

Glacier response to the transition from lake- to land-terminating following a glacial lake outburst flood at Jinwu Glacier, southeastern Tibet

The 2020 abrupt drainage of Jinwuco triggered lake to land terminus transition and lagged slowdown of Jinwu Glacier, southeastern Tibet

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Abstract. ~~Glacier-p~~Proglacial lake ~~interactions~~ can accelerate terminus retreat and dynamic thinning of lake-terminating glaciers. However, glacier responses to abrupt lake disconnection following glacial lake outburst floods (GLOFs) remain poorly quantified. Here, we investigate ~~the dynamic evolution~~changes in surface velocity (2017–2025) and elevation (2002–2025) of Jinwu Glacier (~~in~~ southeast Tibet, China), whose proglacial lake (Jinwuco) drained during a GLOF on 26 June 2020, shifting the glacier from lake- to land-terminating conditions. Following this loss of lake contact, the downstream glacier tongue showed clear changes in ice flow and frontal morphology. Mean ice-flow velocity within the 100–200 m section from the terminus decreased from $42.97 \pm 4.38 \text{ m a}^{-1}$ during 2017–2020 to $24.09 \pm 6.71 \text{ m a}^{-1}$ during 2021–2025, corresponding to a reduction of approximately 44%. During 2020–2021, the glacier front was characterized by ice-cliff degradation and collapse, together with a slight terminus advance of approximately 23 m. Rapid lake drainage triggered a pronounced but lagged dynamic response. Ice-flow velocities within 0–200 m of the terminus decreased by ~49%, from ~40 m a⁻¹ (2017–2020) to ~20 m a⁻¹ (2022–2025). In contrast, the upstream glacier tongue showed no pronounced velocity change, with mean velocity in the 650–1600 m section changing from $37.55 \pm 9.56 \text{ m a}^{-1}$ during 2017–2020 to $35.24 \pm 9.37 \text{ m a}^{-1}$ during 2021–2025. These results show that GLOF-induced lake disconnection can substantially alter ice flow and frontal morphology near the glacier terminus. velocity reductions in the upstream reach (600–1550 m) were smaller (~14%). Surface elevation thinning in the terminal 0–550 m

section intensified from -2.90 m a^{-1} during 2002–2014 to -3.71 m a^{-1} during 2014–2025, whereas surface lowering in the 600–1550 m section slowed from -1.17 to -0.87 m a^{-1} , with a slight surface elevation increase in the topographic transition zone (500–750 m). Following the GLOF, the glacier terminus underwent slight advance and localized ice calving. These patterns suggest a short lived longitudinal extension at the glacier terminus, followed by a shift toward a more compressive regime in the 500–750 m zone as downstream ice flux demand weakened. This study provides the first quantitative evidence of glacier dynamic adjustment following a GLOF driven transition from lake- to land-terminating conditions.

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1 Introduction

Since the 20th century, most glaciers in most regions earth have retreated and thinned, and mass loss has accelerated over recent decades under continued global warming (Zemp et al., 2025; Hugonnet et al., 2021; Maurer et al., 2019). Glacier retreat promotes the ~~formation and~~ expansion of glacial lakes, driving rapid increases in both lake number and area (Wang et al., 2020; Shugar et al., 2020; Zhang et al., 2024). This development has increased the hazard and risk associated with glacial lake outburst floods (GLOFs) (Zhang et al., 2025; Zhang et al., 2023b; Zheng et al., 2021c). The sudden and catastrophic release of lake water during GLOFs can produce severe socio-economic and geomorphic impacts far downstream, threatening critical infrastructure such as hydropower facilities, bridges, and roads (Richardson and Reynolds, 2000). This problem is particularly acute in High Mountain Asia (HMA), where several high-magnitude GLOFs have been documented and glacial lakes continue to expand (Zheng et al., 2021a; Lützow et al., 2023; Zhang et al., 2025). Many studies have assessed GLOF hazard and risk at regional and local scales (Zhang et al., 2023b; Zheng et al., 2021c; Ahmed, 2025), with a primary focus on downstream process chains and impacts. In contrast, the role of GLOFs as event-driven perturbations to ~~upstream~~ source glaciers remains poorly quantified, particularly for glaciers in contact with proglacial lakes.

Proglacial lakes typically form through two main pathways. One involves the coalescence of multiple supraglacial ponds near the glacier terminus (Carrivick and Tweed, 2013). The other occurs when ~~a retreating~~ glacier retreat exposes deeply carved subglacial depressions that progressively store water ~~as retreat continues~~ (Schomacker, 2010). Proglacial lakes can enhance glacier mass loss by

promoting thermal undercutting and calving at the glacier terminus (Watson et al., 2020). Lake water also exerts buoyant support on the ice and reduces effective pressure, thereby lowering basal friction and increasing ice flow (Benn et al., 2007). Lake-terminating glaciers often behave differently from land-terminating glaciers. Several studies report faster ice flow and stronger velocity variability for lake-terminating glaciers (Pronk et al., 2021; Liu et al., 2020; King et al., 2018), as well as more pronounced mass loss near the terminus (Tsutaki et al., 2019; Maurer et al., 2019). Importantly, glacier–lake contact is not constant (Luo et al., 2025). As the glacier retreats and the proglacial lake expands into the exposed basin, the lake may eventually occupy the available over deepened basin completely, leading to a progressive separation between the lake and the glacier terminus. Lake expansion and changes in terminus geometry can lead to progressive detachment from the lake. In contrast, GLOFs can lower the lake level abruptly, shifting glaciers from lake-terminating to land-terminating conditions. Unlike gradual detachment during long-term retreat, GLOF-driven detachment represents a rapid short-lived but strong perturbation to the terminus boundary conditions, yet the magnitude and temporal evolution of the glacier dynamic response remain poorly understood. A previous recent study reported a dynamic response after the risk-mitigation lowering of a proglacial lake by 5.6 m at Longbasaba Glacier. The glacier subsequently experienced accelerated retreat, with ice speed exceeding its long-term mean in 2006, before entering a period of decelerating accelerated above its mean flow speed, followed by a period of deceleration (Liu et al., 2020). This suggests that abrupt changes in terminus boundary conditions can induce strong short-term dynamic adjustments in lake-terminating glaciers.

Climate projections indicate a trend toward warmer and wetter conditions in monsoon-influenced parts of High Mountain Asia (HMA) (Kitoh, 2017; Sanjay et al., 2017), alongside an expected increase in GLOF occurrence (Zheng et al., 2021c). It is therefore important to quantify how GLOFs affect glacier termini and to characterize the subsequent short-term adjustment and longer-term evolution. Improved understanding of these processes is critical for constraining projections of glacier response and informing hazard assessment and risk management.

