Abstract  Fluvial incision, regarded as one of the fundamental geomorphic processes, drives the evolution of mountainous landscapes. The transitional landscape from low-relief to high-relief in the central Tibetan Plateau is rapidly evolving as it is influenced by river dynamics, climate change and tectonic uplift. Combining cosmogenic \(^{10}\)Be depth profile dating and topographic analysis, this study provides new constraints on the formation and destruction of low-relief surfaces in the central Tibetan Plateau. We find that the high-relief landscape in the Suoqu area (a major tributary of the upper Nu River) shows a rapid fluvial incision rate of \(710 \pm 70\) mm kyr\(^{-1}\) since the late Pleistocene, while the low-relief topography in the adjacent Xiaziqu area presents an order of magnitude lower incision rate of \(70 \pm 10\) mm kyr\(^{-1}\). These results are consistent with the long-term (multi-million-year) exhumation rates derived from low-temperature thermochronology, suggesting that this region has experienced an evolving incision history. We interpret that the higher relief was caused by enhanced fluvial incision, and the lower relief was slowly developed by sedimentation and relatively steady low exhumation rate. The presence of a knickzone appears to mark the boundary between these differentially incising landscapes, which may be caused by rapid headward retreat and higher river discharge in the Suoqu River. The coincidence of fluvial terraces ages with climate-driven events, in addition to paleodenudation rates, indicates that the formation of fluvial terraces in the Xiaziqu and Suoqu areas might be associated with the quick sedimentation of weathered materials in early warming periods.

Plain Language Summary  Understanding the formation mechanisms of topographic relief is crucial for interpreting the contrast between the low-relief surfaces in the central Tibetan Plateau and the high-relief plateau margin. In this study, fluvial terraces, which directly record landscape incision history, are used to interpret the relationship between fluvial incision and topographic relief in two tributaries of the upper Nu River. We present new terrace formation ages derived from cosmogenic \(^{10}\)Be, a rare radioisotope produced in quartz. We find that the formation ages of terraces since the late Pleistocene are coincident with regional climate change events. Fluvial incision rate in the relatively low-relief Xiaziqu area calculated by dividing the terrace height above the modern river by the exposure age is \(70 \pm 10\) mm kyr\(^{-1}\), which is an order of magnitude lower than that in the high-relief Suoqu area \((710 \pm 70\) mm kyr\(^{-1}\)). Together with river profile analysis in the Xiaziqu and Suoqu areas, we interpret that relief is probably controlled by the sharp change in channel slope, which, in turn, is influenced by tectonically driven rapid headward retreat and climate-driven river discharge.

1. Introduction  High-elevation and low-relief surfaces are widely distributed in the central Tibetan Plateau (Figures 1 and 2), which is in sharp contrast to the high-relief topography of the southeastern Tibetan Plateau (Fielding et al., 1994). The high-elevation and low-relief surfaces of the southeastern Tibetan Plateau have been explained by two models. One model proposes that a pre-uplift continuous low-relief surface formed at low elevation has been uplifted to account for the relict landscape (M. K. Clark et al., 2006). An alternative hypothesis posits that low-relief topography is the result of recent external processes involving relief-reduction by surface exhumation and basin filling due to inefficient drainage (Han et al., 2019; Liu-Zeng et al., 2008).

Fluvial incision is often used as a proxy for tectonic and climate signals and used to unravel the formation and destruction of low-relief landscapes (M. K. Clark et al., 2006; Liu-Zeng et al., 2008; Whipple et al., 2017; R. Yang et al., 2015). In the past two decades, the incision history of the eastern Tibetan Plateau has been
Figure 1.
relatively well-constrained by short-term (multi-millennial) cosmogenic nuclide dating (Cook et al., 2018; Godard et al., 2010; Henck et al., 2011; Ouimet et al., 2009; Schäfer et al., 2002; Y. Yang et al., 2019) and long-term (multi-million-year) low-temperature thermochronology (Kirby et al., 2002; Liu-Zeng et al., 2018; Nie et al., 2018; Tian et al., 2018; R. Yang et al., 2016). However, the central Tibetan Plateau has so far received little attention due to its rugged and complex environment, especially in transitional landscapes from low-relief to high-relief topography in the headwater regions of the Three Rivers (the Jinsha, Lancang, and Nu rivers) (Liu-Zeng et al., 2008).

Previous thermochronological studies suggested that the central and eastern parts of the Tibetan Plateau have experienced relatively low exhumation rates ($<50$ mm kyr$^{-1}$) since the early Eocene (Dai et al., 2013; Rohrmann et al., 2012), resulting in a relatively stable landscape. However, thermochronological data from valley bedrock (M. K. Clark et al., 2005; Nie et al., 2018; Ouimet et al., 2010) indicate that incision rates may have increased since the middle Miocene, carving deep river canyons in the eastern Tibetan Plateau. The channel slope and drainage area are widely used to estimate the bedrock river incision as described in the stream power model (Whipple & Tucker, 1999). Even if a river channel is not in a steady-state condition, Chi plots have been successfully used to identify transient erosional signals (Perron & Royden, 2013; R. Yang et al., 2015). This topographic analysis allows us to explore the regional spatial variation of river dynamics. Despite these advances in studying long-term (thermochronology, multi-million-year) landscape evolution, data on short-term scale (cosmogenic nuclides, multi-millennial) are still sparse in the central Tibetan Plateau, and the timing and mechanisms of low-relief landscape evolution at different scales remains elusive. Especially since empirical and numerical models are difficult to verify with natural processes occurring over a multi-millennial timescale.

Figure 1. (a) Simplified geologic structures of the Tibetan Plateau. IYS, Indus-Yarlung zangbo suture; BNS, Bangonghu-Nujiang suture; LSS, Longmucuo-Shuanghu suture; JSS, Jinsha suture; KLS, Kunlun suture; QLS, Qilian suture; ATF, Altyn Tagh Fault. (b) Hillshade showing river systems and suture zones in the study region. Colored thick lines show channel steepness $k_n$ (Wobus et al., 2006). Red stars represent AHe (red number) and ZHe (black number) ages (in Ma) (Dai et al., 2013). (c) Chi plot for the upper Nu River (Perron & Royden, 2013; Schwanghart & Scherler, 2014). $\theta$ was calculated using the assumed concavity index $\theta = 0.45$. (d and e) Watersheds of sampling sites showing the local landscape, modified from Google Earth. (f) Longitudinal profile of the upper Nu River bedrock channel. Orange and blue lines show uncorrected and corrected profiles based on ArcGIS watershed analysis function of fill, respectively. Short red lines show knickzones and corresponding tributary inlets.

