [cm]	$^{10}\text{Be}/^9\text{Be}$ ^(c,d,e) [10^{-15}]	Sample Mass [g quartz]	Carrier Mass [mg]	^{10}Be Conc. ^(e,f) [10^3 at.g^{-1}]
DP_SURF	B5593	33.14162	77.81162	4765	0	22938 ± 165	36.520	0.372	15636 ± 367
DP_20	B5594	33.14162	77.81162	4765	20	20827 ± 193	42.414	0.374	12268 ± 297
DP_40	B5595	33.14162	77.81162	4765	40	21408 ± 180	43.300	0.372	12299 ± 294
DP_60	B5596	33.14162	77.81162	4765	60	18826 ± 130	40.987	0.374	11490 ± 269
DP_80	B5597	33.14162	77.81162	4765	80	17213 ± 189	38.740	0.374	11113 ± 277
DP_100	B5598	33.14162	77.81162	4765	100	13631 ± 78	38.214	0.374	8921 ± 206
DP_120	B5599	33.14162	77.81162	4765	120	13079 ± 90	42.160	0.376	7786 ± 182
DP_140	B5600	33.14162	77.81162	4765	140	14163 ± 90	44.405	0.376	8005 ± 186
DP_160	B5601	33.14162	77.81162	4765	160	11323 ± 65	45.619	0.377	6250 ± 144
PTOP	B5602	33.11190	77.79445	4804	0	14343 ± 133	23.665	0.376	15228 ± 369

a) Coordinates determined using hand held GPS and referenced to WGS84 Datum.

b) Topographic shielding was measured in the field and yielded a value of 0.999 for both DP_SURF and PTOp.

c) $^{10}\text{Be}/^9\text{Be}$ ratios were normalised by SRM KN-5-2 (same as 07KNSTD) with a nominal ratio of $8,558 \times 10^{-15}$ (Nishiizumi et al., 2007).

d) Corrected for a mean (n=2) procedural blank of $5.15 \pm 1.34 \times 10^{-15}$.

e) Uncertainties expressed at 1 σ level.

f) Final uncertainties include: i) AMS analytical errors (the larger of the counting statistics errors and the one standard deviation of the repeated measurements), ii) standard reproducibility (2%) and iii) errors in ^9Be -carrier concentration and quartz mass (1%), in quadrature.

Table 3. Estimated erosion of valley-fill volumes, and specific sediment yields (SSY), averaging over 135 ka

Valley	Basin Area [km ²]	Min Volume [km ³]	Mean Volume [km ³]	Max Volume [km ³]	Min SSY [t km ⁻² yr ⁻¹]	Mean SSY [t km ⁻² yr ⁻¹]	Max SSY [t km ⁻² yr ⁻¹]
Sumka	298	2.70	3.59	4.68	127.49	169.50	220.79
Tozai	475	8.61	11.10	14.06	255.03	328.86	416.56
Sumka + Tozai	773	11.31	14.69	18.73	205.86	267.43	341.09
Zanskar*	15,000	7	-	22	90	-	280

Estimates from method #2 (methods and results section), based on 30-m SRTM data, and specific sediment yield (SSY) calculated with a bulk density of 1.9 g.cm⁻³. *Data from Clift and Giosan (2014), and averaged over the past 10 ka.

Figures

Fig. 1. Shaded relief of the western Tibetan Plateau margin and the Transhimalayan mountain ranges (inset) in the Zaskar River headwaters, represented by the Sumka and Tozai catchments, and the adjacent Tso Kar and Tso Moriri endorheic basins. Black points are sampling locations for catchment-averaged cosmogenic ^{10}Be with inferred denudation rates (mm ka^{-1} , Table 1). Contributing catchment areas are shaded blue. White circle locates PTOP ^{10}Be surface grab sample; white circle with black centre locates DP_SURF ^{10}Be surface grab sample and DP_020 to DP_160 depth profile (Fig. 4, Table 2). For geology see Fig. 3.

Fig. 2. Field photos of the More Plains. (A) Northward view of the former valley floor dipping to the north. (B) Downstream view of Sumka River cutting >250 m into the former valley fill. Red polygon (bordered by matching symbols in Fig. 1) is eroded cross section. Sedimentary pillars partly extend down to modern Sumka River level in midground. (C) More Plains section seen from the Sumka channel. Pillars consist mainly of sub-rounded very coarse

gravels (-5 to -6 ϕ) in cemented matrix. (D) View from terrace riser shown in (C) towards north-facing slope dotted by periglacial solifluction lobes. (E) Slope details of valley flank shown in (C). Surface consists of loose gravel dislodged from consolidated earth-pillars. (F) Earth-pillar detail with ~A5-sized field book for scale. (G) Detail of the uppermost ~30m of north-facing valley flank with ~2-m wide earth-pillar. (H) Close-up of valley-fill sediments shown in (G).