This study focuses on the Jinwuco–Jinwu Glacier system in southeastern Tibet, China. On 26 June 2020, the proglacial lake Jinwuco produced a GLOF during a period of high air temperature and heavy rainfall (Zheng et al., 2021b). The event caused a rapid lake-level decline and shifted Jinwu Glacier to a land-terminating glacier. We integrate multi-source historical remote-sensing data to reconstruct changes in terminus geometry, surface velocity, and surface elevation before and after the event. We then quantify

how the GLOF driven step change in terminus boundary conditions affected glacier dynamics. Finally, we discuss the ~~glacier dynamics short-term~~ response, ~~potential lags~~, and the subsequent adjustment toward a new dynamic state.

2 Study area

Jinwu Glacier (30.335° N, 93.634° E) is located in the middle-upper Nidou Zangbo Basin in the eastern Nyainqentanglha, Tibetan Plateau (Figure 1a, b). The glacier lies on the northern flank of Nenang Peak, the highest summit in the basin (Figure 1c). The Nidou Zangbo basin covers ~1268 km² (Lehner and Grill, 2013) ~~and spans an elevation range of ~3666 m a.s.l. based on the SRTM-DEM. Elevations generally decrease from west to east. Basin topography is higher in the west and lower in the east, and~~ the mean elevation exceeds 5000 m a.s.l. Most peaks rise above 6000 m a.s.l., with Nenang Peak reaching 6870 m a.s.l. ~~(Rabus et al., 2003). Owing to this high relief setting, glaciers and glacial lakes are widespread.~~ The basin contains 343 glaciers with a total area of ~385 km² ~~(Consortium, 2023)(RGI 7.0)~~. In 2018, 39 glacial lakes were mapped with a total area of ~3.0 km² (Wang et al., 2020). The regional climate is strongly influenced by the monsoon, with cold winters and cool summers (Yao et al., 2012). Annual mean air temperature increased significantly during 1990–2025, with a linear warming rate of ~0.03 °C a⁻¹ (p = 0.01, R² = 0.16; Figure 1d). Jinwu Glacier is oriented approximately north to south and covers ~8.16 km² in 2025. Its lower tongue is partly debris covered and includes a pronounced icefall. The icefall has an elevation drop of ~300 m, and the maximum slope exceeds 35° ~~, based on the SRTM-DEM (Figure 1e), based on the SRTM DEM (Rabus et al., 2003).~~

A moraine-dammed proglacial lake, Jinwuco, is located at the glacier terminus. Jinwuco produced a GLOF on 26 June 2020. The event was triggered by a large debris landslide from the western lateral moraine that entered the lake (Zheng et al., 2021b). Before the flood, the glacier tongue was in direct contact with the lake. The lake area was ~0.58 km² and its straight-line length was ~1.8 km, making it the largest glacial lake in the basin. ~~Post-GLOF bathymetric measurements indicate that the Jinwuco basin consists of two main depressions (Zhang et al., 2023a). Ordered from the terminal moraine toward the glacier, the glacier terminus has been mainly located near the second depression since 1990. The maximum water depth in this depression was approximately 17 m after the GLOF. Considering an estimated lake-level lowering of approximately 20 m after the GLOF (Zheng et al., 2021b), the pre-~~

GLOF maximum water depth in this depression was inferred to be approximately 37 m. After the outburst, the lake area rapidly decreased to $\sim 0.27 \text{ km}^2$ in 2020, and lake level lowering led to loss of contact between the lake and the glacier terminus.

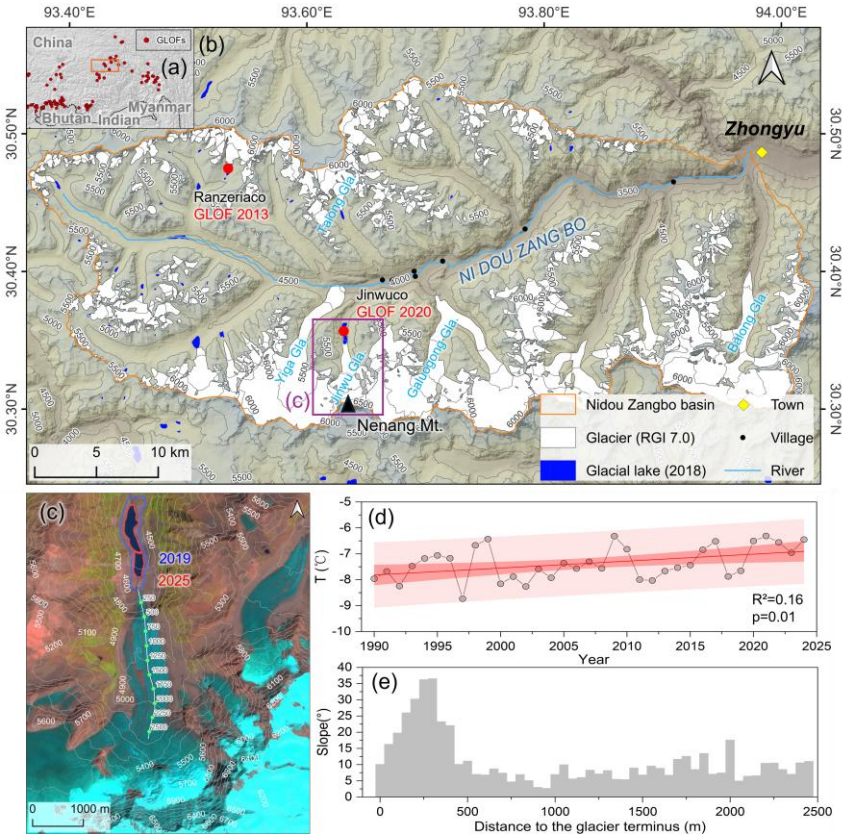


Figure 1 Study area and basic geographic setting of Jinwu Glacier in southeastern Tibet. (a) Distribution of reported glacial lake outburst floods (GLOFs) in southeastern Tibet (Lützow et al., 2023) and the location of the Nidou Zangbo basin. The basemap in panel (a) is Esri World_Shaded_Relief (Powered by Esri). (b) Overview map of the Nidou Zangbo basin showing glaciers (RGI 7.0), glacial lakes (Wang et al., 2020), rivers, and settlements, with Jinwuco and the 2020 GLOF site indicated. The hillshade in panel (b) is based on the SRTM DEM. (c) Sentinel-2 optical image acquired on 25 August 2024, with a close-up view of Jinwu Glacier and its proglacial lake (Jinwu). (d) Interannual variability and linear trend in basin-mean annual air temperature from ERA5 during 1990–2025. (e) Longitudinal slope profile derived from the SRTM DEM along the main flowline of Jinwu Glacier as a function of distance from the glacier terminus. All numbers shown in the figure without explicit units are in meters (m).

3 Data and Methods

3.1 Glacial lake and glacier terminus mapping

Glacial lake area and glacier terminus position were manually mapped from optical satellite images. We reconstructed changes in the extent of Jinwuco from 1990 to 2025 based on historical remote sensing imagery (Table S1) using manual visual interpretation. Landsat imagery (TM/ETM+/OLI, 30 m resolution) was used for the period from 1990 to 2015, and PlanetScope images (3 m resolution) were used for 2016 to 2025.