Figure 2. (a) Simplified geological map. (b) Spatial distribution of annually averaged precipitation derived from TRMM. (c and d) Topographic slope and relief maps derived from 90-m-resolution Shuttle Radar Topography Mission (SRTM) digital elevation model (DEM). The relief values are calculated over a $1 \times 1$ km square area. Low-relief areas ($<300$ m, the elevation change is nominal) are mainly distributed in the headwater regions of the river.
To test the hypothesis that renewed basin-wide denudation and fluvial incision are likely modifying the steady-state high-elevation and low-relief surfaces in the central Tibetan Plateau, we present terrestrial cosmogenic nuclide (TCN) $^{10}$Be dating and channel profile analysis on the transitional landscape from low-relief to high-relief topography in the central Tibetan Plateau to focus on (a) the role of climate and river dynamics in terrace formation in the catchment; (b) deciphering the incision patterns across this transitional landscape; and (c) providing new insights into the mechanism of formation and destruction of the low-relief surfaces in the study region.

2. Geological Setting and Sample Descriptions

2.1. Study Area

As a result of the collision between the Indian and Eurasian continental plates, the Tibetan Plateau consists of five main terranes, including the Qaidam, Songpan-Ganzi, Qiangtang, Lhasa and Himalaya from north to south, which are separated by the Kunlun, Jinsha, Bangong-Nujiang and Indus-Yarlung zangbo suture zones, respectively (Yin & Harrison, 2000) (Figure 1a). The central Tibetan Plateau is typically characterized by low-relief (1 km or less at wavelengths of ~100 km) and upland erosion surfaces relative to the much higher relief (up to 6 km at wavelengths of ~100 km) and greater incision on the plateau margins (Fielding et al., 1994; Liu-Zeng et al., 2008). Fluvial incision has cut deep canyons along the plateau margin and formed fluvial terraces along the valleys. The study area is located in the junction area of the Bangong-Nujiang and Longmucuo-Shuanghu sutures (Figure 1), where the Lhasa and Qiangtang terranes were accreted onto the southern margin of Eurasia during the Late Triassic-Early Jurassic and the Late Jurassic-Early Cretaceous (Yin & Harrison, 2000). The Carboniferous strata, interlayered with basalts and mafic sills, consist of shallow marine sequences of quartzite and carbonate rocks, while the Permian, Triassic and Jurassic strata are dominated by carbonate rocks (Figure 2a) interbedded with terrestrial clastic and volcaniclastic strata (Yin & Harrison, 2000). The complex region of continental fragments and arc terranes was reactivated during the Cenozoic, and Cenozoic deformation has affected all of the surface landscapes.

Based on field observations and Google Earth images, fluvial terraces are sparsely preserved along the main stream of the Nu River. Some of them are hard to access due to the steep topographic escarpment, and the rest are mostly utilized as agricultural and residential lands by human activities. However, some strath and fill-cut terraces are well-preserved along the tributary channel in this region (Figures 3 and 4), in which strath terraces are cut into bedrock and fill-cut terraces have formed as the result of both vertical incision and lateral migration (Bull, 1991). Located in the highland sub-frigid and semi-humid climate zone, the mean annual precipitation ranges from 120 to 848 mm (Figure 2b; data from Tropical Rainfall Measuring Mission (TRMM); http://www.geog.ucsb.edu/~bodo/TRMM/). The mean annual precipitation and temperature in Naqu are 450 mm and −0.7°C for the period of 1981–2010, respectively (data from National Meteorological Information Center, China; http://data.cma.cn/en). The studied tributaries (shaded areas in Figure 1b) of the Suoqu and Xiaqiuqu catchments are located above 4,000 m and have large differences in the average slope (20.6° in Suoqu and 12.8° in Xiaqiuqu) and relief (343 m in Suoqu and 209 m in Xiaqiuqu), being higher in Suoqu and lower in Xiaqiuqu (Figures 2c and 2d). Both catchments (closed curves in Figure 2a) show similar lithologic frameworks of shallow marine carbonate rocks. The evolving transitional landscape from low relief (mostly less than 300 m, white areas in Figure 2d) to high relief (greater than 300 m, yellow and red areas in Figure 2d) shows that the upper reach of the Nu River closer to the internally drained interior of the Tibetan Plateau has a lower relief, while the higher relief (greater than 500 m, red areas in Figure 2d) is mainly distributed along the trunk and main tributaries, as well as the suture zones in the Suoqu region (Figure 2).

2.2. Sampling Strategy

Both strath and fill-cut terraces are developed in the Suoqu catchment, while only fill-cut terraces are observed in the Xiaqiuqu catchment (Figure 3). To explore the formation of terraces by external forcings (tectonic uplift and climate change) rather than completely internal dynamics (autogenic processes due to lateral aggradation and erosion) based on the field observation of thick sediments (SQ-T6: >30 m and XQQ-T2: >17 m), we sampled strath terraces (SQ-T6, SQ-T5, and SQ-T1) and the uppermost fill-cut terraces...
Thin soils (less than 5 cm) developed on the pre-existing fluvial sediments. All the abandoned fluvial sediments are composed of upper fine fractions (clay, silt and fine sand) and lower coarse materials (sand, pebbles, cobbles and boulders) (Figures 4 and 5). The compositions of all depth profiles are relatively homogeneous, excluding the fine fraction of SQ-T6 and XQQ-T2 as the interbedded sand sediments. The preservation of upper fine fractions and the lack of paleosoils for all depth profiles suggest that post-denudational erosion of the terraces was probably minimal. Based on the observation of modern fluvi-al sediments (Figure 4), such as SQ-T1 in Figure 5, we assume that the lag time between the sedimentation of coarse materials and fine fractions is negligible. To determine the timing of formation of these terraces, twenty-four ∼3 kg quartz-rich samples, including amalgamated sands and sandy to pebbly gravels, were collected from fluvial terraces in these two tributaries of the upper Nu River (Figures 1 and 3). To avoid the sediments possibly being from different provenances (Carretier et al., 2015; Codilean et al., 2014; Y. Yang et al., 2019), we processed fine gravel- and sand-size clasts (less than 1 cm diameter) and assumed that the quartzes have the same inherited cosmogenic nuclide concentration. All samples were taken vertically through terraces within 4 m from the topmost surface (Figures 4 and 5), and the sampling thickness of each sample site was less than 5 cm. We collected sediment from well-preserved artificial road cuts, modern river erosion profiles and natural gullies, avoiding effects from any modern surface disturbance and landslides (Figures 4 and 5). Six samples were collected from one depth profile (XQQ-T2) in the Xiaqiuqu area, and 18 samples were taken from four depth profiles (SQ-T1, SQ-T4, SQ-T5, and SQ-T6) in the Suoqu area (Table 2).

