

Active faults and earthquake due to continental deformation

Edited by

Chuanyou Li, Wenjun Zheng, Daoyang Yuan
and Zhuqi Zhang

Published in

Frontiers in Earth Science



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ISSN 1664-8714
ISBN 978-2-83251-041-4
DOI 10.3389/978-2-83251-041-4

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Active faults and earthquake due to continental deformation

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Citation

Li, C., Zheng, W., Yuan, D., Zhang, Z., eds. (2023). *Active faults and earthquake due to continental deformation*. Lausanne: Frontiers Media SA.

doi: 10.3389/978-2-83251-041-4

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OPEN ACCESS

EDITED AND REVIEWED BY

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 28 October 2022

ACCEPTED 01 November 2022

PUBLISHED 30 November 2022

CITATION

Li C, Zheng W, Yuan D and Zhang Z
(2022), Editorial: Active faults and
earthquake due to
continental deformation.
Front. Earth Sci. 10:1082316.
doi: 10.3389/feart.2022.1082316

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Editorial: Active faults and earthquake due to continental deformation

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KEYWORDS

continent dynamics, late Cenozoic deformation, active block boundary, active faults, earthquake hazard

Editorial on the Research Topic

Active faults and earthquake due to continental deformation

Introduction

The kinematics and dynamics of continental deformation are fundamentally different from those of plate boundary deformation. Continental tectonics is complex and as a result we have been unsuccessful in fully understanding them. Deformation of continental lithosphere is not restricted to a narrow belt like plate boundaries, but can be distributed over small or large regions (e.g., [Wernicke et al., 2008](#)). Understanding how structures are distributed in continental tectonic systems is key to interpreting the continental deformation and assessing earthquake hazards. Neotectonic studies have shown that the crustal deformation in continental China is driven by forces due to the motions of neighboring plates ([Molnar and Tapponnier, 1975](#); [Tapponnier and Molnar, 1977](#); [Wesnousky et al., 1984](#); [Ye et al., 1985](#); [Xu et al., 1993](#); [Zhang et al., 1998, 2003](#)).

The primary pattern of late Cenozoic tectonic deformation of continental China is characterized by relative movements and interactions of active tectonic blocks bounded by major active faults ([Deng et al., 2002](#); [Zhang et al., 2003](#)). Tectonics of these blocks shares significant similarities with that of plate tectonics, that is, concentrations of earthquakes and tectonic activity within narrow boundaries, but no significant deformation within the rigid blocks, such as the Ordos block and the active blocks in and around the Tibetan plateau. Historical records also reveal that most of the strong earthquakes have occurred along these block boundaries ([Zhang et al., 2003](#)). A necessary step in understanding the movement of the blocks is to document the late Cenozoic activity and kinematics of the major faults along the block margins. Tibetan Plateau and the active Ordos block are the most active areas of intra-continental deformation in China and serve as natural

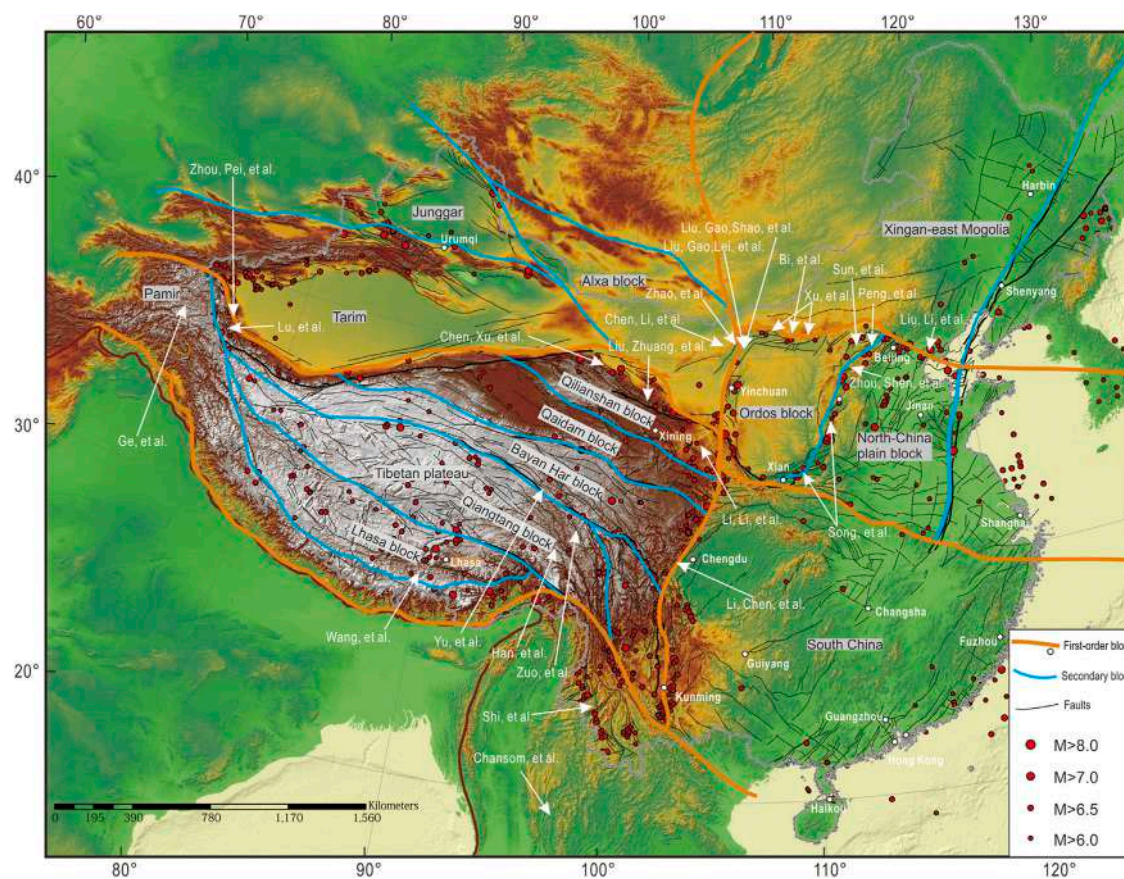


FIGURE 1

Active blocks, active faults and major earthquakes ($M > 6.0$) in China continent. Division of the active tectonic blocks follows the pattern proposed by Zhang et al. (2003). Locations of the 24 studies presented in this Research Topic are shown.

laboratories to better understand continental deformation. The majority of the 24 articles collected in this Research Topic mainly focus on present-day deformation in and around the Tibetan plateau and Ordos regions and try to cover much of the new research progress in this field (Figure 1).

Tectonic deformation along boundaries of active blocks in Tibetan plateau region

Cenozoic tectonics in Tibetan plateau region is clearly controlled by the Indo-Asian collision, which was separated by the roughly E-W-oriented fault zones into a series of blocks. The Bayan Har block in central Tibetan plateau is no doubt the most striking active block due to its intense fault activity. Field investigation obtains new evidence of the later Quaternary activity and recent large earthquake ruptures of the Garzê-Yushu fault that connects the Xianshuihe fault zone and constitutes the southern boundary of the Bayan Har block in

west. The determined left-lateral slip-rate of 6.3 ± 1.9 mm/yr in the Holocene along the fault (Yu et al.) is very close to the estimated average slip rate for the Xianshuihe fault. Surface ruptures with a length of ~ 100 km of the 1,738 earthquake have been discovered along the Dangjiang segment (Yu et al.), which demonstrates that the individual segments of the faults bounding the southern boundary of the Bayan Har block are all able to produce large earthquakes. The eastern margin of the Tibetan plateau is bounded by the Longmen Shan fault zone that is also the eastern boundary of the Bayan Har block and hosted the devastating Wenchuan earthquake in 2008. However, this structure once had been assigned a moderate-to-low seismic hazard rating before the 2008 event because it is slowly slipping (Zhang, 2013). Another impression making this fault zone seem innocuous and less significant seismic hazard is that the co-seismic displacements along its southern segment were very small and thus the southern segment is defined as a minor surface-rupture fault (Li et al.). Using scanning electron microscope and X-ray diffraction methods, Li et al. discovered that the fault gouge of the Dachuan-Shuangshi fault (the

southern Longmen Shan fault zone along which the Lushan M7.0 earthquake occurred in 2013) greatly differs to that of the Yingxiu-Beichuan fault (hosted devastating Wenchuan earthquake and produced significant surface ruptures) in the microstructure and mineral composition. This finding can be used as an auxiliary indicator to identify minor-surface-rupture faults.

The tectonic deformation of northeastern Tibetan plateau also exhibits a unique pattern. For the Haiyuan fault, the characteristics of the crustal deformation before the 2022 Menyuan Ms 6.9 earthquake and the future seismic hazards on this fault zone are analyzed using GPS-derived velocity data combined with the distribution of small earthquakes along the fault (Liu et al.). This study shows that the earthquake may link the Lenglongling and Tuolaishan faults, and that the present-day crustal deformation in the northeastern margin of the Tibetan plateau characterizes as the crust-shortening distributed in the Qilian Shan region and the left-lateral strike slip localized in the Haiyuan fault zone. Work from geomorphologic study to the west shows similar results (Chen et al.). Based on the slip rates derived from slope profiles and dating of the deformed river terraces across the NWW-trending faults in the western Qilian Shan, the study also suggests that the crustal shortening is widely distributed in the northern Qilian Shan region, while the strike-slip faults accommodate the sinistral strike-slip of the Altyn Tagh fault that bounds the northern Tibetan plateau block. To assess the earthquake hazard in the northeastern margin of the Tibetan plateau, Li et al. constructed a three-dimensional fault model concerning eight major faults by using multidisciplinary data and computer-aided modeling and optimization technique. Based on this 3D fault model, this study finds four fault segments with relatively higher potential to rupture in strong earthquakes in this region and illustrates the high complexity of a strike-slip dominant fault system developing along the boundary between active tectonic blocks.

Neotectonics, Cenozoic kinematics and geodynamics of the Tibetan plateau also have been studied recently in the western and southern Tibetan plateau, which can lead to a better understanding of the tectonic settings in these regions. By reviewing and reanalyzing the geologic, geodetic and seismic data, Ge et al. proposed the multi-spatiotemporal evolution of deformation patterns of the Pamir salient in western Tibetan plateau since the Late Cenozoic and thus provide a new view to understand the Cenozoic kinematics and geodynamics in this region. The northward motion of the Pamir salient may have caused the intense uplift of the Northeastern Pamir from 4.9 to 1.9 Ma, according to a detailed magnetostratigraphic investigation at the southwestern margin of the Tarim Basin (Zhou et al.). Using a fault scarp degradation model, Lu et al. constrained earthquake history of the Muztagh Ata and Tahman faults in

the northern Pamir. The recurrence interval of ~1,000 years and the vertical slip rate of 2.2 ± 0.3 mm/yr across the Muztagh and Tahman fault suggest strong active faulting and earthquake activity in this region. In southern Tibet plateau, time constraints from thermochronology dating of the Yadong-Gulu rift supports the viewpoint that these rifts had been formed by the east-westward extension of the Tibetan plateau, which was induced by a combination of strike slipping and thrusting along the main faults in southern Tibetan block (Wang et al.).

Tectonic deformation along the boundary of Ordos block

In the Ordos region, the major driving forces for active crustal deformation are complicated, and the main influence is from both the Indo-Eurasian collision and subduction of the Pacific (Deng et al., 2002; Zhang et al., 2003). The rates of Quaternary crustal deformation are much lower than those in western China. In contrast to the well-studied western and southern margins along which the great historic Pingluo and Huaxian earthquakes occurred, less work has been conducted on active faulting in the northern, northwestern and northeastern margins of the Ordos block. In northwestern margin of the Ordos block, the late Quaternary slip rate of the Zhuozishan West Piedmont fault (ZWPF) and the extension rate across the northwestern edge were determined by using an unmanned aerial vehicle system for measurement and optically stimulated luminescence (OSL) dating methods, which suggests that extensional deformation is mainly concentrated on the western side of the basin (Liu et al.). To assess the regional seismic hazard potential, Liu et al. determined seven earthquake events along the fault through trenching and OSL dating methods. According to the calculated recurrence interval and the elapsed time of the latest earthquake, they suggest that the seismic hazard potential in the northwestern corner of the Ordos block should not be underestimated. To the west, broadband magnetotelluric (MT) exploration across the Linhe basin bounded by the ZWPF on the east revealed that the Langshan Piedmont fault bounds the Linhe basin rift on the west, and is also one of the main boundary faults separating the Alxa and Ordos blocks and that the Linhe basin rift has experienced continuous and strong extension since the Cenozoic era (Zhao et al.). To the southwest, the late Quaternary activity and paleoseismicity of the nearly-north-striking, dextral Bayanhaote fault that separates the Alxa and Ordos blocks in western piedmont of the Helan Mountains were studied on the basis of analysis of the geomorphic features and trench excavations, which helps to better understand present-day deformation of this active boundary (Chen et al.).

By correlating the offset stratigraphic markers exposed in the footwall with those discovered by borehole drilling in the hanging wall, Xu et al. calculated the vertical slip rates of three faults (the Seertengshan piedmont, Wulashan piedmont, and the Daqingshan piedmont faults) bounding the northern margin of the Ordos block, which are $\sim 1\text{--}3$ mm/yr since 65 ka. Based on the slip rates and paleoearthquake results, they assessed a high seismic hazard on these faults. Another study on the Wulashan piedmont fault obtained similar results (Bi et al.). In that study, the paleoseismicity of the fault was constrained through statistical analysis of the morphology of fault scarps extracted from high resolution LiDAR topography, and the average recurrence interval estimated for the identified seven paleoseismic events also indicates a high potential seismic hazard on the Wulashan piedmont fault. Six paleoearthquake events recognized by trench excavations in late-Quaternary along the Liulengshan fault suggest that tectonic deformation in the northeastern edge of the Ordos block is also strong (Sun et al.). The three-dimensional deep structures of this region were revealed by electromagnetic inversion. The results indicate that the deep electrical structure in the western part of the northeastern Ordos is laterally heterogeneous and the migration of mantle thermal material into the middle and upper crust results in active faulting and moderate-strong earthquakes in this region (Peng et al.).

The present-day crustal deformation across the Weihe and Shanxi grabens that constitute the southern and eastern margins of the Ordos block was studied on the basis of a kinematic finite-element model constructed by joint fitting of the geological slip rate, GPS velocity field, and principal compressive stress data, which shows that this region is a low-strain kinematic setting and there is a prominent dextral shear mode along the central Shanxi graben (Song et al.). Recent one short-period seismic array study across multiple rift-basins and their boundary faults within Huai'an-Zhangjiakou region near the north end of the Shanxi graben unveils not only the low-velocity zones at the bottom of the upper crust but also the fractured Moho and crust beneath the graben, which indicates that the upwelling of the locally reheated mantle magma intrudes into the crust and thus may explain how thermal and material circulation in mantle wedge drives the crustal extension and the normal faulting along the Shanxi graben (Zhou et al.).

Active faulting and mid-strong earthquakes within blocks

Some faults in the interior of the active blocks also exhibit strong tectonic and earthquake activity. In contrast to the faults bounding the blocks, these faults within the blocks have drawn relatively less attention. They typically have low slip rates of less than 4 mm/yr and with an averaged recurrence interval of more than 1,000 years. For examples, the Nantinghe fault in the

southeastern Tibetan Plateau region shows such a low slip rate (Shi et al.) and the Yangda-Yaxu fault was constrained to have a recurrence interval of $\sim 4,000$ years for major earthquakes (Han et al.). Although not as active as the faults along the block boundaries, faults within the blocks also play important roles in accommodating crustal deformation. Surface ruptures recognized along these faults suggest that the earthquake hazard on this kind of faults is also substantial (Zuo et al.). Some earthquakes with magnitude of ~ 7.0 occurred within the blocks, such as the Ludian Ms 6.5 earthquake in Yunnan in 2014 (Xu et al., 2015) and the historic earthquakes including the 1948 Litang M7.3 earthquake with long surface ruptures on the Litang fault within the Qiangtang block (Gao et al., 2022). The Tangshan Ms7.8 earthquake occurring in the North China Plain block may be the largest earthquake ever recorded within a block. Through shallow seismic exploration combined with borehole-drilling exploration and trench excavation, Liu et al. obtained a detailed structural model of the seismogenic structure of the Tangshan earthquake and suggests that the shallow part of the structure corresponds to the strike-slip Guye-Nanhu Fault. Study of the Mae Hong Son fault in northern Thailand is another example that moderately active faults with a moderate-strong earthquake potential can develop within the active blocks (Chansom et al.).

Summary

The papers collected in this Research Topic mainly cover the following subjects: 1) late Cenozoic deformation along active block boundary, 2) geometry, surface ruptures and segmentation of active faults, 3) chronology, faulted landforms, and slip rate of active faults, 4) paleo-earthquake and large earthquake studies on the major faults along block boundary, 5) numerical models of geodynamic process of the active blocks, 6) geophysical exploration of three-dimensional deep structure, and 7) great earthquake hazard potential on the major faults along block boundary. Despite the uncertainties, we believe that these studies of this Research Topic will shed important light on the tectonic deformation pattern and dynamic model of the active blocks.

Author contributions

All authors listed have made a substantial, direct, and intellectual contribution to the work and approved it for publication.

Funding

The work throughout the conception, design and publication of this Research Topic was supported by the

National Key Research and Development Program of China (2017YFC1500101), the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901) and the National Natural Science Foundation of China (42072250).

Acknowledgments

We would like to thank all reviewers and authors for their contributions to this Research Topic. We also thank the entire team of Frontiers in Earth Science for their dedicated effort in guiding the revision and detailed editing of the papers of the Research Topic.

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Active Tectonics of the Nantinghe Fault in Southeastern Tibetan Plateau and its Implications for Continental Collision

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OPEN ACCESS

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Reviewed by:

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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 19 November 2021

Accepted: 29 December 2021

Published: 21 January 2022

Citation:

Shi F, He H, Liu Y, Wei Z and Sun H
(2022) Active Tectonics of the
Nantinghe Fault in Southeastern
Tibetan Plateau and its Implications for
Continental Collision.
Front. Earth Sci. 9:818225.
doi: 10.3389/feart.2021.818225

Quantifying the structure and kinematics of active faults is the key to understanding the India-Eurasia continental collision. While the boundary faults of blocks in the southeastern Tibetan Plateau region have been under scrutiny for decades, relatively less attention has been paid to the active faults within the blocks. Here, we use structural, sedimentary, and geomorphic characteristics to constrain the late Quaternary kinematics and slip rates of the Nantinghe fault, a major fault within the southeastern Tibetan Plateau region. The Nantinghe fault is an active, dominantly left-lateral strike-slip fault that can be traced continuously for 500 km. We analyze the late Quaternary slip rates at five sites along the fault trace. The latest Quaternary apparent throw rates are typically less than 1 mm/a. The rate of strike-slip displacement is estimated to be an order of magnitude higher, 4.0 ± 0.6 (2 σ) mm/a. Trenching suggests that active fault behavior is dominated by strike-slip faulting and reveals five earthquake events with refined information of timing in the last 5,000 years, suggesting an averaged recurrence interval of ~1,000 a. These observations agree with GPS-derived estimates and show that late Quaternary slip rates on the Nantinghe fault are comparable to those on other major active strike-slip faults in the southeastern Tibetan Plateau. The newly-obtained slip rates on the Nantinghe fault, integrated with previous estimates on other faults in this region, provide direct evidence that favors the continuum models over the rigid block ones, on the Quaternary deformation mechanism in the India-Asia continental collision.

Keywords: Southeastern Tibetan plateau, nantinghe fault, late quaternary, active faulting, slip rates, India-Asia collision

1 INTRODUCTION

The Indian-Asian continental collision during the past ca. 50 Myr has induced crustal deformation in a vast region that extends more than 2,000 km from the Himalaya Mountains to Central and East Asia (Molnar and Tapponnier, 1975; Royden et al., 2008; Taylor and Yin, 2009). Two end-member views dominate the debates on continental collision dynamics: 1) block models, in which intercontinental deformation is concentrated on major faults that separate a number of relatively rigid blocks (Dewey et al., 1973; Avouac and Tapponnier, 1993; Peltzer and Saucier, 1996; Tapponnier et al., 2001; Meade, 2007; Thatcher, 2007); and 2) continuum models, in which

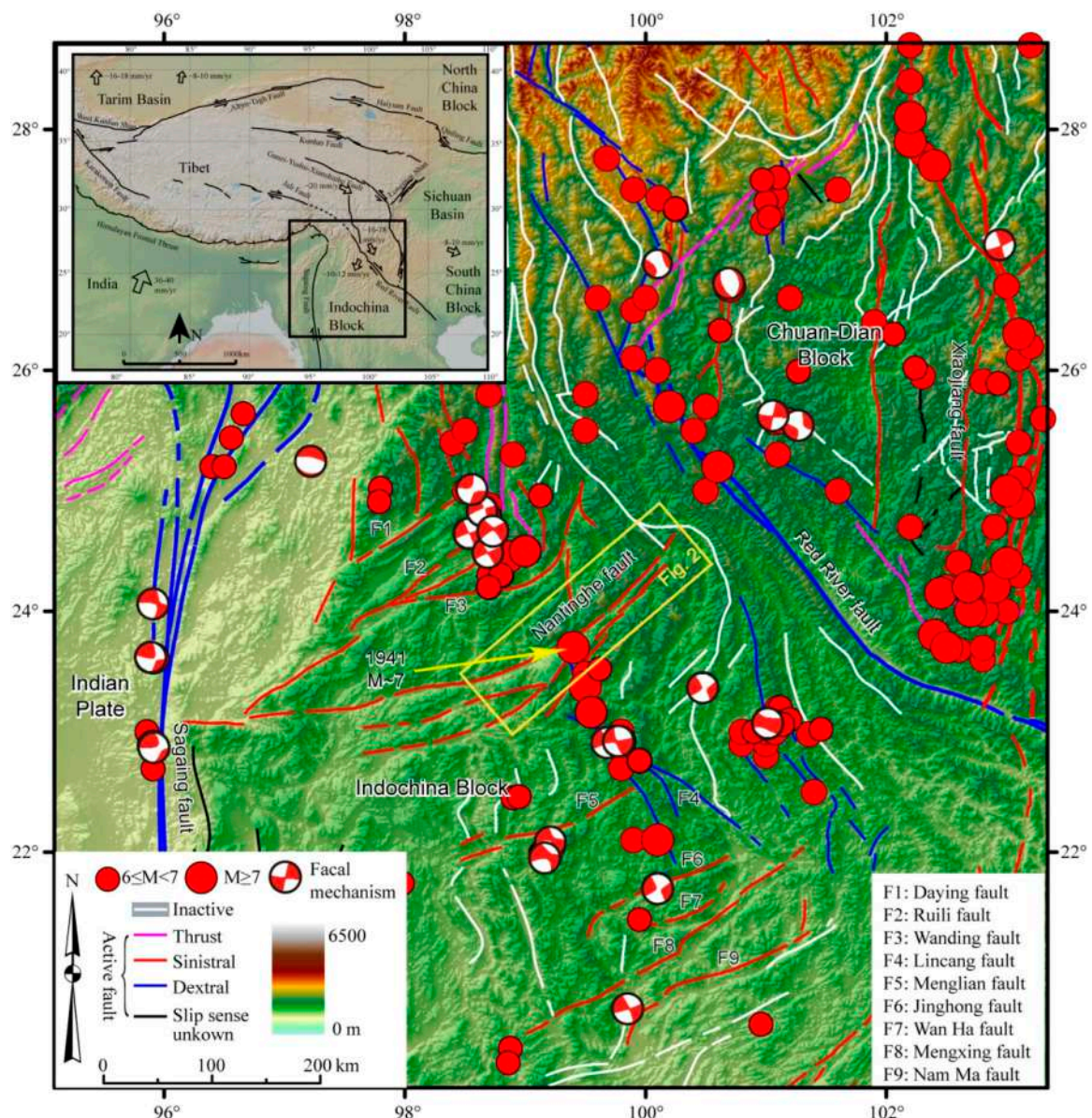


FIGURE 1 | Major faults and earthquakes in southeastern Tibetan Plateau. White lines indicate inactive faults, whereas magenta, red, blue, and black lines represent active faults with different slip senses. Fault locations are modified from (Deng, 2007). Focal mechanism solutions from the ISC-GEM (Storchak et al., 2013; Storchak et al., 2015; Di Giacomo et al., 2018). Red circle indicates instrumentally record earthquakes (China Earthquake Networks Center, 2019) and historical earthquakes (China Earthquake Administration, 1995; China Earthquake Administration, 1999). The shaded topography is created from USGS SRTM data. The yellow rectangle shows the extent of **Figure 2**. The inset at the top left shows the major boundary faults and blocks of the Tibetan Plateau (simplified and modified from (Tapponnier et al., 2001). The black box shows the extent.

deformation is regionally distributed in the shallow brittle crust, and is essentially continuous at depth (Molnar and Tapponnier, 1975; England and McKenzie, 1982; England and Houseman, 1986; Clark and Royden, 2000). When these two schools of view are applied to eastern Asia, large slip rates on major faults are required by the block models, but not by the continuum models. In general, the block models predict large slip rates (mostly over 10 mm/a (Tapponnier et al., 2001; Thatcher, 2007) on the major faults between tectonics blocks and small or negligible slip rates

(far less than the slip rates on the major faults between tectonic blocks (Lacassin et al., 1998; Tapponnier et al., 2001; Thatcher, 2007) on the faults within the blocks. The continuum models, in contrast, do not imply such a difference, and the predicted slip rates are all around or less than 5 mm/a (England and Molnar, 2005).

The southeastern Tibetan Plateau, including the Indochina Block and the Chuan-Dian Block, is a natural laboratory to assess this debate (Tapponnier et al., 2001) (**Figure 1**). The block models

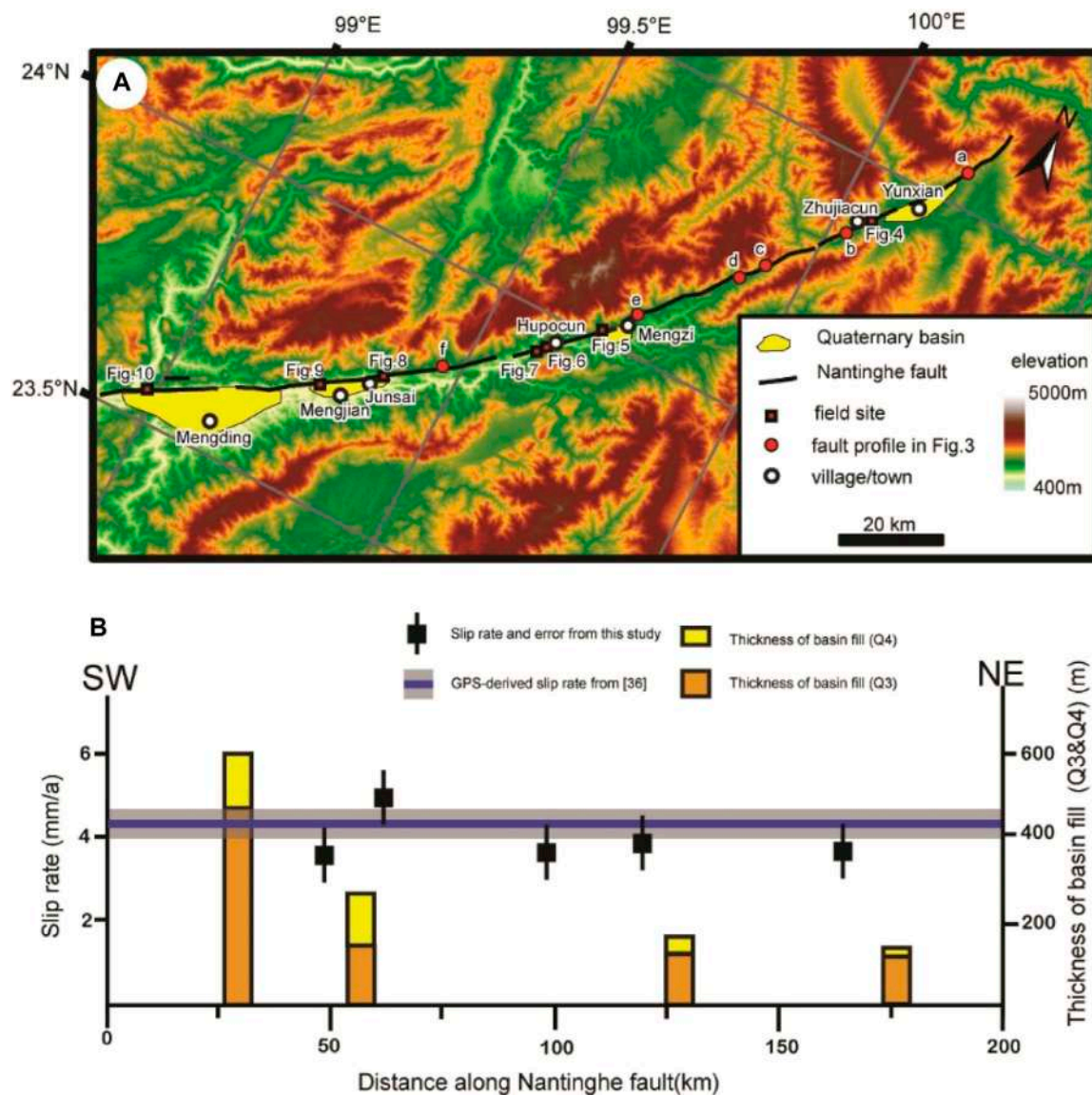


FIGURE 2 | (A) Fault map of the active Nantinghe fault. **(B)**, strike-slip rates and depth of basin along the Nantinghe fault. Blue line with shadows is strike-slip rates and errors estimated from GPS surveys (Wang, 2009). Black squares with black line are strike-slip rates and error bars estimated in this study. Orange rectangle is the thickness of the older basin fill (Q3), yellow rectangle is the thickness of the younger basin fill (Q4).

suggested that the boundary fault, Ailao Shan-Red River shear zone or Red River fault, dominates the extrusion of the Indochina block during the past 40 Myr, and predicted that the slip rates of the strike-slip faults (e.g., Daying, Ruili, Nantinghe, Mengxing, among others) within the Indochina block should be substantially lower than that of the Red River fault (Lacassin et al., 1998). In contrast, in the continuum models, although the Red River fault forms a striking geomorphic feature, it does not dominate the present-day kinematics of the whole block, and the slip rates of the internal faults may be comparable with, or even greater than, that of the Red River fault (Copley, 2008). Because of their obvious geomorphic features, fast slip rates, and the occurrence of devastating earthquakes, the block-bounding

Red River and Sagaing faults have attracted intensive investigations (Allen et al., 1984; Tapponnier et al., 1990; Weldon et al., 1994; Leloup et al., 1995; Wang et al., 1998; He et al., 2002; Shen et al., 2005; Leloup et al., 2007; He and Oguchi, 2008; Wang et al., 2014; Shi et al., 2016; Shi et al., 2018a). In contrast, a lesser amount of attention has been paid to the active faults within the blocks. Seismic station networks and historical documents have recorded numerous earthquakes with magnitudes greater than 6 within the block (China Earthquake Administration, 1995; China Earthquake Administration, 1999; China Earthquake Networks Center, 2019). GPS measurements and geomorphic analysis also indicated that the interior of the Indochina Block is actively deforming (Copley, 2008; Wang et al.,

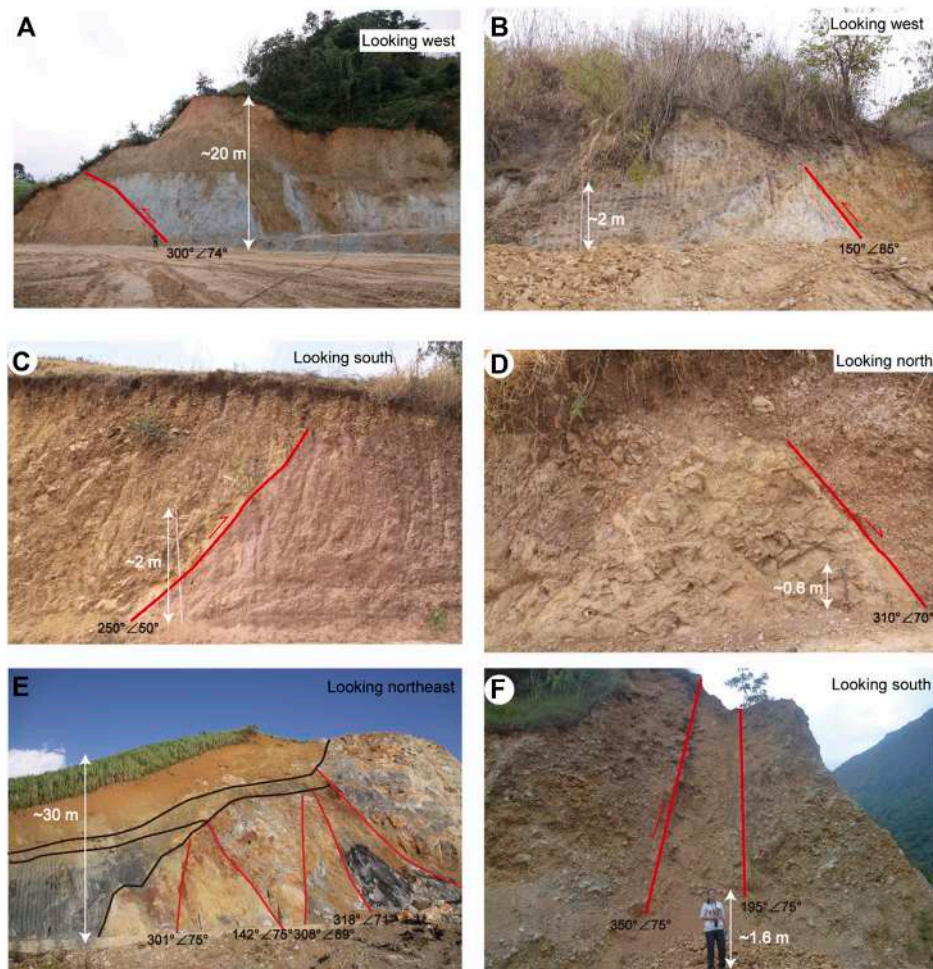


FIGURE 3 | Fault profile photo of the Nantinghe fault. Their locations are shown in **Figure 2**. Fault surface attitudes are reported in the format of dip direction and dip. Slip sense is determined by the stratigraphic ages on both sides of the fault.

2014; Shi et al., 2018b). However, to our knowledge, some faults within the Indochina Block remain enigmatic in terms of their slip rates and the timing and recurrence interval of paleoearthquakes during the Quaternary. This makes it difficult to quantify the kinematics and seismicity of the block deformation, and hence hinders the evaluation of the dynamics of continental collision in the southeastern Tibetan Plateau region.

In this study, we choose the Nantinghe fault, the largest one within the Indochina Block, to assess its activity and testify the competing models. While the GPS-derived sinistral strike-slip rate of the Nantinghe fault (~ 4 mm/a) (Wang, 2009) is close to the field- and GPS-derived sinistral strike-slip rate of the Red River fault (~ 5 mm/a) from decades to tens of thousands years (Allen et al., 1984; Weldon et al., 1994) or even larger than the slip rate of the Red River fault (~ 1 mm/a) from stratigraphic and geomorphologic investigations over the last 30,000–50,000 years (Shi et al., 2018a), how fast it slipped at a greater time scale is still unknown. We focus on the northeastern 200 km of the Nantinghe

fault zone (**Figure 2**). This is because 1) geological evidence is more extensively exposed along the northeastern portion of the fault zone, and 2) the southwestern portion of the fault in Myanmar is relatively inaccessible. We adopt a combination of techniques, including field mapping, image interpretation, surveying of offset geomorphic markers, and trenching, in order to elucidate the history of fault slip over the last few thousand years, and to fill a gap in the paleoearthquake records in this forestry and remote area.

2 GEOLOGICAL SETTING

The Indochina Block lies in the southeast of the Tibetan Plateau. It is bounded by two major strike-slip boundaries, the Red River fault to the east and the Sagaing fault to the west (**Figure 1**). Many strike-slip faults cut the region between these faults, forming rugged terrain whose elevation ranges from 500 to 3,000 m. Most

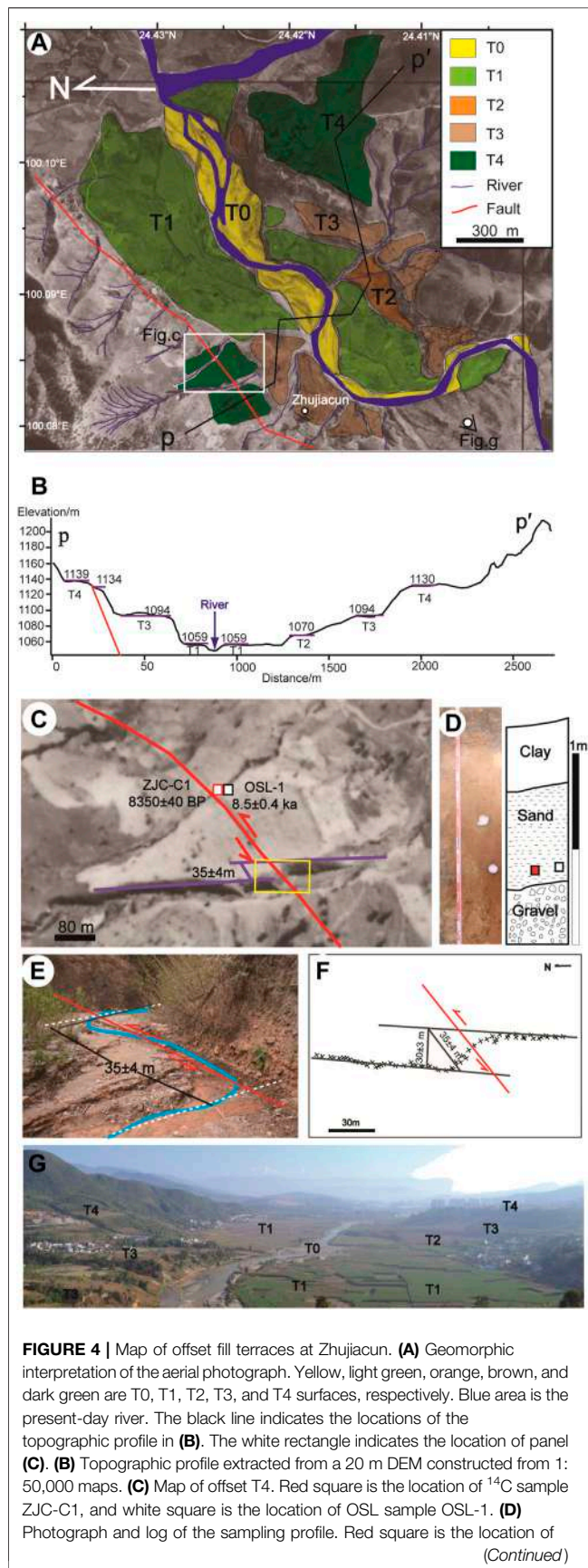


FIGURE 4 | the ^{14}C sample, and white square is the location of the OSL sample. (E) Photograph of the offset, view to NW. (F) Map of offset, obtain from differential GPS measuring system. Cross symbol shows the measured location of the riser. (G), Photograph of terraces, viewing to the northeast.

of these faults strike NE-SW and have arc-shaped surface traces. The Indochina Block moved southeastward relative to the Eurasian plate as an extrusion body along the sinistral strike-slip Ailaoshan-Red River shear zone, and the dextral strike-slip Red River fault during the early stage (between ca. 35 and 17 Ma) of the India-Eurasia collision (Tapponnier et al., 1990; Leloup et al., 1995; Leloup et al., 2007). As the Indian plate continued moving north, an inversion occurred in the Indochina block, switching from east-west compression to east-west extensions. Slip sense on the Red River fault changed to right-lateral before at least 5 Ma (Leloup et al., 2007) or even 8–20 Ma (Wang et al., 2016; Wang et al., 2019; Wang et al., 2020).

As one important fault within the Indochina Block, the Nantinghe fault initiated at the end of the early Paleozoic and was reactivated in different geological periods (Huang and Ren, 1987). Slip sense on the Nantinghe fault that changed to left-lateral is associated with slip sense on the Red River fault that changed to right-lateral (Lacassin et al., 1998). The eastern tip of the Nantinghe fault zone is located northwest of Yunxian County, Yunnan Province, SW China. It extends westward through Mengzhi, Mengjian, Mengding towns in Yunnan, China into Shan State in eastern Myanmar. Where it is detectable on the surface, the fault appears to dip 70–85°NW in the northeastern segment and 70–80°SE in the southwestern segment (Shi, 2014).

Exposures of the active Nantinghe fault show that it coincides with zones of dense fracturing in Pre-Quaternary bedrocks; the width of these zones is generally ~10–50 m but can be occasionally up to a few hundred meters. In remote sensing imagery, strands of the fault form clear linear features associated with scarps, shutter ridges, and offset drainages, characteristic of active strike-slip deformation. The fault movement is dominated by left-lateral strike-slip, with minor net vertical slip (Shi, 2014). Previous estimates of the fault activity vary, especially on the northern segment (Zhu et al., 1994; Wang et al., 2006; Sun et al., 2017). Moreover, some parameters of activity, such as slip rates of all fault and the paleoearthquake events of south segments, are still poorly constrained for the Nantinghe fault.

3 METHODS AND TECHNIQUES

3.1 Fault Mapping

We mapped the active traces of the Nantinghe fault zone (Figure 2) using the Advanced Land Observation Satellite (ALOS) imagery (2.5 m spatial resolution) and Chinese aerial photographs (~1 m spatial resolution) (Chen et al., 2006). We then carried out field observations at sites with geomorphic indicators of late Quaternary activity, including scarps or offset surfaces in Quaternary deposits, offset channels, shutter ridges, and linear valleys (Figure 3). Offset landforms were

TABLE 1 | Optically stimulated luminescence (OSL) data and ages.

Sample	Site name	Counts/ks	K ₂ O (%)	Total dose rate (Gy/ka)	Equivalent dose (Gy)	Age (Ka)
OSL-1	Zhujiacun	12.1 ± 0.4	1.66	3.7 ± 0.2	31.6 ± 0.9	8.5 ± 0.4
OSL-2	Daxueshan	9.2 ± 0.3	1.42	3.0 ± 0.2	5.5 ± 0.1	1.8 ± 0.1
OSL-3	Hupocun	12.6 ± 0.4	2.16	4.2 ± 0.2	55.7 ± 2.1	13.2 ± 0.7
OSL-4	Junsai	5.6 ± 0.2	0.80	1.8 ± 0.1	24.5 ± 0.8	13.3 ± 0.5
OSL-5	Mengjia	8.3 ± 0.3	0.92	2.4 ± 0.1	9.0 ± 0.2	3.7 ± 0.2
OSL-6	Mengding	8.5 ± 0.4	1.23	3.3 ± 0.2	3.6 ± 0.2	1.1 ± 0.1

Note: All sample preparation and analyses were conducted by the State Key Laboratory of Earthquake Dynamics, China. See detail in the **Supplemental Material**.

TABLE 2 | Radiocarbon sample information and analytical results.

Sample	Material	Radiocarbon age	13C/12C ratio	2σ Calendar age
ZJC-C1	Charcoal	8350 ± 40 BP	−25.9‰	7522–7328 BC
HPTC-C1	Charcoal	520 ± 40 BP	−18.7‰	1315–1356 BC (21.3%) 1388–1448 BC (74.1%)
MDTC-C5	Charcoal	4970 ± 40 BP	−26.0‰	3929–3877 BC (9.8%) 3804–3654 BC (84.6%)
MDTC-C17	Charcoal	1710 ± 30 BP	−27.2‰	251–697 AD
MDTC-C38	Charcoal	690 ± 30 BP	−26.2‰	1265–1312 AD (68.9%) 1358–1366 AD (26.5%)

Note: All sample preparation and analyses were conducted by Beta Analytic Inc. All samples were analyzed using accelerator mass spectrometry. All raw radiocarbon ages were calibrated by OxCal 4.3 (Bronk Ramsey, 2009). The calibration calculates probability distributions for raw radiocarbon ages with associated uncertainties (reported by the lab facility). Radiocarbon ages BP, relative to 1950. All samples typically underwent the acid–alkali–acid (AAA) method before radiocarbon dating.

surveyed using a differential GPS (DGPS) measuring system, with a measurement repeatability of finer than ±10 cm. For some large-scale offsets, we estimated it from the ALOS imagery and aerial photographs beside the field survey.

The majority of the evidence for active faulting came from offset or truncated fluvial fill terraces. Fill terraces in this region are typically composed of subhorizontal, crudely- to well-bedded gravel and sand layers. We identified terrace surfaces and assigned relative ages based on the relative height between the surface and the modern river bed. At each site, terraces were numbered in ascending order from the youngest (T0, representing the modern channel bed) to the oldest. We further determined absolute ages of fill terrace deposition and abandonment using ¹⁴C and optically stimulated luminescence (OSL) dating techniques on samples from the sand layers (Aitken, 1998). The ¹⁴C analysis is performed in the Beta Analytic Radiocarbon Dating Laboratory. The raw radiocarbon ages are calibrated by OxCal 4.3 (Bronk Ramsey, 2009). The OSL dating is conducted in the State Key Laboratory of Earthquake Dynamics. We did not obtain sufficient evidence to assess whether terrace ages can be correlated between different sites along the Nantinghe fault, or whether the terrace chronology is site-specific.

While the strike-slip component prevails along the entire fault, along-strike variations exist in terms of the ratios of strike-slip and dip-slip components as well as the strike-slip rates. We describe our observations below in both horizontal and vertical offsets. Throw rates are relatively small (less than

1 mm/a) along each fault segment, and the ratio of strike-slip and dip-slip components varies.

3.2 Trenching

To document the timing of recent slips on the fault, we excavated two trenches across the Nantinghe fault at two sites. Accurate location of fault traces and sufficient sediment accumulation with organic materials are two key factors for trenching. In Mengding township (**Figure 2**), we established a site that is located at a well-developed fault trough valley on an alluvial fan and excavated a 2-m-deep trench across the fault. The steep terrain south of the site has trapped sediment, providing abundant organic materials. To assess possible differences in fault activity between different segments, we excavated another 2-m-deep trench across the fault at Hupocun township (**Figure 2**). This site is also located on an alluvial fan surface. The steep terrain to the north of the site provides sufficient sediment accumulation and organic materials. All trench walls were cleaned and logged at a scale of 1:10 following standard procedures (McCalpin, 1996). Depositional ages of the units in the trench facies were provided by ¹⁴C analysis of charcoal fragments and OSL analysis on sands.

4 RESULTS

The Nantinghe fault can be traced continuously for approximately 200 km in Yunnan Province, SW China. The fault shows clear evidence for Quaternary sinistral strike-slip displacement, with minor components of dip slip. Field-based

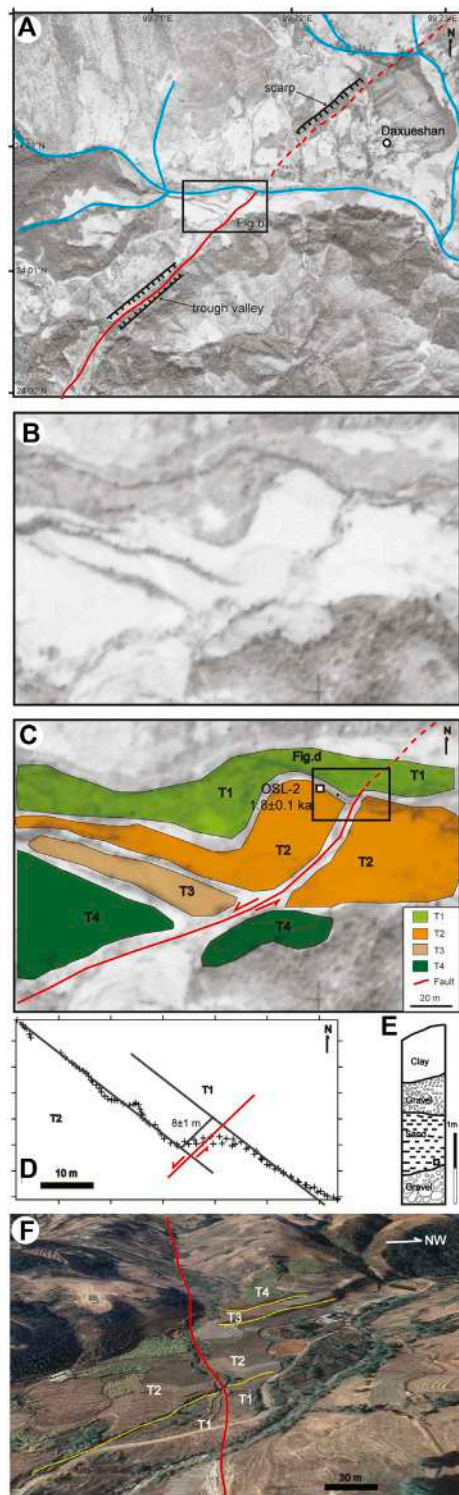


FIGURE 5 | Map of offset fill terraces at Daxueshan. **(A)** Aerial photograph of Daxueshan, the black rectangle indicates the location of panel **(B)**. **(B)** Original aerial photograph. **(C)** Geomorphic interpretation of the aerial photograph. Light green, orange, brown, and dark green polygons denote T1, T2, T3, and T4 surfaces, respectively. White square is the location of OSL sample OSL-2, with its age. The black rectangle indicates the location

(Continued)

FIGURE 5 | of panel **(D)**. **(D)**, Map of offset T1/T2 riser, obtain from differential GPS measuring system. Cross symbol shows the measured location of the T1/T2 riser. **(E)**, Log of the sampling profile. Black square shows the location of the sample OSL-2. **(F)**, Geomorphic interpretation of the Google Earth images, in oblique view with no vertical exaggeration. Location: 24.0234°N, 99.7165°E. Viewing southwest. Red curve denotes the surface trace of the Nantinghe fault.

geology data, especially fault profiles, are key to identifying fault traces. Several fault profiles (**Figure 2**) and their tectonic geomorphology characteristics form the pieces of evidence of fault traces along the Nantinghe fault. Below we document the results from northeast to southwest along the fault.

4.1 Zhujiacun

Zhujiacun is located southwest of Yunxian county. Offsets of fill terraces at this site are clearly shown on aerial photographs (**Figure 4**). We then mapped these terraces in the field investigations. Fill terraces are typically composed of subhorizontal, crudely- to well-bedded gravel and sand layers. We identified five terrace surfaces (T0, T1, T2, T3, and T4) based on the relative height between the surface and the modern river bed (**Figure 4B**). The fault has caused a sinistral offset to the gully of T4 by approximately 35 m, and a vertical offset of the T4 surface by 5 m. The ages of T4 were estimated at 8.5 ± 0.4 ka based on our OSL dating and 8350 B.P. based on ^{14}C dating (**Tables 1, 2**). Both methods yield a similar age, indicating the age of T4 is ca. 8.5 ka. Based on the sinistral strike-slip displacement of the gully of T4, the vertical offset of the T4 surface, and the age of T4, we estimated the mean sinistral-slip rate of 4.1 ± 0.4 mm/a and the throw rate of ~ 0.6 mm/a. This is a minimum estimate of the strike-slip rate because the gully formed after any of the T4 sediments were deposited. No other clear terrace offsets are preserved at this site.

4.2 Daxueshan

Daxueshan is located on the southwestern margin of the Mengzi basin, about 60 km southwest of Yunxian County (**Figure 2**). Immediately south of this site, a trough valley is present, indicative of the Nantinghe fault trace in the Pre-Quaternary bedrocks (**Figure 5**). At this site, four terraces T1-T4 are all cut by the Nantinghe fault. The vertical dislocation between the hanging wall and the footwall is 5 m on the T3 terrace and 1 m on the T2 terrace, but no vertical displacement is observed on T1. The T2/T1 terrace riser is offset in a sinistral sense. Because of the twists and turns of the T3/T2 riser, it is difficult to measure accurately the sinistral offset of the T3/T2 riser.

We used the sinistral offsets of the T2/T1 riser and the T2 age to evaluate the sinistral-slip rate, and the throw rate was constrained from the height of the scarp on T2 and the T2 age. The sinistral and vertical offsets, as obtained from DGPS surveys, are 8 and 1 m, respectively. The ages of T2 at this site are 1.8 ± 0.1 ka based on OSL dating on sample OSL-2 (**Table 1**), thereby the sinistral-slip rate and the throw rate are 4.5 ± 0.8 mm/a and 0.5 ± 0.1 mm/a, respectively. Again, this is a minimum estimate of the slip rate because the riser could only form after the T2 sediments were deposited.

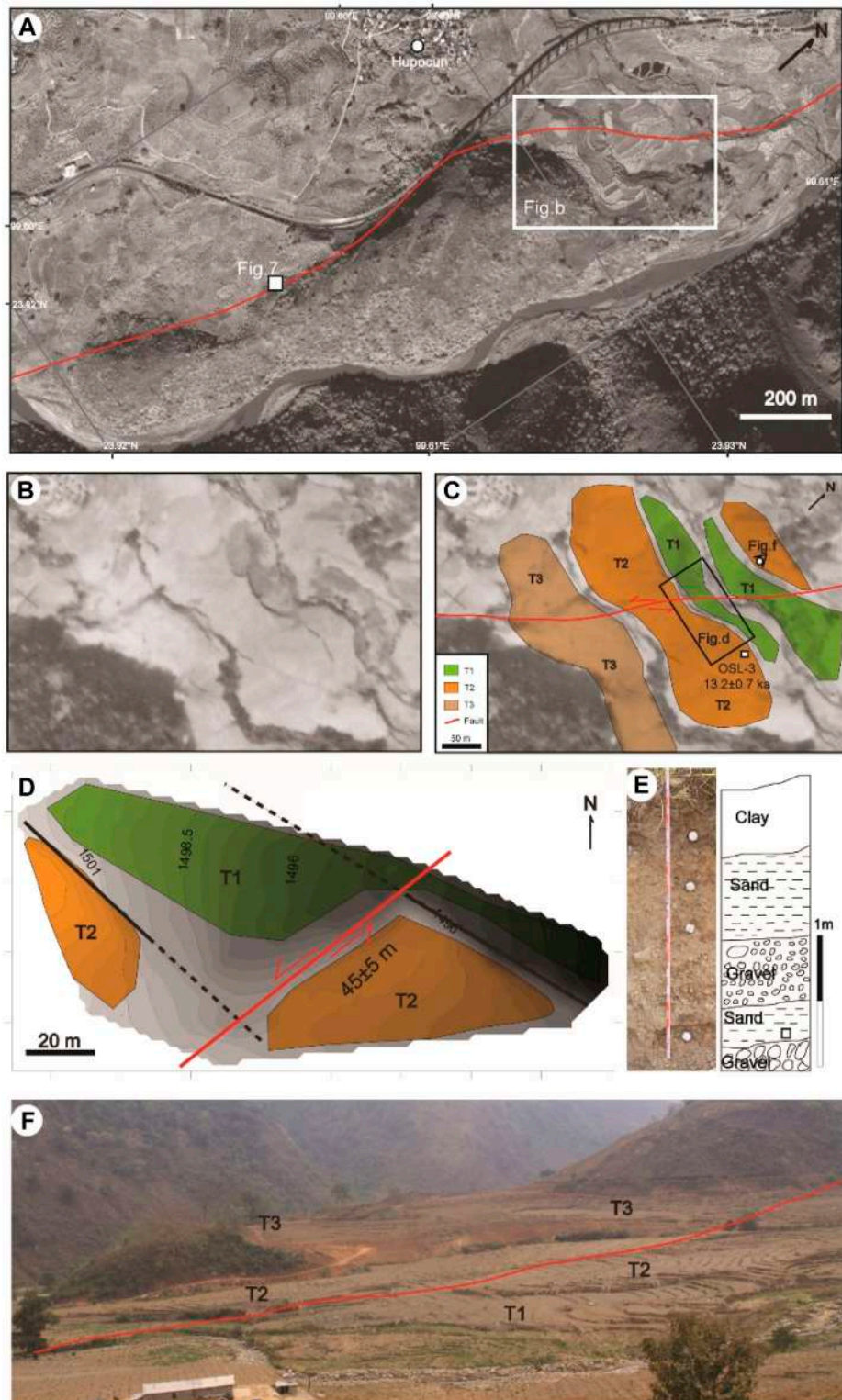


FIGURE 6 | Map of offset fill terraces at Hupocun. **(A)** Gaofen 2 satellite image of Daxueshan, the white rectangle indicates the location of panel **(B)**, white square shows the location of panel **(C)**. **(B)** Original aerial photograph. **(C)** Geomorphic interpretation of the aerial photograph. Light green, orange, and brown polygons represent T1, T2, and T3 surfaces, respectively. Black rectangle indicates the location of panel **(D)**. White square shows the location of the OSL sample (OSL-3) and its age (13.2 ± 0.7 ka). **(D)**, Topographic map of offset T1/T2 riser (0.5 m contour interval), obtained from differential GPS measuring system. **(E)** Photograph and log of the sampling profile. White square is the location of the OSL sample. **(F)** Photograph of terraces, viewing to the south.

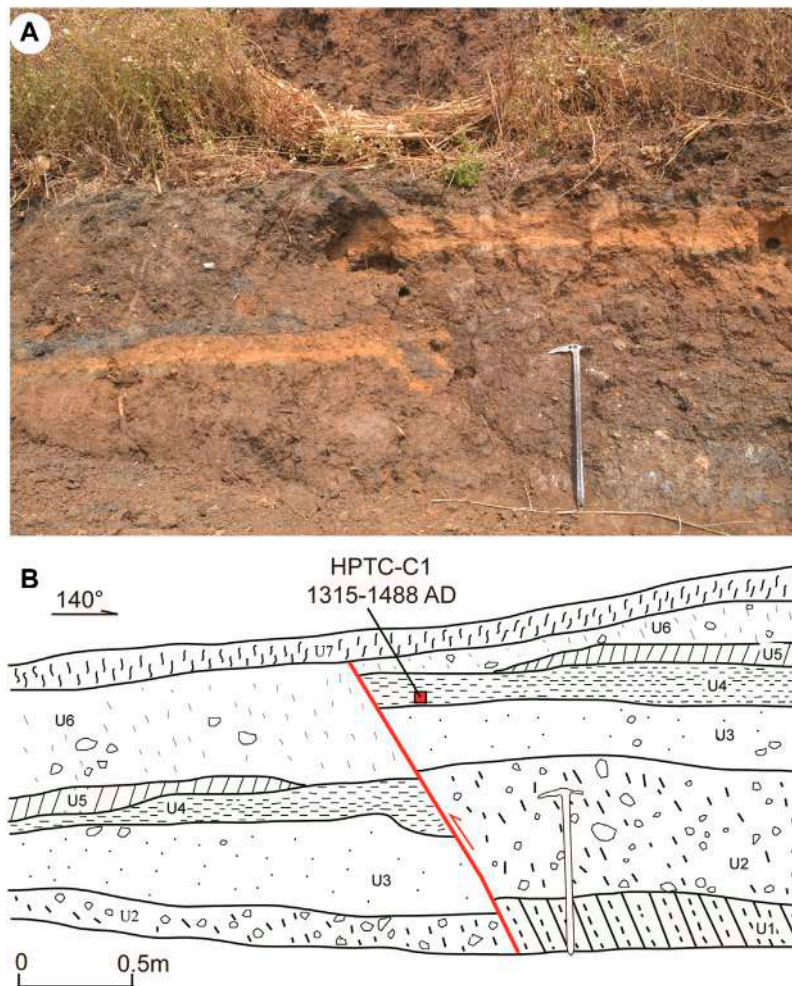


FIGURE 7 | Trench site on the Nantinghe fault at Hupocun. **(A)** Photo of the Hupo trench HPTC. **(B)** Log of the Hupo trench HPTC. Red square shows the location of the ¹⁴C sample HPTC-C1. Ice ax for scale.

TABLE 3 | Lithologic units in the HPTC trench.

Unit	Description
U1	Devonian phyllite, weathered serious
U2	Grey gravel layer. The mean clast diameter is about 5 cm, with rare clasts over 10 cm
U3	Grey sand layer with rare small gravel
U4	Yellow clay layer with rare gravel
U5	Celadon sand layer with rare small gravel
U6	Grey gravel layer. The mean clast diameter is about 5 cm, with rare clasts over 10 cm
U7	plough layer

4.3 Hupocun

Three terraces (T1-T3) are found near Hupocun village, about 6 km southwest of the Daxueshan site (Figure 2). All terraces are cut by the Nantinghe fault (Figure 6). T3 is only preserved on the southern side of a creek. The throw is 10 m on the T3 terrace and 3–4 m on the T2 terrace and the T2/T1 riser is offset left-laterally

by about 50 m. We use the sinistral offsets of the T2/T1 riser and the T2 age to calculate the sinistral-slip rate, and the throw rate is estimated from the height of the scarp on T2 and the T2 age. The sinistral and vertical offsets, as obtained from DGPS surveys, are 45 ± 5 m and 4 ± 0.5 m, respectively. The age of T2 at this site is estimated at 13.2 ± 0.7 ka based on OSL dating on sample OSL-3 (Table 1). Thereby, the sinistral-slip rate and throw rate are 3.4 ± 0.6 mm/a and 0.3 ± 0.1 mm/a, respectively. Again, these rates are conservative estimates.

We also excavated a trench (HPTC) at this site, on the south of the terraces. The fill terraces at this site are typically composed of subhorizontal, crudely- to well-bedded gravel layers with a few sand layers. The HPTC trench was excavated in the alluvial fan with well-bedded sand layers. Seven distinct lithologic units were identified in the trench walls (Figure 7) and are described in Table 3. Only one fault plane was recognized. The fault cuts the units U1 to U6 and is overlain by U7, a plough layer (Figure 7). We collected a charcoal sample, HPTC-C1, from U4 (Figure 6) and obtained its ¹⁴C age, 1315–1488 AD (Table 2).

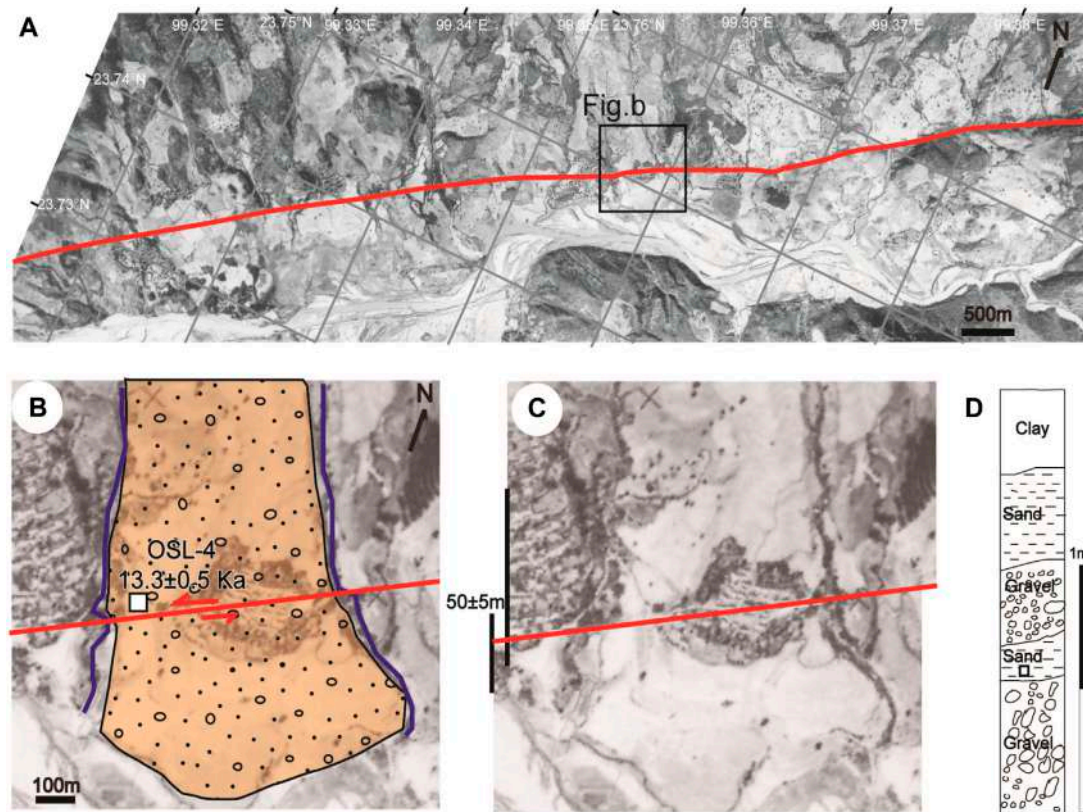


FIGURE 8 | Map of offset alluvial fan at Junsai. **(A)** Aerial photograph of Junsai. Black rectangle indicates the location of panel **(B)**. **(B)** Geomorphic interpretation of the aerial photograph. **(C)** Restoration of the alluvial fan to the pre-offset condition. **(D)** Log of the sampling profile. White square is the location of the OSL sample OSL-4.

4.4 Junsai

Further southwest, about 100 km from Yunxian County, we identified obvious offsets of the alluvial fan and the gully at a site near Junsai township (**Figure 8**). We measured horizontal offsets on the aerial photographs. The alluvial fan is sinistrally offset by about 50 m. OSL dating on sample OSL-4 yields 13.3 ± 0.5 ka for the age of the alluvial fan at this site (**Table 1**). Therefore, the sinistral-slip rate, at its minimum, is 3.8 ± 0.2 mm/a using the age and the displacement of the alluvial fan. Because of the artificial reforming, we could not obtain the vertical offset of the alluvial fan.

4.5 Mengjian

In the same Quaternary basin, west of Mengjian, offsets of an alluvial fan and a gully are clearly shown on the aerial photographs (**Figure 9**). We measured horizontal offsets from aerial photographs. The alluvial fan is left-laterally offset by about 20 m. The age of the alluvial fan at this site is 3.7 ± 0.2 ka based on OSL dating on sample OSL-5 (**Table 1**). Hence, the sinistral-slip rate is 5.4 ± 0.3 mm/a using the age and the sinistral offset of the alluvial fan. Due to artificial reforming, the vertical offset of the alluvial fan is too inexplicit to obtain.

4.6 Mengding

The Mending basin is the largest and deepest Quaternary basin along the Nantinghe fault (**Figure 2**). At the southeast

corner of the Mending basin, a river clearly shows left-laterally offset, marking the fault traces (**Figure 10**). We excavated a trench, MDTC, in the alluvial fan near the river, across to the fault trace (**Figure 11**). Exposed faults on the trench wall have high apparent dips ($>60^\circ$) to the northwest and show a top-NW normal sense of slip. Rocks in the footwall block of the master fault are composed primarily of consolidated gravel layer Unit 1, overlain by Unit 16 (**Table 4**). The presence of a resistant consolidated gravel layer at the footwall has allowed the accumulation of at least two distinct sediment wedges in the hanging wall. The fault system is expressed as a negative flower structure with a general dip to the north, and five separate fault planes can be identified.

The MDTC trench shows a prolific record of fault activity. Fault F1 cuts unit U1-U14 and appears to terminate at the base of U16. Fault F2 cuts U2-U11 and appears to terminate at the base of U12. Fault F3 truncates unit U2-U12 and is overlain by U14. Fault F4 offsets unit U2-U9 and is covered by U10. Fault F5 cuts U2-U13 and is concealed by U14. We infer that five paleoearthquake events are recorded by the faulting shown in this trench, all of which postdated the deposition of U2. Three ^{14}C samples, MDTC-C5, C17, and C38 were collected from U2, U5, and U11 layers, respectively, and their ages are 4970 ± 40 BP, 1710 ± 30 BP, and 690 ± 30 BP, respectively. An OSL sample, OSL-6, from U9 yields 1.1 ± 0.1 ka (**Tables 1, 2**).

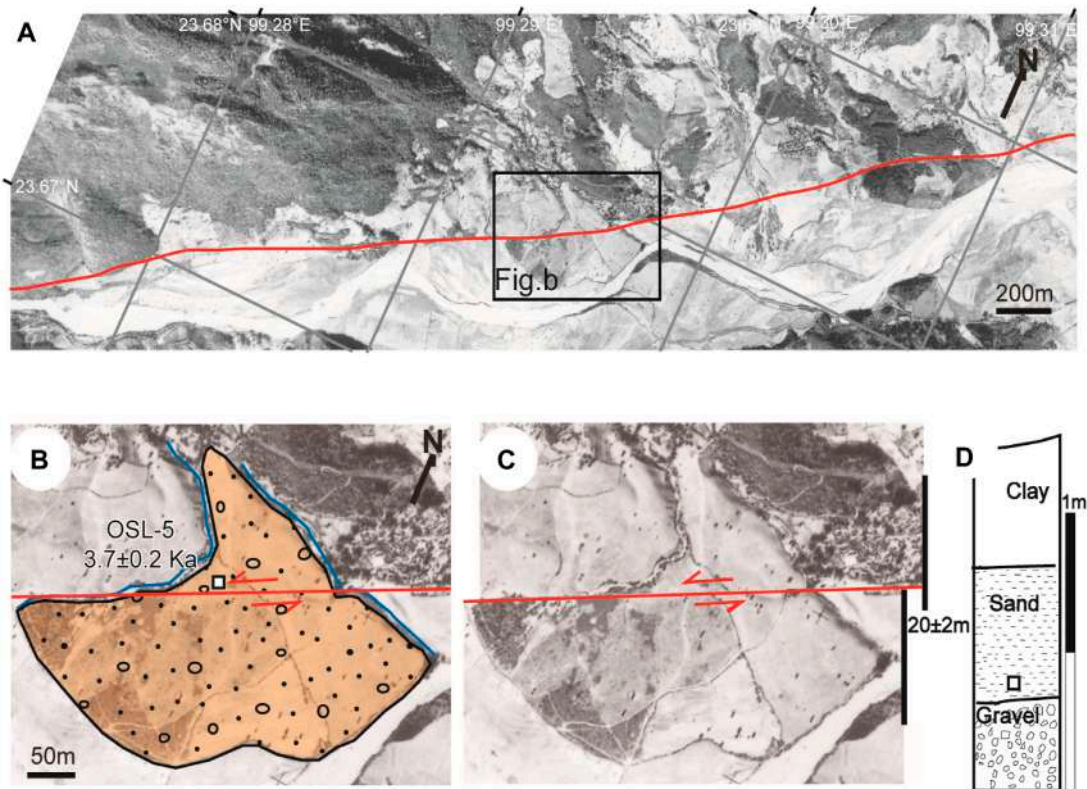


FIGURE 9 | Map of offset alluvial fan at Mengjian. **(A)** Aerial photograph of Mengjian. Black rectangle indicates the location of panel **(B)**. **(B)** Geomorphic interpretation of the aerial photograph. **(C)** Restoration of the alluvial fan to pre-offset condition. **(D)** Log of the sampling profile. White square marks the location of the OSL sample OSL-5.



FIGURE 10 | Satellite image that shows the offset river (blue curve), fault surface trace (red curve), and the location of trench site MDTC (white box) along the Nantinghe fault at the Mengding site.

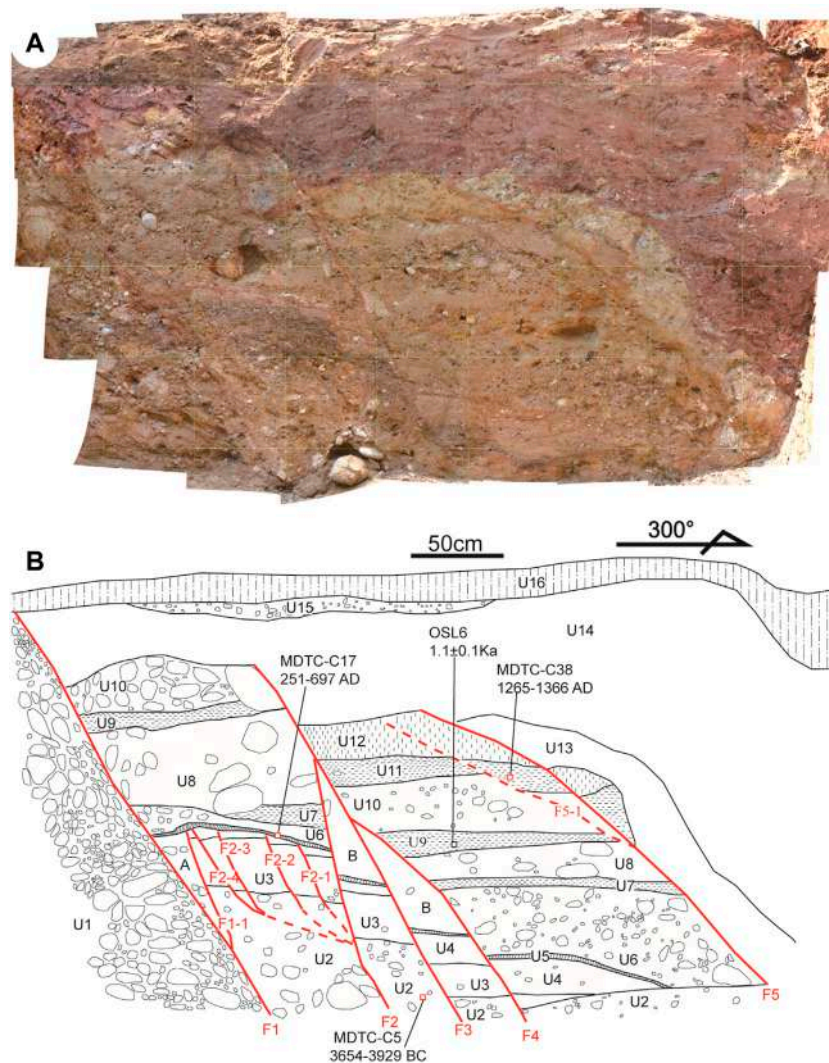


FIGURE 11 | Trench site on the Nantinghe fault at the Mengding site. **(A)** Photo of the Mengding trench (MDTC). **(B)** Log of the Mengding trench MDTC. Red line denotes fault. Sample locations for radiocarbon and OSL dating (MDTC-C5, C17, and C38, and OSL-6, respectively) and their ages are also annotated.

5 DISCUSSION

5.1 Paleoearthquake Records

Previous work revealed a ~45,000 year-long, incomplete paleoearthquake history in the northern segment of the Nantinghe fault (Sun et al., 2017). No earthquake greater than 6 along the Nantinghe fault is recorded in history so far, except for the 1941 M7 earthquake near the Myanmar-China border. In contrast, many earthquakes have occurred in the nearby region (Figure 1).

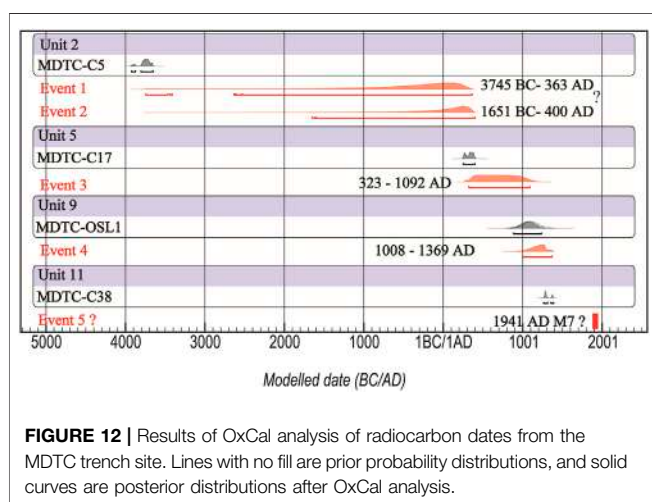
The Mengding trench MDTC provides us an opportunity to document the paleoearthquakes. We use OxCal 4.3 with the three ^{14}C (MDTC-C5, MDTC-C17, MDTC-C38) samples and the OSL sample (OSL-6) to estimate the timing of the paleoearthquake events (Figure 12). At least four events are identified: event 1 (3745 BC–363 AD), event 2 (1651 BC–400 AD), event 3 (323–1092 AD),

and event 4 (1008–1369 AD), while the fifth event that corresponds to the 1941 AD M7 event is not revealed by our data.

We illustrate the sequence of these events in a set of diagrams (Figure 13). The first paleoearthquake event took place after the deposition of U2 and formed U3, which may be a sedimentary wedge (Figures 13A,B). Materials in U3 appear to be derived from U1. The ^{14}C sample MDTC-C5 from U2 indicates that the lower limit time of the first paleoearthquake is 3929–3654 BC (Figure 12). Unit U4 was deposited after the first event (Figure 13C). The second paleoearthquake event occurred after the deposition of U4 and involved both faults F1 and F2 (Figure 13D). Unit U5 was deposited after the second event (Figure 13E). The third paleoearthquake event took place after the deposition of unit U5 and involved faults F1 and F2 (Figure 13F). The ^{14}C sample MDTC-C17 from U5 indicates that the upper limit time of the second paleoearthquake and the

TABLE 4 | Lithologic units in the Mengding trench MDTC.

Unit	Description
U1	Grey-yellow coarse gravel layer, dominated by gravel and cobbles with a diameter of 3–5 cm and 15–20 cm, with rare clasts over 20 cm. The gravel is poorly sorted, the 3–5 cm gravel is poorly rounded but the 15–20 cm gravel is slightly rounded
U2	Grey gravel layer. The mean clast diameter is about 5 cm, with rare clasts over 10 cm
U3	Black sand gravel layer. Rich in organic matter and thins out from east to west
U4	Grey sand layer, thin out from east to west
U5	Prunose clay with rare small gravel
U6	Grey gravel layer. The mean clast diameter is about 5 cm, with rare clasts over 10 cm
U7	Grey sand layer with rare small gravel
U8	Grey-yellow sand gravel layer
U9	Grey sand layer
U10	Grey gravel layer. The gravel is poorly sorted and is slightly rounded
U11	Grey sand layer
U12	Grey fine sand layer
U13	Grey-white sand clay layer
U14	Crimson clay layer with rare grey gravel. The top boundary is nearly horizontal, but the bottom boundary fluctuates and is affected by fault activity
U15	Tawny sandy gravel layer, dominated by gravel and cobbles with a diameter less than 2 cm. The gravel is generally sorted and finely rounded
U16	Artificial accumulation
A	Grey sand and fine gravel wedge. The grain size of the fine gravel is 3–5 cm. Materials in the wedge are similar to those mixtures that form U2, U3, and U4
B	Grey-yellow sand gravel wedge. Materials in the wedge are similar to those that form U8

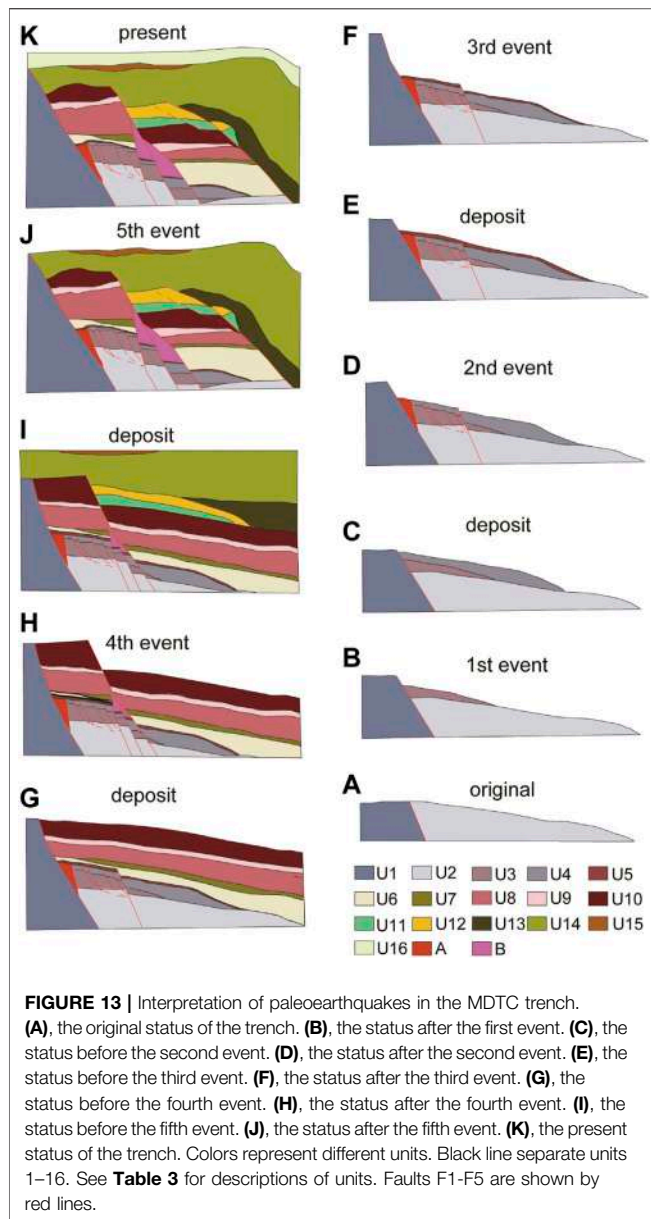


lower limit time of the third paleoearthquake were around 251–697 AD. Unit U6, U7, U8, U9, and U10 were deposited after the third event, which is compatible with the OLS age (ca. 1.1 ± 0.1 ka) from U9 (**Figure 13G**). The fourth paleoearthquake event happened after the deposition of unit U10 and involved faults F1, F2, F3, and F4 (**Figure 12H**). Unit U11, U12, U13, U14, and U15 were deposited after the fourth event (**Figure 12I**). Samples from U9 (OSL) and U11 (radiocarbon) bracket the time of the fourth paleoearthquake between 1.1 ka and 1265–1366 AD. The fifth paleoearthquake event took place after the deposition of U15 and involved faults F1, F2, F3, F4, and F5 (**Figure 13J**). Unit U16 was deposited after the fifth event. Due to the lack of well-documented history in this region prior to the Ming and Qing Dynasties (1368–1911 AD), no destructive paleoearthquake in the Historical Strong Earthquake Catalog of China (China

Earthquake Administration, 1995) can be associated. Thus, the fifth event may be related to the 1941 M7 historical earthquake.