The area uncertainty associated with lake-boundary delineation from optical imagery primarily depends on image quality and can be approximated as ± 0.5 pixel (Salerno et al., 2012). In addition, the uncertainty introduced by manual delineation can be assumed to follow a Gaussian distribution (Hanshaw and Bookhagen, 2014). We estimated the uncertainty (δ) in glacial lake area using Eq.(1) (Hanshaw and Bookhagen, 2014):

$$\delta = \frac{P}{G} \times \frac{G^2}{2} \times 0.6872 \quad (1)$$

where P is the perimeter of the glacial lake (m), G is the spatial resolution of the images used (m), 0.6872 is the revised coefficient under 1σ (i.e. approximately 69% of peripheral pixels are subjected to errors).

3.2 Surface velocity

We used 10 PlanetScope images (Table S1) to derive annual surface velocity fields for Jinwu Glacier from 2017 to 2025. To capture displacement over a full calendar year, we prioritized image pairs acquired in the second half of each year. After October, in autumn and winter, terrain shadows often obscure large parts of the lower glacier (Fig. S1), reducing image texture and degrading feature tracking. We therefore mainly selected scenes from September and early October to minimize shadow effects. We also prioritized images with minimal cloud and snow cover to improve tracking reliability and spatial coverage.

Surface velocities of Jinwu Glacier were derived using the Coregistration of Optically Sensed Images and Correlation (COSI-Corr) technique implemented in ENVI (Leprince et al., 2007). This approach has been widely applied to glacier motion monitoring and has shown robust performance (Gao et al., 2025; Li et al., 2023; Singh Jasrotia et al., 2024). We used a correlation window of 150×150 m

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(50 × 50 pixels) with a step size of 30 × 30 m (10 × 10 pixels). The search range was set to 90 × 90 m (30 × 30 pixels). Systematic biases in the east–west and north–south displacement components were corrected using stable terrain. Stable terrain was defined as non-glacier surfaces unaffected by clouds, lakes, or topographic shadows, and was further restricted to pixels with slopes <30° (Fig. S2). To suppress outliers, we excluded pixels with velocities exceeding four times the normalized median absolute deviation (NMAD) (Fig. S2). Glacier velocity (GV) was then calculated using Eq. (2) and Eq. (3):

$$D = \sqrt{Dx^2 + Dy^2} \quad (2)$$

$$GV = \frac{D \times A}{N} \quad (3)$$

where D is the displacement measured between each image pair, Dx and Dy are the east–west and north–south displacement components, respectively, N is the time interval between the image pairs (days), and A is the number of days per year (365). To further improve robustness, we retained only velocity pixels with a signal-to-noise ratio (SNR) > 0.7. Local extreme outliers were then identified and removed using a 3 × 3 pixel moving window. Subsequently, velocity profiles along the glacier centreline were extracted at 50 m intervals, and profile points showing velocity differences greater than 15 m a⁻¹ relative to adjacent points were excluded. To further evaluate the reliability of the derived velocity fields, we manually selected visually traceable surface features from the original image pairs for independent validation, as described in Sect. 5.2.

We assessed velocity uncertainty using residual displacements over stable, non-glacier terrain (Mattea et al., 2025; Paul et al., 2022). The mean residual displacement over stable terrain was 2.65 m a⁻¹. NMAD values ranged from 0.82 to 2.58 m a⁻¹ (Table S1). We additionally used the open-access global glacier surface velocity dataset ITS_LIVE (annual mosaics, 1986–2022, 120 m resolution) as a benchmark for comparison (Gardner et al., 2025).

3.3 Surface elevation change

The DEMs used in this study were generated from ASTER L1A stereo data using the NASA Ames Stereo Pipeline (ASP) (Beyer et al., 2018). This approach has been widely applied to glacier surface elevation measurements (Brun et al., 2017; Bhattacharya et al., 2021). We produced three DEMs for the

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study area, dated 2 October 2002, 1 November 2014, and 13 June 2025. Surface elevation change was derived by DEM differencing. Because suitable ASTER stereo pairs over this glacier are limited, the available DEM epochs do not provide a strict pre- and post-GLOF comparison. In particular, the 2014–2025 differencing period spans both several years before and after the 2020 GLOF. Therefore, the DEM-differencing results are used to characterize multi-year surface elevation changes, rather than to isolate the direct event-scale elevation response to the GLOF. DEM biases may arise from limitations in DEM generation and from differences in processing strategies across datasets (Ke et al., 2020). To reduce these biases, we co-registered DEMs from different epochs following Nuth and Kääb (2011) and then removed the residual median vertical offset over stable terrain (Fig. S3). Stable terrain was defined as pixels with slopes below 30° and absolute elevation differences smaller than four times the NMAD (Mattea et al., 2025). Pixels with absolute elevation differences larger than 150 m were excluded (Hugonnet et al., 2021).

We quantified the systematic bias and random errors of DEM differencing using stable, non-glacier areas. For 2002–2014, the mean elevation difference over stable terrain was 0.05 m and the NMAD was 5.66 m. For 2014–2025, the mean elevation difference was –1.68 m and the NMAD was 6.46 m. We modelled the dependence of DEM-differencing errors on terrain slope and maximum absolute curvature. The binned NMAD values were interpolated to generate pixel-level elevation-change uncertainty maps. Standardized elevation differences were then used to model the spatial correlation of DEM errors with a double-range exponential Wevriogram (Hugonnet et al., 2022; Mattea et al., 2025). We also used two open-access glacier surface elevation change datasets that cover the study area as independent references, spanning 2000–2019 (Hugonnet et al., 2021) and 2000–2016 (Brun et al., 2017), as described in Sect. 5.2.

4 Results

4.1 Evolution~~Long-term evolution~~ of Jinwuco and the Jinwu Glacier terminus

~~Since 1990, Jinwuco has undergone sustained expansion, followed by an abrupt decrease after the June 2020 GLOF (Fig. 2). The lake area was 0.371 ± 0.004 km² in 1990 and increased steadily to a maximum of 0.577 ± 0.004 km² in 2019, representing an increase of 0.206 km² (approximately +56%) relative to 1990 (Fig. 3). Over 1990–2019, the mean expansion rate was 6.51×10^{-3} km² a⁻¹. Following~~

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216 the 2020 GLOF, the lake area decreased sharply from $0.577 \pm 0.004 \text{ km}^2$ in 2019 to $0.272 \pm 0.004 \text{ km}^2$
 217 in 2020, a reduction of 0.305 km^2 (approximately 53%). During 2020–2025, the lake area remained
 218 relatively stable at $0.25–0.27 \text{ km}^2$ ($0.251 \pm 0.004 \text{ km}^2$ in 2025). During 1990–2025, the terminus of Jinwu
 219 Glacier showed an overall retreating trend (Figure 2). Measurements along the glacier centerline indicate
 220 that the glacier length decreased from 5699.00 m in 1990 to 4938.41 m in 2025, corresponding to a
 221 cumulative retreat of 760.59 m and a mean retreat rate of 21.73 m a^{-1} (Figure 3a). Before the GLOF, the
 222 glacier terminus retreated continuously. From 1990 to 2019, the cumulative terminus retreat was 702.38
 223 m, with a mean retreat rate of 24.22 m a^{-1} . After the GLOF, changes in terminus position became smaller.
 224 In 2021, the glacier terminus advanced downstream by 23.02 m relative to 2020 (Figure 4). It then
 225 retreated again during 2022–2025, with a cumulative retreat of 33.87 m and a mean retreat rate of 8.47
 226 m a^{-1} .