3. Analytical Methods

3.1. TCN Sample Preparation and Analyses

Twenty-four samples were crushed and sieved to 250–500 μm followed by magnetic and gravimetric separation. Standard chemical preparation (Kohl & Nishiizumi, 1992) was performed at the School of Earth System Science, Tianjin University, China. All samples of pure quartz were screened using inductively coupled plasma optical emission spectrometry (ICP-OES) to confirm the purity of quartz (Al < 150 ppm) before dissolution. The purified quartz samples were dissolved in 5:1 HF/HNO₃ and spiked with ~270 μg of ⁹Be standard. Beryllium was separated and purified by ion exchange chromatography, hydroxylation and oxidation.
BeO was mixed with niobium (BeO:Nb = 1:6) and loaded into the ion source of the 5MV accelerator mass spectrometer (AMS) at the Scottish Universities Environmental Research Centre (Xu et al., 2010). The primary standard NIST SRM 4325 with a nominal value of $^{10}\text{Be}/^{9}\text{Be} = 2.79 \times 10^{-11}$ was used for normalization. Two secondary standards with $^{10}\text{Be}/^{9}\text{Be}$ ratios in levels of $10^{-12}$ and $10^{-13}$ were simultaneously measured to assess the uncertainties of the sample which are typically 3%–5% on $^{10}\text{Be}/^{9}\text{Be}$ ratios of $\sim 10^{-13}$–$10^{-14}$. Two full chemistry procedural blank ratios ranged from 0.3% to 4.5% of the sample $^{10}\text{Be}/^{9}\text{Be}$ ratios (Table 2).

3.2. TCN Model Setup

3.2.1. Exposure Age, Denudation Rate, and Inheritance Calculations

The measured TCN concentration at or near the Earth’s surface can be described as a function of exposure age ($t$), denudation rate ($\varepsilon$) and inheritance ($C_0$) (Lal, 1991; Y. Yang et al., 2019):

Figure 4. Field photographs of Suoqu (a–d) and Xiaqiuqu (e–f) showing locations of sampling sites and fluvial terrace levels.
where \( C(x, \varepsilon, t) \) is the nuclide concentration, \( \lambda \) is the decay constant \( 5.00 \times 10^{-7} \text{ yr}^{-1} \) for \(^{10}\text{Be} \) (Chmeleff et al., 2010; Korschinek et al., 2010), \( P_i \) is the surface production rate (at \( \text{g} \text{ cm}^{-2} \text{ yr}^{-1} \)), \( \rho \) is the integrated density (\( \text{g} \text{ cm}^{-3} \)), \( \Lambda_i \) is the effective attenuation length (\( \text{g} \text{ cm}^{-2} \)), and \( x \) is the depth (cm) (Table 1). The sea level and high latitude (SLHL) spallogenic production rate of 4.01 at \( \text{g} \text{ cm}^{-2} \text{ yr}^{-1} \) (Borchers et al., 2016) for \(^{10}\text{Be} \) is used to scale the local production rate. In order to consider the variation of density with depth and contribution from inheritance, a three-dimensional-graph visualization Chi-square inversion approach (Y. Yang et al., 2019) has been used to determine the best fit values of denudation rate, exposure age, and inheritance. The inversion model allows a precise estimation of the surface denudation rate and inheritance but only a minimum exposure age due to the tradeoff in fitting the denudation rate and exposure age (Delmas et al., 2018; Y. Yang et al., 2019). The surface denudation rates calculated from the inversion approach show that the best fitting denudation rates are around 0 mm kyr\(^{-1} \) (Figure S1 in Supporting Information S1). According to the field-observed well-preserved depth profile surface (Figure 4) and the insignificant modeled surface denudation rates, the assumption of no terrace surface erosion was used to constrain the inversion model in this study.

Figure 5. Field photographs and modeling results of cosmogenic \(^{10}\text{Be} \) depth profiles. Red and black curves show fitting solutions within 1σ and 2σ confidence levels, respectively. Due to the scattered data of SQ-T1 and SQ-T5, equivalent 1σ confidence level (\( \chi^2 \) min + Degree of freedom) and best fitting solutions are depicted by red and blue curves, respectively. See detailed modeling results in Figure S2 in Supporting Information S1.
Table 1
Sampling Sites, Shielding Factors, Scaled Surface $^{10}$Be Production Rates, and Effective Attenuation Lengths

| Sample | Latitude (N) | Longitude (E) | Elevation (masl) | Height (m) | Shielding factor$^b$ | $P_{10}^{spal}$ (at g$^{-1}$ yr$^{-1}$) | $P_{10}^{slow}$ (at g$^{-1}$ yr$^{-1}$) | $P_{10}^{fast}$ (at g$^{-1}$ yr$^{-1}$) | $\Lambda_{slow}^{P}$ (g cm$^{-2}$) | $\Lambda_{fast}^{P}$ (g cm$^{-2}$) |
|--------|--------------|---------------|-----------------|------------|----------------------|--------------------------------------|--------------------------------------|--------------------------------------|-------------------------------------|--------------------------------------|
| SQ-T1  | 31.929       | 93.8825       | 4,044           | 4 ± 1      | 0.9969               | 50.62                                | 0.4615                               | 0.1475                               | 803                                 | 2,767                                |
| SQ-T4  | 31.9269      | 93.9640       | 4,114           | 21 ± 2     | 0.9969               | 52.47                                | 0.4716                               | 0.1501                               | 800                                 | 2,749                                |
| SQ-T5  | 31.9308      | 93.9959       | 4,122           | 30 ± 3     | 0.9969               | 52.70                                | 0.4728                               | 0.1504                               | 799                                 | 2,747                                |
| SQ-T6  | 31.9336      | 93.8986       | 4,123           | 67 ± 5     | 0.9969               | 52.72                                | 0.4730                               | 0.1504                               | 799                                 | 2,747                                |
| XQQ-T2 | 31.7592      | 92.7281       | 4,293           | 17 ± 1     | 0.9994               | 57.33                                | 0.4983                               | 0.1568                               | 792                                 | 2,706                                |

