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Pre- and co-seismic stress loading promoted low-angle splay fault during the 2025 M_w 7.1 Tingri earthquake

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Major ($M \geq 7$) continental normal-fault earthquakes are rare, and the potential for low-angle ($< 30^\circ$) fault slip in such events remains highly debated. The 2025 M_w 7.1 Tingri earthquake provides a pivotal case study. Here our Bayesian inversion of geodetic data reveals complex slip on a west-dipping main fault dipping at $45\text{--}55^\circ$ and a synthetic, low-angle splay fault dipping at 27° . Aftershock seismicity confirms the coexistence of both steeply and shallowly dipping fault segments. Coulomb stress modeling shows the main fault slip partially loaded the splay, while stress inversion of regional historical seismicity indicates a pre-stress regime that promoted instability on the low-angle splay. This demonstrates a cascading process, enabled by the pre-stress environment and co-seismic stress transfer, which together facilitated the low-angle faulting. Our findings suggest the mechanical viability of low-angle faulting in major normal-fault earthquakes, refining models of fault activation and seismic rupture dynamics in continental extensional regions.

Major ($M_w \geq 7$) continental normal-fault earthquakes are uncommon, yet their rupture dynamics carry outsized implications for seismic hazard in extensional regions. Normal-fault systems are typically segmented, exhibiting variable dips and linked synthetic–antithetic pairs, as documented by geomorphology, seismic imaging, and paleoseismology^{1–4}. While the existence of low-angle normal faults (LANFs; dip $< 30^\circ$) in continental extensional regions is well-documented, the mechanics of their formation remain controversial. Classic Andersonian fault theory predicts that such shallow-dipping faults should remain mechanically locked. There is also a notable scarcity of observed LANF slip during continental normal-fault earthquakes. Consequently, LANFs are often interpreted as the result of long-term rotation of originally high-angle faults⁵. A central unresolved issue concerns the conditions under which LANF segments slip during major earthquakes, as well as the role of multi-segment interactions in modulating rupture propagation and aftershock distributions^{6,7}. Recent dynamic-rupture models of an M_w 7 LANF earthquake indicate that shallowly-dipping synthetic splays not only host greater co-seismic slip but also more effectively arrest shallow rupture on the main fault compared to steeper antithetic splays⁸. This rupture dynamic impacts seismic hazard, yet it remains difficult to verify with modern, well-recorded continental events.

Historical records include a few major ($M_w > 7$) continental normal-fault earthquakes, such as the 1887 M_w 7.5 Sonora⁹, 1915 M_w 7.3 Pleasant Valley¹⁰, 1954 M_w 7.1 Fairview Peak¹¹, and 1959 M_w 7.3 Hebgen Lake,

Montana¹² events. Over the past ~30 years, high-resolution space-based geodetic measurements, such as Interferometric Synthetic Aperture Radar (InSAR) and optical remote sensing, have enabled detailed inversions of segmented ruptures. During this period, continental strike-slip or thrust earthquakes have frequently exceeded M_w 7.5^{13–16}, whereas major continental normal-fault earthquakes are rare, and their maximum magnitude appears mechanically constrained to the lower M_w 7 range^{17,18}. Representative examples include the 2008 M_w 7.1 Yutian earthquake in northwest Tibet (Fig. 1a), where InSAR inversions revealed variable-dip segments within a north–south trending graben¹⁹, and the 2016 M_w 6.6 Mt. Vettore–Mt. Bove event in central Italy, which ruptured antithetic splays resolved by joint InSAR, GNSS, and strong-motion data²⁰. The scarcity of LANFs in continental normal-fault earthquakes remains notable. This may be partly because multi-fault ruptures can mask LANF slip in seismic nodal-plane solutions, as energy from shallow-dipping segments is overprinted by concurrent slip on steeper faults. Alternatively, LANF slip may also occur in early post-seismic phases when a LANF is triggered as a secondary structure.

The prolonged northward movement of the Indian plate over tens of millions of years led to its tectonic collision with Eurasia, forming the Tibetan–Himalayan orogen²¹. While Cenozoic crustal shortening dominates due to ongoing India–Eurasia convergence, extensional structures have emerged as a significant deformation feature within the orogen²². Among these, the prominent N–S trending rifts represent the most

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distinctive active extensional tectonics (Fig. 1a). On 7 January 2025, at 01:05:16 UTC, a destructive M_w 7.1 normal-fault earthquake struck Tingri County in southern Tibet, as reported by the global Centroid-Moment-Tensor (CMT) catalog^{23,24} (Fig. 1b). This event occurred in the southern Xainza-Dinggye rift (XDR) and is the largest normal-fault earthquake on the Tibetan Plateau since the 2008 M_w 7.1 Yutian earthquake²⁵. It offers a rare opportunity to understand how continental normal-fault systems accommodate major earthquakes as well as the extension in the southern XDR, which includes Lagoi, Dingmuco, Sangkar, Cuomai, Mabja and Dinggye faults²⁶. Field observations indicate that the earthquake ruptured along the pre-existing north-south trending Dingmuco–Lagoi fault system, with a maximum vertical offset of 2.5–3.0 m and segmented surface ruptures^{27,28}. The throw rate of the Dingmuco fault is determined at 0.6 mm/yr²⁹ to 1.1 mm/yr²⁶ during the Quaternary. Combining long-term deformation data with co-seismic data suggests an average recurrence interval of ~800 years for earthquakes of similar magnitude to the 2025 event along the Dingmuco fault²⁶. Initial InSAR inversions assumed a single planar fault geometry³⁰. A subsequent grid search using a two-fault model with a fixed top edge suggested a conjugate normal fault system³¹. However, such graben-like conjugate normal faults are difficult to reconcile with the high-elevation setting and the north–south asymmetry of aftershocks³².

Here, we explore the rupture geometry of the Tingri earthquake using a comprehensive dataset comprising historical seismicity, co-seismic InSAR, and seismic waveforms, along with aftershock catalogues. First, we conduct a Bayesian probabilistic inversion of InSAR data to resolve the detailed geometry of the multi-segment rupture. Utilizing the derived fault geometry, we then analyze the distributed slip and slip evolution by inverting the InSAR and seismic data, respectively, and evaluate the resulting co-seismic stress transfer. Next, we invert historical seismic data to determine the regional pre-seismic stress regime, thereby assessing the tectonic controls on fault activation. Finally, we employ an automatic algorithm to detect and parameterize fault structures from aftershock hypocenters, facilitating an analysis of how rupture geometry controls aftershock seismicity. This study advances our understanding of major earthquake ruptures on segmented normal-fault systems and provides insights into seismic hazard in active continental extensional regions.

Results

SAR deformation and rupture geometry

Range offsets from Sentinel-1A (one ascending and two descending tracks; see Methods) reveal segmented surface ruptures as discontinuities in offset maps (Supplementary Fig. S1), outlining a fault system approximated by three segments (S1–S3). Because pixel-offset noise degrades parameter stability, we excluded these offsets from fault inversions and instead mapped line-of-sight (LOS) InSAR deformation using three Sentinel-1A scenes and one ALOS-2 descending-track image (Supplementary Fig. S2). While C-band Sentinel-1A data decorrelate near the fault, the L-band ALOS-2 interferogram retains coherence and resolves a more complete co-seismic displacement field. The combined datasets yield a high-quality deformation field with asymmetric patterns across the fault and a localized signal west of the epicenter. This latter feature is especially prominent in the ascending Sentinel-1A track (T12A), whose narrower color map range accentuates the complexity of this normal faulting event.