For the Hupocun trench HPTC, based on the age of U4 there (^{14}C age 520 ± 40 B.P.) (**Figure 6; Table 2**) and the only record of the strong earthquake (1941 M7) in this areas, we attribute the youngest deformation also to the 1941 M7 event.

Our results also place important constraints on the occurrence times of paleoearthquake events on the Nantinghe fault. One paleoearthquake event is recorded in the trench at Hupocun (**Figure 7**). The event occurred after 1315–1448 AD. Based on our work on the Mengding trench, there are five paleoearthquake events in the last 5,000 a, and four paleoearthquake events between 3929 BC and 1366 AD. Since the distance between Hupocun and Mengding is only 60 km and there is no obvious step or discontinuity in the fault trace over this distance, we suggest that the faults in Hupocun and Mengding are likely to have similar rupture histories. If this is true, then we can analyze earthquake occurrence times using paleoearthquakes from the two trench sites and historical earthquakes. One documented historical earthquake occurred in AD 1941. Thus, the youngest events in the MDTC trench may be equivalent to that in the HPTC trench, which postdated 1315–1448 AD. Since the faults almost ruptured the surface in both trenches, we speculate that the youngest event in the two trenches corresponds to the 1941 M7 event. Five paleoearthquakes occurred during the last 5,000 years (yielding an approximate recurrence interval of $\sim 1,000$ a). Given slip rates of 3.5–5.4 mm/a, the millennial displacement should be 3.5–5.4 m. The magnitude of the earthquake generated by the latest event is an important parameter in evaluating earthquake hazards in a region. It depends on the coseismic rupture length, displacement, and history records. Associate with the history records, the last earthquake along the Nantinghe fault was estimated to be an M7 event (China Earthquake



Administration, 1995). According to the empirical relationships between the magnitude, surface rupture length, and displacement on strike-slip faults around the Tibetan Plateau (Deng et al., 1992; Ye et al., 1996), a surface rupture length of 60 km and a maximum displacement of 3.5–5.4 m suggest a magnitude ranging from Ms 7.4 to 7.6. Additionally, the empirical relationship of (Wells and Coppersmith, 1994) points to Ms 7.2–7.7. Moreover, the 1988 Ms 7.6 and 7.2 Lancang-Gengma earthquakes in the nearby region produced two surface ruptures of ~50 and ~14 km, respectively (Yu et al., 1991). Therefore, the magnitude of the latest earthquake is likely to be in the range of Ms 7.0–7.7.

5.2 Slip Rate Estimate

The methodologies for estimating slip rates are often debated due to the various proxies and time scales. For the large strike-slip faults

within the Indo-Asian collision zone and other regions around the world (Van Der Woerd et al., 2002; Tong et al., 2014), geodetically-determined slip rates are generally different from those reported using reconstructions of offset landforms, and it is unclear if this discrepancy reflects true secular variation in slip history, systematic errors in interpretation, or both. Even slip rates derived from different reconstructions of offset landforms along the same fault can be different. A major potential source of uncertainty when assessing slip rates from offset fill terraces is the underlying geomorphic model of terrace formation and abandonment that must be assumed (Mériaux et al., 2004; Cowgill, 2007). One evolutionary model suggests that the erosion of the flood plain and its banks continues until the river begins to incise and new terraces form (Van Der Woerd et al., 2002). According to this model, the offset of a terrace riser only begins to accumulate once the lower terrace is abandoned. An alternative model allows for differences in erosion patterns between the sides of the river, in which case riser offsets may begin to accumulate after the upper terrace is abandoned. For example, at a site on the Altyn Tagh fault, a study used the offset of a terrace riser and the abandonment age of the lower terrace to obtain a slip rate of 25.9 ± 1.8 mm/a (Mériaux et al., 2004). Using the abandonment age of the higher terrace, a competing study yielded a slip rate of 9–10 mm/a (Cowgill, 2007). Thus, the choice of a conceptual model is the key to slip rate evaluation but remains controversial. Using the alluvial fan to calculate the slip rate, however, can avoid this issue.

At the Zhujiacun, Daxueshan, and Hupocun sites, we used the ages of the higher terraces to determine the slip rate of the Nantinghe fault. At the Junsai and Mengjian sites, slip rates are calculated from the offset alluvial fan. The slip rates from the two approaches agree with each other (between 3.5 and 5.4 mm/a), indicating that the higher terrace approach is more suitable.

We calculated the mean slip rate using the maximum likelihood regression *via* IsoplotR (Vermeesch, 2018) based on the modified error-weighted least-squares algorithm of York et al. (2004) (Figure 14A). Due to the lack of slip rates, previous studies have debated whether different segments of the Nantinghe fault have different activities. From the depth of the basin, which decreases northeastward along the fault, and the occurrence of the 1941 M7 earthquake, it seemed that the northeastern segment may be less active than the southwestern segment (Zhu et al., 1994; Wang et al., 2006) (Figure 2). Nevertheless, our results of 4.0 ± 0.6 (2σ) mm/a reveal a similar Quaternary activity along the Nantinghe fault (Figures 2, 14A).

5.3 Tectonic Implications

Our results provide clear evidence for the late Quaternary oblique sinistral-thrust activity on the Nantinghe fault; its millennial strike-slip rate is ca. 3.9 ± 0.6 (2σ) mm/a, in agreement with the GPS measurements (4 mm/a) (Wang, 2009). The Nantinghe fault also shows a 0.3–0.6 mm/a slip rate in the dip-slip orientation. These slip rates help us constrain the applicability of the competing tectonic models for the Tibetan Plateau. The block models require large strike-

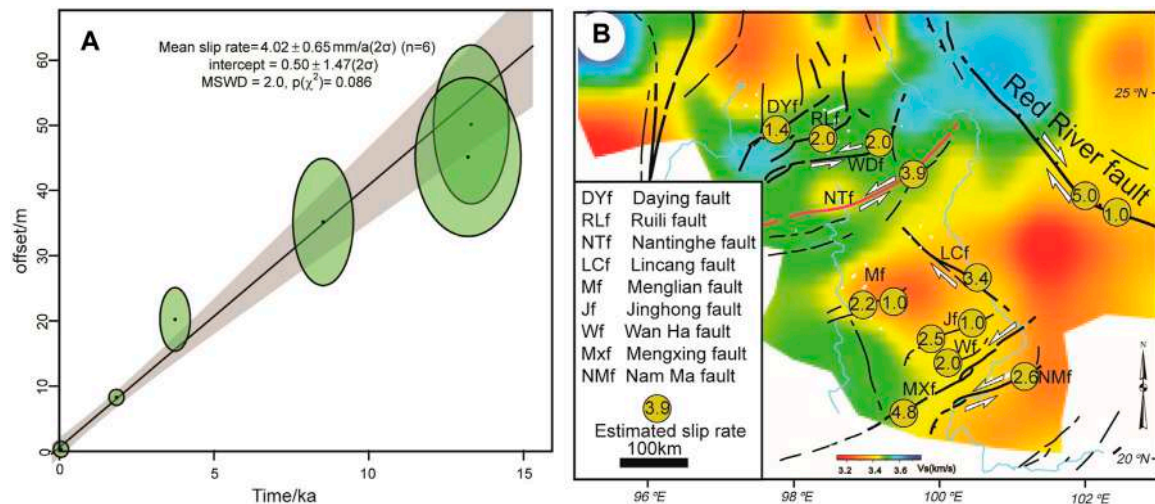


FIGURE 14 | (A) Mean left-lateral slip rate of the Nantinghe fault. The maximum likelihood regression is calculated via IsoplotR (Vermeesch, 2018) based on the modified error-weighted least-squares algorithm of York et al. (2004). **(B)** Integrated map of estimated slip rates (mm/yr) in the study area. Base map is the shear-wave velocity model at 21 km from (Bao et al., 2015). Sources for slip rates on the faults, from north to south, include: Daying fault (Chang et al., 2011); Ruili fault (Huang et al., 2010; Wang et al., 2014); Wanding fault (Chang et al., 2012); Nantinghe fault, this study; Lincang fault (Wang et al., 2014); Menglian fault (Wang et al., 2014; Liu et al., 2019); Jinghong fault (Shi et al., 2018c); Wan Ha fault (Wang et al., 2014); Mengxing fault (Lacassin et al., 1998; Wang et al., 2014); Nam Ma fault (Lacassin et al., 1998; Wang et al., 2014); and Red River fault (Replumaz et al., 2001; Shi et al., 2018a).

TABLE 5 | Summary of Quaternary slip rates along the major faults in Southeastern Tibetan Plateau.

Fault name	Sense-of-motion	Strike-slip rate (mm/a)	Dating method ^a	References
Daying	sinistral	1.4	TL	Chang et al. (2011)
Ruili	sinistral	2.0	OSL/TL	Huang et al. (2010)
Wanding	sinistral	2.0	TL	Chang et al. (2012)
Nantinghe	sinistral	4.0	OSL/ ¹⁴ C	This study
Lincang	dextral	3.4	estimation	Wang et al. (2014)
Menglian	sinistral	2.2	OSL	Liu et al. (2019)
Jinghong	sinistral	1.0	estimation	Wang et al. (2014)
		2.5	¹⁴ C	Shi et al. (2018c)
		1.0	estimation	Wang et al. (2014)
Wan Ha	sinistral	2.0	estimation	Wang et al. (2014)
Mengxing	sinistral	4.8	estimation	Lacassin et al. (1998), Wang et al. (2014)
Nam Ma	sinistral	2.6	estimation	Lacassin et al. (1998), Wang et al. (2014)
Red River	dextral	5.0	Estimation	Replumaz et al. (2001)
		1.0	¹⁴ C	Shi et al. (2018a)

^aTL, thermoluminescence; OSL, optically stimulated luminescence.

slip rates at the boundaries between different tectonic blocks, but small strike-slip rates within tectonic blocks (Leloup et al., 1995; Tapponnier et al., 2001). On the contrary, the continuum models (England and Molnar, 2005; Copley, 2008) do not necessarily predict such large slip rates at boundaries between different tectonic blocks, but larger strike-slip rates within tectonic blocks.

Moderate to high Quaternary slip rates have been found on the major strike-slip faults within the Indochina Block, including the Daying, Ruili, Wanding, Nantinghe, Lincang, Jinghong, Wan Ha, Mengxing, and Nan Ma faults (Table 5). Their strike-slip rates range from 1.0 to 5.0 mm/a. By contrast, the Red River fault, albeit a plate boundary fault in geologic history, shows a comparable dextral-slip

rate of up to 1–5 mm/a over decadal, millennial, and million-year timescales (Allen et al., 1984; Weldon et al., 1994; Copley, 2008; Shi et al., 2018a; Replumaz et al., 2001; Schoenbohm et al., 2006) (Figure 13B). Moreover, as pointed out by Shi et al. (2018b), the distinct arcuate geometry of the sinistral strike-slip faults and the curved GPS velocity field are difficult to reconcile with the block models but can be readily explained in the continuum models.

An asymmetric, tongue-shaped GPS velocity pattern is found across the Indochina Block from Sunda to the Eastern Himalayan Syntaxis (Shi et al., 2018b). In turn, Shi et al. (2018b) proposed that asthenospheric flow exerts basal traction to the overlying lithosphere and induces the faulting in the upper crust. The Nantinghe fault, with its 3.5–5.4 mm/a millennial slip rate, is

located in a swath of the largest slip rate on the velocity profile by Shi et al. (2018b). Interestingly, this swath, containing the Daying, Ruili, Wanding, and Nantinghe faults on the surface, is located between two localized, NE-SW-trending channels of low-velocity anomalies at the mid-crustal depths from a high-resolution shear-wave seismic tomography model of this region (Bao et al., 2015), (Figure 13B). All these results indicate that the Red River fault does not dominate the present-day kinematics of the Indochina Block, and hence the continuum models could play a more dominant role in the Quaternary tectonics of this region. Pre-existing crustal weakness (such as the Red River and Nantinghe faults) and crustal heterogeneities promote the lateral variations in the lower crustal strength, which shaped the upper-crustal deformation and topography (Tan et al., 2019; Penney and Copley, 2021).

6 CONCLUSION

We have documented the Quaternary activity of the Nantinghe fault in the Southeastern Tibetan Plateau region, using a combination of field observation, satellite and aerial imagery interpretation, trenching, and ^{14}C and OSL geochronology. The Nantinghe fault has been active in the latest Quaternary with an oblique sinistral-thrust sense of slip. Detailed geochronology and fault kinematic measurements on the sinistral displacement reveal that the sinistral strike-slip rates along the fault are ca. 4.0 ± 0.6 (2σ) mm/a over the past 13 ka. Apparent throw rates are typically less than 1 mm/a. Our trench investigations demonstrate at least four 4) Holocene earthquakes along the Nantinghe fault, suggesting an averaged recurrence interval of $\sim 1,000$ a. The newly-obtained slip rates on the Nantinghe fault, integrated with previous estimates on other faults in this region, provide direct evidence that favors the continuum models over the rigid block ones, which advances our knowledge on the deformation mechanism in the India-Asia continental collision.

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DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/**Supplementary Material**, further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

Conceptualization was performed by FS and HH, the fieldwork was done by FS, ZW, and HS, the writing and preparation of the original draft were conducted by FS, the review and editing of the final manuscript were performed by FS and YL. All authors have read and agreed to the published version of the manuscript.

FUNDING

This work was jointly supported by the National Key Research and Development Program of China (2021YFC3000601), the National Natural Science Foundation of China (41702221) and the China Active Fault Survey Project (201102001).

ACKNOWLEDGMENTS

We thank the two Reviewers Drs. Hu Wang and Bo Zhang for providing prompt and constructive reviews. YL acknowledges the Governor's University Research Initiative (GURI) fund from the State of Texas and the University of Houston.

SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2021.818225/full#supplementary-material>

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Tectonic Deformation of the Western Qilian Shan in Response to the North–South Crustal Shortening and Sinistral Strike-Slip of the Altyn Tagh Fault Inferred From Geomorphologic Data

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OPEN ACCESS

Edited by:

Wenjun Zheng,
Sun Yat-sen University, China

Reviewed by:

Daoyang Yuan,
Lanzhou University, China
Manoj Kumar Jaiswal,
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 04 November 2021

Accepted: 18 January 2022

Published: 10 February 2022

Citation:

Chen Z, Xu W, Liu R, Li A and
Koronovsky NV (2022) Tectonic
Deformation of the Western Qilian
Shan in Response to the North–South
Crustal Shortening and Sinistral Strike-
Slip of the Altyn Tagh Fault Inferred
From Geomorphologic Data.
Front. Earth Sci. 10:808935.
doi: 10.3389/feart.2022.808935

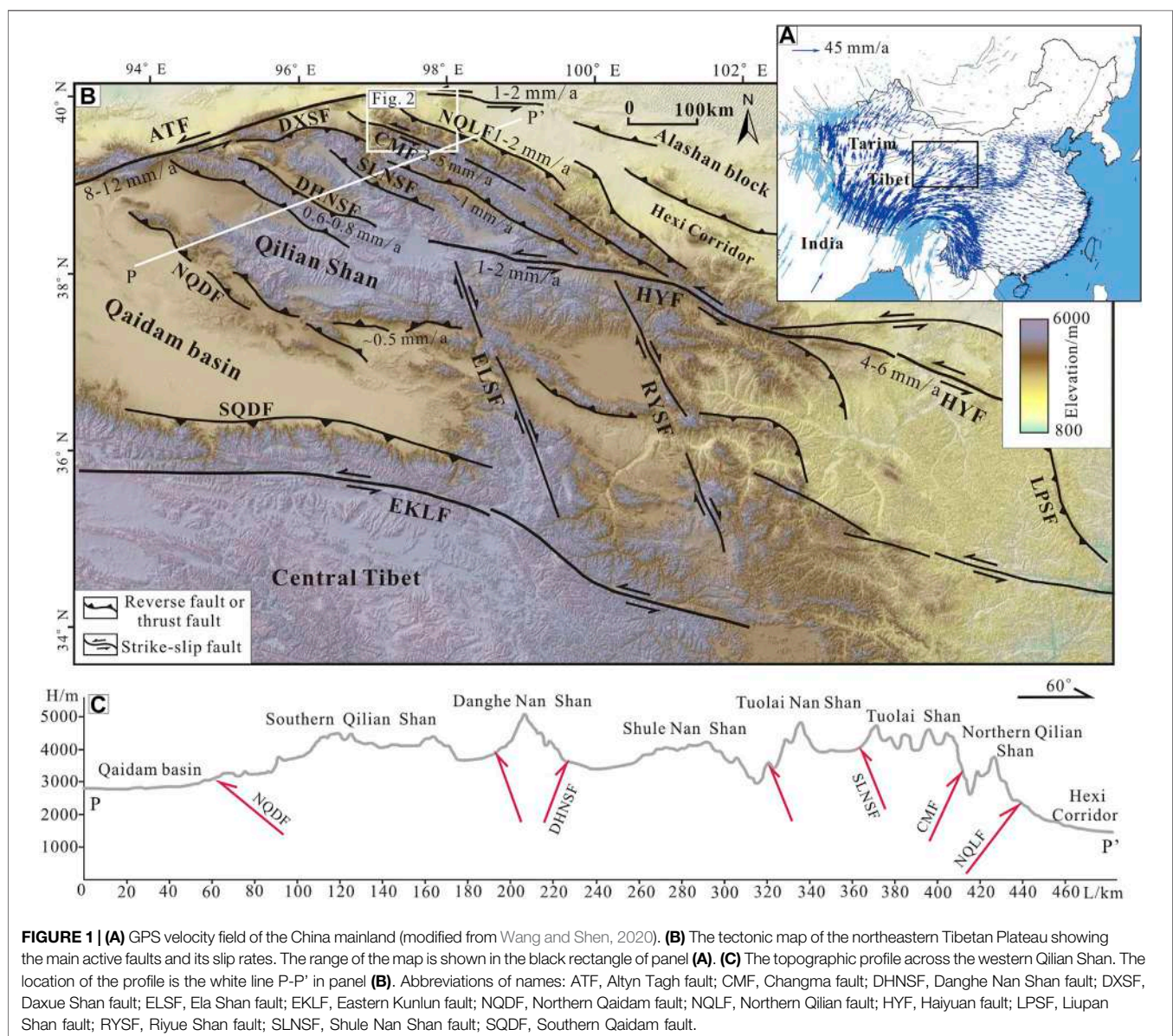
The tectonic deformation of a series of NWW-trending compressional band-shaped mountains and fault zones in the Qilian Shan, northeastern Tibet, plays an important role in absorbing the north–south crustal shortening of the Tibetan Plateau and accommodating the sinistral strike-slip of the Altyn Tagh fault. However, the deformation pattern of these mountains and fault zones in response to crustal shortening and sinistral strike-slip remains to be investigated. In this study, we present the slope map and swath profiles of the western Qilian Shan extracted from high-resolution digital elevation model (DEM) data and the fluvial geomorphologic characteristics of the Baiyang river in the northwestern Qilian Shan. Our data indicate that the crustal uplift of the northern Qilian Shan is stronger than that of the central and southern ranges of the Qilian Shan because of the high-terrain relief and stronger erosion. In addition, the deformed terraces of the Baiyang river documented the activity of NWW-trending faults in the western Qilian Shan. Based on the longitudinal profiles and ages of the deformed river terraces, the vertical slip rates of the Changma, Yumen, and Bainan faults in the western Qilian Shan since ~60 ka were constrained to be 0.31 ± 0.06 , 0.33 ± 0.02 , and 0.24 ± 0.02 mm/a, respectively. Based on the comparison with previous studies, we noticed that the rate of the vertical slip since the Late Quaternary of the northern Qilian fault of 1.5–2 mm/a is significantly higher than that of the Changma fault of 0.3–0.6 mm/a, which matches the terrain relief in the Qilian Shan and demonstrates that the northern Qilian Shan has been rising faster than its southern ranges. From the slip rates of the NWW-trending faults in the western Qilian Shan, we suggest that the crustal shortening is widely distributed in the NWW-trending compressional orogenic belts and fault zones across the Qilian Shan, among which the northern Qilian Shan and its frontal thrust system absorb the greatest shortening amount, whereas strike-slip faults within the Qilian Shan accommodate most of the sinistral strike-slip of the Altyn Tagh fault.

Keywords: western Qilian Shan, geomorphologic characteristics, river terrace, slip rate, crustal shortening, sinistral strike-slip

1 INTRODUCTION

The collision and post-collisional convergence between India and Eurasia have caused crustal shortening and thickening and led to the generation of the magnificent mountain ranges of the Himalayas and Tibetan Plateau (Molnar and Tapponnier, 1975; England and Houseman, 1986; Harrison et al., 1992; Yin and Harrison, 2000). Since the onset of the Indo-Asian collision at 65–50 Ma, the Himalayan–Tibetan orogen has absorbed at least 1,400 km of crustal shortening (Yin and Harrison, 2000; Ding et al., 2017), which have been transformed into the rapid uplift of mountains (e.g., Harrison et al., 1992; Wang et al., 2014), thrusting and strike-slip of faults (e.g., Molnar and Tapponnier, 1975; Zhang et al., 2004), lateral extrusion of blocks (e.g., Tapponnier and Molnar, 1976; Hodges et al., 2001), and shrinking and extinction of basins (e.g., Wang et al., 2014;

Ding et al., 2017). The northward expansion of the Tibetan Plateau led to the growth of a series of compressional orogenic belts in central Asia. Among them, the Qilian Shan (Shan means “mountains” in Chinese), an orogenic belt in the northeastern margin of the Tibetan Plateau, is a typical example (Wang et al., 2016). The Qilian Shan is located in the convergence zone of the Tibetan, Tarim, and Alshan blocks, that is, the transition region from the plateau ranges to basins in the north, and plays an important role in accommodating the tectonic deformation in the northern plateau (Zuza et al., 2016). Recent results based on the stratigraphy and apatite fission track chronology suggest that the Qilian Shan began to uplift during the Late Miocene (14–10 Ma) (Wang et al., 2016; Zheng et al., 2017), much later than the Indo-Asian collision (Wang et al., 2014; Ding et al., 2017). Although the Qilian Shan with an average altitude of >4,000 m was formed relatively later, its area exceeds 2,000 km². Its enormous size



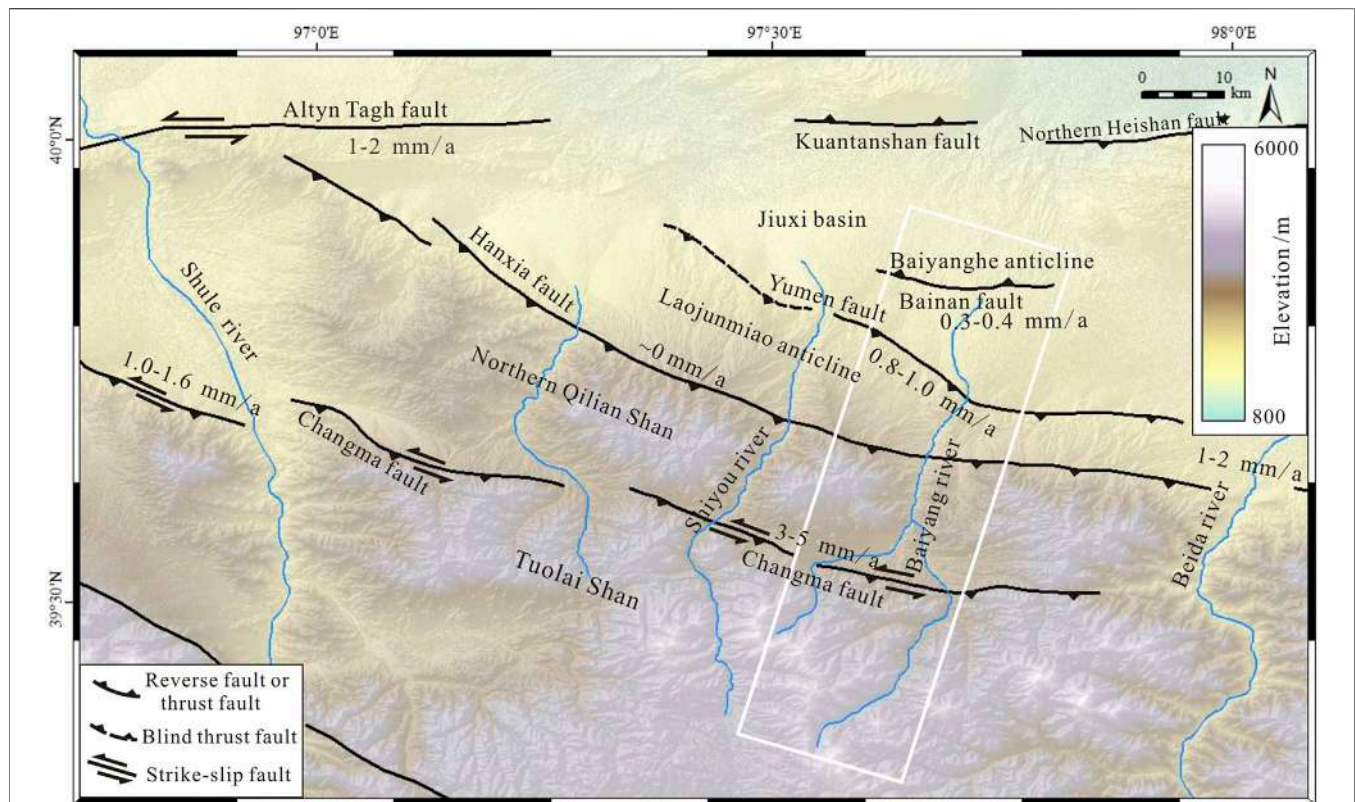


FIGURE 2 | The tectonic map of the northwestern Qilian Shan showing the slip rates of the active faults. The range of the map is shown in the white rectangle of **Figure 1B**. The white frame in this map shows the Baiyang river study area covering the range of **Figures 4A, 8A**.

makes it a non-negligible part of studies of the tectonic deformation of the plateau (**Figure 1B**).

The Qilian Shan absorbs the north–northeast (NNE) crustal shortening of the Tibetan Plateau and accommodates the sinistral strike-slip of the Altyn Tagh fault. Based on the Global Positioning System (GPS) velocity field, the present-day shortening rate between India and Eurasia is 40–50 mm/a (millimeters per year) of which 15–20 and 20–30 mm/a are absorbed by the Himalayan tectonic belt and the deformation within the Tibetan Plateau and surrounding mountains, respectively (**Figure 1A**) (Zhang et al., 2004; Wang and Shen, 2020). The crustal shortening rate across the Qilian Shan is 5.5 ± 1.5 mm/a, accounting for 15%–17% of the total convergence rate between India and Eurasia, which is mainly absorbed by the uplift of mountains and deformation of five large-scale folded fault belts (Zhang et al., 2004). The Altyn Tagh fault on the west side of the Qilian Shan is a continental sinistral strike-slip fault in the northern plateau. The GPS measurements revealed that the sinistral slip rate of the Altyn Tagh fault decreases from 8 to 12 mm/a in the middle section to 1–2 mm/a in the easternmost section after crossing the Qilian Shan (**Figure 1B**) (Zhang et al., 2004; Liu J. et al., 2020; Wang and Shen, 2020), indicating that the Qilian Shan accommodates ~ 8 mm/a of the sinistral strike-slip of the Altyn Tagh fault. However, the focus of previous studies was mainly placed on the tectonic deformation of boundary fault belts in the Qilian Shan, especially the northern Qilian Shan and its

frontal thrust (e.g., Hetzel et al., 2002; Hetzel et al., 2004; Zheng et al., 2013a; Hu et al., 2015; Liu R. et al., 2017; Liu X.-W. et al., 2017; Yang et al., 2018a; Yang et al., 2018b; Liu X.-w. et al., 2019; Liu X. et al., 2019; Hetzel et al., 2019; Yang et al., 2020; Hu et al., 2021; Li et al., 2021). The knowledge of the sinistral strike-slip and the compressional deformation within the Qilian Shan is insufficient, which restricts the overall understanding of the tectonic deformation in the northern plateau.

Spatial analysis and geomorphological quantitative studies are powerful tools for studying tectonic deformation within the Qilian Shan. Based on high-resolution digital elevation model (DEM) data and geomorphic parameters, such as the slope and relief extracted in this study, the spatial variation of the terrain relief can be determined. A series of north–south rivers flow perpendicularly through the band-shaped mountains and fault zones in the western Qilian Shan and have inevitably been affected by tectonic activity (**Figure 2**). The geomorphologic characteristics of these rivers reflect the tectonic deformation of the mountain ranges and tectonic belts (Chen and Koronovskii, 2020). It has been proven that the slip rates of the faults and fold belts can be constrained by studying deformed landforms along the active Qilian Shan front and in its foreland (e.g., Hetzel et al., 2002; Hetzel et al., 2004; Hu et al., 2015; Liu X.-W. et al., 2017; Yang et al., 2018a; Yang et al., 2018b; Liu X.-w. et al., 2019; Liu X. et al., 2019; Hetzel et al., 2019; Chen and Koronovskii, 2020; Liu R. et al., 2020). In this study, field surveys

and topographic measurement were applied to analyze the fluvial geomorphologic characteristics of the western Qilian Shan. By measuring the displacements and ages of deformed river terraces, we obtained the long-term slip rates of the main faults in the study area. All above-mentioned results were used to explain the tectonic deformation, strain distribution, and deformation pattern in the western Qilian Shan.

2 TECTONIC SETTING OF THE STUDY AREA

The Qilian Shan is the northeastern leading edge of the Tibetan Plateau, resulting from the northward expansion of the plateau. The overall tectonic trend of the Qilian Shan is approximately N65°W, almost perpendicular to the direction of the relative movement between the Indian and Eurasian plates (Zhang et al., 2004). A series of band-shaped mountains, compressional basins, and fault and fold zones in the Qilian Shan are distributed in the NWW direction, demonstrating that the tectonic deformation of the Qilian Shan is characterized by compression (**Figure 1B**). Based on GPS data, the crustal shortening rate between the Qaidam Basin and Alshan Block is 5.5 ± 1.5 mm/a, which is largely absorbed by the NWW-trending mountains and fault zones in the Qilian Shan (Zhang et al., 2004).

Our study area is the western Qilian Shan, which is bounded by the Altyn Tagh fault to the west and the Haiyuan fault to the east. From south to north, five large-scale NWW-trending compressional fault zones can be observed, that, are the northern Qaidam, Danghe Nan Shan, Shule Nan Shan, Changma, and northern Qilian faults (**Figures 1B,C**). The results of previous studies showed that the tectonic deformation of the NWW-trending mountains and fault zones in the western Qilian Shan plays a significant role in absorbing the NNE crustal shortening of the plateau and accommodating a large amount of sinistral strike-slip components of the Altyn Tagh and Haiyuan faults (Zhang et al., 2007; Zheng et al., 2009; Zheng et al., 2013b; Liu J. et al., 2020). These mountains and fault zones obliquely intersect the Altyn Tagh and Haiyuan faults at low angles (**Figure 1B**). Based on GPS data, the sinistral strike-slip rate in the middle section of the Altyn Tagh fault is 8–12 mm/a and decreases to 1–2 mm/a in the eastern section (**Figure 1B**) (Zhang et al., 2007; Zheng et al., 2009; Zheng et al., 2013b; Liu J. et al., 2020). The geological results suggest that the sinistral strike-slip rate of the Altyn Tagh fault in the Holocene was 17.5 ± 2.2 mm/a in the midwest, 11 ± 3.5 mm/a in the middle, and 2.2 ± 0.2 mm/a in the east (Xu et al., 2005). The sinistral strike-slip rate of the Altyn Tagh fault decreased from west to east and considerable sinistral strike-slip was absorbed by the tectonic deformation in the western Qilian Shan. Contrary to the decreasing trend of the Altyn Tagh fault, the slip rate of the Haiyuan fault gradually increased from west to east. Based on GPS data, the sinistral strike-slip rate of the Haiyuan fault is only 1–2 mm/a in the west adjacent to Danghe Nan Shan and 4–5 mm/a in the middle section near the Tuolai Nan Shan, but it reaches 4–6 mm/a at the northern margin of the Qilian Shan (**Figure 1B**) (Zheng et al., 2009; Zheng et al., 2013b), which

suggests that ~ 4 mm/a of the sinistral strike-slip in the western Qilian Shan has been transferred into the Haiyuan fault. The variation in the slip rates of the Altyn Tagh and Haiyuan faults indicates that the western Qilian Shan is an important transition zone for the accommodation of the sinistral strike-slip of these two large-scale faults, where a large proportion of sinistral strike-slip components of the Altyn Tagh fault has been absorbed and partially transferred into the Haiyuan fault by the NWW-trending mountain building and faulting within the western Qilian Shan.

The northwestern Qilian Shan represents the junction zone of the Qilian, Tarim, and Alashan blocks where tectonic deformation is characterized by crustal shortening and surface uplift (Chen and Koronovskii, 2020). A series of north–south rivers flow through the NWW-trending band-shaped mountains and active fault zones in the western Qilian Shan. These rivers provide typical geomorphologic markers for studying the tectonic deformation in this area. In this study, the Baiyang river, one of the largest north–south rivers in the northwestern Qilian Shan, was selected as the research object (**Figure 2**). Four NWW-trending fold-and-thrust zones are developed from south to north in the Baiyang river catchment, that is, the Changma, Hanxia, and Yumen faults and its anticline (Laojunmiao anticline) and the Bainan fault and its anticline (Baiyanghe anticline) (**Figure 2**) (Liu R. et al., 2017; Chen and Koronovskii, 2020). These faults are subject to compressional deformation in the NNE direction and the sinistral strike-slip of the Altyn Tagh fault. Their trend is consistent with the principal stress field of the Qilian Shan. Based on topographic surveys and ^{14}C dating, Zheng et al. (2013a) determined the left-lateral and thrust slip rates of 1.17 ± 0.07 and 0.25 mm/a of the Changma fault. However, Luo et al. (2013) constrained the left-lateral and thrust slip rates of the Changma fault to be 1–3 mm/a and 0.3–0.6 mm/a; Du et al. (2020) estimated the left-lateral slip rates of the Changma fault to be 3–5 mm/a. Previous studies on the constraint of slip rates of the Changma fault differ greatly. Using similar methods, Liu X.-w. et al. (2019) derived a vertical slip rate of 0.9 ± 0.1 mm/a and a shortening rate of 1.9 ± 0.5 mm/a for the Yumen blind thrust in the Qilian Shan foreland and Hetzel et al. (2002, 2019) obtained a vertical slip rate of 0.35 ± 0.05 mm/a for the Bainan fault and a horizontal shortening rate of 2.0 ± 0.3 mm/a for the northern Qilian Shan frontal thrust. Most of studies show that the shortening rates of the northern Qilian Shan fault system vary in the range of 1–2 mm/a (e.g., Hetzel et al., 2004; Zheng et al., 2013a; Hu et al., 2015; Liu X.-W. et al., 2017; Yang et al., 2018a; Yang et al., 2018b; Liu X.-w. et al., 2019; Liu X. et al., 2019; Hetzel et al., 2019).

3 MATERIALS AND METHODS

3.1 Geomorphologic Spatial Analysis Based on the Digital Elevation Model

Geomorphologic parameters, such as the slope and slope direction, extracted from high-resolution DEM data can provide qualitative and quantitative constraints for studying the tectonic deformation in orogenic belts (e.g., Chu, 2015). In

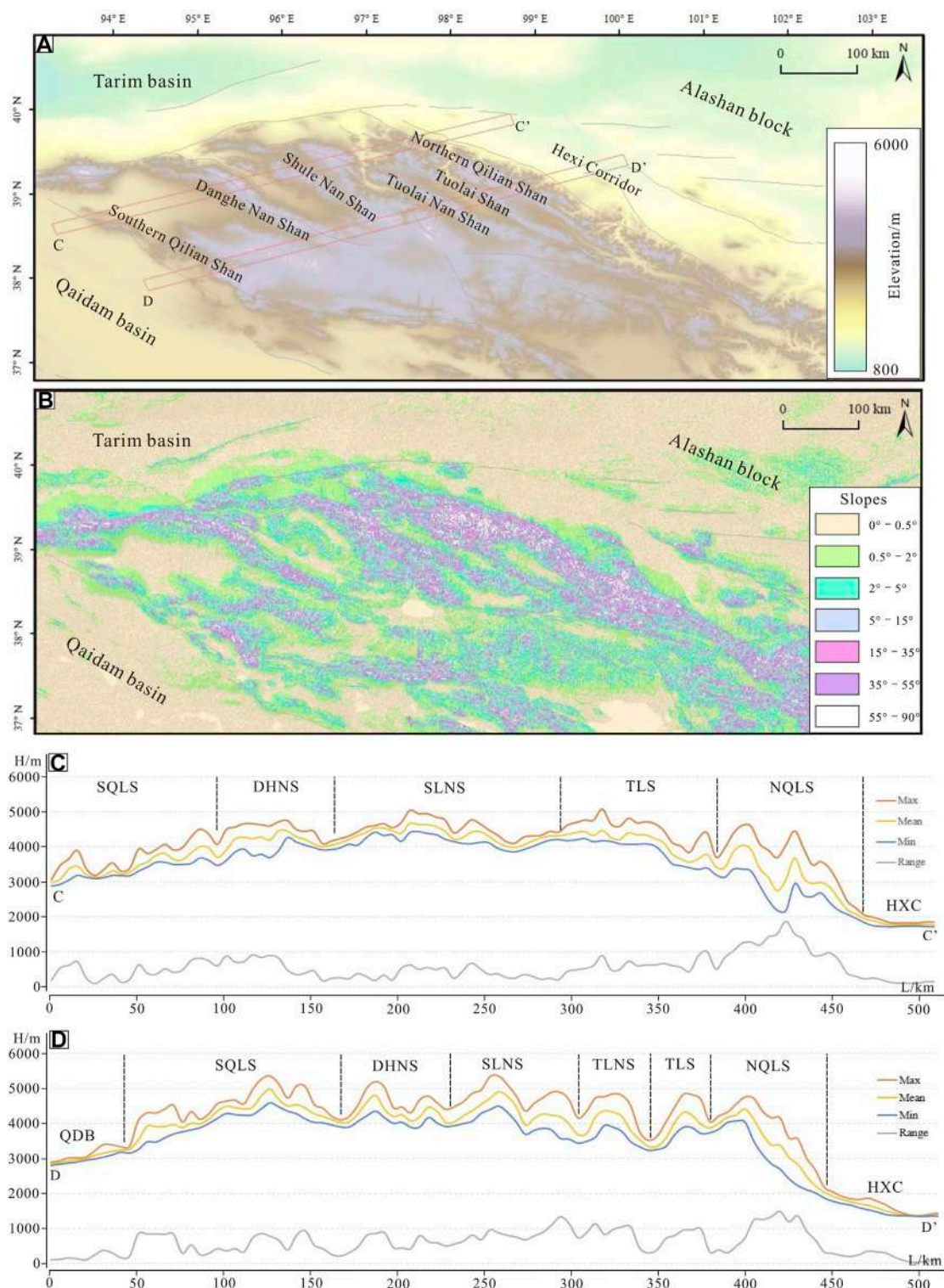


FIGURE 3 | (A) The DEM image of the Qilian Shan. **(B)** The slope map of the Qilian Shan. The gray lines represent active faults in the Qilian Shan. **(C,D)** are swath profiles on the western Qilian Shan. The locations of the swath profiles are shown in the red rectangles of panel **(A)**. The Max, Mean, Min, and Range curves represent the maximum, average, minimum, and standard deviation (difference between the maximum and minimum) of elevations in the swath ranges, respectively. Abbreviations of names: SQLS, Southern Qilian Shan; DHNS, Danghe Nan Shan; SLNS, Shule Nan Shan; TLNS, Tuolai Nan Shan; TLS, Tuolai Shan; NQLS, Northern Qilian Shan; QDB, Qaidam basin; HXC, Hexi Corridor.

this study, we obtained slope parameters based on DEM data with a precision of 30 m and developed the slope map of the Qilian Shan using ArcGIS software (**Figure 3B**), which accurately reflects the spatial variation of the terrain slope in the Qilian Shan. Before analyzing the spatial variation of the terrain slope, a reasonable slope classification system must be established. In this study, based on the slope classification developed by the Geomorphology Survey and Mapping Committee of the International Geographical Union, we classified the slope grades as follows: 0° – 0.5° , 0.5° – 2° , 2° – 5° , 5° – 15° , 15° – 35° , 35° – 55° , and 55° – 90° , representing the plain, gentle slope, moderate slope, steep slope, very steep slope, cliff, and vertical wall, respectively (**Figure 3B**) (Chu, 2015; Cheng et al., 2019). In the slope map, different slopes were expressed using different colors to present the spatial variation of the slopes.

Topographic swath profiles have an advantage over conventional topographic linear profiles with respect to the evaluation of the effects of vertical surface movements in the investigation of fluvially sculpted topography. Based on topographic swath profiles derived from DEM data, a simple linear profile can be expanded into a rectangular swath profile and the elevation in a given swath area can be calculated as statistical parameters (maximum, mean, minimum, and standard deviation), which reflect the terrain relief in the study area and compensate for the deficiencies of single linear topographic profiles (Chu, 2015). In this study, we used the 3D analyst tools of ArcGIS to extract two swath profiles across the western Qilian Shan, perpendicular to the strike of the mountains and faults. Each profile has a width of 5 km and length exceeding 500 km (**Figure 3A**). Then, we divided each profile into 100 units, each of which is a square of $5 \times 5 \text{ km}^2$. Based on the elevation from DEM, we derived the maximum (Max), average (Mean), minimum (Min), and Range values (the difference between Max and Min) of elevation for each unit and plotted four curves in the swath ranges, respectively. (**Figures 3C,D**).

3.2 Topographic Surveys of the Fluvial Geomorphology

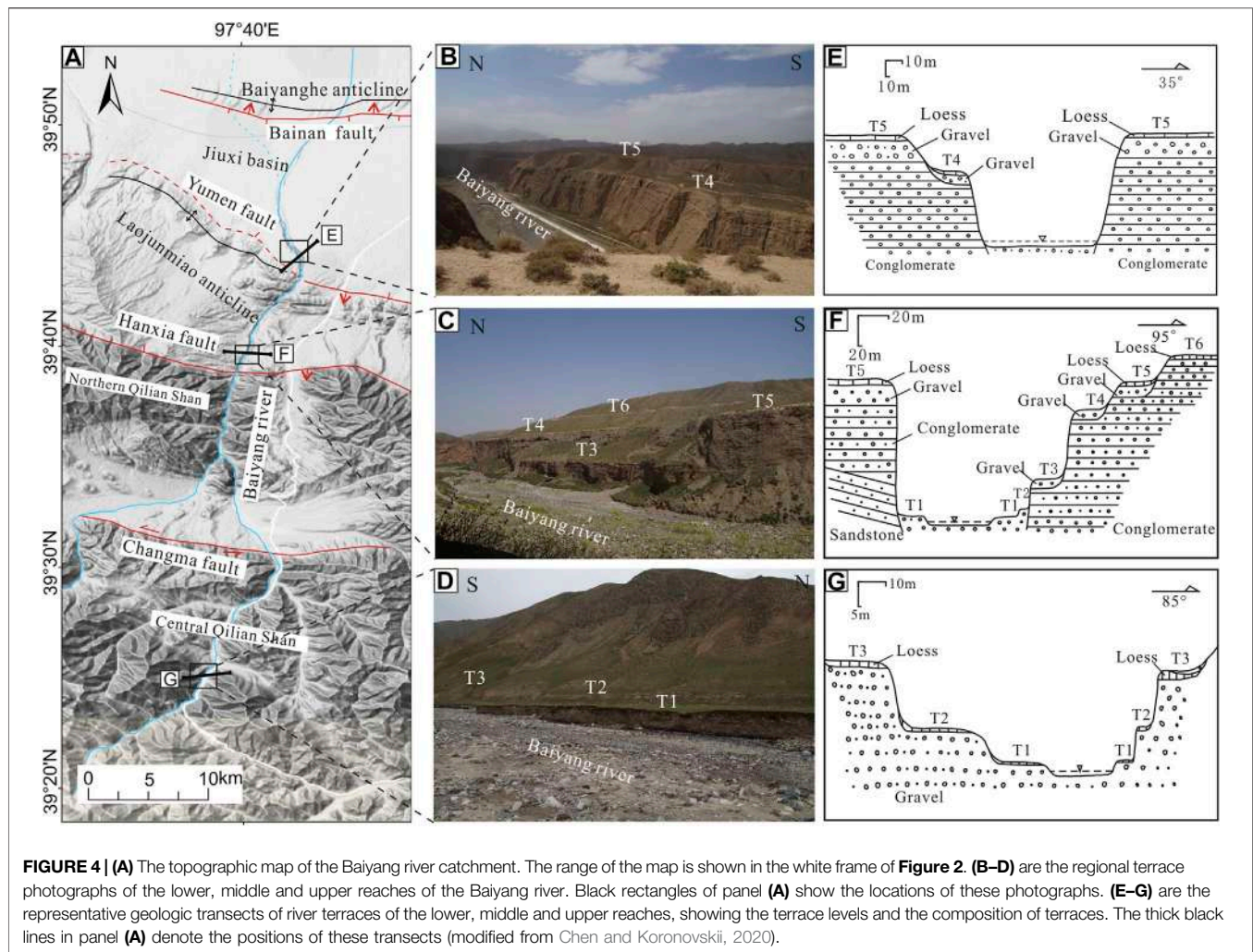
The fluvially sculpted topography on the Earth's surface is the result of interactions among the tectonics, erosion, and climate (Lavé and Avouac, 2001). To analyze the coupling between the fluvial topography and tectonic movements, it is crucial to accurately identify geomorphic markers that can reflect tectonic deformation. The results of previous studies suggested that climate change is the driving force of deep incision and the formation of terraces of rivers in the northern Qilian Shan because the regional fluvial incision rates are significantly higher than the vertical slip rates of the northern Qilian thrust (e.g., Hetzel et al., 2006; Yang et al., 2020). Despite the inconsistencies of the incision and vertical slip rates, numerous fluvial geomorphic markers can be used to reflect the tectonic deformation in orogenic belts such as hanging valleys, bedrock gorges, steep V-shaped channels, fluvial strath terraces, etc. (Chen and Koronovskii, 2020). Through a field investigation, we comprehensively analyzed the geomorphic

features of the reaches of the Baiyang river catchment, including the valley contour, riverbed characteristics, heights of the valleys, types, and composition of terraces, etc. Based on the distribution of mountain ranges and faults and fluvial geomorphologic features, the Baiyang river can be divided into upper, middle, and lower reaches. The sections south of the Changma fault, from the Changma fault to the Yumen fault, and north of the Yumen fault were classified as the upper, middle, and lower reaches, respectively (**Figure 4A**).

3.3 Measurement of the Displacement and Ages of Deformed Terraces

River terraces have been deformed due to the continuous activity along faults in the northern Qilian Shan. In turn, the displacement of the river terraces reflects the movement of these faults. By measuring the displacement of the terraces and determining their ages, the rates of the long-term slip of these faults since the Late Quaternary can be obtained (e.g., Liu R. et al., 2017; Yang et al., 2018a; Yang et al., 2018b; Liu X.-w. et al., 2019; Liu X. et al., 2019; Chen and Koronovskii, 2020). The key to deriving the long-term slip rates of faults is to accurately measure the offsets and ages of terraces. In this study, the differential GPS with a centimeter accuracy was used to obtain the elevation of terraces above the riverbed near faults, based on which the longitudinal profiles of the river terraces were established. We first investigated the development of river terraces near the faults and divided the river terraces into different levels based on the elevation, distribution areas, and composition. Along the Changma fault, we identified three terrace levels and named them T1 to T3 (**Figures 5A,B**). Along the Hanxia and Yumen faults, we divided the terraces into 5–6 levels (**Figures 6A,B,D**). Along the Bainan fault and its anticline, we identified five terrace levels (**Figures 7B,C**). Then, we used the Trimble R10 GNSS measurement system to obtain the elevation of terraces. In this way, we obtained the elevation data of T1–T3 terraces near the Changma fault, T1–T5 near the Yumen fault, T1–T5 near the Bainan fault and the riverbed (T0). At last, the longitudinal profiles of the terraces were established to derive the offsets of the terraces due to faulting. T3 along the Changma fault and T5 along the Yumen fault represent the regional terraces in the upper and middle reaches and its ages are also similar, demonstrating that these two terraces are comparable (**Table 1**).

We employed two dating methods (optically stimulated luminescence, OSL, and terrestrial cosmogenic nuclide dating, ^{10}Be) to estimate the ages of the offset terraces. Near the Changma fault, we collected the OSL dating sample from the loess layer approximately 2 m below the surface of T3 (**Figure 5A**). Between the Hanxia fault and Yumen fault, we collected the ^{10}Be dating samples of T3 and T5 from the terrace well-developed section (**Figure 6A**). On the north side of the Bainan fault, we collected ^{10}Be samples from all terraces (**Figure 7B**). All sampled terraces are well-developed regional terraces in their respective sections. The dating procedures can be referred to our previous study of Liu R. et al. (2020). Some results of offsets and ages of fault terraces have been published in our previous works (Liu R. et al.,



2017; Chen and Koronovskii, 2020). In this study, we reorganized these data for our research purpose and listed them in **Table 1**.

4 RESULTS

4.1 Slope Map and Swath Profiles Based on the Digital Elevation Model

To characterize the topographic features of the Qilian Shan, a slope map and two swath profiles based on high-resolution DEM data of the Qilian Shan were extracted in this study (**Figures 3A–D**). The slope map shows that plains, gentle slopes, and moderate slopes are mainly distributed in the basins and valleys within and around the Qilian Shan and are widely distributed in the south of the Qilian Shan. However, they are limitedly distributed in band-shaped intermontane basins in the central and northern Qilian Shan (**Figure 3B**). Areas with steep slopes are strictly limited to the foothills of mountain ranges. Cliffs and vertical walls best represent the high-relief topography and are widespread in all NWW-trending band-shaped mountains of the Qilian Shan, but their distributions differ. In the northern Qilian

Shan, they are widely distributed, extending from the northern margin of the Qilian Shan to the Tuolai Nan Shan, accounting for more than 60% of the entire distribution of cliffs and vertical walls (**Figure 3B**). In the central Qilian Shan, they are in the top areas of the Shule Nan Shan and the Danghe Nan Shan, accounting for ~30% of the distribution of cliffs and vertical walls. However, in the southern Qilian Shan, cliffs and vertical walls are rare; they are distributed at the peaks of mountain ranges, accounting for less than 10% of the distribution of cliffs and vertical walls (**Figure 3B**). Therefore, the spatial variation of the slopes shows that the distribution of areas with high slopes in the northern Qilian Shan is significantly larger than that in the southern and central Qilian Shan.

We extracted two swath profiles with an extension direction perpendicular to the strike of the mountains and plotted four curves (Max, Mean, Min, and Range curves) based on calculating the elevations in the swath ranges (**Figures 3B–D**). The range curve best represents the terrain relief in the study area. In the western profile (**Figure 3C**, swath profile C–C'), the range is comparatively low in the southern and central areas, reaching ~800 m in the southern Qilian Shan, ~1,000 m in the Danghe Nan

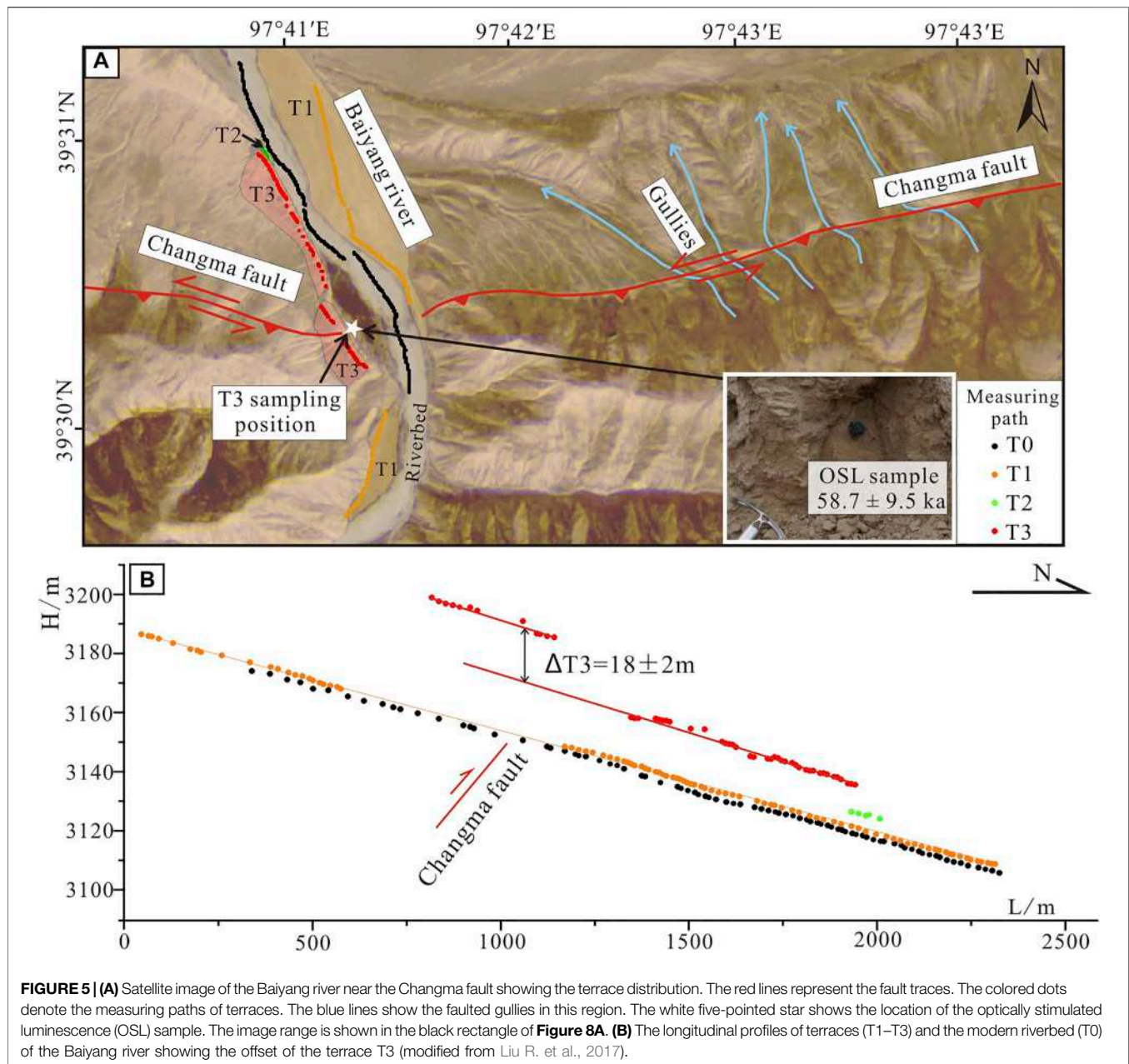


FIGURE 5 | (A) Satellite image of the Baiyang river near the Changma fault showing the terrace distribution. The red lines represent the fault traces. The colored dots denote the measuring paths of terraces. The blue lines show the faulted gullies in this region. The white five-pointed star shows the location of the optically stimulated luminescence (OSL) sample. The image range is shown in the black rectangle of **Figure 8A**. **(B)** The longitudinal profiles of terraces (T1–T3) and the modern riverbed (T0) of the Baiyang river showing the offset of the terrace T3 (modified from Liu R. et al., 2017).

Shan, and the minimum of 700 m in the Shule Nan Shan. From the Tuolai Nan Shan to the northern Qilian Shan, the range significantly increases. In the Tuolai Nan Shan, the range steadily varies between 700 and 1,000 m, but that is above 1,000 m within a width of 70 km in the northern Qilian Shan and rapidly reaches the maximum of 2,000 m in the middle of the northern Qilian Shan (**Figure 3C**). In the eastern profile (**Figure 3D**, swath profile D–D'), the range is generally larger than that in the western profile, especially in the central Qilian Shan, and has a greater advantage of 200–300 m in the same mountain ranges. However, the range reaches maxima in the northern Qilian Shan in both profiles. In the eastern profile, the range reaches up to 1,000 m within a width of 35 km in the

northern Qilian Shan, with a maximum of 1,400 m (**Figure 3D**).

In summary, the topographic slope map and two swath profiles of the Qilian Shan show that the distribution of the high-relief terrain and the range value reach maxima in the northern Qilian Shan, demonstrating that the terrain relief in the northern Qilian Shan is much greater than that in the central and southern regions. The high terrain relief, in general, results from intense erosion, which is closely related to tectonics and climate. However, located in a same region, the mountains in the Qilian Shan have similar climatic conditions. Thus, the terrain relief in the northern Qilian Shan might be related to the rapid surface uplift in the northern Qilian Shan. Furthermore, the

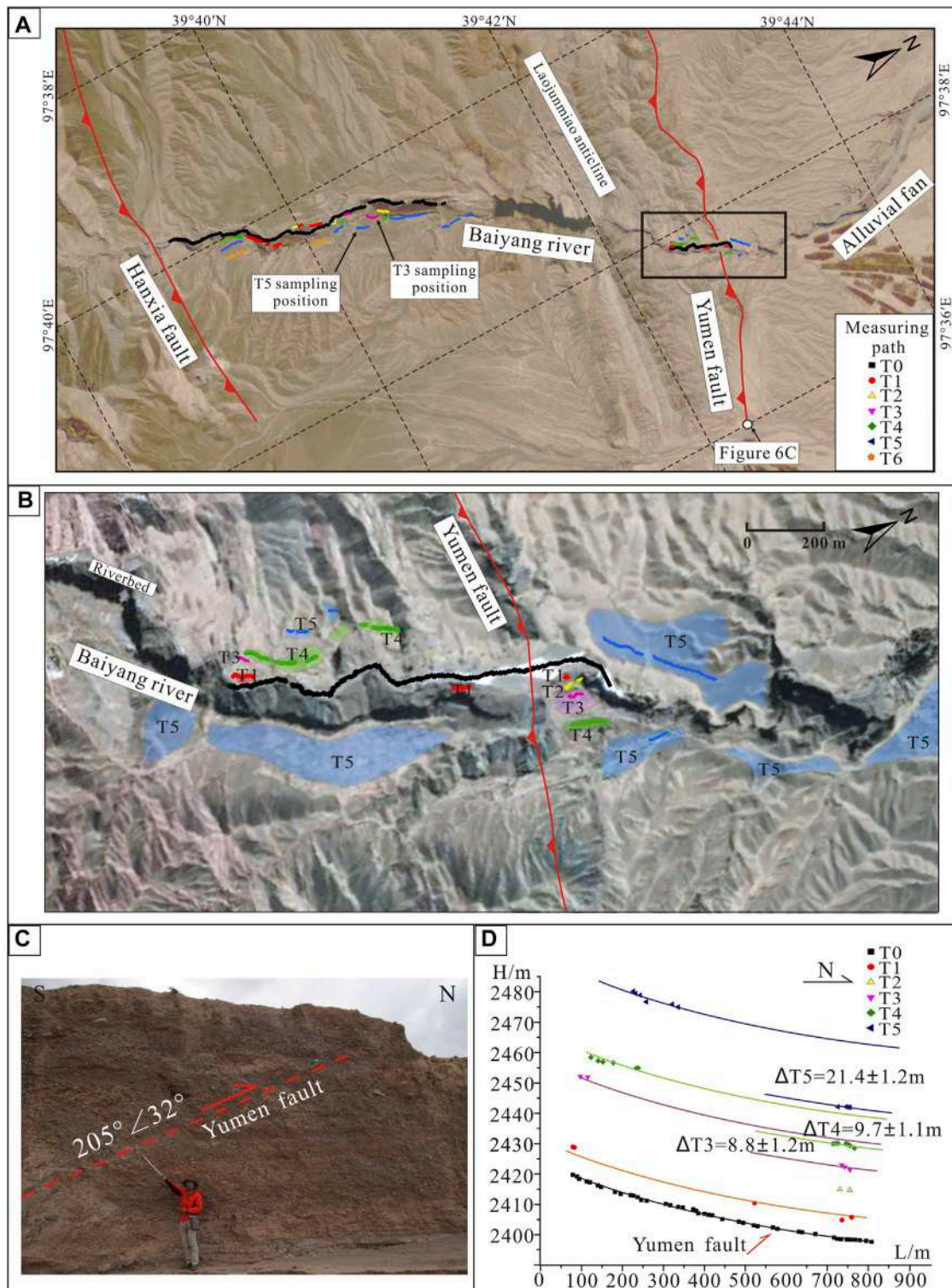
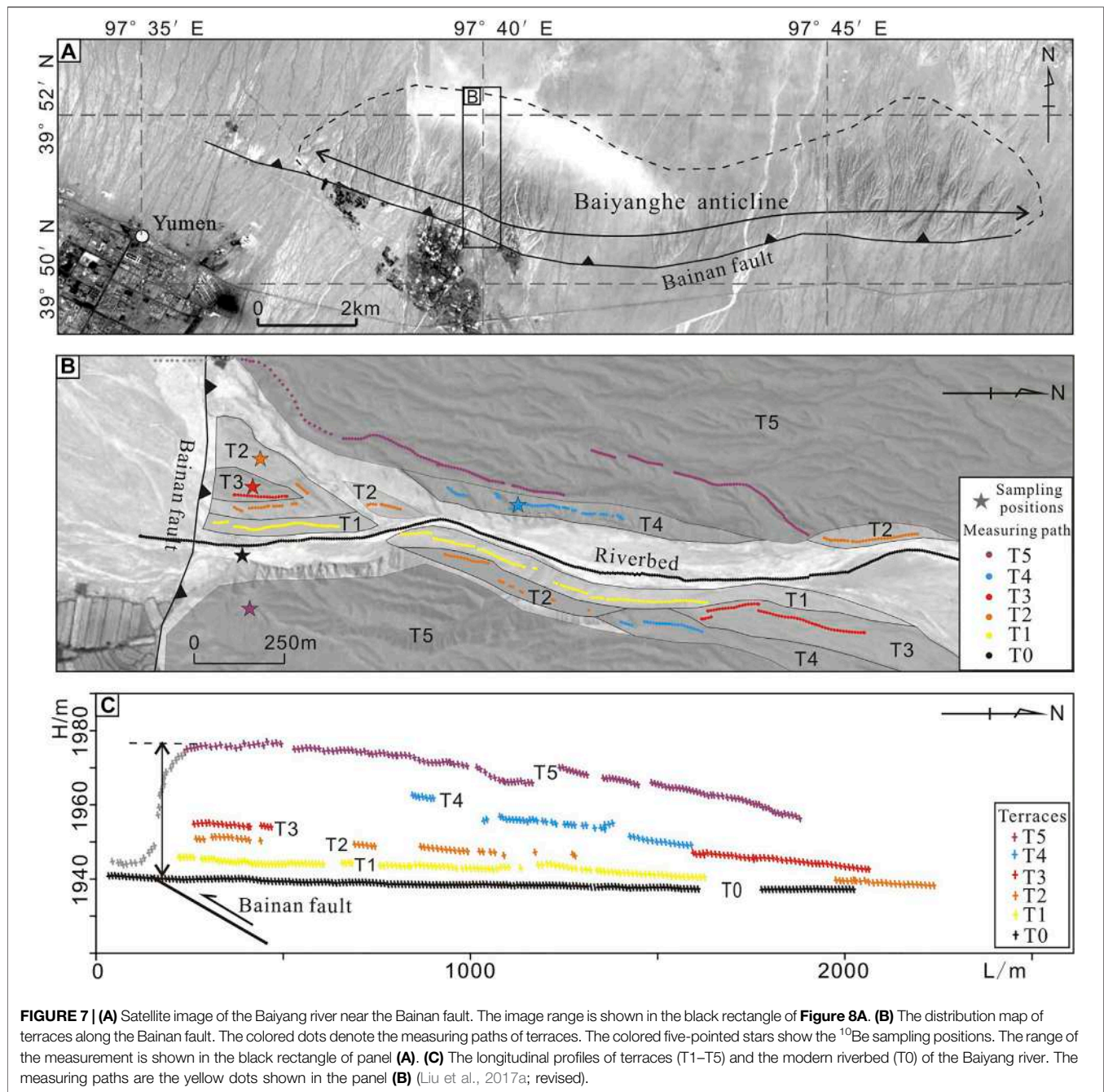


FIGURE 6 | (A) Satellite image of the Baiyang river in the northern Qilian Shan front. The red lines represent the fault traces. The colored dots denote the measuring paths of terraces. The ^{10}Be sampling positions of T3 and T5 terraces have been exhibited. The image range is shown in the black rectangle of **Figure 8A**. **(B)** Satellite image of the Baiyang river near the Yumen fault showing the terrace distribution. The image range is shown in the black rectangle of **(A)**. The colored dots denote the measuring paths of terraces. **(C)** The photograph of fault scarp of the Yumen fault in the alluvial fan. The location of photograph is shown in the **(A)**. **(D)** The longitudinal profiles of terraces (T1–T5) and the modern riverbed (T0) of the Baiyang river near the Yumen fault. The measuring paths have been shown in the **(B)** (Chen and Koronovskii, 2020; revised).



research results of the Longmen Shan, the eastern margin of the Tibetan Plateau, also suggested that crustal deformation and topographic relief in mountain ranges are largely coupled (Hubbard and Shaw, 2009).

4.2 Geomorphologic Characteristics of the Baiyang River

The Baiyang river is a north–south river in the northern Qilian Shan. It originates within the Qilian Shan and flows northward through the northern mountain ranges of the Qilian Shan,

roughly perpendicular to the strike of the mountains and faults. After crossing the mountain ranges, it continues to incise piedmont alluvial fans and flows into the Jiuxi Basin (**Figure 4A**). The Baiyang river flows through four active faults, which from south to north are the Changma, Hanxia, Yumen, and Bainan faults (**Figure 4A**). In this study, the Baiyang river was divided into upper, middle, and lower reaches based on the distribution of faults and fluvial topographic characteristics (Liu R. et al., 2017; Chen and Koronovskii, 2020).

In the upper reaches of the Baiyang river, wide and U-shaped valleys are developed. The riverbeds are wide (up to 50–150 m

TABLE 1 | The displacements and ages of terraces and slip rates of faults in the Baiyang river.

Faults	Terraces	Offsets (m) ^a	Ages (ka) ^a	Vertical slip rate (mm/a)
Changma	T3	18 ± 2	58.7 ± 9.5	0.31 ± 0.06
Yumen	T3	8.8 ± 1.2	13.0 ± 4.5	0.67 ± 0.2
	T5	21.4 ± 1.2	64.4 ± 9.0	0.33 ± 0.02
Bainan	T1	5.0	12.0 ± 2.0	0.41 ± 0.04
	T2	10.5	39.9 ± 5.8	0.27 ± 0.04
	T3	14.5	73.4 ± 8.8	0.20 ± 0.02
	T4	23.5	85.0 ± 9.8	0.28 ± 0.03
	T5	36	174.6 ± 18.2	0.21 ± 0.02

^aThe age of terrace T1 near the Bainan fault from Hetzel et al. (2002) and other offset and age data of terraces from our previous studies of Liu R. et al. (2017) and Chen and Koronovskii, (2020).

wide) and flat with a coarse gravel cover. We observed two to three terraces upstream and named them T1 to T3 from youngest to oldest (below naming terraces in the same way). Terraces T1 and T2 are widely distributed, but T3 can only be observed locally. The surfaces of terraces T1, T2, and T3 are 2–3, 7–10, and 20–30 m higher than the modern riverbed. The T1, T2, and T3 are accumulation terraces, which are mainly composed of fluvial gravel and covered with eolian loess (**Figures 4D,G**).

The middle reaches of the Baiyang river flow through the main mountain ranges of the northern Qilian Shan in which the valleys significantly narrow and the riverbeds that are more than 100 m wide upstream quickly narrow down to a width of 10 m in the midstream. Compared with the wide valleys upstream, deep and V-shaped valleys are developed in the midstream and the terrace type gradually changes from accumulation to erosion and basement terraces. Unconformities occur between the fluvial gravel and bedrock. The bedrock is composed of Paleozoic limestone, Mesozoic limestone and sandstone, Cenozoic orange sandstone, and Pleistocene conglomerate and the overlying layers consist of fluvial gravel and aeolian loess (Liu X.-w. et al., 2019). With the gradual increase in the terrace levels, high terraces are widely developed. In the south of the midstream, the number of terrace levels is 3–5. In the north of the midstream near the Hanxia fault, the terrace levels increase to 6–7. Terraces T1 to T7 are 2–3, 5–10, 20–30, 45–50, 65–75, 85–90, and up to 110 m above the modern riverbed, respectively (**Figures 4C,F**).

The lower reaches of the Baiyang river incise the piedmont alluvial fans of the northern Qilian Shan. In the range of the alluvial fans, the valleys widen and the types of valleys change from V-shaped valleys midstream to U-shaped valleys in the alluvial fan area. However, the gravel layer in the fan is relatively loose; thus, the valleys are still very deep. Low terraces downstream are poorly developed and only two high terraces are preserved. Based on the height of the river terraces, the two high terraces downstream can contrast with the T4 and T5 terraces in the midstream, 45–50 and 70–80 m above the modern riverbed, respectively. These terraces are mainly basement terraces. The basement is composed of Pleistocene conglomerates and the overlying layers consist of fluvial gravel and aeolian loess. After crossing the alluvial fan, the river flows

into the Jiuxi Basin and dries up in the middle of the basin (**Figures 4B,E**).

Based on the fluvial geomorphological characteristics of the Baiyang river, the fluvial geomorphic features in the middle reaches notably differ from those in the upper and lower reaches, mainly manifested by narrow, deep, and steep V-shaped valleys, an increase in the terrace level, wide distribution of high terraces, and different terrace types (mainly erosional and basement terraces). The above-mentioned characteristics indicate that the erosion in the middle reaches of the Baiyang river is more intense than that in the other reaches. As a regional river at the northern margin of the Qilian Shan, the Baiyang river has a similar regional climate and hydrodynamic environment, but the notable differences in the fluvial topography among the three reaches may be related to different intensities of tectonic movement, indicating that the northern Qilian Shan was subject to more intense tectonism.

4.3 Kinematics of Faults in the Baiyang River Catchment

From south to north, the Baiyang river flows through the Changma fault, northern Qilian fault (including the Hanxia fault and Yumen fault and associated Laojunmiao anticline), and Bainan fault and associated Baiyanghe anticline. The long-term slip rates of the faults since the Late Quaternary were determined (Liu R. et al., 2017; Chen and Koronovskii, 2020).

4.3.1 Changma Fault

The Changma fault with a total length of ~120 km is a sinistral strike-slip fault in the Qilian Shan with reverse components. Its western end is connected to the Altyn Tagh fault. The Changma fault is NWW-striking and inclined to the southwest, with a dip angle of 50°–70°. In the Baiyang river catchment, a fault plane can be observed along the Changma fault, with the occurrence of 165°/70° (Liu R. et al., 2017). Based on the fault plane, the Changma fault shows reverse faulting characteristics, that is, the Precambrian gray gneiss thrusts over the Miocene orange sandstone (Liu R. et al., 2017). Based on the longitudinal profile of the terraces near the Changma fault, we observed a significant vertical displacement of 18 ± 2 m for terrace T3 (**Figure 5B**). Combined with the age of T3 of 58.7 ± 9.5 ka obtained by the OSL dating method, the vertical slip rate of the Changma fault since ~59 ka was estimated to be 0.31 ± 0.06 mm/a and the horizontal shortening rate was determined to be 0.11 ± 0.02 mm/a (**Table 1**) (Liu R. et al., 2017). In addition, notable sinistral offset ridges and gullies can be observed on the eastern bank of the Baiyang river, indicating that the Changma fault is affected by sinistral strike-slip (**Figure 5A**). Based on the offsets and ages of the deformed terraces, the estimated sinistral strike-slip rate of the Changma fault is 3–5 mm/a in the Baiyang river catchment, which is in the eastern segment of the Changma fault (**Figure 2**) (Luo et al., 2013; Du et al., 2020).

4.3.2 Northern Qilian Fault

The northern Qilian fault is the boundary between the northern Qilian Shan and adjacent Jiuxi Basin, which controls the

geomorphologic contour of the study area. In the Baiyang river catchment, this fault can be divided into two NWW-striking secondary faults, that is, the Hanxia fault in the south and the Yumen thrust fault in the north (**Figure 6A**). Based on the fault scale, the Hanxia fault is the main boundary fault between the northern Qilian Shan and Jiuxi Basin. The Yumen fault is considered to be a newly formed thrust due to the northward expansion of the Qilian Shan (Chen and Koronovskii, 2020). Between the Hanxia and Yumen faults, the fault-related NWW-trending anticline fold (Laojunmiao anticline) is 7–10 km wide from south to north and ~50 km long from west to east (**Figure 6A**). The Hanxia fault, Yumen fault, and Laojunmiao anticline jointly constitute the northern Qilian frontal thrust system.

Based on topographic profiles, the river terraces along the Hanxia fault are insignificantly deformed, but distinct fault surfaces were observed in natural outcrops where the Ordovician limestone thrusts over the Yumen conglomerate of the Lower Pleistocene and the occurrence of the fault plane is $178^{\circ}\angle 75^{\circ}$ (Liu R. et al., 2017). Based on geological mapping, the Hanxia fault has been inactive since the Pleistocene and active geomorphological evidence has not been found in the Baiyang valleys (Institute of Geology and Lanzhou Institute of Seismology, 1993). In addition, notable deformation cannot be observed for the terraces of the Shiyu river, a south–north river on the west side of the Baiyang river (Hetzl et al., 2006).

The activity of the Yumen fault and Laojunmiao anticline caused significant deformation the terraces of in the Baiyang river catchment. Although the Yumen fault is regarded as a blind thrust at the western front of the Qilian Shan (Liu X.-w. et al., 2019), we observed notable fault traces of the Yumen fault in the gullies on the east side of the Baiyang river, where the fault cuts through the gravel deposits of alluvial fans and its occurrence was measured to be $205^{\circ}\angle 32^{\circ}$ (**Figure 6C**) (Chen and Koronovskii, 2020). Thrusting leads to the vertical displacement of the terraces of the Baiyang river. The longitudinal profile of the terraces of Baiyang river shows that T3, T4, and T5 along the Yumen fault were significantly deformed, with displacements of 8.8 ± 1.2 , 9.7 ± 1.1 , and 21.4 ± 1.2 m, respectively (**Figures 6B,D**). The abandonment ages of T3 and T5 obtained by terrestrial cosmogenic nuclide dating (^{10}Be) are 13.0 ± 4.5 and 64.4 ± 9.0 ka, respectively (**Table 1**) (Liu R. et al., 2017). Based on the dip angle of the fault, we determined the vertical slip and horizontal shortening rates of the Yumen fault since ~13 ka to be 0.67 ± 0.2 and 1.06 ± 0.5 mm/a, respectively. The vertical slip and horizontal shortening rates of the Yumen fault since ~64 ka were determined to be 0.33 ± 0.02 and 0.53 ± 0.03 mm/a, respectively (**Table 1**) (Chen and Koronovskii, 2020). Liu R. et al. (2017) constrained the vertical uplift and horizontal shortening rates of the Laojunmiao anticline since ~9 ka to be 1.23 ± 0.81 and 0.67 ± 0.02 mm/a, respectively. Based on the combination of the slip rates of the Hanxia fault, Yumen fault, and Laojunmiao anticline, the long-term vertical slip rate of the western part of the northern Qilian fault is 1.5–2 mm/a and the horizontal shortening rate is 1.2–1.7 mm/a, which is consistent with the shortening rate of 1.2–1.9 mm/a determined for the eastern segment of the Yumen fault (Liu X.-W. et al., 2017; Liu X.-w. et al., 2019; Liu X. et al.,

2019), slightly higher than the vertical slip rate of 0.9–1.3 mm/a and horizontal shortening rate of 1.1–1.4 mm/a reported for the middle segment of the northern Qilian fault since the Late Quaternary (Yang et al., 2018a) and significantly higher than the vertical uplift rate of 0.4–0.8 mm/a determined in the eastern part of the northern Qilian fault (Hu et al., 2015).

4.3.3 Bainan Fault

The Bainan fault is in the Jiuxi basin and has a NWW trend, northward dip, dip angle of 30° , and length of ~20 km (Liu R. et al., 2017). A fault-related fold (Baiyanghe anticline) is developed on the northern side of the Bainan fault (**Figure 7A**). The Bainan fault and Baiyanghe anticline, located on the alluvial fan in the center of the basin, considered to represent the reverse structure of the northern Qilian fault (Min et al., 2002). River erosion and tectonic movement may jointly result in the development and uplift of multiple terraces of the Baiyang river. However, its uplifting topographic expression contrasts sharply with the surrounding flat surface of the alluvial fan, indicating that the tectonic activity may play a crucial role in the growth of the fault and anticline. Therefore, we focus on the effects of the internal dynamics rather than the external dynamic process. From the longitudinal profiles, terraces T1–T5 are 5.0, 10.5, 14.5, 23.5, and 36 m above the modern riverbed (**Figures 7B,C**). The abandonment ages of T1–T5 based on terrestrial cosmogenic nuclide (^{10}Be) dating are 12.0 ± 2.0 , 39.9 ± 5.8 , 73.4 ± 8.8 , 85.0 ± 9.8 , and 174.6 ± 18.2 ka, respectively (**Table 1**) (Liu R. et al., 2017). Based on the combination of the heights and ages of the terraces, we determined the uplift rates of the Bainan fault and Baiyanghe anticline since the formation of T1–T5 to be 0.41 ± 0.04 , 0.27 ± 0.04 , 0.20 ± 0.02 , 0.28 ± 0.03 , and 0.21 ± 0.02 mm/a, respectively (**Table 1**) (Liu R. et al., 2017).

5 DISCUSSION

5.1 Deformation Characteristics of the Northwestern Qilian Shan

Based on the fluvial geomorphologic characteristics of the Baiyang river and the slip rates of faults, the vertical uplift rate of the northern Qilian Shan dominated by the northern Qilian fault is greater than that of the central Qilian Shan controlled by the Changma fault. On the one hand, the geomorphologic characteristics of the Baiyang river derived from the slope map and swath profiles of the western Qilian Shan based on DEM data suggest that the river erosion in the northern Qilian Shan is more intense than that in the central Qilian Shan and the terrain relief reaches a maximum in the northern Qilian Shan. On the other hand, the vertical slip rates of the Changma and the northern Qilian faults since the Late Pleistocene were estimated to be 0.31 ± 0.06 and 1.5–2 mm/a, respectively, based on comparing the slip rates of the main faults in the Baiyang river catchment (Luo et al., 2013; Liu R. et al., 2017), demonstrating that the northern Qilian Shan has been uplifted faster than the central Qilian Shan.