227 Over the same period, the area of Jinwuco increased from $0.371 \pm 0.004 \text{ km}^2$ in 1990 to $0.577 \pm$
 228 0.004 km^2 in 2019, representing an increase of approximately 56% (Figure 3b). After the 2020 GLOF,
 229 the lake area rapidly decreased to 0.272 km^2 , corresponding to a reduction of approximately 53%,
 230 consistent with the results of Zheng et al. (2021b). During 2020–2025, the lake area remained relatively
 231 stable at $0.25–0.27 \text{ km}^2$, with an area of $0.251 \pm 0.004 \text{ km}^2$ in 2025.

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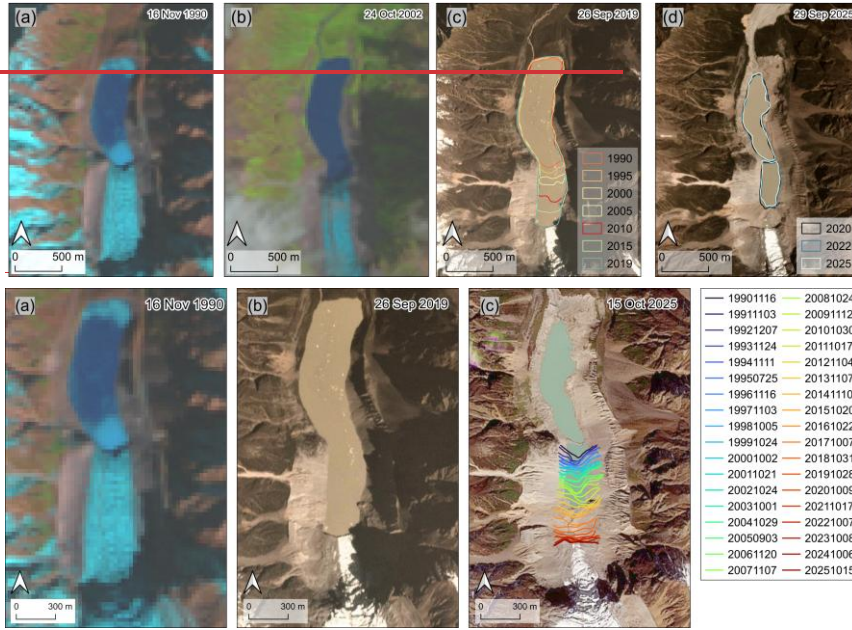


Figure 2 Multi-temporal evolution of the proglacial lake Jinwuoglacier terminal. (a–b) Landsat 5 TM imagery as the basemap acquired on 16 November 1990; (c–d) PlanetScope imagery as the basemap acquired on 26 September 2019 and 15 October 2025, respectively. Colored outlines indicate glacier terminus positions mapped from satellite images between 1990 and 2025 denote lake extents (shorelines) mapped for different years.

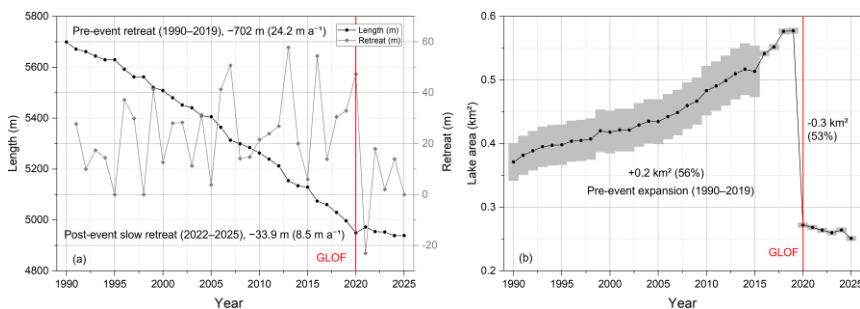
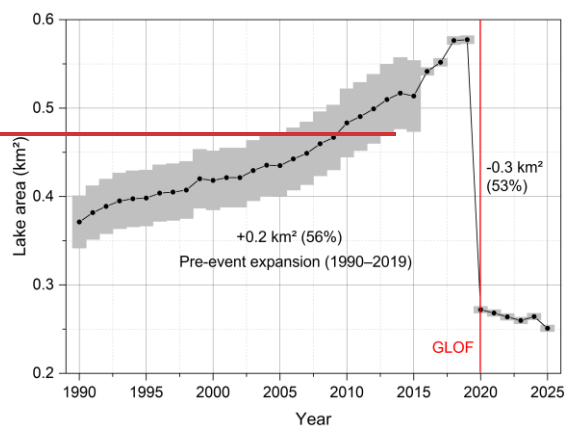


Figure 3 Changes in glacier terminus position and lake area during 1990–2025. (a) Glacier length and annual terminus retreat of Jinwu Glacier measured along the glacier centerline. (b) Lake area changes of Jinwuco during 1990–2025. Black dots denote the mapped lake area for each year. The grey shading indicates the uncertainty of lake area and the uncertainty calculation is described in Sect. 3.1.

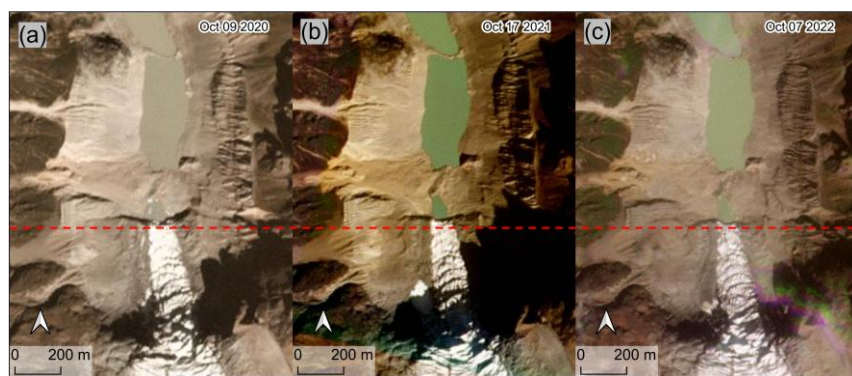


Figure 4 Map of terminus changes of Jinwu Glacier during 2020–2021 based on PlanetScope imagery. The red dashed line is used as a visual reference to facilitate comparison of glacier terminus-position changes among different years.