For the prescribed segments S1–S3, strikes were fixed and each segment was represented by a uniform-slip plane with unknown dip, width, rake, and slip magnitude. We inverted down-sampled pixels from all four LOS maps using a Bayesian framework to estimate the posterior probability distributions of the geometric and kinematic parameters (see Methods). Posterior marginals are shown in Supplementary Fig. S3. The preferred three-fault model, obtained from the medians of these posterior probability distributions, yields strikes of 181°, 192°, and 180° for S1, S2, and S3, with corresponding dip angles of 51°, 46°, and 38° (Supplementary Table S2). The fit quality, measured by the variance reduction (VR; see Methods), is 80%, 85%, 93%, and 94% for the ALOS-2, Sentinel-1A T12A, T121D, and T48D InSAR data, respectively (Supplementary Fig. S4).

Residual deformation persists west of the epicenter, most clearly in the T12A data, indicating that the three-fault model underfits signals in this area. We therefore introduced a fourth segment (S4) west of S1–S3 and adopted a four-fault configuration. Given the absence of mapped surface rupture measurements, fault segment S4 was modeled as a uniform-slip plane with its top edge unconstrained. S4 was parameterized by nine variables: (1) 3D fault center coordinates; (2) geometric attributes (strike, dip, along-strike length, and along-dip width); and (3) slip kinematics (rake and slip magnitude). Priors confined the fault center west of the epicenter (Supplementary Fig. S2 and Table S3), and we used broad uniform priors on strike (0–359°) and dip (1–90°) to avoid geometric bias. A Bayesian inversion for the four-fault model (Supplementary Fig. S5) yielded median dip estimates of 55°, 48°, and 45° for S1–S3, respectively. S4 strikes 159°, dips WSW at 27°, is centered at 4.6 km depth, and accommodates ~0.6 m of oblique slip dominated by normal-fault motion. Relative to the three-fault model, the four-fault solution significantly improves the fits to both Sentinel-1A and ALOS-2 datasets (cf. Fig. 2 and Supplementary Fig. S4), increasing VR from 80% to 86% for ALOS-2, from 85% to 91% for Sentinel-1A T12A, and from 93% to 97% for Sentinel-1A T121D (Supplementary Table S4). The VR for Sentinel-1A T48D remained unchanged (94%), as that track does not cover the area west of the epicenter.

As with three-fault configuration, the four-fault model still underfits near-fault deformation in the ALOS-2 map. This an expected limitation of the uniform-slip assumption. This misfit is mitigated in subsequent distributed-slip inversions (Supplementary Fig. S8). Taken together, the Bayesian inversion of ALOS-2 and Sentinel-1A datasets supports a major, steeply dipping normal fault system (S1–S3) accompanied by a synthetic, shallow-dip splay fault (S4).

Distributed fault slip patterns

As shown in Fig. 2, the ALOS-2 and two Sentinel-1A interferograms (T12A and T121D) were acquired 5–10 days after the mainshock, and Sentinel-1A Track (T48D) does not image the area west of the epicenter. A joint inversion of these InSAR data with seismic records would therefore risk aliasing early post-seismic deformation into the co-seismic model. To avoid this bias, we inverted the InSAR and seismic datasets separately (See Methods).

The InSAR-only inversion locates the principal slip north of the epicenter (Fig. 3a). Peak surface offset reaches ~2.5 m, consistent with the field investigation^{27,28}. Subsurface slip peaks at ~3 m at a depth of 4–6 km and extends ~60 km along strike and ~15 km down dip. The slip on fault S4 peaks at ~0.8 m, concentrated at 5–7 km depth on the northern section of the fault.

For the seismic finite-fault inversion, we used near-field co-seismic strong-motion and GNSS stations (Fig. 1a) together with teleseismic records (Supplementary Fig. S7). Due to the sparse near-field data, the inclusion of S4 not significantly affect the model fit. Consequently, it remains uncertain whether S4 slipped coseismically with the main fault or in the early postseismic phase. The resulting slip evolution model for S1–S3 is shown in Supplementary Fig. S8. It indicates that the rupture initially propagated bilaterally during the first ~10 s. Southward propagation then arrested, while northward rupture continued until ~20 s after the origin time.

Pre-stress and aftershock seismicity

Stress and friction coefficient inversion from regional historical focal mechanisms (see Methods) yields principal stress orientations (σ_1 , σ_2 , σ_3) with azimuths/plunges of 270°/74°, 175°/2°, and 84°/16°, respectively (Fig. 3b). The estimated friction coefficient for faults in the region is 0.45. In a standard Andersonian normal-fault regime, σ_1 is vertical. Here, however, σ_1 is rotated by 16° from vertical. This inclination indicates the horizontal plane is no longer a principal stress plane, revealing a shear stress component acting on the horizontal plane. The estimated stress shape ratio is $R = (\sigma_1 - \sigma_2)/(\sigma_1 - \sigma_3) = 0.17$. A low R value indicates that σ_1 and σ_2 are nearly

