



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

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Abstract. Proglacial lakes 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 of Jinwu Glacier in southeastern 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 100–200 m of 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. In contrast, the glacier tongue upstream of the icefall 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. The icefall may limit dynamic coupling between the upstream section and the terminus. These results show that GLOF-induced lake disconnection can substantially alter ice flow and frontal morphology near the glacier terminus. This study provides the first quantitative evidence of glacier dynamic adjustment following a GLOF driven transition from lake- to land-terminating conditions.

1 Introduction

Since the 20th century, most glaciers on 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 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 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 glacier retreat exposes deeply carved subglacial depressions that progressively store water (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 overdeepened basin completely, leading to a progressive separation between the lake and the glacier terminus. 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 but strong perturbation to the terminus boundary conditions, yet the glacier dynamic response remains poorly understood. A previous 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 deceleration (Liu et al., 2020). This suggests that abrupt changes in terminus boundary conditions can induce strong 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. We then quantify how the GLOF driven step change in terminus boundary conditions affected glacier dynamics. Finally,

we discuss the glacier dynamic response 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 (Fig. 1a and b). The glacier lies on the northern flank of Nenang Peak, the highest summit in the basin (Fig. 1c). The Nidou Zangbo basin covers $\sim 1268 \text{ km}^2$ (Lehner and Grill, 2013). Elevations generally decrease from west to 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). The basin contains 343 glaciers with a total area of $\sim 385 \text{ km}^2$ (RGI Consortium, 2023). In 2018, 39 glacial lakes were mapped with a total area of $\sim 3.0 \text{ km}^2$ (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 $\sim 0.03 \text{ }^\circ\text{C a}^{-1}$ ($p = 0.01$, $R^2 = 0.16$; Fig. 1d). Jinwu Glacier is oriented approximately north to south and covers $\sim 8.16 \text{ km}^2$ in 2025. Its lower tongue is partly debris covered and includes a pronounced icefall. The icefall has an elevation drop of $\sim 300 \text{ m}$, and the maximum slope exceeds 35° (Fig. 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 $\sim 0.58 \text{ km}^2$ and its straight-line length was $\sim 1.8 \text{ km}$, 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.