$^a$Heights represent the terrace surface heights above the modern river. $^b$Shielding factors used to correct the topographic or geometric obstructions are determined by the CRONUS online calculator (Balco et al., 2008). $^c$Scaled surface spallation production rates (Stone, 2000) using the SLHL spallogenic production rate of 4.01 at g$^{-1}$ yr$^{-1}$ of Borchers et al. (2016). $^d$Scaled surface slow and fast muons production rates are derived from Balco (2017) using the SLHL slow and fast muons production rates of 0.0959 and 0.0450 at g$^{-1}$ yr$^{-1}$, respectively. $^e$Slow and fast effective attenuation lengths are described in Y. Yang et al. (2020) using the approximate exponential attenuation lengths.

3.2.2. Basin Paleodenudation Rate Calculation

Quartz sediment that was sourced from a catchment and mixed in a fluvial system retains a concentration of $^{10}$Be (inheritance) (Lal, 1991):

$$C_0(\varepsilon_p) = \frac{P_{basin}}{\lambda + \frac{P_{basin}}{\Lambda_{spal}}}$$

where $C_0(\varepsilon_p)$ is the inherited concentration, $P_{basin}$ is the averaged surface production rate of spallation (at g$^{-1}$ yr$^{-1}$) (Table 3), $\rho$ is the averaged density of 2.6 g cm$^{-3}$, and $\varepsilon_p$ is the basin paleodenudation rate. The decrease of elevation caused by surface denudation can reduce the cosmogenic nuclide production rate. However, for the time scale less than 1 Ma, the difference of production rate caused by changed elevation is negligible. For example, the elevation decrease of $\sim$50 m, derived from assumed surface denudation rate of $\sim$50 mm kyr$^{-1}$ (the most frequent denudation rate in the eastern Tibetan Plateau [Y. Yang et al., 2019]), only reduces the cosmogenic nuclides production rate from 72.1 (4,764 m, XQQ-T2) to 70.7 at g$^{-1}$ yr$^{-1}$ (4,714 m).

3.3. River Profile Analysis

In this study, the trunk channel and main tributaries of the upper Nu River were analyzed using a 90 m DEM from SRTM (Figure 1b) using TopoToolbox (Schwanghart & Scherler, 2014). Channel steepness $k_m$ and upstream integral of drainage area, $\chi$, for the upper Nu River were computed based on the methods of Perron and Royden (2013) and Schwanghart and Scherler (2014). A concavity index $\theta$ of 0.45 was used.

4. Results

The measured cosmogenic $^{10}$Be results are listed in Table 2. The $^{10}$Be concentrations vary from 1.60 $\times$ 10$^3$ to 59.93 $\times$ 10$^3$ at g$^{-1}$ with analytical uncertainties less than 4%. The $^{10}$Be concentrations in depth profiles SQ-T4, SQ-T5, SQ-T6, and XQQ-T2 exponentially decrease with depth from the topmost surfaces, with one outlier in SQ-T5-3 (Figure 5 and Table 2). The depth profile SQ-T1 shows greater scatter due to the relatively recent abandonment of the fluvial terrace, which led to lower accumulations of $^{10}$Be that are more difficult to measure precisely. Due to the microscopic heterogeneity in profiles, the abnormal data points SQ-T1-3 and SQ-T5-3 might be also caused by the difference in grain size of the collected samples, in which coarse- or fine-grained quartz are derived from different provenances with variable inheritance (Carretier et al., 2015; Codilean et al., 2014; Y. Yang et al., 2019). The results of SQ-T6-1 and XQQ-T2-1 collected from the fine fractions also exponentially decrease with depth (Figure 5), supporting our assumption of negligible lag time between the deposition of fine and coarse fractions.
Table 2