Based on this study, the age of T3 along the Changma fault, that of T5 along the Yumen fault, and the average age of T2 and T3 along the Bainan fault are comparable (~60 ka). Based on the

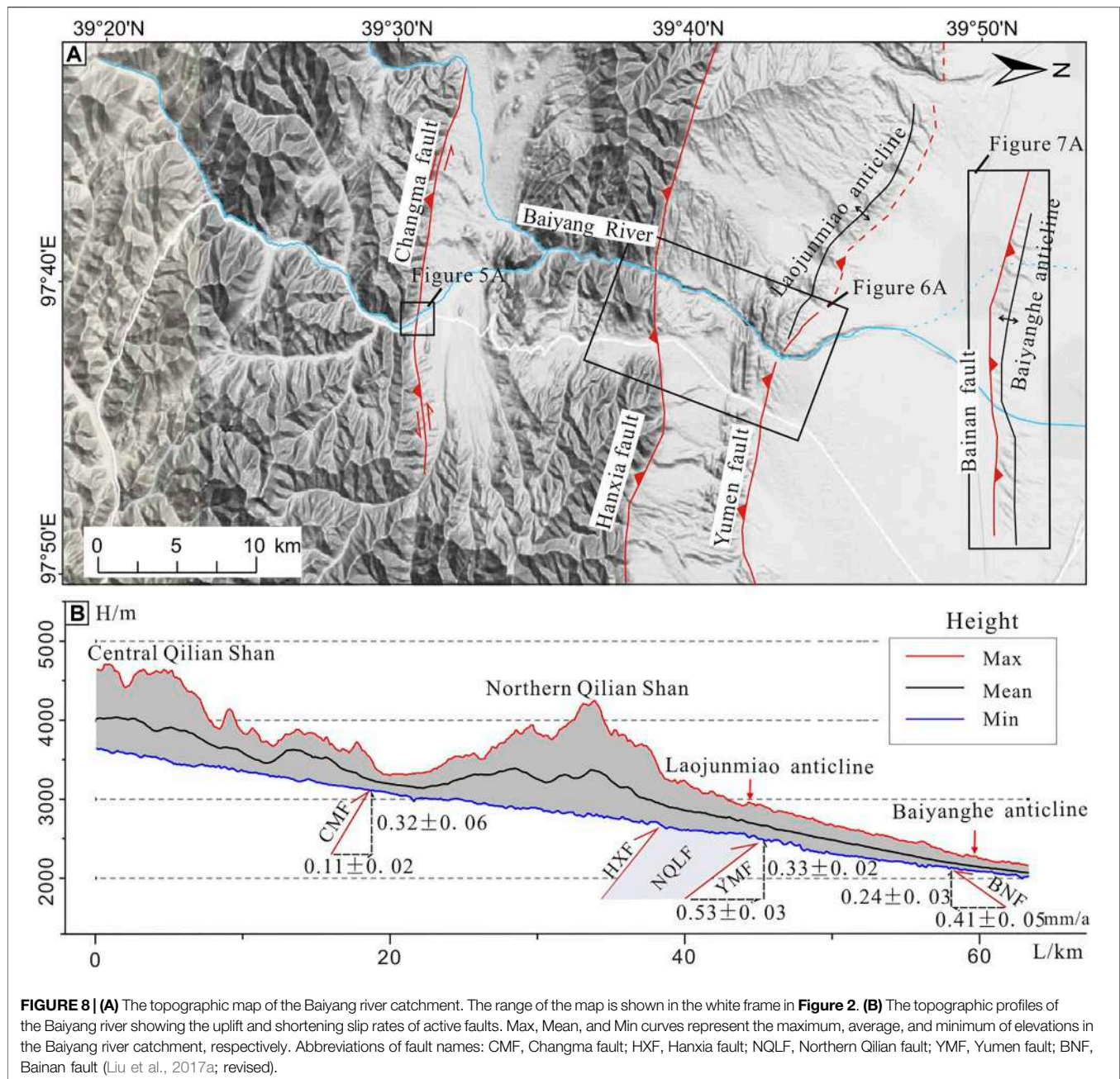


FIGURE 8 | (A) The topographic map of the Baiyang river catchment. The range of the map is shown in the white frame in **Figure 2**. **(B)** The topographic profiles of the Baiyang river showing the uplift and shortening slip rates of active faults. Max, Mean, and Min curves represent the maximum, average, and minimum of elevations in the Baiyang river catchment, respectively. Abbreviations of fault names: CMF, Changma fault; HXF, Hanxia fault; NQLF, Northern Qilian fault; YMF, Yumen fault; BNF, Bainan fault (Liu et al., 2017a; revised).

combination of the estimated fault slip rates with the dip angles, the vertical uplift rates of the Changma, Yumen, and Bainan faults since ~60 ka were estimated to be 0.31 ± 0.06 , 0.33 ± 0.02 , and 0.24 ± 0.03 mm/a, respectively, and the horizontal shortening rates were determined to be 0.11 ± 0.02 , 0.53 ± 0.03 , and 0.41 ± 0.05 mm/a, respectively (**Figures 8A,B**). Although the Hanxia fault might be the main fault based on its scale, a significant fault trace cannot be observed along the Hanxia fault (Hetzel et al., 2006; Liu R. et al., 2017; Liu X.-W. et al., 2017; Liu X.-w. et al., 2019). Instead, the Yumen fault and its anticline (Laojunmiao anticline) were notably deformed and their uplift and shortening rates since the Late Pleistocene are 1.5–2 and 1.2–1.7 mm/a,

respectively (**Figure 9A**) (Liu R. et al., 2017). Thus, both vertical uplift rates and shortening rates of the northern Qilian fault are significantly higher than those of the Changma fault within the Qilian Shan. However, it should be noted that the activity intensity of the Yumen fault at the northern Qilian Shan front is greater than that of the Hanxia fault (Hetzel et al., 2006). The reason for this phenomenon may be that the active area at the front of the northern Qilian Shan migrated from the Hanxia fault to the Yumen fault and associated fold belt due to the northward expansion of the Qilian Shan (**Figure 9B**).

Despite the relatively faster vertical uplift and horizontal shortening rates along the northern Qilian fault, a notable

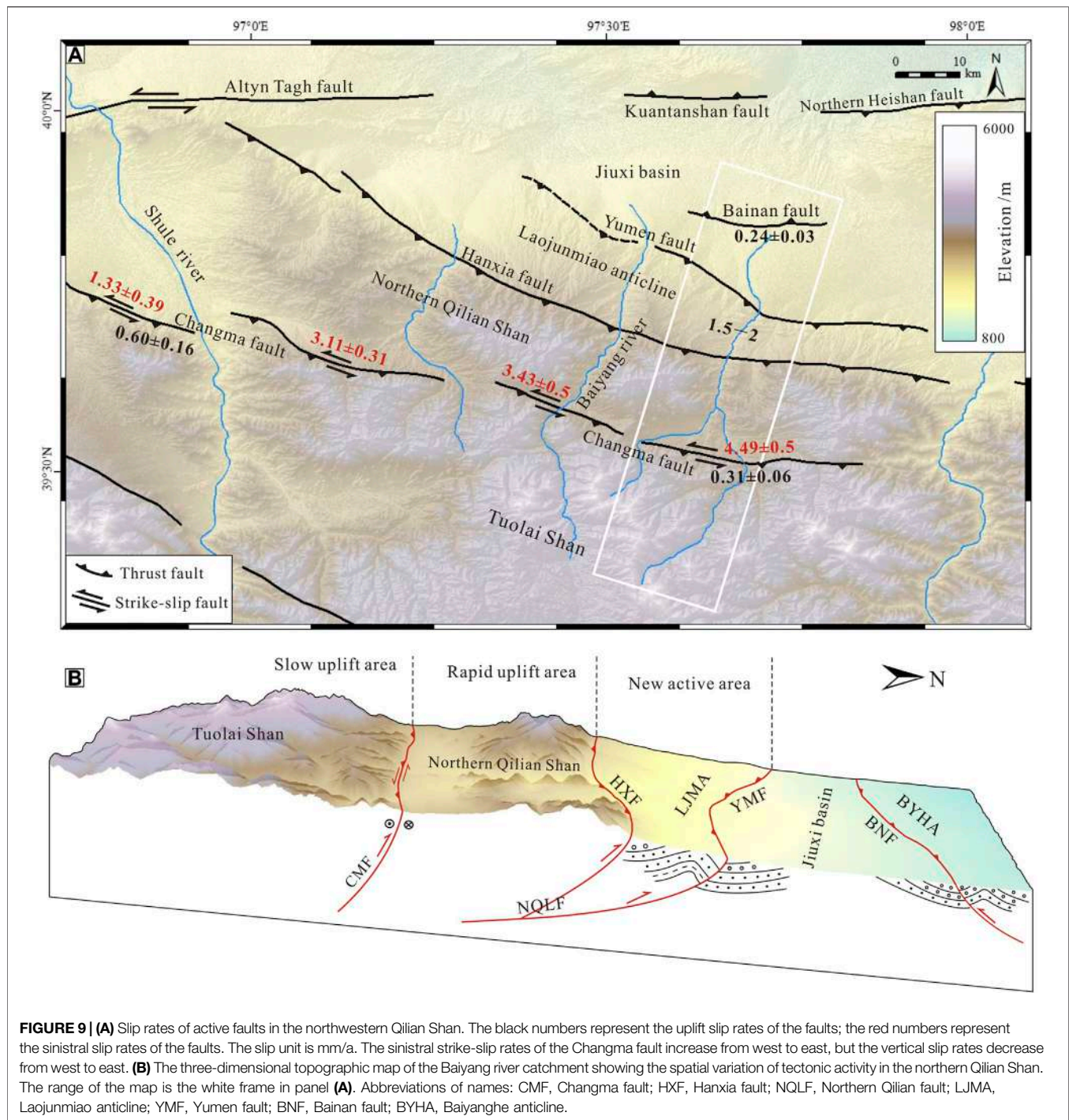


FIGURE 9 | (A) Slip rates of active faults in the northwestern Qilian Shan. The black numbers represent the uplift slip rates of the faults; the red numbers represent the sinistral slip rates of the faults. The slip unit is mm/a. The sinistral strike-slip rates of the Changma fault increase from west to east, but the vertical slip rates decrease from west to east. **(B)** The three-dimensional topographic map of the Baiyang river catchment showing the spatial variation of tectonic activity in the northern Qilian Shan. The range of the map is the white frame in panel (A). Abbreviations of names: CMF, Changma fault; HXF, Hanxia fault; NQLF, Northern Qilian fault; LJMA, Laojunmiao anticline; YMF, Yumen fault; BNF, Bainan fault; BYHA, Baiyanghe anticline.

strike-slip displacement has not been observed for this fault (Yang et al., 2018a). In contrast, the tectonic movement of the Changma fault resulted in notable sinistral displacements of ridges and gullies in the Baiyang river catchment. Based on the deformation and ages of the terraces along the Changma fault, the sinistral slip rates of the Changma fault in the western, middle, mid-eastern, and eastern segments were estimated to be 1.33 ± 0.39 , 3.11 ± 0.31 , 3.43 ± 0.5 , 4.49 ± 0.5 mm/a, respectively (Luo et al., 2013; Du et al., 2020). Conversely,

the vertical slip rate was estimated to be 0.60 ± 0.16 mm/a in the western segment and 0.31 ± 0.06 mm/a in the eastern segment (Figure 9A) (Luo et al., 2013; Liu R. et al., 2017). The sinistral strike-slip rates of the Changma fault gradually increase from west to east, whereas the vertical slip rates gradually decrease from west to east. Consequently, in the northwestern region of the Qilian Shan, crustal uplift and shortening largely occurred in the northern Qilian Shan and the range-bounding faults, but strike-slip occurred along the

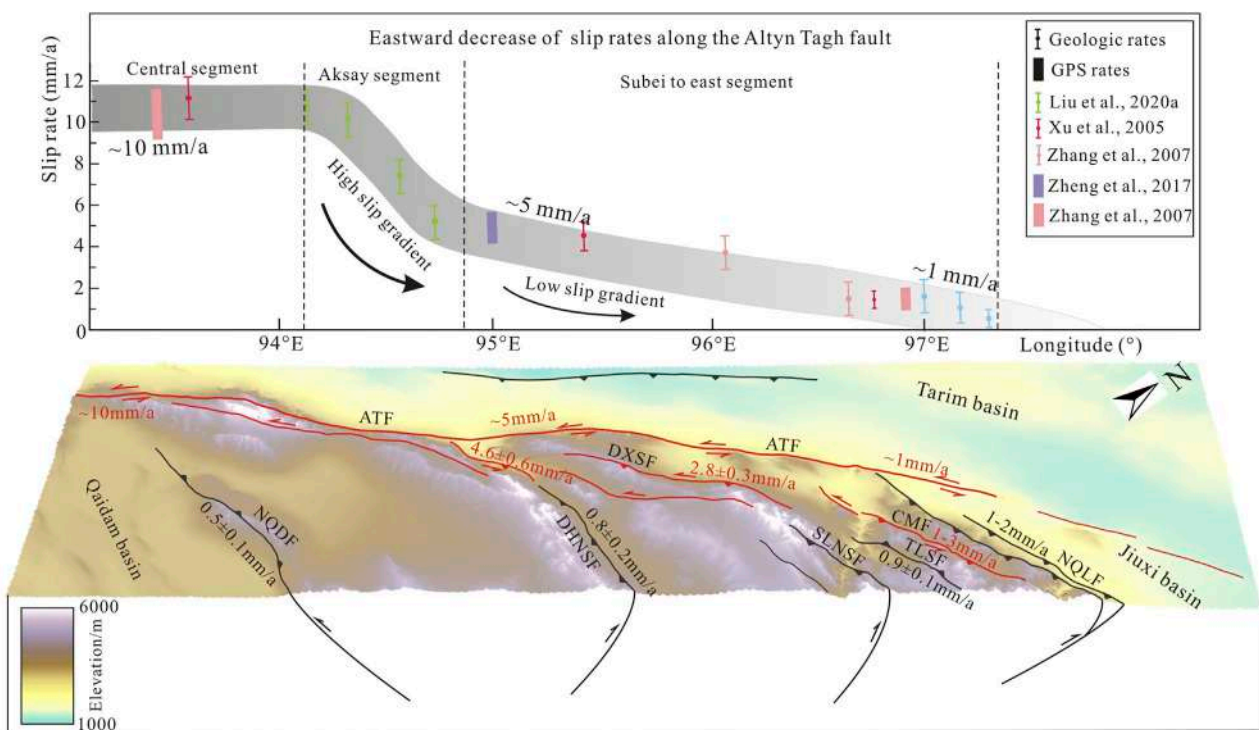


FIGURE 10 | Spatial variation of slip rates along the Altyn Tagh fault and three-dimensional DEM map of the western Qilian Shan. The upper figure shows an eastward decrease of slip rates along the Altyn Tagh fault, exhibiting steady slip rates in the central segment, a high slip gradient in the Aksay segment and a low slip gradient from Subei to the east segment (modified from Liu J. et al., 2020). In the DEM map, the black lines and numbers represent reverse faults and thrusts and its shortening rates; the red lines and numbers represent sinistral strike-slip faults and its sinistral slip rates. Abbreviations of fault names: ATF, Altyn Tagh fault; NQDF, Northern Qaidam fault; DHNSF, Danghe Nan Shan fault; SLNSF, Shule Nan Shan fault; TLSF, Tuolai Shan fault; CMF, Changma fault; NQLF, Northern Qilian fault.

Changma fault within the Qilian Shan, accommodating 1–3 mm/a of the sinistral strike-slip of the Altyn Tagh fault.

5.2 Deformation Pattern of the Western Qilian Shan in Response to the Crustal Shortening and Sinistral Strike-Slip of the Altyn Tagh Fault

The northern margin of the Tibetan Plateau consists of the Altyn Tagh fault, Qilian orogenic belts, and Haiyuan fault, which are characterized by large-scale lateral strike-slip, the uplift of NWW-trending mountains, and thrusting. However, the deformation pattern of the Qilian Shan in response to crustal shortening and sinistral strike-slip remains controversial because the strain partitioning in this region is poorly understood. The Qilian Shan is regarded to be a continental stepover between the Altyn Tagh and Haiyuan faults (Zhang et al., 2007). The GPS observations and geological investigations reveal that the Qilian Shan accommodates the sinistral slip rate component of ~ 8 mm/a of the Altyn Tagh fault and the strain is partitioned into the compressional deformation and sinistral strike-slip of the faults in the Qilian Shan (e.g., Zhang et al., 2007; Zheng et al., 2013b; Liu J. et al., 2020; Xu et al., 2021). Moreover, the GPS data also show NNE shortening at a rate of 5.5 ± 1.5 mm/a between the Qaidam Basin and Hexi Corridor, which is mainly absorbed by NWW-

trending mountain ranges and fault zones in the western Qilian Shan (Zhang et al., 2004).

The Altyn Tagh fault, as a large-scale boundary fault between the Tarim block to the north and the Tibetan Plateau to the south, plays an important role in controlling the tectonic deformation of the Tibetan Plateau during its northward growth. Two models, that is, the “eastward extrusion model” and “crust thickening model,” can be used to explain the deformation of the plateau. The “eastward extrusion model” is supported by the very high slip rates on the main boundary faults of the rigid blocks and the “crust thickening model” is characterized by continuous deformation distributed within the blocks as well as on the boundary faults, that is, a relatively low slip rate on the Altyn Tagh fault (Liu J. et al., 2020; Zhang et al., 2007). In recent years, results from geological studies, GPS observations, and interferometric synthetic aperture radar measurements suggested that the average slip rate of the Altyn Tagh is 8–10 mm/a on different timescales (e.g., Zhang et al., 2004; Xu et al., 2005; Zhang et al., 2007; Zheng et al., 2013b). This implies that the previous estimation of 20–30 mm/a may present an overestimation (e.g., Molnar et al., 1987; Peltzer and Tapponnier, 1989; Tapponnier et al., 2001). Most results show that the slip rate in the central part of the Altyn Tagh fault is relatively high and can reach 9–12 mm/a (Figure 10) (e.g., Zhang et al., 2007; Zheng et al., 2013b; Liu J. et al., 2020). Researchers

have noted an ununiform eastward decrease in the slip rate along the eastern segment of the Altyn Tagh fault, specifically at the junction of the Altyn Tagh fault and western Qilian Shan (e.g., Zhang et al., 2007; Zheng et al., 2013b; Liu J. et al., 2020). Liu et al. (2020a) revealed that the slip rate of the Altyn Tagh fault begins to significantly decrease in the Aksay segment in which the Altyn Tagh fault connects to the Danghe Nan Shan in the central Qilian Shan. In the Aksay–Subei segment, the slip rate of the Altyn Tagh fault decreases from 10 ± 1 to 5.1 ± 0.8 mm/a within a distance of 50 km. However, subsequently, the slip rate slowly decreases from ~ 5 mm/a in Subei to 1–2 mm/a in the easternmost segment of the Altyn Tagh fault (Figure 10) (Liu J. et al., 2020). Relatively low slip rates obtained along the Altyn Tagh fault suggest that the activity of the Altyn Tagh fault is not compatible with the “block extrusion model,” but supports the “crust thickening model.” In addition, the high decay of the slip rate in the eastern segment of the Altyn Tagh fault may be related to crustal thickening, thrusting, and left-lateral strike-slip in the Qilian Shan, resulting in blocking at this position (Liu J. et al., 2020).

The NNE-ward crustal shortening and sinistral strike-slip of the Altyn Tagh fault are mainly accommodated by NWW-trending mountains and fault zones in the western Qilian Shan. Understanding the strain partitioning of these faults is important for studying the deformation patterns in the northern plateau. From south to north, five large-scale NWW-trending fault zones are distributed in the western Qilian Shan, that is, the northern Qaidam, Danghenan Shan, Shulenan Shan, Changma, and northern Qilian faults (Figures 1B,C). In most of the previous studies in the region, the focus was placed on boundary faults, especially the northern Qilian fault system, and the faulting or folding kinematics within the Qilian orogen were poorly documented. However, it is necessary to evaluate the deformation pattern in northeastern Tibet (Hu et al., 2021). The northern Qilian fault has a vertical slip rate of 0.8–1.4 mm/a and a shortening rate of 1–2 mm/a due to Late Quaternary geomorphic deformation (Figure 10) (e.g., Hetzel et al., 2002; Hetzel et al., 2004; Hetzel et al., 2006; Hu et al., 2015; Liu X.-W. et al., 2017; Yang et al., 2018a; Yang et al., 2018b; Hetzel et al., 2019; Liu X.-w. et al., 2019; Liu X. et al., 2019; Chen and Koronovskii, 2020; Liu R. et al., 2020; Yang et al., 2020; Hu et al., 2021), which is consistent with GPS data (e.g., Zheng et al., 2009; Pan et al., 2020) and the long-term shortening rate inferred from balanced cross-sections (Zuza et al., 2016). Notable horizontal strike-slip displacement has not been observed along the northern Qilian fault (Hetzel et al., 2006; Yang et al., 2018a). The results of previous studies showed that the northern Qilian fault has an eastward decreasing shortening rate, varying from 1.2–2 mm/a along the western segment to 0.8–1.4 mm/a in the central and eastern segments (e.g., Chen and Koronovskii, 2020; Hetzel et al., 2004; Hetzel et al., 2006; Hu et al., 2015; Liu R. et al., 2017; Liu X.-W. et al., 2017; Liu X. et al., 2019; Liu R. et al., 2020; Yang et al., 2018a; Yang et al., 2018b; Hetzel et al., 2019; Yang et al., 2020). The Changma fault, which is located in an intramontane basin within the western Qilian Shan, is characterized by left-lateral strike-slip with a thrust component. Its left-lateral strike-slip rates are 1.0–1.6 mm/a in the western segment and 3.0–4.9 mm/a in the eastern segment; however, the vertical slip rates vary from

0.6 ± 0.2 mm/a in the west to 0.3 ± 0.1 mm/a in the east (Figure 9A) (Luo et al., 2013; Du et al., 2020). Hu et al. (2021) derived a vertical slip rate of 1.5 ± 0.1 mm/a and a shortening rate of 0.9 ± 0.1 mm/a based on the deformation and ages of the terraces along the North Tuolai Shan fault (Figure 10), which is the boundary fault between the central and northern Qilian Shan mountains. This deformation rate is similar to the rates determined along the leading-edge fault in the northern Qilian Shan. To the west, a shortening rate of 2.1 ± 0.6 mm/a has been estimated for the northern frontal thrust system of the Daxue Shan (Meyer et al., 1998), which is aligned on the Tuolai Nan Shan and Shulenan Shan, jointly constituting the boundary between the central Qilian Shan and southern Qilian Shan (Hu et al., 2021). By applying ^{10}Be exposure dating to samples from deformed river terraces, Xu et al. (2021) obtained a mean uplift rate of 0.6 ± 0.2 mm/a and a shortening rate of 0.8 ± 0.2 mm/a for the Danghe Nan Shan frontal thrust fault (Figure 10). The southern Zongwulong Shan thrust, the southern margin of the southern Qilian Shan, has a Holocene shortening rate of 0.5 ± 0.1 mm/a (Shao et al., 2019). Hu et al. (2021) suggested that the widely distributed shortening across the Qilian Shan supports the kinematics of distributed upper crustal shortening across the Qilian Shan, rather than a large, propagating crustal-scale deformation wedge.

In spite of the shortening rates of the main NWW-trending faults in the Qilian Shan varying in a similar range and the crustal shortening widely distributed across the Qilian Shan, the terrain relief shows that the crustal shortening varies in different orogenic belts within the Qilian Shan. The results of previous studies verified that crustal shortening, structural relief, and topographic relief correlate in mountain ranges, suggesting that crustal shortening is a primary driver of the uplift and topography of the Longmen Shan at the eastern margin of the Tibetan Plateau (Hubbard and Shaw, 2009). The slope map and swath profiles of the western Qilian Shan established in this study show that the terrain relief in the northern Qilian Shan is significantly higher than that in the central and southern orogenic belts. Specifically, the distribution of high-relief terrain and the relief values reach a maximum in the northern Qilian Shan, indicating that more intense mountain uplift and crustal shortening may occur in the northern Qilian Shan. Thus, we suggest that the crustal shortening in the northern Qilian Shan is more intense than that in the central and southern mountain ranges.

The northern Qilian Shan accounts for considerable crustal shortening in the northeastern Tibetan Plateau, whereas notable strike-slip has not been observed in the northern Qilian frontal region. Previous studies showed that the Qilian Shan accommodates the sinistral rate component of ~ 8 mm/a of the Altyn Tagh fault and transfers the strain into compressional deformation and sinistral strike-slip faults within the Qilian Shan (Liu J. et al., 2020; Zhang et al., 2007). It is necessary to determine the positions accommodating the eastward extrusion of the Altyn Tagh fault in the Qilian Shan. At the junctions between the Altyn Tagh fault and the western Qilian Shan, particularly the Danghe Nan Shan, Daxue Shan, and Changma faults, the slip rates of the Altyn Tagh fault significantly decrease (Hu et al., 2021; Zhang

et al., 2007). In contrast to the northern Qilian frontal thrust, notable sinistral strike-slip has been observed in these regions (Luo et al., 2013, 2015; Shao, 2010). Shao (2010) estimated the Holocene sinistral slip rate of the Danghe Nan Shan fault to be 4.55 ± 0.6 mm/a (**Figure 10**). Luo et al. (2013, 2015) determined the average left-lateral slip rate of the Daxue Shan fault and the slip rate of the Changma fault to be 2.81 ± 0.32 and $1\text{--}3$ mm/a, respectively (**Figure 10**). These results show that the Danghe Nan Shan, Daxue Shan, and Changma faults within the Qilian Shan accommodate almost all sinistral slip rates of the Altyn Tagh fault (**Figure 10**). Liu J. et al. (2020) further suggested that the slip rate decrease along the Altyn Tagh fault is not uniform in junction segments. The slip rate decreases faster in the Aksay–Subei segment, at the intersection between the Altyn Tagh fault and Danghe Nan Shan, but it decreases slowly from Subei to the easternmost segment. Accordingly, the strike-slip of the three NWW-trending faults within the Qilian Shan, rather than the range-bounding faults, account for most of the strike-slip of the Altyn Tagh fault. However, the focus of previous studies was mainly placed on the northern boundary faults of the Qilian Shan and faulting or folding kinematics within the Qilian Shan were poorly documented, which limits our understanding of the tectonic deformation and strain partitioning in the Qilian Shan (Hu et al., 2021).

In summary, the compressional deformation of the Qilian Shan and strike-slip faults within the orogens is due to the NNE-ward crustal shortening of the northeastern Tibetan Plateau and the rapid sinistral strike-slip of the Altyn Tagh and Haiyuan faults. The crustal shortening across the Qilian Shan may be widely distributed in the NWW-trending compressional orogenic belts among which the northern Qilian Shan plays the most important role in absorbing crustal shortening because of its high-relief terrain and more rapid slip rate along the northern Qilian fault. Strike-slip faults within the Qilian Shan are key areas for the accommodation of the strike-slip of the Altyn Tagh fault through the sinistral strike-slip of faults and regional compressional deformation within the Qilian Shan. In addition, it is necessary to enhance the research of tectonic deformation within the Qilian Shan.

6 CONCLUSION

Spatial analysis and fluvial geomorphological studies are powerful tools for studying tectonic deformation within orogens. The topographic slope map and two swath profiles of the western Qilian Shan extracted from high-resolution DEM data show that the terrain relief in the northern Qilian Shan is significantly higher than that in the central and southern Qilian Shan. In addition, the fluvial geomorphologic characteristics of a north–south regional river (Baiyang river) illustrate that the stronger erosion in the middle reaches correlates with the rapid uplift of the northern Qilian Shan, demonstrating that the northern Qilian Shan has been uplifted more rapidly than the central Qilian Shan since the Late Quaternary. Based on high-resolution topographic measurements and chronological

constraints of deformed river terraces along faults in the western Qilian Shan, the vertical slip rates of the Changma, Yumen, and Bainan faults in the Baiyang river catchment since ~ 60 ka were estimated to be 0.31 ± 0.06 , 0.33 ± 0.02 , and 0.24 ± 0.02 mm/a, respectively. The comparison of our data with previously published results shows that the uplift rate of $1.5\text{--}2$ mm/a of the northern Qilian fault is significantly higher than the uplift rate of $0.3\text{--}0.6$ mm/a of the Changma fault within the Qilian Shan since the Late Quaternary, which matches the terrain relief in the Qilian Shan and demonstrates that the uplift rate of the northern Qilian Shan is higher than that of the central Qilian Shan. The slip rates of active faults in the western Qilian Shan reveal that the crustal shortening across the Qilian Shan may be widely distributed in the NWW-trending compressional orogenic belts and the northern Qilian Shan plays the most important role in absorbing crustal shortening; however, the rapid sinistral strike-slip of the Altyn Tagh fault is largely accommodated by strike-slip faults within the Qilian Shan.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding authors.

AUTHOR CONTRIBUTIONS

ZC is responsible for making the research plan, field work, sampling, writing the manuscript, and drawing the figures. WX is responsible for writing and revising the manuscript. RL is responsible for field work, sampling and dating, and drawing the figures. AL is responsible for the research plan, field work, and professional guidance. NK is responsible for professional guidance.

FUNDING

This research was financially supported by the National Key Research and Development Program of China (grant No. 2018YFC1504201), the National Natural Science Foundation of China (grant No. 41402185), the National Natural Science Foundation of China (grant No. 42174023), the National Key R&D Program of China (grant No. 2019YFC1509205), and China Scholarship Council (CSC).

ACKNOWLEDGMENTS

We are grateful to the editor, Wenjun Zheng, and three reviewers for providing valuable comments and suggestions on the manuscript. We thank Geospatial Data Cloud (<http://www.gsccloud.cn/>) and USGS (<https://earthexplorer.usgs.gov/>) for providing the high-resolution DEM data and satellite images for this study.

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Conflict of Interest: Author RL was employed by Northwest Electric Power Design Institute Co. Ltd.

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Co-Seismic Surface Ruptures of the CE 1738 *M* 7.6 Dangjiang Earthquake Along the NW Continuation of the Xianshuihe Fault Zone and Tectonic Implications for the Central Tibetan Plateau

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OPEN ACCESS

Edited by:

Daoyang Yuan,
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 08 November 2021

Accepted: 24 January 2022

Published: 15 February 2022

Citation:

Yu Z, Yin N, Xiao P and Chen B (2022)
Co-Seismic Surface Ruptures of the
CE 1738 *M* 7.6 Dangjiang Earthquake
Along the NW Continuation of the
Xianshuihe Fault Zone and Tectonic
Implications for the Central
Tibetan Plateau.
Front. Earth Sci. 10:810891.
doi: 10.3389/feart.2022.810891

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The role of large-scale strike-slip faults in high-elevation areas in absorbing the strain resulting from plate convergence has yet to be scientifically understood. The Dang Jiang Fault (DJF), as the NW continuation of the Xianshuihe Fault Zone (XFZ) in the central Qinghai-Tibetan Plateau, may provide an excellent testing ground for this question, given its high slip rate, sparse vegetative cover, minimal modification, and possible relationship with the CE 1738 Dangjiang destructive earthquake. However, co-seismic surface ruptures and seismotectonics remain in dispute because of inconvenient transportation and lack of oxygen at high altitudes. Thus, field investigations are conducted here to determine co-seismic surface ruptures. The newly synthesized data from geologic observations, historical record reviews, geomorphic mapping, trench logging, and sample dating indicate that the CE 1738 Dangjiang earthquake produced an ~100 km-long surface rupture that includes offsets of gullies, linear scarps and troughs, sag ponds, *en echelon* fractures, and pressure ridges. The magnitude is re-estimated as *M* 7.6, with average and maximum strike-slip displacements of $\sim 2.1 \pm 0.1$ m and $\sim 3.5 \pm 0.1$ m, respectively. The DJF has undergone multiple seismic faulting events, and the linear fitting surface displacement rate in the Holocene is $\sim 6.3 \pm 1.9$ mm/yr with a 95% confidence interval. This study implies that the seismic hazard of the DJF cannot be underestimated given that its elapsed time is close to or beyond the recurrence interval of major earthquakes and that the oblique convergence of the Qiangtang Block might be accommodated by the clockwise rotation of the block through repeated left-lateral strike-slip movements along the southern boundary of the Bayan Har Block.

Keywords: co-seismic surface rupture, CE 1738 dangjiang earthquake, central tibetan plateau, repeated left-lateral strike-slip movement, seismic hazard

INTRODUCTION

The large-scale strike-slip faults in continental interiors have played important roles in the process of the adjustment and absorption of tectonic deformation that has resulted from the collision between the Indian plate and the Eurasian plate since 45 ± 5 Ma (Molnar and Tapponnier, 1975; Rowley, 1996; Tapponnier et al., 2001; Lin et al., 2002; Yin, 2010), although their geodynamic processes have yet to be thoroughly understood. The ~1,400 km-long Xianshuihe Fault Zone (XFZ), extending from southern Yunnan Province northwest through Sichuan into Qinghai in the central Tibetan Plateau, is a typical example among these great strike-slip faults and is probably currently the most active strike-slip fault system within China (Allen et al., 1991; Deng et al., 2002; Deng et al., 2007; Zhang et al., 2003) (**Figure 1**). In recent years, the XFZ has released strain energy with an average left-lateral slip rate of ~10 mm/yr (Wen et al., 2003) and more than 14 destructive earthquakes ($M \geq 7$) (**Figure 1**). Although many achievements in the middle and eastern sections of the XFZ have been attained by geologists around the world, the active tectonics and seismogenic

behavior, recurrence interval of great earthquakes, slip rate, and maximum magnitude estimation along the western portion of the XFZ remain debated (Wang et al., 2007). In fact, these basic data are crucial for assessing the seismic hazard and activity habit along the whole structure, as well as for deciphering the strain distribution and tectonic deformation model in the central Tibetan Plateau.

The Dang Jiang Fault (DJF, also called the Dangjiang section of the Ganzi-Yushu Fault Zone), which is the westernmost continuation of the XFZ in the central Qinghai-Tibetan Plateau, provides a unique natural laboratory for the acquisition of these data. Historical accounts and paleoseismic studies show that more than seven great earthquakes have occurred along the western portion of the XFZ, all with distinctive co-seismic surface ruptures. These studies also suggest that co-seismic rupture lengths at the surface vary from ~30 to 180 km and that the average and maximum co-seismic left-lateral displacements are approximately 1.5–2.5 and 5 m, respectively (Zhou et al., 1996; Zhou et al., 1997; Wen et al., 2003; Chen et al., 2010; Lin et al., 2011). In particular, the CE 1738 Dangjiang earthquake, despite matching the area crossed by the

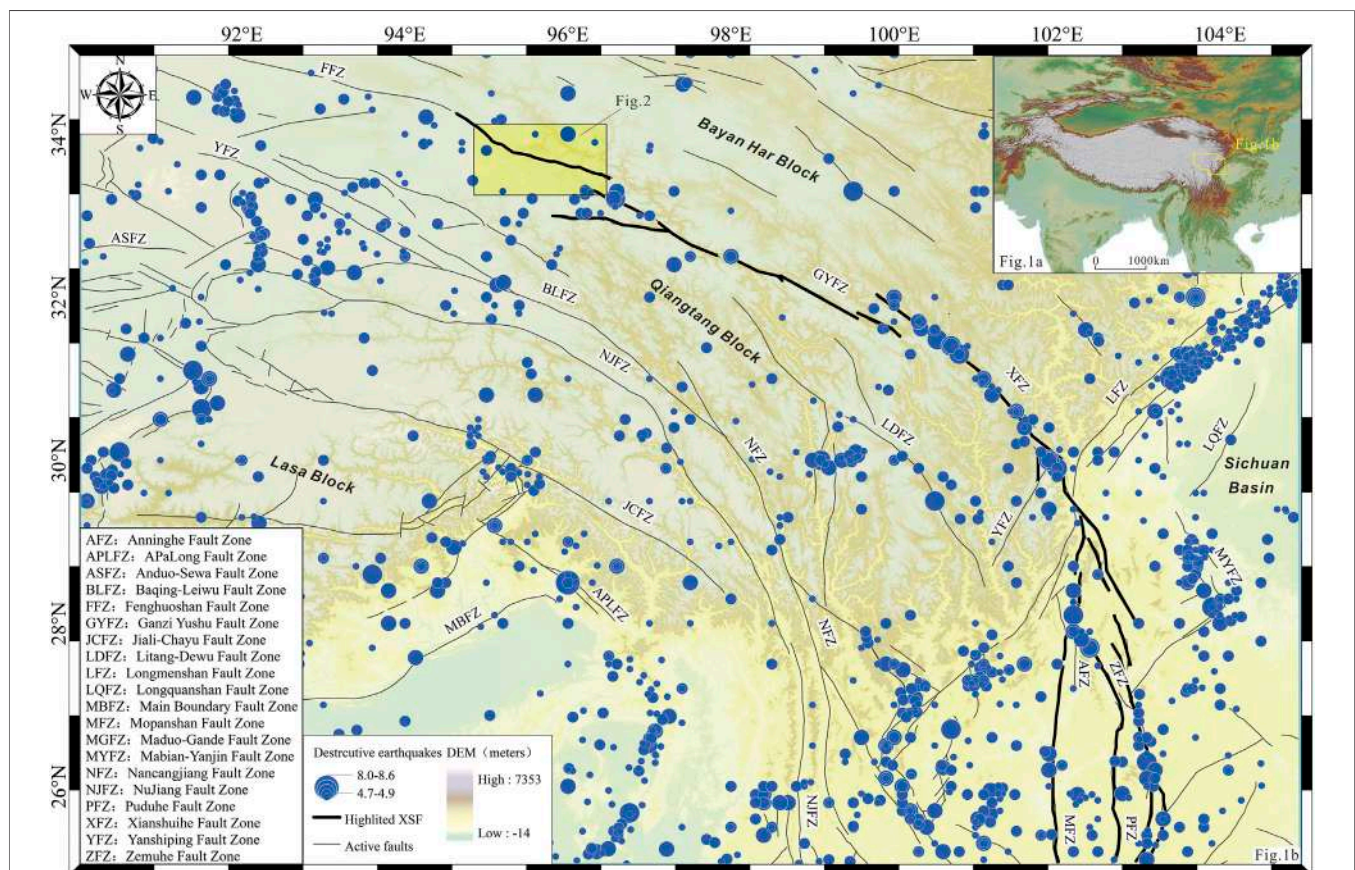


FIGURE 1 | Sketch map of earthquake distribution and geological background of the study area. **(A)** Topographic relief of the Tibet Plateau. **(B)** Seismotectonic map of the Xianshuihe Fault Zone (XFZ) and the surrounding area. Black bold lines in the map represent the XFZ structure. Blue solid circles represent the destructive earthquake epicenters ($M \geq 4.7$, epicenters are from the China Earthquake Networks Center, 2020). Red solid lines represent the main active fault traces (Deng et al., 2007). The light yellow rectangle on the north side of the image represents the scope of the study region. The inset at the lower left shows the abbreviation of regional active faults illustrated in the map.

TABLE 1 | Measured ages of samples from trenches and outcrops along the DJF.

No. in the field	No. in laboratory	Equivalent dose(Gy)	Dose rate (Gy/ka)	Age (ka BP)
DJFZOSL-01	FZ-22	3.23 ± 0.32	3.62 ± 0.12	1.10 ± 0.09
DJFZOSL-02	FZ-14	30.75 ± 1.86	3.62 ± 0.10	10.47 ± 0.58
DJFZOSL-07	FZ-31	0.93 ± 0.22	3.60 ± 0.10	0.32 ± 0.06
DJFZOSL-10	FZ-15	1.17 ± 0.39	4.77 ± 0.10	0.40 ± 0.08
DJFZOSL-11	FZ-12	7.04 ± 2.13	2.99 ± 0.13	2.35 ± 0.71
DJFZOSL-12	FZ-07	9.74 ± 0.55	3.99 ± 0.10	2.44 ± 0.14
DJFZOSL-20	FZ-04	1.04 ± 0.25	4.01 ± 0.10	0.26 ± 0.06
DJFZOSL-22	FZ-28	0.97 ± 0.20	3.84 ± 0.10	0.33 ± 0.05
DJFZOSL-29	FZ-09	52.88 ± 3.21	4.78 ± 0.15	11.18 ± 1.07
DJFZOSL-30	FZ-19	2.00 ± 0.36	3.63 ± 0.10	0.68 ± 0.10
DJFZOSL-41	FZ-27	14.23 ± 0.63	4.62 ± 0.17	3.08 ± 0.50
DJFZOSL-42	FZ-29	12.24 ± 1.10	3.54 ± 0.15	3.46 ± 0.45
DJFZOSL-50	FZ-47	1.53 ± 0.07	3.35 ± 0.13	0.46 ± 0.03
DJFZOSL-51	FZ-48	13.95 ± 0.48	3.71 ± 0.16	3.76 ± 0.20
No. in the field	Analyzed Material	IRMS $\delta^{13}C$	Conventional Age	2 Sigma calibration age
DJFZ- ¹⁴ C-04	Organic sediment	−25.5‰	770 ± 30 BP	Cal. BP 729 to 668
DJFZ- ¹⁴ C-06	Organic sediment	−24.3‰	3,380 ± 30 BP	Cal. BP 3696 to 3,650
DJFZ- ¹⁴ C-10	Charred material	−22.0‰	1,270 ± 30 BP	Cal. BP 1287 to 1,175

DJF, has some still unverified issues: the characteristics of co-seismic surface deformation and the possible associated maximum magnitude. Both of these factors directly influence our understanding of the geodynamic process of strain adjustment and active tectonics of the whole XFZ, as well as the assessment of the earthquake potential along the structure and in the neighboring area of the central region of the Qinghai-Tibetan Plateau.

Thus, the main purposes of this study are to investigate the co-seismic surface ruptures of the 1738 Dangi Jiang earthquake, to study the active behavior of the DJF in the late Quaternary, and to discuss its tectonic implications for the seismic hazard and strain partition along the southern boundary of the Bayan Har Block in the central Tibetan Plateau area.

TECTONIC SETTING

Broadly speaking, the XFZ (in the broad sense) is a well-known NW-SE trending seismogenic structure within the Tibetan Plateau, and more than 60 earthquakes with magnitudes greater than 6.0 have occurred since CE 1500 because of repeated left-lateral slippage along the structure (Allen et al., 1991; Lin et al., 2002; Wen et al., 2003; Wang et al., 2007; Wen et al., 2008; Li et al., 2012) (**Figure 1**). The structure is composed of many subsections from southeast to northwest, including the Zemuhe Fault Zone (ZFZ), the Anninghe Fault Zone (AFZ), the Xianshuihe Fault Zone (XFZ, in the narrow sense), and the Ganzi-Yushu Fault Zone (GYFZ) (**Figure 1**). Previous studies on seismology (Wang and Burchfiel, 2000), thermochronology (Arne et al., 1997; Xu and Kamp., 2000), geodesy (King et al., 1997), and granites or volcanic rocks (Roger et al., 1995; Wang et al., 1998) have shown that their formation times, evolutionary processes, and deformation features are not exactly the same.

Our study area is located in a high mountain region along the northwestern section of the large-scale left-lateral strike-slip GYFZ, which is viewed as the northwestern continuation of the XFZ, at an

average elevation of ~5,300 m in the central Tibetan Plateau (**Figure 1**). The GYFZ strikes NW-SE with a length greater than ~550 km and forms the southern tectonic boundary of the Bayan Har Block with other branches of the XFZ (**Figure 1**). The GYFZ overlaps the NW end of the XFZ with a left *en echelon* stepover of approximately 40 km (Allen et al., 1991). Field geological observations and GPS velocity inversion results reveal that the average left-lateral strike-slip rate of the GYFZ in the Holocene is 4–13 mm/yr (Wang et al., 2001; Lin et al., 2002; Peng et al., 2006; Wang et al., 2007). The basement in this study area consists mainly of Triassic limestone, sandstone, mudstone, and shale with volcanic rocks (Lin et al., 2011), while the Cenozoic strata are dominated by Paleogene purple coarse-grained sediments, Neogene purple fine-grained lacustrine sediments, and Quaternary unconsolidated deposits. Previous seismological studies have revealed that modern and historical destructive earthquakes, together with paleoearthquakes along the faults, have ruptured these branches of the GYFZ with variable co-seismic surface rupture sizes and distinctive geomorphic features (Zhou et al., 1997; Lin et al., 2011; Li et al., 2012).

METHOD AND DATA

In this study, high-resolution satellite images and digital topographic data are first used to interpret the morphology of the DJF. These data mainly include web-based Google Earth images and 30 m resolution SRTM (Shuttle Radar Topography Mission) data, both of which are used to digitally map individual structures and tectonic features using ESRI ArcMap software. Then, we investigated the Holocene activity of the DJF through tectonic geomorphic surveys, trench excavation, and log and laboratory chronology tests. Finally, we analyzed the co-seismic surface ruptures of the CE 1738 *M* 7.6 Dangi Jiang earthquake and discussed its slip rate in the Holocene and the implications for earthquake hazards along the whole XFZ, as well as for tectonic deformation characteristics and possible strain-distribution processes in the central Tibetan Plateau.

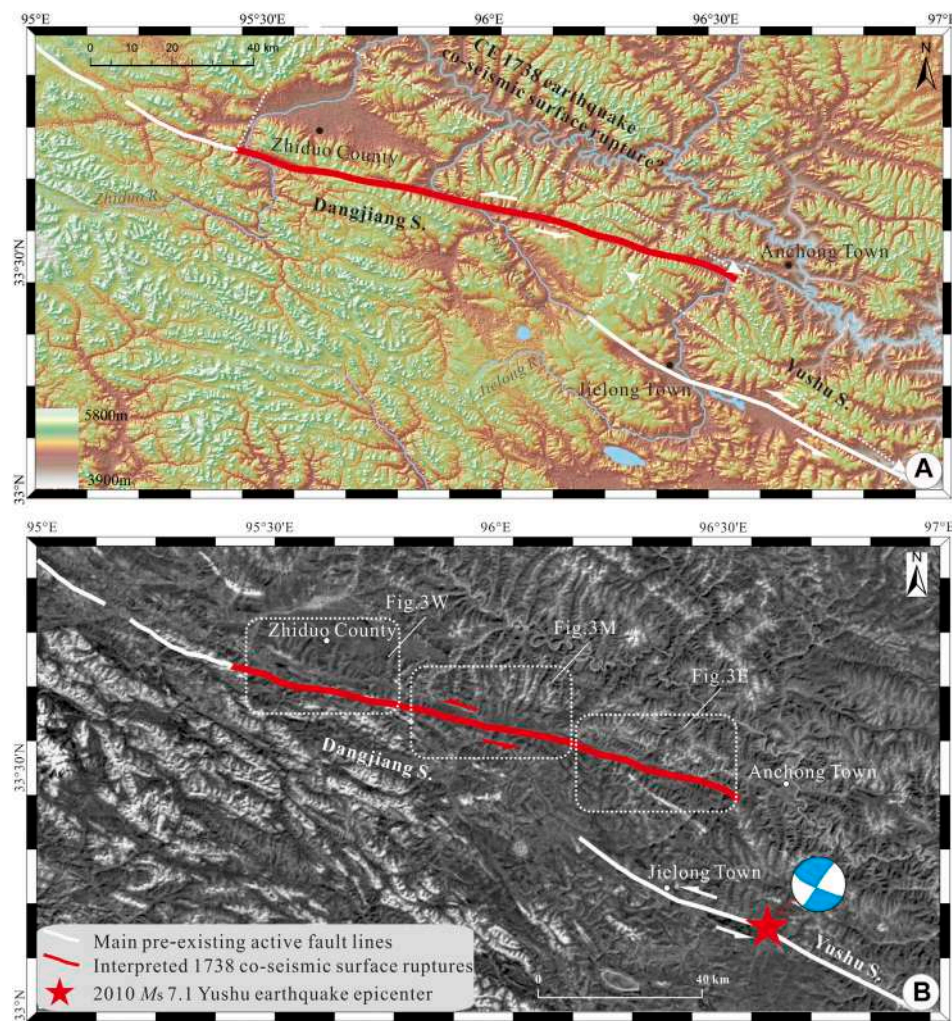


FIGURE 2 | Satellite images showing the tectonic topographies of the NW continuation of the Xianshuihe Fault Zone (XFZ). **(A)** Color relief map showing that the Dangjiang Fault Zone (DJF) and the Yushu Fault Zone (YFZ) are distributed in a right step oblique pattern in the plane and that the DJF is developed along a linear tectonic valley. The inset legend at the lower left shows the range of topographic relief height variation. **(B)** Google image showing three key substudy areas along the DJF. The red line represents the co-seismic surface rupture zone of the CE 1738 Dangjiang historical earthquake (Zhou, et al., 1997; Lin et al., 2011), while the red star represents the epicenter of the 2010 *M*_s 7.1 Yushu earthquake (modified from Lin et al., 2011; Li et al., 2012). The beach ball at the bottom right represents the focal-mechanism solution (Li et al., 2012), which indicates that the 2010 *M*_s 7.1 Yushu earthquake is mainly sinistral shear rupture.

Geomorphic markers, such as fault scarps, stream channels, gullies, small ditches, and river terraces, were mapped and integrated by trench excavation and field surveys. Trench walls were cleaned and logged in detail and then interpreted in the field and laboratory. Offsets were jointly measured by a laser rangefinder and a steel tape. In particular, UAV (unmanned aerial vehicle, Phantom four Manufactured by SZ DJI Technology Co., Ltd.) aerial survey technology is applied in this field survey, which is the first to obtain high-precision displacement-distribution data and tectonic-deformation images in this structure thus far.

Two kinds of dating methods are used in this research: optically stimulated luminescence (OSL) and ¹⁴C, and a total of seventeen valid dating results were used in this paper (Table 1). The OSL samples were jointly measured by LinYi University, Institute of Geology of China Earthquake Administration, the

Institute of Disaster Prevention in China, and the sensitivity-corrected multiple aliquot regenerative-dose (SMAR) method was adopted. All tests were conducted using a Daybreak 2,200 automated OSL reader equipped with a combined blue (470 ± 5 nm) and infrared (880 ± 80 nm) LED OSL unit and an Am-241 alpha source ($0.062 \mu\text{m}^{-2} \text{s}^{-1}$) for irradiation (Yu et al., 2018). ¹⁴C dating was completed at the Beta Analytic Laboratory (United States) with four in-house NEC accelerator mass spectrometers and four Thermo IRMSs. Conventional radiocarbon ages were calculated using the Libby half-life (5,568 years) corrected for the total isotopic fraction and were used for calendar calibration where applicable.

Unless otherwise specified, all earthquake records and active faults were collected from the China Earthquake Networks Center (2019), Gu, (1983), the Editorial Board of Annals of Sichuan

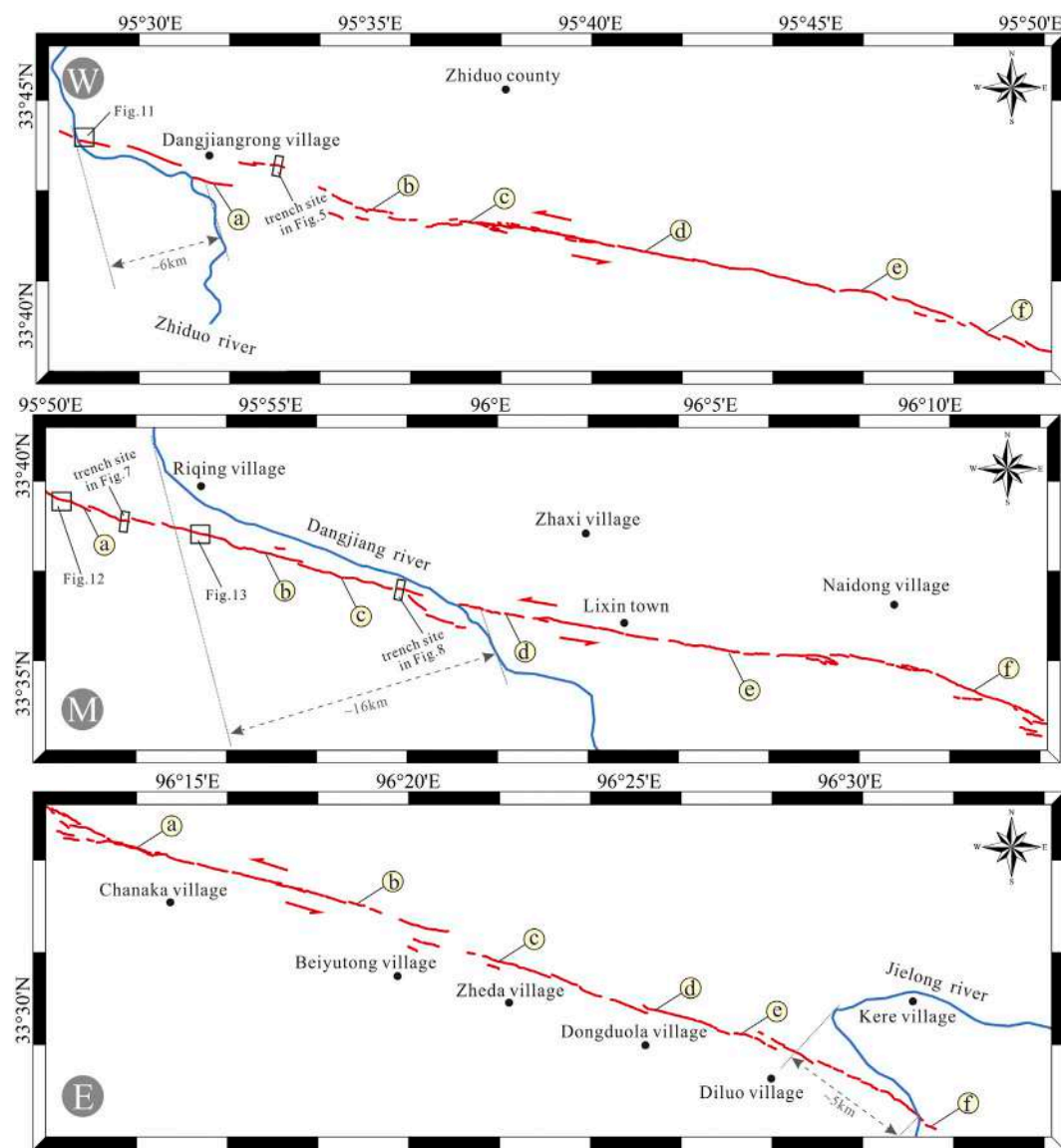


FIGURE 3 | Fine fault geometries interpreted based on high-resolution satellite images and field observations. The red lines represent the co-seismic surface rupture traces of the CE 1738 Dangjiang historical earthquake. The blue solid lines represent the regional deflected rivers. The capital letters of W, M, and E in the map represent the western, middle, and eastern sections of the DJF, respectively, and the marks from (a, b, c–f) to correspond to the locations of the photographs in **Figure 4**, **Figure 6**, and **Figure 9**, respectively.

Province (1998), and Deng et al. (2007), and previous research data also provide appropriate references in this study (Zhou et al., 1996; Zhou et al., 1997; Wen et al., 2003; Chen et al., 2010; Lin et al., 2011; Li et al., 2012; Li et al., 2016; Wu et al., 2017).

RESULTS

Field investigation reveals that the strike of the DJF is oriented more east–west than other branches of the XFZ and can be traced for ~170 km in length (**Figure 1** and **Figure 2**). The DJF ends to the east near Anchong town (**Figure 2**), and an ~20 km wide

uplift area can be observed near Jielong town because the Yushu fault overlaps the southeastern end of the DJF with a right *en echelon* stepover of approximately 25 km. Previous studies show that this uplift might have hampered the surface rupture propagation of the *M*_s 7.1 Yushu earthquake on 14 april 2010 (Lin et al., 2011) (**Figure 2**). At its northwestern extremity, the DJF connects with the FFZ (Fenghuoshan Fault Zone) and then extends into the complex and little-studied Hoh Xil region (Allen et al., 1991) (**Figure 1**).

The historical earthquake that occurred in Yushu near Qinghai Province in CE 1738 caused ~157 family members to be seriously injured and approximately 500 or 261 people to die



FIGURE 4 | Diagrammatic features of the CE 1738 co-seismic surface ruptures along the western section of the DJF. Red arrows represent the distribution of the latest co-seismic characteristics, while the black arrows delineate the distribution of the older fault scarp. The azimuth angles in the maps represent the section direction. Note the scale difference in each picture. Pictures from, to correspond to the investigated sites in **Figure 3 W**. **(A)** Fault scarp, trough, and depressions developed within unconsolidated alluvial and pluvial sediments, which are duplicated in a preexisting structure zone. Note the youngest gully terrace was vertically dislocated by approximately 0.5 ± 0.1 m, while the height of the older fault scarp is of $\sim 1.6 \pm 0.2$ m; **(B)** A gully was offset (by 9.4 ± 0.4 m) by a preexisting fault and co-seismic surface ruptures where a fault scarp was observed with a height of $\sim 3.0 \pm 0.2$ m at a site located ~ 1 km northwest of the gully; **(C)** Co-seismic surface ruptures in a preexisting fault depression developed along a fault trough accompanied by a geomorphic scarp with a height of 0.9 ± 0.1 m; **(D)** Fresh alternating *en echelon* fractures and pressure ridges (marked by white arrows) are observed in the same direction with the preexisting fault (delineated by red arrows); **(E)** One gully was horizontally offset (by 2.5 ± 0.1 m) by the preexisting fault, while the older gully was abandoned with accumulated dislocations $\sim 5.1 \pm 0.2$ m. Additionally, an $\sim 2.8 \pm 0.2$ m-high geomorphic scarp and depression were observed along fault; **(F)** Co-seismic surface ruptures in a preexisting graben with a breadth of $\sim 12.0 \pm 0.5$ m along the valley.

(Compiling Committee of Annals of Qinghai Province, 1998; Yuan et al., 2011), and textual research and field investigations hint that the earthquake resulted in at least ~ 50 km surface rupture along the Dangjiang section of the GYFZ (Zhou et al., 1996; Zhou et al., 1997). In this study, the co-seismic surface rupture traces of the CE 1738 Dangjiang earthquake were first mapped through integrated data, such as the interpretation of satellite images (**Figure 2** and **Figure 3**), previous studies (Zhou

et al., 1996; Zhou et al., 1997; Yuan et al., 2011), and our field observations. The strike-slip fault activity is shown by deflected gullies, dislocated ditches, linear scarps, fault troughs, shutter ridges, sag ponds, and mole tracks. Furthermore, according to their main differences in co-seismic deformation and geomorphic characteristics, the whole co-seismic rupture line can be divided into three different sections: the western segment, the middle segment, and the eastern

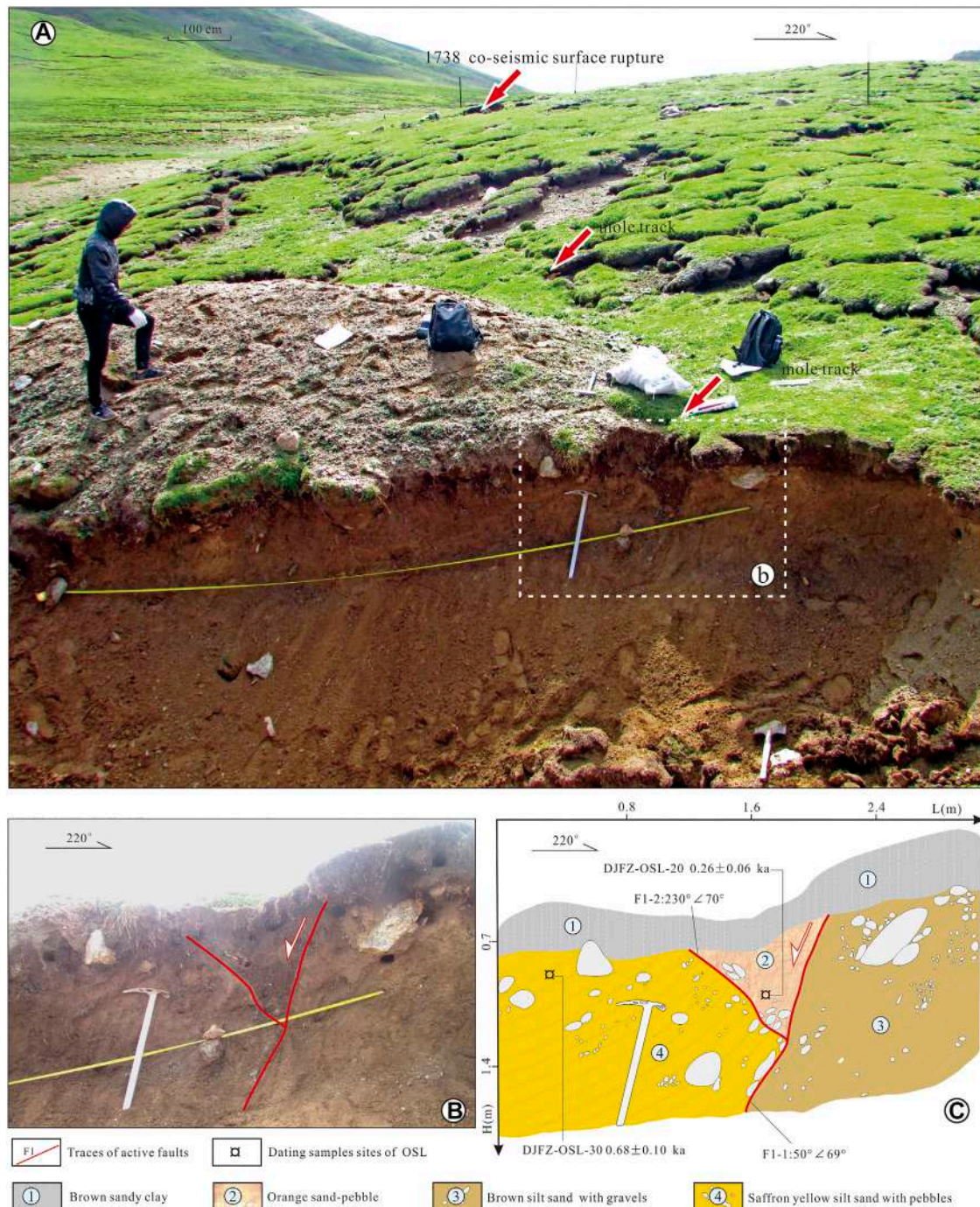


FIGURE 5 | West wall logs of the trench excavated near Dangjiangrong village, and the site of the trench is marked in **Figure 3 W**. **(A)** Tectonic geomorphic features of the DJF near Dangjiangrong village; **(B)** Original profile of the west wall (Courtesy of Li et al., 2012, unpublished results); **(C)** Interpretation map of the west wall. Fault F1 cuts the sediments of Units three to four and is overlain by Unit 1, while the wedge shape of Unit 2 may represent rapid transient accumulation after the earthquake (see the text for details). Fresh mole track structures are observed in the wall. Note the ice-axe in maps for scale.

segment (**Figure 2** and **Figure 3**). Each fault segment is made up of several strands with changing rupture lengths from dozens of meters to several kilometers, most of which are spread out in an *en echelon* pattern at the surface with an average width ranging from 5 to 35 m (**Figure 3**).

The Western Segment of the DJF

High-resolution satellite images and field observations reveal that the western segment of the DJF terminates at Dangjiangrong village at its northwestern tip with a length of ~36 km (**Figure 3 W**). The co-seismic surface rupture features produced by the CE

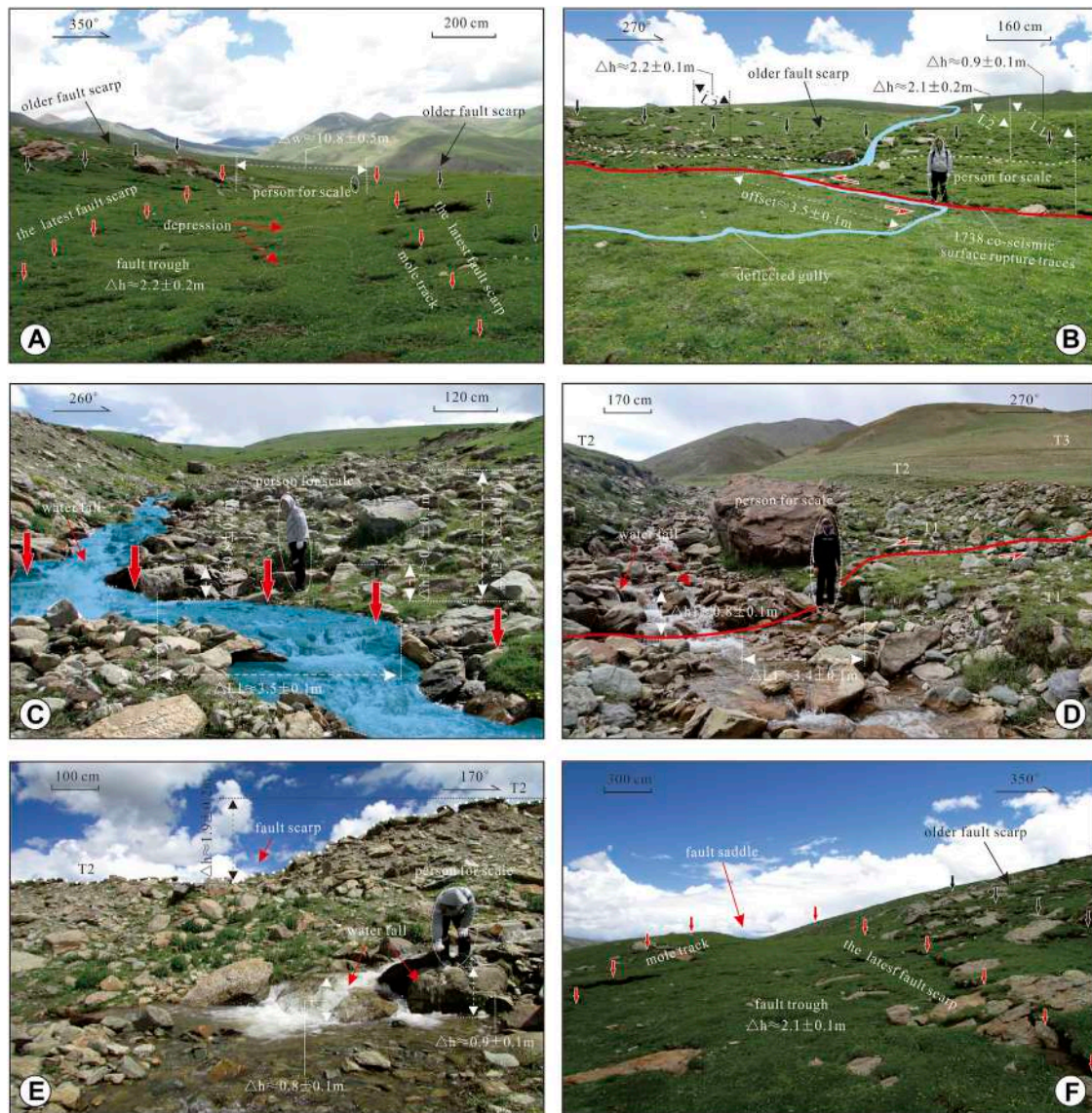
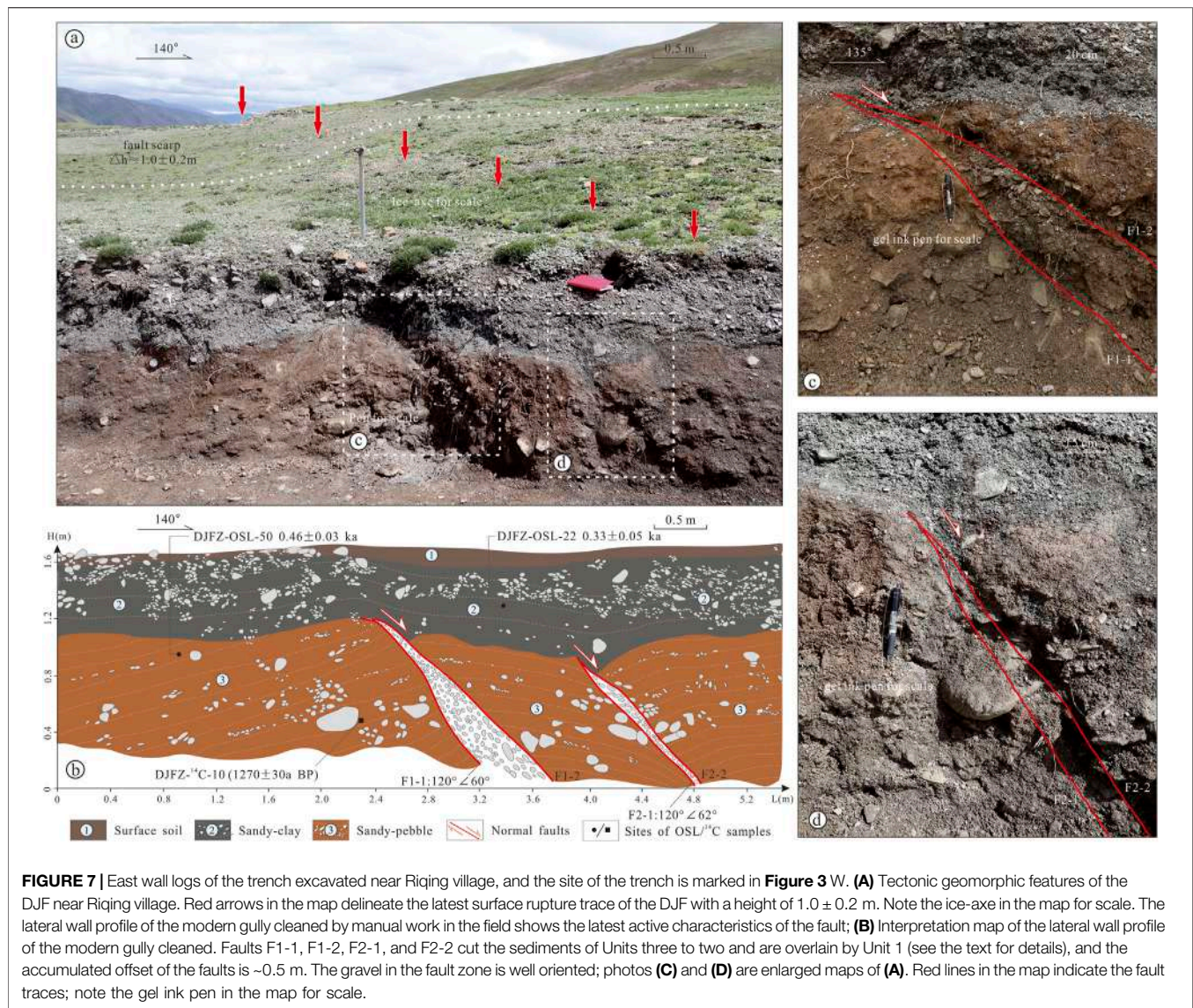


FIGURE 6 | Photographs of representative tectonic landforms of the DJF and CE 1738 co-seismic surface ruptures in the middle section of the structure. Arrows represent the distribution of tectonic characteristics such as fault scarps, mole tracks, and depressions, while the red arrows delineate the latest tectonic deformation. The azimuth angles in the photographs represent the section direction. Note the difference in each scale of the picture. Pictures from, to correspond to the locations of the photographs in **Figure 3 M**. **(A)** An $\sim 2.2 \pm 0.2$ m-high co-seismic surface scarp in a preexisting graben with a breadth of $\sim 10.8 \pm 0.5$ m along the fault valley; **(B)** fault scarp and depressions developed within unconsolidated alluvial and pluvial deposits, which are duplicated in a preexisting structure zone. Note the youngest gully terrace was vertically dislocated by approximately 0.9 ± 0.1 m, which was horizontally deflected or offset by $\sim 3.5 \pm 0.1$ m; **(C)** A modern stream was deflected $\sim 3.5 \pm 0.1$ m and vertically offset $\sim 0.8 \pm 0.1$ m (the height of the waterfall drop in the map basically represents the vertical displacement of the latest co-seismic surface rupture), while the higher geomorphologic surface that consists of boulders and gravels was vertically displaced $\sim 1.8 \pm 0.1$ m on the preexisting structure zone; **(D)** A modern valley flat was deflected $\sim 3.4 \pm 0.1$ m and vertically offset $\sim 0.8 \pm 0.1$ m (the height of the waterfall drop in the map basically stands for the vertical displacement of the latest co-seismic surface rupture), note the person in the map for scale; **(E)** A stream was vertically dislocated (the height of the waterfall drop in the map is measured at 0.9 ± 0.1 m) by a preexisting fault and co-seismic surface ruptures where a fault scarp was observed with a height of $\sim 1.9 \pm 0.2$ m at a site located ~ 3 m east of the stream; **(F)** Fresh mole tracks are observed in the same direction with the preexisting fault zone, and the height of the fault trough is measured at $\sim 2.1 \pm 0.1$ m.

1738 Dangi Jiang earthquake are investigated at six representative observation points and one trench (**Figure 4** and **Figure 5**).

Although nearly all the co-seismic surface ruptures from the CE 1738 Dangi Jiang earthquake are oriented in the same direction as the preexisting fault zone, the new fracture traces can still be clearly distinguished from the preexisting structure (**Figure 4**). A

series of fault scarps, troughs, and depressions has developed within unconsolidated alluvial and pluvial sediments along the fault, and the heights of the scarps vary from 0.5 ± 0.1 m to 3.0 ± 0.2 m (**Figure 4**). In particular, fresh alternating *en echelon* fractures and pressure ridges are observed in the field (**Figure 4D**), and the thickness of the blocks is mainly affected



by the depth of frozen ground in this region (Allen et al., 1991). Additionally, the long-term left-lateral strike-slip movement of the fault is well displayed by the repeated deflection or offset of gullies across the structure. A modern gully is offset by $\sim 9.4 \pm 0.4$ m because of repeated fault movements where a fault scarp is observed has a height of $\sim 3.0 \pm 0.2$ m at a site located ~ 1 km northwest of the gully (**Figure 4B**). Another modern gully is horizontally offset by $\sim 2.5 \pm 0.1$ m, while an older gully is abandoned with an accumulated sinistral displacement of $\sim 5.1 \pm 0.2$ m; moreover, an $\sim 2.8 \pm 0.2$ m-high scarp is observed along the fault zone (**Figure 4E**). Here, it is proposed that the $\sim 0.5 \pm 0.1$ m vertical offset and the $\sim 2.5 \pm 0.1$ m horizontal offset may be attributed to the recent faulting event that postdates the sedimentation of the unconsolidated alluvial and pluvial deposits, which very likely corresponds to the CE 1738 Dangjiang event. Meanwhile, the $\sim 9.4 \pm 0.4$ m horizontal displacement and $\sim 3.0 \pm 0.2$ m vertical offset may represent the accumulated displacements that resulted from repeated sinistral

slip along the preexisting fault zone and a recent co-seismic surface faulting event.

Furthermore, to study the co-seismic surface ruptures of the CE 1738 Dangjiang historical event and the active tectonics of the DJF in the late Quaternary, a trench was excavated and cleaned across the surface rupture traces near Dangjiangrong village (see **Figure 3 W** for location details, associate Professor Li excavated the original trench in 2010 but did not publish it, and he generously provided it to this study). Here, the west wall of the trench is relogged (**Figure 5**).

The interpreted results reveal that the sediments exposed in the walls consist of alluvial deposits of sand-pebbles and silty mudstone layers with gravels containing organic material with small flakes of charcoal (**Figure 5**). In general, four distinct stratigraphic units can be identified according to the difference in color and structure of the constituent materials. Unit 1 is the youngest formation and consists mainly of loose brown sandy clay with a thickness of approximately 20 cm. Unit 3, located on

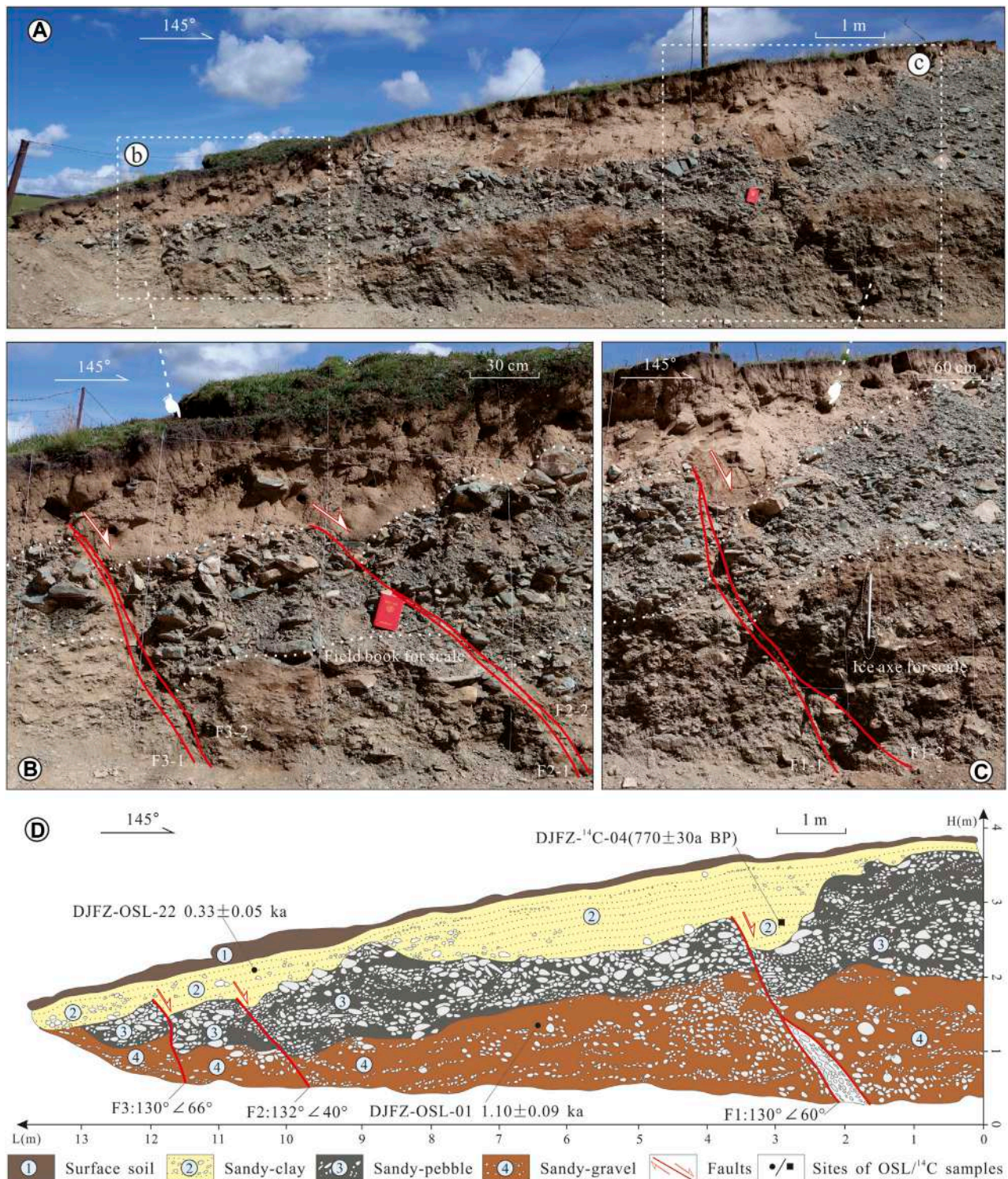


FIGURE 8 | Trench excavated near the east of Riqing village on the middle section of the DJF, and the site of the study point is marked in **Figure 3 M**. **(A)** Original east wall of the trench; photographs **(B,C)** are enlarged maps of **(A)**. Red lines in the map indicate the fault traces; note the ice axis in the map for scale; **(D)** interpretation map of the west wall. Faults F1, F2, and F3 cut the sediments of Units four to two and are overlain by Unit 1. Three chronological test results at this point hint that the latest rupture should occur in the CE 1738 Dangjiang historical earthquake with an accumulated co-seismic offset of ~0.9 m (see the text for details).