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 (Table S1 in the

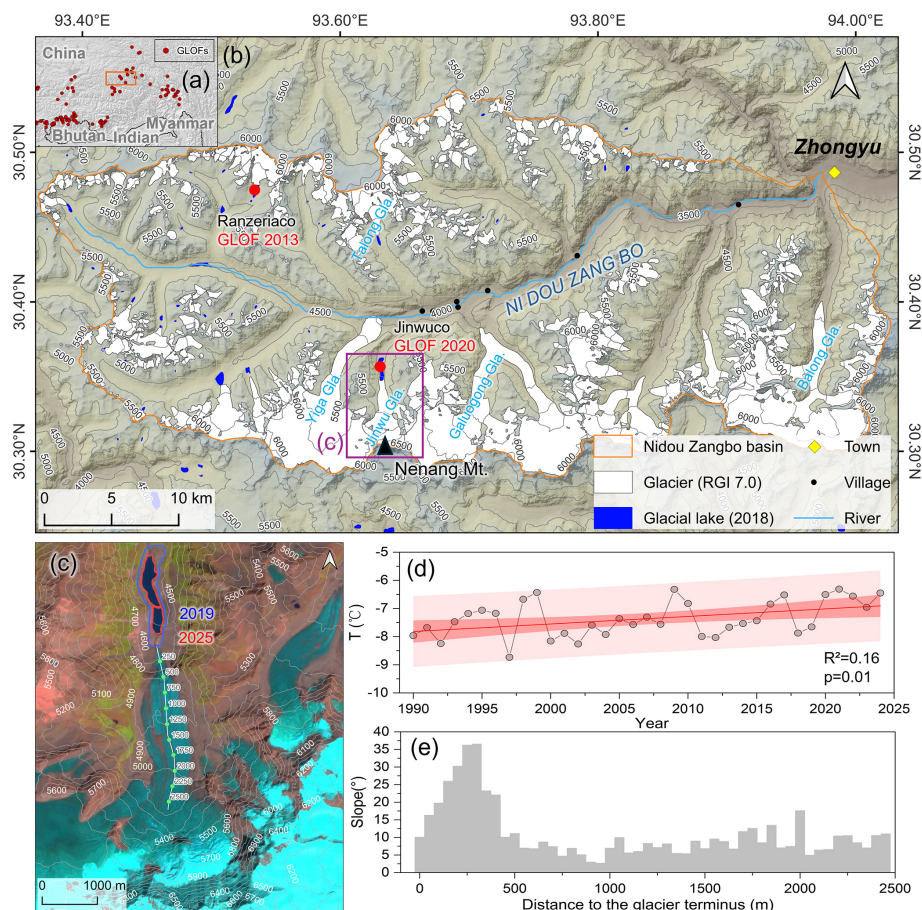


Figure 1. Study area and basic geographic setting of Jinwu Glacier in southeastern Tibet. (a) Distribution of reported glacial lake outburst floods (GLOFs) in southeastern Tibet (Lützow et al., 2023) and the location of the Nidou Zangbo basin. The basemap in panel (a) is Esri World_Shaded_Relief (Powered 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 Jinwu 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).

Supplement). 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 S2) 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, terrain shadows often obscure large parts of the lower glacier (Fig. S1 in the Supplement), 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 $150\text{ m} \times 150\text{ m}$ (50×50 pixels) with a step size of $30\text{ m} \times 30\text{ m}$ (10×10 pixels). The search range was set to $90\text{ m} \times 90\text{ m}$ (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 $< 30^\circ$ (Fig. S2 in the Supplement). 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 Eqs. (2) and (3):

$$D = \sqrt{D_x^2 + D_y^2} \quad (2)$$

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

where D is the displacement measured between each image pair, D_x and D_y are the east–west and north–south displacement components (m), respectively, N is the time interval between the image pairs (days), and A is the number of days per year (da^{-1}), taken as 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. For each window, the centre pixel was identified as an outlier and removed when its absolute deviation from the local median exceeded 1.5 times the median absolute deviation (MAD). 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 S2). 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., 2017a; Bhattacharya et al., 2021). We produced three DEMs for the 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 in the Supplement). 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. Elevation differences over stable terrain were grouped into two-dimensional slope–curvature bins, and the NMAD was calculated within each bin. The binned NMAD values were interpolated to generate pixel-level elevation-change uncertainty maps. Elevation differences were standardized by these local uncertainties, and an empirical variogram was calculated from the standardized residuals and fitted with a double-range exponential model (Fig. S4 in the Supplement). The fitted variogram was used to estimate the effective number of independent samples, which was combined with the pixel-level uncertainties to estimate the uncertainty of spatially averaged elevation changes (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., 2017a), as described in Sect. 5.2.

4 Results

4.1 Evolution of Jinwuco and the Jinwu Glacier terminus

During 1990–2025, the terminus of Jinwu Glacier showed an overall retreating trend (Fig. 2). Measurements along the

glacier centerline indicate that the glacier length decreased from 5699.00 m in 1990 to 4938.41 m in 2025, corresponding to a cumulative retreat of 760.59 m and a mean retreat rate of 21.73 m a^{-1} (Fig. 3a). Before the GLOF, the glacier terminus retreated continuously. From 1990 to 2019, the cumulative terminus retreat was 702.38 m, with a mean retreat rate of 24.22 m a^{-1} . After the GLOF, changes in terminus position became smaller. In 2021, the glacier terminus advanced downstream by 23.02 m relative to 2020 (Fig. 4). It then retreated again during 2022–2025, with a cumulative retreat of 33.87 m and a mean retreat rate of 8.47 m a^{-1} .