Analytical Results of Cosmogenic $^{10}$Be Concentrations

| Sample   | Depth (cm) | Density (g cm$^{-3}$) | Qtz mass (g) | Total $^{9}$Be (μg) | $^{10}$Be/$^{9}$Be $^{\times 10^{-13}}$ | ± (%) | $[^{10}$Be] (×10$^5$ at g$^{-1}$) | ± (%) |
|----------|------------|-----------------------|--------------|---------------------|-------------------------------------|------|---------------------------------|------|
| SQ-T1-1  | 80         | 1.80                  | 19.879       | 484                 | 1.55                                | 3.2  | 2.27                            | 3.5  |
| SQ-T1-2  | 125        | 2.02                  | 20.560       | 281                 | 2.89                                | 3.0  | 2.37                            | 3.3  |
| SQ-T1-3  | 155        | 2.09                  | 18.390       | 278                 | 1.82                                | 3.1  | 1.65                            | 3.4  |
| SQ-T1-4  | 230        | 2.19                  | 25.103       | 278                 | 3.50                                | 3.0  | 2.33                            | 3.3  |
| SQ-T4-1  | 75         | 1.92                  | 20.950       | 278                 | 5.97                                | 2.6  | 4.75                            | 2.9  |
| SQ-T4-2  | 150        | 2.16                  | 19.780       | 277                 | 3.11                                | 3.6  | 2.62                            | 3.9  |
| SQ-T4-3  | 215        | 2.23                  | 19.920       | 277                 | 1.91                                | 3.0  | 1.60                            | 3.3  |
| SQ-T5-1  | 110        | 1.80                  | 17.078       | 277                 | 7.05                                | 2.9  | 6.87                            | 3.3  |
| SQ-T5-2  | 145        | 1.94                  | 13.774       | 275                 | 4.11                                | 2.9  | 4.94                            | 3.2  |
| SQ-T5-3  | 200        | 2.07                  | 15.547       | 275                 | 5.37                                | 2.9  | 5.72                            | 3.2  |
| SQ-T5-4  | 250        | 2.14                  | 23.390       | 277                 | 5.31                                | 2.7  | 3.78                            | 3.1  |
| SQ-T5-5  | 300        | 2.18                  | 25.017       | 277                 | 4.62                                | 3.0  | 3.07                            | 3.3  |
| SQ-T6-1  | 100        | 1.80                  | 7.050        | 267                 | 8.96                                | 2.6  | 20.37                           | 2.9  |
| SQ-T6-2  | 160        | 1.80                  | 19.045       | 266                 | 11.86                               | 2.5  | 9.98                            | 2.9  |
| SQ-T6-3  | 195        | 1.91                  | 19.636       | 269                 | 8.55                                | 3.8  | 7.05                            | 3.1  |
| SQ-T6-4  | 230        | 1.98                  | 18.608       | 267                 | 5.92                                | 2.3  | 5.10                            | 2.7  |
| SQ-T6-5  | 270        | 2.04                  | 19.790       | 266                 | 4.61                                | 2.8  | 3.73                            | 3.1  |
| SQ-T6-6  | 350        | 2.13                  | 20.888       | 267                 | 3.20                                | 3.0  | 2.46                            | 3.3  |
| XQQ-T2-1 | 75         | 1.80                  | 8.770        | 272                 | 32.12                               | 1.8  | 59.93                           | 2.3  |
| XQQ-T2-2 | 120        | 1.80                  | 16.920       | 267                 | 41.83                               | 2.4  | 39.69                           | 2.8  |
| XQQ-T2-3 | 170        | 1.98                  | 17.822       | 267                 | 23.41                               | 2.9  | 21.07                           | 3.2  |
| XQQ-T2-4 | 220        | 2.07                  | 14.203       | 266                 | 11.95                               | 2.9  | 13.48                           | 3.2  |
| XQQ-T2-5 | 280        | 2.14                  | 19.907       | 269                 | 9.58                                | 2.3  | 7.78                            | 2.7  |
| XQQ-T2-6 | 340        | 2.19                  | 19.921       | 268                 | 8.99                                | 3.0  | 7.27                            | 3.3  |

$^{a}$The sampling thickness is less than 5 cm, thus, the depth error is regarded as 2.5 cm. $^{b}$Integrated bulk densities are calculated by estimating the proportion of silt (1.5 g cm$^{-3}$) and clast (2.7 g cm$^{-3}$) fraction using field photos and macroscopic description of the horizons (Hancock et al., 1999), resulting in 1.6 g cm$^{-3}$ for upper fine materials and 2.3 g cm$^{-3}$ for lower coarse materials. $^{c}$Normalized $^{10}$Be/$^{9}$Be ratio which was corrected by blank ratio (7.3 $\times$ 10$^{-15}$ for SQ-T1, SQ-T4, and SQ-T5 and 1.2 $\times$ 10$^{-14}$ for SQ-T6 and XQQ-T2).

Table 3

The Results of Exposure Age, Incision Rate, and Paleodenudation Rate

| Profile | Age (ka) | Incision rate $^{a}$ (mm kyr$^{-1}$) | Average basin elevation $^{b}$ (masl) | $P_{\text{tr}}$ $^{c}$ (at g$^{-1}$ yr$^{-1}$) | Inheritance $^{d}$ (×10$^4$ at g$^{-1}$) | Basin paleodenudation rate (mm kyr$^{-1}$) |
|---------|----------|--------------------------------------|-------------------------------------|---------------------------------|---------------------------|------------------------------------------|
| SQ-T1   | 1.4 ± 0.8| 2,900 ± 1,800                        | 4,863                               | 75.8                            | 19.3 ± 0.9                | 241 ± 12                                 |
| SQ-T4   | 16.4 ± 1.8| 1,300 ± 200                           | 4,863                               | 75.8                            | 11.7 ± 1.6                | 399 ± 64                                 |
| SQ-T5   | 23.8 ± 2.5| 1,300 ± 200                           | 4,863                               | 75.8                            | 31.8 ± 1.5                | 146.4 ± 7.5                              |
| SQ-T6   | 100.4 ± 6.4| 670 ± 70                              | 4,863                               | 75.8                            | 17.7 ± 2.8                | 263 ± 49                                 |
| XQQ-T2  | 234.2 ± 12.6| 70 ± 10                               | 4,764                               | 72.1                            | 52.5 ± 4.9                | 84.2 ± 8.8                                |

$^{a}$Incision rates are calculated by dividing the terrace height above the modern river by the exposure age. $^{b}$The average basin elevations are derived from the DEM of each drainage basin. $^{c}$Surface production rates include terrestrial cosmogenic nuclide production via spallation.
The fluvial terrace exposure ages and fluvial incision rates are listed in Table 3. The precise denudation rates derived from the inversion approach indicate that almost all best fit surface denudation rates (SQ-T4: $0.70^{+7.3}_{-0.0} \text{ mm kyr}^{-1}$, SQ-T5: $0.28^{+6.2}_{-0.0} \text{ mm kyr}^{-1}$, SQ-T6: $0.73^{+6.2}_{-0.0} \text{ mm kyr}^{-1}$, and XQQ-T2: $1.7^{+1.9}_{-1.7} \text{ mm kyr}^{-1}$; see Figure S1 in Supporting Information) are negligible compared to the youngest fluvial terrace (SQ-T1: 634 $^{+422}_{-654}$ mm kyr$^{-1}$; see Figure S1 in Supporting Information) due to its scattering of the data and small number of samples. Combined with the field observation of well-preserved surface silt and the lack of surface paleosoil, as well as very low modeled terrace surface denudation rates, the assumption of no surface erosion for the abandoned terraces was used to constrain the inversion approach. Thus, the results constrain the exposure ages to 1.4 ± 0.8, 16.4 ± 1.8, 23.8 ± 2.5, 100.4 ± 6.4, and 234.2 ± 12.6 ka for the sites SQ-T1, SQ-T4, SQ-T5, SQ-T6, and XQQ-T2, respectively (Figure 5 and Table 3). The integrated incision rate since the late Pleistocene (from T6 to modern river) for the Suoqu River is 710 ± 70 mm kyr$^{-1}$ (single terrace incision rates range from 670 to 2,900 mm kyr$^{-1}$, Table 3), which is an order of magnitude higher than the incision rate of 70 ± 10 mm kyr$^{-1}$ for the Xiaqiuqu River (Figure 6a and Table 3).