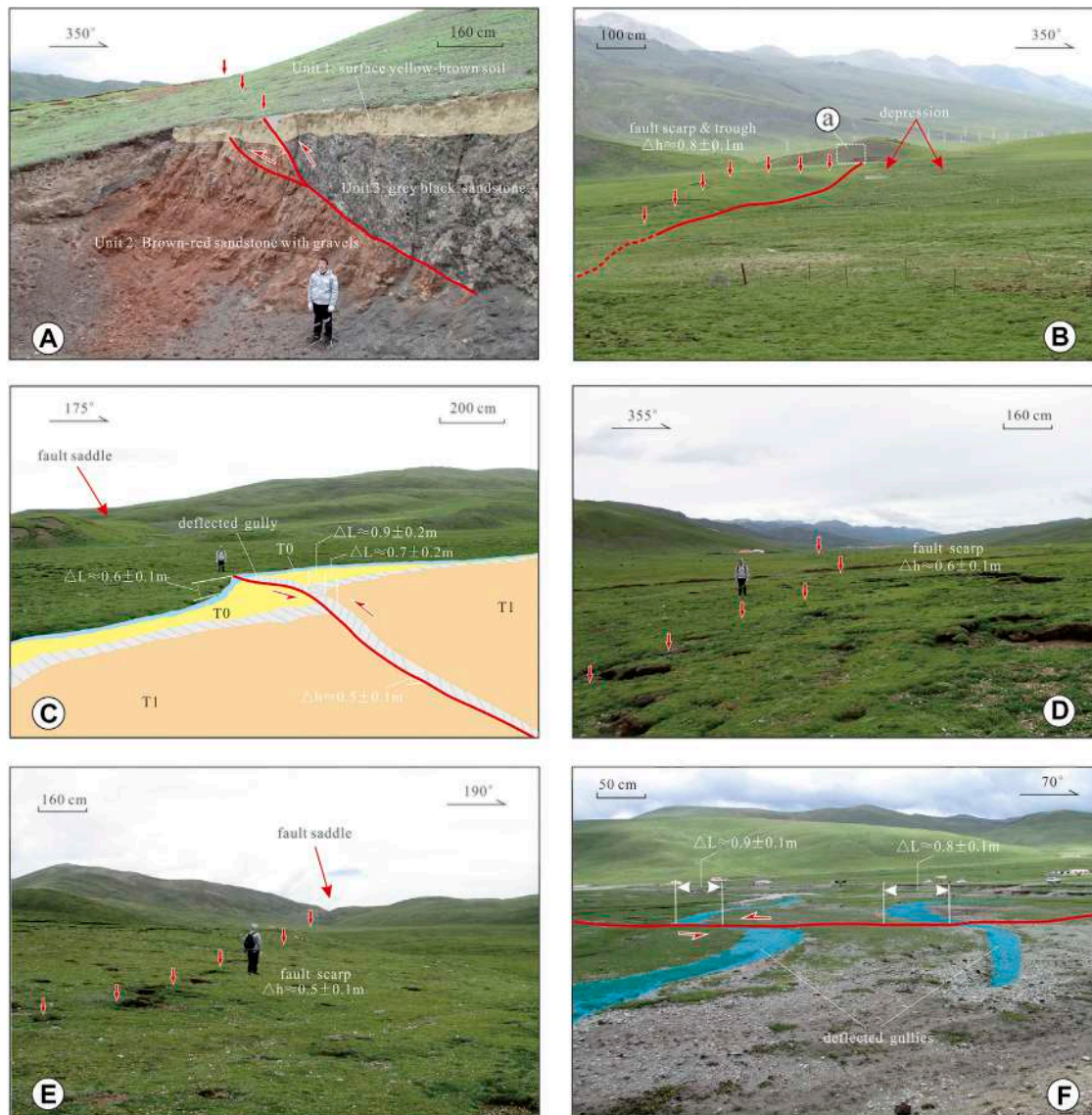


FIGURE 9 | Photographs of representative tectonic landforms of the DJF and CE 1738 co-seismic surface ruptures in the eastern section. Red arrows represent the distribution of co-seismic tectonic characteristics such as fault scarps, mole tracks, and depressions. The azimuth angles in the photographs represent the section direction. Note the difference in each scale. Pictures from, to correspond to the locations of the photographs in **Figure 3 E**. **(A)** Paleogene brown-red sandstone with gravels in fault contact with Triassic gray black sandstone, and the latest activity of the fault has ruptured the surface modern yellow-brown soil in a thrust way. Note the person in the map for scale; **(B)** fault scarp, trough and depressions developed within unconsolidated alluvial and pluvial sediments, which are duplicated in a preexisting structure zone. Note that the height of the scarp is -0.8 ± 0.1 m; **(C)** a gully was deflected (by 0.7 ± 0.2 m) by co-seismic surface ruptures where a fault scarp was observed with a height of -0.5 ± 0.1 m on average; **(D,E)** fault scarp developed within unconsolidated alluvial deposits, and the scarp height is -0.5 ± 0.1 m. Fresh mole tracks can be observed in the same direction as the fault zone; **(F)** two modern gullies were deflected with average horizontal displacements of -0.8 ± 0.1 m.

the right side of the section, is composed mainly of brown silt sand with gravels. Unit 4 is distributed on the left side of the section and mainly consists of orange medium and fine sand mixed with gray pebbles. Both the occurrence of Unit 3 and Unit 4 reveal that they have recently undergone intense tectonic deformation. Unit 2 is mainly composed of disordered orange sand-pebbles, whose boundaries are controlled by Unit 3 to the right and Unit 4 to the left and overlaid by Unit 1 (**Figure 5**).

According to the formation occurrences and cross-cutting relationships between the stratigraphic units mentioned above, the fault traces are sketched and recognized (**Figure 5**). Both faults are perpendicular to the general trend of the CE 1738 Dangjiang co-seismic surface rupture zone. The main fault F1-1 cuts the sediments of Units 3–4 and is overlain by undeformed Unit 1. Fault F1-2, as the branch of main fault 1-1, has the opposite inclination. Wedge-shaped Unit 2 in **Figure 5** should represent rapid transient accumulation shortly after the

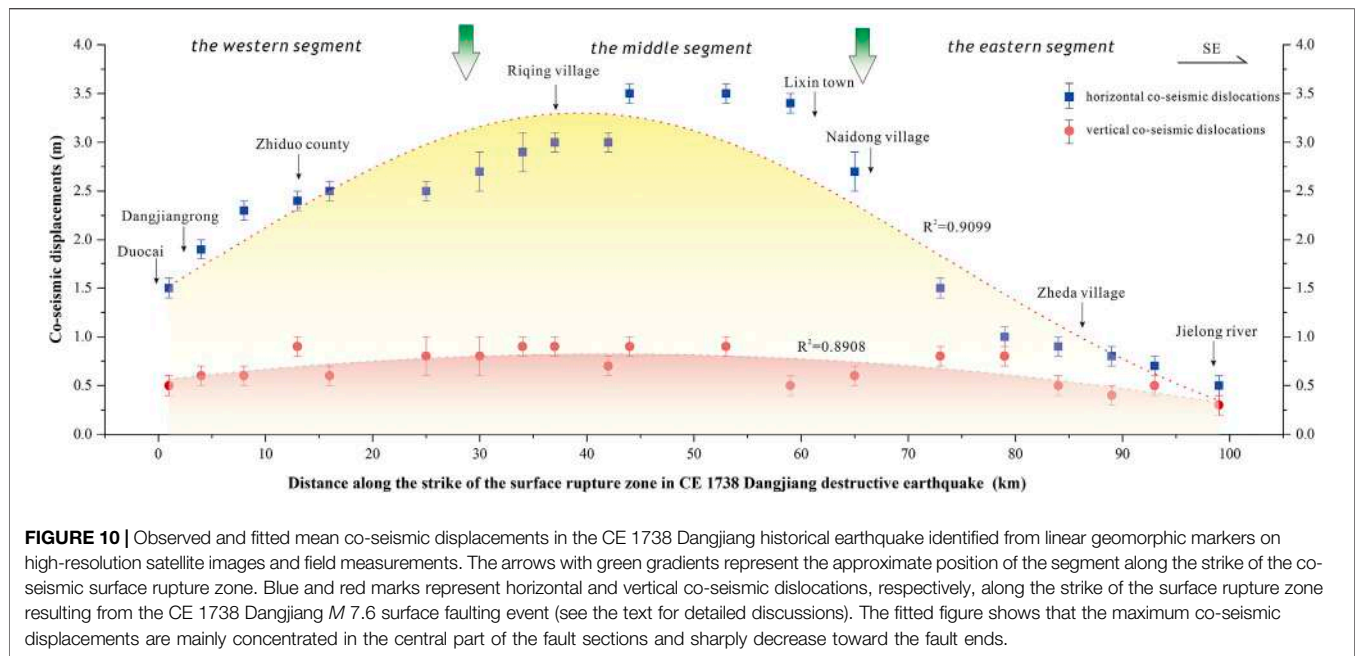


FIGURE 10 | Observed and fitted mean co-seismic displacements in the CE 1738 Dangjiang historical earthquake identified from linear geomorphic markers on high-resolution satellite images and field measurements. The arrows with green gradients represent the approximate position of the segment along the strike of the co-seismic surface rupture zone. Blue and red marks represent horizontal and vertical co-seismic displacements, respectively, along the strike of the surface rupture zone resulting from the CE 1738 Dangjiang M 7.6 surface faulting event (see the text for detailed discussions). The fitted figure shows that the maximum co-seismic displacements are mainly concentrated in the central part of the fault sections and sharply decrease toward the fault ends.

earthquake if the unique colluvial wedge structure and a stacking pattern without stable sedimentary sequences in the units are taken into consideration.

Here, it is proposed that the latest deformation should have occurred after the sedimentation of Units 3–4 and before the deposition of Unit 1. The age of Unit 2 may approximately represent the age of the latest activity of the DJF. The results of laboratory chronological dating further support this suggestion. The ages of samples from Unit 4 imply that the latest rupture of fault F1-1 should have occurred later than 0.68 ± 0.10 ka BP (DJFZ-OSL-30) (Figure 5). The wedge-shaped Unit 2 is represented by sample DJFZ-OSL-20 (0.26 ± 0.06 ka BP), which hints that the latest surface rupture faulting event should correspond to the CE 1738 great Dangjiang earthquake.

In summary, the latest activity of the western segment along the DJF occurred after 0.68 ± 0.10 ka BP and was closer to 0.26 ± 0.06 ka BP, which most likely corresponds to the CE 1738 great historical earthquake. Field observations also reveal that there is little correspondence between the two sides of the fault zone, which implies that the kinematic properties of the fault are dominated by left-lateral motion. This observation is in good agreement with the above results of geomorphic investigations along the whole sections of the DJF (Figure 4).

The Middle Segment of the DJF

The middle ruptured segment of the DJF starts from Riqing village at its northwestern tip and terminates near Naidong village at its southeastern tip with a total length of ~ 35 km (Figure 3 M). The co-seismic surface ruptures are revealed at eleven representative investigation sites, the locations of which are indicated in Figure 3 M.

The middle segment is highlighted by a linear fault trough, mole or extensional tracks, waterfalls, deflected gullies, offset ditches, and fault scarps that developed within unconsolidated

alluvial and pluvial deposits, which duplicate a preexisting structural zone (Figure 6). Notably, the co-seismic displacements, both horizontal and vertical, along the middle segment are larger than those along the eastern and western segments.

Field geomorphologic observations in this segment show that the Quaternary unconsolidated sediments in the alluvial-proluvial fans and river terraces are faulted and duplicate the preexisting structural zone (Figure 6A). Three modern gullies are vertically displaced by approximately 0.9 ± 0.1 m but are horizontally deflected or offset by $\sim 3.5 \pm 0.1$ m (Figures 6B–D). The height of the waterfall drops on the map is measured at 0.9 ± 0.1 m, which may represent the co-seismic vertical offset in the latest faulting event (Figures 6C–E). In addition, a series of fresh mole tracks is observed along the fault zone, and the average height of the fault troughs is measured at $\sim 2.1 \pm 0.1$ m (Figure 6F). Although the lack of datable material makes it difficult to precisely determine the time of the last seismic event, the geomorphic features in Figure 6 may indicate that the fault slipped at the surface recently, which most likely resulted from the CE 1738 Dangjiang historical event.

During this segment, two trenches are excavated in the field to define the co-seismic deformation characteristics of the CE 1738 Dangjiang great event and the active tectonics of the DJF (Figures 7, 8), and the sites of the trenches are marked in Figure 3 M.

One trench, located in the southwest corner of Riqing Village (Figure 3 M), is perpendicular to the general trend of the CE 1738 Dangjiang co-seismic surface rupture traces (Figure 7A). Field geomorphologic observations near this site show that the Quaternary unconsolidated sediments in the alluvial-proluvial fans and river terraces are faulted and duplicate the preexisting structural zone (Figure 7A). The youngest geomorphic surface is vertically offset by 1.0 ± 0.2 m, and this offset may correspond to the CE 1738 Dangjiang great earthquake (Figure 7A). The lateral

TABLE 2 | Distribution of co-seismic displacements in CE 1738 Dangjiang earthquake.

No	Latitude	Longitude	L(m)	Marker	Brief description	References
1	33.701389	95.608420	1.5 ± 0.1	Gully	The gully on the lower terrace (T1) is sinistral offset of ~1.5 m	This research
2	33.701539	95.624868	1.9 ± 0.1	Small ditch	The modern ditch is deflected of ~1.9 m	This research
3	33.700460	95.634369	2.3 ± 0.1	Small ditch	The modern ditch is deflected of ~2.3 m	This research
4	33.699114	95.644469	2.4 ± 0.1	Small ditch	Small ditches on the piedmont alluvial fan are synchronized displaced of ~2.4 m	This research
5	33.698314	95.650614	2.5 ± 0.1	Small ditch		This research
6	33.697348	95.655285	2.5 ± 0.1	Small ditch	The ditch is deflected of ~2.5 m while the older gully is abandoned (Figure 4E)	This research
7	33.695716	95.662287	2.7 ± 0.2	Gully	Gully on the lower terrace (T1) is sinistral offset of ~2.7 m	This research
8	33.688691	95.698120	2.9 ± 0.2	Small ditch	The modern ditch is deflected of ~2.9 m	This research
9	33.682455	95.737597	3.0 ± 0.1	Small ditch	The modern ditch is deflected of ~2.9 m	This research
10	33.660987	95.818018	3.0 ± 0.1	Stone ridge	The stone ridges on the lower terrace (T1) are sinistral offset of ~3.0 m	This research
11	33.650396	95.847152	3.5 ± 0.1	Small ditch	The modern ditch is deflected of ~3.5 m (Figure 6B)	This research
12	33.648525	95.852919	3.5 ± 0.1	Gully	Gully on the lower terrace (T1) is sinistral offset of ~3.5 m (Figure 6C)	This research
13	33.643550	95.875847	3.4 ± 0.1	Gully	Gully on the lower terrace (T1) is sinistral offset of ~3.4 m (Figure 6D)	This research
14	33.626890	95.931588	2.7 ± 0.2	Small ditch	The modern ditch is deflected of ~2.5 m	This research
15	33.605382	96.034834	1.5 ± 0.1	Small ditch	Small ditches on the piedmont alluvial fan are synchronized sinistral offset	This research
16	33.596249	96.087238	1.0 ± 0.1	Small ditch	The modern ditch is deflected of ~1.0 m	This research
17	33.595581	96.103014	0.9 ± 0.1	Small ditch	The modern ditch is deflected of ~0.9 m (Figure 9C)	This research
18	33.594679	96.108459	0.8 ± 0.1	Gully	Gullies on the lower terrace (T1) are sinistral deflected of ~0.8 m (Figure 9F)	This research
19	33.593837	96.127310	0.7 ± 0.1	Small ditch	The modern ditch is deflected of ~0.7 m	This research
20	33.587972	96.156238	0.5 ± 0.1	Small ditch	The modern ditch is deflected of ~0.5 m	This research
21	33.525030	96.356975	3.3	Small ditch	Small ditches on the piedmont alluvial fan are synchronized sinistral offset	Wu et al. (2017)
22	33.525030	96.356975	4.5	Small ditch		
23	33.525030	96.356975	3.1	Small ditch		
24	33.529896	96.333975	4.0	Gully	Gully depth of 0.5 m	Wu et al. (2017)
25	33.548903	96.277825	4.8	Small ditch		Wu et al. (2017)
26	33.556222	96.246250	4.8	Gully	Gully is nearly perpendicular to the fault scarp which is about 1.5 m in height	Wu et al. (2017)
27	33.589904	96.158384	4.5	Gully	Gully depth of 0.5 m is perpendicular to the fault scarp which is about 1.5 m in height	Wu et al. (2017)
28	33.594581	96.099504	4.9	Gully		Wu et al. (2017)
29	33.594139	96.094972	3.8	Small ditch		Wu et al. (2017)
30	33.594722	96.093944	3.5	Gully		Wu et al. (2017)
31	33.603722	96.043250	5.0	Gully	Gully depth of 0.5 m	Wu et al. (2017)
32	33.604472	96.040083	5.0	Gully	Gully depth of 0.5 m	Wu et al. (2017)
33	33.605650	96.034067	1.8	Small ditch	Small ditch is nearly perpendicular to the fault scarp which is about 1.5 m in height	Wu et al. (2017)
34	33.618815	95.966376	3.6	Gully		Wu et al. (2017)
35	33.634833	95.907717	3.0	Gully	Little yellow flower distribute along the small ditch	Wu et al. (2017)
36	33.635150	95.906833	5.0	Gully	Cumulative displacement of two deformation events?	Wu et al. (2017)
37	33.650345	95.846409	4.6	Gully	Gully on the lower terrace (T1) is sinistral offset of 5.0 m	Wu et al. (2017)
38	33.664119	95.803319	5.0	Gully	Cumulative displacement of two deformation events?	Wu et al. (2017)
39	33.689267	95.692783	3.4	Gully	Gully is nearly perpendicular to the fault scarp which is about 1.0 m in height	Wu et al. (2017)
40	33.697333	95.655194	3.5	Gully	Gullies are sinistral offset of 3.5 m, 11 m and 24 m, respectively; the offset of 11 m and 24 m are cumulative offsets based on the field observations	Wu et al. (2017)
41	33.700333	95.634250	5.0	Gully	Gully depth of 0.5 m	Wu et al. (2017)
42	33.726278	95.507528	4.8	Gully	Gully is nearly perpendicular to the fault scarp which is about 0.5 m in height	Wu et al. (2017)
43	33.826678	95.239822	5.0	Gully	Cumulative displacement of two deformation events?	Wu et al. (2017)
44	33.693888	95.670555	1.7	Stone ridge		(Li et al., 2016)
45	33.644444	95.875833	1.3	Gully	The lower floodplain of gully (T0) are sinistral offset of 1.3 m	(Li et al., 2016)
46	33.611111	96.000833	1.7	Gully	The higher floodplain of gully (T0) are sinistral offset of 1.7 m	(Li et al., 2016)
47	33.608333	96.015555	1.7	Small ditch	Small ditch is sinistral offset of 1.7 m	(Li et al., 2016)
48	33.608333	96.015000	1.4	Small ditch	Small ditch is sinistral offset of 1.4 m	(Li et al., 2016)
49	33.608055	96.018055	1.7	Small ditch	Small ditch is sinistral offset of 1.7 m	(Li et al., 2016)
50	33.595000	96.094444	2.0	Small ditch	Small ditch is sinistral offset of 2.0 m	(Li et al., 2016)
51	33.594722	96.095277	2.1	Small ditch	Small ditch is sinistral offset of 2.1 m	(Li et al., 2016)

wall profile of the modern gully cleaned by manual work in the field shows the latest active characteristics of the fault (**Figure 7B**). Three units can be identified based on the main differences in the material composition and structures. Unit 1, rich in plant roots, is a set of loose gray–brown sandy soils with a thickness of ~20 cm. Unit 2 is bluish-gray laminated gravel with more medium sand traps with a thickness of ~40 cm. Although the top of Unit 2 is generally intact, the bottom of the unit

obviously suffers from deformation. Unit 3, as the main deformed stratigraphic unit in the trench, is mainly composed of reddish-brown medium coarse sand with pebbles with a thickness of ~80 cm.

The fault traces F1-1, F1-2, F2-1, and F2-2 are sketched and recognized based on the formation occurrences and cross-cutting relationships between the stratigraphic units observed in the field (**Figure 7B**). Both faults cut the sediments of Units 2,3 and are

overlain by Unit 1, and the accumulated offset of the faults is ~ 0.5 m. The gravel in the fault zone is well oriented (**Figures 7C,D**).

Three effective chronological data constrict the time of the latest deformation of these faults (**Figure 7B**). One ^{14}C dating result (DJFZ- ^{14}C -10) tested by the Beta Laboratory in the United States reveals that formation three in the trench was formed in the late Holocene ($\sim 1,270 \pm 30$ BP), and two OSL dating results further constrain that the latest deformation time of the fault is between 0.33 ± 0.05 ka BP (DJFZOSL-22) and 0.46 ± 0.03 ka BP (DJFZOSL-50). The dating results of these samples are in good agreement with the surface microgeomorphic deformation characteristics mentioned above, indicating that the latest tectonic deformation was formed during the CE 1738 Dangjiang historical earthquake.

Another trench was located on the southwest bank of the Dangjiang River (**Figure 3 M**). Field geomorphologic observations near this site show that the river terraces faulted and duplicated the preexisting structural zone (**Figure 8A**). The trench in terrace T2 is approximately 15 m long and ~ 5 m high and is also perpendicular to the general trend of the CE 1738 Dangjiang co-seismic surface ruptures (**Figure 8**).

Enlarged maps of the trench wall reveal that the strata mainly consist of alluvial-diluvial material with obvious bedding and grinding features (**Figures 8B,C**). According to the differences of these alluvial-diluvial materials in structure and composition, four distinct units can be recognized. Unit 1, rich in plant roots, is mainly a set of dark-brown fine sand and loess with a thickness of approximately 10 cm. Unit 2 is mainly a set of pale yellow sandy clays with an average thickness of ~ 1.0 m, the bottom of which is displaced by the faults. Unit 3 is a bluish gray pebble interbedded with medium coarse sand, with an average thickness of approximately 1.2 m. Unit 4 is a reddish-brown sandy with obvious bedding and gravels, with a visible thickness of approximately 1.5 m.

Three faults can be identified in the wall. Fault F1 vertically cuts Formation 2, Formation 3, and Formation four systematically, and its placement is ~ 0.5 m. Faults F2, and F3 have similar tectonic features, both of which cut the sediments of Units 4–2 and are overlain by Unit 1, the total displacements of which are ~ 0.4 m. Herein, it is speculated that the co-seismic vertical displacement of the CE 1738 Dangjiang historical earthquake was absorbed simultaneously by these three faults, and thus that the offset of ~ 0.9 m observed in the profile may represent the vertical offset during the CE 1738 Dangjiang historical earthquake.

Three chronological samples are collected at this point to constrain the recent deformation of the structure. The dating test result of sample DJFZOSL-01 hints that the age of Unit 1 is 1.10 ± 0.09 ka BP. Two other data points, DJFZOSL-22 (0.33 ± 0.05 ka BP) and DJFZ- ^{14}C -04 (770 ± 30 BP), in Unit 2 further constrain the latest activity time of the DJF. All these results hint that the latest rupture should have occurred in the CE 1738 Dangjiang historical earthquake with a co-seismic vertical offset of ~ 0.9 m.

In brief, the latest activity of the middle segment along the DJF occurred after 770 ± 30 BP and was closer to 0.33 ± 0.05 ka BP, which most likely corresponds to the CE 1738 great historical earthquake. Tectonic geomorphic features and interpretation of

trench results hint that the kinematic properties of the fault are dominated by left-lateral motion.

The Eastern Segment of the DJF

The ~ 35 km-long eastern ruptured segment of the DJF starts from Chanaka village at its northwestern end and stops near the Jielong River at its southeastern end (**Figure 3 E**). The co-seismic surface ruptures in the CE 1738 Dangjiang great earthquake are delineated by six representative observation points, the locations of which are marked in **Figure 3 E**.

Although the geomorphic characteristics of strike-slip movement, such as consistently deflected gullies and fault saddles along the structure, are still as abundant as those of the western and middle sections, the co-seismic displacements in the eastern segment sharply decrease (**Figure 9**). Field observations reveal that Paleogene brown–red sandstone with gravel is over thrust by Triassic gray–black sandstone. Recent fault activity ruptured the surficial Quaternary yellow–brown soil by thrusting, above which $\sim 0.8 \pm 0.1$ m co-seismic vertical offsets can be observed (**Figure 9A**). Several modern creeks and gully terraces are deflected or offset by the movement of the structure, and the horizontal displacement decreases from 1.5 ± 0.1 m near Chanaka village to 0.5 ± 0.1 m in Dongduola village and then disappears just across the Jielong River; meanwhile, the vertical displacements also decrease from $\sim 0.9 \pm 0.1$ m east of Lixin town to $\sim 0.3 \pm 0.1$ m on the east bank of the Jielong River and then gradually fade away.

DISCUSSION

Co-Seismic Surface Ruptures Produced by the CE 1738 Dangjiang Earthquake

It is widely accepted that co-seismic surface ruptures resulting from great earthquakes are crucial for understanding the activity habits of seismogenic faults. Thus, the first goal in the discussion is to decipher the deformation characteristics of the DJF through the co-seismic surface ruptures produced by the CE 1738 Dangjiang earthquake. According to the observed deformation features of the sedimentary layers, the fault structures, and the dating of samples in this research, a nearly 100 km-long co-seismic surface rupture zone associated with the CE 1738 Dangjiang historical earthquake is proposed here (**Figure 2**). The rupture zone is interpreted as a nearly straight lineament trending E-W to WNW-ESE in the satellite images that were used for field mapping and investigation (**Figures 2, 3**). The surface ruptures are concentrated in a zone with average widths of 5–35 m. The surface fault zone constitutes distinct shear faults, consistent offset or deflected gullies, linear scarps, fault troughs, sag ponds, mole tracks, and depressions, most of which are distributed along the fault lineament, as shown in the satellite image (**Figures 2, 3**).

The distribution of the co-seismic displacements from the CE 1738 Dangjiang earthquake is further measured based on satellite images and field verification (**Figure 10**), and all displacements are measured along the co-seismic rupture zone via displaced linear geomorphic and geologic markers such as fault scarps,

TABLE 3 | Estimated Holocene slip rates of the typical observation sites along the DFZ.

Observation sites	Morphologic units	Results of chronology	Measured horizontal displacements	Estimated slip rate
Dasheng village (Figure 11)	T2 terrace	3.04 ± 0.37 ka BP (Wu et al., 2017)	29.0 ± 1.0 m (Wu et al., 2017) 28.0 ± 1.0 m	7 ± 3 mm/yr (Wu et al., 2017) 9.2 ± 1.1 mm/yr
West of the Riqing Village (Figure 12)	T1 terrace	3.38 ± 0.03 ka BP	23.0 ± 0.2 m	6.8 ± 0.1 mm/yr
	T2 terrace	3.76 ± 0.20 ka BP	28.0 ± 0.5 m	7.4 ± 0.4 mm/yr
	T3 terrace	10.47 ± 0.58 ka BP	73.0 ± 2.0 m	6.9 ± 0.4 mm/yr
South of the Riqing Village (Figure 13)	T2 terrace	3.08 ± 0.50 ka BP	26.0 ± 0.5 m	8.4 ± 1.4 mm/yr
	T1 terrace	2.44 ± 0.14 ka BP	20.0 ± 0.2 m	8.2 ± 0.5 mm/yr
South of Dangjiang (Wu et al., 2017)	T3 terrace	9.88 ± 1.88 ka BP	70 ± 10 m	7.1 ± 1.6 mm/yr
Near the Dangjiang village (Zhou et al., 1997)	T2 terrace	5.8 ± 0.20 ka BP	42.0 ± 5.0 m	7.2 ± 0.8 mm/yr

TABLE 4 | Summarized activity parameters of the GYFZ in Holocene.

Segmentation	Ganzi S	Manigange S	Dengke S	Yushu S	Dangjiang S	References
Estimated magnitude of historical earthquake	1845 Ms 7.3	1854 Ms 7.7 1854 Mw 7.7	1896 Ms 7.5 1896 Ms 7.3	2010 Ms 7.1	1738 Ms 7.3 1738 Mw 6.8–7.2 1738 Ms 7.3 1738 Mw 7.5 1738 Ms 7.6	Zhou et al. (1997) Wen et al. (2003) Chen et al. (2010) Wu et al. (2017) (Li et al., 2016) (Li et al., 2016) This research Wen et al. (2003) Chen et al. (2010) Li et al. (2012)
		180	70	70	31 75	
Co-seismic surface rupture length (km)					85 ~100	(Li et al., 2016) Wu et al. (2017) This research
	3.4 ± 0.3	7.0 ± 0.7 3.3–5.0	7.2 ± 1.2		7.3 ± 0.6 ~7.3	Zhou et al. (1996) Li et al., 1995
	11.5 ± 2.4 14.3 ± 3	12.8 ± 1.7 13.4 ± 2	11.3 ± 1.8			Wen et al. (2003) Xu et al. (2003a)
Estimated slip rate (mm/a)				2–5 2.1–3.7	7 ± 3 6.6 ± 1.5 (GPS) 6.4 (InSAR) 6.3 ± 2.4	Lin et al. (2011) Wu et al. (2017) Huang et al. (2015a) Wang et al. (2011) Liu et al. (2011) This research
Average horizontal co-seismic offset (m)	1.7 ± 0.2	4.3 ± 0.6	2.6 ± 0.3	2.6 ± 0.3 0.5–1	1.7 2.1 ± 0.1	Zhou et al. (1997) Li et al. (2012) (Li et al., 2016) This research
Average repeated time of great earthquakes (a)	500 ± 74	614 ± 105 667 ± 61	361 ± 73 443 ± 67	500 ± 74	356 ± 41 274 ± 30	Zhou et al. (1997) Chen et al. (2010)

stream channels, gullies, and terraces, which generally intersect the surface rupture zone vertically (**Table 2**). Along the ~100 km long surface rupture zone, the co-seismic displacements at 20 locations are jointly measured by a laser rangefinder, a steel tape, and a drone. We also collected and verified data from 31 previous surveys (**Table 2**, Zhou et al., 1997; Li et al., 2016; Wu et al., 2017). What needs to be pointed out is that the coexistence of nearly double-magnitude offsets in the gullies (e.g., displacements measured in No. 37 and No. 40 in **Table 2** and the observation site in **Figure 4E**) at the same site hints that the structure might have experienced multiple stages of repeated sliding along the pre-existing fault line, and the minimum

displacement is regarded as the co-seismic offset of the CE 1738 Dangjiang earthquake. Finally, 20 field survey data points were fitted and then compared with previous studies (**Figure 10**).

The statistical data show that the *in situ* co-seismic offsets owing to the CE 1738 Dangjiang earthquake are left-lateral and vary from 1.0 to 3.5 m horizontally, and the average and maximum left-lateral slip displacements are $\sim 2.1 \pm 0.1$ m and $\sim 3.5 \pm 0.1$ m, respectively (**Figure 10**). The maximum and average vertical displacements of the CE 1738 Dangjiang earthquake were 1.0 and 0.3 m, respectively (**Figure 10**). The fitted figure also shows that the maximum horizontal

displacements are mainly concentrated in the central part of the fault sections, such as at Riqing village, Lixin town, and Naidong village (**Figure 10**), while they sharply decrease toward the fault ends in Dangjiangrong village and the Jielong River, respectively. This particular statistical figure almost approaches that of the schematic illustration of an ideal, isolated earthquake (**Figure 10**), where the displacement profile indicates the maximum offset near the structural center and decreases toward the fault tips (Fossen, 2016). Although the variation gradient of the fitted co-seismic vertical displacement distribution is not as dramatic as that of the horizontal displacement, its overall spatial distribution shows a decreasing trend from the middle toward the ends (**Figure 10**).

Our field investigation did not extend into the area west of Duocai village because there are no interpreted surface traces of the CE 1738 Dangjiang co-seismic deformation in the remote sensing images (**Figure 2**) and because this area is too difficult to access given high and steep mountains, seasonal glaciers, permanent marshes, and inconvenient transportation. If the co-seismic rupture of the earthquake did not spread to the area west of Duocai village, this asymmetrical co-seismic distribution may mean that the co-seismic rupture process of the CE 1738 Dangjiang earthquake might start from Duocai village, propagate unilaterally eastward, and finally terminate in the vicinity of the Jielong River. In fact, previous studies have shown that the 2010 Yushu earthquake has similar unilateral propagation characteristics (Li et al., 2012). Herein, it is thus proposed that the surface rupture of the CE 1738 Dangjiang earthquake may be ~100 km long (**Figure 2**), which is slightly different from previous studies (Institute of Geophysics, China Seismological Bureau, 1990; Zhou et al., 1997; Chen et al., 2010; Yin, 2011).

Magnitude Estimation of the CE 1738 Dangjiang Earthquake

The magnitude of the CE 1738 Dangjiang earthquake is re-estimated based on the newly observed parameters described above. Previously, scientists have suggested that the magnitude estimation is jointly influenced by the co-seismic surface-rupture length and the maximum displacement observed on a continental fault trace (Deng et al., 1992; Wells and Coppersmith, 1994; McCalpin, 2009) and that there are no obvious surface ruptures if the magnitude is smaller than M 6.5 on the Tibetan Plateau (Deng et al., 1992; Yin, 2011). Furthermore, based on more than 60 detailed modern surface-rupture earthquakes distributed in different tectonic regions that exhibit different tectonic features in the East Asia area, Deng et al. (1992) presented different kinds of distinctive fitted functions between the magnitude of an earthquake (M), the length of the surface rupture (L), and the maximum displacement (D). Herein, the following fitted function is adopted according to their research (Eq. 1):

$$M = 6.34 + 0.49 \lg(LD) \quad (1)$$

M : the estimated magnitude of the earthquake;

L : the length of the co-seismic surface ruptures in kilometers;

D : the maximum displacements in meters.

In the above equation, L stands for the length of the surface rupture, which is assigned a value of 100 km in this study according to the satellite image interpretation and field observations. Here, D refers to the maximum horizontal displacement of a single event, which is assigned a value of 3.5 m. As a result, the magnitude of the CE 1738 Dangjiang earthquake can be estimated as M 7.6 based on the co-seismic surface rupture length and the maximum displacement observed in the field. (Deng et al., 1992).

If we use the calculation formula suggested by Wells and Coppersmiths (1994, Eq. 2), the calculation result is M 7.4 based on the ~100 km long co-seismic surface rupture length. (Wells and Coppersmiths, 1994).

$$M = 5.16 + 1.12 \lg(L) \quad (2)$$

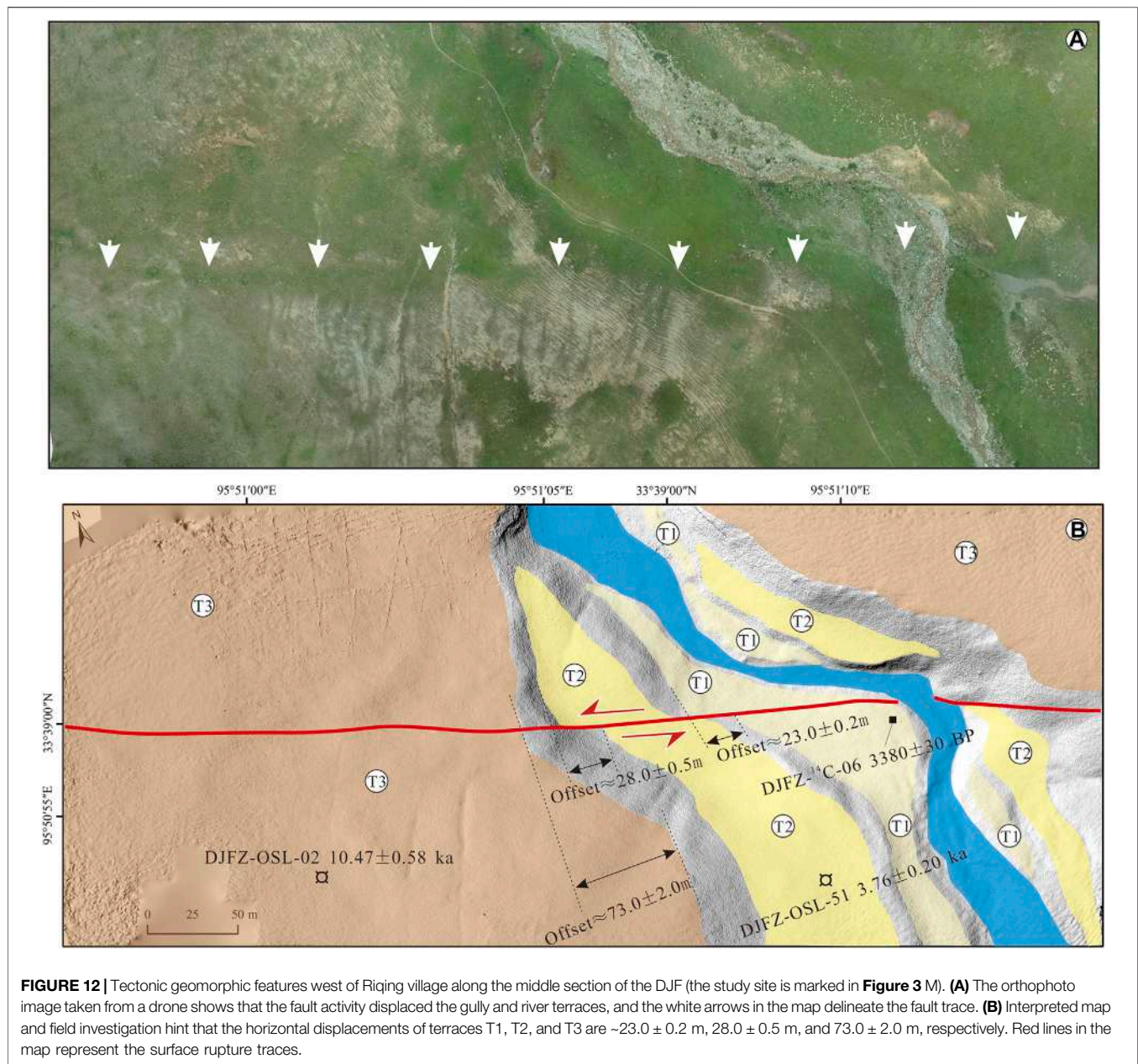
In general, the estimated magnitude of M 7.6 is basically consistent with those of previous studies. Although the traffic and living conditions on the Tibetan Plateau were inconvenient in earlier years and thus would have directly affected field observations, many scientists attempted to estimate the likely magnitude of this great historical earthquake based on geological surveys and historical documents. For example, Zhou et al. (1997) suggested that the possible magnitude of this great historical earthquake was ~ M 7.5 based on an average left-lateral displacement of ~5 m, while Chen et al. (2010) estimated a magnitude of ~ M 7.1 according to a co-seismic surface strike-slip displacement of ~2 m and a co-seismic surface rupture length of ~31 km. Lin et al. (2011), based on a comparative study with the 2010 M_s 7.1 Yushu earthquake proposed that the minimum magnitude of the CE 1738 Dangjiang earthquake was M_s 6.5 and that the magnitude was probably less than M_s 7.3 or perhaps M_s 7.1, with a co-seismic offset of ~1.5–2.0 m. Other scientists speculated that the co-seismic rupture of the CE 1738 Dangjiang great earthquake might have propagated into the Yushu segment of the Ganzi-Yushu Fault Zone (Zhou et al., 1997; Lin et al., 2011) and that the surface rupture length in the CE 1738 Dangjiang earthquake may have reached or exceeded 130 km with an estimated magnitude of $7\frac{1}{2}$ based on a review of historical records (Yuan et al., 2011). All these results mean a larger magnitude than the previously suggested size of 6.5 or $6\frac{1}{2}$ (Institute of Geophysics, China Seismological Bureau, 1990). In addition, the chronological results obtained in recent years may be more reliable than older results if the advances in testing technology over the last 20 years are taken into account. In fact, both historical record research and field geological survey results support that the ~100 km co-seismic rupture length is approximately representative of the $M = 7.5$ + earthquake along the GYFZ (**Table 4**). Thus, it is proposed that the magnitude of the CE 1738 Dangjiang earthquake most likely was M 7.6, with a co-seismic surface rupture of ~100 km in length, and that the average and maximum strike-slip displacements are $\sim 2.1 \pm 0.1$ m and $\sim 3.5 \pm 0.1$ m, respectively.

Active Tectonics of the DJF

Field observations reveal that the DJF is an active structure that has been undergoing repeated left-lateral strike-slip motion

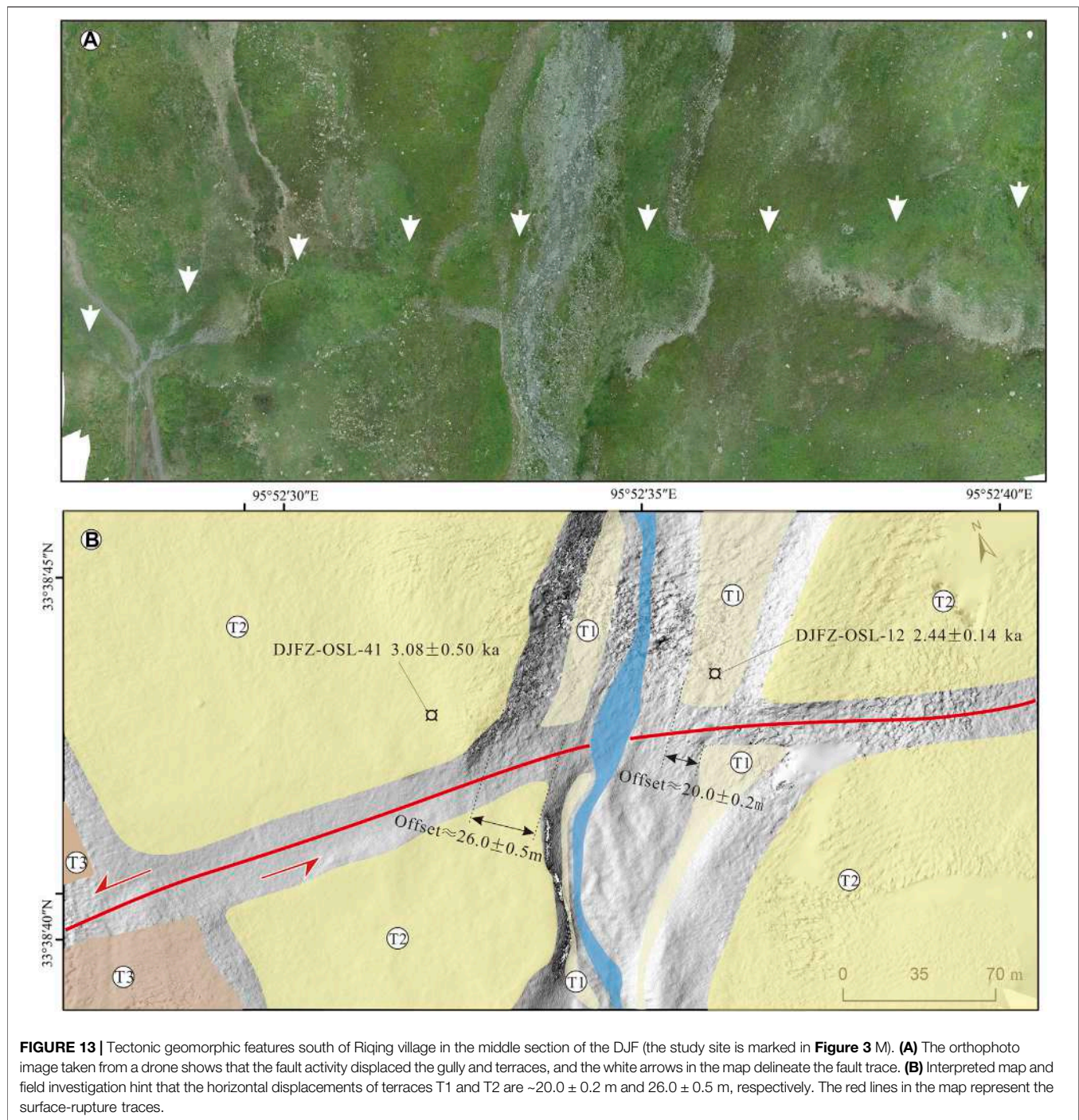


FIGURE 11 | Tectonic geomorphic features in Dasheng village along the western section of the DJF (the study site is marked in **Figure 3 W**). **(A)** Fault activity formed distinct troughs and valleys, and the red arrows in the map delineate the fault trace. **(B)** The orthophoto image taken from a drone shows that river terraces T1 and T2 are displaced by the activity of the DJF. **(C)** Interpreted map and field investigation further reveal that the horizontal displacements of terrace T2 are $\sim 28.0 \pm 1.0$ m.



during the late Quaternary. Although trenches reveal normal features that conflict with the patterns of geomorphic scarps (**Figures 5, 7, 8**) because of the so-called “dolphin effect” along the strike-slip fault or of the gravity collapse of the main fault at the surface, the overall kinematic property of the fault should be left-lateral strike-slip, and there will be deformation of normal and reverse faults in local areas or observation sites. The above inferences are further proven by the following observations at different points along the DJF. In general, the older and larger geomorphic and geologic markers, such as stream channels and terraces, across the DJF structure display greater offsets than younger and smaller markers (**Figures 4–9**). For example, a

gully is offset by a preexisting fault, and the co-seismic surface rupture from the CE 1738 *M* 7.6 Dangjiang historical earthquake, with an accumulated horizontal displacement of $\sim 9.4 \pm 0.4$ m and an accumulated vertical displacement of $\sim 3.0 \pm 0.2$ m (**Figure 4B**). An abandoned gully with an accumulated horizontal displacement of 5.1 ± 0.2 m shows that the fault may have experienced two left-lateral strike-slip movements if the co-seismic displacement in the latest faulting event is estimated at $\sim 2.5 \pm 0.1$ m (**Figure 4E**). Similar cumulative displacement distributions are also well verified by the deformation characteristics on the stream terraces, which show that the displacements of old stream terraces are much greater than those



of young stream terraces. For example, three regional rivers of different sizes that flow through the research area, named the Zhiduo River, Dangjiang River, and Jielong River from northwest to southeast, have obviously experienced long-term left-lateral strike-slip tectonic deformation with accumulated displacements of ~ 6 , ~ 16 , and $\sim 5 \text{ km}$, respectively (**Figures 2, 3**) (Wang et al., 2007). These offsets indicate that the DJF has a long history of repeated activity. All these field observations and the accumulated displacements show

that the DJF has experienced multiple repeated sinistral strike-slip seismic slips.

The estimation of slip rates reveals that this repeated activity is typical of the fault's long-term history. Previous studies show that the whole XFZ is sinistraly displaced at a slip rate of $15 \pm 5 \text{ mm/yr}$ in its northernmost section, decreasing to $\sim 5 \text{ mm/yr}$ in its southern section (Molnar and Deng, 1984; Tang et al., 1984; Allen et al., 1991; Wang

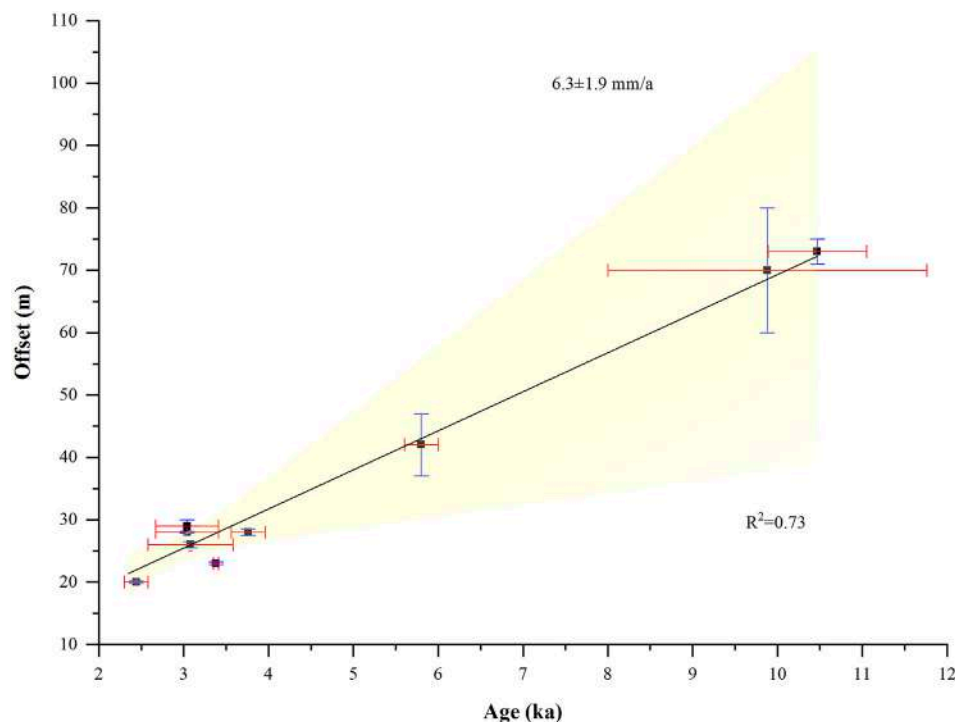


FIGURE 14 | Graph showing ages and offsets for individual sites along the DJF structure in the Holocene (Table 3). Blue and red error bars are the respective offsets and ages measured in the field. The linear fitting result with a 95% confidence interval hints that the horizontal slip rate of the DJF in the Holocene is estimated at 6.3 ± 1.9 mm/yr.

et al., 2001; Lin et al., 2002; Peng et al., 2006; Wang et al., 2007; Wang and Shen, 2020); the strike-slip rate estimations for the GYFZ section vary, including 3.4–7.3 mm/yr (Zhou et al., 1997), ~12 mm/yr (Wen et al., 2003), ~5 mm/yr (Zhang et al., 2004), and 2–5 mm/yr (Lin et al., 2011). These differences may mainly result from the dating methods used and the displacement of the geomorphic surface measured.

Herein, the horizontal slip rate along the DJF in the Holocene is estimated according to the different geomorphic surface exposure ages and the corresponding measured displacements from four typical observation sites (Table 3; Figures 11–13). The most typical survey site is located west of Riqing village in the middle part of the fault (Figure 12). The exposure ages of three geomorphologic surfaces, T3, T2, and T1, are well constrained by three dated samples, DJFZ-OSL-02 (10.47 ± 0.58 ka BP), DJFZ-OSL-51 (3.76 ± 0.20 ka BP), and DJFZ- ^{14}C -06 (3.38 ± 0.03 ka BP), and the left-lateral displacements are measured to be 73.0 ± 2.0 m, 28.0 ± 0.5 m, and 23.0 ± 0.2 m, respectively (Figure 12). Thus, it is further estimated that the horizontal slip rate of the DJF near Dangjiang village is approximately 6.8 ± 0.1 – 7.4 ± 0.4 mm/yr. The same calculation method is applied to other observation sites (Figures 11, 13), and the estimated results are basically consistent (Table 3). The estimated slip rate of the western end of the fault

(Figure 11) is slightly larger than that in the middle segment of the fault (Figures 12, 13). There are two possible reasons for this tiny difference: errors in dating or displacement measurements and local erosion processes near the river. Additionally, Wu et al. (2017) suggested that the slip rate at this site (Figure 11) is approximately 7 mm/yr if the formation time of the upper and lower terrace surfaces is jointly used to constrain the initial time of displacement. Therefore, the linear fitting surface displacement rate with a 95% confidence interval is 6.3 ± 1.9 mm/yr (Figure 14). Collectively, this surface displacement rate falls within the confidence intervals of earlier estimates (Molnar and Deng, 1984; Allen et al., 1991; Wen et al., 2003; Peng et al., 2006; Wang and Shen, 2020). Notably, the estimated slip rate of 6.3 ± 1.9 mm/yr is highly consistent with previous studies from GPS and INSAR observations (Wang et al., 2011; Liu et al., 2011).

Implications for Seismic Hazard and Strain Partitioning Along the Southern Boundary of the Bayan Har Block in the Central Tibetan Plateau

The spatial-temporal distribution of co-seismic ruptures during historical earthquakes is of great importance in assessing the seismic hazard presented by active faults.

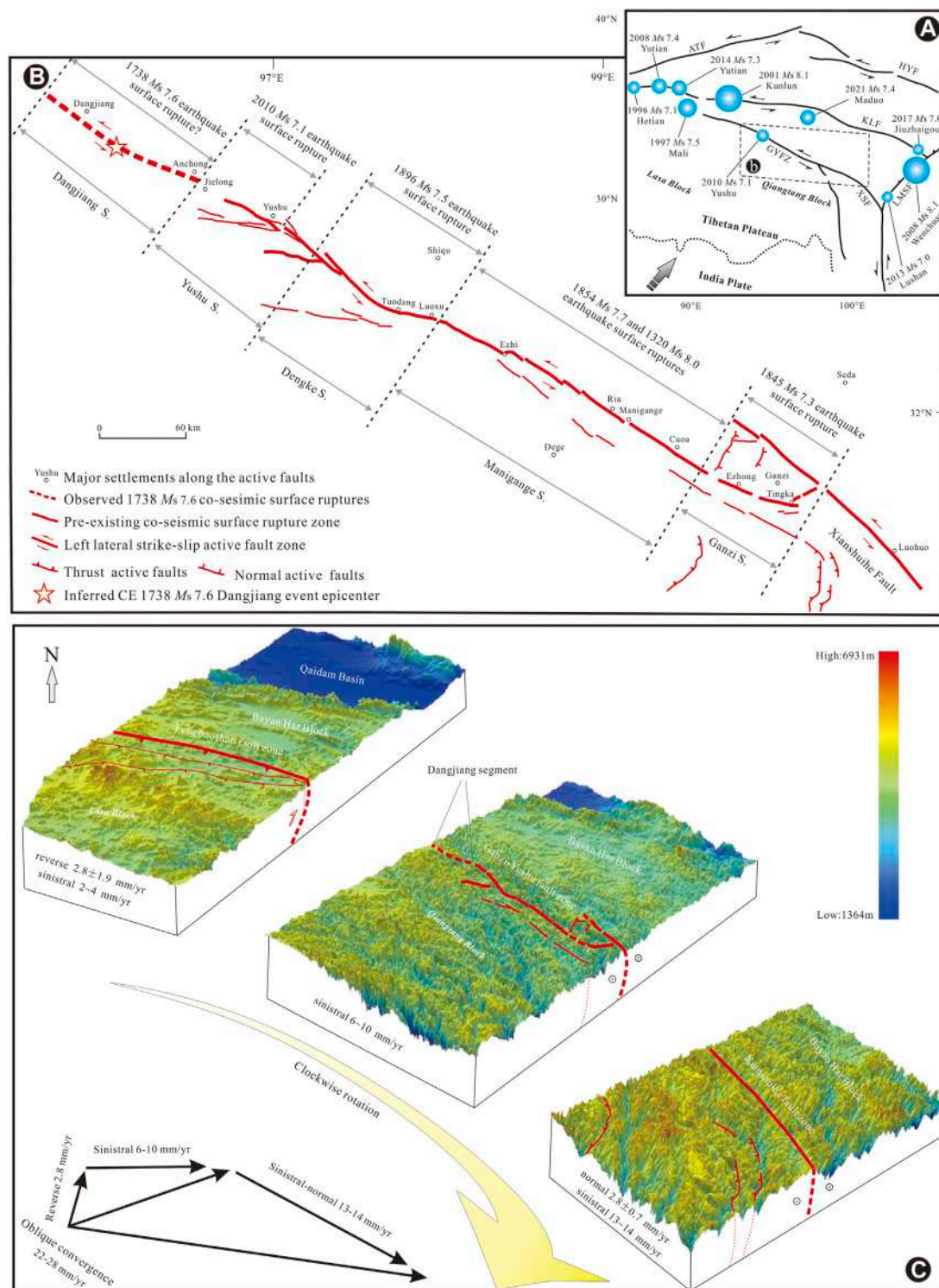


FIGURE 15 | Geometrics of the Ganzi-Yushu Fault Zone and the co-seismic surface ruptures resulting from historical great earthquakes along the northwestern continuation of the Xianshuihe Fault Zone (XFZ) in the high mountain area of the central Qinghai-Tibetan Plateau. **(A)** Index map showing the major active faults and great earthquakes around the Bayan Har Block boundary in the northern Qinghai-Tibet Plateau. ATF, Altyn Tagh Fault; HYF, Haiyuan Fault; KLF, Kunlun Fault; GYZ, Ganzi-Yushu Fault Zone; LMSF, LongMenShan Fault Zone; XFZ, Xianshuihe Fault Zone, FFZ, Fenghuoshan Fault Zone. The gray-black arrow in the lower left corner approximately represents the direction of the northward subduction of the Indian Plate. **(B)** Sketch map showing the geometric distribution of the GYZ and the co-seismic surface ruptures resulting from historical great earthquakes along the structure. Dangjiang S, Dangjiang segment; Yushu S, Yushu segment; Dengke S, Dengke (Continued)

FIGURE 15 | segment; Manigange S, Manigange segment; Ganzi S, Ganzi segment (part of segments refer to Zhou et al., 1996; Zhou et al., 1997; Wen et al., 2003; Deng et al., 2007; Chen et al., 2010; Lin et al., 2011; Li et al., 2012). Traces of co-seismic surface rupture in the CE 1738 *Ms* 7.6 Dangjiang earthquake is delineated by a red dashed line based on satellite image interpretation and field investigation. **(C)** Cartoon model of the strain partition along the southern boundary of the Bayan Har Block. The base map of the digital elevation model is 90 m resolution SRTM (Shuttle Radar Topography Mission) data. Some faults and slip rates refer to previous studies (Zhou et al., 1996; Zhou et al., 1997; Wen et al., 2003; Deng et al., 2007; Chen et al., 2010; Wang et al., 2011; Lin et al., 2011; Li et al., 2012; Li et al., 2016; Wu et al., 2017). The cartoon map shows that the oblique convergence of the Qiangtang Block might be accommodated mainly by clockwise rotation through repeated left-lateral strike-slip movements along the southern boundary of the Bayan Har Block in the central Tibetan Plateau.

Previous studies show that at least ten strong earthquakes ($M_s \geq 7$) have occurred along the boundaries of the Bayan Har Block in the northern Qinghai-Tibetan Plateau (**Figure 15A**). These strong earthquakes hint that tectonic strain is mainly concentrated on the boundaries of the block, with minor deformation in the block interior (Thatcher, 2007; Michale and Yin, 2009). In particular, along the southern boundary of the block, the activity of the northwestern continuation of the XFZ has resulted in numerous destructive earthquakes in history, such as the CE 2010 Yushu *Ms* 7.1, the CE 1896 Dengke *Ms* 7.5, the CE 1845 *Ms* 7.3 in the Ganzi step, the CE 1738 Dangjiang *Ms* 7.6, the CE 1854 *Ms* 7.7, and the CE 1320 *Ms* 8.0 in the Maligange area (**Figure 15A**). Although these great earthquakes have ruptured different segments of the GYFZ with variable co-seismic surface rupture sizes, activity parameters, and distinctive spatial-temporal distribution features (**Figure 15B**; **Table 4**, some co-seismic surface rupture distribution images are after Zhou et al., 1996; Zhou et al., 1997; Wen et al., 2003; Deng et al., 2007; Chen et al., 2010; Lin et al., 2011), previous studies suggest that the tectonic deformation along most sections of the XFZ is indeed consistent with the characteristic earthquake model. Thus, this spatial-temporal distribution of co-seismic ruptures may indicate that the recurrence probability of a strong earthquake along the Dangjiang segment cannot be underestimated given that its ~ 283 years elapsed time of the last great faulting event is basically close to the 350 ± 41 years recurrence interval of major earthquakes (**Table 4**; Zhou et al., 1997) or even slightly beyond its recurrence cycle of 274 ± 30 years (**Table 4**; Chen et al., 2010).

Our study area is located in a transitional area between the Tibetan Plateau to the west and its southeastern margin to the southeast, and the latter area experienced clockwise rotation around the northeastern syntaxis of the Himalayas with a component of southeastward extrusion (Wang et al., 2008). The strain partition in this area is crucial for understanding the southeastward extrusion process. Previous studies indicate that multiple repeated sinistral strike-slip movements along the boundary fault may accommodate the convergent motion of the block. However, how can the strain distribution process be quantitatively understood? This question has important implications for the tectonic deformation of the central region of the Qinghai-Tibet Plateau. The study of the kinematics of the boundary belt is helpful to answer this question.

The southern boundary of the Bayan Har Block consists of three main faults: the Fenghuoshan fault zone in the northwest

segment, the Ganzi-Yushu fault zone in the middle segment, and the Xianshuihe fault zone in the southeast (**Figure 15C**). We observe that the kinematic characteristics along the southern boundary of the Bayan Har Block are different along the strike in two aspects (**Figure 15C**). First, the kinematic characteristics of the structure change from reverse strike-slip in the northwest segment to strike-slip in the middle segment and gradually to normal strike-slip in the southwest segment. In addition, the sinistral slip rate increases from the northwest section to the southeast (**Figure 15C**). The GPS data also show that the northward motion of the plateau decreases across the Fenghuoshan thrust-fold belt, which hints that the belt is still active (Zhang et al., 2004). Obviously, these kinematical characteristics cannot be explained simply by the eastward extrusion. Block rotation, caused by the dextral shear between the Qinghai-Tibet Block and Yangtze block, may lead to a gradual increase in the slip rate from the NW segment to the SE segment along the structure. Thus, the strain partition along the boundary is crucial to discuss the possible transfer process from north-south shortening and east-west extension in the Fenghuoshan segment to oblique rotation in the middle-south segment.

Here, we propose a simple strain-distribution model to explain the tectonic characteristics mentioned above (**Figure 15C**). The oblique convergence of the Qiangtang Block to the south is $\sim 22\text{--}28$ mm/yr based on recent GPS observations (ShenLiao and Wang, 2009). This oblique motion of the Qiangtang Block might be partitioned into the Fenghuoshan Fault Zone with a slip rate of ~ 3 mm/yr, the Ganzi-Yushu Fault Zone sinistral slip rate of $6\text{--}10$ mm/yr, and normal-strike-slip faulting at the Xianshuihe Fault Zone with a slip rate greater than 13 mm/yr (**Figure 15C**). Two possible reasons may be responsible for this strain-allocation process: the strike of the fault obviously changes from nearly E-W in the northwest to nearly N-S in the southeast, and the elevation of the terrain remarkably decreases from $\sim 6,000$ m in the northwest to $\sim 3,000$ m in the southeast. The strike variation is possibly the dominant reason for the kinematic change from compression to extension, while the decrease in altitude may also be more favorable for the material to extrude eastward. In summary, we suggest that the oblique convergence of the Qiangtang Block might be accommodated mainly by clockwise rotation through repeated left-lateral strike-slip seismic motion along the southern boundary of the Bayan Har Block in the central Tibetan Plateau, although the detailed transfer mechanism still needs more investigation.

CONCLUSION

Based on geologic observations, historical literature reviews, geomorphic mapping, trench logging, and sample dating, we conclude that 1) the CE 1738 Dangjiang earthquake produced an ~100 km-long surface rupture zone that includes diagrammatic features of recent displacement, such as consistently offset or deflected gullies, linear scarps and troughs, sag ponds, *en echelon* fractures, and pressure ridges; 2) the magnitude of the CE 1738 Dangjiang earthquake is re-estimated as M 7.6 with average and maximum strike-slip displacements of $\sim 2.1 \pm 0.1$ m and $\sim 3.3 \pm 0.1$ m, respectively; 3) the DJF has undergone multiple seismic faulting events in the Holocene, and the linear fitting surface displacement rate with a 95% confidence interval is $\sim 6.3 \pm 1.9$ mm/yr; 4) oblique convergence of the Qiangtang Block might be accommodated by the clockwise rotation of block through the repeated left-lateral strike-slip movements along the southern boundary of the Bayan Har Block in central Tibetan Plateau.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary material, further inquiries can be directed to the corresponding author.

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AUTHOR CONTRIBUTIONS

ZY: Conceptualization, Methodology, Acquisition of the financial support, Writing-Reviewing and Editing. NY: Original draft preparation and presentation of the published work. PX: Visualization, Field geological survey and analysis and processing of age samples. CX: Surface geomorphology survey and analysis and data collection, Validation.

FUNDING

This work was sponsored by the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901).

ACKNOWLEDGMENTS

We are grateful to professors Pei-zhen Zhang, Wen-jun Zheng, Dao-yang Yuan, Chuan-you Li, and Jun Shen for their great assistance in this research. The discussion with Rong-jun Zhou, Hu Wang, and Yan-bao Li greatly deepened our understanding of the active tectonics and earthquake potential assessment along the western portion of the Ganzi-Yushu Fault Zone. We would like to especially acknowledge the editors and reviewers for their constructive comments and suggestions, which greatly improved this paper.

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Vertical Slip Rates of Normal Faults Constrained by Both Fault Walls: A Case Study of the Hetao Fault System in Northern China

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OPEN ACCESS

Edited by:

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Reviewed by:

Haiyun Bi,
China Earthquake Administration,
China
Yanxiu Shao,
Tianjin University, China

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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 17 November 2021

Accepted: 14 February 2022

Published: 02 March 2022

Citation:

Xu D, He Z, Ma B, Long J, Zhang H and
Liang K (2022) Vertical Slip Rates of
Normal Faults Constrained by Both
Fault Walls: A Case Study of the Hetao
Fault System in Northern China.
Front. Earth Sci. 10:816922.
doi: 10.3389/feart.2022.816922

The slip rate is a fundamental kinematic parameter of active faults. Traditional methods using fault scarps or trenches may produce inaccurate estimates of a fault's vertical slip rate. A normal fault's vertical slip rate requires constraints from the hanging wall and footwall. Here, the vertical slip rate at each measuring point along the fault was calculated by the joint constraints of terraces in the footwall and boreholes in the hanging wall. Nine measuring points were selected along the Sertengshan piedmont fault. The vertical slip rates of this fault since 65 and 12 ka were 0.74–1.81 and 0.86–2.28 mm/a, respectively. Four measuring points were selected along the Wulashan piedmont fault. The vertical slip rates of this fault since 60 and 12 ka were 2.14–3.11 and 1.84–2.91 mm/a, respectively. Seven measuring points were selected along the Daqingshan piedmont fault; the vertical slip rates were 2.5–3.88 and 1.78–2.83 mm/a since 58 and 11 ka, respectively. Analysis of the slip rates, the elapsed time since the last palaeoearthquake and the mean recurrence interval of palaeoearthquakes on each fault segment on the northern margin of the Hetao Basin suggests that the Langshankou and Hongqicun segments of the Sertengshan piedmont fault are at higher risk of earthquakes than the other segments. Among the fault segments of the Wulashan piedmont fault, the Baotou segment is at the highest seismic risk. The seismic risk of the Tuyouxi segment of the Daqingshan piedmont fault should not be ignored, and the Tuzuoqi, Bikeqi and Hohhot segments have high seismic risk. Based on the findings and a dynamic model of the formation and evolution of the Ordos block, it is concluded that the depositional centre of the Hetao Basin has tended to migrate from west to east. The vertical force generated by deep material movement is the dominant factor leading to a greater vertical slip rate in the eastern portion of the northern margin of the Hetao Basin. The modern stress field in the Hetao Basin results in an increase in the vertical slip rate of active faults from west to east along the northern margin of the basin.

Keywords: vertical slip rate, geological borehole, fault displacement, Hetao basin, marker strata, terrace

1 INTRODUCTION

The slip rate is a very important kinematic parameter of active faults and an important index for evaluating the seismic risk of faults. It refers to the velocity of fault dislocation over a certain period of time and represents the long-term and average activity level of the fault (Zhang et al., 2013; Zhang et al., 2017). It can be used to quantitatively compare the relative activity of different fault zones or the same fault zone in different periods and is also a very important parameter for seismic risk evaluation. Accurately obtaining the fault displacement amplitude is difficult, and accurately determining the rate of normal fault vertical slip requires constraints from the footwall and the hanging wall. However, because the strata of the hanging wall are buried underground by sedimentary cover when normal faults form, only the characteristics of the footwall strata exposed by surface faulting can be observed in the field, and it is difficult to find direct indicators of fault displacement. When calculating the slip rate, people usually consider only footwall strata and ignore the information of hanging wall strata buried underground, which leads to underestimating the slip rate of the fault and reducing the expected seismic risk of the fault.

Taking the fault system on the northern margin of the Hetao Basin, China, as an example, the vertical slip rate of the normal fault is constrained by terraces in the footwall of the fault and boreholes in the hanging wall of the fault. Located between the Yinshan fault block and Ordos block (Deng et al., 1999; Feng et al., 2015), the Hetao Basin is an important part of the peripheral Ordos fault basin belt and seismic belt. The northern margin fault system consists of the Sertengshan piedmont fault, Wulashan piedmont fault and Daqingshan piedmont fault from west to east and has controlled the formation and development of the Hetao Basin. The Hetao Basin experienced earthquakes with magnitudes of M7.5–M8 in 849 AD and M8 in 7 BC (He et al., 2018). In the 20th century, the Bikeqi earthquake in 1929 (M6.0), the Wuyuan earthquake in 1934 (M6.75), the Lingeer earthquake in 1976 (M6.3), the Bayinmuren earthquake in 1976 (M6.2), the Wuyuan earthquake in 1979 (M6.0) and the Baotou earthquake in 1996 (M6.0) (Figure 1) successively occurred in the fault zone; some small earthquakes of magnitude 4–5 have also occurred. Such earthquakes are distributed in the fault zone and are obviously controlled by major active faults at the boundary (Chen, 2002).

Since the 1990s, studies have investigated the vertical slip rate of the fault system on the northern margin of the Hetao Basin. The vertical slip rate of the Sertengshan piedmont fault has been 0.88–3.6 mm/a since the late Pleistocene and 0.19–2.2 mm/a during the Holocene (Institute of Crustal and Dynamics and China Earthquake Administration, 1995; Yang et al., 2002); the uplift rate of the Wulashan piedmont fault has been 0.5–2.0 mm/a since the late Pleistocene and 0.4–1.0 mm/a during the Holocene (Ma et al., 1998); and the vertical slip rate of the Daqingshan piedmont fault has been 2.61–4.49 mm/a since the late Pleistocene and 0.37–1.72 mm/a during the Holocene (Wu et al., 1996). In these previous studies, the fault slip rate was calculated considering only the footwall uplift height, that is, the

footwall uplift rate, rather than the complete vertical slip rate of the fault. This approach leads to an underestimation of fault activity and earthquake risk.

In this paper, based on chronological data from marker strata, accurate vertical slip rates are calculated. The kinematic parameters of the faults in the late Quaternary were quantitatively studied by using boreholes in the hanging wall and terraces in the footwall, and the activity characteristics of the fault system in the late Quaternary and the seismic risk in the future were analysed and summarized.

2 REGIONAL SEISMIC TECTONIC SETTING

The Ordos block is located on the northeastern Tibetan Plateau. Due to the continuous influence of the uplift of the Tibetan Plateau and the subduction of multiple plates in the east, faults with different faulting intensities developed in different periods of the Cenozoic and formed a series of graben basins (Deng et al., 1999). The Hetao Basin between the northern margin of the Ordos block and the Yinshan (Figure 1) developed in the Oligocene mainly under the action of the maximum principal compressive stress field oriented in the ENE direction and has accumulated sediments since the early Neogene (Research Group of Active Fault System around Ordos Massif, 1988). Geomorphologically, the striking features of the Hetao fault zone are normal fault tectonic landforms: fault-formed terraces, fault scarps, river terraces, etc. (Wang and He, 2020). From west to east, the fault systems on the northern margin of the Hetao fault zone are the Sertengshan piedmont fault, the Wulashan piedmont fault and the Daqingshan piedmont fault. The Sertengshan piedmont fault extends from Langshan Pass in the west, extends nearly east–west to Wubulangkou, then turns to 120° and continues to the southeast until disappearing to the east of Wulanhudong, with a total length of approximately 150 km. The fault can be divided into four segments from west to east: the Langshankou segment (East Wugai–Dahoudian), the Hongqicun segment (Dahoudian–Wubulangkou) with an E–W trend, the Kuluebulong segment (Wubulangkou–Wayaotan) with a NW trend, and the Dashetai segment (Wulyaotan–Taliang) (Long et al., 2017). The Wulashan piedmont fault starts at Wulateqianqi in the west and ends in northern Kundulun District in Baotou city in the east. The overall trend is nearly east–west, and the total length is approximately 110 km. It can be divided into three segments from west to east: the eastern segment of Wulateqianqi (Wulateqianqi–Gongmiaozi), the Gongmiaozi segment (Gongmiaozi–Heshunzhuang) and the eastern segment of Baotou (Heshunzhuang–Baotou) (Ma et al., 1998; Institute of Crustal and Dynamics and China Earthquake Administration, 2013). The Daqingshan piedmont fault extends from Zhaojunfen on the south bank of the Yellow River in Baotou in the west to the Kuisu area east of Hohhot in the east, with a total length of approximately 200 km. It has a general trend of ENE and a dip to the south with an angle of 45°–75° (He, 2006; Liu, 2012). It is a very important active fault in the Hetao fault zone. It can be divided into five segments from west to east: the Baotou segment (Yellow River–Xuehaigou), the Tuyouxi segment (Xuehaigou–Tuyouqi),

the Tuzuoqi segment (Tuyouqi-Tuzuoqi), the Bikeqi segment (Tuzuoqi-Wusutu) and the Hohhot segment (Wusutu-Kuisu) (Wu et al., 1996; He et al., 2007) (**Figure 1**). In this paper, nine measuring points along the Sertengshan piedmont fault,

four along the Wulashan piedmont fault, and seven along the Daqingshan piedmont fault were selected to calculate the vertical slip rates of three active faults on the northern margin of the Hetao Basin since the late Pleistocene and Holocene.

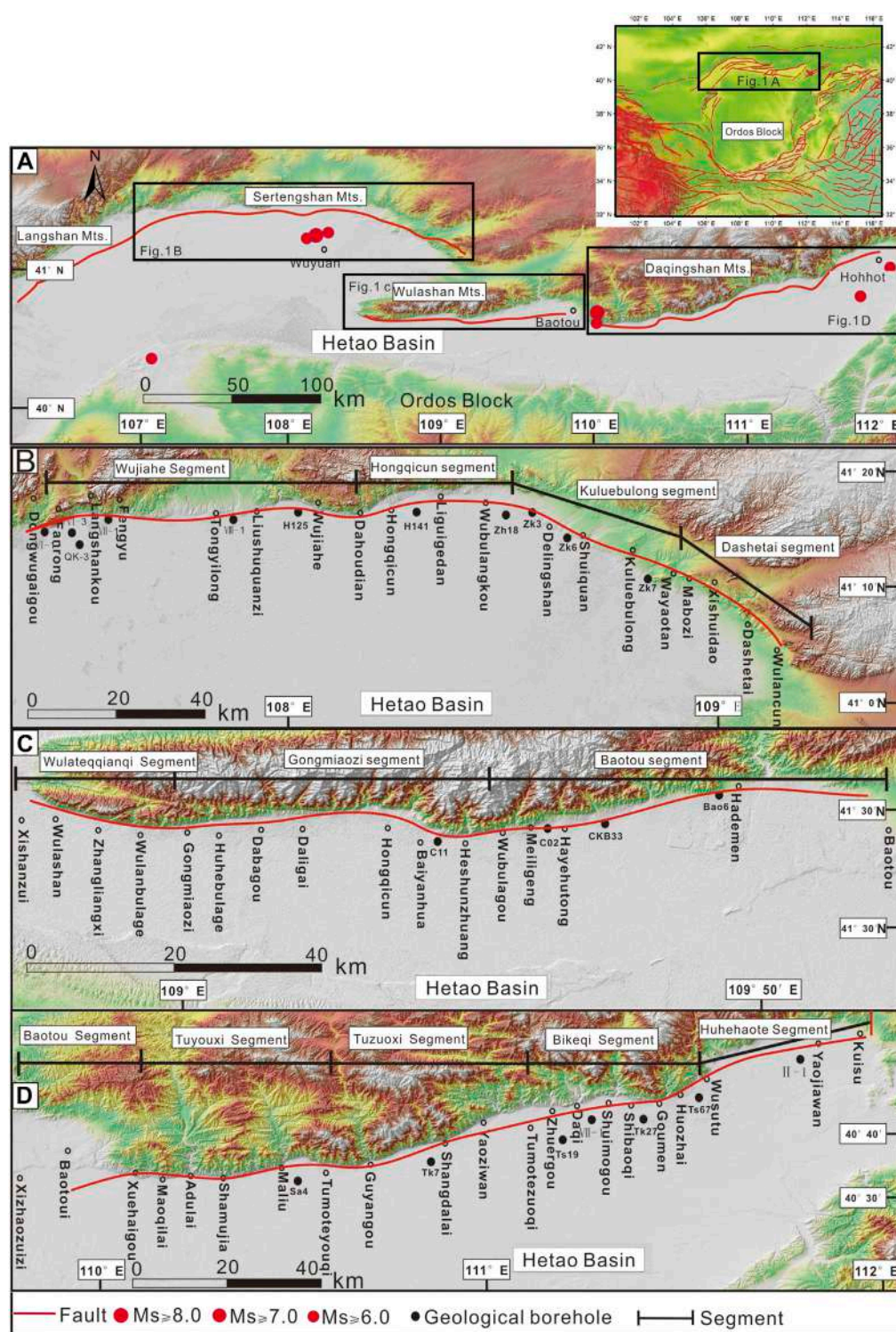


FIGURE 1 | Regional geological structure map: (A) Locations of the three faults; (B) Distribution of the Sertengshan piedmont fault; (C) Distribution of the Wulashan piedmont fault; (D) Distribution of the Daqingshan piedmont fault.

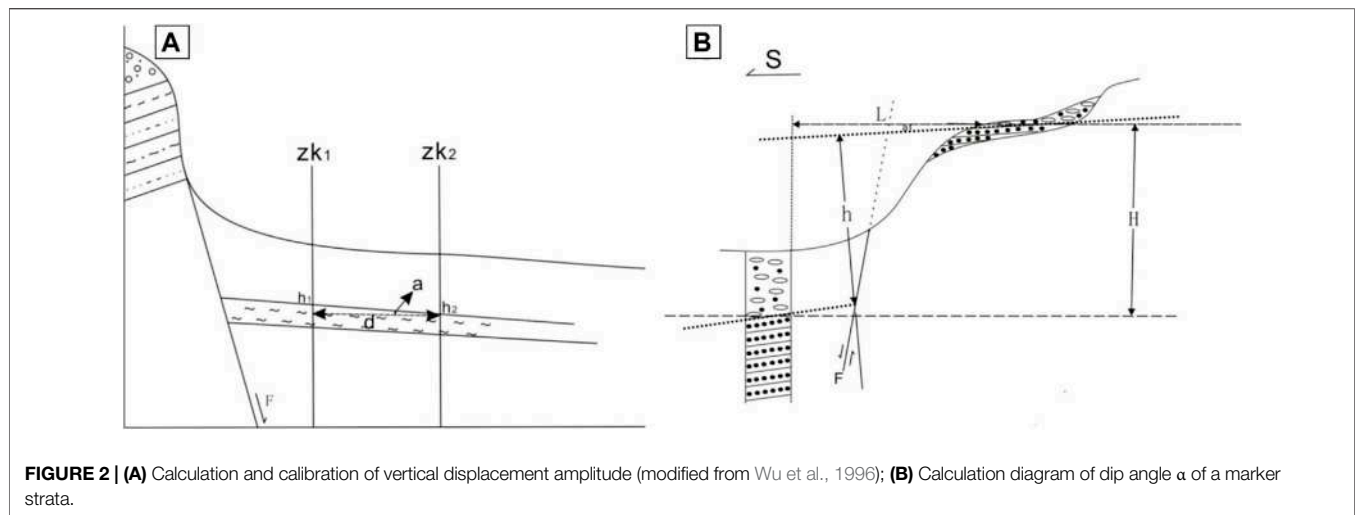


FIGURE 2 | (A) Calculation and calibration of vertical displacement amplitude (modified from Wu et al., 1996); **(B)** Calculation diagram of dip angle α of a marker strata.

3 METHODS

3.1 Vertical Displacement Calculation Method

The vertical displacement of a normal fault must be determined by comparing the same strata on both walls of the fault. The footwall terrace on the Hetao fault system has well-preserved stratigraphic outcrops, but the same strata in the hanging wall have been dissected and buried tens of metres below the surface. Therefore, only the stratigraphic sequence revealed by the borehole in the hanging wall of the fault can be compared to obtain more accurate fault displacement estimates. For the correlation between terrace strata and borehole strata, marker strata should be chosen first. The marker strata should meet the following conditions: (1) special lithology and special sedimentary tectonic strata or special interlayer interfaces; (2) no or very few similar layers or interfaces appear in the longitudinal direction; (3) the strata or interfaces are distributed stably in the lateral direction with little change in lithology or thickness; and (4) obvious features that are easy to identify. In addition to the correlation of marker strata, the ages of strata should be combined so that the activity rate can be calculated more accurately (Wu et al., 1996).

The vertical displacement obtained from the elevation difference of a marker strata is accurate only when the marker strata is horizontal. However, according to the field observations, all strata in this area have certain inclination angles, so it was necessary to calibrate the displacement obtained, as shown in Figure 2A. The real displacement can be calculated as follows: $h = (H - L \tan \alpha) \cos \alpha$, where L is the distance between the borehole and the terrace and α is the dip angle of the marker strata at the fault. H and L can be measured according to the elevation and distance of the borehole and the terrace, while a is measured according to the maker strata in the adjacent borehole near the fault.