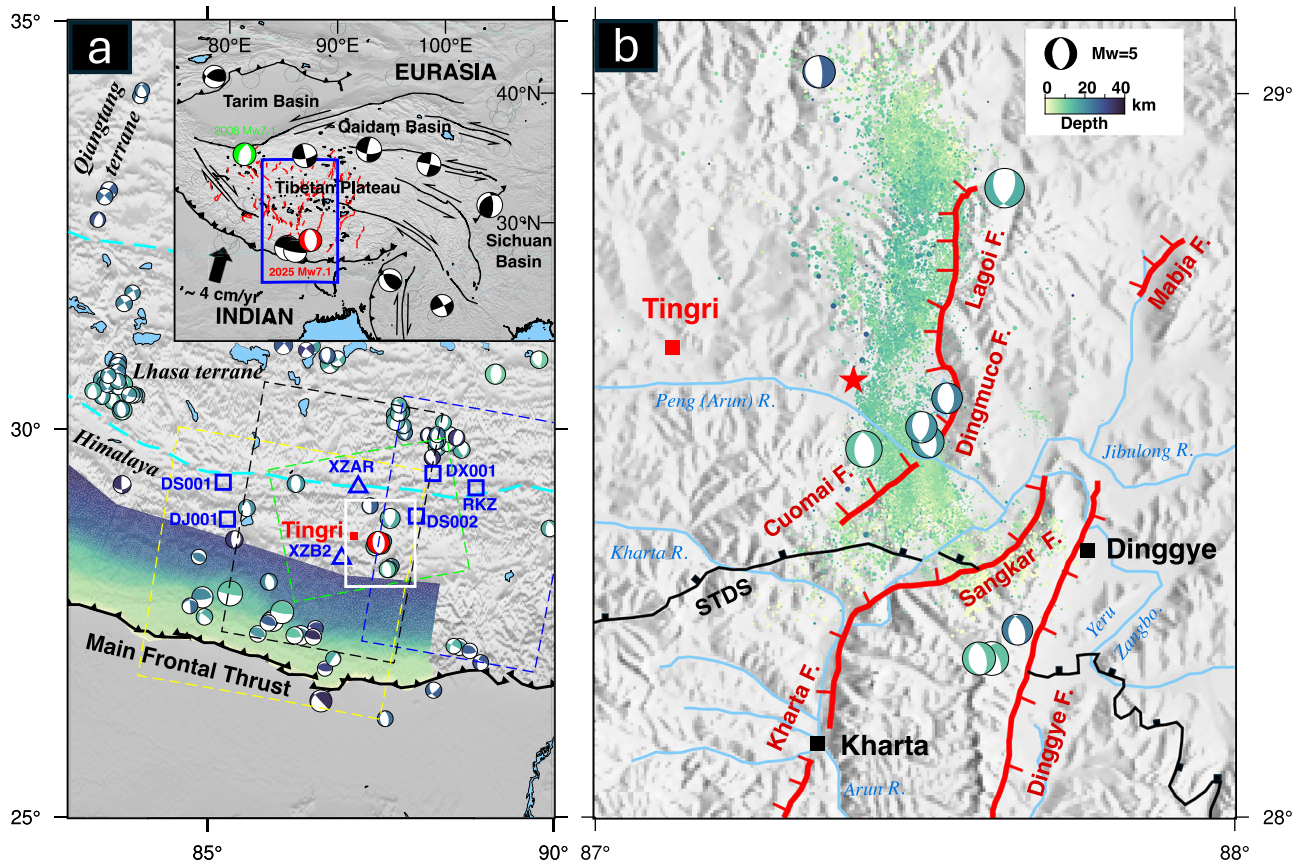


Fig. 1 | Seismotectonic setting of the 7 January 2025 Tingri event. **a** Regional overview outlined by the blue box in the inset. Focal mechanisms of historical $M_w \geq 7.0$ earthquakes from the global CMT project (since 1976) are shown in the inset, with the Tingri $M_w 7.1$ earthquake highlighted in red. Active rift faults on the Tibetan Plateau are marked by red lines. Sutures are shown by cyan dashed lines^{87,88}. Focal mechanisms for $M_w > 5.0$ historical events and the deep Main Himalayan Thrust fault surface⁸⁹ are denoted and color-coded by depth, using the same green-blue scale as the aftershock hypocenters in **(b)**. Tingri city is marked by the filled red square. The coverage of SAR data is outlined with dashed lines: Sentinel-1A

(ascending track 12, green; descending tracks 48, blue; descending track 121, black) and ALOS-2 (descending track, yellow). Instrumentation includes GNSS stations (blue triangles) and strong-motion stations (blue squares). The white box indicates the area enlarged in **(b)**. **b** Detailed view of the epicentral area. The red star denotes the epicenter from the U.S. Geological Survey (USGS) (<https://earthquake.usgs.gov/earthquakes/search/>). Nearby towns are marked by filled black squares. The nearby active fault and the South Tibetan Detachment System (STDS) are marked by bold red and black traces, respectively, adapted from ref. 26.

equal. Given that the trace condition $\sigma_1 + \sigma_2 + \sigma_3 = 0$ holds, this implies both σ_1 and σ_2 are compressional, while only σ_3 is extensional and oriented near east. This result is consistent with a previous seismic moment tensor analysis. Decomposition of the earthquake moment tensor into a dislocation source component and a compensated linear vector dipole component indicates the dislocation source occurred in a mechanical state of near-vertical compression and near east-west extension³³. Given the near-vertical σ_1 , steeper faults would typically require a higher friction coefficient to remain stable. The estimated friction coefficient of 0.45 is slightly less than the typical range of 0.6–0.8 for seismogenic faults³⁴. Therefore, this relatively low value is consistent with conditions favorable to low-angle normal faulting.

We further evaluated the failure tendency of S1–S4 under this stress regime (Fig. 3b). The instability coefficients (I) for S1–S4 are 0.93, 0.99, 0.99, and 0.91, respectively. Values of $I > 0.9$ indicate high instability, signifying that the geometry of these faults is favorably oriented for failure within the prevailing pre-stress field³⁵. The slip directions predicted by the stress model agree with those derived from our kinematic inversion, supporting consistency between the inferred rupture geometry and the regional stress environment.

To examine what promoted activation of the secondary fault S4, we also computed the Coulomb failure stress change (ΔCFS) imparted by slip on faults S1–S3 onto S4 (Fig. 3d; see “Methods”). The upper left and right sections of S4 experience a significant Coulomb stress increase,

which partially correlates with the slip distribution on S4 (Fig. 3c). This correlation suggests that the activation of fault S4 was driven by a combination of the tectonic pre-stress environment and co-seismic stress transfer.

We further analyzed ~30,000 relocated aftershock hypocenters recorded during the first 12 days after the mainshock³². Aftershock density is inversely correlated with the mainshock slip distribution (Fig. 3a). The major surface and underground slip occurred along the Lagoi fault, where the aftershock density is sparse, whereas the minor slip occurred at the southern and northern fault ends, which host denser aftershock activity. This spatial distribution pattern is consistent with residual stress release in the unruptured sections during the mainshock.

The spatial distribution of aftershock hypocenters provides a complementary reference for mapping the geometry of the mainshock fault structures. Using an automatic algorithm³⁶, we parameterized linear fault structures from the aftershock catalog, evaluating cluster quality by planarity, seismic-moment feasibility, and event utilization (see “Methods”). We identify ~90 candidate planes. Their strikes are predominantly north-south, aligned with S1–S4, and their dips span 0° to 90° with a bimodal distribution peaking at the 70°–80° and 10°–20° (Fig. 4).

It is important to acknowledge the limitations inherent in extracting fault geometry from aftershock hypocenters. First, hypocentral precision impacts the reliability of the results, potentially causing misalignments in the interpreted fault structures³⁷. Second, a key uncertainty is whether the