The results of $^{10}$Be inherited concentrations and paleodenudation rates are presented in Table 3. The inheritance derived from depth profiles was used to calculate the basin paleodenudation rates. For the Suoqu area the paleodenudation rates vary from 146.4 ± 7.5 to 399 ± 64 mm kyr$^{-1}$ (Table 3), with a long term mean denudation rate of 262 ± 90 mm kyr$^{-1}$. In contrast, the paleodenudation rate for the Xiaqiuqu area presents a relatively low value of 84.2 ± 8.8 mm kyr$^{-1}$.

5. Discussion
5.1. The Formation of Fluvial Terraces Associated With Climate Change

Terrace formation records the processes of lateral planation and vertical incision of the river, which might be the result of external forcing, such as, climate-driven changes in sediment flux and water discharge to rivers (Hancock & Anderson, 2002; Molnar et al., 1994), tectonically driven changes in elevation/base level (Howard et al., 1994), and/or internal forcing of autogenic processes due to lateral aggradation and erosion (Limaye & Lamb, 2016; Malatesta et al., 2017; Scheingross et al., 2020). The thick sediments (SQ-T6: >30 m and XQQ-T2: >17 m) are unlikely to have been formed by lateral aggradation without external forcing (tectonic uplift or climate change) in such high elevation and low precipitation regions. Recently, numerous studies have proposed that the terraces’ formation ages are regionally consistent with the timing of glacial to interglacial transition across the Tibetan Plateau (Dey et al., 2016; Hetzel, 2013; Huang et al., 2019; Pan et al., 2009; Tao et al., 2020; Y. Yang et al., 2019; Zhang et al., 2018), in which the great changes of flood magnitude and frequency caused by the climatic transitions are responsible for the formation of fluvial terraces (Y. Yang et al., 2019). However, terrace formation linked to climate change has a long history of debate. On one hand, terrace formation is considered as the result of lateral aggradation in cold periods (Molnar et al., 1994; Pan et al., 2009). Conversely, terrace formation has also been interpreted as resulting from the quick sedimentation of weathered materials in early warming periods (Hetzel, 2013; Y. Yang et al., 2019).

In our study, cosmogenic $^{10}$Be exposure ages derived from fluvial terraces show multiple depositional episodes from 234.2 ± 12.6 ka (XQQ-T2) to 1.4 ± 0.8 ka (SQ-T1) (Figure 6). By comparing with paleoclimatic records from the oxygen isotope record of East Asian summer monsoon in Sanbao Cave stalagmites (Cheng et al., 2016), a stacked Benthic oxygen isotope curve (Lisiecki & Raymo, 2005), and summer insolation at 65°N (Berger & Loutre, 1991), all of the terraces’ exposure ages coincide well within error with the climate change from cold to warm, even though the age uncertainties partly extend over the climate variability. Our results are basically consistent with the statistical peaks for the terrace ages ($N = 1,547$) within the Tibetan Plateau and along the surrounding orogenic belts (Tao et al., 2020), such as ∼13 ka (SQ-T4: 16.4 ± 1.8 ka), 25–28 ka (SQ-T5: 23.8 ± 2.5 ka), 100–105 ka (SQ-T6: 100.4 ± 6.4 ka), and possible ∼220 ka (XQQ-T2: 234.2 ± 12.6 ka). The exposure age of XQQ-T2 (234.2 ± 12.6 ka) likely indicates that sediments were abandoned in the valleys when the climate changed from cold (MIS 8) to warm (MIS 7), during termination III (243 ka) (Figures 6b and 6c). Later, the age of SQ-T6 (100.4 ± 6.4 ka) corresponds to the events of the MIS 5d/5c transition from cold to warm (Figure 6b) within 1σ uncertainty. The exposure age of SQ-T5 (23.8 ± 2.5 ka) is plausibly consistent with the Last Glacial Maximum (P. U. Clark et al., 2009) or the Heinrich event 2 (H2) (Hemming, 2004) (Figure 6e). The exposure age of SQ-T4 (16.4 ± 1.8 ka) shows a similar age to the Heinrich event 1 (H1) at about 16.8 ka (Hemming, 2004) (Figure 6e). The most recent events
happened about 1.4 ± 0.8 ka (SQ-T1), approximately coincident with the Himalayan-Tibetan Holocene glacial stage 2 (HTHS 2) (Saha et al., 2019) (Figure 6e), which is one of the regional events of glaciation across the monsoon-influenced and adjacent regions of the Himalayan-Tibetan orogen. The abandoned fluvial terraces in this study seemingly show correspondence within 1σ error with climate change records from cold
to warm conditions, in relation to previous works (Huang et al., 2019; Odom et al., 2020; Tao et al., 2020; Y. Yang et al., 2019). During glacial periods, enhanced frost shattering and degradation of vegetation both caused an excess of sediment on mountain slopes or small valleys (Y. Yang et al., 2019). Subsequently, climate-driven (warmer and wetter during interglacial) debris flow or glaciation transported these sediments to valleys, causing the deposition of these thick fluvial sediments. The LGM deposit (SQ-T5) also supports the bedrock surface denudation event (27.0^{+9.5}_{-7.6} ka) by glacier retreat in the eastern Tibetan Plateau (Y. Yang et al., 2019). However, the absence of fluvial terraces between 100.4 ± 6.4 ka (T6) and 23.8 ± 2.5 ka (T5) in the Suoqu area may suggest that not all terraces would be well-preserved and/or even terraces might have not been formed due to topographic constraints in the local area or limited sediment supply (Bull, 1991). Previous studies also proposed that the absence of fluvial terraces might be caused by the reworking and/or recycling of the sediments by subsequent erosion events (Y. Yang et al., 2019 and references there in). In addition, the ages of T2 and T3 in the Suoqu area may be consistent with the climate change events between HTHS 2 and Heinrich event 1 (Figure 6d). Therefore, our study confirmed that the formation ages of fluvial terraces in the study area are coincident with the quick sedimentation of weathered materials in early warming periods.