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4.2 Surface velocity changes after lake disconnection

Following the June 2020 GLOF, ice flow, the near the glacier terminus shifted toward a lower velocity flow regime transitioned to relatively low velocities. However, the dynamic response was clearly lagged, with the main deceleration emerging after 2021 and subsequently stabilizing (Fig. 4Figure 5a,b). Specifically, with in the 0100–200 m section from distance of the glacier terminus, mean ice-flow velocities decreased gradually from $48.43 \pm 2.53 \text{ m a}^{-1}$ in 2017 to $38.53 \pm 3.36 \text{ m a}^{-1}$ in 2020, followed by a marked decline to $27.03 \pm 7.63 \text{ m a}^{-1}$, accompanied by a change in the along-flow velocity gradient (Figure 5b,c). The mean velocity was $42.97 \pm 4.38 \text{ m a}^{-1}$ were $41.82 \pm 7.63 \text{ m a}^{-1}$ during 2017–2020, but decreased markedly to 24.09 ± 6.71 $21.28 \pm 7.48 \text{ m a}^{-1}$ (~4944%) during in 20221–2025 (Fig. 4e). In parallel, PlanetScope imagery during 2020–2021 indicates that the ice-cliff morphology progressively degraded, while the terminus position showed a slight advance (Fig. 5). In the upstream 600650–4550 1600 m section, no pronounced slowdown was observed after the GLOF (Figure 5dFig. 4d); the The mean velocity changed also decreased from $37.55 \pm 9.56 \text{ m a}^{-1}$ $38.25 \pm 9.12 \text{ m a}^{-1}$ during 2017–2021 2020 to 35.24 ± 9.37 $32.87 \pm 8.37 \text{ m a}^{-1}$ during 20222021–2025, representing a 14% reduction of approximately 6%. Over the longer term, ITS_LIVE suggests an overall deceleration of Jinwu Glacier since 2000 (Figure 5c,dFig. 4e, d), although ice speed, the magnitude of deceleration in the downstream 0–200 m zone differs between ITS_LIVE and our measurements. These discrepancies are further discussed in Sect. 5.2.

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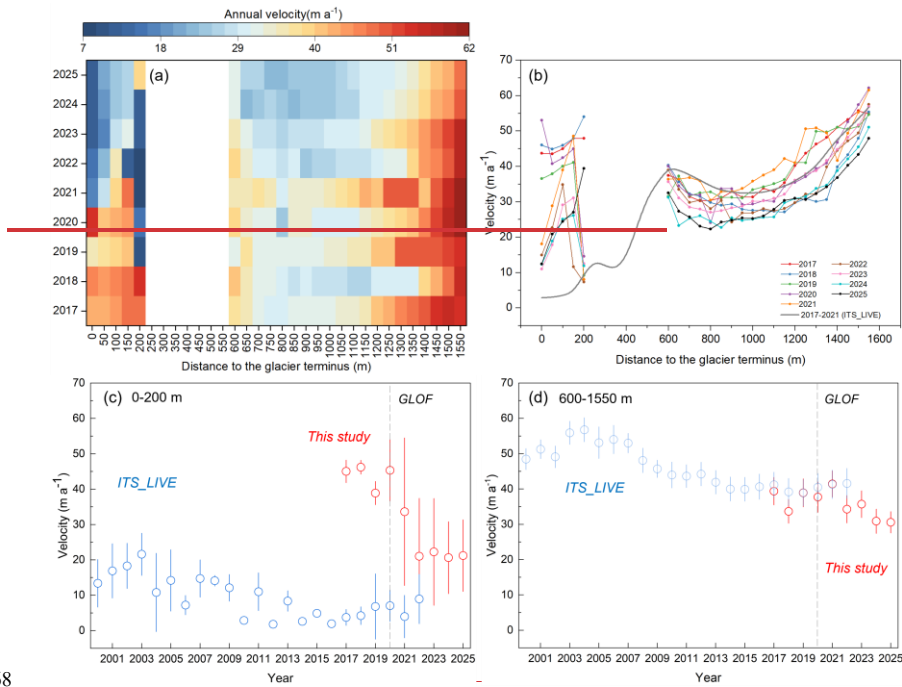
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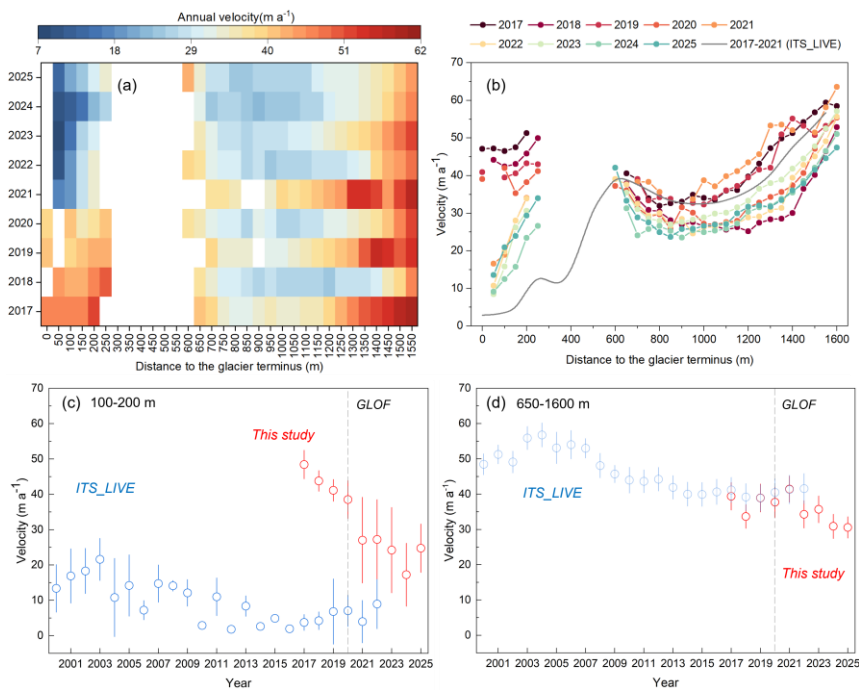


Figure 5

Figure 4 Annual surface ice velocity of Jinwu Glacier along the glacier centerline. (a-b) Spatiotemporal variations in annual velocity along the centerline, calculated using 50 m distance bins. (b) Longitudinal velocity profile derived from PlanetScope feature tracking during 2017–2025, with the ITS LIVE 2017–2021 mean profile shown for comparison. of surface ice velocity along the center flowline (50m bin). Interannual variability in the mean surface velocity within 0–200 m (c) and 600–1550–1600 m (d) from the glacier terminus. Red circles denote results from this study, blue circles denote ITS LIVE data, and the vertical dashed line marks the June 2020 GLOF.