However, according to the formula $h = (H - L \tan \alpha) \cos \alpha$, L has little effect, while α has a great effect on the correction of the vertical displacement. The α referred to by Wu et al. (1996) is

obtained from the field estimation of the dip angle of the marker strata in the terrace profile, which is quite different from the dip angle of the borehole lacustrine strata in the basin. In this paper, the ratio of the elevation difference between the borehole maker strata (m) and the spacing between boreholes (d) in the basin is adopted to calculate the dip angle of the maker strata, namely, $\alpha = \arctan(m/d)$, as shown in Figure 2B.

Additionally, $h_1 - h_2 = m$

$\tan \alpha = m/d$

$\alpha = \arctan(m/d)$

3.2 Real-Time Kinematic Field Terrace Survey

With the development of measurement technology, GPS RTK technology has gradually been applied to surveying and mapping in various industries (Wang et al., 2011). A Trimble R10 global navigation satellite system (GNSS) was used in this work. The system is a multichannel, multiband GNSS that integrates receivers, antennas and digital transmission stations. The horizontal accuracy of the instrument can reach $\pm 10 \text{ mm} + 1 \text{ ppm}$, and the vertical accuracy can reach $\pm 20 \text{ mm} + 1 \text{ ppm}$ under the RTK measurement mode. The initialization time is generally less than 10 s, and the initialization reliability is greater than 99.9%. Based on a large number of field explorations, field investigations and comprehensive remote sensing interpretations, to accurately determine the relative height of the terrace, the distribution of the terrace and the height of the steep ridge on the footwall of the fault on the northern margin of the Hetao Basin were systematically measured, and a large-scale topographic map was produced in this study.

3.3 Drilling and Terrace Chronological Methods

To obtain a complete stratigraphic age profile since the late Pleistocene on the hanging wall of the fault, drilling was

performed within 3 km of the fault, and the cores were described and sampled. The type of drilling rig used in this drilling was an XY-300, and the depth of a single hole was generally less than 100 m. During drilling, the holes were required to cross the fault, and the elevation of the top of the holes was recorded. After the core was removed, the bottom of each section of the core was marked with labels to indicate the depth and lithologic characteristics drilled thus far to facilitate core cataloguing. During core sampling, reliable radioactive carbon isotope dating materials, such as carbon fragments and terrestrial plant residues, that were found in the core were dated using ^{14}C dating, and samples of quartz-rich sand-bearing layers in the core were dated using optically stimulated luminescence (OSL) (Sun, 2012; Tao, 2020).

To determine the age of the piedmont terrace, photoluminescent samples were taken from the sedimentary strata of the piedmont terrace. The sampling principle of light-emitting samples was strictly followed during the sampling, and the preservation in the later period was also ensured using a strip that completely blocked light. All OSL sample analyses were completed at the OSL Laboratory of the National Institute of Natural Hazards. Sediment photoluminescence dating can be expressed as follows: age (A) = equivalent dose (DE)/environmental dose rate (D). Pretreatment with test sample preparation at the Key Laboratory of Crustal Dynamics, China Earthquake Administration, National Institute of Natural Hazards, in a dark room under weak light involved opening the packing and removing the iron pipe ends that may have been exposed and retaining the centre without contamination. For the samples that were used to determine the equivalent dose of exposure, 20 g were taken to measure the moisture content and saturation coefficient. Then, dry grinding was performed. The contents of U, Th and K in the samples were determined after the samples passed through a 63 μm sieve. For the measurement of equivalent dose, the sample photoluminescent irradiation and signal were measured on the Danish Risoeda-20-C/D thermal/photoluminescent automatic measurement system of the Crustal Dynamics Laboratory of China Earthquake Administration. The natural photoluminescent dose (i.e., palaeodose) of the sample was measured by a monolithic reproduction method. For the environmental dose rate measurement, the ambient radiation dose absorbed by the sample was provided by the ionizing radiation generated by the alpha, beta and gamma decay of the radionuclides (^{238}U , ^{232}Th and ^{40}K) in the sample itself and the surrounding sediments, with a small contribution from cosmic rays. The contents of U, Th and K in the samples were determined by element plasma mass spectrometry at the Beijing Research Institute of Uranium Geology. Finally, the age of the sample was calculated (Aitken, 1998; Murray and Wintle, 2003).

Since the development of the ^{14}C dating method, both theory and technology have matured from initial applications in archaeology to applications in geology (Qian, 2014; Zhao et al., 2017; Sanchez et al., 2018). The basic principle of the ^{14}C dating technique is to calculate the age of a sample according to the decay rate of ^{14}C atoms in the sample. Every living organism on Earth is involved in the exchange of carbon dioxide, and ^{14}C is present. Once an organism dies, there is no replacement of ^{14}C in the organism, and its original ^{14}C

concentration decreases exponentially over time. During sample dating, the age of biological death is calculated based on the existing ^{14}C content of the sample to determine the age of the sample (Gao et al., 2004). Sample preparation methods were as follows: first, a proper amount of samples was taken, and roots, leaves and other plant residues and carbon fragments were selected. The second step was to put the selected loose samples into a 15 ml centrifuge tube and stir them with dilute sulfuric acid to remove inorganic carbonate. The third step was to add 2% NaOH solution to remove humic acid and clean the sample until neutral. The fourth step was to dry the sample at a constant temperature of 55°C and grind and wrap it for testing (Tao, 2020).

4 RESULTS

4.1 Fault Displacements and Slip Rates Along the Sertengshan Piedmont Fault

4.1.1 Piedmont Terrace Ages

The sedimentary strata of the Sertengshan piedmont terrace are mainly alluvial-diluvial sand and gravel layers and yellow-green lacustrine layers (Figure 3D). According to the dating data of the terrace strata, the age of the yellow-green lacustrine layer top is approximately 70 ka (Figure 3B). Chen et al. (2008) dated the top of the lacustrine layer in the HTS4 borehole to 71.90 ± 7.34 ka, which is similar to the dating results in this paper in the front terrace. As is apparent from **Supplementary Table S1**, the sedimentary age of the lacustrine layer in the T3 terrace is between 50 and 70 ka. The sedimentary age of the T2 terrace lacustrine layer is between 20 and 40 ka. The age of the T1 terrace sedimentary strata is approximately 10 ka, and the sedimentary strata are mainly interbedded with clay and gravel. The age of the clay layer on the T1 terrace in northeastern Wujiahe town is 9.29 ± 1.11 ka (Figure 3A). The sedimentary strata on the T1 terrace in northern Dongquanzicun are mainly composed of sand and gravel layers, and the two OSL ages are 8.32 ± 0.96 ka and 9.79 ± 1.44 ka. Moreover, several faults can be found in the outcrop. The faults may extend to the surface (Figure 3C). Liu et al. (2014) carried out OSL dating of sedimentary strata in the Hetao Basin through drilling. We used the QK3 borehole, which had complete stratigraphic ages, as the standard borehole for comparison. According to the selection basis of marker strata, the yellow-green lacustrine top was regarded as the marker strata, and the upper and lower stratigraphic sequences are completely different, with obvious characteristics. Therefore, the late Pleistocene slip rate could be calculated based on the displacement and age of the lacustrine strata, while the Holocene slip rate could be calculated based on the Holocene stratigraphic thickness and T1 terrace height within the basin.

4.1.2 Vertical Slip Displacement of the Sertengshan Piedmont Fault

The multistage activities of the Sertengshan piedmont fault resulted in the formation of a multistage terrace, and the height of each terrace reflects the uplift height of the footwall of the faults in different periods. According to the RTK measurement results, the height of terraces in front of Sertengshan was calculated in this

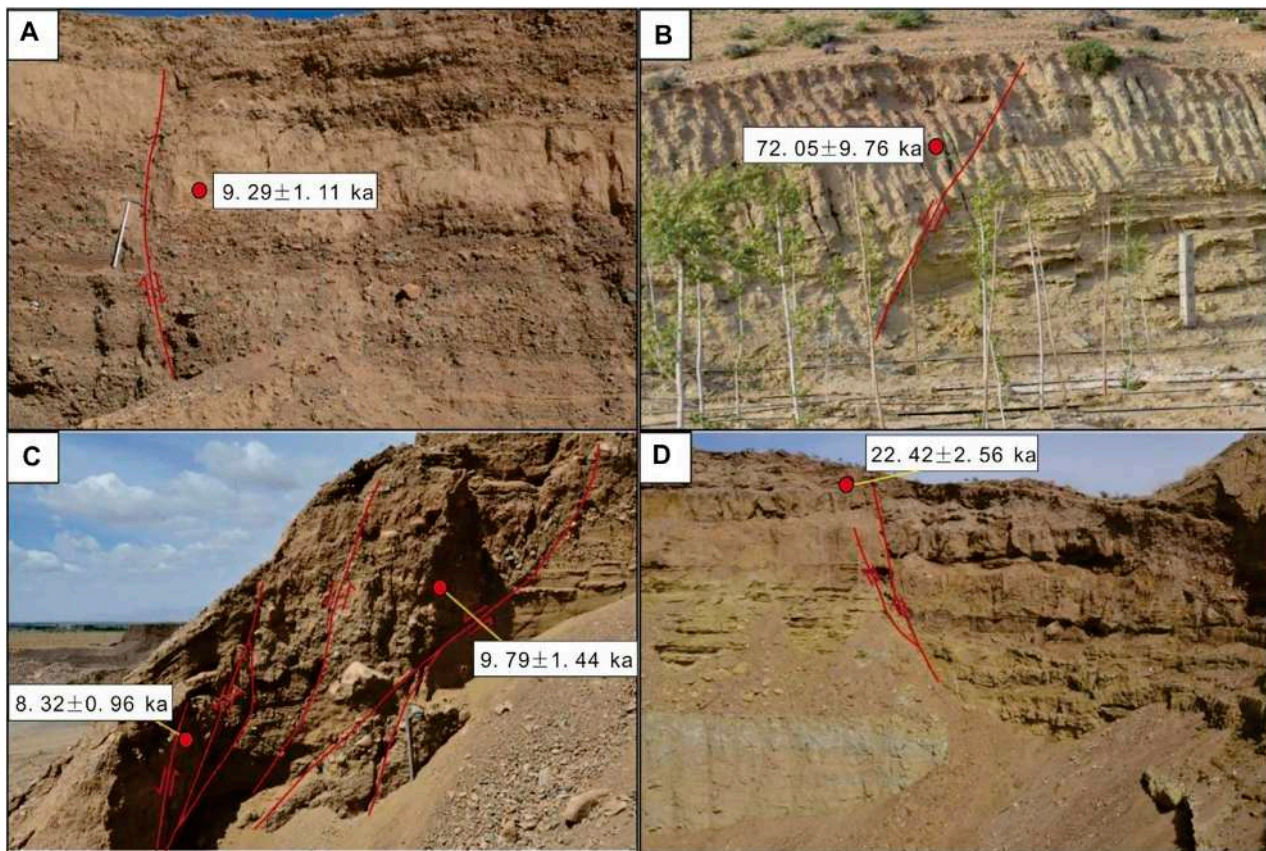


FIGURE 3 | Typical sampling points and ages in (A) northeastern Wujiahe town (looking SE); (B) Xiliuquanzixi (looking NW); (C) northern Dongquanzi (looking W); and (D) northeastern Dashetai town (looking SE).

paper. The results are shown in **Supplementary Table S2**. **Supplementary Figures S1–S3** show geological maps of the terraces associated with the Langshankou segment, Hongqicun segment and Kuluebulong segment.

As shown in **Supplementary Table S2**, the height of T1 terrace is approximately 7 m. Some of T1 terrace is relatively low because there are flood channels around it, and the height of terrace has been reduced due to flood erosion. The height of T2 terrace is approximately 14 m, the height of T3 terrace is approximately 26 m, and the height of T4 terrace is approximately 40 m. The Wujiahe segment and Hongqicun segment of the Sertengshan piedmont fault have more fourth-grade terraces and higher terraces, while the Kuluebulong segment and Dashetai segment have more developed third-grade terraces. Because the progression and height of the terrace can reflect the relative strength of activity of each segment, the activity of the Wujiahe and Hongqicun segments is the strongest, that of the Kuluebulong segment is weak, and that of the Dashetai segment is the weakest.

To obtain the displacement of the hanging wall, we used the borehole data of the Hetao Basin for comparative analysis. The stratigraphic age of the Hetao Basin is based on the age of QK3 (Liu et al., 2014). Boreholes VI-3 and QK3 are close to each other, so it can be considered that the sedimentary sequences of the two

boreholes are similar. We describe the lithology of borehole VI-3 in **Supplementary Table S7**. According to the borehole lithologic profile, it can be found that the Holocene strata are gravel layer and clay sand layer, and the Holocene strata in VIII-1 and VIII-1 are mainly fine sand layer and clay sand layer, indicating that the sedimentary environment in this area is relatively stable. In the four boreholes Zh18, Zk3, Zk6 and Zk7, the Holocene strata are relatively fine, with sand and gravel layers in the middle, indicating that the formation was formed under the interaction of strong flowing water and static water environment. The strata of VI-1, H12 and H141 are all gravel beds since 65 ka in the late Pleistocene, indicating that these three places have strong activity and strong flowing water environment. Due to the inhomogeneity of the activity of piedmont fault zone in space, the elevation of the top of lacustrine layer in the borehole is different, and the thickness of sedimentary strata is different in different time periods. According to the lithologic characteristics of each borehole and the selection principle of marker strata, we select the top of silt layer, clay sand layer and yellow-green fine sand layer as the top of lacustrine layer (**Figure 4A**). The ancient, large Jilantai-Hetao lakes existed from 50 to 60 ka (Chen et al., 2008). According to the borehole data on the Hetao Basin, the age of the top of the lacustrine layer is approximately 65 ka.

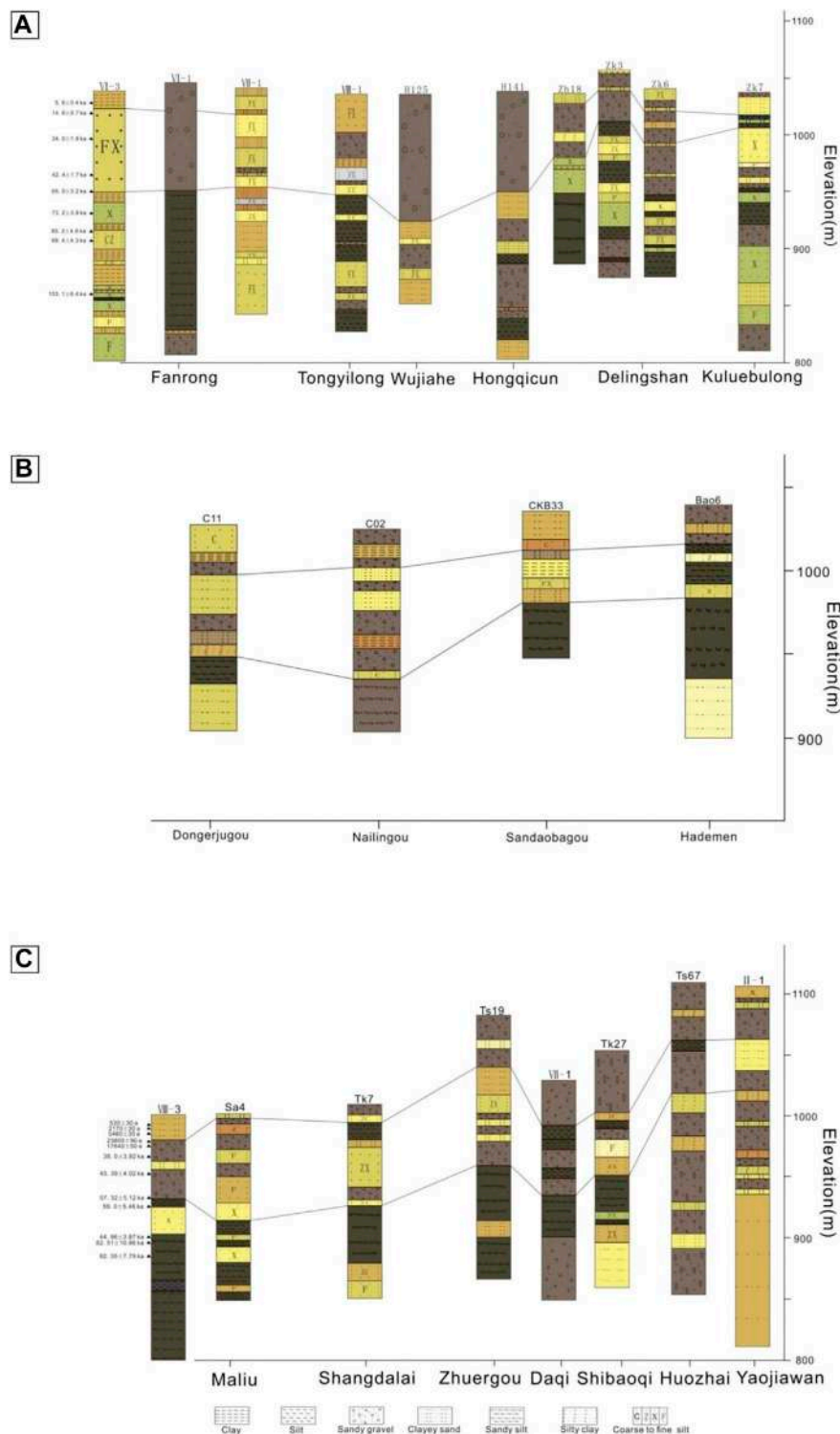


FIGURE 4 | (A) Column diagram of boreholes in front of Sertengshan; **(B)** Column diagram of boreholes in front of Wulashan; **(C)** Column diagram of boreholes in front of Daqingshan.

4.1.3 Slip Rate of the Sertengshan Piedmont Fault

According to the displacement calculation method, we used the above data from nine boreholes, combined with the RTK data and

the measured terrace height, to calculate the displacement of the fault. **Figure 5A** shows the elevation distribution of the terraces and boreholes. Based on these data and the elevation of the 65.0 ± 3.2 ka

surface of the lacustrine layer top in the borehole, the displacement ranges of the Sertengshan piedmont fault since the late Pleistocene can be calculated for different places. The displacement was calibrated according to the distance L between the borehole and the terrace and the dip angle α of the marker strata.

Table 1 shows the displacement at the nine measuring points since 65.0 ± 3.2 ka in the late Pleistocene, and the displacement h of each measuring point is as follows: 113.7 m in Fanrong, 94.2 m in Fengyucun, 102.6 m in Tongyilong, 117.4 m in Wujiahe, 108.2 m in Hongqicun, 63.5 m in Wubulangkou, 57.6 m in northern Delingshan, 61.1 m in Xishuiquancun and 48.2 m in Kuluebulong. The average slip rates are 1.75 mm/a in Fanrong, 1.45 mm/a in Fengyucun, 1.58 mm/a in Tongyilong, 1.81 mm/a in Wujiahe, 1.66 mm/a in Hongqicun, 0.98 mm/a in Wubulangkou, 0.89 mm/a in northern Delingshan, 0.94 mm/a in Xishuiquancun, and 0.74 mm/a in Kuluebulong. The Fanrongcun measuring point is located at the westernmost end of the Sertengshan piedmont fault. According to the trend of fault activity, its slip rate should be low, but in fact, its slip rate is high, which may be related to its proximity to the Langshan piedmont fault.

The stratigraphic thickness since the start of the Holocene cannot be determined by the sedimentary strata in three boreholes (VIII-1, H125 and H141), which may lead to errors in the correlation of marker strata. In addition, the Holocene strata are mainly alluvial-diluvial strata, so it is impossible to correct the vertical displacement of the Holocene marker strata for all boreholes according to the dip angle of the strata. A borehole without correction of Holocene displacement gives the maximum displacement amplitude during the Holocene. **Table 1** lists the displacement ranges in the Holocene at six measuring points. The displacement at each measuring point is 17.7 m in Fanrongcun, 20.6 m in Fengyucun, 10.5 m in Wubulangkou, 24.4 m in northern Delingshan, 12.6 m in Xishuiquancun and 22.2 m in Kuluebulong. The corresponding average slip rates during the Holocene are 1.48 mm/a in Fanrongcun, 1.72 mm/a in Fengyucun, 0.86 mm/a in Wubulangkou, 2.03 mm/a in northern Delingshan, 1.05 mm/a in Xishuiquancun, and 1.85 mm/a in Kuluebulong. In the Tongyilong, Wujiahe and Hongqicun boreholes, the Holocene stratigraphic boundary cannot be clearly demarcated, but according to the slip rate since the late Pleistocene, the Holocene stratigraphic thickness is approximately 15 m in the Tongyilong, approximately 18 m in Wujiahe, and approximately 16 m in Hongqicun. Considering the elevation of the borehole and terrace, the average slip rates at Tongyilong, Wujiahe and Hongqicun during the Holocene are approximately 2.03, 2.28 and 1.86 mm/a, respectively.

4.2 Fault Displacements and Slip Rates Along the Wulashan Piedmont Fault

4.2.1 Elevations and Stratigraphic Ages of the Piedmont Terrace

To analyse the characteristics of the piedmont terrace of Wulashan, we conducted a field exploration (**Figure 6**). Due to the lack of data along the Wulateqianqi segment, we summarized the elevations and ages of terraces around the other two segments. Among them, the T1 terrace from

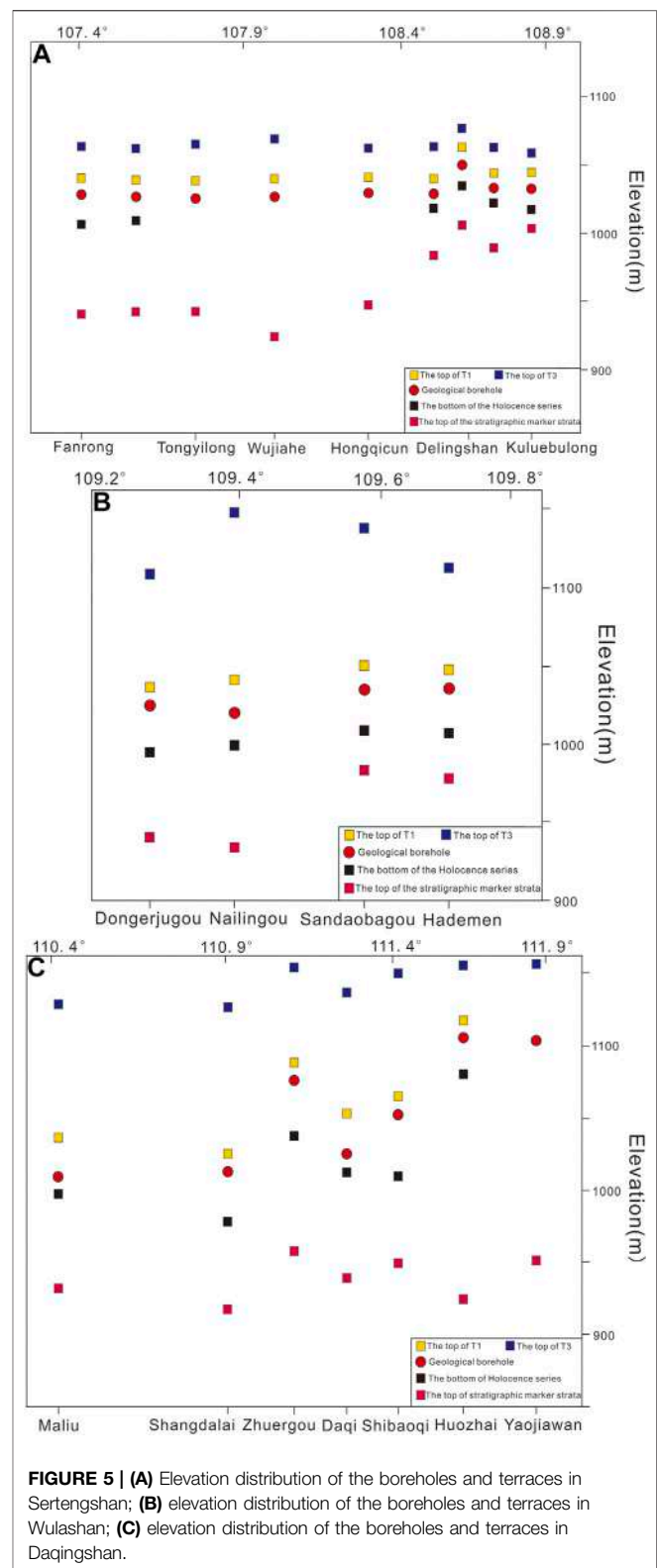
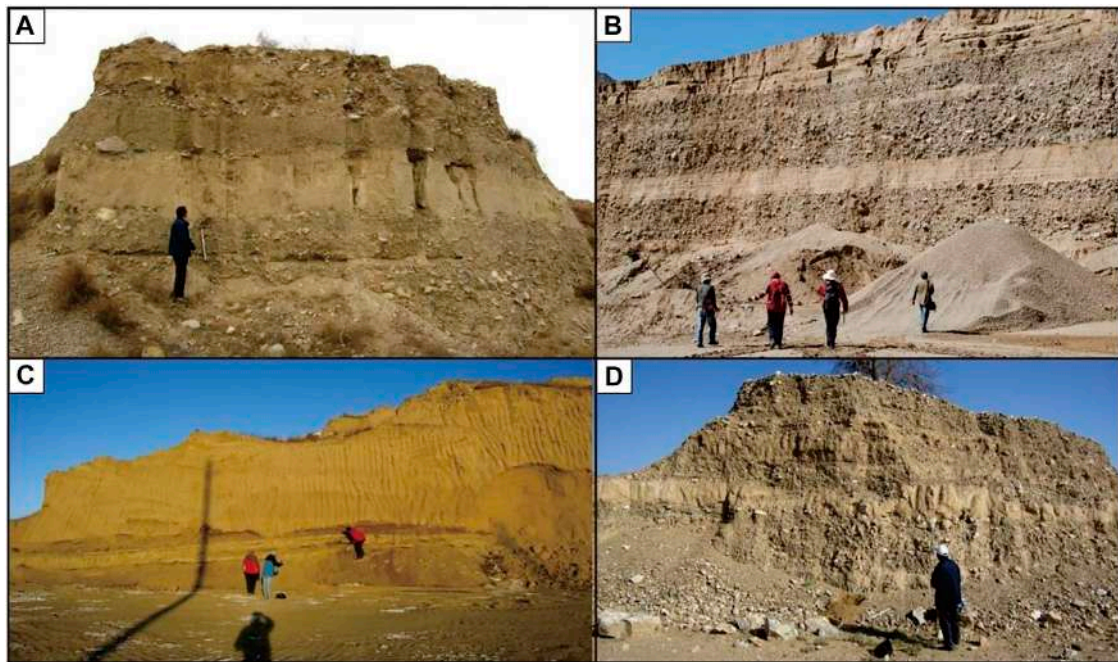


FIGURE 5 | (A) Elevation distribution of the boreholes and terraces in Sertengshan; (B) elevation distribution of the boreholes and terraces in Wulashan; (C) elevation distribution of the boreholes and terraces in Daqingshan.

Gongmiaozi to Heshunzhuang is not well developed, the yellow-green sand layer with horizontal bedding is exposed on the front edge of the terrace in some sections, and the elevation of

TABLE 1 | Vertical displacement of the Sertengshan piedmont fault.

Location	Fanrong	Fengyu	Tongyilong	Wujiahe	Hongqicun	Wubulangkou	Delingshan	Xishuiquan	Kuluebulong
Borehole	VI-1	VII-1	VIII-1	H125	H141	ZH18	ZK3	ZK6	ZK7
T3 elevation (m)	1,062.5	1,060.4	1,063.0	1,064.0	1,060.0	1,061.0	1,068.0	1,060.0	1,058.0
T1 elevation (m)	1,042.5	1,042.0	1,045.0	1,045.5	1,043.3	1,041.0	1,061.0	1,043.2	1,043.0
Wellhead elevation (m)	1,040.6	1,036.2	1,035.6	1,036.1	1,037.0	1,037.3	1,055.7	1,039.8	1,036.4
Elevation of the Holocene series in the footwall (m)	1,017.9	1,021.4	—	—	—	1,030.5	1,035.2	1,026.2	1,018.2
Vertical displacement of the Holocene series (m)	24.6	20.6	—	—	—	10.5	25.8	17.0	24.8
Holocene vertical displacement after correction (m)	17.7	—	—	—	—	—	24.4	12.6	22.2
Elevation of the top of the lacustrine strata (m)	941.9	942.3	942.6	925.0	947.6	984.1	1,009.0	994.5	1,007.2
Displacement of the stratigraphic marker strata since 65 ka (m)	120.6	118.1	120.4	139.0	112.4	76.9	59.0	65.5	50.8
Vertical displacement after correction (m)	113.7	94.2	102.6	117.4	108.2	63.5	57.6	61.1	48.2

**FIGURE 6 |** (A) Holocene stratigraphic section in Shaota (NW), (B) Hademengou diluvial fan stratigraphic section, (C) East Kundulun River stratigraphic section (NE), and (D) Huhe Braggol Holocene stratigraphic section.

the terrace is 1,022–1,035 m. T2 is a pedestal terrace with granite gneisses in the basement and fluvial and lacustrine conglomerate in the upper part. It forms a continuous belt along the Wulashan piedmont fault with an elevation of 1,060–1,100 m. The T3 terrace is distributed intermittently in front of Wulashan, and the elevation is approximately 1,135 m. (dos Massif, 1988; Institute of Crustal and Dynamics and China Earthquake Administration, 2013; Research Group of Active Fault System around Or).

There have been many studies on the formation age of the terrace (**Supplementary Table S3**). Based on the analysis of the dating results of the samples collected in this work, we

preliminarily determined that the formation age of the T1 terrace was the early Holocene, approximately 7–12 ka; the T2 terrace formed in the late Pleistocene, approximately 30–40 ka; and the T3 terrace formed in the middle of the latest Pleistocene, approximately 50–60 ka (He et al., 2020).

4.2.2 Vertical Slip Displacement of the Wulashan Piedmont Fault

To accurately determine the relative heights of the terraces, large-scale topographic maps of Xishanzui, Gongmiaozi, Dalagou, Meigeng, Hayehutong, Hademen and other places were produced with GPS RTK. The relative heights of various terraces along the

Wulashan piedmont fault are listed in **Supplementary Table S4**. **Supplementary Figures S5–S7** show the topographic maps and terrace data around Heshunzhuang, Hayehuutong and Dabagou.

To calculate the vertical displacement amplitude at each fault point, in this paper, large-scale GPS RTK-based topographic maps and borehole locations were used to estimate the corresponding elevations of T1 and T3 in Dongerjugou, Nailinggou, Sandaobagou and Hademen. The results show that the elevations of T1 and T3 in Dongerjugou are 1,028.6 and 1,114.5 m, respectively. The elevations of T1 and T3 in Nailinggou are 1,040 and 1,150 m, respectively. In Sandaobagou, the elevation of T1 is 1,050 m, and the elevation of T3 is 1,140 m. In Hademen, the elevation of T1 is 1,045 m, and that of T3 is 1,120 m.

To obtain the displacement of the Wulashan piedmont fault, four boreholes in the hanging wall of the fault are used for comparative analysis (**Figure 4B**). According to **Figure 4B**, the Holocene strata in the four boreholes are mainly gravel, coarse sand, clay and clay sand. The Bao6 borehole is mainly sand and gravel layer, indicating that the sedimentary environment of the borehole is strong flowing water. The clay layer in C11 and C02 is overlying the sand and gravel layer, which indicates that the sedimentary environment tends to be stable in this period. The coarse sand layer and gravel layer in the uppermost layer are the result of alluviation. The Holocene strata of CKB33 are mainly coarse sand and clay sand, indicating that the sedimentary environment is relatively stable. The top of lacustrine layer of the four boreholes is the top of silt and sandy silt layer, with an elevation of 900–1000 m, which is close to the top of lacustrine layer of the boreholes in Daqingshan and Sertengshan. A great deal of work has been performed to date the terrace in front of Wulashan; the age of the T1 terrace is approximately 10–12 ka, and that of the T3 terrace is approximately 60 ka. According to the analysis of the piedmont boreholes in Daqingshan and Sertengshan, the ages of the top of the lacustrine layer are 65 ± 3.2 ka and 57.32 ± 5.12 ka, respectively, which are in the same period. Although the borehole on the hanging wall of the fault lacks chronological information, it belongs to the northern margin of the Hetao fault system and is located on the northern margin of the ancient lake. Therefore, it can be inferred that the age of the top of the lacustrine layer in the borehole on the southern side of the Wulashan piedmont fault is approximately 60 ka.

4.2.3 Slip Rate of the Wulashan Piedmont Fault

Figure 5B shows the elevation distribution of the terrace and borehole. Based on the elevation of the T3 terrace and the lacustrine layer in the borehole, the displacement of each measuring point can be calculated since the late Pleistocene. The displacement was calibrated according to the distance L between the borehole and the terrace and the dip angle α of the marker strata (**Table 2**).

Table 2 shows the displacement at four measuring points in the latest Pleistocene. Since the latest Pleistocene (60 ka), the displacement h at each measuring point is as follows: 145.3 m in Dongerjugou, 186.8 m in Nailinggou, 153 m in Sandaobagou and 128.24 m in Hademen. The corresponding average slip rates are 2.42 mm/a in Dongerjugou, 3.11/a in Nailinggou, 2.55/a in

Sandaobagou and 2.14 mm/a in Hademen. Dongerjugou is located in the Gongmiaozi segment. In western Dongerjugou, the slip rate cannot be calculated due to the lack of borehole data. Analysis of the Heshunzhuang-Baotou segment shows that the fault slip rate has been decreasing from west to east since the late Pleistocene (60 ka).

Similar to the Holocene strata near Sertengshan, the Holocene strata in the basin on the southern side of Wulashan are mainly alluvial and diluvial. However, the Holocene strata in the borehole are easy to distinguish, so the displacement of Holocene marker strata can be corrected. The displacements at the four Holocene measuring points in **Table 2** are 21.5 m in Dongerjugou, 29.5 m in Nailinggou, 34 m in Sandaobagou and 22.4 m in Hademen. The corresponding Holocene slip rates are 1.84 mm/a in Dongerjugou, 2.52 mm/a in Nailinggou, 2.91 mm/a in Sandaobagou, and 1.91 mm/a in Hademen. Different from the late Pleistocene slip rate, the Holocene slip rate increases from west to east, but the slip rate is relatively small. The activity of the Wulashan piedmont fault since the late Pleistocene is stronger than that during the Holocene.

4.3 Fault Displacements and Slip Rates Along the Daqingshan Piedmont Fault

4.3.1 Vertical Slip Displacement of the Daqingshan Piedmont Fault

Since the late Pleistocene, the Daqingshan piedmont fault has formed three stages of terraces in the piedmont. The T1 terrace is an accumulative terrace, and the front of the terrace is a Holocene fault scarp. The T2 and T3 terraces are pedestal terraces (The Institute of Crustal and Dynamics and China Earthquake Administration, 1994) carried out research, including 1:50000 geological mapping, on the active Daqingshan piedmont fault and collected thermoluminescence (TL) age samples to determine the formation age of the terrace. The dating results suggest that the formation age of the T3 terrace was late Pleistocene. The age of the T1 terrace is approximately 11 ka, indicating Holocene formation. Therefore, the elevations of the T3 terrace and T1 terrace of the Daqingshan piedmont fault were used as marker elevations to calculate the vertical displacement of the Daqingshan piedmont fault in the late Pleistocene and Holocene.

In this paper, according to the measurement results of previous studies, the elevations of the T1 and T3 terraces in the Daqingshan piedmont fault were collected, and the results are shown in **Supplementary Table S5**.

As shown in **Supplementary Table S5**, the T3 terrace is continuously distributed along the Daqingshan piedmont fault and is largely complete. The relationship between the T2 and T3 terraces is erosional, and the development is discontinuous along the front of Daqingshan. T1 terraces are relatively well developed in the western segment of Tuyuqi and the western segment of Tuzuqi, but the T1 and T2 terraces are not well developed in eastern Bikeqi.

To obtain the elevation of the marker strata in the hanging wall, we used boreholes in the Hetao Basin to carry out a comparative analysis. Because borehole VIII-3 is close to ZK2, it can be considered that the sedimentary sequences of the two boreholes are similar. **Supplementary Table S21** shows the

TABLE 2 | Vertical displacement of the Wulashan piedmont fault.

Location	Dongerjugou	Nailingou	Sandaobagou	Hademen
Borehole	C11	C02	CKB33	Bao6
T3 elevation (m)	1,114.5	1,150	1,140	1,120
T1 elevation (m)	1,028.5	1,040	1,050	1,045
Wellhead elevation (m)	1,026.1	1,024.4	1,036.9	1,040.9
Elevation of the Holocene series in the footwall (m)	998	1,005	1,014	1,011
Vertical displacement of the Holocene series (m)	30.5	35	36	34
Holocene vertical displacement after correction (m)	21.5	29.5	34	22.4
Elevation of the top of the lacustrine strata (m)	947	935	985	980
Displacement of the stratigraphic marker strata since 65 ka (m)	167.5	215	155	140
Vertical displacement after correction (m)	145.3	186.8	153	128.24

lithology of borehole VIII-3. **Supplementary Figure S7** shows field photographs and a lithologic histogram for borehole ZK2. **Figure 4C** shows 7 borehole lithologic profiles of Daqingshan piedmont fault. The Holocene strata in Sa4 are mainly sandy clay layers, indicating that the sedimentary environment is relatively stable. The Holocene strata in the five boreholes Tk7, Ts19, VIII-1, Tk27 and TS67 are mainly gravel and middle coarse sand, with large grain size, indicating that they were formed under the action of strong water flow. The upper part of the II-1 is composed of fine sand and silty clay, and the lower part of it is composed of sand and gravel, indicating that the sedimentary environment in the late Holocene was relatively stable after the strong flowing water. The top of lacustrine layer from Sa4 to Zk27 is silt layer or sandy silt layer. The top of yellow-green sand layer is selected as the top of lacustrine layer in Ts67 and II-1.

The ages from borehole ZK2 were adopted. The age of the top of the lacustrine layer is approximately 58 ka, and the age of the bottom of the Holocene layer is approximately 11 ka.

4.3.2 Slip Rate of the Daqingshan Piedmont Fault

Figure 5C shows the elevation distribution of the terrace and borehole at 7 measuring points, and the displacement h at each measuring point since the late Pleistocene (57.32 ± 5.12 ka) is as follows: 195.21 m in Maliu, 145 m in Shangdalai, 186 m in Zhuergou, 187.98 m in Daqi, 188.15 m Shibaoqi, 225 m in Huozhai and 198.89 m in Yaojiawan (**Table 3**). The corresponding average slip rates are as follows: 3.37 mm/a in Maliu, 2.50 mm/a in Shangdalai, 3.21 mm/a in Zhuergou, 3.24 mm/a in Daqi, 3.25 mm/a in Shibaiqi, 3.88 mm/a in Huozhai, and 3.43 mm/a in Yaojiawan. Due to the lack of borehole data in the Baotou segment of the Daqingshan piedmont fault, the vertical slip rate has not been measured. However, according to the slip rates measured at seven measuring points, the average slip rate of the Daqingshan piedmont fault has no obvious regularity in terms of spatial distribution. Wu et al. (1996) calculated the average slip rates along the Baotou segment of the Daqingshan piedmont fault, with those in Wudangou (4.49 mm/a) and Wanshuiquan (2.61 mm/a) being consistent with those in this study.

The Holocene strata in the borehole are easy to distinguish, so the displacement of Holocene marker strata can be corrected. Due to the lack of the T1 terrace in the Hohhot segment of the Daqingshan piedmont fault, the vertical displacement cannot be calculated. Therefore, we use only six measuring points to

calculate the Holocene vertical displacement. The h at each measuring point is as follows: 26.21 m in Maliu, 31 m in Shangdalai, 27 m in Zhuergou, 19.58 m in Daqi, 31.15 m in Shibaoqi and 27 m in Huozhai. The corresponding average slip rates are 2.37 mm/a in Maliu, 2.82 mm/a in Shangdalai, 2.45 mm/a in Zhuergou, 1.78 mm/a in Daqi, 2.83 mm/a in Shibaoqi, and 2.45 mm/a in Huozhai.

5 DISCUSSION

5.1 Interpretation of Fault Slip Rate and Revelation of Seismic Hazard Along the Northern Margin of the Hetao Basin

Much work has been conducted on the slip rates of faults along the northern margin of the Hetao Basin. However, the calculation of fault slip in this region has been simplified. In the 1990s, the state seismological bureau organized the study entitled “1:50000 Geological Mapping and Comprehensive Research of Langshan-Sertengshan Piedmont Fault.” For the Sertengshan piedmont fault, the calculated maximum slip rate since the late Pleistocene was 3.6 mm/a, and the calculated slip rate during the Holocene was 2.2 mm/a (Institute of Crustal and Dynamics and China Earthquake Administration, 1994). Ma et al. (1998) divided the piedmont terrace of Wulashan into high terrace and low terrace areas. They believed that the relative height of the high terrace areas represents the uplift displacement of the fault since the late Pleistocene and that of the low terrace areas represents the uplift displacement of the fault during the Holocene. Based on the relative height and thermoluminescence dating data of the high and low terraces, the uplift rate of the fault footwall is 0.5–2 mm/a since the late Pleistocene and 0.4–1.0 mm/a during the Holocene. Wu et al. (1996) obtained the displacement of the fault in the latest Pleistocene by comparing the footwall terrace of the Daqingshan piedmont fault with the corresponding marker strata in a borehole in the hanging wall; they then obtained the corrected displacement as the vertical displacement of the fault according to the dip angle of the marker strata. According to the thermoluminescence age data of the terrace, the vertical slip rate of the fault since the latest Pleistocene is 2.61–4.49 mm/a. To approximate the Holocene displacement, the free-face ratio of height to the distance between the upper and lower fault scarps is calculated, and the vertical slip rate of the fault during the

TABLE 3 | Vertical displacement of the Daqingshan piedmont fault.

Location	Maliu	Shangdalai	Zhuergou	Daqi	Shibaoqi	Huozhai	Yaojiawan
Borehole	Sa4	TK7	TS19	VII-1	TK27	TS67	II-1
T3 elevation (m)	1,128	1,120	1,154	1,143	1,150	1,155	1,170
T1 elevation (m)	1,039	1,056	1,083	1,052	1,058	1,112	
Wellhead elevation (m)	1,014.1	1,011	1,078	1,030.7	1,054.4	1,109.7	1,108.21
Elevation of the Holocene series in the footwall (m)	1,008	980	1,048	1,020	1,015	1,080	—
Vertical displacement of the Holocene series (m)	31	76	35	32	43	32	—
Holocene vertical displacement after correction (m)	26.21	31	27	19.58	31.15	27	—
Elevation of the top of the lacustrine strata (m)	928	930	960	942.6	950	925	952.21
Displacement of the stratigraphic marker strata since 65 ka (m)	200	190	194	200.4	200	230	217.79
Vertical displacement after correction (m)	195.21	145	186	187.98	188.15	225	198.89

Holocene is 0.37–1.72 mm/a. The Langshan piedmont fault is located on the northwestern margin of the Hetao Basin. Much work has been conducted on the fault activity of the Langshan piedmont fault. The fault dislocated the bottom layer of the Holocene and forced the Yellow River to migrate southward in modern times, with a vertical slip rate of 0.47–2.2 mm/a in the Holocene (Sun et al., 1990; Deng et al., 1999). He et al. (2014) inferred that the uplift rate since 47.4 ka was 0.81–1.8 mm/a by studying typical profiles of terraces along the Langshan piedmont fault. He et al. (2015) conducted a systematic study on ten debris flow gullies in the central and southern parts of the region, combined with an assessment of tectonic movement, and concluded that the uplift rate of the middle segment of the Langshan piedmont fault was 2.8 mm/a. (Dong, 2015) excavated multiple trenches in the Langshan piedmont fault and concluded that the vertical slip rate of Qingshan town was 1.33–1.44 mm/a in the Holocene. However, we believe that the vertical slip rate estimated based on the dislocation of the same strata in trenches or terrace elevations are not representative of the entire fault, and the calculated rate is too small. Liang (2019) re-estimated the vertical slip rate of the Langshan piedmont fault since the late Pleistocene based on the throw between the top of the same lacustrine strata on both sides of the fault and concluded that the vertical slip rate of the Langshan piedmont fault since the late Pleistocene was 1.8–3.2 mm/a, which is more accurate.

In previous studies of the fault slip rate on the northern margin of the Hetao Basin, calculations were mostly based on the uplift rate of the faults. In addition, the calculation of the fault uplift rate by directly using terrace height and age without considering the influence of terrace sedimentary cover will lead to inaccurate results. Moreover, due to the limitations of dating technology, thermoluminescence dating is the most common method to obtain age estimates. The dating accuracy of this method is lower than that of OSL dating. These limitations lead to the underestimation of fault activity and earthquake risk. Most previous studies calculated the slip rates of faults by the cumulative dislocation of palaeoseismic events in profiles and troughs but did not consider the creep slip rate of faults between ruptured earthquakes. Studies have shown that long-term deformation monitoring across faults using high-precision measurement arrays, such as short-range leveling, short baseline and creep meter surveys, is an effective method for obtaining the current activity mode and behaviour of faults (Savage et al., 1979; Bevis and Isacks, 1981; Lisowski and

Prescott, 1981; Schulz et al., 1982; Lee et al., 2001; Lienkaemper et al., 2001, 2013, 2014; Du et al., 2010; Zhang et al., 2018). Yang et al. (2021) analysed the activity of the Kouzhen-Guanshan fault in the Weihe Basin from the late Quaternary to the present based on field investigation results and cross-fault deformation data and concluded that the vertical creep rate of the eastern segment of the fault is 1.56 mm/a, suggesting that a long period of cross-fault deformation monitoring is needed to thoroughly study the activity of a fault. The Hetao Basin and Weihe Basin in this study area both fault-controlled basins around the Ordos block, and their fault activity models are similar. Therefore, we believe that the vertical slip rates of these normal faults constrained by terraces and boreholes consist of both seismic stick-slip and interseismic creep slip rates. Considering the monitoring data of cross-fault deformation in the study area, we will further study the fault creep rates of the faults in the future.

The vertical displacement calculated in this paper by using the marker strata of the borehole and terrace can represent the real displacement of the fault and is not affected by the surface sediment cover. Therefore, the calculated vertical slip rate is the complete vertical slip rate of the fault, which can provide support for evaluating the seismic hazard of the fault. In addition, the vertical slip rates obtained in this paper include the interseismic creep slip rates of the faults, which will be further studied.

Table 4 shows that since 65 ka in the late Pleistocene, the average slip rates of the three segments of the Sertengshan piedmont fault are 1.65 mm/a in the Langshankou segment, 1.32 mm/a in the Hongqicun segment and 0.86 mm/a in the Kuluebulong segment, showing a decreasing trend from west to east. Since the late Pleistocene (60 ka), the average slip rates of the two segments of the Wulashan piedmont fault have been 2.42 mm/a in the Gongmiaozi segment and 2.6 mm/a in the eastern Baotou segment. The average slip rates of the segments show an increasing trend from west to east. Since the late Pleistocene (58 ka), the average slip rates of the four segments of the Daqingshan piedmont fault are as follows: 3.37 mm/a in the Tuyouxi segment, 2.50 mm/a in the Tuzuoqi segment, 3.40 mm/a in the Bikeqi segment and 3.43 mm/a in the Hohhot segment. Among them, the average slip rate in the Bikeqi segment is the highest, and that in the Tuzuoqi segment is the lowest. Moreover, the maximum and minimum slip rates at each measuring point also appear in the Bikeqi segment and the Tuzuoqi segment, respectively. The vertical slip rate at each measuring point in the Bikeqi segment presents an increasing trend

TABLE 4 | Comparison of late Pleistocene slip rates of active faults on the northern margin of the Hetao Basin.

Name of the fault	Segment	Measuring point	Vertical slip rate at the point	Average slip rate of the segment	Average slip rate of the fault
Sertengshan piedmont fault	Langshankou segment	Fanrong	1.75	1.65	1.28
		Fengyu	1.45		
		Tongyilong	1.58		
		Wujiahe	1.81		
	Hongqicun segment	Hongqicun	1.66	1.32	
		Wubulangkou	0.98		
	Kuluebulong segment	Delingshan	0.89	0.86	
		Xishuiquan	0.94		
		Kuluebulong	0.74		
Wulashan piedmont fault	Gongmiaozi segment	Dongerjugou	2.42	2.42	2.51
	Baotou segment	Nailingou	3.11	2.6	
		Sandaobagou	2.55		
		Hademen	2.14		
Daqingshan piedmont fault	Tuyouxi segment	Maliu	3.37	3.37	3.34
	Tuzuoxi segment	Shangdalai	2.50	2.50	
	Bikeqi segment	Zhuergou	3.21	3.40	
		Daqi	3.24		
		Shibaoqi	3.25		
		Huozhai	3.88		
	Hohhot segment	Yaojiawan	3.43	3.43	

from west to east. By comparing the average slip rates of the three faults since the late Pleistocene, it can be found that the average slip rate of the Daqingshan piedmont fault is the highest (3.34 mm/a), and the average slip rate of the Sertengshan piedmont fault is the lowest (1.28 mm/a). According to the spatial distribution of the three faults on the northern margin of the Hetao Basin, the activity of the three active faults increases from west to east.

Table 5 shows that since 12 ka, the average slip rates of the three segments of the Sertengshan piedmont fault are 1.88 mm/a (Langshankou segment), 1.36 mm/a (Hongqicun segment) and 1.64 mm/a (Kuluebulong segment), showing no obvious spatial pattern. The average slip rates of the two segments of the Wulashan piedmont fault are 1.84 mm/a in Gongmiaozi and 2.45 mm/a in Baotou, showing an increase from west to east. Since 11 ka, the average slip rates of the three segments of the Daqingshan piedmont fault are 2.37 mm/a in the Tuyouxi segment, 2.82 mm/a in the Tuzuoxi segment and 2.38 mm/a in the Bikeqi segment. The average slip rate of each segment increases from west to east. By comparing the average slip rates of three active faults on the northern margin of the Hetao Basin, it can be found that since the Holocene, the average slip rate of the Daqingshan piedmont fault is the highest (2.45 mm/a), and that of the Sertengshan piedmont fault is the lowest (1.68 mm/a). Similar to that during the late Pleistocene, the activity of the three faults during the Holocene increases from west to east.

By comparing **Tables 4, 5**, it can be found that the average slip rate of each segment of the Sertengshan piedmont fault since the late Pleistocene is less than that during the Holocene; that is, the fault activity of the Sertengshan piedmont fault increased from the late Pleistocene to the Holocene. The average slip rates of the Daqingshan piedmont fault and Wulashan piedmont fault since the late Pleistocene are greater than those during the Holocene; that is, the fault activity of the Daqingshan piedmont fault and

Wulashan piedmont fault decreased from the late Pleistocene to the Holocene.

Table 6 summarizes the elapsed time and mean recurrence interval of the last palaeoseismic event of the main active fault segments in the Hetao fault zone. By consulting the recurrence characteristics of palaeoearthquakes and the vertical slip rates obtained in this paper, it can be found that in the Langshankou and Hongqicun segments of the Sertengshan piedmont fault, the elapsed time since the latest earthquake is close to or longer than the recurrence interval and that the activity of the fault segments increased from the late Pleistocene to the Holocene, suggesting that the future seismic risk of the two fault segments is high. In contrast, the elapsed times of the Kuluebulong and Dashetai segments are less than the average recurrence interval, and the late Pleistocene-Holocene fault segments are not currently active, so we can conclude that the Kuluebulong and Dashetai segments will have low earthquake risk in the future. It is difficult to estimate the seismic risk along the Wulateqianqi segment of the Wulashan piedmont fault because of the lack of a complete palaeoearthquake sequence and earthquake recurrence interval. However, the elapsed time since the last palaeoearthquake in the Baotou segment is close to the average recurrence interval, and the slip rate is high, so the possibility of future earthquakes is high. In addition, the elapsed time since the last palaeoearthquake in the Gongmiaozi segment is far less than the average recurrence interval, and the slip rate of the fault segment is not high, so the risk of future earthquakes is low. Among the fault segments of the Daqingshan piedmont fault, the characteristic earthquake recurrence interval and slip rate of the Baotou segment were not obtained, so it is difficult to estimate the seismic risk for this segment. Although the elapsed time since the latest palaeoearthquake in the Tuyouxi segment is less than the recurrence interval, the slip rate of this fault section is high, so the seismic risk of this segment cannot be ignored. In addition, the

TABLE 5 | Comparison of the Holocene slip rates of active faults on the northern margin of the Hetao Basin.

Name of the fault	Segment	Measuring point	Vertical slip rate at the point	Average slip rate of the segment	Average slip rate of the fault
Sertengshan piedmont fault	Langshankou segment	Fanrong	1.48	1.88	1.68
		Fengyu	1.72		
		Tongyilong	2.03		
		Wujiahe	2.28		
	Hongqicun segment	Hongqicun	1.86	1.36	
		Wubulangkou	0.86		
	Kuluebulong segment	Delingshan	2.03	1.64	
		Xishuiquan	1.05		
		Kuluebulong	1.85		
Wulashan piedmont fault	Gongmiaozi segment	Dongerjugou	1.84	1.84	2.30
	Baotou segment	Nailingou	2.52	2.45	
		Sandaobagou	2.91		
		Hademen	1.91		
Daqingshan piedmont fault	Tuyouxi segment	Maliu	2.37	2.37	2.45
	Tuzuoxi segment	Shangdalai	2.82	2.82	
	Bikeqi segment	Zhuergou	2.45	2.38	
		Daqi	1.78		
		Shibaoqi	2.83		
		Huozhai	2.45		
		Hohhot segment	Yaojiawan		

TABLE 6 | The elapsed time since the last palaeoseismic event and mean earthquake recurrence interval for segments along the northern margin of the Hetao Basin.

Name of the fault	Segment	Elapsed time (a B.P.)	Mean recurrence interval (a)	Data source
Sertengshan piedmont fault	Langshankou segment	4190 ± 250	4125	Chen (2002)
	Hongqicun segment	3250 ± 250	4105	Chen (2002)
	Kuluebulong segment	2320	4105	Zhang (2017)
	Dashetai segment	3590	4093	He et al. (2018)
Wulashan piedmont fault	Wulateqianqi segment	<17000		Previous study ^a
	Gongmiaozi segment	1655 ± 185	8533	Ran et al. (2003)
	Baotou segment	4130 ± 78	4314	Chen (2002)
Daqingshan piedmont fault	Baotou segment	3390 ± 2140		Ran et al. (2003)
	Tuyouxi segment	1165	1788	He and Ma (2015)
	Tuzuoxi segment	>10,790	2948 ± 560	Ran et al. (2003)
	Bikeqi segment	2867 ± 679	2763 ± 518	He (2006)
	Hohhot segment	4500 ± 230	2642 ± 412	He (2006)

^aInstitute of Crustal and Dynamics and China Earthquake Administration (2013).

elapsed times since the last palaeoearthquake in the Tuzuoxi, Bikeqi and Hohhot segments are longer than the average recurrence intervals, and the slip rates of those fault segments are high; thus, we conclude that these three segments have high seismic risk.

5.2 Fault Slip Rate and Regional Tectonic Dynamics Along the Northern Margin of the Hetao Basin

The faulted basins around the Ordos block are considered to be the common result of the westward subduction of the Pacific plate and the collision between the Indian plate and Eurasian plate (Liu et al., 2004; Northrup et al., 1995; Tian et al., 1992; Xu et al., 2019). The sedimentary-tectonic evolution history of Cenozoic faulted basins around the Ordos block and the shift in tensile stress direction clearly record the remote effects of the convergence of different plates and

deep tectonothermal activity (Shi et al., 2020; Zhang et al., 2006; Zhu et al., 2012). Zhang et al. (2019) stated that the remote effect of the Indo-Eurasian plate collision and deep mantle upwelling induced by Pacific subduction interacted in North China and jointly drove the extensional deformation around the Ordos block. Deng et al. (1999) proposed a dynamic model of the formation and evolution of the Ordos block and its adjacent fault basins. They concluded that several late Mesozoic to early Cenozoic uplift belts around the Ordos block developed under NE-ESE compression of the Tibetan Plateau. On the axis of these uplifts, the upper crust was subjected to secondary tensile stress, easily resulting in ruptures. Once positive strike-slip faults or normal faults began to occur along the rupture, they gradually controlled the development of a series of fault basins. Thus, a discontinuous shear-extensional fault basin belt composed of multiple graben or half-graben basins formed. Furthermore, isostatic adjustment occurred once a faulted basin or faulted basin

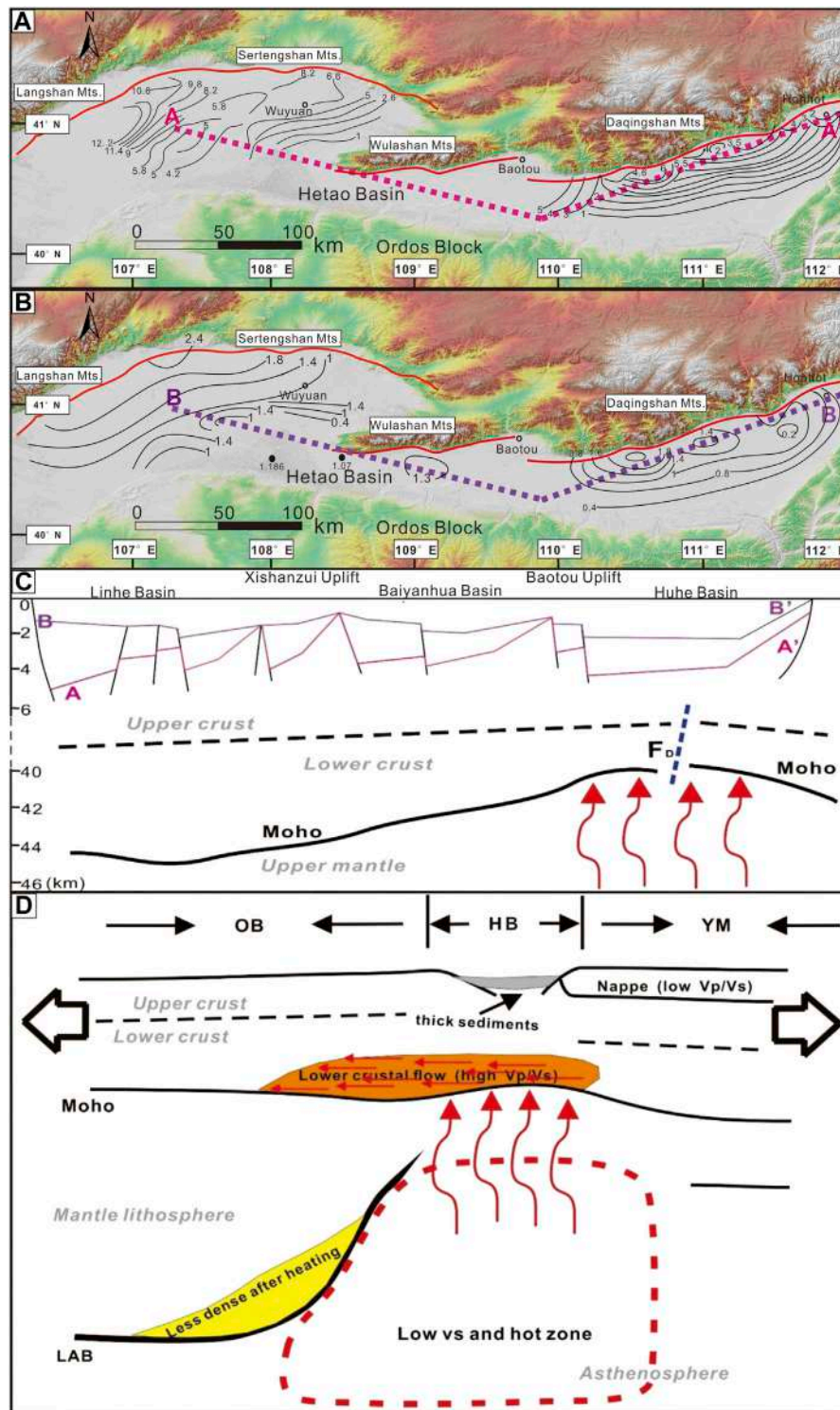


FIGURE 7 | (A) Cenozoic floor isopach (km), thickness from (Li and Nie, 1987); **(B)** Quaternary floor isopach (km), thickness from (Research Group of Active Fault System around Ordos Massif, 1988); **(C)** AA', Cenozoic floor isopach profile in the Hetao Basin; BB', Quaternary isopach profile in the Hetao basin. The FD is the deep fault below Daqingshan (Feng et al., 2015). The red arrow represents the upwelling of deep material below the fracture. The Moho surface depth comes from (Feng et al., 2015) **(D)** Dynamic processes beneath the study area (modified from Teng et al., 2010); OB, Ordos block; HB, Hetao Basin; YM, Yinshan Mountains. The dotted line is the upper and lower crustal boundary of the WRR structure (Teng et al., 2010). LAB represents the lithosphere-asthenosphere boundary, and its depth comes from (Chen et al., 2009; Huang et al., 2009). The region surrounded by the red dotted line represents the lower V region and heat region of the upper mantle; the red arrow

(Continued)

FIGURE 7 | represents the upwelling of the upper mantle pushing the lower crust of the fluvial graben upwards; the yellow area is the area where the density decreases after heating at the bottom of the lithosphere in the northern Ordos block; the orange area shows the ductile flow of the lower crust, which thickens the lower crust of the Ordos block; orange also indicates a high V_p/v_s ratio; the red arrows in the thickened lower crust indicate the direction of ductile flow; and grey indicates thick sedimentary basins beneath the Hetao Basin (Tian et al., 1992).

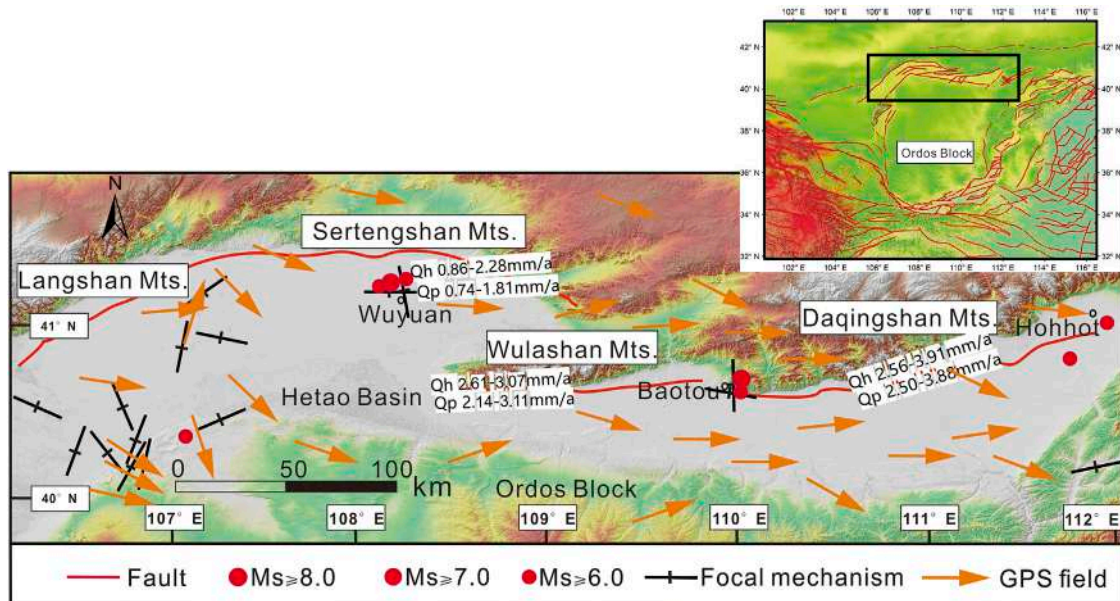


FIGURE 8 | Modern regional tectonic stress field in the Hetao Basin.

belt began to form. Here, upwelling of deep material beneath the basin played an important role (Figure 7D). Therefore, the combined action of the regional horizontal stress field and vertical force generated by deep material movement characterized the dynamic conditions associated with neotectonic activity in this area (Deng and You, 1985).

During the Cenozoic period, the Hetao Basin experienced heavy sedimentation, with the thickness of deposited sediments reaching thousands of metres. Li and Nie (1987) estimated the thickness of the Cenozoic floor based on petroleum geology, hydrology and seismic data and field data (Figure 7A). In the Quaternary period, the boundary faults of the Hetao Basin were highly active, and the faulted basin was constantly subsiding. The depositional centre was always located near the piedmont fault in the north. According to petroleum geology and seismic data, the thickness of the Quaternary strata in the Hetao Basin is large in the west and small in the east (Research Group of Active Fault System around Ordos Massif, 1988) (Figure 7B). Figures 7A,B show that the sedimentary thickness of the Linhe Basin has been much greater than that of the Huhe Basin since the Cenozoic, indicating that the Linhe Basin was the depositional centre of the Hetao Basin in the Cenozoic. However, during the Quaternary, the fault activity on the northern margin of the Hetao Basin has been strong. During this time, the sedimentary thickness of the Linhe Basin has been similar to that of the Huhe Basin, which indicates that the faults on the northern margin controlled the deposition in the Hetao Basin to a certain extent. Based on the increasing fault slip rate from west to east along the northern margin of the Hetao Basin, we infer that the depositional centre of the Hetao

Basin has migrated from west to east during the Quaternary. In addition, the depth of the Moho surface in the Hetao Basin varies greatly, with obvious block characteristics. Overall, the crustal thickness gradually increases from east to west, consistent with its geological structural characteristics (Feng et al., 2015) (Figure 7C). The Yinshan range is mainly characterized by an upper mantle zone, and two mantle zone centres have formed in the Langshan and Daqingshan areas. The crustal thickness of the Langshan area is greater than that of the Daqingshan area, and in the Yinshan range, the crustal thickness is thicker in the west and thinner in the east, as is the overall isobath (Yang et al., 2016). Feng et al. (2015) studied the deep seismic reflection profile of this basin and found that a lithospheric deep fault was located beneath the Daqingshan piedmont fault, which extended from the upper crust, cut the middle-lower crust and Moho surface and entered the upper mantle. The existence of deep faults provides channels for upwelling and intense energy exchange with deep hot material (Figure 7C). The movement of upwelling magmatic material in the asthenosphere caused the melting or softening of the lithosphere in the lower crust and upper mantle of the Hetao fault zone (Yao, 2014). Therefore, we conclude that the depositional centre of the Hetao Basin has tended to migrate from west to east. The vertical force generated by deep material movement may have led to a greater vertical slip rate along the fault zone in the eastern portion of the northern margin of the Hetao Basin (Figure 7C).

Previous studies on the tectonic stress field of the fault zone around the Ordos block suggest that the tectonic stress on the

northern margin of the Ordos block is NW-SE tensile stress (Deng et al., 1999; Fan et al., 2003; Xie et al., 2004; Zhao et al., 2016; Qu et al., 2017). Based on the Crustal Stress Database (2015) of the National Institute of Natural Hazards, Ministry of Emergency Management, we prepared a map of the average modern stress field in the Hetao Basin area. The modern tectonic stress field shows that the Hetao Basin is controlled by NE-SW compression and NW-SW extension. The main compressive stress axis and the tensile stress axis are approximately horizontal, and the direction of the tensile stress axis is 138.66° (Figure 8).

Although the active faults on the northern margin of the Hetao Basin are in the same tectonic stress field, the stress trends of the active faults are different. Therefore, the influence of tensile stress on active faults is different. In addition, the angle between the tensile stress axis and fault strike affects the slip component and dip slip rate of the tensile stress. When the angle between the tensile stress axis and the fault strike is 90° , the slip component and dip slip rate of the tensile stress are the maximum. When the angle between the tensile stress axis and the fault strike increases, the dip slip rate of the tensile stress increases, and the strike-slip rate of the extensional stress decreases. As shown in **Supplementary Table S6**, by comparing the angle between the extensional stress axis and the fault strike in the four faults and the corresponding maximum vertical slip rate, we find that the vertical slip rates of the Sertengshan piedmont fault, Wulashan piedmont fault and Daqingshan piedmont fault gradually increase from west to east under the influence of the modern stress field, which strictly agrees with the theory. However, the angle between the extensional stress axis and the fault strike for the Daqingshan piedmont fault is smaller than that for the Langshan piedmont fault, and the maximum vertical slip rate is larger than that of the Langshan piedmont fault. We believe that this is consistent with our prediction that the vertical force generated by deep material movement may lead to a greater vertical slip rate along the fault zone in the eastern portion of the northern margin of the Hetao Basin. That is, the vertical force generated by deep material movement is the dominant factor that leads to the larger vertical slip rate in the eastern portion of the Hetao Basin. Under the combined action of the modern stress field in the Hetao Basin and vertical force generated by deep material movement, the vertical slip rates of the Sertengshan piedmont fault, Wulashan piedmont fault and Daqingshan piedmont fault increase from west to east, exhibiting spatial variation.

6 CONCLUSION

In this paper, the vertical slip rates were calculated using the vertical displacements of fault-adjacent footwall terraces and the corresponding strata in hanging wall boreholes. The vertical displacement calculated by using the marker strata of both fault walls is not affected by the surface sediment cover, and the calculated vertical slip rate is complete and accurate. However, our vertical slip rate includes the interseismic creep slip rates of the fault. The vertical slip rates of the three active faults on the northern margin of the Hetao Basin since the late Pleistocene and Holocene increase from west to east. The fault activity of the Sertengshan piedmont fault increased from the late Pleistocene to the Holocene, but that of the Daqingshan piedmont fault and Wulashan piedmont fault decreased

during this period. Based on the recurrence characteristics of palaeoearthquakes and the vertical slip rates obtained in this paper, it can be found that the Langshankou and Hongqicun segments of the Sertengshan piedmont fault are at higher risk of earthquakes than other segments. Among the fault segments of the Wulashan piedmont fault, the possibility of future earthquakes in the Baotou segment is high. The seismic risk of the Tuyouxi segment of the Daqingshan piedmont fault should not be ignored, and the Tuzuoxi, Bikeqi and Hohhot segments have high seismic risk. Based on the dynamic model of the formation and evolution of the Ordos block and its adjacent fault basins, it is concluded that the depositional centre of the Hetao basin has tended to migrate from west to east during the Quaternary. The vertical force generated by deep material movement is the dominant factor leading to a greater vertical slip rate along the fault zone in the eastern portion of the northern margin of the Hetao Basin. The modern stress field in the Hetao Basin results in an increasing vertical slip rate of active faults from west to east on the northern margin of the basin.

DATA AVAILABILITY STATEMENT

The raw data supporting the conclusion of this article will be made available by the authors, without undue reservation.

AUTHOR CONTRIBUTIONS

ZH designed the project. ZH, DX, JL, HZ and KL performed the fieldwork. DX and ZH analysed all the data and wrote the first draft. All authors discussed the results, provided feedback and commented on the manuscript.

FUNDING

This work was supported by a research grant from the National Institute of Natural Hazards, Ministry of Emergency Management of China (ZDJ 2019-21), and grants from the National Science Foundation of China (Nos. 41872227 and 41602221).

ACKNOWLEDGMENTS

We thank the reviewers and editors for their help with improving the manuscript. The optically stimulated luminescence (OSL) samples were analysed by Zhao-Junxiang at the Key Laboratory of Crustal Dynamics (National Institute of Natural Hazards, Ministry of Emergency Management of China). ^{14}C sample testing was conducted by the BETA Laboratory in the United States.

SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.816922/full#supplementary-material>

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Modelling Fault Scarp Degradation to Determine Earthquake History on the Muztagh Ata and Tahman Faults in the Chinese Pamir

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OPEN ACCESS

Edited by:

Daoyang Yuan,
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Reviewed by:

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China
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 18 December 2021

Accepted: 18 February 2022

Published: 09 March 2022

Citation:

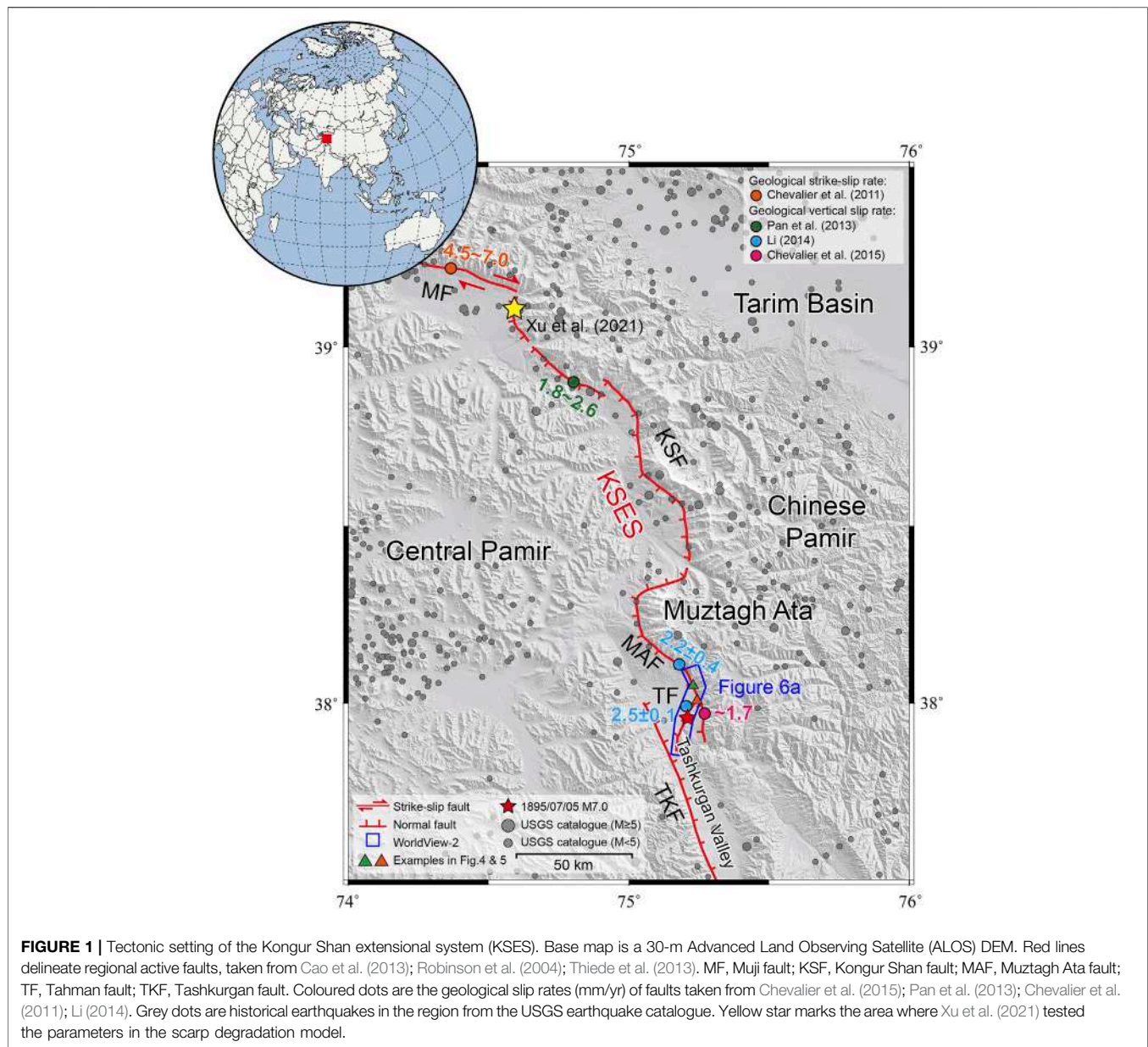
Lu L, Zhou Y, Zhang P and Cheng X
(2022) Modelling Fault Scarp
Degradation to Determine Earthquake
History on the Muztagh Ata and
Tahman Faults in the Chinese Pamir.
Front. Earth Sci. 10:838866.
doi: 10.3389/feart.2022.838866

The frequency of earthquakes on active faults is usually revealed by palaeoseismic trenching and geological dating. Nonetheless, field trenching and sampling are sometimes limited by difficulties in logistics, therefore impeding palaeoseismological studies. As surface rupture earthquakes create fault scarps that evolve with time, they may also provide important information about the timing of earthquakes. In this study, we provide a new approach to determine earthquake history based on the morphology of fault scarps. We select the Muztagh Ata and Tahman normal faults in the Chinese Pamir as a test area, on which fault scarps are well preserved but the earthquake history has not been quantified yet. We first simulate fault scarp degradation using a nonlinear transport model to obtain the empirical relationship between scarp width and its evolution time. We then measure 68 fault scarps from high-resolution topographic data derived from WorldView-2 stereo imagery. The measured fault scarps are clustered in four groups, each group possibly representing one earthquake event. Combining the time-width relationship and scarp width measurements, we infer that four earthquakes occurred at 0.1 ± 0.4 kyrs before the present (BP), 2.1 ± 0.5 kyrs BP, 2.9 ± 0.4 kyrs BP and 4.9 ± 0.5 kyrs BP, respectively. The first event is likely to be the most recent earthquake occurred on the faults, i.e., the 1895 Tashkurgan earthquake (~ 0.13 kyrs BP). Based on the estimated earthquake recurrence intervals and vertical offsets, we obtain a vertical slip rate of 2.2 ± 0.3 mm/yr on the Muztagh Ata and Tahman faults, conforming to previous geological rate estimates.

Keywords: scarp degradation modelling, scarp width, Muztagh Ata and Tahman faults, earthquake history, 1895 earthquake, Chinese pamir

1 INTRODUCTION

The frequency of earthquakes that have occurred along a fault ultimately determines its seismic potential and is therefore important for hazard assessment. A direct way to reveal the number of surface rupture events is palaeoseismic trenching (Deng and Liao, 1996; Marco et al., 2005; Rockwell and Ben-Zion, 2007), but it is often limited by difficulties in logistics in the field. Since earthquakes are recorded on the earth's surface in the form of offset geomorphic features, such offsets can be used as an alternative method to characterise repeating earthquakes and fault slip behaviour. Under the assumption that similar-aged offset geomorphic features experienced the same number of



earthquakes, offset measurements appear as groups, which can be used as an indicator of repeating earthquakes (Sieh, 1978; McGill and Sieh, 1991). This method, now known as “cumulative offset probability distribution, COPD”, requires a large amount of offset measurements for statistical analysis (e.g., Zielke et al., 2015; Bi et al., 2018). During the last decade, high-resolution remote-sensing data promote the application of the “COPD” method in various places around the world (e.g., Zielke et al., 2010; Klinger et al., 2011; Li et al., 2012; Ren et al., 2016; Bi et al., 2018; Choi et al., 2018). For instance, Zielke et al. (2010) measured 149 offsets along the Carrizo segment of the San Andreas fault using light detection and ranging (LiDAR) data, and identified five peaks (5.3, 9.8, 14.9, 20.1, and 24.5 m, respectively) from the offset measurements, with the first peak corresponding to the slip of the 1857 M_w 7.9 Fort Tejon earthquake. Klinger et al. (2011)

measured five slip peaks that are multiples of 6 m along the Fuyun fault in the south of the Altay Mountains, Northeastern Xinjiang, China, based on Quickbird images, and concluded that ruptures on the Fuyun fault exhibits characteristic slip behaviour. Nonetheless, a recent study by Lin et al. (2020) pointed out that the COPD method is only applicable to a few large mature strike-slip faults such as the San Andreas and Fuyun faults.

Surface offsets are important for estimating the magnitude of slip in past earthquakes. Several early studies, e.g., Arrowsmith et al. (1998); Avouac (1993); Colman and Watson (1983); Culling (1960); Jayangondaperumal et al. (2013); Jayangondaperumal et al. (2017), have noted that the morphology of a scarp is useful for constraining the evolution time since its formation. However, previous studies based on the COPD method depend solely on the offset itself, and ignore the shape of the geomorphic features. In this study, we

propose a new method to constrain earthquake history by modelling scarp degradation. The method is applied to the Muztagh Ata and Tahman faults in the Chinese Pamir, providing new information about regional earthquake records.

2 TECTONIC BACKGROUND OF THE KONGUR SHAN EXTENSIONAL SYSTEM

The Pamir, located in the western end of the Indo-Asian collision zone, is a tectonically active region in central Asia (Arnaud et al., 1993; Brunel et al., 1994; Burtman and Molnar, 1993; Cao et al., 2013; Schurr et al., 2014; Sobel et al., 2011) (**Figure 1**). The deformation of the eastern Pamir is dominated by E-W extension along the Kongur Shan (Robinson et al., 2004; Yuan et al., 2013), a 250-km-long NW-SE-striking fault system, which can be divided into four segments from north to south: the right-lateral strike-slip Muji fault, the Kongur Shan and Muztagh Ata normal faults, the Tahman normal fault, and the Tashkurgan normal fault (Robinson et al., 2004; Thiede et al., 2013).