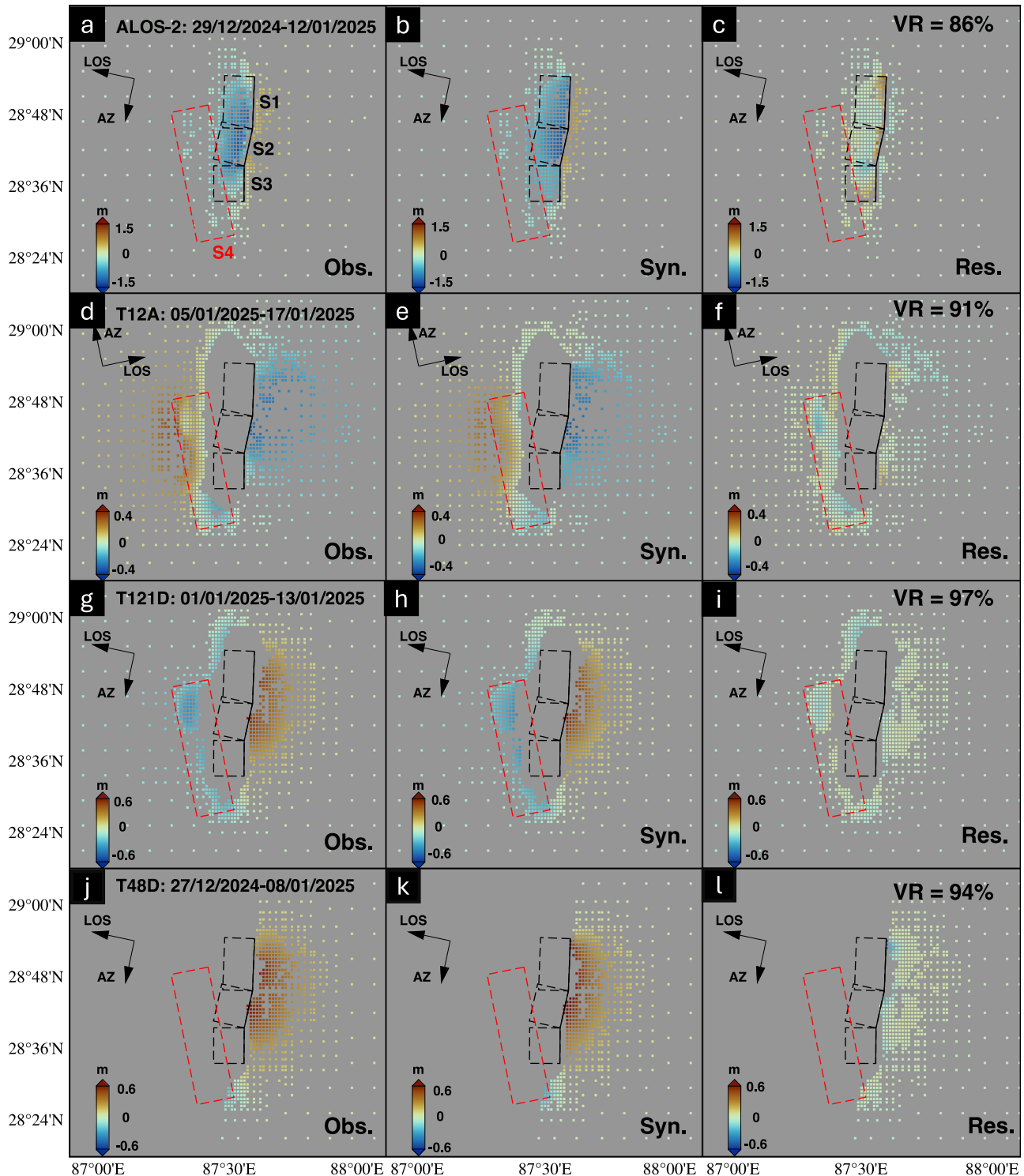


Fig. 2 | Bayesian inversion results for a four-fault geometry scenario. Observed and modeled surface displacements across four datasets. Each row of panels corresponds to a single satellite track: **a–c** ALOS-2 descending track, **d–f** Sentinel 1 A ascending track 12, **g–i** Sentinel-1A descending track 121, and **j–l** Sentinel-1A descending track 48. For each track, the left panel (**a, d, g, j**) shows the observed line-of-sight (LOS) surface displacement, the center panel (**b, e, h, k**) shows the synthetic

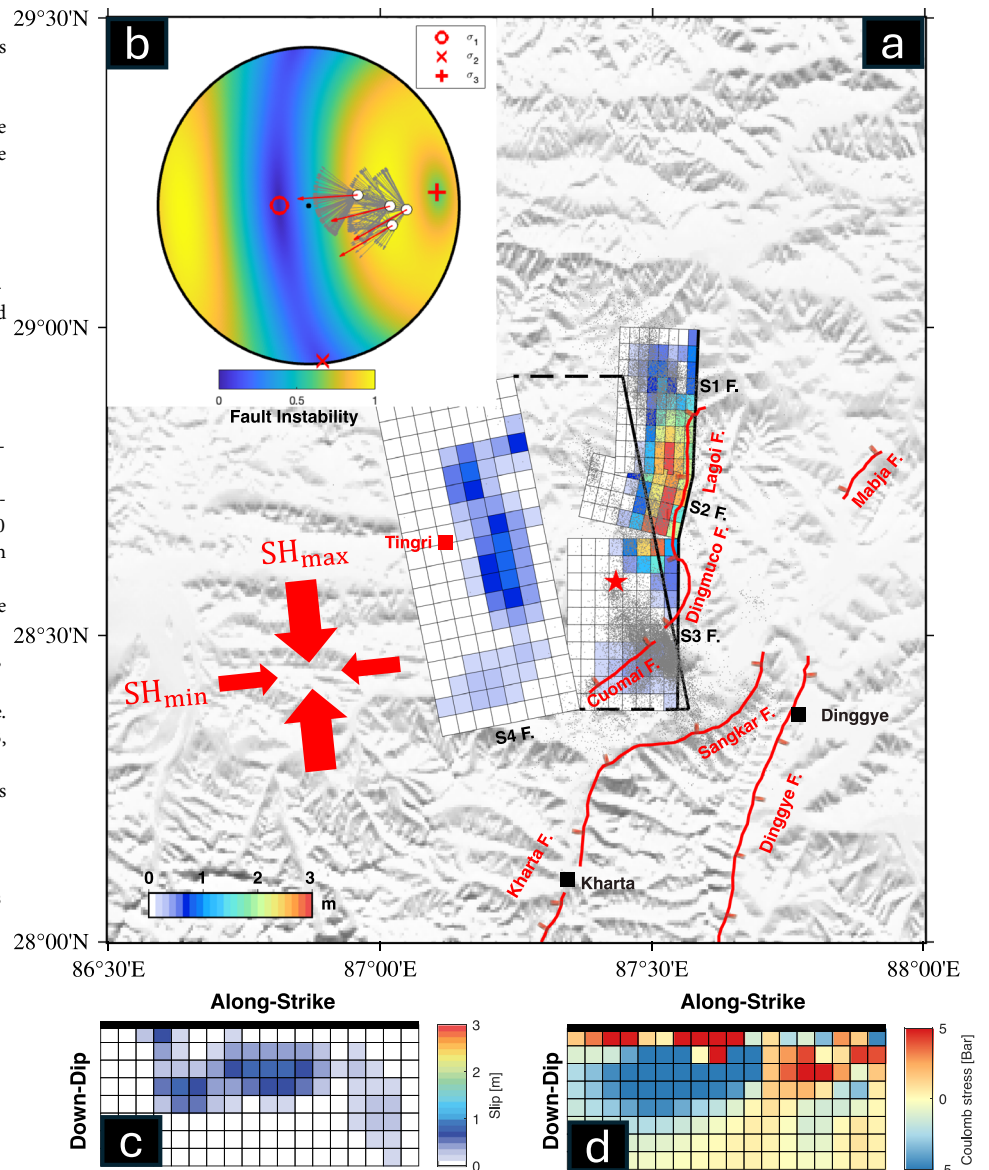
displacement from the preferred model, and the right panel (**c, f, i, l**) shows the residuals. All panels in each row share the same color scale. Surface traces of faults S1–S3 (solid black lines) are fixed from pixel-offset maps (Supplementary Fig. S1). Dashed black and red lines denote the surface projections of fault planes S1–S3 and S4, respectively. AZ azimuthal direction, LOS line of sight. Positive values indicate motion toward the satellite. Dates are expressed as DD/MM/YYYY.

aftershocks occurred directly on the mainshock fault plane or on off-fault fractures³⁸. These factors may lead to an overestimation, meaning our ~90 identified fault planes likely include spurious candidates. Nevertheless, the statistical distribution of their strike and dip angles provides an independent

reference. Although the dominant dip angles in the bimodal distribution from aftershock analysis do not perfectly match those from the Bayesian inversion of InSAR data, both methods reveal the coexistence of steeply and shallowly dipping structures.

Fig. 3 | Distributed fault slip and stress regime.

a The distributed fault slip projected onto the Earth's surface. The amplitude of fault slip is color-coded. Gray dots show aftershock locations. The red star marks the epicenter. Solid red traces denote surface faults. Solid black straight lines indicate the top edge of the S1, S2, S3, and S4 fault planes. The S4 fault plane is offset to the west by 0.2° along the longitudinal direction to clearly display the slip distribution of the S1, S2, and S3 faults. The offset distance is indicated by a dashed black line. Orientations of SH_{min} and SH_{max} in the region are derived from the regional stress regime (**b**). **b** Lower-hemisphere pre-stress stereograph. Red symbols denote the orientations of the principal stress directions (σ_1 , σ_2 , σ_3), inverted from regional historical earthquake focal mechanisms. Colors represent fault instability coefficient I^{35} for fault planes with varying strikes and dips under the inverted pre-stress regime. Instability coefficient I ranges from 0 to 1, indicating how optimally aligned a fault is with the stress field, calculated using a static friction coefficient of 0.45 from stress inversion. At $I = 0$, the fault plane is misoriented relative to the prestress, maximizing stability and suppressing slip. At $I = 1$, optimal alignment between the fault plane and prestress minimizes slip resistance, favoring failure. White dots show the poles of fault planes S1, S2, S3, and S4 projected onto the lower hemisphere. Red arrows indicate the expected slip directions on faults S1–S4 under the pre-stress regime. Gray arrows show the slip directions of each subfault from the distributed slip inversion. **c** Tile view of the slip distribution on fault S4. **(d)** Coulomb failure stress changes on fault S4 due to slips on faults S1, S2, and S3.