5.2. Spatial and Temporal Evolution of the Central Tibetan Plateau Surface Landscape

The TCN-derived incision rates and basin-averaged paleodenudation rates in this transitional landscape were used to explore the timing and mechanisms of strath and fill-cut terraces. Spatially, the Suoqu and Xiaqiuqu areas show similar average precipitation amounts of 316 and 382 mm yr\(^{-1}\) (TRMM), respectively, and similar lithologies, mainly composed of carbonate interbedded with terrestrial clastic and volcaniclastic strata. However, the fluvial incision rate of 70 ± 10 mm kyrr\(^{-1}\) (Figure 6a) in the relatively low-relief Xiaqiuqu area is an order of magnitude lower than that in the high-relief Suoqu area (710 ± 70 mm kyrr\(^{-1}\)). The higher incision rate of 710 ± 70 mm kyrr\(^{-1}\) is consistent with the fluvial incision rates of 380–6,120 mm kyrr\(^{-1}\) estimated for the eastern Tibetan Plateau (Godard et al., 2010; Y. Yang et al., 2019; Zhang et al., 2018). The averaged paleodenudation rate of 262 ± 90 mm kyrr\(^{-1}\) in the Suoqu area (Figure 6b) is much higher than the paleodenudaition rate of 84.2 ± 8.8 mm kyrr\(^{-1}\) in the Xiaqiuqu area and most basin-wide denudation rate data in the eastern Tibetan Plateau, 47% (n = 90) of which are lower than 130 mm kyrr\(^{-1}\) (Y. Yang et al., 2019). Those observations suggest that the fluvial incision and paleodenudation rates in the Suoqu area are much higher than in the Xiaqiuqu area. Although the fill-cut terrace incision rate could be higher than the bedrock incision rate, the coincidence of incision rate (70 ± 10 mm kyrr\(^{-1}\)) and paleodenudation rate (84.2 ± 8.8 mm kyrr\(^{-1}\)) within 1σ confidence level (Figures 6a and 6b) suggests that the Xiaqiuqu landscape has been in a state of dynamic equilibrium since the late Pleistocene. However, in the Suoqu area, the high incision rate (710 ± 70 mm kyrr\(^{-1}\)) relative to the paleodenudation rate (262 ± 90 mm kyrr\(^{-1}\)) indicates that the landscape has not yet reached a steady state. Variability in paleodenudation rates in the Suoqu area, ranging from 146.4 ± 7.5 to 399 ± 64 mm kyrr\(^{-1}\) (Figure 6b and Table 3), is potentially the result of climatically or seismically (Hovius et al., 2011; Li et al., 2017; W. Wang et al., 2017) induced sediment transport events, which can trigger landslides to expose deeply buried low-concentration materials. Temporal variations in fluvial incision since the late Pleistocene are shown in Figures 6a and 6d in the Suoqu area, in which an increase of incision rate from 480 ± 90 mm kyrr\(^{-1}\) (integrated rate from ~100 to ~24 ka with 1σ error; Figure 6a) to 1,270 ± 140 mm kyrr\(^{-1}\) (integrated rate since ~24 ka; Figures 6d and 7) and an average rate of 710 ± 70 mm kyrr\(^{-1}\) (integrated rate since ~100 ka) are observed. However, the seemingly enhanced fluvial incision on short-term scales (multi-millennial) might be caused by various processes such as a relative high incision rate of fill-cut terraces than bedrock, autogenically narrowing channels and accelerating
downcutting (Malatesta et al., 2017), a Sadler effect artifact due to the short averaging time interval (Sadler, 1981), lag time between uplift and incision (Ouimet et al., 2010; Whipple, 2004) or episodic incision hiatuses (Finnegan et al., 2014) as a proposed unsteady river bed evolution model in Y. Yang et al. (2019).

Based on the field observations (Figures 3 and 4) and TCN data, we interpret the terraces formation may be that the SQ-T6 first deposited due to the transition from cold (MIS 5d) to warm (MIS 5c), followed by the river cutting down to the depth of the boundary between the SQ-T5 deposit and bedrock with an incision rate of 480 ± 90 mm kyr$^{-1}$ (ignoring the incision hiatuses during the deposition and incision of the materials of SQ-T5). Next, SQ-T5, SQ-T4, SQ-T3, and SQ-T2 terraces were continuously formed, subsequently, fill-cut terraces are discontinuously abandoned with a higher incision rate of 1,270 ± 140 mm kyr$^{-1}$, finally forming and incising SQ-T1. These abandoned terraces are likely connected to the quick sedimentation of weathered materials in early warming periods. The incision hiatuses during the sedimentation and incision of the materials of SQ-T5 can largely decrease the incision rate of SQ-T6, which might plausibly result in an enhanced fluvial incision in a short-term scale.