In this study, we focus on the Muztagh Ata and Tahman faults in the southern part of the Kongur Shan extensional system (**Figure 1**). The Muztagh Ata fault, a 70-km-long NW-SE-striking normal fault, forms the western boundary of the Muztagh Ata Mountain. Chevalier et al. (2015) combined geomorphic analysis and ^{10}Be dating of alluvial terrace risers to determine the late Quaternary vertical slip rate of the southern Muztagh Ata fault. They obtained a rate of ~ 1.7 mm/yr, corresponding to a NE-SW extension rate of 1.3–2.0 mm/yr for a fault dip of 35° – 45° (Chevalier et al., 2015) (**Figure 1**). The Tahman fault, extending 20 km north-south, is separated into two segments by a 0.5-km step-over zone (Li et al., 2011). The vertical slip rate was estimated to be 2.5 ± 0.1 mm/yr from ^{10}Be dating of alluvial terraces (Li, 2014) (**Figure 1**).

Both the Muztagh Ata and Tahman faults have been seismically active since the Quaternary (Yuan et al., 2013; Li, 2014; Chevalier et al., 2015). The USGS earthquake catalogue shows that hundreds of $M_w \geq 2$ earthquakes occurred in the region since 1900, including 46 $M_w \geq 5$ earthquakes. The largest event that occurred in the Tashkurgan Valley is the 1895/07/05 M_s 7.0 earthquake, which reached a maximum intensity of IX (Li et al., 2011). Up to now, the fault scarps created by the 1895 M_s 7.0 Tashkurgan earthquake are well-preserved along the southern Muztagh Ata and Tahman faults (Li et al., 2011; Chevalier et al., 2015). Nonetheless, due to difficulties in palaeoseismic trenching and sampling in the area, the regional earthquake history remains largely unknown.

3 MODELLING FAULT SCARP DEGRADATION

The scarps along the Muztagh Ata and Tahman faults are potentially useful for constraining the timing of surface rupture events. We use the well-established scarp degradation model (Culling, 1960, 1963; Kokkalas and Koukouvelas, 2005; Pelletier and Cline, 2007) to analyse the relationship between

scarp morphology and its age. The derivation of the model can be found in many studies, (e.g., Andrews and Bucknam, 1987; Roering et al., 1999; Pelletier and Cline, 2007; Xu et al., 2021). Here, we briefly introduce the set-up of the model.

Sediment flux $S(x)$ down surface topography h is related to local slope $\frac{dh}{dx}$ and a diffusion coefficient $f(h)$:

$$S(x) = f(h) \cdot \frac{dh}{dx} \quad (1)$$

where we take a nonlinear diffusion coefficient $f(h)$ that follows the general form (Andrews and Bucknam, 1987; Pelletier and Cline, 2007):

$$f(h) = \frac{k}{1 - \min\left(c, \frac{dh}{dx}/S_c\right)^n} \quad (2)$$

where k is a constant, S_c is the critical slope, n is an empirical exponent, and c is a threshold to limit the maximum ratio between the hillslope and the critical slope S_c .

The continuity equation is used to relate sediment flux and the evolution of topography with time ($\frac{\partial h}{\partial t}$):

$$-\rho_s \frac{\partial h}{\partial t} = \rho_s \nabla \cdot S(x) + \rho_r C_0 \quad (3)$$

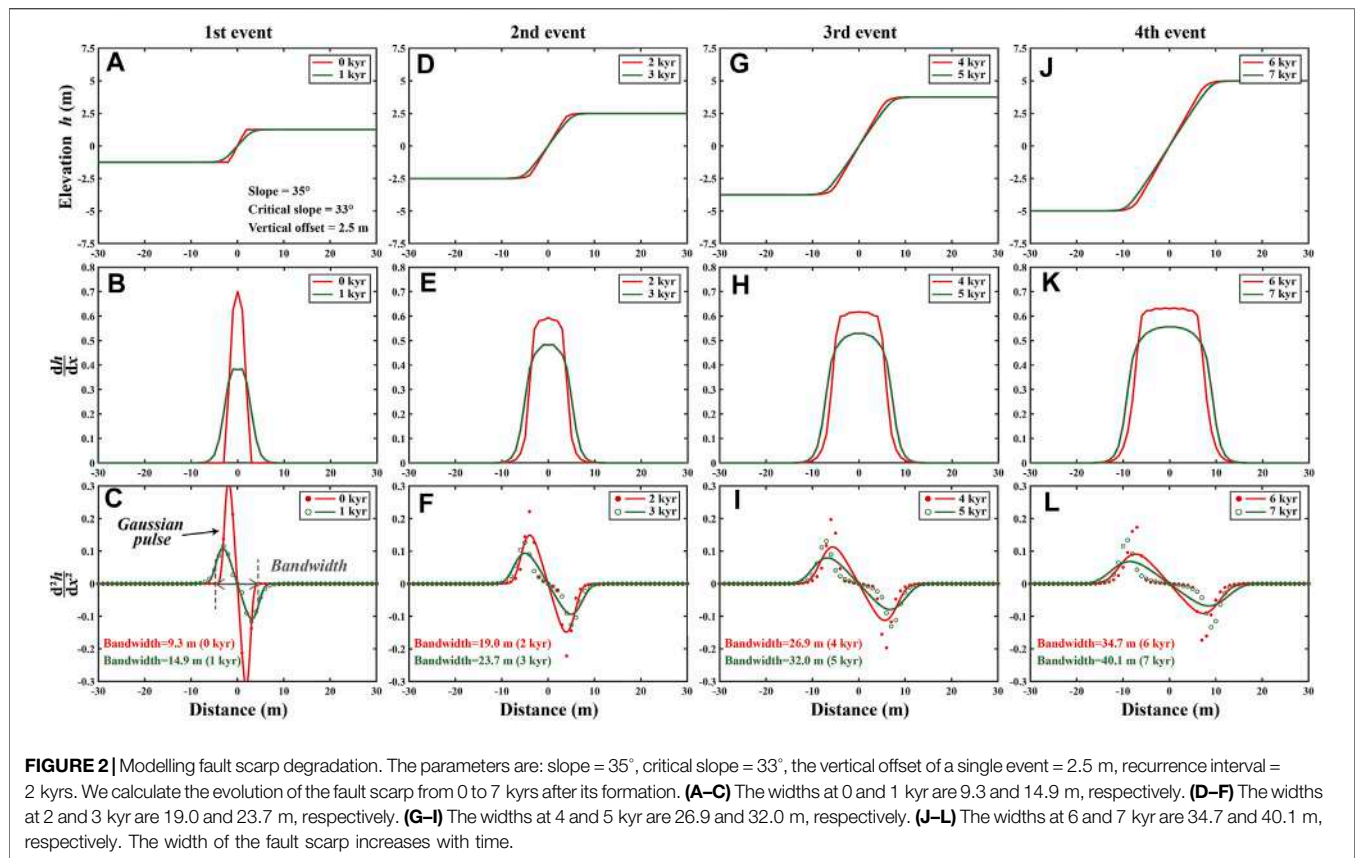
where ρ_s and ρ_r are the bulk densities of sediment and rock, and C_0 is the subsidence/uplift rate.

Combining **Equations 1–3**, and ignoring the subsidence/uplift term $\rho_r C_0$, we derive the nonlinear model for scarp degradation:

$$-\frac{\partial h}{\partial t} = \nabla \cdot \left(\left[\frac{k}{1 - \min\left(c, \frac{dh}{dx}/S_c\right)^n} \right] \frac{dh}{dx} \right) \quad (4)$$

The method is best suitable for normal fault scarps, the shape of which is less affected by gravity. Reliable estimates of the geomorphological parameters in the diffusion model (k , S_c , n , and c) need to be obtained first by local geological survey. Xu et al. (2021) have analysed the fluvial terrace riser degradation along the Kongur Shan fault (~ 130 km north of our study area, yellow star in **Figure 1**) in great detail. They tested a variety of parameters in the diffusion model, and obtained the best-fit by comparing the ages derived from morphological modelling to those from geological dating. We use their best-fitting parameters ($k = 1$, $n = 2$, $c = 0.99$, $S_c = \tan 33^\circ$) in our study because the sediment transport processes should be similar given the close distance. The geometry of a scarp at its formation is determined by the initial slope angle and vertical offset. Also, a scarp degrades faster with a steeper slope. The typical vertical offset is associated with earthquake slip. We use an initial slope angle of 35° [average value from Li. (2014)] and a typical vertical offset of 2.5 m from a single event [from Li et al. (2011)].

Using **Equation 4**, we model the evolution of fault scarp with time. We assume repeating earthquakes every 2 kyrs, i.e., earthquakes occur at 0, 2, 4, 6 kyr with a vertical offset of 2.5 m (**Figures 2A,D,G,J**). This value is chosen based on the geomorphological dating in Chevalier et al. (2015), which suggests that the oldest geomorphic surface along the Muztagh Ata fault (~ 20 kyrs) may have recorded ~ 10 large earthquakes.

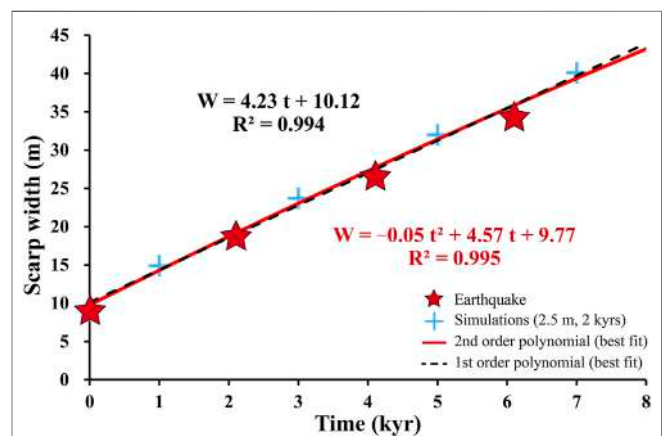


We calculate the slope (first derivative of topography with respect to horizontal distance, $\frac{dh}{dx}$) (Figures 2B,E,H,K), as well as the second derivative $\frac{d^2h}{dx^2}$ in order to remove the regional topographic gradient of the natural earth's surface (Figures 2C,F,I,L). As shown in Figure 2, the most significant feature is that the width of the scarp widens. We find that the second derivative of the scarp can be described by a modified Gaussian pulse function:

$$g(x) = e^{\frac{1}{2}} \cdot \frac{x}{\sigma} \cdot \exp\left(-\frac{\left(\frac{x}{\sigma}\right)^4}{2}\right) \quad (5)$$

where x is the distance, and σ is an exponential factor to be determined.

We use the Gaussian pulse function to define the width of fault scarp: the distance between two points with $g(x)$ equals to 10^{-4} and -10^{-4} ($g(x) = 0$ gives an infinite x , grey line in Figure 2C). The second derivative is advantageous over topography and slope (the first derivative) because it provides a mathematical description of the scarp morphology and therefore avoids artificial identification of scarp end points. The scarp width (also referred to as the bandwidth of the best-fitting Gaussian pulse function) increases from the initial value of 9.3 m after the first event, to 19.0 m after the second event, to 26.9 m after the third event, and to 34.7 m following the fourth event. We plot the bandwidth-time in Figure 3 and find that the changes in



scarp widths can be fitted well with a 1st (black dash line) and 2nd order (red solid line) polynomial. We did not use the 1st order polynomial given that the degradation process is unlikely to be at a constant rate, but should slow down with increasing widths.

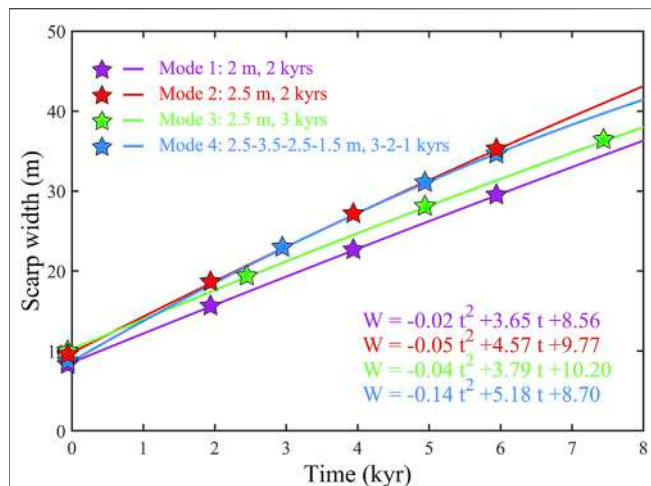


FIGURE 4 | Scarp time-width relationships from different earthquake modes. Mode 1 (purple): offset = 2 m, recurrence interval = 2 kyrs. Mode 2 (red): offset = 2.5 m, recurrence interval = 2 kyrs. Mode 3 (green): offset = 2.5 m, recurrence interval = 3 kyrs. Mode 4 (blue): offsets = 2.5, 3.5, 2.5, and 1.5 m, with corresponding recurrence intervals of 3, 2 and 1 kyrs, respectively. Stars indicate the occurrence of earthquakes.

4 TESTING THE SENSITIVITY OF SCARP WIDTH TO DIFFERENT EARTHQUAKE SLIP MODES

To investigate how the initial parameters in the model including earthquake recurrence intervals and offsets affect the time-width relationship, we simulated four different earthquake modes: 1) offset = 2 m, recurrence interval = 2 kyrs; 2) offset = 2.5 m, recurrence interval = 2 kyrs; 3) offset = 2.5 m, recurrence interval = 3 kyrs; 4) offsets = 2.5, 3.5, 2.5, and 1.5 m, with recurrence intervals of 3, 2, and 1 kyr, respectively. The time-width relationships of the four earthquake modes are shown in **Figure 4**.

Comparing modes 2 (red line) and 3 (green line), we noted that scarp width increases more rapidly with shorter recurrence intervals. For a scarp width of 35.2 m (the maximum scarp width observed in this study), the inferred ages from modes 2 and 3 are 6.0 kyrs before the present (BP) and 7.1 kyrs BP, respectively. Given that most scarps in our study area are narrower than 30 m, the age uncertainties caused by the initial recurrence interval in the model set-up should be within ~1 kyr. The difference between modes 1 (purple line) and 2 (red line) in **Figure 4** shows that larger offsets create wider scarps. Based on the time-width relationships, the age estimates of the widest scarp (35.2 m) are 7.6 kyrs BP (mode 1) and 6.0 kyrs BP (mode 2).

The actual earthquake rupture history is often more complicated, with different slips and recurrence intervals (Zielke et al., 2015). To test this scenario, we vary the offsets from 1.5 to 3.5 m, and recurrence intervals from 1 kyr to 3 kyrs. Interestingly, the time-width relationship of mode 4 is similar to that from mode 2, the long-term fault slip rates of which are equal (1.25 mm/yr). This means that the change in width of a fault scarp is ultimately determined by the fault slip rate. The simulations also suggest that although the time-width relationship varies with

the assumed earthquake offsets and recurrence intervals, the age uncertainty for scarps of narrower than 30 m seems to be less than 1.4 kyr. By analysing a large amount of fault scarps, we may further bring down the uncertainty.

5 DATA AND RESULTS

5.1 Mapping Fault Scarps From High-Resolution Stereo Data

We purchased 0.5 m resolution WorldView-2 stereo imagery covering a 30 km long segment of the southern Muztagh Ata and Tahman faults to investigate the regional earthquake history, including the 1895 M_s 7.0 Tashkurgan rupture (**Figure 1**). The Leica Photogrammetry Suite in ERDAS was used to process the stereo imagery. We generated 63 tie points automatically and removed those mismatches, yielding a root-mean-square error of ~0.1 pixels. A pixel-by-pixel matching procedure was performed to produce a dense point cloud with a search window size of 5×5 pixels (Zhou et al., 2015). The extracted point cloud was then gridded with 1-m spacing to derive the WorldView-2 digital elevation model (DEM) (coordinate system: WGS84 UTM 43 N). The WorldView-2 images were then orthorectified using the 1-m DEM.

Based on the WorldView-2 orthoimage and DEM, we mapped a 26-km-long fault trace (**Figure 6A**). We extracted elevation stacking profiles (20-m wide, 100-m long) across the mapped fault scarps (**Figures 5A,B,E,F** for example). To make sure that the offsets are estimated correctly, we removed those profiles that cross different geomorphic surfaces (Bi et al., 2018). We measured the offsets by fitting two straight lines to each side of the fault, and calculated the minimum and maximum distances between the two lines (see **Figures 5C,G** for example). A total of 68 vertical offsets ranging from 1.2 to 14.8 m were obtained (**Figure 6E** and **Table 1**).

5.2 Calculating Scarp Width

Based on the elevation stacking profiles extracted from the WorldView-2 DEM, we calculated $\frac{d^2h}{dx^2}$ for each of the 68 scarps, and then estimated the scarp width by fitting a Gaussian pulse function. Since some scarps have recorded multiple earthquakes, $\frac{d^2h}{dx^2}$ may show more than one peak (as shown in **Figure 5H**), in which case we use a double pulses function:

$$g'(x) = a_1 \cdot \frac{x+b_1}{\sigma_1} \cdot \exp\left(-\left(\frac{x+b_1}{\sigma_1}\right)^2 / 2\right) + a_2 \cdot \frac{x+b_2}{\sigma_2} \cdot \exp\left(-\left(\frac{x+b_2}{\sigma_2}\right)^2 / 2\right) \quad (6)$$

where a_1 , b_1 , σ_1 , a_2 , b_2 and σ_2 are the coefficients that can be determined by fitting the equation to the profile.

After solving for the best-fitting pulse function using a least-squares adjustment, we compute the widths of all the scarps along the faults (see **Figures 5D,H** for example). The scarp widths range from 7.1 to 35.2 m (**Figure 6C**). We use 1- σ as the uncertainty of the width estimates.

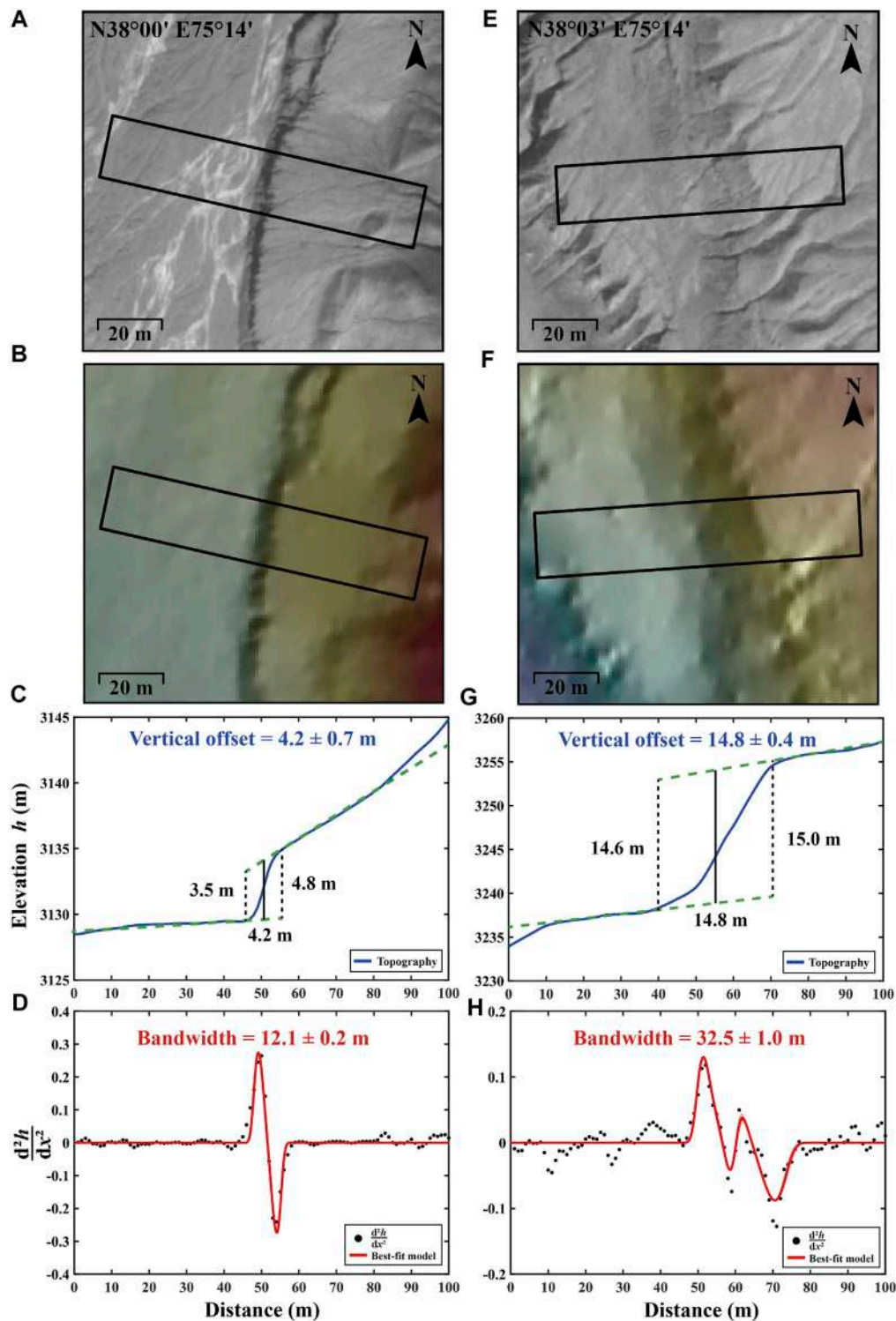


FIGURE 5 | Examples of scarp width calculation. **(A)** WorldView-2 panchromatic orthoimage. **(B)** Shaded WorldView-2 DEM of the scarp. The polygon shows the swath profile. **(C)** Estimating the vertical offset of the scarp. Blue solid line shows the topography across the scarp. Green lines are the best-fitting straight lines to each side of the scarp that are used to calculate the vertical offsets. **(D)** Estimating the scarp width by a single Gaussian pulse. Black dots trace the $\frac{d^2h}{dx^2}$ profile of the observed scarp. Red solid line is the best-fitting Gaussian pulse function. **(E–H)** Same as **(A)–(D)**, but for an older scarp. The positions of the scarps are shown in **Figure 1** with green triangles.

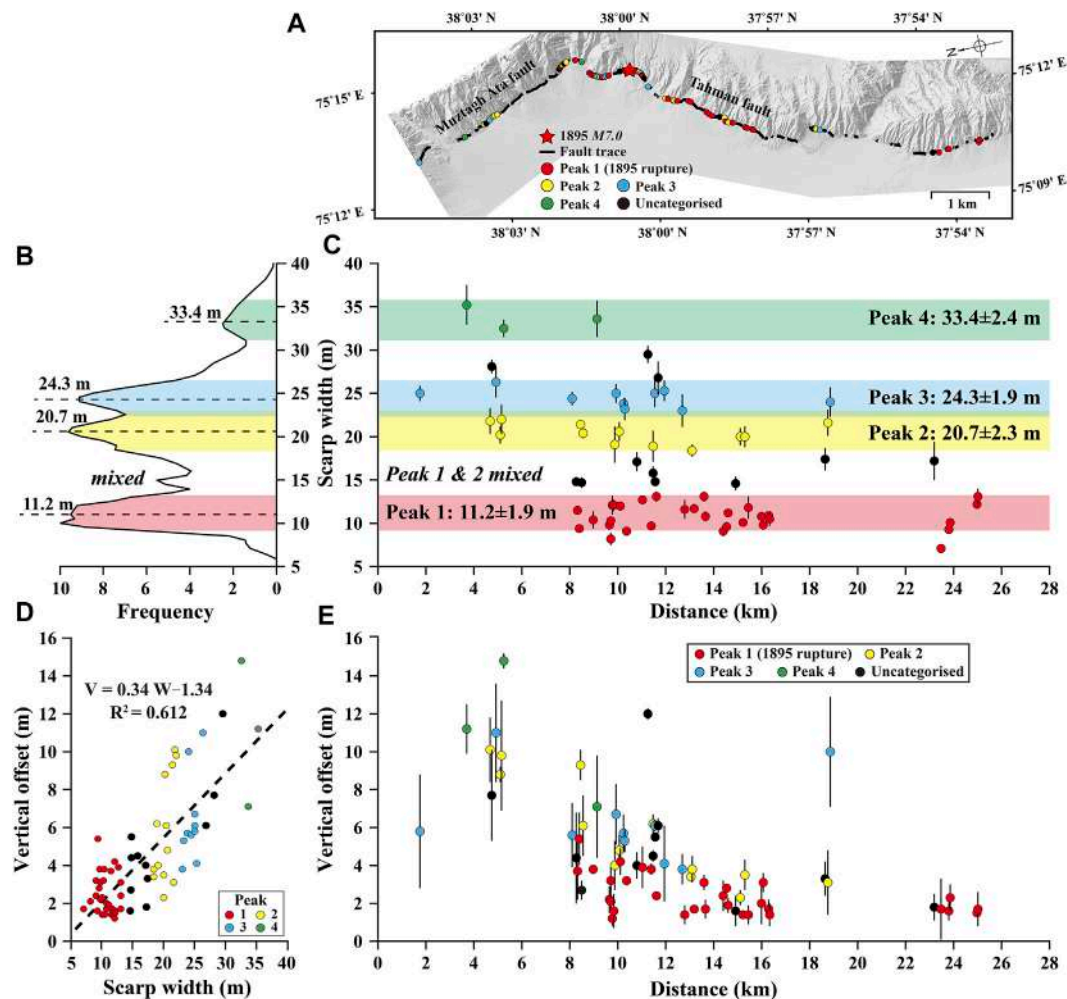


FIGURE 6 | Scarp width probability distribution (SWPD) and cumulative offset probability distribution (COPD). **(A)** WorldView-2 DEM shows the fault trace (black solid lines). Coloured dots are the locations of the scarp measurements. Red star marks the epicentre of the 1895 earthquake, inferred by Li et al. (2011). **(B)** SWPD of the southern Muztagh Ata and Tahman faults, showing four peaks at 11.2, 20.7, 24.3 and 33.4 m. **(C)** The distribution of scarp width along the faults. Measurements (with 1-σ) are clustered in four groups (11.2±1.9, 20.7±2.3, 24.3±1.9 and 33.4±2.4 m). **(D)** The empirical width-offset relationship. **(E)** Along-fault distribution of vertical offsets.

5.3 Probability Density Function of Scarp Widths

We calculated the scarp width probability distribution (SWPD) by summing up the Gaussian probability density of the 68 width measurements. The peak value and uncertainty (1-σ) are determined by applying a normal distribution function. As shown in **Figures 6B,C**, the SWPD shows four peaks centred at 11.2±1.9, 20.7±2.3, 24.3±1.9 and 33.4±2.4 m, respectively, indicating that the widths are clustered in four groups. If we assume that the scarps have experienced similar sediment transport processes, the clustering implies that each group of scarps may have been formed at the same time, possibly during the same earthquake. We calculated the ages of the four peaks of scarp widths from our initial empirical time-width relationship (**Figure 3**) and obtained: 0.3±0.4 kyrs BP (peak 1), 2.5±0.5 kyrs BP (peak 2),

3.3±0.5 kyrs BP (peak 3) and 5.5±0.6 kyrs BP (peak 4), respectively (red markers in **Figure 7A**).

6 DISCUSSION

6.1 Earthquake History on the Muztagh Ata and Tahman Faults

Given that the parameters in the initial time-width relationship (recurrence interval = 2 kyrs and characteristic slip = 2.5 m) are to some extent random, we refine the relationship using the estimated earthquake timing derived from the initial model. Also, we find that the vertical offsets increase linearly with scarp widths (**Figure 6D**), allowing us to estimate the average vertical slip of each earthquake for a given scarp width. With the new constraints on slip and timing, we refined the scarp degradation model (**Figure 7A**) and obtained four new ages: 0.1±0.4 kyrs BP, 2.1±0.5 kyrs BP, 2.9±0.4 kyrs BP and

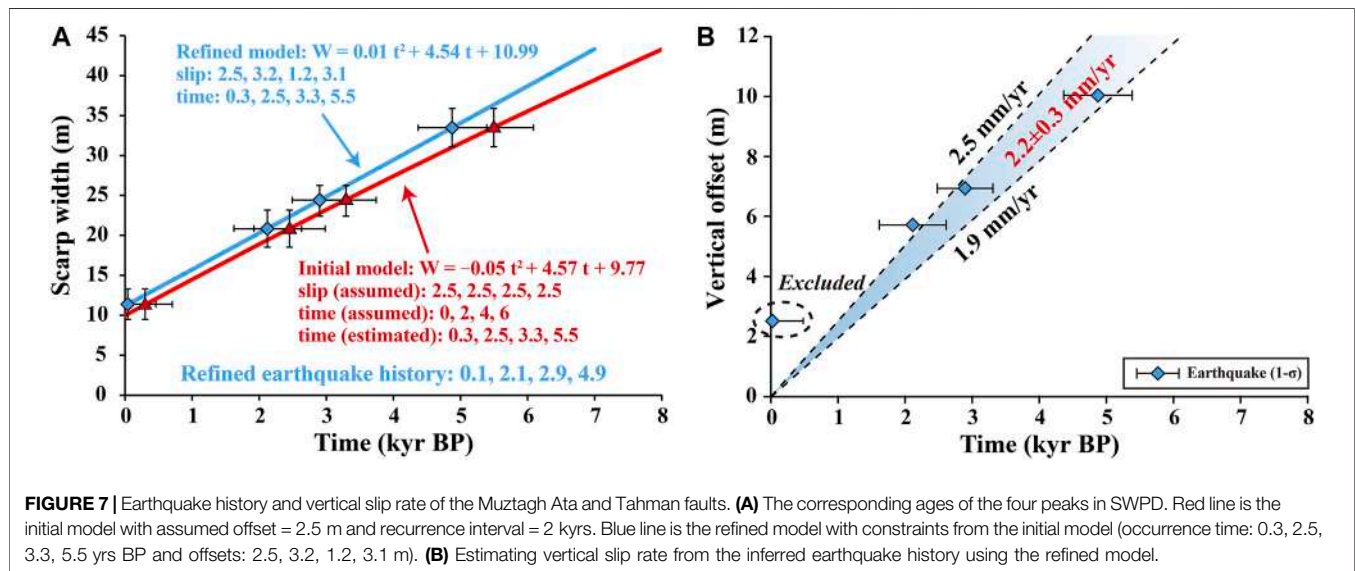
TABLE 1 | Measurements of 68 fault scarps. Profiles 8 and 25 are the selected examples shown in **Figure 5**.

Profile id	Latitude (°)	Longitude (°)	Distance (km)	Width (m)	1- σ (m)	Vertical offset (m)	1- σ (m)
1	38.077 20	75.216 02	1.750	25.0	0.9	5.8	3.0
2	38.060 38	75.222 49	3.700	35.2	2.3	11.2	1.3
3	38.052 40	75.226 11	4.678	21.8	1.5	10.1	1.7
4	38.051 95	75.226 39	4.747	28.1	0.8	7.7	2.3
5	38.050 57	75.227 54	4.920	26.3	1.8	11.0	2.6
6	38.049 19	75.228 73	5.100	20.2	1.0	8.8	0.3
7	38.048 75	75.228 92	5.152	22.0	1.7	9.8	2.9
8	38.047 82	75.229 12	5.245	32.5	1.0	14.8	0.4
9	38.023 51	75.242 31	8.100	24.4	0.8	5.6	1.7
10	38.022 84	75.243 26	8.265	14.8	0.4	4.4	2.4
11	38.022 54	75.243 82	8.321	11.5	0.4	3.7	1.5
12	38.022 13	75.244 28	8.398	9.4	0.4	5.4	1.4
13	38.021 66	75.244 39	8.446	21.4	0.5	9.3	0.8
14	38.021 21	75.244 69	8.492	14.7	0.6	2.7	0.5
15	38.020 76	75.245 15	8.555	20.4	0.6	6.1	1.6
16	38.017 76	75.245 32	8.975	10.4	1.0	3.8	0.2
17	38.016 38	75.244 88	9.134	33.6	2.1	7.1	2.7
18	38.013 84	75.238 58	9.650	9.8	0.5	2.2	0.7
19-a	38.013 54	75.237 97	9.710	10.3	0.5	3.2	0.7
19-b	38.013 54	75.237 97	9.710	8.2	0.7	2.1	0.5
20	38.012 71	75.237 18	9.770	12.1	1.1	1.2	0.4
21	38.012 23	75.236 75	9.836	12.1	0.7	1.6	0.9
22	38.011 78	75.236 47	9.873	19.1	2.1	4.0	1.3
23	38.011 33	75.236 31	9.925	25.0	1.1	6.7	1.6
24	38.010 40	75.235 82	10.063	20.6	1.1	4.8	0.8
25	38.009 94	75.235 56	10.106	12.1	0.2	4.2	0.7
26	38.008 58	75.235 57	10.241	23.8	0.6	5.7	1.0
27	38.008 09	75.235 62	10.291	23.2	1.3	5.3	0.6
28	38.007 66	75.235 66	10.365	9.1	0.2	3.2	0.1
29	38.003 23	75.236 51	10.793	17.1	1.1	4.0	0.7
30	38.001 47	75.235 86	11.026	12.7	0.5	3.9	1.1
31	37.999 18	75.235 48	11.253	29.5	1.0	12.0	0.3
32	37.997 95	75.235 10	11.393	9.7	0.2	3.8	0.3
33-a	37.997 05	75.234 56	11.472	18.9	1.8	6.2	0.5
33-b	37.997 05	75.234 56	11.472	15.8	0.8	4.5	0.3
34	37.996 68	75.234 00	11.542	25.0	1.6	6.1	0.3
35	37.996 41	75.233 42	11.560	14.8	0.4	5.5	0.2
36	37.996 29	75.232 84	11.610	13.1	0.6	2.4	0.2
37	37.995 93	75.231 70	11.683	26.8	1.9	6.1	0.4
38	37.995 09	75.227 40	11.940	25.3	1.2	4.1	2.0
39	37.995 09	75.227 40	12.692	23.0	1.9	3.8	0.8
40	37.989 11	75.220 35	12.792	11.6	1.1	1.4	0.5
41	37.986 99	75.218 97	13.048	18.4	0.4	3.4	0.2
42	37.986 55	75.218 78	13.104	18.4	0.7	3.8	0.7
43	37.986 08	75.218 59	13.184	11.7	0.4	1.7	0.1
44	37.982 35	75.217 55	13.594	13.1	0.6	3.1	0.4
45	37.981 88	75.217 34	13.656	10.8	0.6	1.7	0.5
46	37.977 15	75.211 37	14.395	9.1	0.6	2.4	0.8
47	37.976 46	75.210 37	14.536	9.6	0.6	2.8	0.2
48	37.976 09	75.209 79	14.599	11.2	0.4	1.9	0.4
49	37.972 87	75.207 84	14.912	14.6	0.8	1.6	0.8
50	37.971 55	75.207 54	15.108	20.0	1.0	2.3	0.4
51	37.971 42	75.205 78	15.222	10.1	0.5	1.4	0.2
52	37.970 26	75.204 73	15.300	20.0	1.2	3.5	0.8
53	37.969 33	75.204 53	15.443	11.8	1.3	1.4	0.5
54	37.964 95	75.201 25	15.992	10.8	0.5	2.0	1.1
55	37.964 48	75.200 68	16.065	9.8	0.5	3.1	0.5
56	37.962 88	75.199 74	16.288	10.9	0.5	1.7	0.5
57	37.962 49	75.199 28	16.331	10.5	0.8	1.4	0.6
58	37.942 10	75.194 37	18.645	17.4	1.3	3.3	0.9
59	37.941 18	75.193 94	18.757	21.6	1.5	3.1	1.7
60	37.941 18	75.193 94	18.858	24.0	1.7	10.0	2.9
61	37.941 18	75.193 94	23.195	17.2	2.2	1.8	0.7
62	37.901 17	75.172 75	23.478	7.1	0.5	1.7	1.6

(Continued on following page)

TABLE 1 | (Continued) Measurements of 68 fault scarps. Profiles 8 and 25 are the selected examples shown in Figure 5.

Profile id	Latitude (°)	Longitude (°)	Distance (km)	Width (m)	1- σ (m)	Vertical offset (m)	1- σ (m)
63	37.901 17	75.172 75	23.802	9.3	0.3	1.6	0.5
64	37.901 17	75.172 75	23.859	10.1	0.2	2.3	0.7
65	37.901 17	75.172 75	24.976	12.2	0.5	1.5	0.2
66	37.901 17	75.172 75	25.015	13.1	0.9	1.7	0.9



4.9±0.5 kyrs BP, respectively. The refined model shows a good consistency with the initial model, with a difference of < 1 kyr.

The SWPD reveals four earthquakes along the southern Muztagh Ata and Tahman faults in the past 6,000 years. The inferred most recent event (~0.1 kyrs BP) is likely to be the 1895 Tashkurgan rupture (~0.13 kyrs BP). We use the ages and offsets from the SWPD to estimate the fault slip rate. We excluded the most recent event as it will lead to an overestimate. We obtained a vertical slip rate of 2.2±0.3 mm/yr for the Muztagh Ata and Tahman faults (**Figure 7B**), in agreement with previous geological estimates of 1.7–2.5 mm/yr from Chevalier et al. (2015) and Li. (2014).

6.2 Distribution of Vertical Displacements Along the Fault and Fault Segmentation

The vertical offsets that we obtained along the fault trace are a mixture of slip from the 1895 Tashkurgan earthquake and older events. Because the scarps formed during the most recent event should have a relatively narrow width given the shorter evolution time, we use this as a criterion to distinguish the 1895 rupture from older ruptures. We considered the scarps with a width of less than 13.1 m as the 1895 rupture and plotted the measurements in red in **Figure 6E**. We found that the slip of the 1895 earthquake is concentrated along the Tahman fault (between 8–26 km). The average vertical displacement of the 1895 rupture is 2.4±1.2 m (1- σ) (consistent with our initial assumption of 2.5-m characteristic slip) with a maximum displacement of 5.5±0.2 m (1- σ). According to the empirical relationships between the moment

magnitude and each of the maximum and average displacements in Wells and Coppersmith. (1994), we obtained M_w 7.2 and M_w 7.3 for the 1895 Tashkurgan earthquake, respectively.

Slip distributions depicted by **Figures 6A,C,E** manifest a remarkable feature of fault rupture segmentation. The surface ruptures associated with the 1895 Tashkurgan earthquake distribute only along the North-northeast-trending Tahman fault. No evident fresh scarps has been found along the North-northwest-trending Muztagh Ata fault. Although old scarps may also record the most recent rupture, it would be extraordinary that earthquake propagates along the entire fault segment without any fresh scarps. Therefore, it is likely that the intersection region acts as a fault segmentation boundary that impedes earthquake rupture from one fault segment to another (Schwartz and Coppersmith, 1984). The Tahman fault seems to have experienced less earthquake events than the Muztagh Ata fault as most of the high and wide fault scarps are present along the Muztagh Ata fault (**Figures 6C,E**), suggesting different behaviours of different fault segments.

7 CONCLUSION

In this study, we proposed a new approach of using fault scarp degradation model to constrain earthquake history, with applications to the Muztagh Ata and Tahman faults in the Chinese Pamir. We mapped fault scarps based on WorldView-2 stereo imagery and determined the widths of the scarps using a

modified Gaussian pulse function. The measurements show that the widths of the scarps are clustered in four groups, each group representing a surface rupture event. Given the empirical time-width relationship, we calculated the timing of the earthquakes, with the youngest cluster, 0.1 ± 0.4 kyrs BP, corresponding to the 1895 Tashkurgan earthquake. Based on the inferred earthquake recurrence intervals and offsets, we estimated the vertical slip rate (2.2 ± 0.3 mm/yr) of the Muztagh and Tahman faults, in agreement with the geological estimates of 1.7–2.5 mm/yr from Chevalier et al. (2015); Li. (2014). The consistent results suggest that the proposed method provides an innovative means for investigating earthquake history and can be potentially applied to other active faults where surface scarps are well-preserved.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author.

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AUTHOR CONTRIBUTIONS

LL and YZ conceptualised and designed the research. LL performed the simulations, processed and analysed the data. YZ was responsible for the overall supervision of the data processing and analysis. The manuscript was drafted by LL and then reviewed and edited by YZ, PZ, and XC.

FUNDING

This work was supported by the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901), National Natural Science Foundation of China (41874020), Guangdong Province Introduced Innovative R&D Team of Geological Processes and Natural Disasters around the South China Sea (2016ZT06N331) and Deep Earth Exploration and Resource Environment (2017ZT07Z066).

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Microscopic Characteristics of Fault Gouge in Minor-Surface-Rupture Faults: A Case Study in the Longmenshan Fault Zone, Eastern Tibetan

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OPEN ACCESS

Edited by:

Daoyang Yuan,
Lanzhou University, China

Reviewed by:

Yueren Xu,
China Earthquake Administration,
China
Jiawei Pan,
Chinese Academy of Geological
Sciences (CAGS), China

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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 21 December 2021

Accepted: 24 February 2022

Published: 06 April 2022

Citation:

Li Y, Chen L, Ran Y and Chang Y
(2022) Microscopic Characteristics of
Fault Gouge in Minor-Surface-Rupture
Faults: A Case Study in the
Longmenshan Fault Zone,
Eastern Tibetan.
Front. Earth Sci. 10:840667.
doi: 10.3389/feart.2022.840667

Active faults with potential earthquake magnitude of 6–7 are often incorrectly identified as non-Holocene active faults by traditional geological methods because their co-seismic displacements were very small or even did not reach to surface, and are defined as minor-surface-rupture faults. Geological studies associated with the 2013 Lushan Ms 7.0 earthquake show that the Dachuan–Shuangshi fault (DSF) in the southern segment of the Longmenshan fault zone (LFZ) is a minor-surface-rupture fault. This study focuses on the microstructure and mineral composition of the fault gouge in the DSF using optical microscopy, a scanning electron microscope (SEM), and X-ray diffraction (XRD) methods; then, we compare our results with the previous achievements in the Beichuan–Yingxiu fault (BYF), a major seismogenic fault of the 2008 Wenchuan earthquake, in the middle–northern segment of the LFZ. The results show that the microscopic characteristics of the fault gouge of the DSF are obviously different from those of the BYF in the following aspects: 1) the thickness of the fault gouge produced by one fault event is less than 5 mm; 2) under the microscope, no obvious micro-cracks were examined in surrounding rocks around the fault gouge, and discontinuous micro-cracks and untypical S-C fabrics in the fault gouge were observed; 3) under the SEM, reworked fragments were rare in the fault gouge; 4) the XRD mineral analysis reveals that the total clay content is less than 50%, the content of kaolinite is obviously higher than that of clinocllore, and the content of illite/smectite mixed layer is less than 30%. A contrastive analysis reveals the differences between the microscopic features of the fault gouge of the DSF and the BYF, which are systematical. Therefore, the abovementioned microscopic characteristics identified from the fault gouge in the DSF may be used as auxiliary indicators to identify minor-surface-rupture faults.

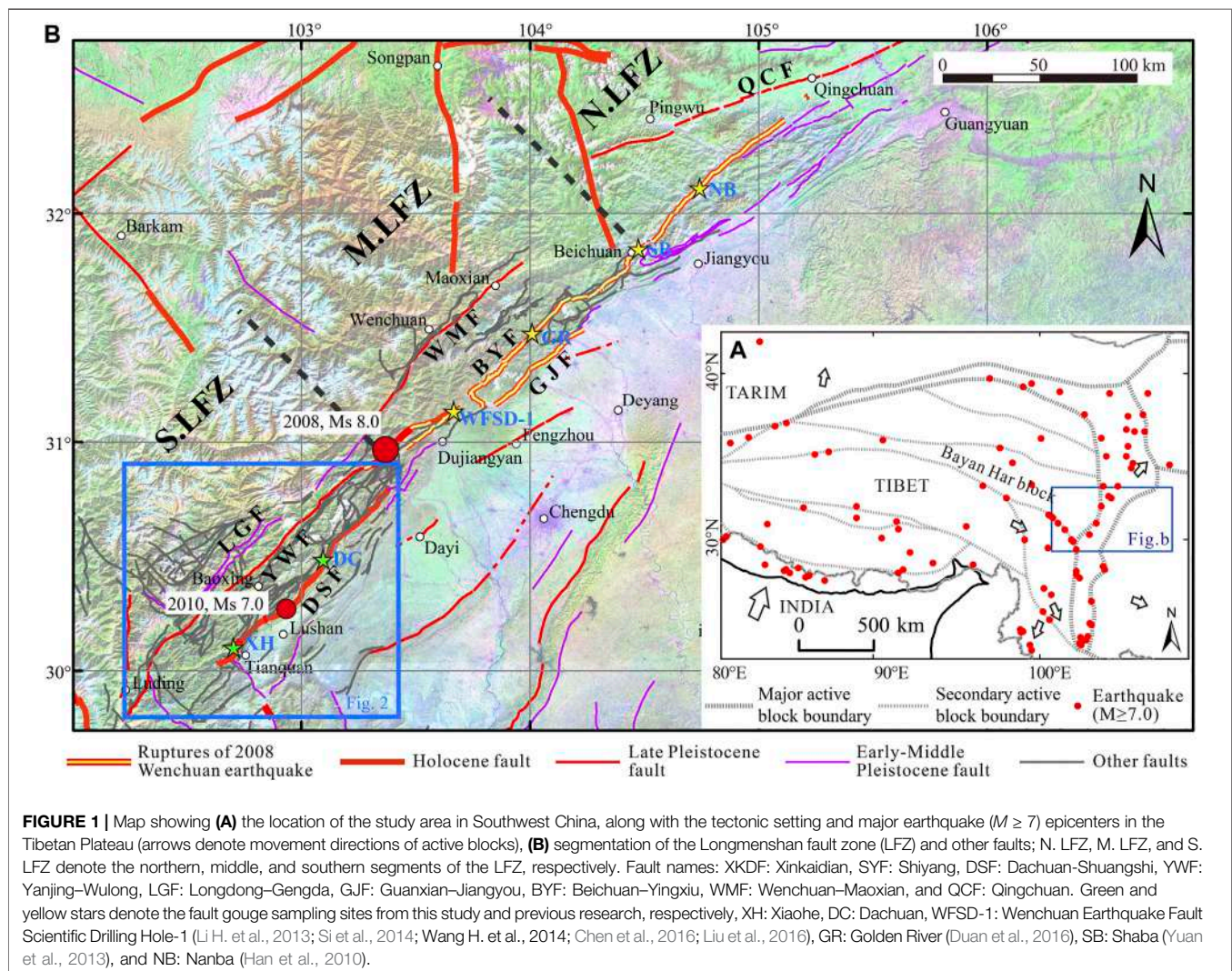
Keywords: fault gouge, microstructure, mineralogical composition, minor-surface-rupture fault, southern Longmenshan fault zone

INTRODUCTION

The offsetting and overlying relationship between faults and Late Quaternary strata is the most immediate evidence to identify the age of fault activity. According to the present achievement, the fault that does not offset the Late Pleistocene or Holocene strata is unlikely to trigger earthquakes with surface ruptures in the future. Although some faults could offset the surface, their co-seismic displacements are very small, only a few centimeters to a few tens of centimeters, and the remains can easily be eroded, and hardly be preserved and identified in the geological record. For example, the 2013 Lushan Ms 7.0 earthquake that occurred along the southern segment of the Longmenshan fault zone (LFZ) did not produce obvious surface rupture (Li C. et al., 2013; Xu et al., 2013a; Xu et al., 2013b; Lei et al., 2014). Such type of active faults is classified as minor-surface-rupture active fault (Chen et al., 2013a), and by traditional geological methods, they are easily identified as non-Holocene active faults, which leads to the neglect of the future strong earthquake risk along these faults.

The fault gouge is the direct lithological product of the fault activity and is formed *via* relatively complex physical and chemical processes related to the low-grade metamorphism that accompanies fault slip (Bos et al., 2000). Thus, the microstructure of the gouge contains information on the fault slip and behavior as well as the nature of the physical and chemical environment during faulting (Reinen, 2000; Zhang et al., 2002; Yuan et al., 2013). Although, so far, no explicit relationship between the microstructures of the fault gouge and the intensity of seismic activity has been reported, we can still find some implications and study cases from the observation of the fault gouge microstructures of the known minor-surface-rupture active faults and the comparison with those of intense active faults.

Geological observations show that the seismic activity of the southern segment of the LFZ is obviously weaker than that of the middle–northern segments, and the Dachuan–Shuangshi Fault (DSF), the unique fault in the southern segment with evidence of Holocene offset, is considered as a minor-surface-rupture fault (Yang et al., 1999; Densmore et al., 2007; Chen et al., 2013b; Wang



et al., 2013; Chen et al., 2014; Dong et al., 2017). We have found fault gouges from the outcrop and drill holes along the DSF. Furthermore, the 2008 Wenchuan Ms 8.0 earthquake that occurred along the middle–northern segments of the LFZ produced significant surface-rupture zones with a total length of more than 300 km (**Figure 1B**) (Xu et al., 2009). An abundance of gouges was observed at the surface ruptures, and the microstructures of these gouges have recently been investigated (Fu et al., 2008; Han et al., 2010; Yuan et al., 2013; Yuan et al., 2014). In particular, the “Wenchuan Earthquake Fault Scientific Drilling” (WFSD) project collected a large volume of fault gouge samples from borehole cores, again providing valuable microstructural information (Li H. et al., 2013; Si et al., 2014; Wang et al., 2014; Chen et al., 2016; Liu et al., 2016).

This study presents an analysis of the microstructure and mineral composition of the fault gouge samples collected along the DSF in the southern segment of the LFZ. We compared our results with results from the middle–northern segments of the LFZ to provide evidence and study case regarding differences between less-intense and intense active fault, and then benefit for identifying the minor-surface-rupture by the fault gouge microanalysis.

LATE QUATERNARY SURFACE ACTIVITY OF THE SOUTHERN SEGMENT OF THE LFZ

The LFZ is over 500 km long and forms the eastern boundary of the Bayan Har Block in the Tibetan Plateau. In general, the fault zone is divided into the southern, middle, and northern segments, bounded by the Yingxiu town of Wenchuan County and Beichuan County (**Figure 1B**). The fault zone is 30–50 km wide, with an overall trend of NE 30°, and dips toward the NW at 50–80°. The fault zone consists of several imbricate thrust faults with strike-slip components that are roughly parallel and known as the back-range, central, fore-range, and range-front (buried) faults (Zhang et al., 2008). The geometry of the LFZ is relatively simple in the middle–northern segments, but branches into a series of roughly parallel fault sets toward the south. Here, the fault zone intersects other NW-trending faults near the Baoxing County and further diverges to the NW and SE, with the width increased to about 80 km (**Figure 1B**) (Chen et al., 2013b; Chen et al., 2014).

The geomorphological analysis suggests that differential uplift in the southern segment of the LFZ is noticeably weaker than that in the middle–northern segments (Gao et al., 2016). The southern segment consists of moderately elevated mountains, with few Quaternary basins. Structurally, this area consists of a series of subparallel NE-striking faults and arc-like thrusts, which can be recognized from local linear features in satellite imagery. In the southern segment of the back-range fault, also referred to as the Gengda–Longdong Fault, there is no evidence of the Late Quaternary activity. Geological surveys have shown that the southern segment of the central fault (the Yanjing–Wulong Fault) may have been active in the early part of the Late Pleistocene (Chen et al., 2013b). The range-front fault is mostly buried in the southern segment, and no geological

profile showing offsets of Late Pleistocene–Holocene lithologies was found. However, traces of active folds do appear, and they are inferred to represent a buried fault system active since the Late Pleistocene that does not reach the surface (Xu et al., 2013b; Dong et al., 2017).

The southern segment of the aforementioned fore-range fault is dominated by thrusting and is known as the DSF. The fault extends from the northeast of Dayi County in the north to the southwest of Tianquan County in the south, with a length of over 115 km, trending at NE 40° and dipping toward the NW (**Figure 2**). Several trenches along the DSF show that it has offset Holocene strata, with a co-seismic vertical displacement of ~0.3 m, and with unclear fault planes. Therefore, the DSF is classified as a minor-surface-rupture fault (Densmore et al., 2007; Chen et al., 2013b; Chen et al., 2014; Dong et al., 2017).

MICROSTRUCTURE AND MINERALOGICAL COMPOSITION OF FAULT GOUGE IN THE DSF

Sampling and Methods

The DSF, the unique fault with evidence of the Holocene offset in the southern segment of the LFZ, extends through the township of Dachuan, Lushan County. Based on previous seismic surveys, six boreholes (DC1–DC6) were drilled at this site (site DC in **Figures 1B, 2**). As shown in **Figure 3**, boreholes DC4, DC5, and DC6, located on the hanging wall of the fault, penetrate the fault plane and reveal two fractured sections. Borehole DC4 (30.4876°N, 103.1023°E) reveals a fault fracture zone at depths of 41.8–75.5 m, consisting predominantly of cataclastic sandstone or conglomerate and cataclasite rock, with a soft fault gouge of 1 cm thickness at a depth of 60.3 m that is gray-white in color and dips at 60°. Another fracture zone with evidence of compressional and folded deformation is observed at depths of 90.7–97.7 m. In borehole DC5 (30.4891°N, 103.1029°E), the two fractured sections are at depths of 98.2–99.5 m and 103.6–108.0 m, respectively. Oriented rock fragments and cataclastic, lens-like, and schistose fabrics were observed in the two fractured zones, which dip up to 70°. Borehole DC6 (30.4901°N, 103.1038°E) revealed one fractured section at a depth of 91.9–96.6 m and another at 103.0–104.5 m. The uppermost section also contains noticeable cataclasis, lenses, and schistositicities, with clear compression surfaces and a thin fault gouge that dips at 65–70°. The lower section is characterized by coal beds with folds. Fault gouge samples were collected from these two sections, as shown in **Figure 3** and **Table 1**.

A very clear fracture zone is also evident in the DSF on a slope west of Xiaohe town, Tianquan County. At an abandoned limestone quarry pit (site XH in **Figure 1B, 2**; 30.1079°N, 102.7243°E), the fracture zone is composed of a fault gouge, fault breccia, cataclasites, and folds. The fault gouge is loose with slightly variable thickness (20–50 cm) at different locations, and the fault is exposed at the surface, which implies an activity in the Late Quaternary; the exposed fault trace dips toward 300° at 65–75°. The hanging wall consists of Lower Permian limestone,

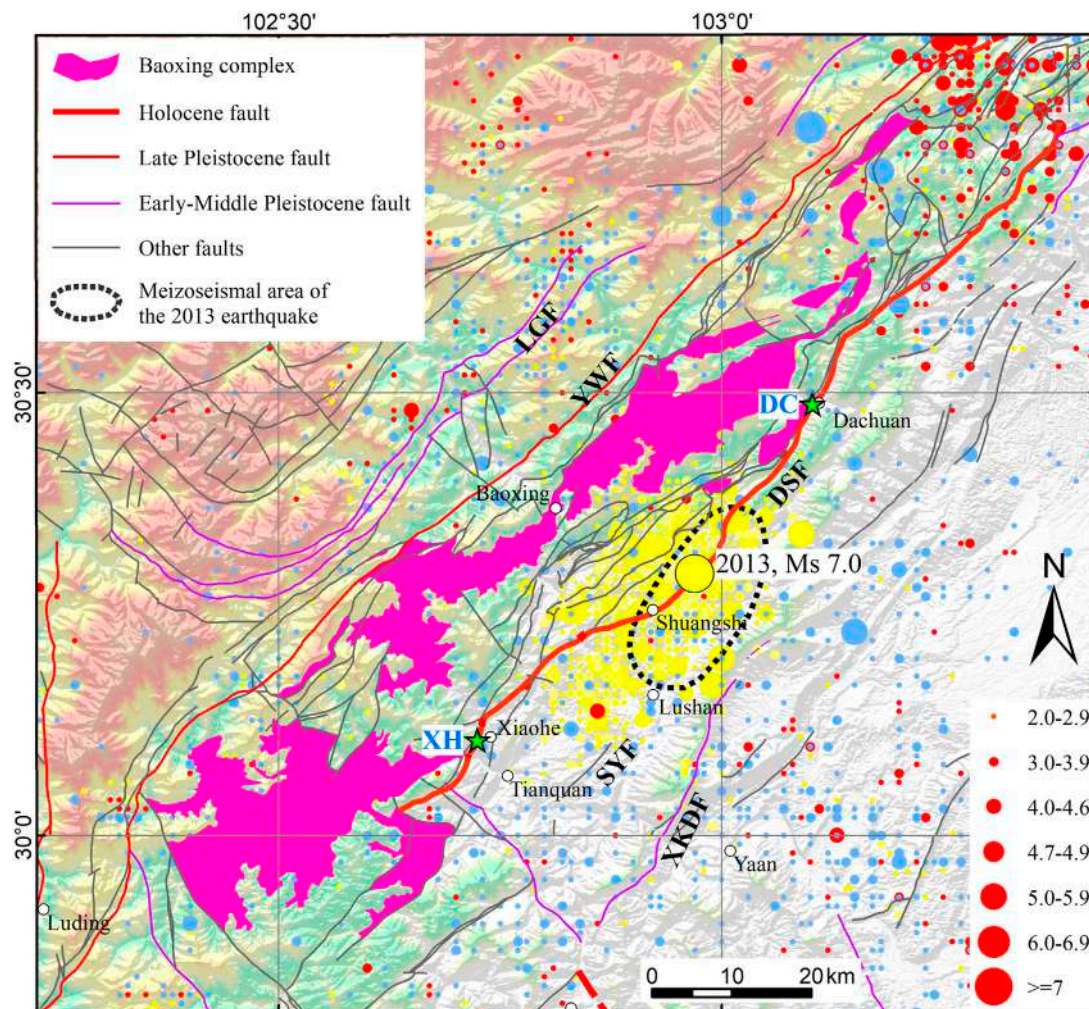


FIGURE 2 | Map showing the topographic distribution of the southern segment of the LFZ and earthquake epicenters ($M \geq 2.0$). The blue circles denote events before the 2008 Wenchuan earthquake, red circles denote events after the 2008 Wenchuan earthquake but before the 2013 Lushan earthquake, and yellow circles denote relocated aftershocks of the 2013 Lushan earthquake. Fault names: XKDF: Xinkaidian and SYF: Shiyang (see **Figure 1** for other symbolic meanings).

and the footwall consists of Upper Permian basalt. Samples of this fault belt were collected (**Figure 4; Table 1**).

Fault gouge samples were processed according to the method adopted by Tanaka et al. (2001). Each sample was carved into four slabs (for preservation, mineralogical sampling, hand specimens, and thin sections). The thin sections of the core samples from the Dachuan site were made by cutting parallel to the long axis of the core and perpendicular to the fault plane. For the outcrop samples from the Xiaohe site, the cuts were made perpendicular to the strike and plane of the fault. Then the thin sections were examined by optical microscopy (**Figures 5–7**), with selected samples analyzed by a scanning electron microscope (SEM) (**Figure 8**).

To investigate the mineralogical composition of the fault gouges, X-ray diffraction (XRD) was used in this study. XRD analyses for the bulk-rock relative mineral content and semi-quantitative analyses for clay minerals were performed with a Dmax X-ray powder diffractometer (12 kW, 45 kV, and 100 mA)

at the Beijing Micro Structure Analytical Laboratory at Peking University Science Park. Unoriented powder bulk samples were scanned with 2θ ranging from 3° to 45° with a step size of 0.02° by CuK- α radiation (0.15418 nm) and a scan rate of $2^\circ/\text{min}$. The relative amount of bulk clays in the bulk powder sample was evaluated, the proportions of which were further determined by analyzing the centrifugation-separated clay-sized portion ($b2 \mu\text{m}$). The clay-water slurry was air-dried on glass slides, and ethylene glycol was added to detect swelling components. Then the samples were heated to 550°C to distinguish kaolinite from chlorite.

Microstructural Observations Dachuan Borehole Site

Sample DC4-1

Photographs taken under a stereoscope show that the fault gouge between the two walls of the fault (sandstone W1 and mudstone W2) occurs in three different colors: gray-black (X2), gray (X3),

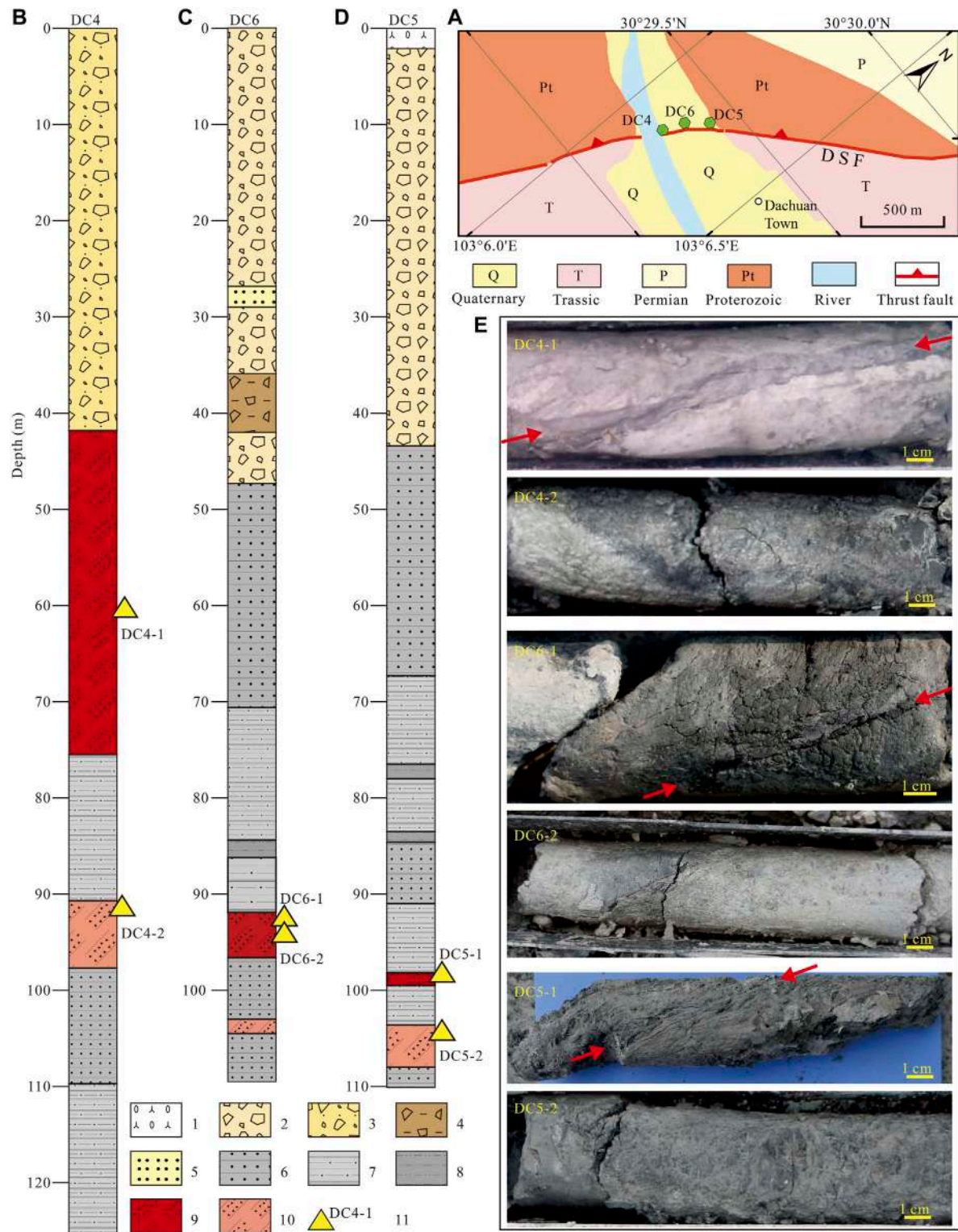


FIGURE 3 | (A) Geological map showing the borehole positions (green dots) at the Dachuan site (DC in Figure 1B and Figure 2) relative to the DSF. **(B–D)** Core logs of boreholes DC4, DC6, and DC5. Legends: 1: artificial soil fill, 2: boulder, 3: gravel intercalated with clay, 4: sandy gravel, 5: sand bed, 6: sandstone, 7: fine sandstone, 8: silty sandstone, 9: active fault belt, 10: old fault fracture belt, and 11: sampling site and number. **(E)** Photographs of core samples with red arrows showing locations of localized, soft, and thin fault gouge in fractured zones.

TABLE 1 | Descriptions of fault gouge samples from the DSF.

Sampling sites	Sample number	Description of samples
Dachuan site	DC4-1	From depths of 60.2–60.4 m in borehole DC4 and the upper fracture belt, with visible ~1-cm-thick loose soft fault gouge and detrital rock on either side
	DC4-2	From depths of 91.4–91.6 m in borehole DC4 and the lower fracture belt, with detrital rock intercalated with coal beds
	DC5-1	From depths of 98.2–98.4 m in borehole DC5 and the upper fracture zone, with broken cataclastic mudstone and thin layers of loose soft fault gouge
	DC5-2	From depths of 105.1–105.3 m in borehole DC5 and the lower fracture belt, with detrital rock
	DC6-1	From depths of 92.7–92.9 m in borehole DC6 and the upper fracture belt, with broken cataclastic mudstone and 3–5-cm-wide loose soft fault gouge
	DC6-2	From depths of 94.7–94.9 m in borehole DC6 and the upper fracture belt, with broken cataclastic mudstone
Xiaohe site	XH1	From cataclastic bedrock in the hanging wall of the fault and ~6 cm from the interface of the fault gouge belt
	XH2	From cataclastic country rock in the hanging wall of the fault, contacting with the fault gouge
	XH3	From the fault gouge and ~2 cm from the boundary between the fault gouge belt and hanging-wall cataclastic bedrock, as gray-white fault gouge
	XH4	From near the center of the fault belt (slightly closer to the footwall) and as gray-white fault gouge, with schistosity oblique to the slip plane
	XH5	From the gray-black breccia belt in the footwall of the fault, contacting with the fault gouge, with compressive schistosity visible
	XH6	From the gray-black breccia belt in the footwall of the fault and ~8 cm from the interface of the fault gouge belt

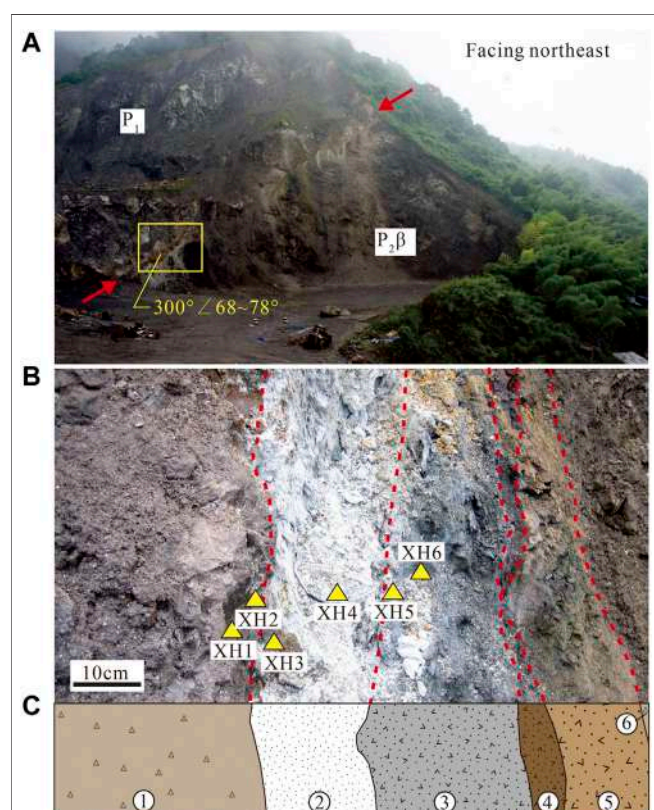


FIGURE 4 | (A) Photograph of the Xiaohe site (location shown in Figure 1B) with the fault trace denoted by red arrows. P₁: Lower Permian limestone and P₂β: Upper Permian basalt. The sampled area is shown by the yellow box, which is enlarged in (B) and contains the fault gouge sample sites XH1–XH6 (yellow triangles). (C) Interpretation of the fault belt. ① Cataclasite at the hanging wall, ② gray-white fault gouge, ③ gray-black fault breccia, ④ ferruginous vein, ⑤ yellow breccia, and ⑥ cataclasite at the footwall.

and light-gray (X4). The gouge zone is 2.5 cm thick, of which X2, X3, and X4 are 0.5–1.0, ~10, and ~15 mm, respectively (Figures 5A,B). Under an optical microscope, no obvious microscopic deformation structures were observed in walls W1 and W2 (Figures 5C,D). In fault gouges X2, X3, and X4, there were a series of subparallel and discontinuous fracture cleavages; of these, the scale and intensity in gouge X2 were greatest, followed by X3 and then X4. Inside these fault gouges, pores and fissures likely related to fluid migration were observed, and some pores developed in the solution seams (Figures 5E–G). Under the SEM, the boundary between the fault gouge and the sandstone cataclasite was obvious. The fragments, appearing principally as long angular strips, were mainly quartz grains with rare ilmenite and microcline. In the sandstone cataclasite, the debris was mostly larger than 20 μm and not arranged, and no second deformation or fracture was observed. On the boundary between cataclasite and the gouge, the particles were significantly reduced and generally smaller than 15 μm, the amount of clay increased, and the fragments were oriented and mostly clean and intact, with tiny amounts of clasts seen as reworked micro-cracks, or fragments (Figures 8A,B). Inside the gouge, particles larger than 15 μm were rare, fine grains were oriented (Figure 8B), and micro- to nanometer-sized grains were observed (Figure 8C).

Sample DC4-2

This sample contains black and light-gray fault gouges with stripe-like features and clear edges. The parent rock of the black gouge is coal, and that of the light-gray gouge is mudstone. A series of discontinuous cracks are present in the light-gray fault gouge, parallel to the long-axis of the drill core (arrows in Figures 5H,I). Cracks healed by calcite can be occasionally observed along the fault (Figure 5H).

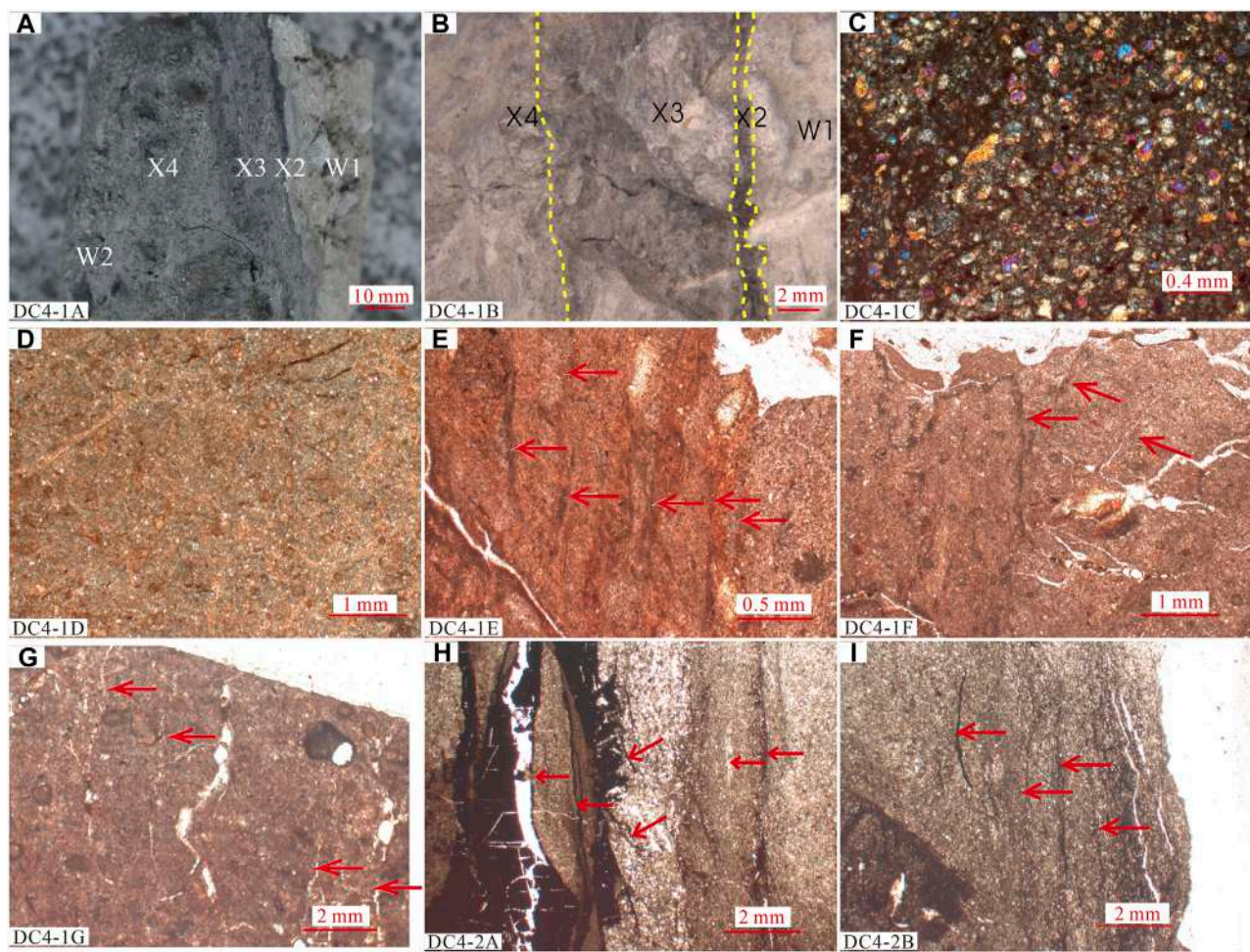


FIGURE 5 | Photographs of sample DC4-1 and DC4-2. **(A)** Core sample containing fault gouges X2, X3, and X4. W1 and W2 denote the country rock (sandstone and mudstone, respectively) on either side of the gouge. **(B)** Stereoscopic image showing the fine textures of the fault gouge samples X2, X3, and X4. **(C,D)** Photomicrograph of country rock W1 and W2, respectively, showing the lack of deformation. **(E–G)** Photomicrographs of sample DC4-1. **(H,I)** Photomicrographs of sample DC4-2. Arrows show discontinuous fracture cleavage or fissures.

Sample DC5-1

The fault gouge in DC5-1 contains en echelon fold structures, which are presumably formed under compression, and a quasi-S-C fabric (letters “S” and “C” in **Figure 6A**). This sample also contains discontinuous fracture cleavage (arrows in **Figure 6B**) and straight, fine striations (arrows in **Figure 6C**). As shown in **Figure 6D**, the debris and mineral fragments between the mudstone and fine sandstone show a strong orientation, indicating the sharply defined fault contact. Structural deformation is concentrated in the mudstone on the left, with two periods of activity being evident (arrows F1 and F2 in **Figure 6D**); F2 is cut by F1 at an inclination of $\sim 30^\circ$. In addition, pressure solution seams can be observed locally in this sample (**Figure 6A**).

Sample DC5-2

This fault gouge also has an irregular, striped appearance with quasi-S-C fabrics (letters “S” and “C” in **Figure 6E**). Some of this S-C fabric folded during a compression event (**Figure 6F**).

Sample DC6-1

This gouge comprises two main colors, gray-black and light gray, and also displays a striped appearance. This feature is indicative of the newer fault trace, which is distinct and straight, at the interface of the two sets of gouge (**Figure 6G**). In the gray-black gouge, there are many subparallel fracture cleavages (arrows in **Figure 6G**), which transect the black stripes and intersect the newer fault plane at an angle of $\sim 18^\circ$. In the light gray gouge, there

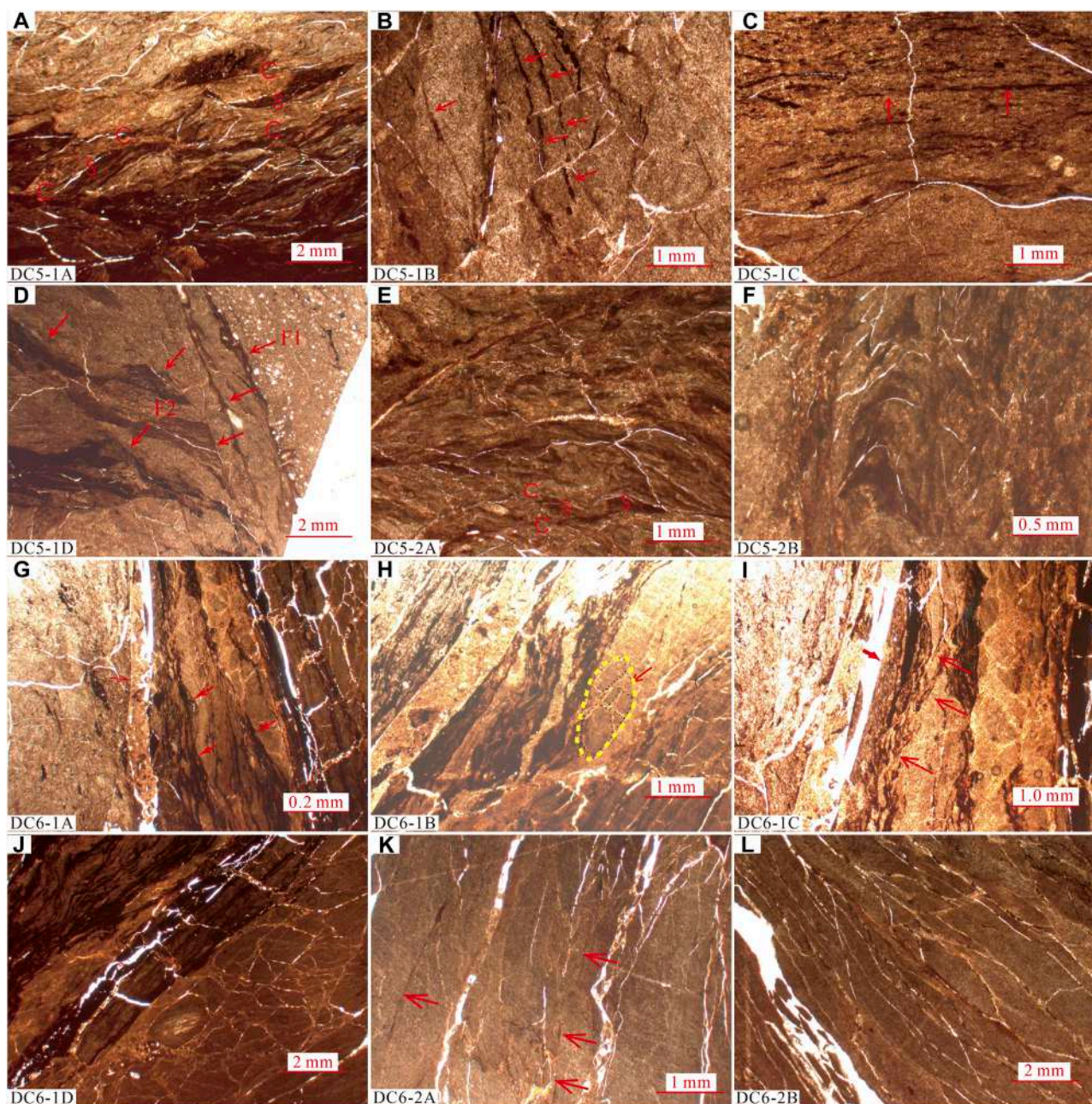


FIGURE 6 | Photomicrographs of samples DC5-1 (A–C), DC5-2 (D–F), DC6-1 (G–J), and DC6-2 (K,L). S and C denote quasi-S-C fabrics, and arrows indicate cracks or fracture cleavage. Yellow dashed lines mark reworked clasts.

are some mortar textures, with the long axis being subparallel to the fault plane (**Figure 6H**). Both sets of the fault gouge contain cracks, of which the density in the light gray gouge is higher than that in the gray-black gouge (**Figures 6I,J**).