Discussion

Resolving the complex rupture geometry of normal-fault systems is critical for understanding seismic rupture dynamics and for hazard assessment in continental extensional regions. For the 2025 Tingri earthquake, earlier studies inverted InSAR data using a single-plane approximation, yielding a single west-dipping fault with a strike of $\sim 187^\circ$ and a dip of $\sim 55^\circ$ ³⁰. To better fit localized deformation west of the epicenter, a subsequent study introduced a secondary fault, fixing its top edge based on the InSAR deformation gradients observed west of the epicenter. They identified an antithetic east-dipping plane and proposed a conjugate, graben-like system³¹. However, this two-fault model presents inconsistencies. In the proposed configuration, the secondary fault is unloaded by the main fault at depth. Furthermore, the graben-like geometry is difficult to reconcile with the unusually high topography of the purported down-dropped block. Classic grabens are crustal sections lowered relative to surrounding blocks³⁹. They commonly form low-lying basins and rift valleys that host normal-fault earthquakes^{19,20}.

In contrast, our Bayesian inversion method does not prescribe a fixed top edge for a secondary fault and explores more fault strike-dip combinations and positions (Supplementary Fig. S2 and Table S3). In other words, our search range for fault locations, strike and dip angles not only encompasses the setting of the previous study³¹ but also includes more potential

candidates. Therefore, our results represent a global optimization within a broader parameter space. Our results indicate a synthetic, low-angle ($\sim 27^\circ$) secondary fault (S4). Furthermore, the major slip occurs beneath areas of high surface elevation (> 5 km) (Fig. 5a, b). These observations demonstrate that major normal-fault earthquakes can nucleate and propagate without a classic graben structure. This finding carries implications for how rupture scenarios are represented in seismic hazard models for extensional terrains.

Mechanically, S4 is favorably oriented in the background prestress field (Fig. 3a). The ΔCFS calculations also suggest that S4 was partially loaded by major faults S1–S3 (Fig. 3d). S4 may reflect either a newly formed structure or the reactivation of a pre-existing one at the co- or early post-seismic phase. The bimodal dip distribution inferred from aftershocks also supports the coexistence of steep and shallow fault segments (Fig. 4).

The activity of LANFs during earthquakes has been controversial, as Andersonian mechanics would favor frictional locking at such shallow dips⁴⁰. Global statistics also indicate that events with 30° – 60° dips are about an order of magnitude more frequent than those with dips $< 30^\circ$ ⁴¹. Nevertheless, recent paleoseismic evidence demonstrates that LANF segments can rupture during $M_w > 7.0$ normal-fault events⁷. In addition, 3-D dynamic rupture simulations show that LANFs can rupture simultaneously with pre-existing splay faults⁸. In such simulation, a LANF (dipping at 16 – 24°) acts as

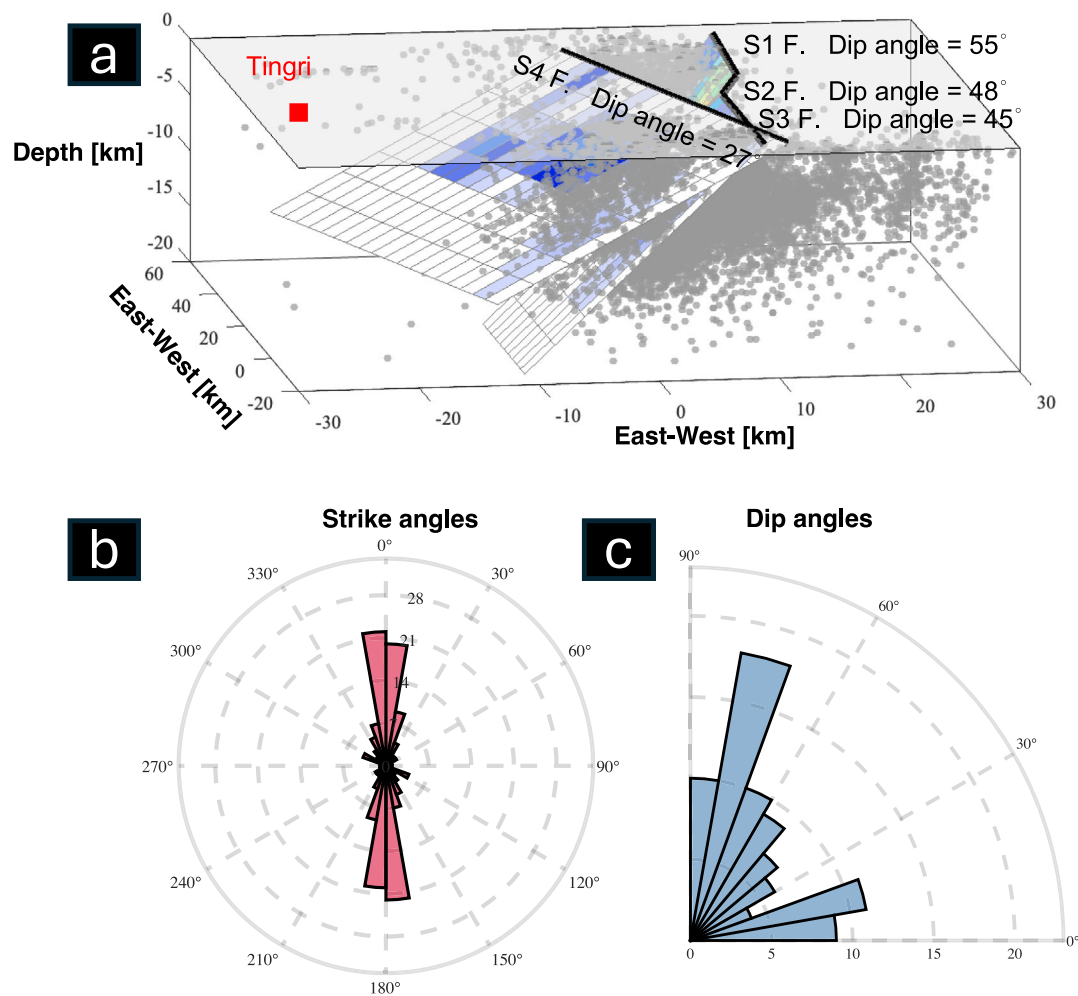


Fig. 4 | Aftershock seismicity and rupture geometry. **a** Perspective view of the inverted finite-fault planes and relocated aftershocks (gray dots). Lower panels show the frequency distributions of strike (**b**) and dip (**c**) angles for fault segments

extracted from the aftershock sequence. The radius of each sector reflects the number of strike and dip candidates within that interval.

the main fault, while synthetic and antithetic splay faults dip at 45–75°. Results show that shallowly-dipping synthetic splays not only host greater co-seismic slip but also more effectively arrest shallow rupture on the main fault compared to steeper antithetic splays. In the Tingri earthquake, the low-angle fault (S4) acted as a synthetic splay fault, while the main fault (S1–S3) was steeper (45°–55°). Here, shallow slip on the LANF was likely arrested, highlighting a potential mechanical competition between a low-angle splay and a steeper main fault. These findings underscore the necessity for future dynamic rupture scenarios in which a LANF is modeled as a synthetic splay alongside a steeper main structure.

In multi-fault ruptures, LANF slip can be masked by simultaneous slip on steeper faults, as moment-tensor solutions in global catalogs (e.g., Global CMT) represent the aggregate signal from all segments. Furthermore, the seismic moment tensor of the mainshock may not account for a secondary, low-dip fault if it slips shortly after the primary co-seismic rupture. The scarcity of modern, well-recorded $M_w > 7.0$ normal-fault events may therefore contribute to an underestimation of LANF involvement. Our findings underscore the critical role of high-resolution geodetic observations (e.g., InSAR), combined with probabilistic Bayesian inversion and dense aftershock analyses, in resolving hidden shallow-dip participation in major continental normal-fault earthquakes. By revealing such structures, our work demonstrates how these techniques provide a more complete basis for understanding how strain is partitioned and released across multiple faults.