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

4.2 Surface velocity changes after lake disconnection

Following the June 2020 GLOF, ice flow near the glacier terminus shifted toward a lower velocity regime (Fig. 5a). In the 100–200 m section from 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 (Fig. 5b and c). The mean velocity was $42.97 \pm 4.38 \text{ m a}^{-1}$ during 2017–2020, but decreased to $24.09 \pm 6.71 \text{ m a}^{-1}$ ($\sim 44\%$) during 2021–2025. In the upstream 650–1600 m section, no pronounced slowdown was observed after the GLOF (Fig. 5d). The mean velocity changed 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, representing a reduction of approximately 6 %. Over the longer term, ITS_LIVE suggests an overall deceleration of Jinwu Glacier since 2000 (Fig. 5c and d), although ice speed in the downstream differs between ITS_LIVE and our measurements. These discrepancies are further discussed in Sect. 5.2.

4.3 Multi-year surface elevation change

Over the past two decades, Jinwu Glacier has experienced sustained surface lowering (Fig. 6). 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. (2017a) for 2000–2016 (Fig. 6). Notably, relatively large discrepancies are found among the different datasets within the terminal 0–550 m section (Fig. 6c), as further discussed in Sect. 5.2.

5 Discussion

5.1 Dynamic adjustment following GLOF-induced loss of lake contact

Following the 2020 GLOF, the downstream tongue of Jinwu Glacier showed a clear slowdown (Fig. 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 (Fig. 5c). In addition, PlanetScope imagery indicates degradation of the terminus ice cliff after the GLOF, while the glacier terminus position exhibited a slight advance during 2020–2021 (Fig. 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. These observations indicate that the geometry and stability of the glacier front changed after the drainage event. A comparable response has been reported at Longbasaba Glacier, where 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 was removed and a new terminus force balance established,

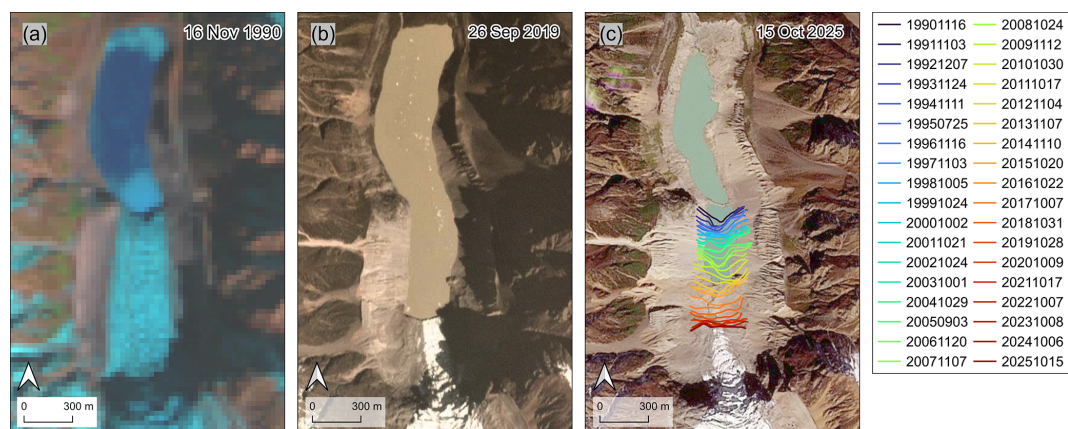


Figure 2. Multi-temporal evolution of the glacier terminus. (a) Landsat 5 TM imagery as the basemap acquired on 16 November 1990; (b, c) PlanetScope imagery as the basemap acquired on 26 September 2019 and 15 October 2025, respectively. Colored lines indicate glacier terminus positions mapped from satellite images between 1990 and 2025.