Fig. 3. Geological map of the Rupshu Plateau, featuring the More Plains (MP), the Sumka and Tozai catchments, and the internally drained Tso Kar and Tso Moriri lake basins (dotted with white outline). Phitse river sediments form an alluvial dam at the southern end of Tso

Moriri near the drainage divide between the Indus and Sutlej rivers. The dam is subject to episodic channel avulsions, thus partly draining into the Sutlej and shifting the draining divide (dashed with question marks). Black points are ^{10}Be -sampling locations with contributing catchment areas shown using black outlines. Paired brackets are transfluence passes showing elevation in metres. Arrows indicate flow directions. PS = Tso Kar, highest former shoreline detectable from satellite imagery is at 4635 m asl. Base map after Steck (2003); for formations #4, 5 (forearc basin) see Garzanti & Van Haver (1988) and Henderson et al. (2010), for formations #19-28 (Tethys Himalaya) see Garzanti et al. (1986), Gaetani and Garzanti (1991), and Sciunnach and Garzanti (2012).

Fig. 4. Cosmogenic ^{10}Be concentration of the More Plains aggradation surface as a function of depth. Empty circles are measured DP_ samples (Table 2) with 1σ uncertainties; filled circles are corresponding best-fit predictions. Inset: χ^2 plot shows best-fit result of exposure age ($\chi^2_{\min} = 13.8$) at ~ 135 ka assuming no surface erosion, and a bulk density of 1.9 g cm^{-3} .

Fig. 5. Reconstructed former valley floors of (A) Sumka and (B) Tozai rivers, Zanskar headwaters. Grey points are elevations of raster cells with slopes $<15^\circ$ extracted from 2500-m long valley cross sections with 500-m spacing along dashed lines T1 – MP (More Plains), and S1 – MP in (C). Dashed bold lines show former surfaces from second-order polynomial fit (method #2 in Section 3.2); dashed thin lines show hypothetical surfaces reconstructed from GIS-based cut-and-fill routine (method #1 in Section 3.2; rejected in 5.1) and estimated eroded volumes with 2σ errors. (C) Boundaries of Sumka (red), Tozai (yellow), and current endorheic Tso Kar (white) catchments. Horseshoe-shaped dashed line marks pre-135-ka drainage route across MP and down the Zara River. Star is position of 3 x 10 m (d x l) trench, devoid of granitic clasts; white circle is PTOP ^{10}Be surface grab sample; white circle with black centre is position of DP_SURF ^{10}Be surface grab sample and corresponding DP_020 to _160 depth profile (Figs. 1, 2, 4), TK = Tso Kar lake, TM = Tso Moriri lake (endorheic). Background image is pan-sharpened Landsat 7 ETM+ false-colour composite, RGB 432, September 2000 (<http://glcf.umd.edu/data/landsat/>).

Pre river capture

Post river capture

Fig. 6. Reconstructed former (A) and (B) modern topography of the More Plains and Sumka and Tozai catchments; black arrows indicate flow directions (Figs. 1, 3). Colour ramp repeats every 20 vertical metres, and highlights $<10^\circ$ -slopes of inferred former and modern More Plains surface; colour spacing scales inversely with slope gradient. White circle is PTOP ^{10}Be surface grab sample, white circle with black centre is position of DP_SURF ^{10}Be surface grab sample and corresponding DP_020 to _160 depth profile samples (Figs. 1, 2, 4).

Fig. 7. Longitudinal profiles of major rivers descending from the Indus-Sutlej drainage divide (profile colours applies to all panels). (A) River profiles projected back to back illustrate

differences in vertical drop per unit length between upper Indus and Sutlej rivers, with steeper profiles towards the Zaskar (Indus). (B) Longitudinal profiles with stacked slope/elevation pixels along 2-km wide swath cross-profile, respectively. 'p' and 'p₁' mark up- and downstream end of horseshoe-shaped Zara river segment that was cut short by stream capture dashed in (B) and (C), and offset in (C) for better graphical representation; 'p₂' marks recent Zara – Tozai confluence. (C) Location of rivers (MP = More Plains).

Fig. 8. Transhimalayan late Quaternary landscape evolution – synopsis of selected key geomorphic markers. (A) Regional overview of the Transhimalayan ranges. Squares are terrace surface ages; triangle is a fan surface of oldest dated glacial succession in the Himalaya-Tibet orogen (Owen et al., 2006; recalculated by Blöthe et al., 2014); circles are

moraines used for glacial chronologies; lavender areas are modern glaciers (Pfeffer et al., 2014). (B) Monsoon proxies compiled from Wang et al. (2001, Hulu cave), Wang et al. (2008, Sanbao cave), Cheng et al. (2012, Kesang cave), and Berger and Loutre (1991, insolation); marine isotope stage substages after Winograd (1997, duration), and Lisiecki and Raymo (2005). (C)–(E) Key geomorphic marker ages. Blue dots are mean ages of glacial advances with reported uncertainties (except for 'In', where error bars extend to youngest and oldest age); grey circles are scaled to maximal glacier advance relative to recent extent; orange bars are major damming phases. Squares are aggradation surface ages with external uncertainties. For consistency, we recalculated the data with the CRONUS online calculator (<http://hess.ess.washington.edu>; Balco et al., 2008) using the ^{10}Be SLHL spallation production rate of Heyman (2014). Data compiled from Dortch et al. (2010, 'Do'), Scherler et al. (2014); Owen et al. (2006, 'In'), Hedrick et al. (2011, 'P' and 'K'), Taylor and Mitchell (2000, 'Z'), and Blöthe et al. (2014).