This is critical for developing accurate fault models and simulating the spatial and temporal evolution of seismicity, thereby advancing long-term seismic hazard assessment for the region.

The largest documented continental normal-fault earthquakes typically occur in extensional plate boundary zones, such as the western U.S. Basin and Range, Italy, the eastern Mediterranean, and the East African Rift¹⁷. The 2025 $M_w 7.1$ Tingri earthquake occurred in the southern Xainza-Dinggye rift, ~250 km north of the 2015 $M_w 7.9$ Gorkha thrust event (Fig. 1a). Its location coincides with the systematic south-north transition in focal mechanisms across the Himalayan-Tibetan orogen (Fig. 1a). Thrusting dominates near the Main Himalayan Thrust, whereas normal faulting prevails to the north, reflecting fundamental contrasts in the regional stress regime^{22,42}. Our stress inversion confirms an extensional environment in the study area (Fig. 3b).

With respect to rupture geometry, our estimated dip angles of 45°–55° for major faults S1–S3 align with typical normal-fault earthquakes on the Tibetan Plateau¹⁹. Multi-segment rupture is a fundamental characteristic of large continental earthquakes in extensional settings, as increasingly revealed by high-resolution space geodetic observations. While the mechanics of extension are broadly shared, rupture geometries can vary significantly depending on local fault properties and crustal heterogeneity^{4,9,10}. This diversity underscores the need to understand multi-segment ruptures not as uniform phenomena, but as context-dependent

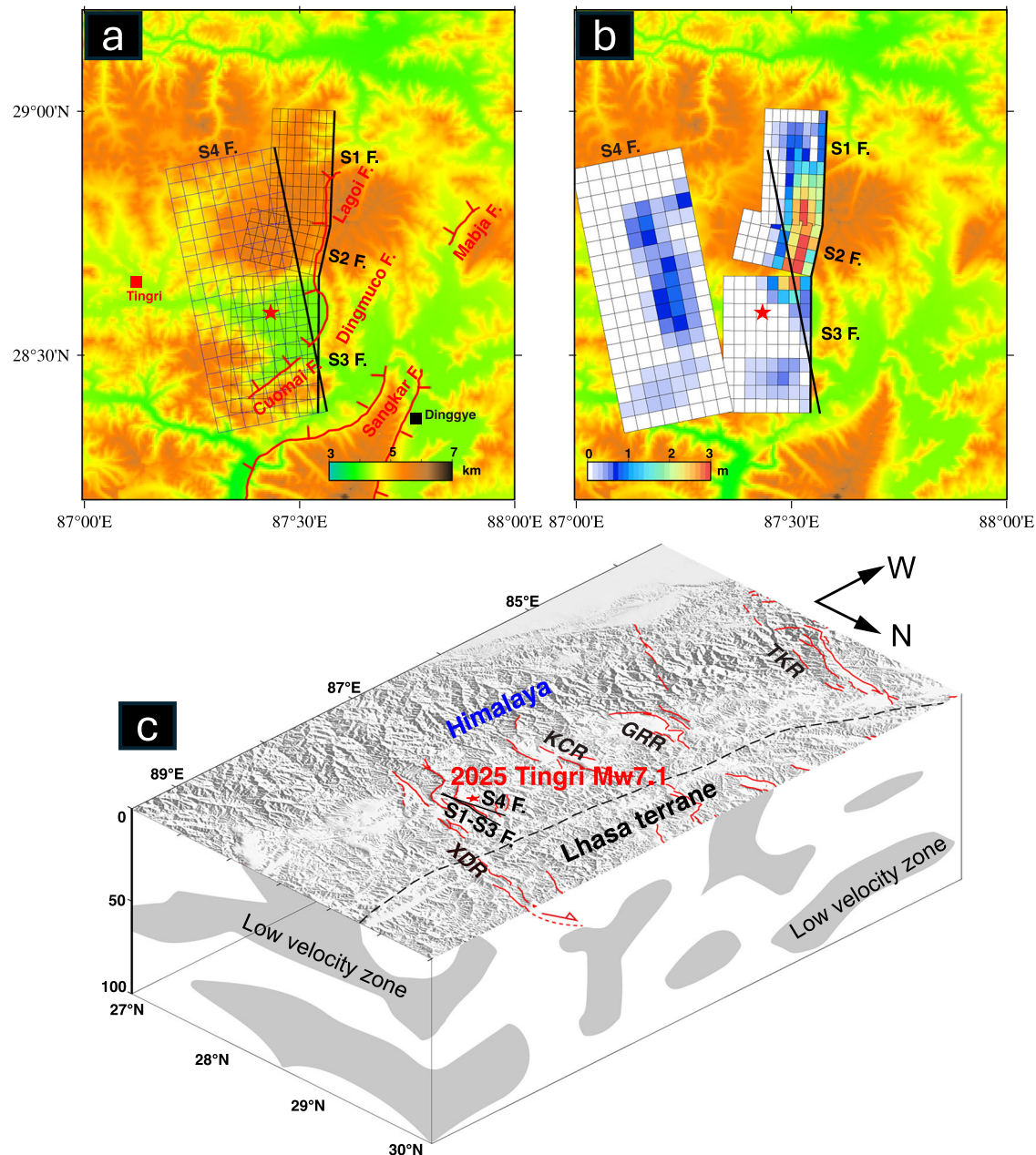


Fig. 5 | The Tingri earthquake versus topography and crustal low-velocity zone. **a** Divisions of finite-fault patches projected onto the surface, along with elevations around the earthquake. The top edge of finite-fault planes are marked by bold black straight lines. Thick red lines indicate the surface segments of regional active faults. **b** Same as (a), but with slip distribution displayed. The S4 fault plane is offset westward by 0.2° along the longitudinal direction to clearly show the slip distribution of the S1, S2, and S3 faults. **c** West-dipping fault geometry (red dashed line) of the earthquake beneath the Xainza-Dinggye rift (XDR), along with the low-velocity zone

beneath the southern Tibet rift from seismic tomography⁵⁰. Dashed black lines indicate sutures. Red traces represent normal faults of the southern Tibet rift. The red star marks the earthquake epicenter, and the black straight lines denote the top edge of S1–S4 fault planes. The inferred deep fault structure of the southern XDR, interpreted from the focal mechanism of the 2025 Tingri event, is shown as dashed red lines. The red arrow indicates the motion direction of the hanging wall. Abbreviations: KCR Kung Co rift, GRR Gyirong rift, TKR Thakkhola rift.

expressions of regional extensional processes. For instance, the 2008 $M_w 7.1$ Yutian normal-fault earthquake ruptured in a localized east–west trending extensional field near the southern edge of the Tarim Basin. Geodetic modeling shows it ruptured multiple segments with significant strike changes and a steep splay fault dominated by strike-slip motion⁴³. Another representative example, the 2016 $M_w 6.6$ central Italy earthquake, occurred within a heterogeneous stress regime characterized by NE–SW extension⁴⁴. Its rupture occurred not only along the primary Mt. Vettore–Mt. Bove fault but also along two secondary antithetic faults within the hanging wall²⁰. Both cases demonstrate multi-fault involvement, yet their distinct patterns reflect

differences in fault network complexity, stress field heterogeneity, and crustal layering.

The 2025 Tingri event is distinguished by a low-angle splay fault. In mature continental rifts, extension is typically accommodated by high-angle normal rift faults, and LANFs in these settings are often interpreted as the result of long-term rotation of initially high-angle faults⁵. The multi-fault rupture of the Tingri event challenge classic Andersonian fault theory and contributes to the ongoing debate about whether such faults can initiate at low angles and slip seismically, thereby advancing our understanding of tectonic evolution in rift systems. Thus, while multi-segment ruptures

appear universal in continental extensional earthquakes, their specific geometries and governing mechanics are regionally contingent. The Tingri event expands the spectrum of possible rupture configurations of continental extensional earthquakes and encourages reevaluation of seismic hazard models in other extensional regions, particularly where low-angle structures or complex fault interactions may have been previously underestimated.