The long-term scale (multi-million-year) landscape evolution of this area has been well constrained by low-temperature thermochronology, showing that base level changes due to tectonic uplift and/or intensified monsoon precipitation as climate change probably influence the evolution of the eastern and central Tibetan Plateau (M. K. Clark et al., 2005; Dai et al., 2013; Liu-Zeng et al., 2018; Nie et al., 2018; Ouimet et al., 2010; Replumaz et al., 2020; Rohrmann et al., 2012; Tian et al., 2018; Y. Wang et al., 2018; R. Yang et al., 2016). In the Xiaqiuqu area, the apatite (U-Th)/He (AHe) ages ranging from 35.7 ± 9.4 to 65.9 ± 6.4 Ma (Figure 1b) are substantially older than the age of 15.0 ± 0.8 Ma in the Suoqu area (Dai et al., 2013), suggesting that the Suoqu area experienced a high exhumation rate of 100–200 mm kyr$^{-1}$ from ∼20 Ma relative to the lower exhumation rate of 20–40 mm kyr$^{-1}$ from ∼50 Ma in the Xiaqiuqu area (Figure 7). According to the similarly low exhumation rates (<50 mm kyr$^{-1}$) from about 45 Ma in the central Tibetan Plateau (Rohrmann et al., 2012), Dai et al. (2013) suggested that the central and eastern parts of the Tibetan Plateau have been in a stable landscape with low exhumation rates (<50 mm kyr$^{-1}$) since the early Eocene. However, the enhanced fluvial incision rate (100–500 mm kyr$^{-1}$) since the middle Miocene in the eastern Tibetan Plateau associated with sediment transport by external drainages, such as the Yellow, Yangtze, Lancang and Nu rivers, are proposed according to low-temperature thermochronological data in valley bedrock (M. K. Clark et al., 2005; Liu-Zeng et al., 2018; Nie et al., 2018; Ouimet et al., 2010; E. Wang et al., 2012; R. Yang et al., 2016). The much higher exhumation rates (600–1,400 mm kyr$^{-1}$) since the Quaternary were presented in a tectonic active region of the middle Nu River valley (Replumaz et al., 2020). Increasing exhumation rates from the central to eastern Tibetan Plateau may show that the headward retreat of external drainage mainly controls the deeply carved landscape (Liu-Zeng et al., 2008), partly enhanced by local tectonic uplift. From combining of these different timescales exhumation rates with our TCN results, we suggest that (a) the slightly increasing incision rates from 20 to 40 mm kyr$^{-1}$ (multi-million-year timescale) to 70 ± 10 mm kyr$^{-1}$ (multi-millennial timescale) (Figure 7) and the equivalent paleodenudation...
rate of 84.2 ± 8.8 mm kyr⁻¹ might slowly develop the low-relief landscape by local sedimentation in the Xiaqiuqu area; (b) but the increasing incision rates from 100 to 200 mm kyr⁻¹ (multi-million-year timescale) to 710 ± 70 mm kyr⁻¹ (multi-millennial timescale) may destroy the previously formed low-relief landscape to build the high-relief surface in Suoqu area. The higher exhumation, fluvial incision and paleodenudation rates in the Suoqu area at different time scales show that the high relief areas, including the eastern Tibetan Plateau, are demonstrably more active than those in the central Tibetan Plateau that are characterized by low relief and lower exhumation rates. The consistency of increased relief of ~128 m since past ~200 ka (~710 mm kyr⁻¹ minus ~70 mm kyr⁻¹, then multiply by 200 ka) and the difference of modern relief of 134 m between the Xiaqiuqu and the Suoqu regions (209 m in Xiaqiuqu and 343 m in Suoqu, see detail in Section 2.1) supports that the low-relief surfaces in the central Tibetan Plateau have been recently incised.

Knickzones in channel profiles are regarded as the boundaries between downstream segments that have reached their stable gradients and upstream segments that have not yet steepened in response to rapid base level changes (Whipple et al., 2017; Whipple & Tucker, 1999). Previous studies proposed that a migrating knickzone can largely influence the local exhumation rate (Nie et al., 2018; Replumaz et al., 2020; R. Yang et al., 2016; Zhang et al., 2018). Here, we present the channel gradient analysis and find that the existence of a knickzone (black box in Figure 1b) between the outlets of the Suoqu and Xiaqiuqu rivers (high k_sn values in Figure 1b and abrupt river base level rise of upstream in Figure 1f) probably controls the differential surface incision process (70 ± 10 mm kyr⁻¹ for the Xiaqiuqu River and 710 ± 70 mm kyr⁻¹ for the Suoqu River). Although there is no obvious difference in bedrock properties (Figure 3a) or the precipitation (Figure 3b) between the two study areas, the modern glaciated area of the Suoqu watershed is much larger than that in the Xiaqiuqu watershed. The higher river discharge supplied by the melting of more widely distributed modern glaciers in the Suoqu area has been observed (Figures 1d and 1e). Thus, we propose that the knickzone is probably caused by the rapid incision and higher paleodenudation in the Suoqu River catchment, which are controlled by both rapid headward retreat (associated with local tectonic uplift [Liu-Zeng et al., 2008] and/or the intensified East Asian monsoon since the middle Miocene [Nie et al., 2018]) and higher river discharge of the Suoqu River. In the Xiaqiuqu area (upper reach of the knickzone), both the low fluvial incision rate of 70 ± 10 mm kyr⁻¹ and paleodenudation rate of 84.2 ± 8.8 mm kyr⁻¹ are influenced by the relatively higher local base level restricted to the knickzone, indicating that the landscape is likely to be in steady state and the low-relief surface slowly developed. Integrating the evidence from incision rates (TCN), paleodenudation rates (TCN), exhumation rates (AHe), Chi values, and k_sn values in study areas, we propose that the discrepancy of surface landscape evolution between the low and high relief areas is likely caused by knickzones, which, in turn, are controlled by base level changes due to long-term tectonic uplift and/or short-term river discharge change.

6. Conclusions

We present new cosmogenic ⁹Be exposure ages, paleodenudation rates, and fluvial incision rates, as well as previous long-term exhumation rates, in order to constrain surface landscape evolution in the upper Nu River in the central Tibetan Plateau. Our results show that (a) the formation ages of fluvial terraces since the late Pleistocene are coincident with transition duration from glacial to interglacial; one during glaciation but due to large errors our speculation is tentative.; (b) a steady state topographic relief has been reached in the Xiaqiuqu area, but not in the adjacent Suoqu area; (c) the fluvial incision rate of 70 ± 10 mm kyr⁻¹ in the relatively low-relief Xiaqiuqu area is an order of magnitude lower than that in the high-relief Suoqu area (710 ± 70 mm kyr⁻¹); (d) the high relief areas including the eastern Tibetan Plateau is demonstrably more unstable than that in the central Tibetan Plateau characterized by low relief and lower exhumation rate; and (e) the discrepancy of surface landscapes between the low-relief area and high-relief area is probably caused by the knickzone, which, in turn, are controlled by base level changes due to long-term tectonic activity and/or short-term river discharge changes. These findings support our hypothesis that high-elevation and low-relief surfaces in the central Tibetan Plateau were recently incised by rapid headward retreat of external drainage and higher river discharge.
Data Availability Statement

Data sets for this research are available on Dryad archive at https://doi.org/10.5061/dryad.9cpn5hqj1. The three-dimensional-graph visualization Chi-square inversion approach is described in Y. Yang et al. (2019) and associated supplementary. Topographic analysis is done using MATLAB-based software TopoToolbox 2 in Schwanghart and Scherler (2014).

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