Sample DC6-2

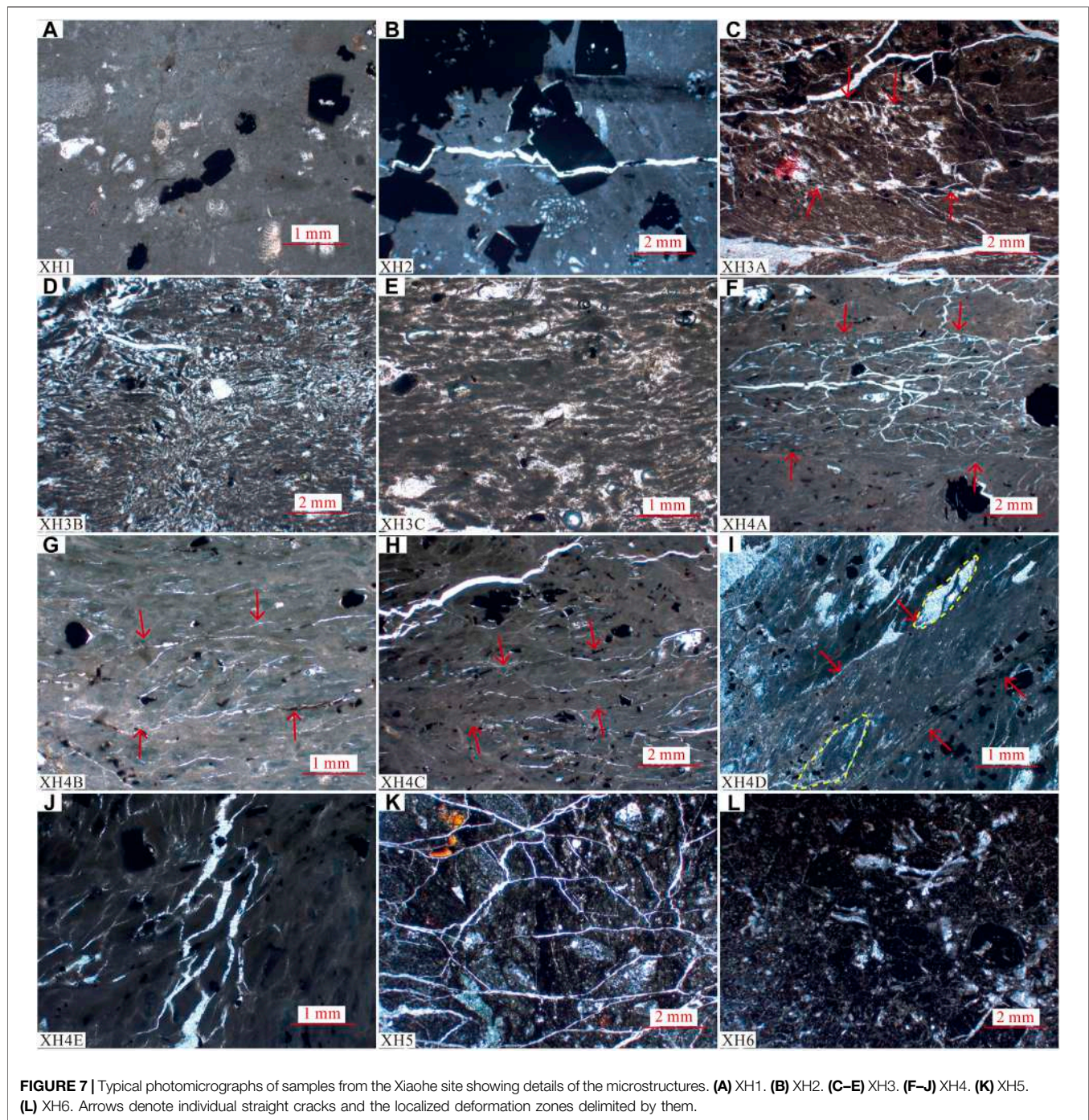
This sample comprises the fault gouge formed in mudstone and is light gray in color. Simple deformation is evident, with subparallel

but discontinuous fracture cleavage (**Figure 6K**) or dense subparallel fissures (**Figure 6L**).

Xiaohe Outcrop Site

Sample XH1

Microscopic observation of this gouge reveals only minor cataclasis of mineral grains, with no preferred orientation or arrangement of long axes visible (**Figure 7A**).



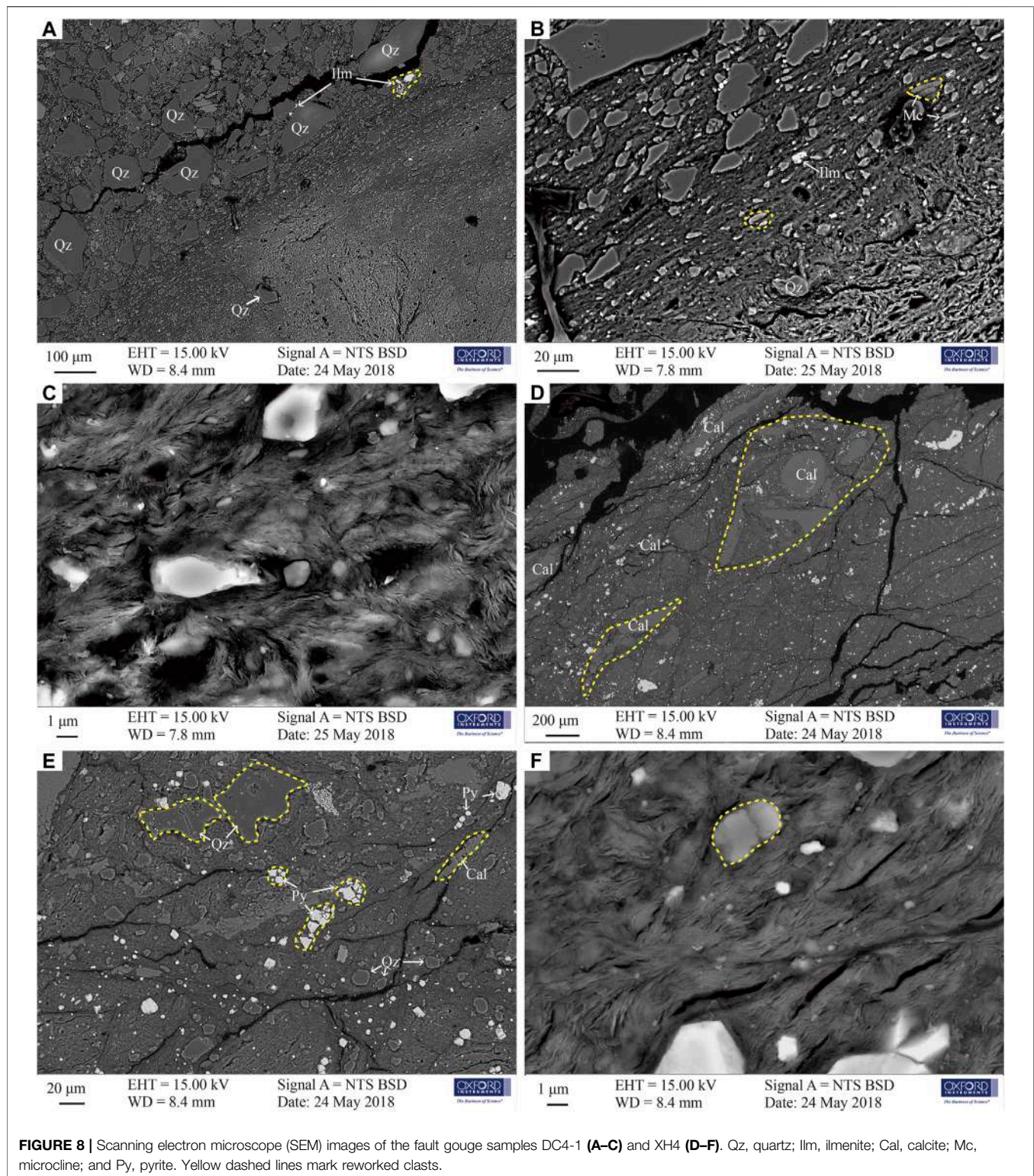
Sample XH2

This gouge contains relatively long but irregular and unfilled cracks that transect sheet-like mineral grains locally (Figure 7B).

Sample XH3

This sample contains numerous sets of cracks and foliations (Figures 7C–7E), the broadest of which are filled with calcites, thus forming vein-like textures. The cracks are mostly disordered

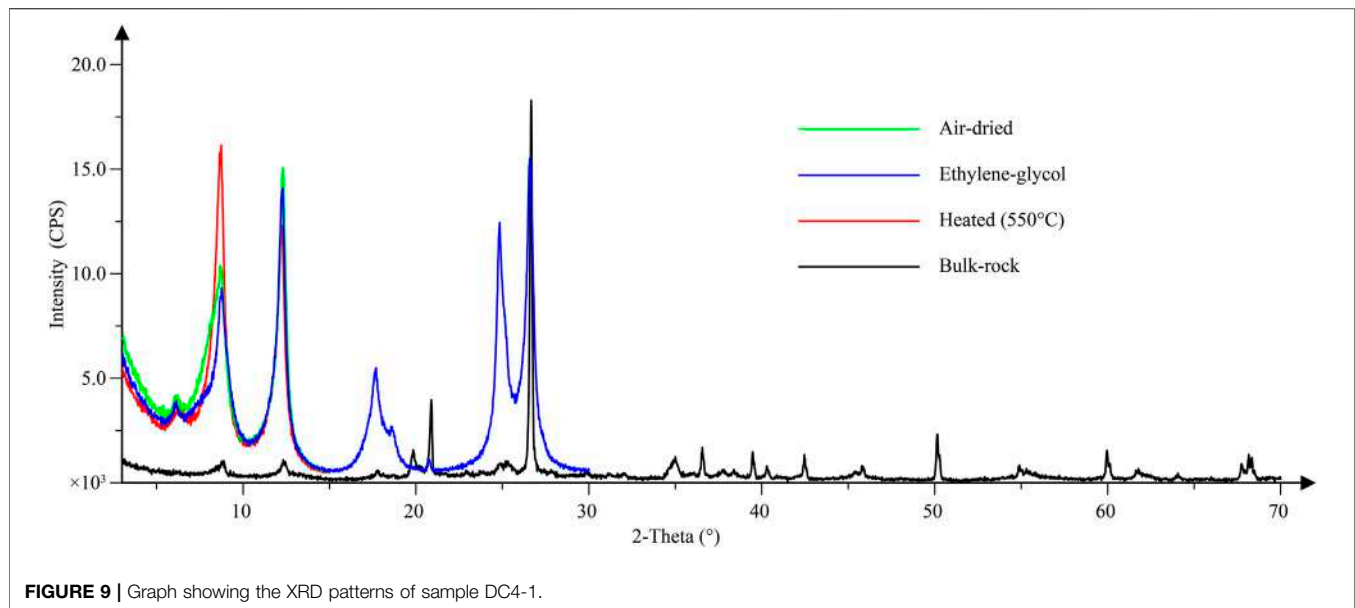
and v are not individually pervasive. Only two cracks (arrows in Figure 7C) are relatively straight, but are not very long, and bound by a domino-like structure consisting of rock fragments. The two sets of cracks mentioned before constitute a quasi-S-C fabric. In addition, it was observed that preexisting foliations were deformed by later cleavage in which mineral grains were oriented, with disorderly mineral arrangement at the intersection of the two foliations (Figure 7D). Pores and healed fissures related to fluid migration were observed (Figures 7C–7E).



Sample XH4

Under an optical microscope, this sample has a much denser distribution of foliations and cracks than sample XH3. In addition, porphyroclasts, cataclastic minerals, and

porphyrotopes are present in the microstructure. **Figure 7F** shows a zone of crack concentration that is limited by two straight cracks, outside of which several oblique fissures form a quasi-S-C fabric in conjunction with the principal foliations. As



shown in **Figure 7G**, there are rock fragments bounded by two straight cracks, which overall constitutes a domino-like fabric. These cracks intersect the principal foliations at low angles with evidence of cataclastic mineral grains (**Figure 7H**). **Figure 7I** displays numerous foliations and quasi-S-C fabrics in which debris is delimited by two straight cracks, forming an oriented domino-like fabric. In addition, some fragments are reworked and show fractures or micro-cracks. As shown in **Figure 7J**, the preexisting foliations are transected by quite broad younger cracks that contain replacement filling of calcite and intersect the foliations at high angles. Pores and fissures possibly related to fluid migration can be observed in this gouge, and pores are occasionally developed in the healed fissures (**Figures 7F–I**). According to the SEM observation, the fragments in sample XH4 are mainly composed of calcite, quartz, and pyrite, and the directional arrangement and reworked deformation or fragmentation are obvious. Calcite grains, which are most abundant, are large and mainly vein-shaped. Compared with the calcite, the quartz and pyrite particles are smaller (generally $<20\ \mu\text{m}$), subangular, and irregular in shape. The phenomenon of clasts being wrapped by clay is apparent, and micro- to nanometer-sized grains could be observed (**Figures 8D–F**).

Sample XH5

This sample also contains preexisting foliations transected by later cracks, along with cataclastic rocks, and relatively straight but branching cracks (**Figure 7K**).

Sample XH6

Almost no deformation is visible in the microstructure of this sample, but several fissures or some cataclases are evident (**Figure 7L**).

Overall, the country rock samples from the Xiaohe site display very little deformation in their microstructure, while inside the fault gouge, sample XH4, collected from near the center of the fault, shows relatively abundant deformation concentration.

Mineralogical Results From XRD Analyses

XRD mineralogical analyses were performed on the fault gouge from core sample DC4-1 and outcrop sample XH4, based on the abovementioned microstructural observation results.

As shown in **Figure 9** and **Table 2**, the mineralogical analysis identified quartz and clays as the major minerals, which reach contents of 49% and 44%, respectively, for the fault gouge of sample DC4-1. In addition, small amounts of albite and microcline were detected. The clay in sample DC4-1 is composed of illite, and mixed layers of illite/smectite, kaolinite, and clinocllore, and no dissociative smectite was found. Among these clay minerals, the content of illite is the highest (39%), followed by mixed layers of illite/smectite, and the clinocllore content is lowest at 13%. The ratio of the illite/smectite mixed layer is only 15%.

Fault gouge sample XH4 has a richer variety of minerals than sample DC4-1, including clay, calcite, quartz, pyrite, and small amounts of albite, microcline, and gypsum. The content of calcite is higher ($\sim 29\%$) than that of quartz ($\sim 16\%$), but clay content is 36%. The clay minerals, similar to that in sample DC4-1, are composed of illite, mixed layers of illite/smectite, kaolinite, and clinocllore. No dissociative smectite was found. Of the clay minerals present, illite has the highest content, reaching 50%, followed by the mixed layers of illite/smectite, and the clinocllore content is lowest at only 6%. The ratio of the illite/smectite mixed layer is also 15% (**Figure 10** and **Table 3**).

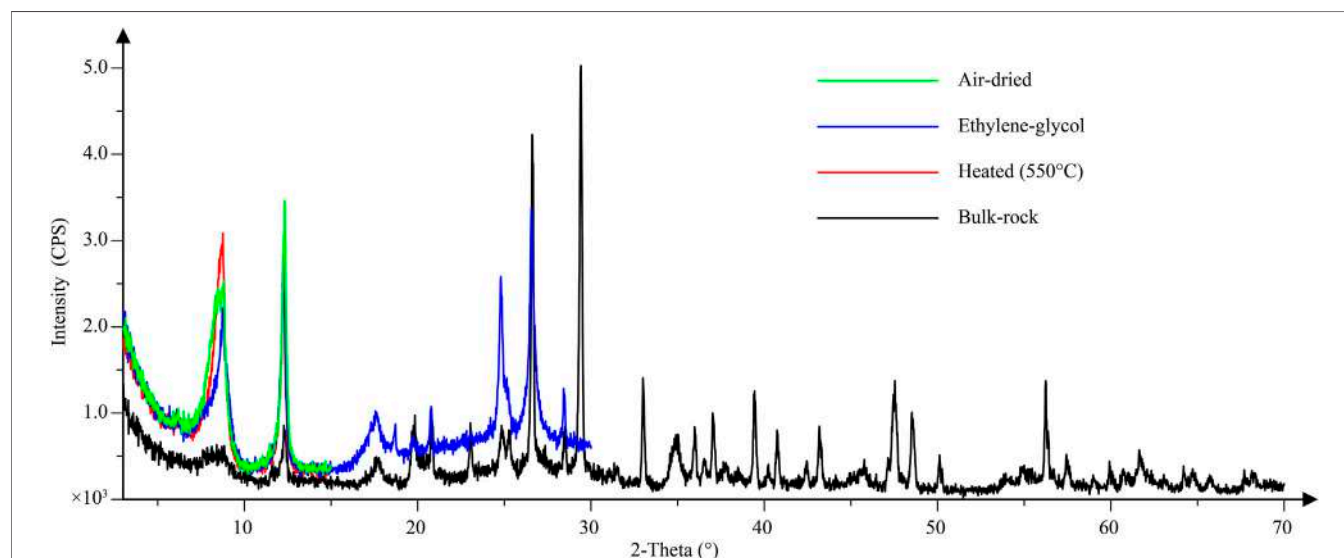
DISCUSSION

Slip Behavior of the DSF Revealed by Fault Gouge Microanalysis

Previous studies about experimental gouge and nature fault gouge have suggested that irregular striations, quasi-S-C fabrics, and

TABLE 2 | Bulk-rock and clay mineral compositions of fault gouge sample DC4-1.

Bulk composition (%)							
Quartz	Albite	Microcline	Calcite	Pyrite	Gypsum	Total clay	
49	3	4	—	—	—	44	
Clay composition (%)				Mixed-layer ratio (%S)			
Smectite	Illite/smectite	Illite	Kaolinite	Clinochlore	Clinochlore/smectite	Illite/smectite	Clinochlore/smectite
—	27	39	21	13	—	15	—

**FIGURE 10 |** Graph showing the XRD patterns of sample XH4.**TABLE 3 |** Bulk-rock and clay mineral compositions of fault gouge sample XH4.

Bulk composition (%)							
Quartz	Albite	Microcline	Calcite	Pyrite	Gypsum	Total clay	
16	2	4	29	10	3	36	
Clay composition (%)					Mixed-layer ratio (%S)		
Smectite	Illite/smectite	Illite	Kaolinite	Clinochlore	Clinochlore/smectite	Illite/smectite	Clinochlore/smectite
—	28	50	16	6	—	15	—

small folds in the fault gouge are associated with fault creep, whereas localized brittle deformation features such as discontinuous fracture cleavages, straight cracks or striations, concentrated zones of cracking, and other deformation are caused by stick-slip during faulting (Moore et al., 1989; Reinen, 2000; Zhang et al., 2002; Yuan et al., 2013).

The microstructure observations mentioned before demonstrate that samples DC4-1, DC5-1, and DC6-1, collected from the upper fracture zone shown in the cores at the Dachuan site, have more abundant signs of deformation such

as straight fracture planes, cracks or striations, subparallel fracture cleavage, and preferred orientations of porphyroclasts, and an overall higher degree of deformation than samples DC4-2 and DC5-2 (from the lower fracture zone). It can reasonably be concluded that the upper fracture zone has been the most recently active part of the DSF. Sample DC4-1 shows almost no microstructural deformation in the country rocks on either side of the fault gouge (Figures 5C,D). Thus, the structural deformation is concentrated in the ~2.5-cm-thick fault gouge in the center of the fracture zone (Figures 5A,B).

In the country rock on either side of the fault gouge at the Xiaohé site, samples XH1, XH2, XH5, and XH6 show little or no microstructural deformation, which is in sharp contrast to samples XH3 and XH4 within the fault gouge (**Figure 7**). In particular, XH4, which was collected from near the center of the fault gouge, has the most conspicuous fractures or micro-cracks, oriented fragments, and reworked clasts, which imply a decreasing trend of deformation from the center to either side of the fault plane. Such deformation is unevenly distributed, with a 1- to 5-mm-thick zone composed of cleavages or micro-cracks or domino-like arranged fragments bounded by two subparallel straight cracks, within which mineral grains or debris have a preferred orientation. Outside of this zone, only a few micro-cracks developed.

According to Han et al. (2010) and Duan et al. (2016), the newest fault gouge in a fault belt is the softest, thinnest, and deepest in color, with almost no breccia inside and a clear, flat boundary with other parts. Therefore, it can be speculated that the gray-black and relatively loose and soft gouge X2 with 0.5- to 1.0-mm thickness (**Figures 5A,B**) in sample DC4-1 and the localized brittle deformation zone bounded by two straight cracks in sample XH4 are the result of the most recent stick-slip event (**Figures 7F-I**).

Previous studies suggest that the illite-rich fault gouge is the product of the fault slip that occurs during an earthquake, whereas the smectite-rich fault gouge is the product of fault creep (Chen et al., 2007; Isaacs et al., 2007). In this study, relatively abundant illite contents were found in the fault gouge at both the Dachuan boreholes and the Xiaohé outcrop sites, and almost no smectite was present (**Tables 2, 3**).

Stick-slip event of the fault is generally accompanied by co-seismic frictional heating, and experimental research works showed that the flash heating can cause thermal decomposition of minerals to form nanoparticles (Han et al., 2007). Then the existence of micro- to nanometer-sized grains can greatly reduce the friction strength of the fault and play a role of lubrication and drag reduction in the fault slip, enabling the fault slip in a large scale (Yuan et al., 2014). As mentioned before, abundant micro- to nanometer-sized grains were developed in the fault gouge of the samples DC4-1 (**Figures 8A-C**) and XH4 (**Figures 8D-F**).

Friction experiments show that the existence of water can lead to the transformation of frictional sliding from stable sliding to unstable sliding, which would result in seismic nucleation. Therefore, a more intense fluid activity in the fault gouge may correspond to more seismic events in the fault history (He et al., 2006; He et al., 2007; Han et al., 2010). Abundant pores could relate to fluid migration developed in the fault gouge in samples DC4-1 and XH4 from the two study sites.

Based on a combination of the microstructure observations and mineralogical analysis results, the DSF presents obvious stick-slip behavior and has the ability to produce devastating earthquake confirmed by paleoseismological studies (Densmore et al., 2007; Chen et al., 2013b; Chen et al., 2014; Dong et al., 2017). It should be pointed out, of course, that other mechanisms in addition to fault slip may involve the development of microstructures of our fault gouge. For example, the development of micro- to nanometer-sized particles observed

in our fault gouge may involve thermal decomposition in parts, and the development of healed fluid migration fissures may relate to the pressure solution.

Differences Between DSF and BYF in Fault Gouge Microstructures and Mineralogical Composition

As shown in **Table 4**, features of the fault gouge of the DSF in the southern segment of the LFZ, as revealed by borehole and outcrop observations, differ significantly from observations made in trenches and drill-holes in the middle-northern segments. For instance, in the southern segment, the thickness of the fault gouge produced by single events is considerably less than that in the middle-northern segments, and the country rock is less fractured on a microscopic scale. Additionally, the fracture system in the southern segment is relatively simple and mainly consists of micro-cracks or discontinuous fracture cleavage. However, oriented fragments or clasts are distinct in the fault gouge from the southern LFZ, unlike in the fault gouge from the middle-northern segments in which reworked clasts are also evident.

With the exception of clay, the mineralogical composition of fault gouge does not seem to be comparable at study sites in the LFZ. The total clay content in the relatively new fault gouge in the middle-northern segments of the LFZ is generally more than 50%, or even more than 70% (Yuan et al., 2013; Wang et al., 2014; Duan et al., 2016), whereas that of our two study sites in the southern segment was only around 40%. In addition, the clay mineral composition in the fault gouge also differs significantly between the southern and the middle-northern segments. For example, the contents of the illite/smectite mixed layer and chlorite are significantly lower in the southern segment, whereas the content of kaolinite is significantly higher. The illite/smectite mixed layer ratio in the southern segment is significantly smaller than that in the middle-northern segments (**Tables 2-4**).

In terms of ductile deformation features, the microstructures of the fault gouge from the south segment of the LFZ is different from that from the middle-northern segments. For instance, the S-C fabric from the gouge of the middle-northern segments is relatively strong and obvious, while that from the gouge of the southern segment are very untypical and we call it the quasi-S-C fabric.

Previous research works on structural geology revealed that the DSF is unique in the southern segment of the LFZ because it contains records of Holocene offset at the surface (Densmore et al., 2007; Chen et al., 2013b; Chen et al., 2014; Dong et al., 2017). However, in contrast to the middle-northern segments of the LFZ, which show clear fracture surfaces and >1-m co-seismic displacement (Xu et al., 2009; Ran et al., 2010a; Ran et al., 2010b; Ran et al., 2013; Ran et al., 2014), the DSF produced only tiny offsets near the surface. Therefore, these studies suggested that, in contrast to middle-northern LFZ, the single faults in southern LFZ are less active and lead to lower seismic activity because they have more branches distributed over a larger area, allowing tectonic deformation to be

TABLE 4 | Comparison of fault gouge microanalysis results between the DSF and BYF.

DSF			BYF
Fault gouge thickness of single fault event	0.5–1.0 mm* or 1–5 mm [#]		>10 mm ^{*,#a,b,c}
Country rock around fault gouge	No notable trace of structural deformation* or a few discontinuous non-straight micro-cracks [#]		Many (micro-) cracks, diamond-like cracks or shear planes ^{*,#b,d} ; local severely fractured breccias, and widely distributed fissures ^{#,c}
Features of debris or mineral clasts	Oriented arrangement ^{*,#} or a few porphyroblastic features [#]		Obvious oriented arrangement, many dislocations of debris or mineral clasts ^{*,#b,d,e}
Ductile deformation	Local features of quasi-S-C fabrics and small folds ^{*,#}		Strong S-C fabrics, notable P-foliations, extended debris grains, and asymmetric drag structures ^{*,#a,b}
Total clay content	44%* or 36% [#]		~50% or higher ^{*,f,g} or 72.8% ^{#,b}
Clay mineral composition	Relative content	I > I/S > Kln > Chl ^{*,#}	I/S (~50%) > I > Chl (~30%) > Kln (almost 0%) ^{*,f,g} , or I > I/S (40%) > Chl (>7%) > Kln (1%) ^{*,b,f}
	I/S mixed layer ratio	15% ^{*,#}	>50% ^{*,f,g} or 35% ^{*,b,f}

Note: Symbol * denotes fault gouge (rock) from boreholes, and [#] is fault gouge exposed on the surface.

^aLi et al. (2013b).

^bYuan et al. (2013).

^cHan et al. (2010).

^dSi et al. (2014).

^eChen et al. (2016).

^fWang et al. (2014).

^gDuan et al. (2016).

The sampling sites of those study are shown in **Figure 1**. I: illite, I/S: illite/smectite mixed layer; Kln: kaolinite, Chl: clinochlore.

accommodated by more faults (Densmore et al., 2007; Chen et al., 2013b; Chen et al., 2014; Dong et al., 2017). Coincidentally, the microstructure, clay content, and clay mineral content of the fault gouge varied greatly between the southern and middle–northern LFZ. That is, the results of our microanalysis of the fault gouge further support the inference that the southern segment of LFZ is not like the middle–northern segments and is unlikely to experience an *M*_s 8.0 or larger earthquake.

Implications to identify minor-surface-ruptures by microscopic analysis of fault gouge.

According to the comparison of the microscopic characteristics of the abovementioned gouges, whether collected from near the surface or from boreholes, it can be seen that the DSF with minor co-seismic displacement is systematically different from the BYF with intense activity. These differences may be caused by the intrinsic differences in the active characteristics of the two types of faults; that is to say, the information identified from the fault gouge of the DSF may serve as a preliminary marker for identifying minor-surface-rupture faults. To sum up, the obvious signs about fault gouge are as follows:

- 1) fault gouge thickness of a single fault event is less than 5 mm;
- 2) under a microscope, no obvious micro-crack was examined in surrounding rocks around the fault gouge; and observed micro-cracks are discontinuous and S-C fabrics are not very typical in the fault gouge.
- 3) under SEM, mineral debris reprocessing phenomenon is rare.
- 4) XRD mineral analysis shows that the total clay content is less than 50%, the content of Kaolinite is obviously higher than

that of Clinochlore, and the content of Illite/Smectite mixed layer is less than 30%.

These markers may provide useful evidence for identifying minor-surface-ruptures and make up for the limitations of traditional geological methods.

CONCLUSION

This study investigated the microstructures and mineral composition of the fault gouge exposed in outcrop and from boreholes in the DSF in the southern segment of the LFZ. The microstructural results revealed a series of localized brittle deformations mainly produced by stick-slip of the fault, such as fracture cleavages, straight cracks, and striations, and strongly deformed zones consisted of these deformation styles. In addition, fluid migration pores or fissures and micro- to nanometer-sized grains, related to earthquake occurrence and possibly involved by pressure solution and thermal decomposition, respectively, were also observed. The mineralogical composition revealed by the XRD analysis shows total clay content of 44% in the fault gouge from the Dachuan borehole site and 36% from the Xiaohe outcrop site. Abundant illite content, considering the result of fault slip when an earthquake occurs, was found, with 39% and 50% at Dachuan and Xiaohe sites, respectively. In conclusion, the microanalysis results of the fault gouge show that the DSF has the behavior of stick-slip, which supported by paleo-seismic researches along the fault.

A comparison analysis demonstrated that the DSF, one minor-surface-rupture fault in the southern segment of the LFZ, has

significantly different fault gouge microstructure and mineralogical composition from those in BYF with intense activity in the middle–northern segments of the LFZ. We summarized several representative characteristics of DSF that are different from BYF: 1) Macroscopically, fault gouge thickness of a single fault event is thinner. 2) Microscopically, micro-cracks in surrounding rocks and the debris reprocessing phenomenon in the gouge were rare, and discontinuous micro-cracks and untypical S-C fabrics in the gouge are common. 3) In terms of mineral composition, the total clay content is less than 50%, the content of kaolinite is obviously higher than that of clinocllore, and the content of illite/smectite mixed layer is less than 30% in the gouge. These features are systematic; therefore, comparing these fault gouge markers can help us to identify minor-surface faults.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author.

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AUTHOR CONTRIBUTIONS

YL mainly participated in the field investigation, sample collection, and analysis, and completed the writing of the article. LC mainly participated in the fieldwork, formed the idea of the article, and supported the completion of the article. YR helped with the writing of this article. YC participated in the microstructural observations of the fault gouge under a microscope. All authors contributed to the article and approved the submitted version.

FUNDING

This work was supported financially by the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901); Basic Scientific Work of the Institute of Geology, China Earthquake Administration (IGCEA1418); and National Key R&D Program of China (2019YFC0604901). The topography data used in this study (SRTM data V4) were obtained from the following website: <http://srtm.csi.cgiar.org>.

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Three-Dimensional Fault Model and Activity in the Arc-Shaped Tectonic Belt in the Northeastern Margin of the Tibetan Plateau

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OPEN ACCESS

Edited by:

Zhuqi Zhang,
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 10 March 2022

Accepted: 10 May 2022

Published: 30 May 2022

Citation:

Li L, Li Z, Sun C, Wang W, Dai X, Zhang Y, Liang H, Xu B, Wu X and Lv L (2022) Three-Dimensional Fault Model and Activity in the Arc-Shaped Tectonic Belt in the Northeastern Margin of the Tibetan Plateau. *Front. Earth Sci.* 10:893558. doi: 10.3389/feart.2022.893558

The arc-shaped tectonic belt, located in the northeastern margin of the Tibetan Plateau, is one of the leading edges of the plateau's outward growth and uplift expansion, with a large number of active faults and frequent seismic activity. Researchers have carried out numerous studies on active faults in this region, and a wealth of reliable basic data has been accumulated. However, integrating multidisciplinary data to establish a 3D geometrical structure model that is concerned about seismogenic tectonics and can be tested, has become the key to restricting the regional seismic hazard evaluation. Based on a series of published active tectonic research, we analyze in detail the surface and deep coupling relationships of the major active faults in this region and establish three sets of 3D fault structure models, which are built respectively by active fault mapping and dip angles (the V1 model), 7 magnetotelluric profiles and 7 auxiliary profiles (the V2 model), and multi-source data (the V3 model) and continuously close to the real geological facts. From the model perspective, it is suggested that the controversial Haiyuan fault is a crustal-scale left-slip fault and the shape of the Liupanshan fault reflects the absorption of the left-slip component of the Haiyuan fault. Comparing the same fault plane of these three models, we find that the V3 model is more consistent with geological facts, showing that by assisting the multi-source data 3D geological modeling technique we can establish a 3D geological model closest to the real regional structure. Finally, combining the V3 model, the fault segmentation, and the empirical formulas of the moment magnitude-rupture parameters, we segment the faults and calculate the potential moment magnitudes of the major active faults in the study region. The faults with relatively higher seismic hazards are the Liupanshan fault, the southeastern segment of the Xiangshan-Tianjingshan fault, the westernmost segment of the Haiyuan fault, and the West Qinling fault, of which the estimated potential moment magnitudes are generally more than 7.0. Our study provides a referenced 3D geological model for exploring the deep structures of the region, regional geological research, and earthquake disaster prevention.

Keywords: active fault, 3D fault structure, 3D structural modeling, seismic hazard, Tibetan Plateau

1 INTRODUCTION

A large number of studies have shown that faults have non-single planes and complex geometric structures in 3D space, which control the rupture behavior of strong and large earthquakes (e.g., Evans et al., 2020; Ross et al., 2020). How to visually describe the fine 3D geometry of the fault underground has always been an unavoidable problem for structural geologists (e.g., Fischer and Wilkerson, 2000; Plesch et al., 2007; Guan et al., 2010). Since the concept of 3D geological modeling was proposed in the 1990's (Houlding, 1994), with the rapid development of computer technology, the establishment of 3D geological models that can reveal information about rock mass structure relationships and strata spatial distribution in computers has also become more and more sophisticated. 3D geological models have been the product of the mutual intersection of geography, geology, computer graphics, mathematics, computational geometry, environmental science, and other disciplines (Li et al., 2016b).

3D geological modeling methods are classified according to model data sources and can be divided into drilling-based, field-based geological data, geological section, and multi-source data modeling (eg, Carena and Suppe, 2002; Riesner et al., 2017; Yue et al., 2005; Wu et al., 2015; Wu et al., 2016). If it is only explained by the geometrically constrained fault structure, the results are generally uncertain and diversified (Evans et al., 2020; Li et al., 2007; Li et al., 2019b; Ramsay and Huber, 1987; Ross et al., 2020; Tavani et al., 2005). At present, the technology of building 3D geological structure models by comprehensively using a series of multi-source data such as shallow geology, artificial earthquake, and drilling data has become increasingly mature, and it has great application value whether it is for a single geological structure or complex regional active faults (e.g., Hubbard et al., 2016; Jia et al., 2010; Li et al., 2010, Li et al., 2013, Li et al., 2014; Li et al., 2019a; Li et al., 2021; Plesch et al., 2007; Shaw and Plesch, 2012; Shaw et al., 2015; Wang et al., 2020), which provides a new idea for the study of geometric forms and distribution in the 3D underground space of faults.

The arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau is one of the frontal parts of the entire plateau growing outward and uplifting expansion (e.g., Tapponnier et al., 2001; Zhang et al., 2003), located at the northernmost of the North-South seismic zone and has bred numerous active faults (Deng, 2007). Researchers have carried out a large number of active fault surveys, geophysical explorations, and seismological research in this area, so abundant and reliable original data has been accumulated. But at present, no researcher has used these data to carry out detailed 3D geological modeling work of this region. In addition, due to the boundaries of multiple active blocks in this area, the intracontinental tectonic movement and fault activity is strong, and a series of historical strong earthquakes have occurred, such as the 734 A.D. Tianshui M 7^{1/2} earthquake, the 1125 A.D. Lanzhou M 7 earthquake, the 1622 A.D. North Guyuan M 7 earthquake and the 1920 A.D. Haiyuan M 8^{1/2} earthquake (Lei et al., 2007; Mei et al., 2012; Yuan et al., 2013).

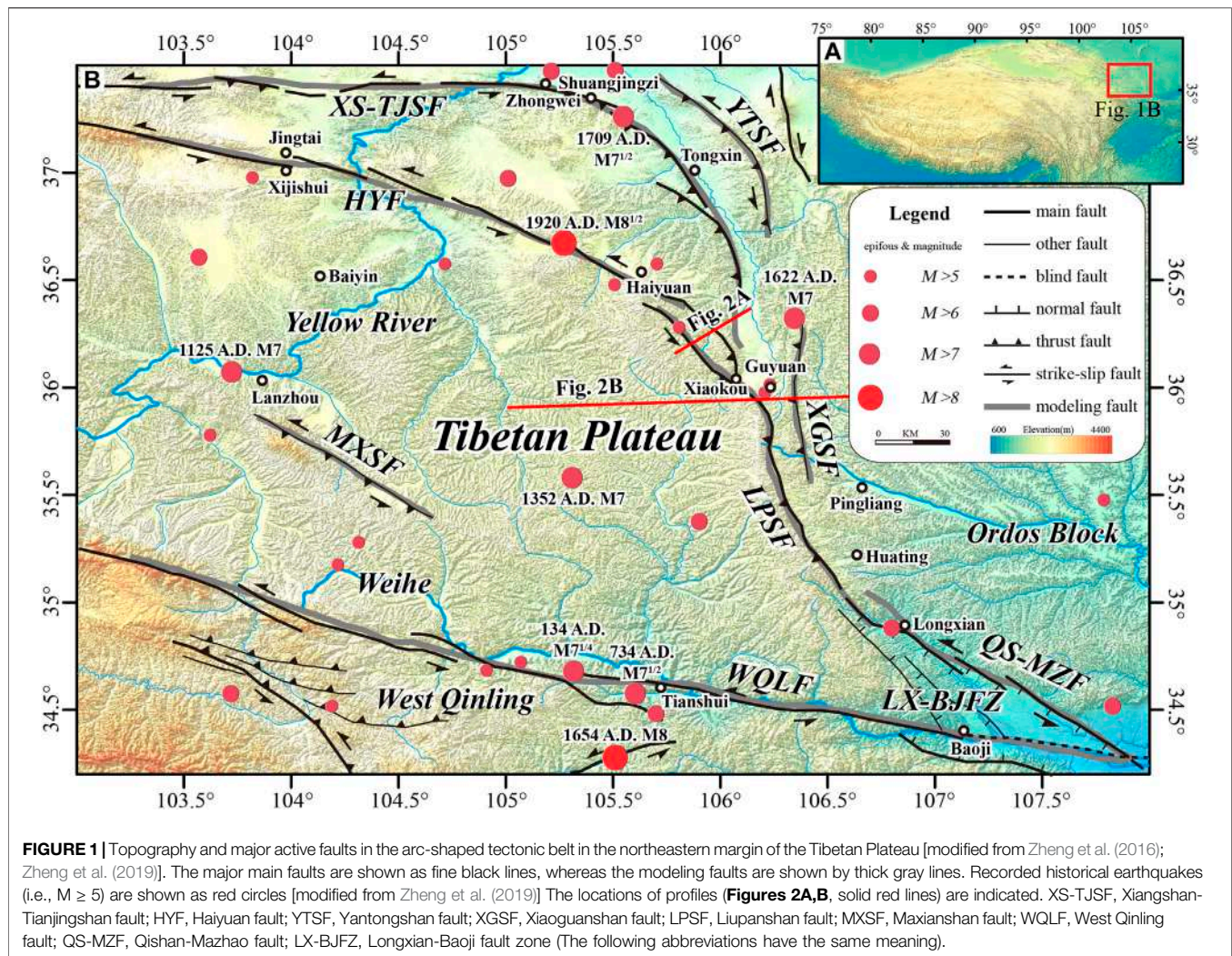
At present, some of the structures of the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau are still

somewhat controversial. For example, for the extended depth of the Haiyuan fault, different researchers have used different methods to reveal whether it cuts through the Moho, and the deep structure of the Haiyuan fault needs further research. Zhan et al. (2005) and Jin et al. (2012) revealed that the Haiyuan fault is a deep fault that cuts the Moho by magnetotelluric profiles, the comprehensive geophysical profile shows that the Moho beneath the Haiyuan fault is a complex transition zone (Li et al., 2002), and the offset of the Moho beneath the Haiyuan fault can be directly observed by receiver function profile (Ye et al., 2015). Based on field geological investigation, Burchfiel et al. (1991) conclude that the Haiyuan fault is a fault terminating at the base of the upper crust, and some seismic reflection profiles (Gao et al., 2013; Guo et al., 2016) and magnetotelluric profiles (Min, 2012; Xia, 2019) also indicate that the Haiyuan fault does not cut through the Moho. Through 3D geological modeling, we can effectively clear the structure of the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau and provide the most credible 3D geological model of this region, providing a basis for subsequent regional research work. Similar to the above problem, although there is a large amount of data accumulated in the region, a paradox of uncoupled data in the modeling process. How to efficiently integrate multiple sources of data to make the model close to reality is also a key challenge in the multi-source data 3D geological modeling technology. Ultimately, our model will serve earthquake prevention and disaster reduction in the study area with frequent earthquakes. Use the 3D geological model to study the seismic activity of major active faults, then focus on the prevention in advance in cities with large earthquake risks to reduce the loss of life and property. Therefore, the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau has become an ideal area for studying 3D geological modeling and fault activity of active faults.

To establish the 3D fault model of the main active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau, we will collect 2D profiles revealing the coupling relationship between the surface and deep in the study area, geological surveys of active faults, and regional Digital Elevation Model (DEM) data. Then we will use the above data to extract the 3D information of faults and detachment layers, assist the SKUA-GOCAD software to generate models, and explore the reasonable method of integrating multi-source data to generate the 3D geological models. Based on the models, we will update the geological understanding of the relevant active faults, analyze the deep segmental properties of the main active faults in the study area, and estimate the seismic moment magnitude of each segment of the fault, to provide a theoretical basis for regional seismic risk assessment.

2 GEOLOGICAL SETTING

In this research, 3D geological modeling will focus on the main active faults developed in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau, including the West Qinling fault (WQLF), the Maxianshan fault (MXSF), the Haiyuan fault (HYF), the Xiangshan-Tianjingshan fault (XS-

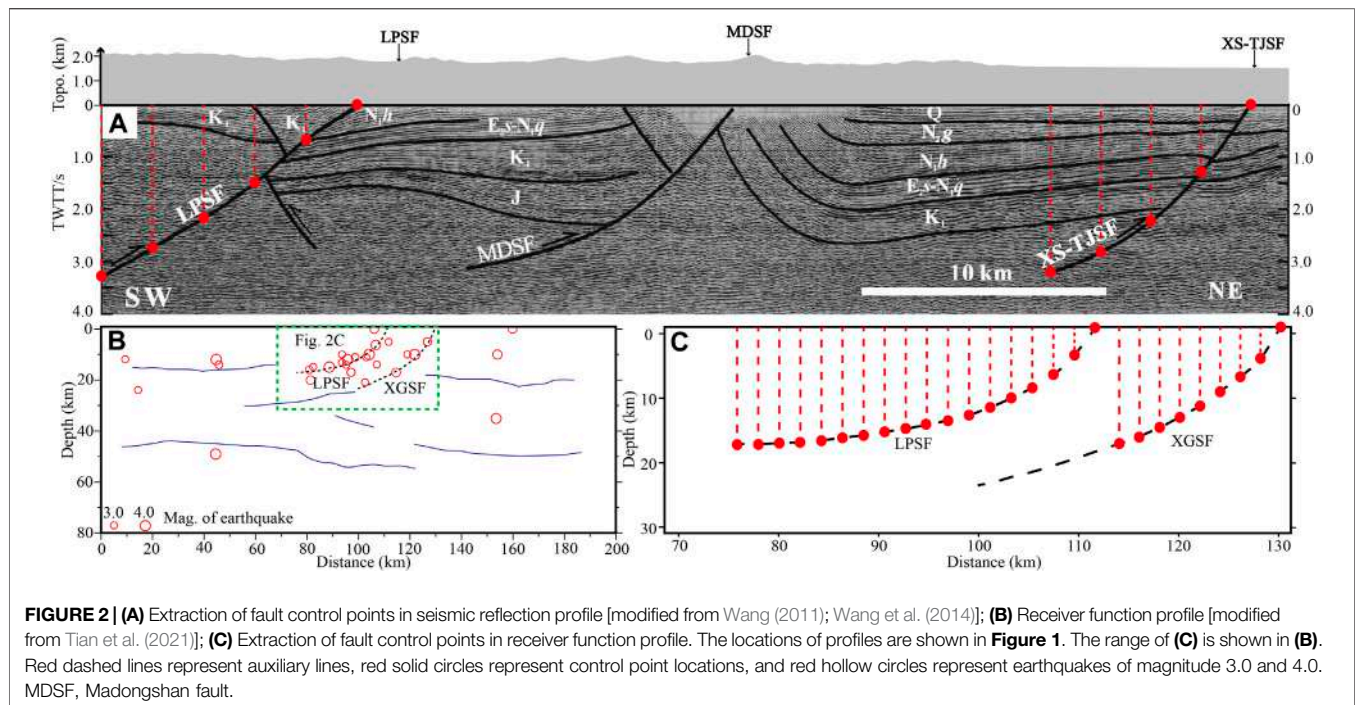


TJSF), the Yantongshan fault (YTSF), the Liupanshan fault (LPSF), and the Xiaoguanshan fault (XGSF) and the Qishan-Mazhao fault (QS-MZF) (**Figure 1**). Due to the special geographical location and strong Late Quaternary tectonic deformation in this area, it has always been a hotspot for the study of neotectonics and active tectonics (Zheng et al., 2016). Researchers have studied the geometric structure, motion properties, activity periods, and historical earthquakes of the above-mentioned faults to a considerable extent.

The WQLF is located between the HYF and the eastern Kunlun fault, which is an ultra-crustal fault that cuts through the Moho in the southern part of the study area, with a depth of about 50 km (Guo et al., 2016; Huang et al., 2017; Ye et al., 2015; Zhan et al., 2014). The fault extends from the vicinity of Baoji in the west of the Weihe Basin to Tongren, Qinghai Province in the west of Linxia, with a total length of about 420 km, a general strike of 280° – 310° , and dip angles of 50° – 85° (Shao et al., 2011). Since the Quaternary, it has been mainly dominated by left-slip with a relatively small dip-slip component. According to historical records, there have been eight historical strong earthquakes along the WQLF, including five earthquakes of magnitude 6 or higher, which are the 128

A.D. Gangu $M 6^{1/2}$ earthquake, the 143 A.D. Gangu $M 7^{1/4}$ earthquake, the 734 A.D. Tianshui $M 7^{1/2}$ earthquake, the 1765 A.D. Gangu-Wushan $M 6^{1/2}$ earthquake, and the 1936 A.D. Kangle $M 6^{3/4}$ earthquake (Lei et al., 2007; Yuan et al., 2007; Zhang, 2020b).

The MXSF zone is located in the southern mountainous area of Lanzhou and consists of four secondary faults, of which the northern margin fault of the Maxianshan is the main fault, with the largest scale and the strongest activity (Yuan et al., 2003; Yuan et al., 2013). The northern margin fault of the Maxianshan has a total length of about 115 km, a general strike of $N60^{\circ}W$, a tendency of SW, and a steep dip, which is a lower crustal-scale left-slip fault extending to about 40 km with a thrust component (Gao et al., 2013; Guo et al., 2016; Huang, 2017; Xin et al., 2021; Yuan et al., 2003; Yuan et al., 2013; Zhao, 2014). According to the textual research of historical seismic data, many earthquakes of moderate strength or above have occurred in this area, such as the 1125 A.D. Lanzhou $M 7$ earthquake, the 1590 A.D. Southeast Yongjing $M 6^{1/2}$ earthquake, the 1524 A.D. Maxianshan $M 5$ earthquake, the 1629 A.D. Lanzhou $M 5^{1/2}$ earthquake, the 1901 A.D. West Lanzhou $M 5$ earthquake, and others (Yuan et al., 2013).



The HYF starts from Xiaokou, Guyuan in the eastern Liupanshan piedmont, passes through Haiyuan, crosses the Yellow River, and extends northwest to the Xijishui left-stepped pull-apart basin in Jingtai, with a total length of about 240 km. The HYF is mainly left-slip motion with thrust component and dips to SSW with a large dip angle, of which the strike gradually changes from NNW to NWW from east to west and generally presents a slight arc protruding NE. The 1920 A.D. Haiyuan M 8^{1/2} earthquake occurred on this fault, forming a maximum left-lateral displacement of 10 m and a 230 km long surface rupture zone.

The XS-TJSF is also known as the Zhongwei-Tongxin fault with a total length of about 200 km, whose overall shape is similar to the HYF, presenting an arc slightly protruding NE in the middle. In this study, taking the compression stage area of two secondary faults intersecting at an angle of 17° at Shuangjingzi, Zhongwei as the boundary, the fault is divided into the northwest segment and the southeast segment (Li, 2014). The northwestern segment of the XS-TJSF strikes near E-W, dipping to S with steep dip angles, and is dominated by left-slip motion with either normal or thrust component, which also was the seismogenic fault of the 1709 A.D. Zhongwei M 7^{1/2} earthquake. The southeastern segment is a thrust fault dipping to SW with left-slip property, and the general strike is near NW (Min et al., 2001; Li, 2014; Li et al., 2016a).

The YTSF is the third arc-shaped fault protruding northeastward in the northeastern margin of the Tibetan Plateau. The fault is mainly dominated by thrusting motion with a total length of over 150 km, a tendency of NW-NNW, a dip direction of SW, and dip angles of 40°–70° (Li and Shen, 2001; Qin et al., 2017). The linear contour of the fault is not obvious, and there are no active profiles and various

geomorphological manifestations in the Holocene, indicating that the activity has been weak since the Late Quaternary (Min, 1998).

The LPSF is NNW-near SN as a whole with a length of about 175 km. The northernmost end of the fault is connected to the left-slip HYF near Xiaokou, Guyuan, and extends southward through Huating, Gansu Province to the vicinity of Longxian, Shaanxi Province, and is connected to the Longxian-Baoji fault zone. The LPSF is the main active fault in the Liupanshan area, dipping to SW and mainly dominated by thrust motion. The measured dip angles in the north are about 50°–55° and those in the south are 40°–60° (Shi, 2011b; Wang, 2018). The Liupanshan area has experienced the 600 A.D. Qinlong M 6^{3/4} earthquake, the 1219 A.D. Guyuan M 7 earthquake, the 1306 A.D. Guyuan M 7 earthquake, and the 1622 A.D. North Guyuan M 7 earthquake, which are mainly concentrated in the northern segment of the fault. There are few historical records of strong earthquakes in the middle and south segments, which are unbroken active faults with the potential risk of strong earthquakes (Du et al., 2018; Lei et al., 2007; Mei et al., 2012; Wang et al., 2018).

The XGSF is located on the eastern side of the Liupanshan, nearly parallel to the eastern Liupanshan piedmont fault. It is a W-dipping thrust fault with a length of about 60 km and dip angles of 40°–50°. The Cretaceous Liupanshan group thrusts onto the Tertiary red bed (Zhang, 1990).

Located on the southwestern edge of Ordos, the QS-MZF is the largest, longest, and most active fault in the Longxian-Baoji fault zone (Shi, 2011a; Li, 2017). The fault has a total length of about 180 km, a strike 300°–320°, and dip angles of 70°–85°. According to field surveys, the features of various geomorphological bodies along the QS-MZF trace show that the fault movement mode is mainly left-slip, with a normal fault motion component (Li, 2017).

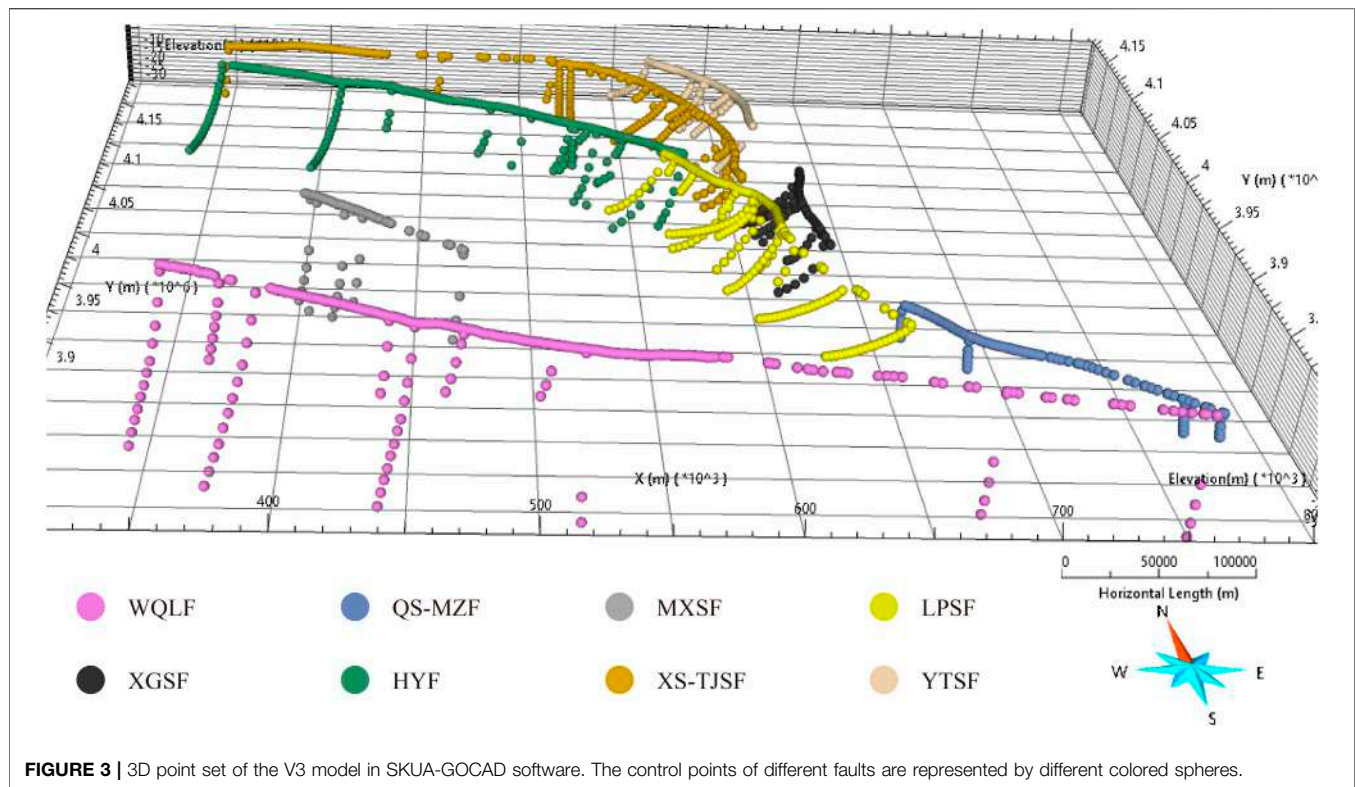


FIGURE 3 | 3D point set of the V3 model in SKUA-GOCAD software. The control points of different faults are represented by different colored spheres.

3 RESEARCH METHOD

3.1 Data Source and Filtering

The data used in this study can be mainly divided into two parts: 1) surface traces of active faults, and 2) deep structures of active faults. The surface traces are derived from the accurate and reliable geological mapping of active faults (Deng et al., 2002; Zheng et al., 2016, 2019). The deep structures of active faults come from published research of seismic reflection profiles, geological profiles, magnetotelluric profiles, tomographic imaging profiles, receiver function profiles, accurate location of small earthquakes, and focal mechanism solutions in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau. The geological mapping of active faults provides accurate fault surface traces for this study, which is not only the highest priority data for limiting fault surface locations but also the prerequisite for constraining the deep structural profile data of faults. With current technology, it is impossible to directly observe the deep structure of the earth. The indirect observation of the deep geometric structures of the faults through numerous 2D profiles published by researchers is the main way to constrain the deep structures of active faults in this study.

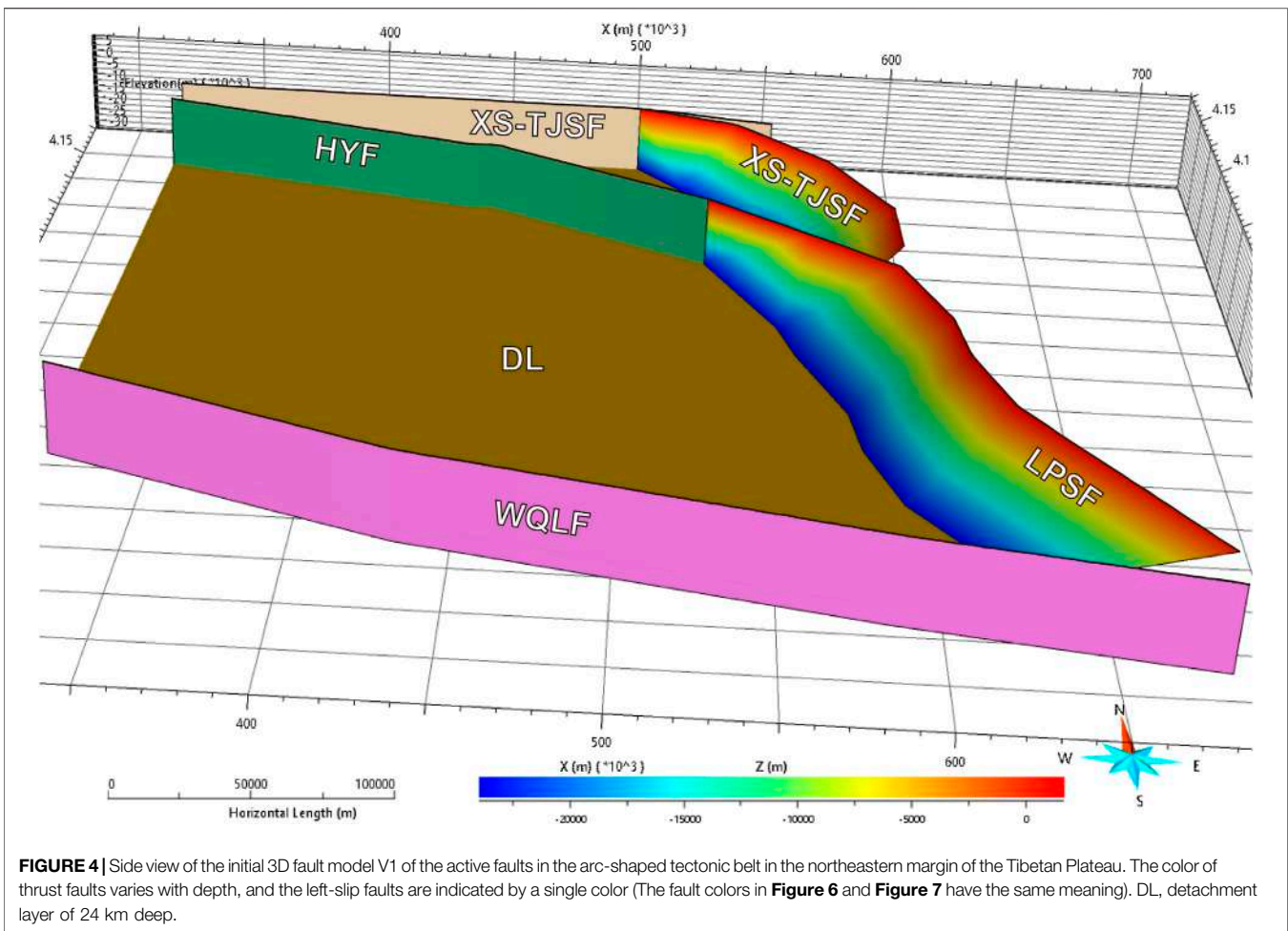
Before the establishment of the 3D geological models, this study fully collected a series of data revealing the deep geometric structures of active faults in the study area, mainly including digital image data of the profile locations and the 2D profiles. Digital image data involves multiple disciplines, technical

equipment updates, and different processing institutions, making the collected data have the properties of multiple sources, multiple resolutions, and multiple standards. Therefore, through the evaluation and weighting of multi-source data, this research takes the data closer to the true deep geometric structure of the fault as the main body, extracts control points at a density of about 2 km horizontally, and extracts control points of low-priority data at a horizontal interval of about 4 km, to obtain the 3D space coordinate sets corresponding to the faults and detachment layers.

3.2 3D Geological Modeling Process

3.2.1 Vectorization of Active Fault Surface Traces

According to the most accurate geological mapping of active faults, the surface traces of each active fault are outlined in the geological map. Considering that each active fault has segmental properties, and the current resolution of the 2D profiles cannot reach the accuracy of revealing the segmental structure of deep fault sections, the fault trace of each fault is fitted into a continuous trace distribution in this study (Figure 1). The geological map is registered through ArcGIS software and corrected to the CGCS2000 GK CM 105E projection coordinate system. Add control points according to the surface trace, add latitude and longitude coordinates for the control points and export the coordinate point set. Combined with DEM data, the elevation information is added to the control points of the corresponding surface trace to obtain complete 3D data.



3.2.2 Vectorization of Digital Images

Firstly, the profile locations are usually indicated in regional geological maps of the published literature, by which we can locate the specific distribution of each profile. Through ArcGIS software, register the corresponding regional geological map, project the longitude and latitude coordinate point set of active fault surface trace control points extracted in step (1), correct the profile location, and extract the profile distribution coordinates. Secondly, import 2D profile images into Adobe Illustrator software, add active fault control points with the corresponding horizontal point density according to the proportional scale in the figures and count the depth information (**Figure 2**). Then, the 2D profile images are registered with the profile distribution coordinates, and the latitude and longitude coordinates of the control points are extracted *via* ArcGIS software. Data processing in ArcGIS software is unified to CGCS2000 GK CM 105E projected coordinate system. Finally, based on the above data, 3D coordinate point set files of the control points of each corresponding fault and detachment layer are sorted out.

3.2.3 Build 3D Models

The extracted 3D data point set was imported into SKUA-GOCAD software (**Figure 3**). Then assist the function of the “Structure and Stratigraphy” module to draw strata and fault

planes (Li et al., 2016b). Due to the possible sparsity of the modeling data sources, the constructed 3D geological fault plane model is relatively rough. The SKUA-GOCAD software uses a combination of discrete smoothing and Kriging interpolation methods to interpolate the sparse and empty area of the data to smooth the fault plane model (Franke, 1982; Mallet, 1992).

3.3 Theory of Earthquake Magnitude Estimation

In the past decades, to analyze the potential seismic hazards, a series of formulas of the moment magnitude-rupture parameters have been proposed by researchers, among which the most widely used is the empirical formulas proposed by Wells and Coppersmith (1994) that the earthquake magnitude corresponds to the surface rupture length (SRL), subsurface rupture length (RLD), rupture area (RA), and average displacement (AD) respectively. Cheng et al. (2020) collected seismic data from 91 earthquakes in mainland China and adjacent areas and used general orthogonal regression to derive the magnitude-rupture scaling relations. Their results were close to the statistical results given by global data without considering the five major earthquakes in eastern

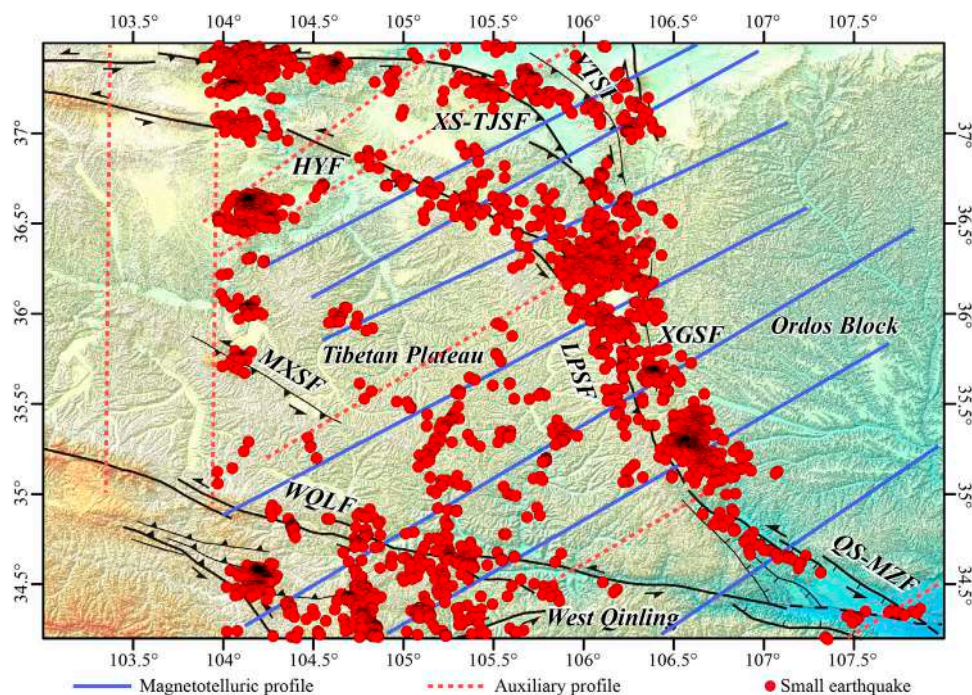


FIGURE 5 | Distribution of modeling profiles and small earthquakes in the V2 model. Small earthquakes occurred from 1 January 1990 to 30 June 2018, with magnitudes less than 6.7 (Zhan et al., 2020).

China before 1900 (Blaser et al., 2010; Hanks and Bakun, 2002; Wells and Coppersmith, 1994). Therefore, in this study, Eqs 1–5, which are more representative of the empirical seismic magnitude-rupture scale relationship in China, were used to estimate the seismic moment magnitudes (Cheng et al., 2020).

For strike-slip earthquakes

$$M_w = 3.77 + 0.98 \log(RA) \quad (M_w < 6.7) \quad (1)$$

$$M_w = 1.86 + 1.64 \log(RA) \quad (M_w \geq 6.7) \quad (2)$$

$$\log(RLD) = (-2.45 \pm 0.17) + (0.61 \pm 0.03)M_w \quad (3)$$

For thrust earthquakes

$$M_w = 4.41 + 0.83 \log(RA) \quad (4)$$

$$\log(RLD) = (-3.27 \pm 0.66) + (0.72 \pm 0.10)M_w \quad (5)$$

4 RESULT

4.1 3D Fault V1 Model

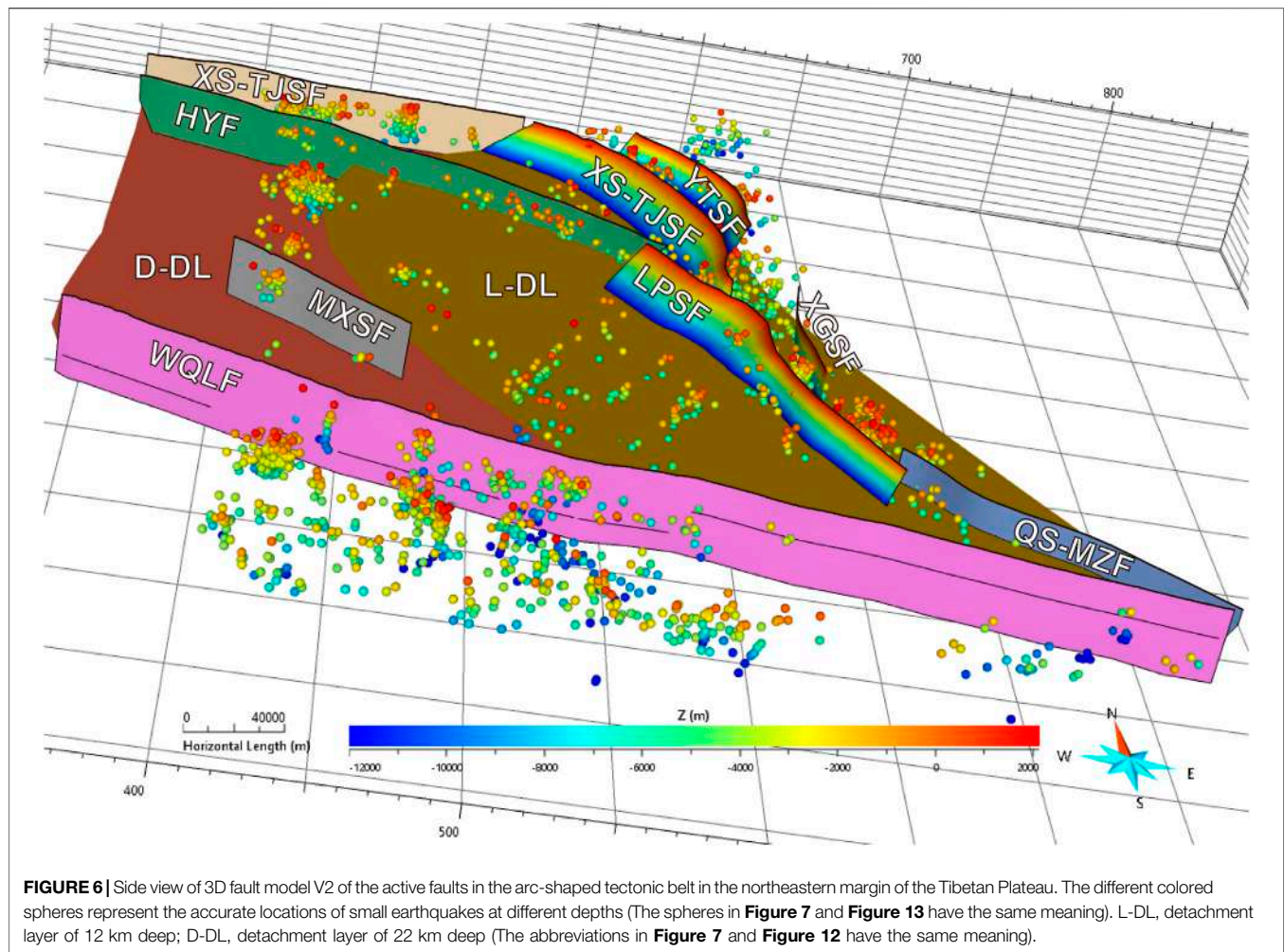
At the early stage of this study, the initial 3D fault model V1 of active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau was established (Figure 4, hereinafter referred to as the “V1 model”). The V1 model involves the deep structures of four large-scale active faults, namely, the HYF, the XS-TJSF, the LPSF, and the WQLF, and a set of detachment layers with continuously changing depths is also set up. The surface part of the initial V1 model is controlled by the main 3D control points extracted from the fault surface

trace. The deep structure setting of the model is as follows. For the HYF, the WQLF, and the northwestern segment of the XS-TJSF, which are left-slip in nature, the dip angle is uniformly set to 90°. For the LPSF and the southeastern segment of the XS-TJSF with a thrust nature, the dip angle variation along the strike of the faults is determined according to the actual geological surveys. In this model, it is considered that the detachment layer with a depth of about 24 km controls the development of all faults in this region.

In the initial V1 model, firstly, the density of control points extracted from the surface traces of each fault is insufficient, which cannot accurately reflect the surface strike change of each fault, and secondly, the deep structures of each fault are only constrained by the geological dip angles. Therefore, the constraints at all levels of the V1 model are not accurate enough, and only the structure of active faults in the region can be briefly displayed. The model is still far from reaching our modeling goal, but it verifies for us the feasibility of establishing the 3D fault model of major active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau.

4.2 3D Fault V2 Model

Given the shortcomings of the V1 model mentioned in Section 3.1 above, rational optimization of this model is the main direction of this research. The most effective way to optimize the fault surface trace is to encrypt the control points, to improve the accuracy of the fault surface trace. Therefore, we encrypt and extract the latitude and longitude coordinates of the control points at the turning points and curvature changes of the trace, then project them onto DEM to obtain the complete 3D



spatial coordinates with depth information. For the deep structure information of each fault, this study combines published geological, geophysical, and seismic data to interpret the deep structures of main active faults and detachment layers in 7 magnetotelluric profiles and 7 auxiliary profiles (**Figure 5**) in the study area. Then extract 3D spatial coordinates of the deep structures, and obtain the accurate location of small earthquakes in the study area simultaneously (Zhan et al., 2020). Finally, with the aid of the SKUA-GOCAD software modeling function, the 3D fault model V2 of the active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau was obtained (**Figure 6**, hereinafter referred to as the “V2 model”).

The modeling scope of the V2 model has been extended to more active faults in the study area, such as the MXSF, the YTSF, the XGSF, and the QS-MZF, which are relatively small-scale faults. According to the published research, two sets of detachment layers with depths of 12 and 24 km were installed in the V2 model to replace the detachment layer with a depth of 24 km in the V1 model (Du et al., 2018; Tian et al., 2021; Xia, 2019). In addition, the accurate locations of small earthquakes are added to the model to visualize the relationship between the deep geometry of active faults and the distribution of small earthquakes, which provides model support for

studying the relationship between the fault structure and earthquakes and evaluating the regional seismic risk.

4.3 3D Fault V3 Model

The optimization direction of the third model is mainly about the deep structure of the relevant active faults. To achieve this purpose, we searched a large number of published literature and obtained a series of 2D profiles passing through the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau. The surface traces, accurate location of small earthquakes, and detachment layers distribution of the third model follow the extracted information of the V2 model. For the deep structures of the faults, this study adopts 7 seismic reflection profiles, 3 geological profiles, 22 magnetotelluric profiles, 9 accurate location profiles of small earthquakes, 4 tomographic imaging profiles, and 1 receiver function profile from many collected profiles (**Figure 7**, **Table 1**), aiming to use multidisciplinary data to finely characterize the 3D subsurface structures of active faults. Based on the above data, we built the 3D fault model V3 of the active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau (**Figure 8**, hereinafter referred to as the “V3 model”).

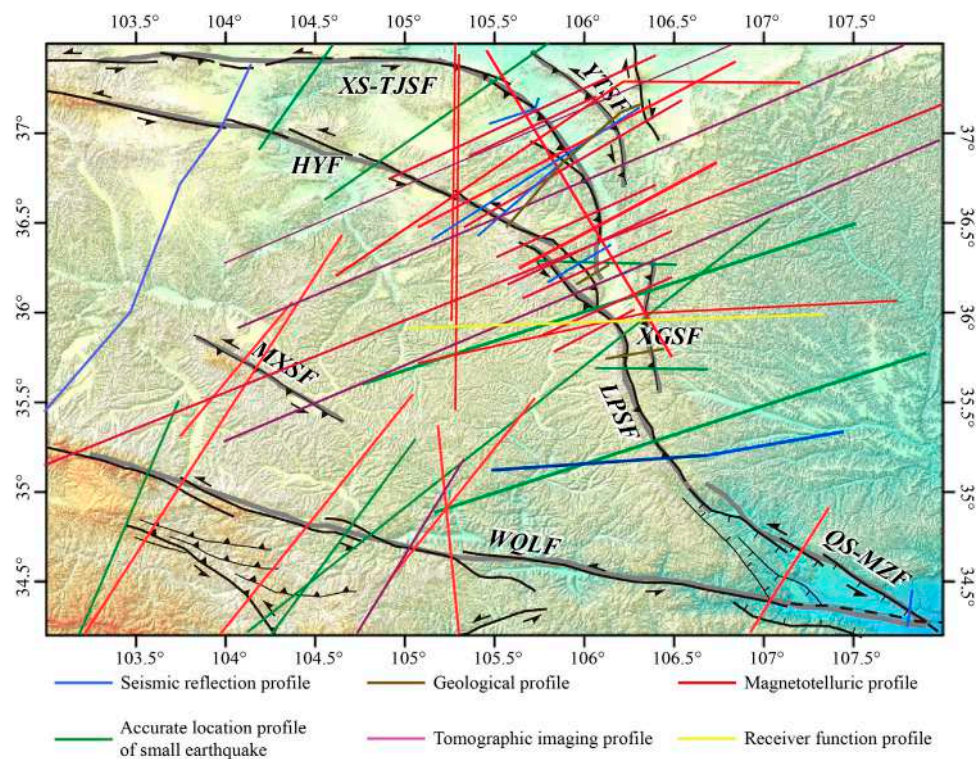


FIGURE 7 | Distribution of modeling profiles in the V3 model. The modeling profiles corresponding to each fault, profile types, and references are shown in the Table 1.

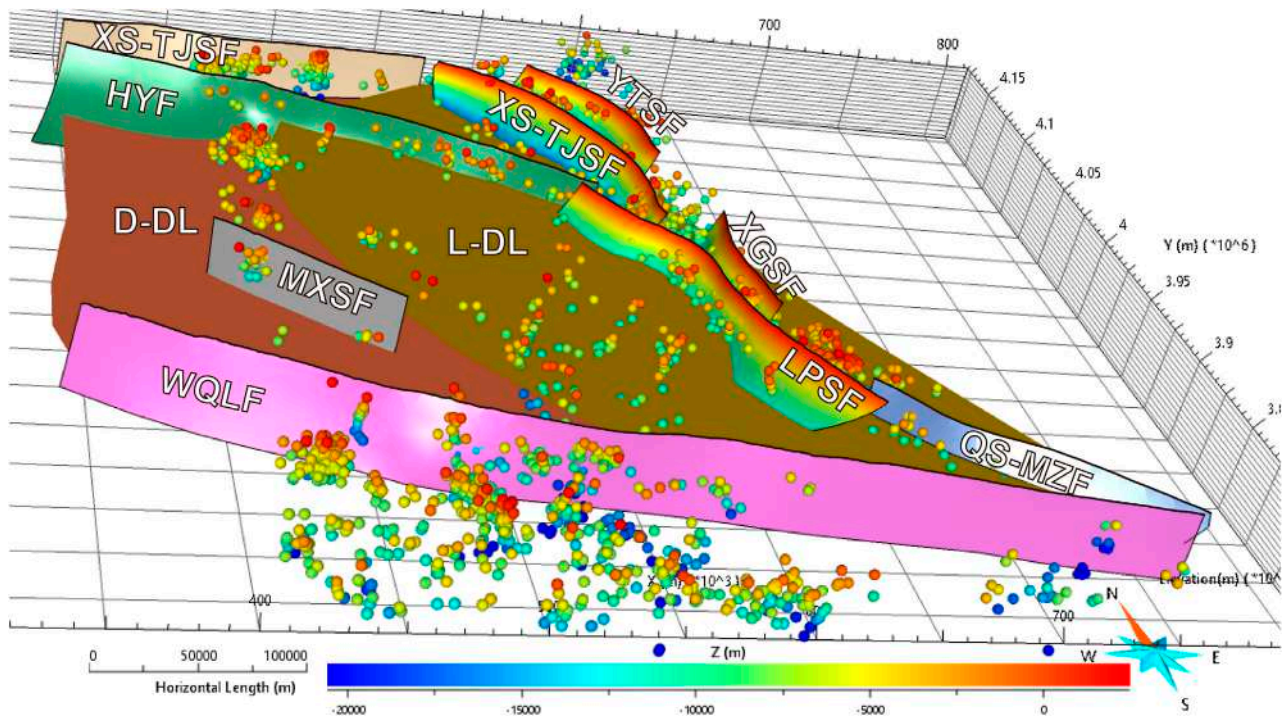


FIGURE 8 | Side view of 3D fault model V3 of the active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau.

TABLE 1 | Modeling profiles in the V3 model.

Fault	Fault modeling depth/km	Type of profiles	References
WQLF	30	MP ALP TIP	Zhan et al. (2014); Zhao, (2014); Xin et al. (2021) Zhang, (2020a) Zhang et al. (2021)
MXSF	30	MP	Zhan, (2008); Zhao, (2014); Xin et al. (2021)
HYF	30	SRP GP MP ALP TIP	Wang, (2011); Guo et al. (2016); Zeng et al. (2020) Wang et al. (2011); Shi et al. (2015) Zhan et al. (2004); Zhan, (2008); Min, (2012); Xia, (2019) Xin et al. (2020) Zhan et al. (2020)
XS-TJSF	15–22	SRP GP MP ALP	Li and Lai, (2006); Wang, (2011); Zeng et al. (2020) Shi et al. (2015) Zhan, (2008); Min, (2012); Xia, (2019) Zhan et al. (2020)
YTSTF	15	SRP GP MP TIP	Li and Lai, (2006); Zeng et al. (2020) Shi et al. (2015) Zhan, (2008); Min, (2012); Xia, (2019) Xin et al. (2020)
LPSF	12–22	SRP GP MP ALP TIP RFP	Zhan, (2008); Wang, (2011); Guo et al. (2015); Zeng et al. (2020) Zhang et al. (1991) Xia, (2019) Xin et al. (2010); Du et al. (2018); Zhan et al. (2020) Sun et al. (2021) Tian et al. (2021)
XGSF	15	GP MP ALP TIP RFP	Zhang et al. (1991) Xia, (2019); Zeng et al. (2020) Xin et al. (2010); Du et al. (2018) Sun et al. (2021) Tian et al. (2021)
QS-MZF	12	SRP	Li, (2017)

SRP, Seismic reflection profile; GP, Geological profile; MP, Magnetotelluric profile; ALP, Accurate location profile of small earthquake; TIP, Tomographic imaging profile; RFP, Receiver function profile.