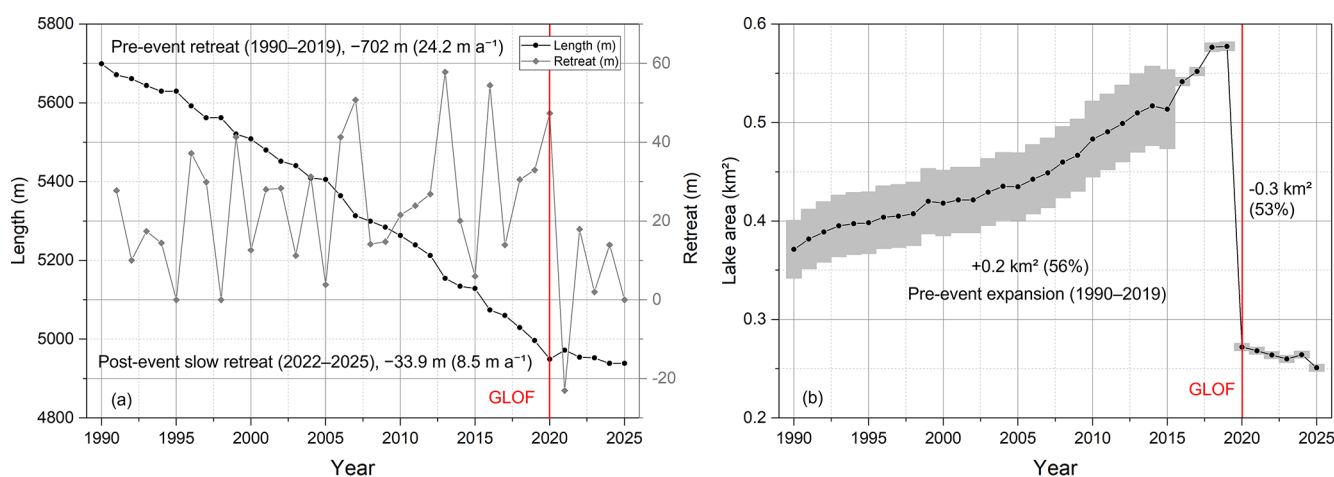


Figure 3. Changes in glacier terminus position and lake area during 1990–2025. (a) Glacier length and annual terminus retreat of Jinwu Glacier measured along the glacier centerline. (b) Lake area changes of Jinwu during 1990–2025. Black dots denote the mapped lake area for each year. The grey shading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

subsequent glacier changes may have become increasingly governed by climatic forcing.

Surface elevation changes in the upstream sector are more robust than those near the terminus (Sect. 5.2). Within the 500–750 m reach, 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 (Fig. S5 in the Supplement). Ice-flow velocity in the upstream sector remained broadly stable after the GLOF (Fig. 5d). An icefall separates the near-terminus region from the upstream glacier tongue. Previous studies have shown that steep icefalls and abrupt changes in ice thickness or bed geometry may reduce longitudinal stress transmission between glacier sections and limit the upstream propagation of dynamic perturbations originating near the terminus (McNabb et al., 2012;

Durkin et al., 2017). Observations at Yachtse Glacier showed that interannual dynamic changes originating near the terminus were strongly attenuated above an icefall, suggesting that the icefall may inhibit upstream transmission of stress perturbations (Durkin et al., 2017). Regional observations have also shown that the upstream limit of seasonal velocity variations often coincides with the base of icefalls, suggesting that icefalls may separate glacier sections with different dynamic responses (Armstrong et al., 2017). Jinwu Glacier shows a spatial pattern consistent with these observations. The near-terminus section below the icefall (100–200 m) experienced a pronounced slowdown, whereas mean velocity within the 650–1600 m section upstream of the icefall changed only slightly. Given the potential for dynamic separation across the icefall, the upstream velocity record is therefore insufficient to infer a direct response to the GLOF.