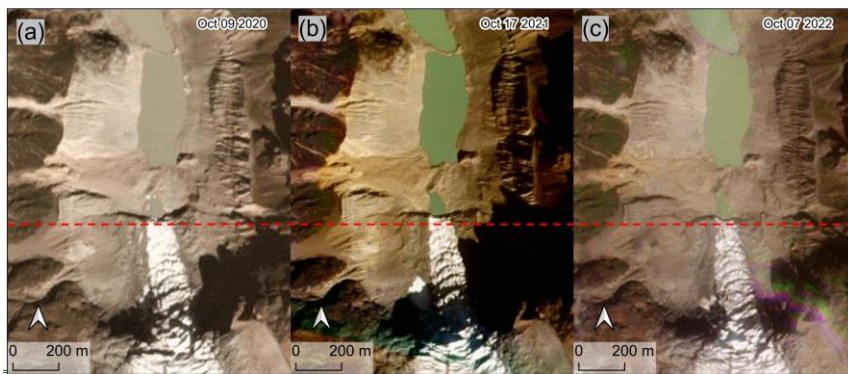


Figure 5 Map of terminus changes of Jinwu Glacier during 2020–2021 based on PlanetScope imagery.

4.3 Multi-year Surface-surface elevation change

Over the past two decades, Jinwu Glacier has experienced sustained surface lowering (Figure 6). Based on DEM differencing sampled at 50 m intervals within a 400 m wide swath, The terminal 0–550 m section showed enhanced thinning, with the mean surface elevation change rate changing from -2.90 m a^{-1} during 2002–2014 to -3.71 m a^{-1} during 2014–2025. By contrast, thinning in the upstream 600–1550 m section slowed, with the mean rate changing from -1.17 to -0.87 m a^{-1} over the same period. This reduction was mainly driven by a slowdown in thinning within the 500–750 m section, where the mean surface elevation change rate changed from -1.20 m a^{-1} during 2002–2014 to -0.57 m a^{-1} during 2014–2025. A similar spatial pattern is also evident in the dataset of Hugonnet et al. (2021) for 2000–2019 and in that of Brun et al. (2017) for 2000–2016 (Figure 6). Notably, relatively large discrepancies are found among the different datasets within the terminal 0–550 m section (Figure 6c), as further discussed in Sect. 5.2.

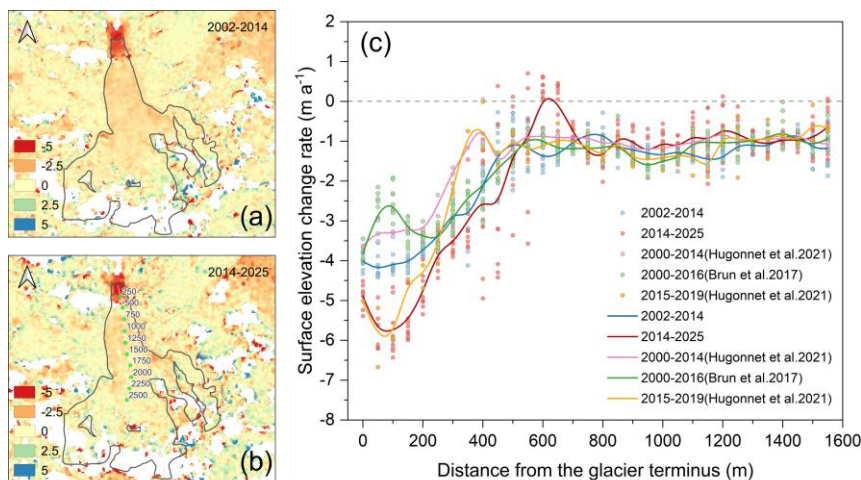


Figure 6 Spatial patterns and longitudinal profiles of surface elevation change rates (dh/dt) of Jinwu Glacier. Maps of dh/dt for the periods 2002–2014 (a) and 2014–2025 (b). (c) shows longitudinal profiles sampled every 50 m along the glacier centerline and averaged within a 400 m-wide swath.

5 Discussion

5.1 ~~Glacier dynamic adjustment following response to~~ GLOF-induced ~~loss of lake contact detachment~~

For Jinwu Glacier, terminus dynamics prior to the GLOF were likely dominated by lake conditions, particularly lake depth (Minowa et al., 2023). Our analysis shows that when lake depth exceeded 10 m (measured after the GLOF) (Zhang et al., 2023a), ice-flow velocity (600–1550 m) was significantly positively correlated with lake depth ($p < 0.001$, $R^2 = 0.78$, Fig. S4). Following the 2020 GLOF, the downstream tongue of Jinwu Glacier showed a clear slowdown, the loss of ice-water contact likely increased basal resistance at the glacier terminus, contributing to its subsequent deceleration (Figure 5a). The rapid loss of lake contact likely reduced hydrostatic pressure and buoyant support at the glacier front, thereby increasing basal and frontal resistive stresses, which contributed to the reduction in ice-flow velocity (Sutherland et al., 2020; Truffer and Motyka, 2016; Benn et al., 2007). After 2021, the near-terminus section maintained a relatively stable low-velocity state. However, a more pronounced and persistent low-velocity state appeared only after 2022 (Figure 5c), suggesting that the glacier terminus likely experienced a rapid but short-lived adjustment phase. In addition, Lowering of the lake level reduces hydrostatic pressure and buoyant support (Benn et al., 2012). The glacier terminus may therefore

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have temporarily maintained, or even locally enhanced, extensional flow. As a result, the deceleration signal may have been delayed. This is consistent with geomorphic evidence from both satellite imagery and field observations. PlanetScope imagery indicates marked degradation of the terminus ice cliff after the GLOF, while the glacier terminus position exhibited a slight advance during 2020–2021 (Figure 4). This slight advance occurred together with ice-cliff degradation observed in PlanetScope imagery and ice-cliff collapse documented by UAV and field investigations on 30 October 2021 (Peng et al., 2023). These observations suggest reduced frontal stability shortly after the loss of lake contact and indicate that the mapped terminus change during 2020–2021 likely reflected short-term frontal morphological adjustment rather than a sustained advance of the glacier. In addition, UAV surveys and field investigations conducted by Peng et al. (2023) in 2021 showed that ice calving occurred at the glacier terminus after the GLOF. These observations indicate that the geometry and stability of the glacier front changed after the drainage event. A comparable response has been reported at a similar nonlinear response has been reported for other lake-terminating glaciers. For example, Longbasaba Glacier, where showed above-average acceleration in 2006 after engineering works lowered the lake level by about 5.6 m in 2005. The glacier subsequently experienced accelerated retreat during 2005–2008 and an increase in ice speed above its long-term mean in 2006 (Liu et al., 2020). Liu et al. (2020) suggested that lake-level lowering may increase the surface slope near the grounding line, enhance grounding-line flux, and promote calving into the lake. Although Jinwu Glacier experienced a more abrupt and complete loss of lake contact, this case illustrates that rapid lake-level lowering can perturb glacier terminus dynamics. Taken together, the loss of lake contact modified the stress balance at the glacier terminus. The removal of hydrostatic pressure and buoyant support likely increased basal and frontal resistive stresses, contributing to the observed near-terminus slowdown, while also reducing the stability of the frontal ice cliff. Ice-cliff degradation, collapse activity, and the short-lived mapped terminus advance during 2020–2021 may therefore reflect the early adjustment of the glacier front to the changed terminus forcing. As the direct lake forcing removed and a new terminus force balance established, subsequent glacier changes may have become increasingly governed by climatic forcing. Although surface-elevation lowering near the terminus (0–200 m) intensified over 2014–2025 (Fig. 6e), the dataset of Hugonnet et al. (2021) indicates that thinning in this sector had already been increasing during 2000–2019 (Fig. S5). Enhanced terminus thinning therefore appears to reflect a longer-term retreat trend, rather than a signal initiated

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only after the GLOF. Furthermore, poor agreement among elevation change datasets in the 0–550 m zone (Sect. 5.2) limits robust assessment of post-GLOF surface elevation changes near the glacier terminus.