Modern Himalayan deformation is often cited as a textbook case of syn-convergent extension, in which the plateau interior extends as arc-parallel despite the ongoing India–Asia collision^{21,45}. The most prominent expressions of this extension are the N–S trending rifts defined by normal faults^{22,46} (Fig. 1a). The 2025 M_w 7.1 Tingri earthquake provides a valuable opportunity to probe active extension within the Himalayan orogen and to better constrain the geometry and kinematics of these N–S rifts. The spatiotemporal evolution of rift initiation and propagation has been widely investigated due to its potential links to deep lithospheric processes^{47–51}. Yet, despite extensive work, the deep structure of these rifts and the coupling between upper-crustal extension and deeper geodynamic processes remain debated.

A variety of mechanisms have been proposed to explain the N–S rifts on the Tibetan Plateau, including radial spreading⁵², oroclinal bending of the Himalayan arc⁵³, lateral slab detachment⁵⁴, convective thinning of the lithospheric mantle⁵⁵, changes in boundary conditions⁵⁶, oblique convergence of the Indian plate⁵⁷, gravitational potential energy⁵⁸, slab tearing⁵⁹, and divergent orthogonal thrusting along the curved Himalayan front^{46,60}. While these models provide frameworks for long-term evolution, direct observations of ongoing rift processes are scarce, and the fault geometries/kinematics remain poorly constrained. Our analysis of the Tingri event shows that the rupture terminated at a depth shallower than 20 km, abutting a mid- to lower-crustal low-velocity zone (LVZ) identified by geophysical profiles⁵⁰ (Fig. 5c). This LVZ is commonly interpreted as a domain of ductile flow in the mid- to lower-crustal crust. Stress inversion analysis for the Tingri region indicates that the orientations of the maximum (σ_1) and minimum (σ_3) principal stresses are rotated by approximately 16° from the vertical and horizontal, respectively, suggesting the presence of significant horizontal shear stress. Such a deviation from an ideal Andersonian regime may reflect shear associated with southeastward mid- to lower-crustal flow^{58,61,62}. Taken together, the west-dipping rupture geometry of the Tingri earthquake and the evidence for ductile mid- to lower-crustal flow beneath the Xainza–Dinggye rift suggest that the southern segment of the rift cuts obliquely westward through the upper crust but likely does not penetrate the full crustal thickness. More broadly, these observations support a model in which segmented upper-crustal normal faulting is mechanically coupled to, yet partly decoupled from, ductile mid- to lower-crustal flow, providing direct constraints on the style of syn-convergent extension in the Himalaya.

Methods

SAR data processing

We use synthetic aperture radar (SAR) data from the European Space Agency's (ESA) C-band Sentinel-1A satellite and the Japan Aerospace Exploration Agency's (JAXA) L-band ALOS-2/PALSAR-2 to image the coseismic deformation of the Tingri earthquake. The dataset comprises three Sentinel-1A tracks (ascending track 12 and descending track 48 and 121) along with one descending ALOS-2 track, each acquired within a few days before and after the earthquake. Details of these images, including acquisition dates, spatiotemporal baselines, and orbit directions, are provided in Supplementary Table S1. A multilook factor of 20×5 is applied in the range and azimuth directions. Topographic contributions are removed from each interferogram using 30-m resolution Shuttle Radar Topography Mission (SRTM) DEM data⁶³. To suppress noise and enhance coherence, we apply the Goldstein filter⁶⁴, masking pixels with coherence values below 0.4 or exhibiting severe decorrelation. Phase unwrapping is performed using the minimum-cost flow (MCF) algorithm implemented in SNAPHU⁶⁵, with a maximum allowed deformation of 4 cycles per pixel. The unwrapped phase

is then converted to line-of-sight (LOS) displacement maps shown in Supplementary Fig. S2.

We also derive range offset maps for the three Sentinel-1A tracks using the pixel offset tracking (POT) technique^{66,67}, which estimates surface displacements in range and azimuth directions. Although noisier and less precise than interferograms, POT does not require phase unwrapping and can accommodate larger deformation gradients, providing reliable constraints on near-fault slip, typically decorrelated in interferograms^{68,69}. For our analysis we use an offset window of 64×64 pixels and a search window of 24×24 pixels, with offsets sampled every 32 pixels in both range and azimuth. Pixels with signal-to-noise ratios (SNR) below 5 are excluded after filtering and smoothing. The range offset maps are shown in Supplementary Fig. S1. All interferograms and offset maps are generated using the InSAR Scientific Computing Environment (ISCE) software⁷⁰.

Bayesian inversion of uniform-slip fault planes

The InSAR LOS displacement maps are used to invert for the geometric parameters of multiple faults. Each fault segment is modeled as a plane with uniform slip. While the uniform slip model cannot capture detailed variations in fault slip, it effectively represents fault geometry. Additionally, it avoids the subjectivity of distributed slip models, which require numerous parameters and regularization in Bayesian inversion. InSAR displacement maps contain millions of data points. To reduce noise and make the dataset suitable for inversion, we downsampled the InSAR displacement maps using the quadtree method implemented in the Kite software⁷¹. The downsampling process selects data points based on the spatial gradient of deformation. Ultimately, we obtained 1153 quadtrees for the ALOS-2 displacement map, 1208 quadtrees for the Sentinel-1A ascending track (path 12) displacement map, 794 quadtrees for the Sentinel-1A descending track (path 48) displacement map, and 1,318 quadtrees for the Sentinel-1A descending track (path 121) displacement map (Fig. 2).

The inversion of geometric parameters employs a statistical framework that relates the observations $\mathbf{d}$ to the forward model $\chi(\theta, \mathbf{m})$. This relationship is expressed as $\mathbf{d} = \chi(\theta, \mathbf{m}) + \boldsymbol{\varepsilon}$, where $\boldsymbol{\varepsilon}$ represents the observation error vector, typically modeled as a zero-mean Gaussian distribution. The forward model $\chi(\theta, \mathbf{m})$ is computed within an elastic environment⁷². Using Bayes' theorem, the posterior probability distribution of the fault parameters ($\theta, \mathbf{m}$) and the noise hyperparameter (σ) can be jointly estimated from their prior distributions and the likelihood of the observed data.

$$p(\theta, \mathbf{m}, \sigma | \mathbf{d}) \propto p(\theta, \mathbf{m}) \exp \left\{ -\frac{1}{2\sigma^2} [\mathbf{d} - \chi(\theta, \mathbf{m})]^T \mathbf{C}_d^{-1} [\mathbf{d} - \chi(\theta, \mathbf{m})] \right\} \quad (1)$$

Here, $\mathbf{C}_d$ is the displacement data covariance matrix, and σ scales the covariance. Subsequently, using the sequential Monte Carlo (SMC) method⁷³, we sampled the marginal posterior probability distributions of all model parameters. We started the SMC algorithm with 1200 Markov chains, with each undergoing 400 steps to sample the posterior distributions of model parameters.

We first inverted for a three-fault scenario (S1, S2 and S3) with the top edge of fault fixed. Then we estimate the dip angle, down-dip width, rake angle and slip of each fault plane. The setting of prior distribution of model parameters is shown in Supplementary Table S2. The sampled marginal posteriors of model parameters are shown in Supplementary Fig. S3. The data fit of the preferred three-fault model are shown in Supplementary Fig. S4.