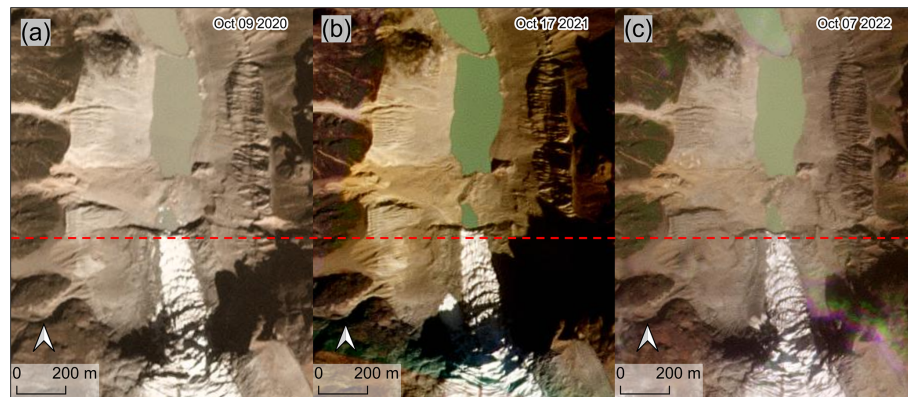


Figure 4. Map of terminus changes of Jinwu Glacier during 2020–2022 based on PlanetScope imagery. The red dashed line is used as a visual reference to facilitate comparison of glacier terminus-position changes among different years.