Compared with the terminus sector, ~~In contrast,~~ surface elevation changes in the upstream sector (600–1550 m) are comparatively more robust reliably (Sect. 5.2) ~~assessed due to better agreement among different datasets (Fig. 6c).~~ Within the 500–750 m reach, ~~our results show that~~ surface lowering slowed from -1.20 m a^{-1} during 2002–2014 to -0.57 m a^{-1} during 2014–2025. Hugonnet et al. (2021) reported a rate of -1.09 m a^{-1} for the same reach during 2015–2019. ~~Taken together, these results suggest that surface lowering in this reach may have slowed further after the GLOF, with a~~ A simple time weighted calculation combining these estimates yields an estimate of approximately lowering rate of -0.14 m a^{-1} for 2020–2025, although this value should be regarded as an indirect estimate rather than a direct DEM-differencing result. Ice-flow velocity in the upstream 600–1550 m sector did not show a pronounced post-GLOF change ~~decreased after the flood (Figure 5d), but the magnitude of the reduction~~ ($\sim 14\%$) was smaller than that at the glacier terminus ($\sim 49\%$). The much stronger slowdown near the glacier terminus, compared with the weak velocity change upstream, indicates that the along-flow velocity gradient changed after the GLOF. This may have reduced longitudinal stretching in the 500–750 m reach and partly contributed to the observed weakening of surface lowering.

As terminus flow slowed, the downstream ice-flux demand imposed on the upstream ice flow likely decreased, favouring the development of compressive flow in the 500–750 m section. Additionally, the reduced interannual variability in velocity suggests that, following the disappearance of the proglacial lake, the glacier terminus became more stable. After a short period of dynamic adjustment, glacier changes may have gradually shifted back toward being primarily controlled by climatic forcing.

Proglacial lakes are commonly regarded as a significant non-climatic forcing factor influencing lake-terminating glaciers (Carrivick and Tweed, 2013). In contrast, a GLOF represents an event-driven ~~is a sudden~~ perturbation that can abruptly rapidly alter ~~reset~~ glacier terminus boundary conditions by lowering lake level and removing ice–lake contact. The ~~lagged and spatially variable~~ response of Jinwu Glacier further suggests that such events can modify longitudinal velocity structure coupling, and dynamic stress coupling ~~transfer, and ice-flux demand~~ between the upstream and downstream sections. Given the increasing hazard potential of ~~As moraine-dammed GLOFs,~~ moraine-dammed glacial lakes become more frequent under ongoing climate warming (Zheng et al., 2021c; Zhang et al., 2023b; Zhang

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et al., 2025)(Zhang et al., 2025), GLOFs they should be considered regarded as event-driven processes in future assessments of lake-terminating glacier dynamics and associated hazards.

5.2 Uncertainties and limitations analysis

In this study, uncertainties in surface velocity and elevation change were assessed using from residual statistics over stable terrain adjacent to the glacier, which allowed thereby quantifying both systematic and random errors to be quantified. Nevertheless, these such estimates do cannot account for all potential error sources. In the case of For surface velocity, image velocity, image matching can be degraded by repetitive or self-similar glacier surface features, such as crevasses, seracs, and landslide or avalanche deposits, which can may increase noise and produce mismatches (Paul et al., 2015). For surface elevation change, additional uncertainties may arise from terrain dependent biases linked to slope and curvature, as well as from residual co-registration errors (Li and Hajnsek, 2025; Nuth and Kääb, 2011; Hugonnet et al., 2022)(Li and Hajnsek, 2025; Nuth and Kääb, 2011).

A high noise region is observed in In the velocity field, relatively high noise occurs within 200 to 450 m from of the glacier terminus. This reach coincides with likely because image matching is degraded by an icefall (Figure 1e) and dense crevassing (Figure 4-), where image matching is likely less stable. These conditions may reduce the stability of image matching. Moreover, Jinwu Glacier is north facing, and this reach is more frequently affected by topographic shadowing (Fig. S1). Although we prioritized image pairs with minimal shadow influence, residual shadow effects are difficult to fully eliminate. This may partly explain the further elevated the noise level in this area and could also explain the lower ITS_LIVE velocities observed there (Figure 5a). To further constrain the accuracy of our velocity estimates, we performed an independent validation based on manual interpretation (Sato et al., 2022; Hyde et al., 2025). WWee selected 36 clearly identifiable tie points on the image pairs (Fig. S6S5). The validation shows a root mean square error (RMSE) of 3.61 m a⁻¹ between the PlanetScope velocities and the manually derived estimates (Fig. S7S6). Two targeted validations further confirmed the robustness of the centerline velocity profiles and the near-terminus slowdown signal. Along the lower glacier 0-500m tongue, 20 repeatedly traceable points covering the 2017–2025 annual velocity fields showed strong agreement between manual and automatic estimates, with an R² of 0.92 and an RMSE of 3.84 m a⁻¹ (Fig. S7a, b). In the near-terminus reach where the most pronounced slowdown was observed, nine additional validation points showed even higher consistency, with an R² of 0.96 and an RMSE of 2.14 m

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a⁻¹ (Fig. S7c, d). These validations suggest that, despite locally elevated noise in parts of the velocity field, the main velocity patterns reported in this study, especially the slowdown in the near-terminus section (Fig. S8), are robust. In addition, we tested the robustness of the near-terminus velocity-gradient change using manually identifiable feature points along the glacier centerline from the 2017–2018 and 2023–2024 image pairs (Fig. S9). Before the GLOF, velocities were relatively high and varied only slightly along the centerline, whereas after the GLOF they clearly decreased toward the glacier terminus. This distinct near-terminus deceleration gradient supports stronger resistive effects after lake drainage and provides direct evidence that the loss of lake–ice contact affected the near-terminus dynamic adjustment.