We then inverted for a four-fault scenario. The introduced fourth fault S4 is modeled as a uniform-slip plane without constraints on the top edge. For the fourth fault, we estimate the 3-D location of the fault plane center, its strike, dip, rake, and slip, alongside the fault length (along strike) and width (along dip). The setting of prior distribution of model parameters for the four-fault inversion is shown in Supplementary Table S3. The sampled marginal posteriors of model parameters are shown in Supplementary Fig. S5. The sampled marginal posteriors of noise hyperparameters are

shown in Supplementary Fig. S6. The data fit of the preferred four-fault model are shown in Fig. 2.

Variance reduction calculation

The goodness-of-fit of the preferred model relative to four independent InSAR datasets is quantified by the variance reduction (VR) between observed displacements and synthetic displacements at all data points

$$VR = 100\% \times \left[1 - \frac{\sum_{i=1}^N (d_{\text{obs},i} - d_{\text{syn},i})^2}{\sum_{i=1}^N d_{\text{obs},i}^2} \right] \quad (2)$$

where $d_{\text{obs},i}$ and $d_{\text{syn},i}$ denote the observed and synthetic displacements at i -th data points.

Inversion of fault slip kinematics

We obtained P-wave records from 36 broadband stations (Supplementary Fig. S7) from the IRIS Data Management Center. These stations provide complete azimuthal coverage, with epicentral distances ranging from 30° to 90°. After removing the instrument response to obtain displacement waveforms, we calculated theoretical P-wave arrival times using the 1-D IASP91 velocity model⁷⁴ via ObsPy⁷⁵. The waveforms were then decimated to a 1 Hz sampling rate and bandpass-filtered (0.01–0.1 Hz) for inversion.

Additionally, we collected 1 Hz displacement waveforms from two GNSS stations and velocity waveforms from five strong-motion stations (Fig. 1a) provided by the China Earthquake Administration. The GNSS displacement data were filtered with a corner frequency of [0.001, 0.2] Hz, while the strong-motion velocity records were decimated from 100 Hz to 1 Hz and bandpass-filtered at [0.02, 0.1] Hz.

Based on the four-fault geometry from the Bayesian estimation, we then allocate smaller sub-faults to parameterize the fault geometry. For fault S1, we allocated 9 sub-faults along the strike and 8 sub-faults along the down-dip direction, with each sub-fault measuring 2.9 km × 3.1 km. For fault S2, we allocated 4 sub-faults along the strike and 8 sub-faults along the down-dip direction, with the same sub-fault size of 2.9 km × 3.1 km. For fault S3, we allocated 10 sub-faults along the strike and 8 sub-faults along the down-dip direction, with slightly larger sub-fault dimensions of 3.1 km × 3.5 km. As a result, the bottom edges of faults S1–S3 reach a depth of 20 km. For fault S4, we extended the fault plane to the surface, allocating 18 sub-faults along the strike and 8 sub-faults along the down-dip direction, each measuring 3.4 km × 3.5 km. Due to its relatively smaller dip angle, the bottom edge of fault S4 reaches a shallower depth of 13 km.

For the inversion, we employed a 1-D velocity model based on IASP91⁷⁴, with its shallow layers replaced by the Crust 1.0 model⁷⁶. Green's functions for InSAR displacements were synthesized using the frequency-wavenumber integration method⁷⁷, while teleseismic Green's functions were computed via a propagator matrix approach⁷⁸ and band-pass filtered to match the observed data's frequency range.

Slip evolution was estimated using a multi-time-window linear inversion approach^{79,80}. Each sub-fault was parameterized with five symmetric triangular time windows (2-second half-duration, staggered by 1 s) and two slip vectors (rake angles of −45° and −135°) for non-negative least squares inversion. To mitigate the ill-posed nature of the problem, we applied spatial Laplacian regularization and temporal first-derivative regularization. The optimal spatial smoothing parameters were determined using Akaike's Bayesian Information Criterion (ABIC)⁸¹. The maximum rupture velocity in the inversion was set empirically to 2.8 km/s (0.8 times the S-wave velocity)⁸².

The fault slip distribution inverted from InSAR deformation maps is shown in Fig. 3a. The inverted fault slip distribution fits the observations well (Supplementary Fig. S8). The slip evolution, jointly inverted from GNSS, strong-motion, and teleseismic waveform data, is shown in

Supplementary Fig. S9. Our kinematic slip inversion fits the GNSS, strong-motion, and teleseismic waveform data well, including the teleseismic data (Supplementary Figs. S7b, S10 and S11).

Stress inversion from regional historical earthquakes

To assess tectonic controls on the earthquake, we derived the regional pre-stress regime from historical focal mechanisms in the global CMT catalog within the study area (Fig. 1b). The detailed information of the historical earthquake focal mechanisms used for stress inversion is listed in Supplementary Table S5. Assuming (1) homogeneous tectonic stress and (2) earthquake slip vectors aligning with shear stress according to the Wallace-Bott hypothesis^{83,84}, we resolved the orientations of the major principal stress (σ_1), the intermediate principal stress (σ_2), the minimum principal stress (σ_3), the shape ratio ($R = (\sigma_1 - \sigma_2)/(\sigma_1 - \sigma_3)$), and the fault friction coefficient (μ) using the iterative inversion method³⁴. We then evaluated fault stability using the instability coefficient I from³⁵, a dimensionless measure ranging from 0 for fully stable faults to 1 for maximum instability, which quantifies the likelihood of fault activation under the given stress conditions.

Coulomb failure stress change

The change in Coulomb failure stress change (ΔCFS) on a fault plane is given by ref. 85

$$\Delta CFS = \Delta \tau - \mu' \Delta \sigma \quad (3)$$

where $\Delta \sigma$ and $\Delta \tau$ are the normal and shear stress change, respectively. μ' denotes the effective friction coefficient. Using an effective friction coefficient of 0.45 from stress inversion and the Poisson ratio of 0.25 for Earth's crust, we calculated the ΔCFS on fault plane S4 due to slip on fault planes S1–S3. The primary calculation used a rake of −130° for S4, based on our Bayesian inversion (Supplementary Fig. S5). To evaluate the sensitivity to this parameter, we also computed ΔCFS for rakes of −90°, −110°, −150°, and −170° (Supplementary Fig. S12).

Fault geometry structure from aftershock seismicity

Aftershock catalogs contain information on fault geometry structure. We use an aftershock catalog containing ~30,000 hypocenters and magnitudes³². To analyze the potential geometry structures within the aftershock seismicity, we applied an automatic fault geometry extraction algorithm (E2F)³⁶. The E2F algorithm estimates fault parameters in three main steps: (1) identifying linear segment features using a 3D Hough transform, (2) clustering fault candidates with a fault width-based clustering radius (defined by parameter C in the software), and (3) extracting fault strike and dip angles. The reliability of the extracted strike and dip angles is evaluated by three indices: the seismic event utilization rate, a 3D aspect ratio (which assesses how planar a fault zone is), and the ratio of seismic moment (RSM, which evaluates whether the seismic moment from the fitted fault region is representative).

In our implementation, after identifying linear segments with the 3D Hough transform, we tested different clustering radius by varying the C parameter (Supplementary Fig. S13). We found that a clustering radius with $C = 100$ yielded satisfactory results for all three indices. Using this parameter, we extracted ~90 fault plane candidates and calculated their strike and dip angles (Fig. 4d, c).

Data availability

The GNSS and strong-motion data in SAC format are stored on Zenodo (<https://doi.org/10.5281/zenodo.15873406>). The processed SAR data are also available on Zenodo (<https://doi.org/10.5281/zenodo.17790351>). The P-wave records used in this study were obtained from the IRIS Data Management Center (https://ds.iris.edu/wilber3/find_event).

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Author contributions

W.G. performed the finite-fault inversion and stress calculations and drafted the initial manuscript. K.C. conducted data analysis and contributed to the writing. M.L. processed the pixel offset and InSAR data. K.H. contributed to figure production and tectonic analysis. Q.Z., L.Y., and L.D.Z. participated in finalizing the manuscript.

Competing interests

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Additional information

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