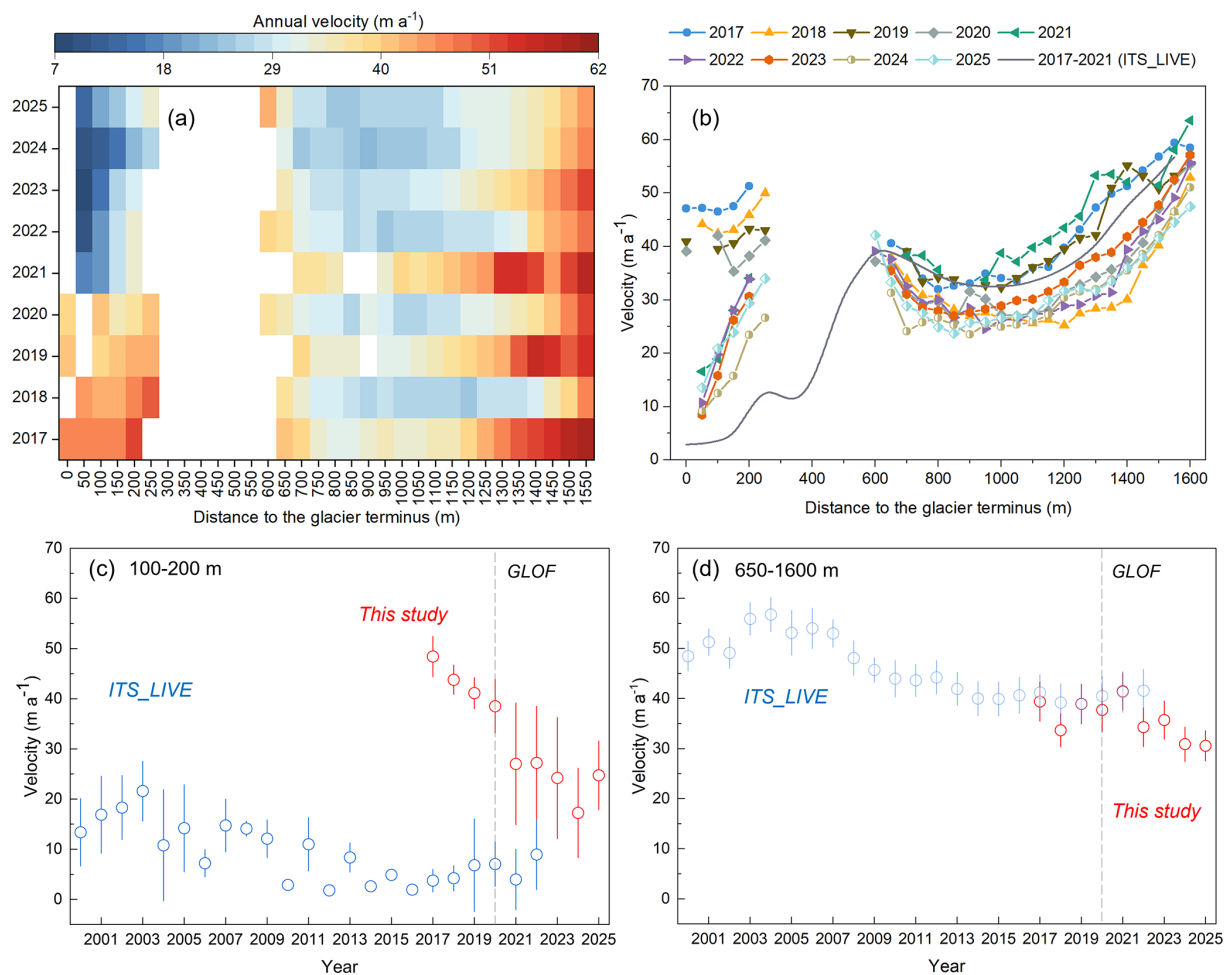


Figure 5. Annual surface ice velocity of Jinwu Glacier along the glacier centerline. **(a)** Spatiotemporal variations in annual velocity along the centerline, calculated using 50 m distance bins. **(b)** Longitudinal velocity profile derived from PlanetScope feature tracking during 2017–2025, with the ITS_LIVE 2017–2021 mean profile shown for comparison. Interannual variability in the mean surface velocity within 100–200 m **(c)** and 650–1600 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.

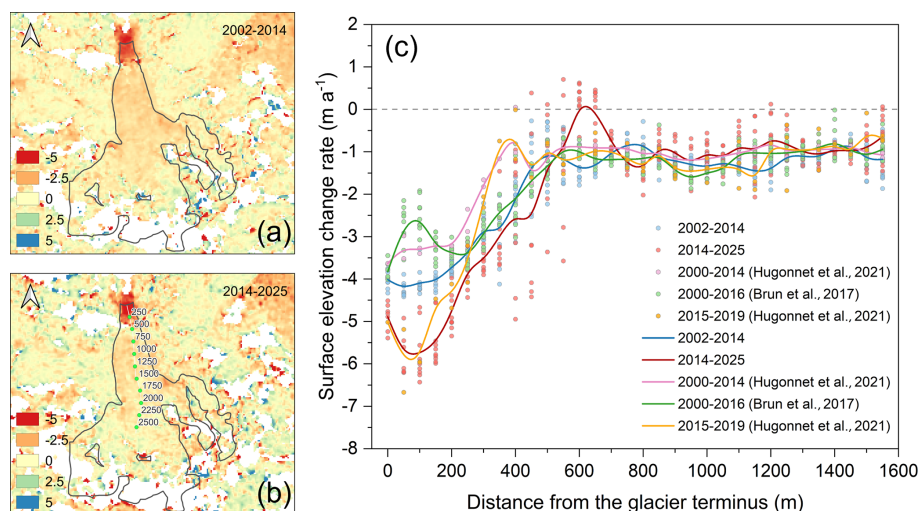


Figure 6. Spatial patterns and longitudinal profiles of surface elevation change rates (dh/dt) of Jinwu Glacier. Maps of dh/dt for the periods 2002–2014 (a) and 2014–2025 (b). (c) shows longitudinal profiles sampled every 50 m along the glacier centerline and averaged within a 400 m-wide swath.

Because the 2014–2025 DEM-differencing period spans several years both before and after the GLOF, the slower thinning in the 500–750 m section cannot be directly attributed to either the GLOF or the near-terminus slowdown.