Surface elevation change estimates from different data sources show relatively poor agreement in the terminus sector (0–550 m, Figure 6c). The discrepancy with Brun et al. (2017) is mainly expressed in the elevation-change gradient within the 0–200-m zone, whereas that with Hugonnet et al. (2021) is primarily reflected in the thinning magnitude across the 0–550-m sector. To quantify the consistency among datasets, we calculated the pairwise RMSE between elevation-change profiles at corresponding centerline sampling points and used the mean standard deviation among the profiles to represent their overall spread. Quantitatively, pairwise RMSE values within 0–550 m from 0.75 to 0.97 m a⁻¹, and the mean standard deviation among the three profiles is 0.532 m a⁻¹. By contrast, agreement is substantially improved in the upstream sector (600–1550 m), where RMSE decreases to 0.18–0.29 m a⁻¹ and the mean standard deviation drops to 0.162 m a⁻¹. This contrast indicates that surface elevation change estimates are subject to markedly greater uncertainty near the glacier terminus, whereas estimates from the upstream sector are comparatively robust. A likely explanation for this discrepancy is the complex and steep topographic setting of the Jinwu Glacier terminus. DEM elevation errors generally increase with slope (Hugonnet et al., 2022), and residual horizontal offsets in DEM differencing can be amplified over steep terrain, where they are translated into larger apparent elevation differences (Nuth and Kääb, 2011; Li and Hajnsek, 2025). In addition, the highly crevassed surface, pronounced local shadowing, and complex surface texture of the terminus may degrade DEM extraction, image matching and DEM co-registration quality (Berthier et al., 2014). These effects likely accumulate in the terminus sector and reduce the consistency of elevation-change estimates among datasets. Importantly, although different datasets yield somewhat different thinning magnitudes near the terminus sector, they consistently indicate substantial surface lowering. Therefore, This suggests that the conclusion of pronounced thinning near the glacier terminus area is robust, but it remains difficult to isolate the surface

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elevation change specifically induced by the GLOF even though the exact magnitude of thinning should be interpreted with caution. We further assessed the statistical uncertainty of our DEM-differencing results using stable, non-glacier terrain (Hugonnet et al., 2022; Mattea et al., 2025) (Fig. S10). The uncertainty is generally larger in the terminus sector than in the upstream section, consistent with the stronger disagreement among independent datasets near the glacier front. In the 500–750 m section where thinning weakened most clearly, the uncertainty ranges of the two periods are largely separated, suggesting that the reduced thinning signal is unlikely to be caused solely by DEM-differencing uncertainty.

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6 Conclusion

Using historical remote sensing data, we reconstructed the dynamic evolution of Jinwu Glacier in southeastern Tibet, China, during its transitioned from a lake-terminating to a land-terminating glacier following the glacial lake outburst flood (GLOF) on 26 June 2020. Annual surface velocities were derived from multi-temporal PlanetScope imagery (2017–2025) using feature tracking, and surface elevation changes were quantified from ASTER stereo DEM differencing for the periods 2002–2014 and 2014–2025.

Our results show that the glacier tongue experienced clear changes in ice flow and frontal morphology after the GLOF. The near-terminus region showed a clear slowdown, accompanied by a change in the along-flow velocity gradient. After 2021, this region remained in a relatively stable low-velocity state. The GLOF-driven change in glacier terminus boundary conditions triggered a pronounced but lagged dynamic response. Ice-flow velocities within 100–200 m from the glacier terminus decreased from $42.97 \pm 4.38 \text{ m a}^{-1}$ during 2017–2020 to $24.09 \pm 6.71 \text{ m a}^{-1}$ during 2021–2025, significantly after 2021, corresponding to with an overall reduction of approximately 44.49%. During 2020–2021, the glacier front was characterized by ice-cliff collapse and degradation, together with a slight terminus advance of approximately 23 m. In contrast, the upstream showed no pronounced velocity change. The mean velocity in the 650–1600 m section changed only slightly from $37.55 \pm 9.56 \text{ m a}^{-1}$ during 2017–2020 to $35.24 \pm 9.37 \text{ m a}^{-1}$ during 2021–2025.

Velocities within 600–1550 m also declined, but the magnitude of change was smaller (approximately 14%). Surface elevation changes reveal strong spatial contrasts. Thinning within 0–200

m of the terminus intensified after 2014, whereas thinning within 600–1550 m slowed. Notably, a slight surface elevation increase was observed in the topographic transition zone at 500–750 m. In addition, the glacier terminus showed slight advance and local ice calving after the GLOF. Following the loss of proglacial lake influence, the glacier terminus underwent rapid geometric and stress adjustment. Near the terminus, the adjustment likely involved enhanced longitudinal extension. Farther upstream, reduced ice-flow velocity and weakened downstream ice flux demand may have promoted a shift toward a more compressive strain regime. After this short-lived dynamic adjustment, subsequent glacier change may have become increasingly governed by climatic forcing.

Overall, our study provides the first quantitative constraints on the glacier response to a GLOF-induced loss of lake contact-driven reset of terminus boundary conditions, advancing understanding of lake-terminating glacier behaviour and its future evolution under uncertain climate scenarios. The results highlight that event-driven glacial lake drainage can substantially modify glacier terminus dynamics and should be considered in future assessments of lake-terminating glacier evolution and associated hazards.

Data availability

The Landsat TM/ETM+/OLI imagery used in this study is publicly available from the U.S. Geological Survey (USGS) archive (<https://www.usgs.gov/landsat-missions/landsat-data-access>, last access: 24 January 2026). ASTER L1A stereo data are publicly available from the NASA Land Processes Distributed Active Archive Center (LP DAAC; <https://www.earthdata.nasa.gov/data/catalog/lpcloud-ast-l1a-004>, last access: 11 February 2026). The SRTM DEM is publicly available from the USGS/NASA archive (<https://www.usgs.gov/centers/eros/science/usgs-eros-archive-digital-elevation-shuttle-radar-topography-mission-srtm>, last access: 3 February 2026). PlanetScope imagery was provided by Planet Labs Inc. through the Education and Research Programme and accessed via Planet Explorer (<https://www.planet.com/>, last access: 27 January 2026). The Hugonnet et al. (2021) dataset is available at <https://www.sedoo.fr/theia-publication-products/?uuid=c428c5b9-df8f-4f86-9b75-e04c778e29b9> (last access: 16 February 2026), and the Brun et al. (2017) dataset is available at <https://doi.org/10.1594/PANGAEA.876545> (last access: 29 January 2026). The derived datasets supporting the main findings of this study, including proglacial lake outlines from 1990 to 2025, glacier

485 surface velocity data from 2017 to 2025, and DEMs covering the glacierized area, are openly available
486 at <https://doi.org/10.5281/zenodo.19450085> (Luo, 2026).

487 **Author contributions**

488 Y.L. designed the study, developed the methodology, performed the analysis, and wrote the manuscript.
489 Q.L. provided funding, support, and supervision, and contributed to writing and revising the manuscript.
490 Y.F. contributed to data production and analysis, and to writing and revising the manuscript. Xue.L.,
491 Y.Y., J.Y., and G.Z. contributed to data production and analysis. Xuy.L. contributed to discussion and
492 manuscript revision. All authors contributed to the final version of the paper.

493 **Competing interests**

494 The contact author has declared that neither they nor their co-authors have any competing interests.

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