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 perturbation that can abruptly alter glacier terminus boundary conditions by lowering lake level and removing ice–lake contact. The response of Jinwu Glacier further suggests that such events can modify the velocity structure and frontal morphology of the terminus sector. Given the increasing hazard potential of moraine-dammed glacial lakes under ongoing climate warming (Zheng et al., 2021c; Zhang et al., 2023b; Zhang et al., 2025), GLOFs should be considered event-driven processes in future assessments of lake-terminating glacier dynamics and associated hazards.

5.2 Uncertainties and limitations

In this study, uncertainties in surface velocity and elevation change were assessed using residual statistics over stable terrain adjacent to the glacier, which allowed both systematic and random errors to be quantified. Nevertheless, these estimates cannot account for all potential error sources. For surface 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 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).

In the velocity field, relatively high noise occurs within 200–450 m from the glacier terminus. This reach coincides with an icefall (Fig. 1e) and dense crevassing (Fig. 4), where image matching is likely less stable. Moreover, Jinwu Glacier is north facing, and this reach is more frequently affected by topographic shadowing (Fig. S1). This may partly explain the elevated noise in this area and the lower ITS_LIVE velocities observed there (Fig. 5b). To constrain the accuracy of velocity estimates, we performed an independent validation based on manual interpretation (Sato et al., 2022; Hyde et al., 2025). We selected 36 clearly identifiable tie points on the image pairs (Fig. S6 in the Supplement). 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. S7 in the Supplement). Two targeted validations further confirmed the robustness of the centerline velocity profiles and the near-terminus slowdown signal. Along the lower 0–500 m glacier tongue, 20 repeatedly traceable points covering the 2017–2025 annual velocity fields showed strong agreement between manual and automatic estimates, with an R^2 of 0.92 and an RMSE of 3.84 m a^{-1} (Fig. S8a and b in the Supplement). In the near-terminus reach where the most pronounced slowdown was observed, nine additional validation points showed even higher consistency, with an R^2 of 0.96 and an RMSE of 2.14 m a^{-1} (Fig. S8c and 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. S9 in the Supplement), 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. S10 in the Supplement). 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, Fig. 6c). 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. Pairwise RMSE values within 0–550 m range 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 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. Although different datasets yield somewhat different thinning near the terminus, they consistently indicate substantial surface lowering. Therefore, the conclusion of pronounced thinning near the glacier terminus is robust. However, because the 2014–2025 DEM-differencing interval spans several years both before and after the 2020 GLOF, it cannot isolate an event-scale elevation response. We use the DEM-derived elevation changes to characterize the longer-term thinning pattern of Jinwu Glacier. 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. S11 in the Supplement). 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.

6 Conclusion

Using historical remote sensing data, we analyzed the dynamic evolution of Jinwu Glacier in southeastern Tibet, China, during its transition from a lake-terminating to a land-terminating glacier following the GLOF on 26 June 2020.

The glacier tongue experienced clear changes in ice flow and frontal morphology after the GLOF. The near-terminus region showed a slowdown of approximately 44 %, accompanied by a change in the along-flow velocity gradient, ice-cliff collapse and a terminus advance of approximately 23 m. In contrast, the glacier upstream of the icefall showed no pronounced velocity change. The observed post-GLOF dynamic adjustment was concentrated in the terminus sector below the icefall. The icefall may have limited the upstream transmission of terminus perturbations.

Our study provides the first quantitative analysis on the glacier response to a GLOF-induced loss of lake contact, advancing understanding of lake-terminating glacier behaviour and its future evolution. 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-11a-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. (2017a) dataset is available at <https://doi.org/10.1594/PANGAEA.876545> (Brun et al., 2017b). The derived datasets supporting the main findings of this study, including proglacial lake outlines from 1990 to 2025, glacier surface velocity data from 2017 to 2025, and DEMs covering the glacierized area, are openly available at <https://doi.org/10.5281/zenodo.23190897> (Luo, 2026).

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