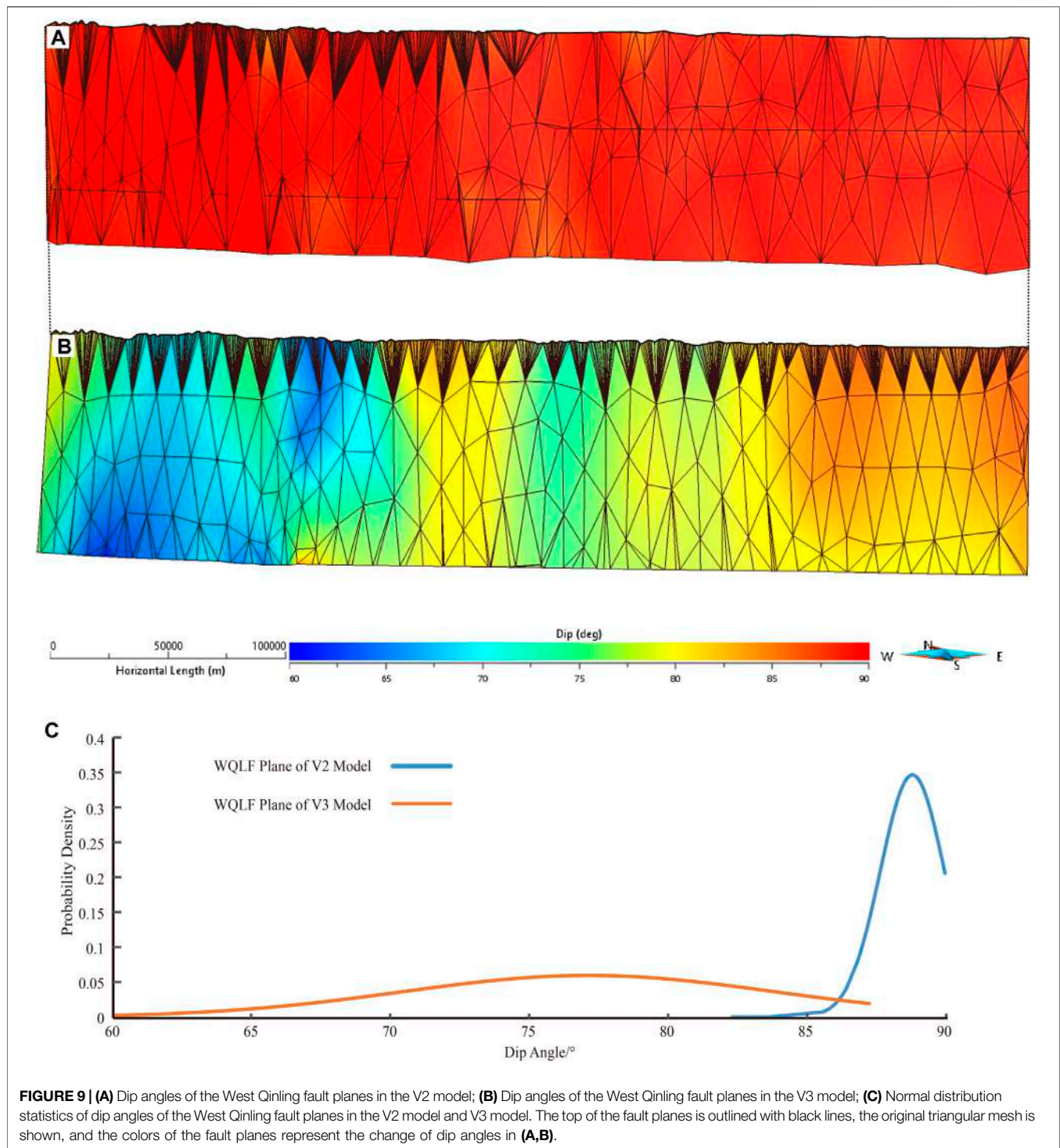
4.4 Model Optimization and Comparison

Through three versions of the models established successively, it is not difficult to see that our results have been gradually transformed from simple modeling to a 3D fine model under multiple constraints. First, from the view of the whole model, the major active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau have been all involved in the modeling process from the V1 model to the V3 model, whether it is a large-scale active block boundary fault across the study area or a small-scale fault that is only under the control of local neotectonic movement. Secondly, starting from the V2 model, referring to the accurate geological mapping of active faults, the surface traces of main active faults in the model are constrained by more 3D control points along the fault strike. Meanwhile, the accurate locations of small earthquakes are considered, and the relative positions are visualized in the model. Finally, the most important deep geometric structure, their constraints are gradually strengthened. The V1 model only depends on the geological profiles, the dip angles of left-slip faults are set to 90° uniformly, and thrust faults refer to the surface dip angles to infer the deep extension of faults. The V2 model mainly

refers to a series of magnetotelluric profiles and auxiliary profiles to reveal the deep geometry of active faults. The deep structure of each fault in the V3 model is constrained by multi-source 2D profiles. In general, our current V3 model has meticulously depicted the distribution along the strike and extension along the longitudinal direction of the main active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau from surface to subsurface.

During our model optimization process, the interpolated formed fault planes have also changed dramatically with more detailed changes as the constraint data increased. Given the fact that in the V1 model only the dip angles from the geological field survey can hardly outline the deep geometric structures of each fault, the following fault plane comparisons only involve the V2 model and the V3 model.

The WQLF with left-slip nature is a deep boundary fault that divides the Tibetan Plateau block from the West Qinling block in the southern part of the study area. The fault cuts through the Moho and extends much deeper than our modeling scale. The top of the WQLF plane constructed by fault surface traces, geological profiles, and DEM data shows the elevation variation of the fault



surface trace along the strike. The WQLF planes of the V2 model and V3 model are established from different deep geometric information and the plane of the V3 model has more abundant changes in dip angles by contrast (**Figures 9A,B**). By analyzing and comparing the normal distribution of dip angles on the two fault planes, the dip angles are mainly distributed in the high-angle area (**Figure 9C**). For the dip angles of the WQLF plane in

the V2 model, the average value is 88.8° , the standard deviation is 1.15° , and 99.7% of the dip angles are distributed in the range of 85.4° – 90.0° . It shows that the distribution of dip angles is concentrated the nearly vertical. For the dip angles of the WQLF plane in the V3 model, the average value is 77.4° , the standard deviation is 6.99° , and 99.7% of the dip angles are distributed in the range of 60.0° – 87.3° . Although the dip angles

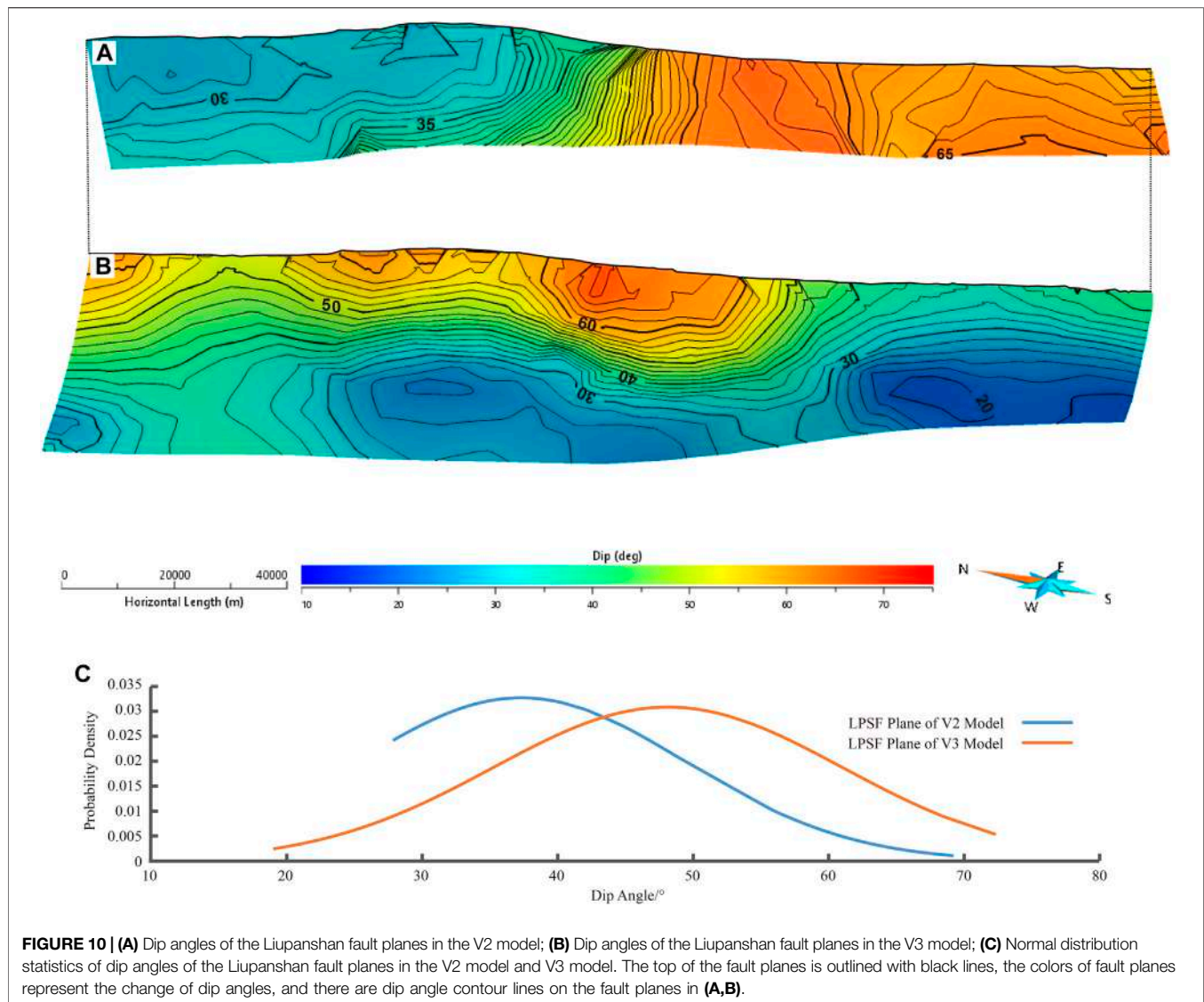


FIGURE 10 | (A) Dip angles of the Liupanshan fault planes in the V2 model; **(B)** Dip angles of the Liupanshan fault planes in the V3 model; **(C)** Normal distribution statistics of dip angles of the Liupanshan fault planes in the V2 model and V3 model. The top of the fault planes is outlined with black lines, the colors of fault planes represent the change of dip angles, and there are dip angle contour lines on the fault planes in **(A,B)**.

of the WQLF plane in the V2 model have a more concentrated distribution, the dip angles of the WQLF plane in the V3 model constrained by multi-source data are more in line with the geological surveys (Shao et al., 2011).

The LPSF with a thrust nature is the boundary fault, which divides the Tibetan Plateau block from the Ordos stable block in the east of the study area. Similar to the analysis of the WQLF plane above, the LPSF surface traces of the V2 model and V3 model show consistent elevation relief, and the shape of fault planes is significantly different (**Figures 10A,B**). The normal distribution of dip angles shows that the distribution center of dip angles in the V3 model is about 16° higher than that of the V2 model, but the distribution widths of dip angles are extremely similar (**Figure 10C**). For the dip angles of the LPSF plane in the V2 model, the average value is 37.3°, the standard deviation is 12.17°, and 99.7% of the fault dip angles are distributed in the range of 27.9°–70.0°. For the dip angles of the LPSF plane in the V3 model, the average value is 53.6°, the standard deviation is

12.05°, and 99.7% of the fault dip angles are distributed in the range of 19.1°–72.3°.

In the V2 model, the near-surface dip angles of northern LPSF are about 40°, and those of southern LPSF are about 60°. The dip angles of the LPSF gradually increase southward along the strike. Meanwhile, as the LPSF extends deeper, the dip angles decrease, and the LPSF plane gradually merges into the detachment layer with a depth of about 12 km (**Figure 10A**). However, there is a completely different situation in the V3 model. The dip angles of LPSF gradually decrease southward along the strike. The near-surface dip angles of northern LPSF are about 60° and the northern LPSF plane gradually merges into the detachment layer with a depth of about 22 km. The near-surface dip angles of southern LPSF are about 40°, and the southern LPSF plane gradually merges into the detachment layer at a depth of about 12 km. According to geological investigations (Shi, 2011b; Wang, 2018), the shallow dip angles in the V3 model are more in line with geological facts.

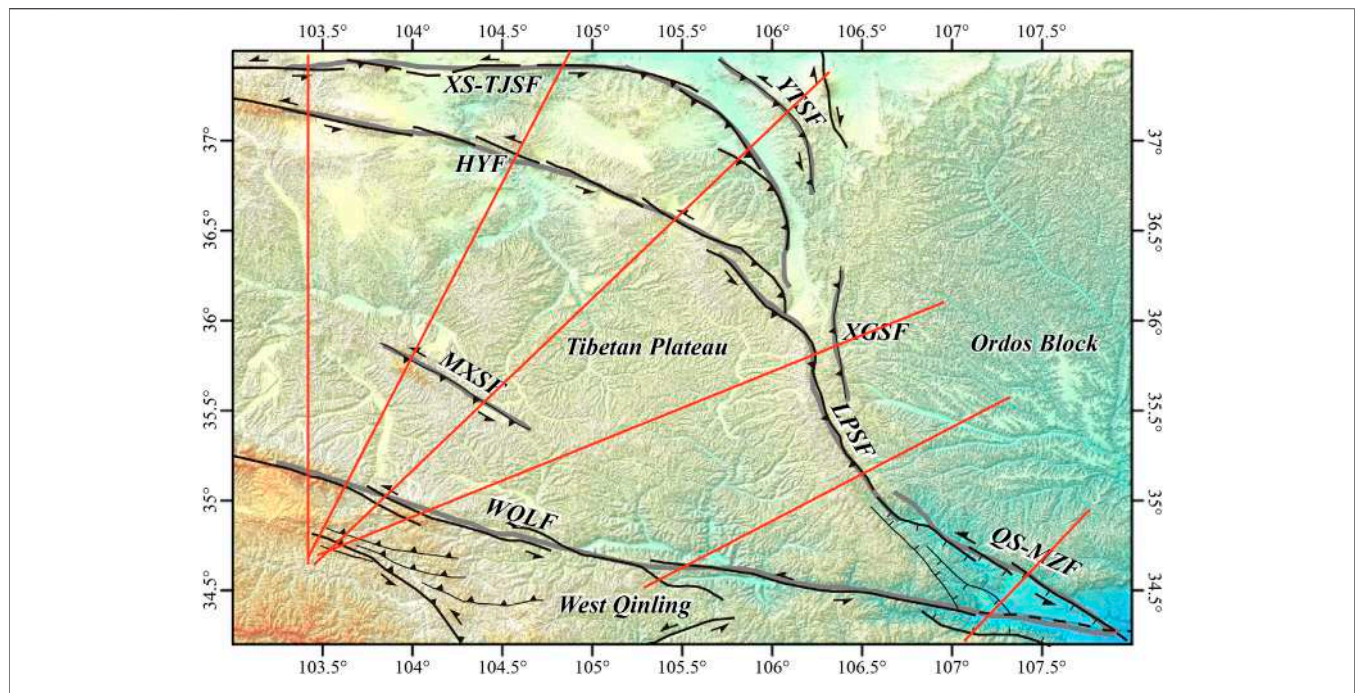


FIGURE 11 | Figure cut profile distribution of the V3 model. Red solid lines represent the locations of profiles.

4.5 Modeling Fault Characteristics

Through the establishment from the V1 model to the V3 model, we have the following insights into the main active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau.

- (1) The WOLF is a large left-slip fault cutting the Moho, which has a depth of about 50 km, a total length of about 443.94 km, an overall strike of 278° – 287° , a tendency of SSW, and the overall dip angles of 60° – 90° . Along the strike, the near-surface dip angles show the characteristics of low in the west and high in the east. Vertically, the dip angles do not change much and the fault plane extends deeper maintaining the original trend.
- (2) The MXSF is at least a lower crustal-scale left-slip fault, extending to about 40 km deep near the Moho, which has a total length of about 91.26 km, a strike of 300° , and a tendency of SW. The dip angles are distributed between 73° and 87° with a decreasing trend towards deeper, and the overall shape of the fault plane is steep.
- (3) The HYF is a crustal-scale left-slip fault, with a depth of about 30 km and a total length of about 248.36 km. The surface trace is slightly convex to NE as a whole and the strike gradually changes from 297° in the easternmost end to 280° in the westernmost end, dipping to SW. The dip angles of the HYF are generally distributed at 30° – 83° . From west to east, the near-surface dip angles increase from 60° to about 80° , and the fault plane becomes steeper. The dip angles of the HYF gradually decrease towards the deeper part, from 60° to 30° in the western segment of the HYF, and from 80° to 50° in the middle-eastern segment. It is suggested that the HYF has been influenced by the thrust component, forming a gentle west-east steep, nearly shovel-shaped fault plane.
- (4) The LPSF is a thrust fault with a total length of about 194.25 km, a tendency of SW, and an overall strike of 310° – 328° . However, the fault strike changes near Guyuan, and the surface trace shows a slightly NE convex arc. About north of 35.5° N, the near-surface dip angles of LPSF range from 50° to 70° . With increasing depth, the LPSF finally merges into the deep detachment layer with a depth of about 22 km, and the dip angles gradually decrease to about 30° . About south of 35.5° N, the near-surface dip angles of LPSF are about 40° . With the extension deeper, the LPSF finally merges into the shallow detachment layer with a depth of about 12 km, and the dip angles gradually decrease to about 20° .
- (5) The QS-MZF is an upper crustal-scale left-slip fault, approximately 145.88 km in length, with a strike of 300° , a tendency of SW, and overall dip angles of 60° – 70° . The fault plane extends down to the shallow detachment layer with a depth of about 12 km.
- (6) In this study, the XS-TJSF is divided into two segments, the northwestern and the southeastern segments. The northwestern segment is an upper crustal-scale left-slip fault, with a total length of about 170.30 km, a strike of E-W, a tendency of S, and dip angles of 70° – 80° . The extended depth is mainly about 22 km, but it gradually decreases to about 18 km towards the easternmost end. The southeastern segment is an SW dipping thrust fault, with a total length of about 171.72 km. The surface trace is an arc of NE convexity, with the northernmost trending 298° and gradually deflecting clockwise to the southernmost trending near S-N. The dip

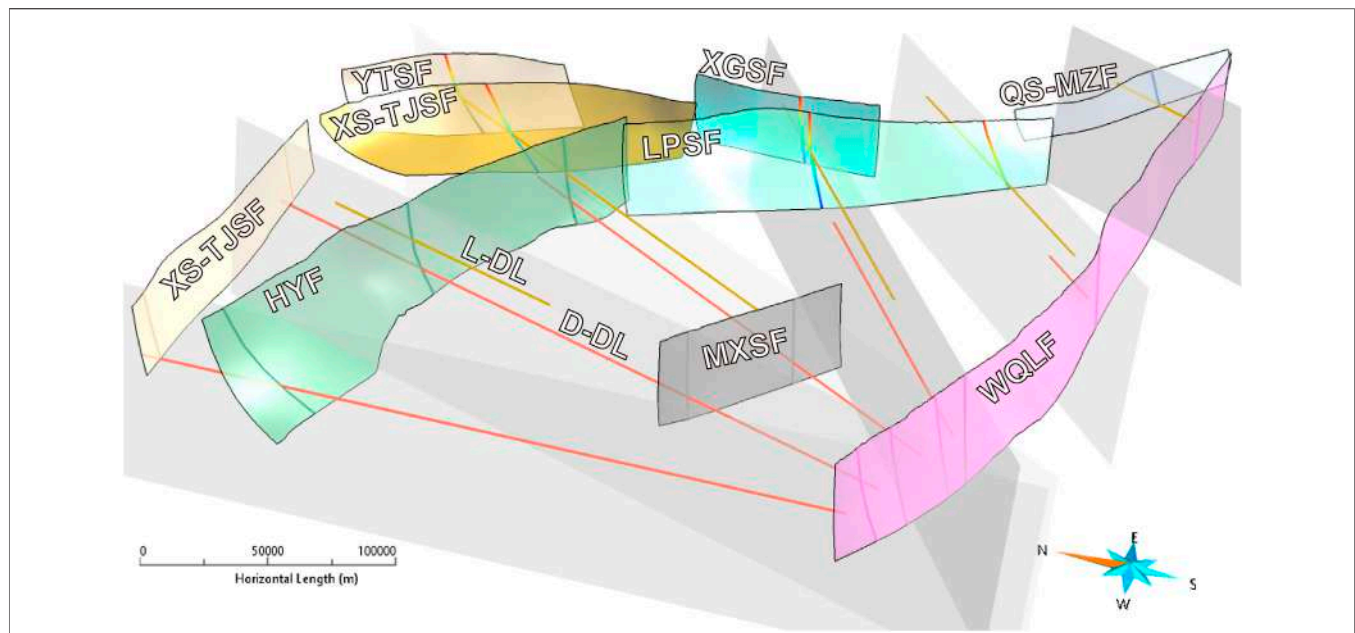


FIGURE 12 | Cut into slices of the V3 model. The fault planes are transparent displays and their contours are outlined by black lines.

angles of the near-surface fault are about 40° as a whole, but the southern tail of the southeastern segment has larger dip angles of about 60° . Extending to the deeper part, the dip angles eventually decrease to 20° – 30° and the fault plane terminates at a depth of about 15 km.

- (7) The YTSF is an SW dipping thrust fault, with a total length of about 101.57 km. The surface trace is an arc of NE convexity, with a strike of 305° at the northernmost end and gradually deflecting clockwise to the southernmost end near S-N. The northern segment of the YTSF has near-surface dip angles of 50° – 60° , while the dip angles decrease southward and finally are up to 40° – 50° . With increasing depth, the YTSF plane slows down, extending to about 15 km deep, and the dip angles decrease to about 30° .
- (8) The XGSF is a W dipping thrust fault, with a total length of about 81.55 km and a strike of near S-N. The dip angles are about 60° near-surface, which gradually decrease to about 30° as the fault extends deeper, and the fault plane ultimately terminates at the depth of about 15 km.

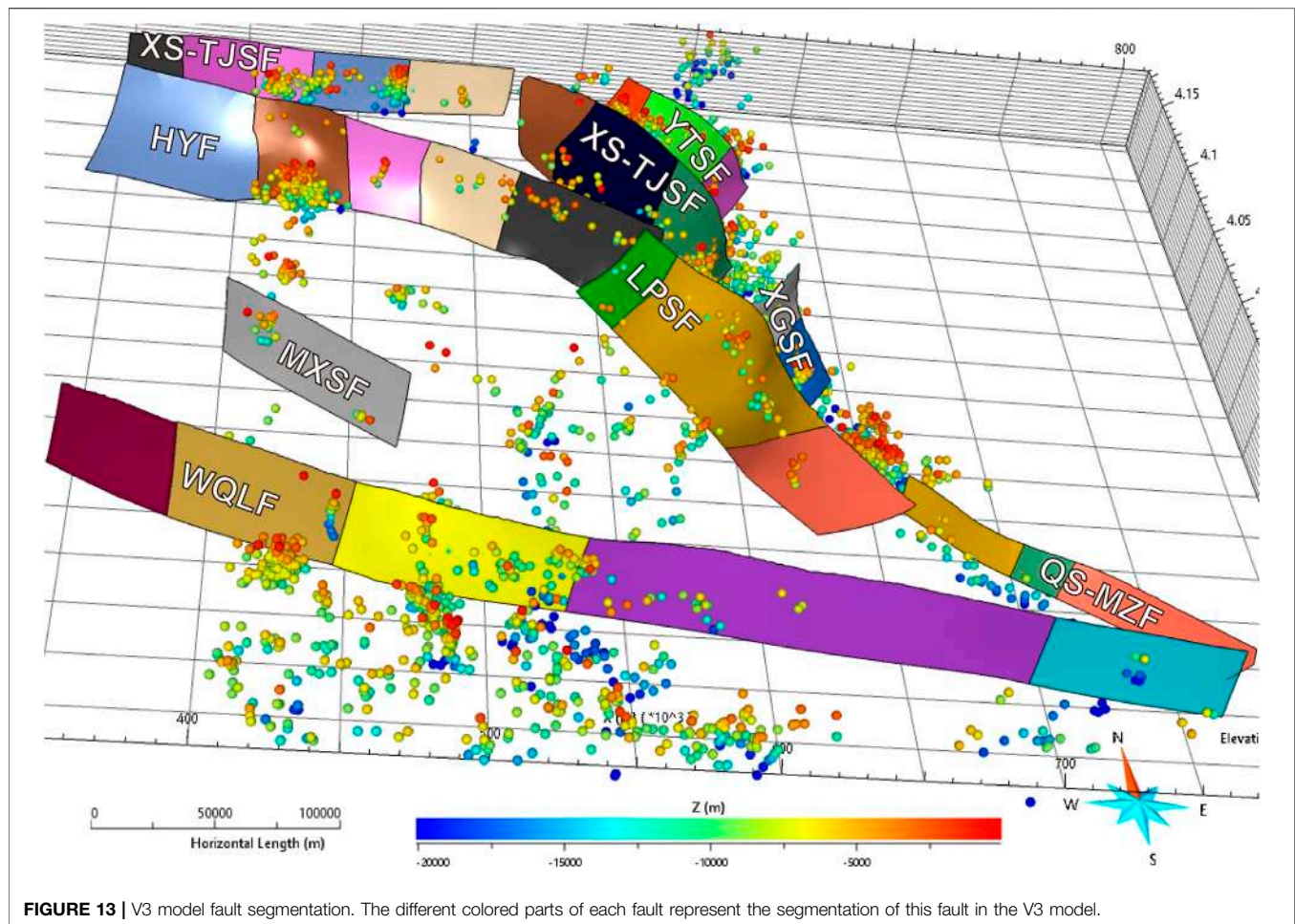
We suggest that the major active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau are largely controlled by two sets of detachment layers at depths of about 12 and 22 km. Except for the WQLF in the scale of super crust and the HYF and the MXSF in the scale of at least lower crust, thrust faults terminate between 12 and 22 km in depth, indicating that they are mainly controlled by the above two sets of detachment layers. And the QS-MZF and the northwestern segment of the XS-TJSF with left-slip nature extend to about 12 and 22 km, respectively. In this paper, our research is limited to the HYF (narrow sense), through which most of the 2D profiles show that it has not yet cut through the Moho. However, some

researchers have suggested that the western segment of the HYF (broad sense) may be a deep fault cutting through the Moho (Shi et al., 2017; Zhan et al., 2008; Zhao et al., 2019; Zhao et al., 2020). The LPSF continues southeastward along the tail of the HYF, of which the northwestern segment has a similar extended depth to the tail of the HYF and surface dip angles of about 60° , while the southeastern segment has an extended depth of only about 12 km and gentler dip angles of about 40° . The change of the deep structure of the LPSF from northwest to southeast probably reflects the fact that the strike-slip component of the near E-W trending HYF is absorbed by the near NW-SE trending LPSF, and the nature changes from predominantly left-slip to predominantly thrust (Zhang et al., 1991; Zheng et al., 2013, 2016).

5 DISCUSSION

5.1 3D Fault Model Uncertainty

To a certain extent, the V3 model has visualized the geometric structure of main active faults in the study area from the overall model. However, there are still a series of uncertainties in the V3 model, which affect the accuracy of the V3 model and determine whether it has practical application value. First of all, the multi-source 2D profiles, which are the basis for establishing the V3 model, have different geological interpretations for the same fault, and the resolution of profiles obtained by different methods is different. Through unified evaluation and screening of these data in the early stage, we can slightly reduce the impact of these uncertainties. Secondly, as the ordinate of some seismic profiles is the two-way travel time, we uniformly adopt the empirical seismic wave velocity of 5 km/s to make the time-depth



conversion, instead of using the actual seismic wave velocity. This approach will inevitably distort the converted fault plane in the model and cannot make it closest to the real structure of the fault plane. Finally, during the modeling process, we found that the density of control profiles in different parts of the study area is different, such as 2D profiles through the convergence area of the LPSF, HYF, and southeastern segment of the XS-TJSF are extremely dense. In contrast, the control profiles through the northwestern segment of the XS-TJSF and the western HYF are relatively sparse. As a result, the control points in different areas of the V3 model are unevenly distributed, and different parts of the fault plane constructed by interpolation also have different accuracy.

5.2 3D Fault V3 Model Application

Through the visual V3 model, we can easily obtain the cutting profiles of the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau in any direction, which reflects the deep geometry of regional active faults (Figures 11, 12). Therefore, the V3 model can provide pre-study results for subsequent researchers who plan to carry out relevant geological research in this area, such as collecting long-distance 2D profiles in a

certain direction, helping them to arrange and adjust subsequent fieldwork.

In addition, the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau is the front of the northeastward expansion of the plateau with active tectonic movements, developing a series of active faults. The evolutionary history of this region since the late Cenozoic is of great significance for studying the expansion of the Tibetan Plateau. At present, the geological research of this area is only limited to the regional surface geological evolution and the deep structures of several active faults (Lei et al., 2016; Shi et al., 2013; Wang et al., 2013; Wang et al., 2014; Zheng et al., 2013), and have not combined the deep structures of the overall active fault and surface geology to jointly explain the regional geological evolution history. The 3D fault model may serve as a bridge to combine the two, effectively linking regional surface geology, deep structures, most of the active faults, and so on. However, if we want to study the small-scale geological structures and the finer structure of a single fault in this area, our model is far from reaching the corresponding resolution and accuracy. The actual distribution of active faults in this region is far more complex than the V3 model. To reveal the deep segmental and geometric features of active faults more

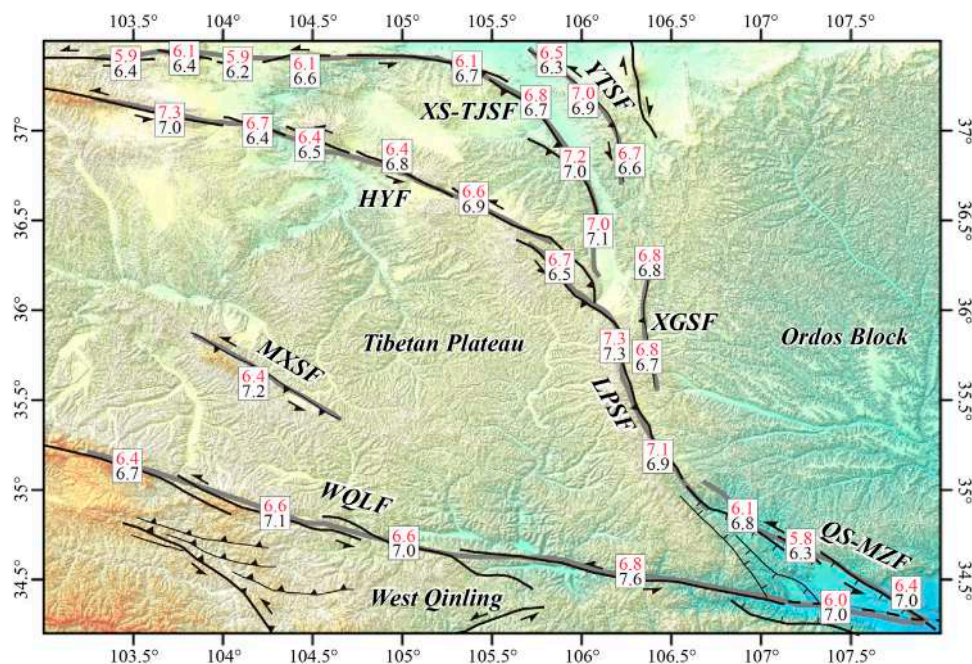


FIGURE 14 | Estimated moment magnitude of major active faults in the northeastern margin of the Tibetan Plateau. Red numbers represent results from the rupture area-moment magnitude empirical formula, and black numbers represent results from the rupture length-moment magnitude empirical formula.

precisely in the 3D fault model depends on the improvement of the regional 2D profiles coverage density.

5.3 Moment Magnitude Estimation of Main Active Faults

In the modeling process of this study, considering that the resolution and density of the modeling data cannot identify the segmental properties of active faults, the discontinuous surface traces of the faults are fitted into continuous traces. However, in the earthquake occurrence, not the entire fault is active at the same time, but only the energy release of the partial fault. Given this, it is necessary to clarify the segmentation of faults before quantifying the potential seismic hazard (Ding, 1993; Ding, 1995). For strike-slip faults, this study segments the fault planes according to the segmentation of fault surface traces (Howarth et al., 2021; Wesnauky, 1988; Wesnauky, 2006). For thrust faults, the dip angle contour of the fault planes was generated based on SKUA-GOCAD software, and the changing interfaces of dip angles are used to segment the fault planes (Qiu et al., 2016) (Figure 13). The accurate locations of small earthquakes indicate that there are still a considerable number of small-earthquake distributions within the region at least 30 km deep (Du et al., 2018; Sheng, 2015; Xin et al., 2010; Zhan et al., 2020; Zhang B., 2020), whereas our model sets a uniform cut-off of the fault plane extension depth to 30 km, for which the statistical fault plane area can be used to quantify the potential seismic moment magnitude.

Assuming that the planes and surface traces of the seismogenic faults all rupture during an earthquake, we can calculate the

potential seismic moment magnitude of each fault segment according to the empirical formulas proposed by Cheng et al. (2020) (Figure 14). If the major active faults in the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau produce earthquakes, the estimated moment magnitudes are generally higher than magnitude 6.0.

- (1) The WQLF can be divided into five segments. From W to E, the moment magnitudes of each segment are 6.4, 6.6, 6.6, 6.8, and 6.0 deduced from the RAF (RAF = rupture area-moment magnitude empirical formula), and 6.7, 7.1, 7.0, 7.6, and 7.0 according to the RLF (RLF = rupture length-moment magnitude empirical formula).
- (2) The MXSF is unsegmented as a whole. The moment magnitude is inferred to be 6.4 by the RAF, and 7.2 inferred by the RLF.
- (3) The HYF can be divided into five segments. From SWW to SE, the moment magnitudes of each segment are 7.3, 6.7, 6.4, 6.4, and 6.6 deduced from the RAF, and 7.0, 6.4, 6.5, 6.8, and 6.9 according to the RLF.
- (4) The LPSF can be divided into three segments. From NNW to SSE, the moment magnitudes of each segment are 6.7, 7.3, and 7.1 deduced from the RAF, and 6.5, 7.3, and 6.9 according to the RLF.
- (5) The QS-MZF can be divided into three segments. From NW to SE, the moment magnitudes of each segment are 6.1, 5.8, and 6.4 deduced from the RAF, and 6.8, 6.3, and 7.0 according to the RLF.
- (6) The northwestern segment of the XS-TJSF can be divided into five segments. From W to E, the moment magnitudes of

- each segment are 5.9, 6.1, 5.9, 6.1, and 6.1 deduced from the RAF, and 6.4, 6.4, 6.2, 6.6, and 6.7 according to the RLF.
- (7) The southeastern segment of the XS-TJSF can be divided into three segments. From NW to SSE, the moment magnitudes of each segment are 6.8, 7.2, and 7.0 deduced from the RAF, and 6.7, 7.0, and 7.1 according to the RLF.
 - (8) The YTSF can be divided into three segments. From NW to SSE, the moment magnitudes of each segment are 6.5, 7.0, and 6.7 deduced from the RAF, and 6.3, 6.9, and 6.6 according to the RLF.
 - (9) The XGSF can be divided into two segments. From NW to SSE, the moment magnitudes of each segment are 6.8 and 6.8 deduced from the RAF, and 6.8 and 6.7 according to the RLF.

The moment magnitudes respectively estimated from the rupture area and rupture length of thrust faults are in good agreement, with the difference basically within ± 0.3 , while for left-slip faults the moment magnitudes calculated from the rupture length are mostly larger than those calculated from the rupture area, with significant deviation. In combination with the deep structures of the main active faults shown in the V3 model, the thrust faults are relatively shallow in extended depth. The potential seismic hazards of thrust faults in this region are greater than those of left-slip faults with high dip angles. Therefore, we pay more attention to the possible seismic hazards caused by thrust faults in this region.

For thrust faults, the estimated moment magnitudes are all less than or equal to 7.3. Except for the southernmost part of the southeastern segment of the XS-TJSF, all moment magnitudes estimated from the RAF are greater than or equal to those estimated from the RLF. These thrust faults, which have experienced many historical strong earthquakes (Figure 1), have the potential for large earthquakes of magnitude 7 or greater. Therefore, the estimations based on the RAF are closer to reality and more reliable. Among thrust faults, the LPSF and the southeastern segment of the XS-TJSF, where the maximum seismic moment magnitude can be respectively up to 7.3 and 7.2. Moreover, the historical records of strong earthquakes in the middle and south segments of the LPSF are few and the passing time is long, so there is a risk of strong earthquakes (Du et al., 2018; Lei et al., 2007; Mei et al., 2012; Wang et al., 2018).

At the same time, the seismic hazards caused by left-slip faults, which also have the potential for a strong earthquake of magnitude 7 or more, should not be ignored. For the WQLF, the QS-MZF, the MXSF, and the northwestern segment of the XS-TJSF, all moment magnitudes estimated from the RLF are greater than those estimated from the RAF. The eastern segment of the HYF is consistent with this phenomenon, but the western segment is opposite. We suggest that this may be because the western segment of the HYF, which is constrained by seismic reflection profile, has smaller dip angles of about 60° compared to other left-slip faults (Guo et al., 2016). Therefore, the estimations based on the RLF are more reliable for the left-slip faults with steep planes, and conversely, the estimations based on the RAF are more reliable. Among the left-slip faults, the westernmost segment of the HYF and the WQLF are at relatively higher

seismic risks. The potential moment magnitudes of the westernmost segment of the HYF can reach up to 7.3. Considering that the WQLF is beyond the modeling range, it extends continuously to the west of the model, and the surface trace is fitted into a continuous trace. Therefore, the V3 model underestimates the length of the westernmost segment of the WQLF, and the moment magnitude calculated by the RLF is small, which is most likely larger than 7.0. All segments of the WQLF have potential seismic risks of moment magnitude 7.0 or greater, and one segment can reach a maximum of magnitude 7.6.

Compared with the historical earthquake records, we find that the moment magnitudes estimated by the V3 model are small. We believe that this error mainly comes from the following two reasons. Firstly, the fault traces are simplified in the model, and the rupture length and rupture area of the seismogenic fault are underestimated. Secondly, we calculate the moment magnitudes under the assumption that only a single fault segment ruptures, without considering the cascade rupture of active faults, and the seismogenic structure generally involves more than a single fault segment. At present, the accuracy of the V3 model is not yet sufficient to solve the problem. However, our result reveals the relative seismic hazard and the minimum seismogenic potential of each rupture in the region, which is of the reference value.

6 CONCLUSION

In this paper, three versions of 3D fault models of the arc-shaped tectonic belt in the northeastern margin of the Tibetan Plateau have been successfully established separately using geological data, magnetotelluric profiles, and multi-source data. Comparing these three models, it is suggested that the multi-source data 3D geological modeling technology is more advantageous in establishing 3D models close to the real regional structure than the single-source data 3D geological modeling technology. Meanwhile, the structures of the main active faults in the study area are reconfirmed from the model perspective. In particular, the V3 model shows that the Haiyuan fault does not cut through the Moho and is a crustal-scale left-slip fault. The shape of the Liupanshan fault plane, which is steep and deep in the north and opposite in the south, may reflect that the left-slip component of the Haiyuan fault with nearly E-W strike is absorbed by the near NW-SE trending Liupanshan thrust fault. Based on the V3 model, the fault segmentation, and the empirical formulas of moment magnitude-rupture parameters, the modeling faults are segmented and the potential moment magnitudes of each segment are calculated. The thrust faults with relatively higher seismic hazards are the Liupanshan fault and the southeastern segment of the Xiangshan-Tianjingshan fault, of which the maximum moment magnitudes are respectively estimated to be 7.3 and 7.2. Among the left-slip faults in the study region, the most westernmost segment of the Haiyuan fault has a potential moment magnitude of 7.3, and potential moment magnitudes of each segment of the West Qinling fault are at least 7.0, of which one segment can be up to 7.6.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

LGL: modeling, drawing the figures, and writing. ZL: modeling, writing, and verification. WW, CS, YZ, and HL: guidance and suggestion. BX, XW, XD, and LWL: suggestion and communication.

FUNDING

This study was jointly funded by the National Key Research and Development Program of China (2017YFC1500100), the

National Natural Science Foundation of China (42172233, 41772209, and 41902201), and Guangdong Province Introduced Innovative R&D Team of Geological Processes and Natural Disasters around the South China Sea (2016ZT06N331).

ACKNOWLEDGMENTS

We especially thank the two reviewers Renqi Lu and Zhongtai He, Profs. Wenjun Zheng and Zhuqi Zhang gave us vitally important suggestions to improve the manuscript. We thank GEBCO Compilation Group (2021) GEBCO 2021 Grid (doi:10.5285/c6612cbe-50b3-0cff-e053-6c86abc09f8f) and Geospatial Data Cloud site, Computer Network Information Center, Chinese Academy of Sciences (<http://www.gscloud.cn>) for providing the high-resolution DEM data and satellite images for this study.

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Conflict of Interest: The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

The handling editor ZZ declared a shared research program with the author ZL at the time of review.

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Deep Electrical Structure of the Langshan Mountain-Linhe Basin Area on the Northwest Edge of the Ordos Block, China

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OPEN ACCESS

Edited by:

Wenjun Zheng,
Sun Yat-sen University, Zhuhai
Campus, China

Reviewed by:

Zhen Guo,
Southern University of Science and
Technology, China
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Sun Yat-sen University, China
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Specialty section:

This article was submitted to
Geohazards and Georisks,
a section of the journal
Frontiers in Earth Science

Received: 08 April 2022

Accepted: 16 May 2022

Published: 01 June 2022

Citation:

Zhao L, Wang Q, Cao C and Hao M
(2022) Deep Electrical Structure of the
Langshan Mountain-Linhe Basin Area
on the Northwest Edge of the Ordos
Block, China.
Front. Earth Sci. 10:916044.
doi: 10.3389/feart.2022.916044

A series of fault depression structures have developed around the Ordos Block. The Langshan Mountain-Linhe Basin area (LLA), located on the northwest edge of the Ordos Block, is a typical, normal tension fault system. A geological survey shows that the Langshan Piedmont fault (LPF) in this area has a large slip rate and indicates risk of earthquake preparation. Broadband magnetotelluric (MT) exploration research was recently carried out across the LLA in the NW–SE direction, and the three-dimensional deep electrical structure thus obtained revealed that the LPF in the LLA is an evident electrical boundary zone on the whole crustal scale and is the main boundary fault of the primary structural block of the Alxa and Ordos Blocks. The MT results also show that the Linhe Basin and Ordos Block belong to the same tectonic basement. The Linhe and Dengkou faults belong to the internal faults of the Ordos Block. The upper crust of the Langshan Mountain on the west side of LPF is characterized by high-resistivity, the middle and lower crust have a low-resistivity layer, and the Linhe Basin on the east side has a Cenozoic low-resistivity sedimentary layer of approximately 10 km thick, which reveals that the Linhe Basin is a faulted basin with sedimentary thickness around the Ordos Block. This indicates that the LLA has experienced continuous and strong tension, normal fault depression sedimentary activities since the Cenozoic era. The current Global Positioning System velocity field shows that there is an apparent NW–SE acceleration zone in the LLA. The leveling data indicate that Linhe Basin shows a subsidence trend relative to the Ordos Block, indicating that the area is undergoing continuous NW–SE tension and faulting. It is speculated that there is a risk of earthquake preparation in the LPF.

Keywords: magnetotelluric, electrical structure, Langshan Piedmont fault, Linhe Basin, low-resistivity sedimentary layer

1 INTRODUCTION

There are a series of extensional basin-mountain faulted structural areas around the Ordos Block, such as the Langshan Mountain-Linhe Basin area (LLA) on the northwest edge of the Ordos Block (**Figure 1B**), Helanshan Mountain-Yinchuan Basin faulted structural area on the west edge, Qinling Mountain-Weihe Basin faulted structural area on the south edge, and Lvliangshan Mountain-Linfen

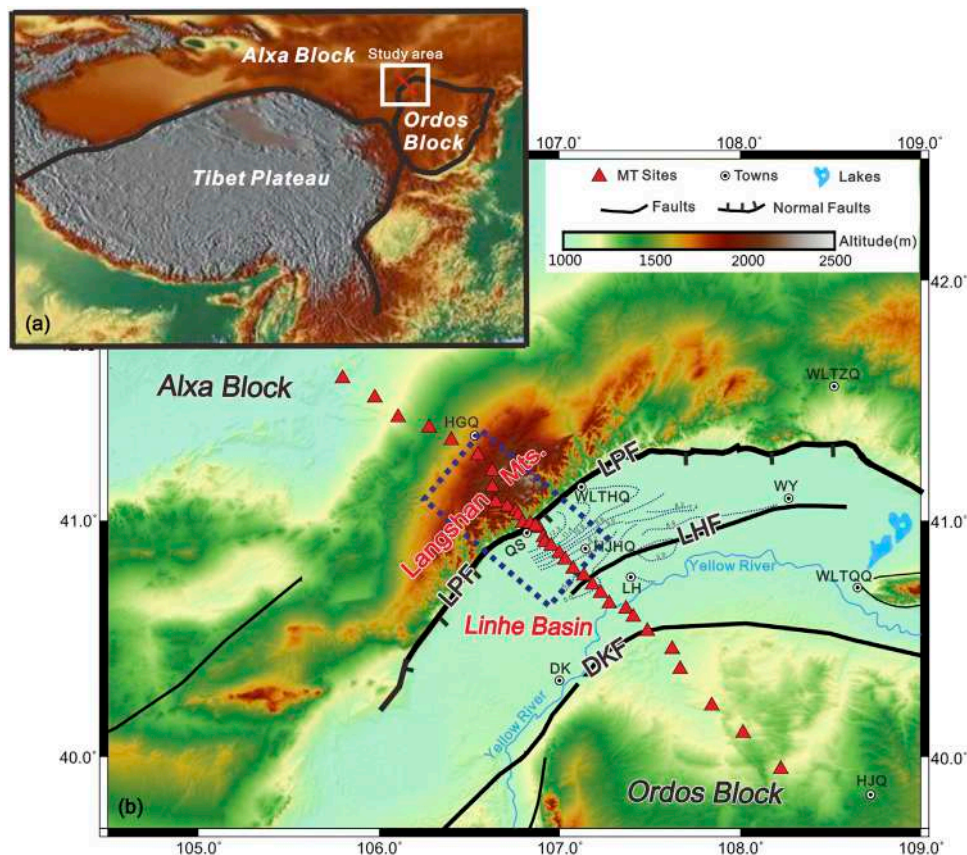


FIGURE 1 | Regional tectonic setting **(A)** distribution map of faults and magnetotelluric (MT) sites on the northwest edge of Ordos Block **(B)** The thickness of the Cenozoic sedimentary layer is quoted from the State Seismological Bureau (1988). LPF, Langshan Piedmont fault; LHF, Linhe fault; DKF, Dengkou fault; QS, Qingshan; HJHQ, Hangjinhouqi; LH, Linhe; DK, Dengkou; WH, Wuhai; WLTHQ, Wulatehouqi; WY, Wuyuan; WLTQQ, Wulateqianqi; WLTZQ, Wulatezhongqi; HJQ, Hangjinqi.

Basin on the east edge. These basin-fault rift tectonic areas are often strong earthquake preparedness zones for the Chinese mainland (State Seismological Bureau, 1988). In 1739, the Pingluo M8 earthquake occurred in the faulted structural area of Helanshan Mountain-Yinchuan Basin, causing more than 50,000 casualties (Lei et al., 2015). The Huashan earthquake occurred in the faulted structural area of the Qinling Mountain-Weihe Basin in 1,556 and was the most serious natural disaster in Chinese historical records (Hou et al., 1998; Rao et al., 2014; Du et al., 2017; Feng et al., 2020). The Hongdong M8 earthquake occurred in the structural area of the Lvliangshan Mountain-Linfen Basin in 1,303 (Deng et al., 2003; Xu et al., 2018). These areas have been the key research areas in the field of Geosciences in recent years. The LLA is located in the northern part of the north-south seismic belt of the Chinese mainland, and it is also subjected to the northeast (NE) pushing of the Tibetan Plateau and northwest-southeast (NW-SE) tensioning caused by deep subduction of the Pacific plate (Northrup et al., 1995; Zhang et al., 2006). The Langshan Piedmont fault (LPF) in this structural area is a large fault that controls the NW boundary of the Hetao Basin (the Linhe Basin belongs to the western region of the Hetao Basin). The Hetao Basin can be divided into three secondary depressions from west to east: Linhe Basin, Baiyanhua Basin, and

Huhe Basin), and may also be the boundary fault between the Alxa and Ordos Blocks (State Seismological Bureau, 1988; Rao et al., 2016; Zhang et al., 2020). Geological surveys show that activity of the LPF, especially in the middle section, is very intense (Dong et al., 2016; Rao et al., 2016; Dong et al., 2018a; 2018b). Dong et al. (2018a) and Rao et al. (2016) revealed that the fault is an important seismic zone in the Hetao Basin through trenching, dating, and other methods. The average recurrence period of strong earthquakes is about 2,500 A, and the average magnitude of each earthquake reaches 7–8 (Dong et al., 2018a, 2018b; Liang et al., 2021). At present, the recurrence period of strong earthquakes at the fault has been exceeded, so there is a great possibility of destructive, strong earthquakes occurring again (Dong et al., 2018a). Li et al. (2015) and Rao et al. (2016) believe that the latest earthquake rupture event of LPF may be related to the great earthquake in 7 BC. Liang et al. (2021) believe that the magnitude of the earthquake in 7 BC reached 8.1, and there is still a risk of magnitude 7.4–8.0 earthquakes in this area. Due to the lack of large earthquakes recorded by modern instruments in the fault depression structural area of LLA, the existing research results in this area focus on geomorphology and paleoseismology (Jia et al., 2016; Rao et al., 2016; He et al., 2017, 2018; Dong et al., 2018a, 2018b, 2018c; He et al., 2020; Zhang

et al., 2020; Li J. et al., 2021; Liang et al., 2021; Xu et al., 2022). The results obtained lack the support of deep geophysical data. The Linhe Basin has had a relatively developed economy and society since ancient times; there are important towns such as Linhe (LH), Hangjinhouqi (HJHQ), and Wulatehouqi (WLTHQ) with large population densities. Therefore, the deep distribution of faults and seismic risk in the structural area of LLA is of great concern.

Wu et al. (2022) recently obtained the large-scale, three-dimensional (3D) S-wave velocity structure of the crust and upper mantle of the Ordos Block and its surrounding areas and found that there is a thickening of the Cenozoic sedimentary layer in the Linhe Basin. Chen et al. (2020) revealed that the Moho depth of Alxa Block is 38–45 km, that of Ordos Block is 38–42 km, and that of LLA in the transition zone is 52 km. Seismology also shows that there are obvious differences in the crustal structure between the Alxa and Ordos Blocks (Chen et al., 2020). It is speculated that the Alxa and Ordos Blocks belong to different tectonic units. Chang et al. (2011) found that the anisotropic fast wave direction in LLA was in a NW–SE direction when summarizing the anisotropic characteristics of the Ordos Block and surrounding upper mantle, which indicates that there may be NW–SE stretching in this area. Numerical simulation shows that the lateral horizontal movement of the stable Ordos Block has good integrity (Chen et al., 2011). Scholars have carried out Coulomb stress statistics (Zhang et al., 2021), anisotropy analysis (Hu et al., 2011), and crustal strain analysis (Cui et al., 2016) in the LLA on the northwest edge of the Ordos Block and have concluded that the LLA is in a state of tensile strain. The results of multi-scale tomography reveal that the northwest edge of the Ordos Block may have experienced mantle thermal material upwelling, transforming the lithospheric structure (Gao et al., 2018). These research results focus on crustal deformation and large-scale geophysical imaging; there is a lack of research on the fine structure of deep faults, especially fine, deep, geophysical exploration near the LPF.

The electrical structural information of the crust obtained using the magnetotelluric (MT) method can provide a basis for the division of block units (Chen, 1990). It is an important deep geophysical method used to detect the deep extension characteristics of the fault zone and has been widely used (Zhao et al., 2012; Cai et al., 2017; Yu et al., 2017; Wang et al., 2018; Hu et al., 2020; Sun et al., 2020; Li L. et al., 2021; Ye et al., 2021; Zhan et al., 2021; Zhao et al., 2021; Chang et al., 2022; Zhao et al., 2022). Predecessors have completed much MT exploration work in the Ordos Block and its surrounding areas. The Ordos Block belongs to a stable block unit and is in contact with the surrounding blocks (Zhao et al., 2004, 2010; Dong et al., 2014; Li et al., 2017; Xu et al., 2017; Zhan et al., 2017; Han et al., 2022). The LLA in the contact area between the Ordos and Alxa Blocks lacks MT exploration results. Therefore, we completed a broadband MT profile with a length of more than 270 km passing through the LLA. The MT profile started from the Alxa Block in the west and extended to the Ordos Block in the east. The data were inversed and used to calculate the deep electrical structural information of the area. From the perspective of electrical

structure, this paper discusses the deep structural characteristics of LLA and its adjacent areas, and further evaluates the contact relationship between the Ordos and Alxa Blocks, and then discusses the dynamic background of LLA and the earthquake breeding risk of LPF in combination with the regional 3D deformation field data. The results obtained can advocate the seismic risk assessment of LLA, and can also provide deep geophysical data for the formation mechanism of faulted basins around the Ordos Block.

2 GEOLOGIC SETTING

LLA is located in the contact area between the Alxa and Ordos Blocks, which is a part of Hetao Graben (Rao et al., 2016; Zhang et al., 2020). The LPF is located at the southeast foot of the NE-trending Langshan Mountain. It intersects with the EW-trending Sertengshan Piedmont fault in the north and terminates in Ulanbuhe Desert in the southwest. The total length of the LPF is approximately 160 km and the overall strike is NE 55°. The overall tendency of the fault is SW (Dong et al., 2018a; 2018b). According to the block division theory of Deng et al. (2003), LPF is the boundary fault between the Ordos and Alxa Blocks. The Langshan Mountain on the west side of the fault belongs to the Alxa Block, and the Linhe Basin on the east side of the fault belongs to the Ordos Block (Rao et al., 2016; Zhang et al., 2020). Previous researchers have summarized (State Seismological Bureau, 1988; Dong et al., 2016; Dong et al., 2018a, 2018b; He et al., 2020; Liang et al., 2021) the evolutionary history of multiple stages in the LLA: 1) Oligocene stage: This was the initial developmental stage of LLA. Due to the activity of LPF and the relative uplifting of the mountain, the Linhe Basin began to form. The sedimentary thickness of the subsidence center was 2,400 m and the subsidence rate was 0.10 mm/a (State Seismological Bureau, 1988). The maximum thickness of the stratum was mainly distributed along the LPF. 2) Late tertiary stage: This was the most active stage in the development of the LLA. The Linhe Basin continued to decline strongly under the control of the Piedmont faults, with a sedimentary thickness of 3,800 m and a subsidence rate of 0.20 mm/a. 3) In the Pliocene stage, the faulting was more intense, the sedimentary range was further expanded, the thickness reached 6,000 m, and the settlement rate was 0.64 mm/a. 4) In the Quaternary period, the Piedmont Fault activity was still relatively strong, generally faulting the Quaternary system, and the fault activity has maintained its characteristics since the late Tertiary. The mountain rose slowly and the basin continued to settle. The thickness of the Quaternary system was 2,400 m and the settlement rate was 1.00 mm/a. Drilling and petroleum geological profile data (State Seismological Bureau, 1988; Liang et al., 2021) show that Linhe Basin is a sedimentary basin with the largest sedimentary layer thickness around the Ordos Block and is much higher than that of other basins around the Ordos Block over the same period, possibly exceeding 10 km near HJHQ. This shows that the Linhe Basin has a high subsidence rate, which should correspond to very strong fault depression sedimentary activities.

3 MT METHOD AND DATA PROCESSING

3.1 Introduction to the MT Method

The core of the MT method is to determine the underground conductivity structure by using the Earth's external magnetic field and electric field, which vary with time. Specifically, conductivity or resistivity can be determined by the dependence of amplitude and the phase relationship between the electrical field and the magnetic field vector on the Earth's surface. These relationships are expressed in the frequency domain by the correlation transfer function between the surface components of the electromagnetic field, which normalizes the induced response of the magnetic field fluctuation intensity. The transfer function is represented by the impedance tensor, Z , which is defined as:

$$E = ZH, \quad (1)$$

where E is the electric field vector and H is the magnetic field vector (Chen et al., 1990).

In recent years, with the application of supercomputers and the increasingly mature and practical MT 3D inversion algorithms (Siripunvaraporn et al., 2005; Egbert et al., 2012; Kelbert et al., 2014), 3D inversion technology has been widely used in the practical research of various MT detections (Dong et al., 2014; Sun et al., 2019, 2020; Zhao et al., 2022).

3.2 MT Profile and Data Acquisition

The field data acquisition of LLA was carried out from September to October 2021 and took approximately 35 days. Four sets of MTU-5A equipment (Phoenix Company, Canada) were used for data acquisition at the same time (frequency band range 320–0.0005 Hz). The completed MT profile started from Alxa Block in the border area between China and Mongolia in the northwest, passed through LLA in the southeast, exited the Langshan Mountain near Qingshan (QS), entered Linhe Basin, passed through the Yellow River, entered the hinterland of the Ordos Block, and finally terminated near HJQ (Figure 1B). The contact area between the Langshan Mountain and the Linhe Basin is the area through which LPF passes, and it is also the key area of this study (as shown by the blue empty frame in Figure 1B). A total of 34 MT sites were obtained along the profile. The profile was ~270 km long and densified to 2–3 km near the LPF, which was mainly used to obtain the fine electrical structural information of the LPF area. Due to the relatively simple structure in the peripheral Alxa and Ordos Blocks, the MT site spacing was increased to 10–20 km, which was mainly used to obtain the large-scale, regional electrical structural information (Figure 1B).

3.3 Electrical Orientation Analysis

The phase tensor decomposition technique can be used to obtain the electrical orientation information of the MT profile (Caldwell et al., 2004). The rose-petal diagram of the electrical orientation of the full frequency band and sub-band of all 34 MT sites along the profile was calculated using the phase tensor decomposition method in the MTP software developed by Chen et al. (2004). The rose petal diagram of electrical orientation indicates that the electrical orientation of the full

frequency band of all sites along the profile is NE 60° or NW 30° (Figure 2A). Combined with the tectonic characteristics of the Langshan mountain, LPF, LHF, and DKF in this area, which mainly show a NE 60° orientation, it is reasonable to judge that the electrical orientation of this area is NE 60° on the whole. The rose-petal diagram of the electrical orientation of the sub-band shows similar characteristics to the full-band, which indicates that the electrical orientation of the study area is relatively uniform from shallow to deep and the linear characteristics are more evident.

3.4 Dimensional Analysis

The phase tensor invariant, φ_2 , of each MT site was obtained using the phase tensor decomposition technique (Caldwell et al., 2004). The distribution characteristics of φ_2 with frequency can indicate the variation pattern of the apparent resistivity value below that MT site, which is actually the variation relationship of the apparent resistivity value with depth due to the correspondence between frequency and depth. When φ_2 is less than 45°, the apparent resistivity value below this MT site increases with depth, and when φ_2 is greater than 45°, the apparent resistivity value below this MT site decreases with depth (Heise et al., 2008). Figure 2B shows the distribution of the phase invariance, φ_2 , with frequency for 34 sites along the profile. The electrical variation trends of the plots along the profile are not consistent, the electrical variation trends of the Ordos and Alxa Blocks are relatively simple, and the variation of φ_2 is mainly from high to low and then to high values, which shows that the stratigraphic characteristics of these two blocks are more clear. The value of φ_2 fluctuates more on both sides of the LPF and the high frequency part of the Langshan Mountain is dominated by a low phase, while the middle and low frequency parts are dominated by high frequency, and the φ_2 is larger than 70°. This shows that there may be an evident low-resistivity structure below the Langshan Mountain. In the Linhe Basin east of the LPF, the middle-high frequency part of φ_2 is dominated by high values (black dashed part in the figure), which indicates that there may be an existing, large-scale, low resistivity structure in the shallow part of the area, which corresponds to the existence of a thick, low-resistivity sedimentary layer in the Linhe Basin (Liang et al., 2021). The φ_2 in the low-frequency part of the Linhe Basin is dominated by low values, indicating that the resistivity values in the Linhe Basin show a significant increasing trend with increasing depth.

The phase tensor decomposition technique can also obtain the two-dimensional (2D) skewness angle ($|\beta|$) with a frequency distribution for each MT site. Considering the interference in the tectonic zone of LLA, it was determined that the subsurface medium could be approximated as a one-dimensional (1D) or 2D case when $|\beta|$ is less than 6, and the subsurface medium can be considered as a 3D case when $|\beta|$ is greater than 6. The larger the value of $|\beta|$, the stronger the three-dimensionality of MT data (Caldwell et al., 2004; Bibby et al., 2005; Cai et al., 2017). The high-frequency part of $|\beta|$ in the Linhe Basin is smaller (black dashed part in the figure), close to 0, and extends all the way to the vicinity of the Ordos Block, indicating that the shallow part of the region is an

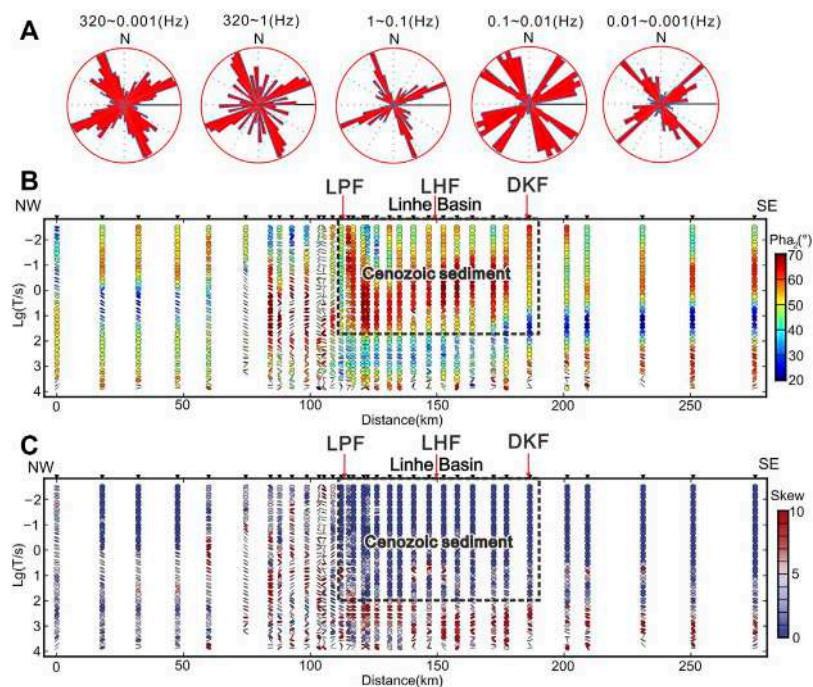


FIGURE 2 | Rose-petal diagram of the electrical orientation (A), phase tensor invariant, φ_2 (B), and two-dimensional skewness angle, $|\beta|$ (C) of different periods of the profile.

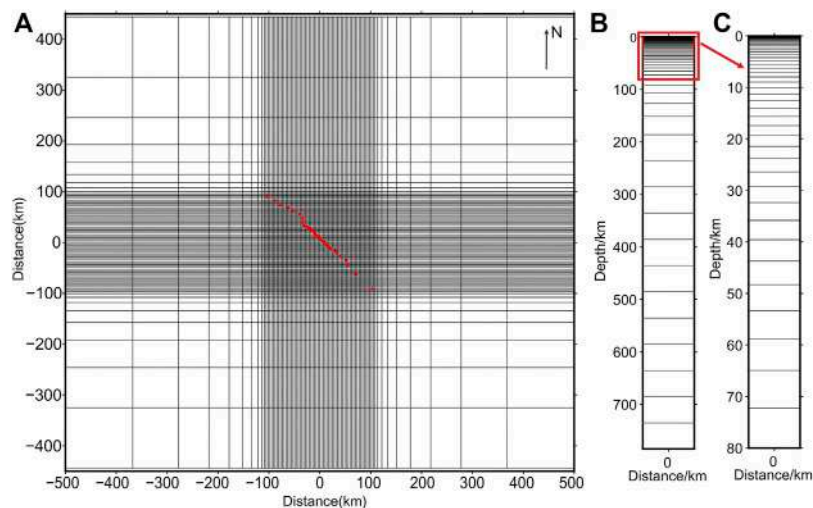


FIGURE 3 | Initial model used for the three-dimensional inversion of the study region (A) horizontal grid, (B) vertical grid, and (C) vertical grid of the target area. The red dots denote magnetotelluric (MT) sites.

approximately 1D or 2D laminar structure, which is presumed to be due to the thicker isotropic Cenozoic sedimentary layers, which is more similar to the distribution characteristics of φ_2 (Figure 2C). The overall $|\beta|$ indicates that the Langshan Mountain is more complex than the Linhe Basin, while the smaller values of $|\beta|$ in the Ordos and Alxa Blocks indicate that the tectonics of these two regions are relatively simple.

4 3D Inversion

In this study, the resistivity result model of the study area was calculated using the 3D inversion algorithm to invert the original data collected in the field. The latest version of the ModEM program (Egbert et al., 2012; Kelbert et al., 2014) was used and the full impedance tensor (four components: Z_{xx} , Z_{xy} , Z_{yx} , and Z_{yy}) was used as the input data for the 3D inversion calculation. A total

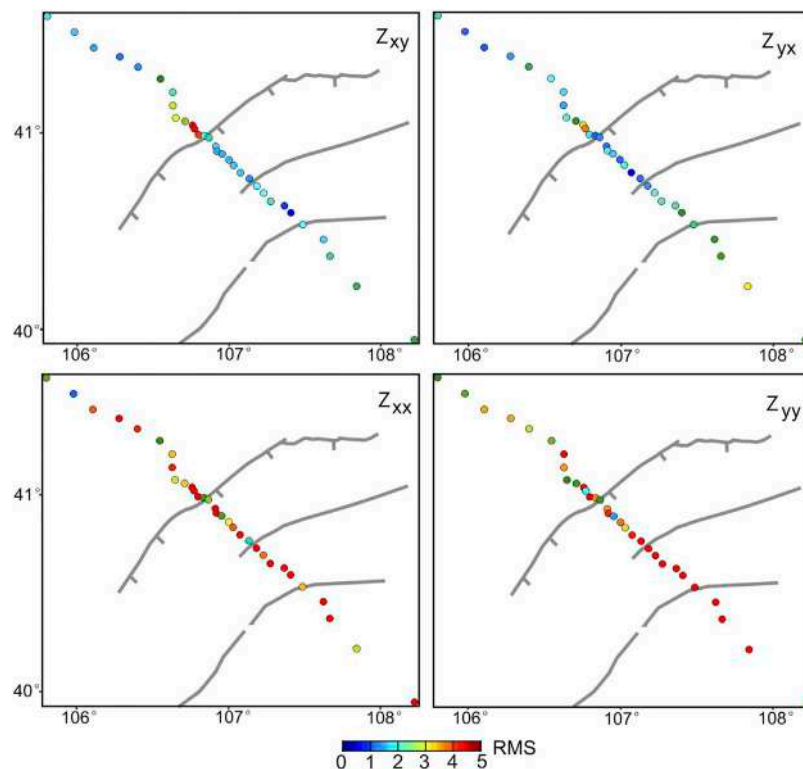


FIGURE 4 | Root mean square error (RMS) distribution of four components (Z_{xy} , Z_{yx} , Z_{xx} , Z_{yy}) at each magnetotelluric (MT) site of the three-dimensional inversion.

of 35 frequency points were used in the 3D inversion calculated from high to low frequency (**Supplementary Figure S5**). Before inversion calculation, data preparation and unified interpolation were carried out using MTP software (Chen et al., 2004) and Topeak software (Liu et al., 2017). The initial model grid was constructed using 3D inversion and was divided into horizontal and vertical grids (**Figure 3**). The horizontal grid was divided into the core part of the MT sites involved in the calculation and the expanded boundary grid. The horizontal grid of the core part was uniformly dissected using 3.0×3.0 km spacing, and a total of 60 (NS) $\times$ 60 (EW) grids were obtained. The extended boundary grid was constructed outside the core part by expanding 10 grids in four directions with 1.5 time multipliers. Finally, a total of 80×80 grids were constructed in the horizontal direction. The vertical grid design was similar to that of Cai (2017), Sun (2020) and Zhao (2022), and a total of 75 layers were obtained in the vertical direction.

In the 3D inversion calculation, the selection of data threshold error was consistent with the previous work of Sun (2020) and Zhao (2022). Since the diagonal components (Z_{xx} , Z_{yy}) were too small and the error was too large, we enlarged the error of the diagonal components in the actual 3D inversion and decreased the inversion weighting. The data threshold errors of diagonal components (Z_{xx} , Z_{yy}) and off-diagonal components (Z_{xy} , Z_{yx}) were set to 10% and 5%, respectively. A 100Ω m uniform half space was used as the initial model and an automatically updated regularization factor was adopted. The initial value of the

regularization factor was 5,000. When the inversion was no longer convergent, the regularization factor was updated to one 10th of the original value and the inversion was continued. The inversion was calculated many times and all data were carefully selected. After 65 iterations to the end of the inversion, the root mean square error (RMS) was reduced from 32 to 3.1. The RMS distribution map of each component for all 34 MT sites is shown in **Figure 4** and the detailed inversion process is shown in **Supplementary Figures S5–S7**; **Table S1**.

5 ELECTRICAL STRUCTURE IN THE LANGSHAN MOUNTAIN-LINHE BASIN AREA

Figure 5 shows the 3D inversion result within the depth range of 0–60 km along the profile. The topographical changes along the profile and the distribution of blocks and faults are drawn above the electrical structure image. The study area was divided into the Alxa Block, Langshan Mountain, Linhe Basin, and Ordos Block (Dong et al., 2016; Dong et al., 2018a, 2018b; He et al., 2020; Liang et al., 2021). Among them, the terrain of the Langshan Mountain fluctuates greatly, with an altitude of 1,500–2,300 m. The terrain of the Linhe Basin is relatively flat, and the average altitude is $\sim 1,000$ m (**Supplementary Figure S2**). The maximum altitude difference between the two blocks is up to thousands of meters, which is the location of LPF at the boundary of the two blocks. To

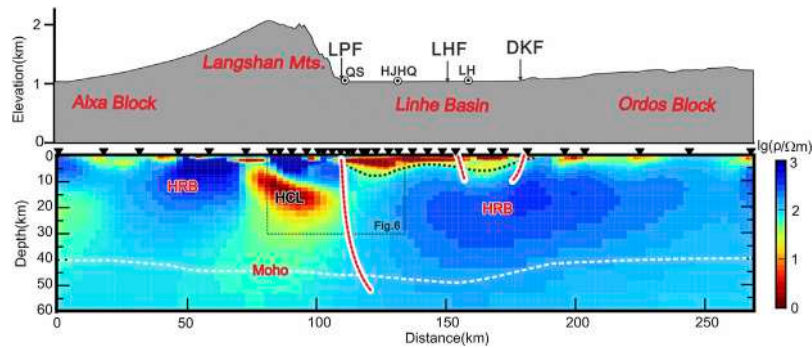


FIGURE 5 | Deep electrical structure map obtained using three-dimensional inversion and the topographical distribution map of the profile. The red dotted lines are the deep distribution of the fault inferred from the magnetotelluric (MT) results, and the black dotted line is the bottom boundary of the Cenozoic sedimentary layer inferred from the MT results. The white dotted line is distributed along the Moho surface (Chen et al., 2020).

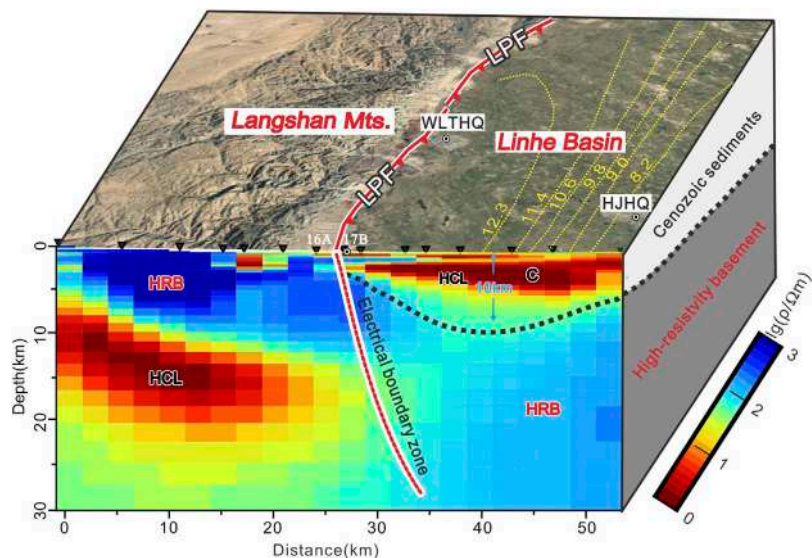


FIGURE 6 | Electrical structure and interpretation map of the core area of the Langshan Mountain-Linhe Basin area (the location is shown in the black dotted box in Figure 1B). The thickness of the Cenozoic sedimentary layer is quoted from the State Seismological Bureau (1988).

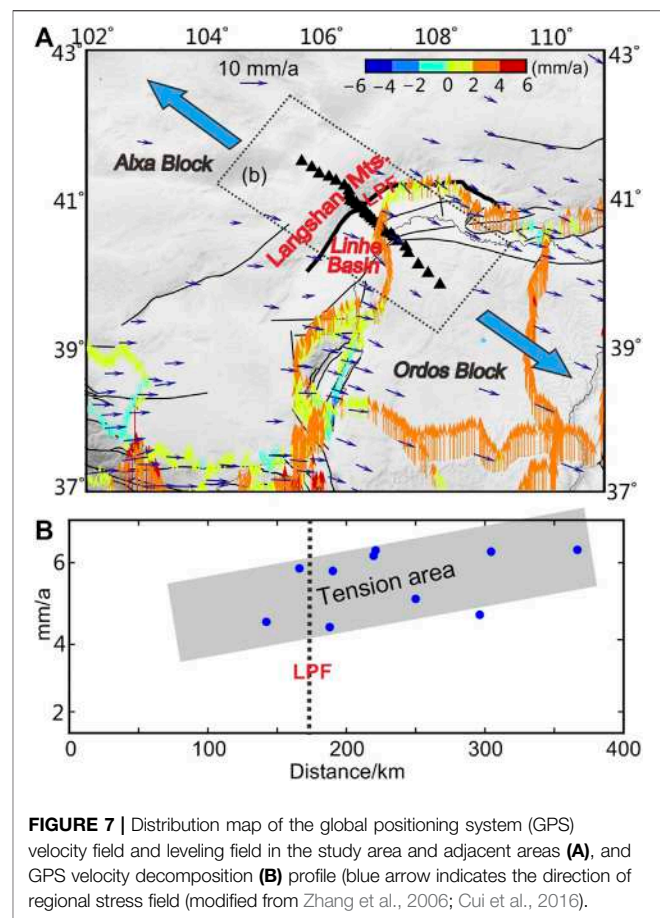
show the fine structure of LPF more intuitively and the sedimentary layer thickness of the Linhe Basin, the electrical structure and geomorphic distribution map within 30 km of the core area of LLA are drawn in **Figure 6** (the range is shown in the blue dotted box in **Figure 1** and **Figure 5**).

The LPF is an obvious electrical boundary zone between the Langshan Mountain and Linhe Basin, whether it is 10 km above or below the area (**Figure 5**). It is not only the boundary fault between the Langshan Mountain area and Linhe Basin, but also the largest main boundary fault in the study area. This corresponds to the view of Wen et al. (2014) that the structural deformation in the LLA is mainly distributed on the LPF. Dong et al. (2018a) revealed that the maximum sliding rate and maximum displacement of the fault are located in the middle of the fault, which is also where the MT profile passes through at

this time. The upper crust in the Langshan Mountain area west of the LPF is a horizontally layered, high-resistivity body, with a depth of nearly 10 km as a whole, showing relatively complete high-resistivity characteristics, corresponding to the Paleozoic bedrock strata exposed in the Langshan Mountain area above (Rao et al., 2016; Dong et al., 2018a; 2018b; Zhang et al., 2020; Liang et al., 2021). The middle and lower crust of the Langshan Mountain area is dominated by a low-resistivity/high-conductivity layer (HCL), and there is a clear high and low electrical boundary zone between the middle and lower crust, and the high resistivity body structure of the upper crust at the bottom to 10 km, which indicates that the deep medium characteristics of the Langshan Mountain have a predominantly layered distribution. The HCL in **Figures 5, 6** is a large-scale, low resistivity structure in the middle and lower

crust below the Langshan Mountain. Based on other geophysical data, such as the results of multi-scale tomography, we speculate that this may be due to the upwelling of mantle thermal materials and the transformation of the lithospheric structure of Langshan Mountain (Gao et al., 2018). There is a low-resistivity layer in the shallow part of the Linhe Basin to the east of LPF. The distribution of the low-resistivity layer is relatively flat on the profile, and the bottom depth is more than 5 km, becoming shallower and shallower to the east and extending near DKF.

To determine the specific location of LPF more accurately, intensive observation was carried out in the contact area between the Langshan Mountain and the Linhe Basin (Figure 6). It was determined that the accurate location of LPF is between the 2 MT sites, LSL1-16A and LSL1-17B (see Supplementary Figure S4 for the apparent resistivity phase curve of all MT sites and the red triangle in Figure 6). The shallow part of the Linhe Basin is dominated by a low-resistivity layer, combined with 2D deviation angle $|\beta|$ distribution (Figure 6). Previous geological survey results (Dong et al., 2016; He et al., 2017; 2018; 2020) speculate that these low-resistivity layers represent the Cenozoic sediments in the area, with evident 1D or 2D characteristics. The thickness of the low resistivity sedimentary layer near HJHQ in the hinterland of the Linhe Basin exceeds 10 km, which confirms the early geological survey that the Linhe Basin is a sedimentary basin with the largest thickness of sediment around the Ordos Block and that the Cenozoic sedimentary layer may reach 5–6 km (State Seismological Bureau, 1988). The latest research of Liang et al. (2021) found that the sedimentary layer of the Linhe Basin may be close to 10 km in the area crossed by the MT profile, which is similar to our results. Wu et al. (2022) recently obtained the large-scale, three-dimensional (3D) S-wave velocity structure of the crust and upper mantle of the Ordos Block and its surrounding areas, and found that there is a huge thickness of Cenozoic sedimentary layer in the Linhe Basin. The extremely thick Cenozoic sedimentary layer further reflects that the LLA has experienced continuous and strong tension, and normal fault depression sedimentary activities since the Cenozoic stage (State Seismological Bureau, 1988; Dong et al., 2018a; 2018b). It is one of the areas with the strongest tension in the Ordos Block. There is an evident high-resistivity body under the low-resistivity sedimentary layer in Linhe Basin. The high-resistivity basement is connected with the Ordos Block and extends to the hinterland of the Ordos Block, indicating that Linhe Basin and Ordos Block may belong to the same structural unit. The Alxa Block to the west of Langshan Mountain has high-resistivity and is a relatively simple structure. Ordos Block was also shown to be a relatively complete block unit, which is similar to the previous MT exploration results (Zhao et al., 2010; Dong et al., 2014; Li et al., 2017; Xu et al., 2017; Han et al., 2022). Through the two north-south MT profiles (Xu et al., 2017; Li et al., 2017) in Hetao Basin, it was found that the scale of the crust in the north of the Ordos Block is mainly a layered high resistivity body, which corresponds to our MT results. There is a low resistivity layer near the mantle. Our MT results this time did not find a similar situation, which may be due to the short profile. Our results reveal that the electrical structural characteristics of the Ordos Block are



similar to those of Han et al. (2022) in the Helanshan Mountain-Yinchuan Basin. In addition, the electrical structural image in Figure 5 shows that LHF and DKF in Linhe Basin are small in scale and do not pass through the high-resistivity body below the Linhe Basin and Ordos Block. These two faults belong to the internal faults of the Ordos Block. It is speculated that it may be an associated fault in the scale of upper crust formed during the process of tension in the structural area of LLA.

6 DYNAMIC ENLIGHTENMENT OF THE LANGSHAN MOUNTAIN-LINHE BASIN AREA

Active normal faults in China mainly develop around the Ordos Block (State Seismological Bureau, 1988). Except that the southwest edge of the Ordos Block is strike slip thrust, the rest are normal fault basins. Hetao Basin, including Linhe basin, is a sinistral shear tension zone (Fan et al., 2003; Chen et al., 2011). Since the Miocene, the Ordos Block has been characterized by NW–SE tensile stress and NE–SW compressive stress (Northrup et al., 1995; Zhang et al., 2006; Cui et al., 2016). Under the current tectonic stress field environment, the normal fault activity of the NE-distributed LPF coordinates the maximum normal strain produced by the regional stress field. In the active fault zone

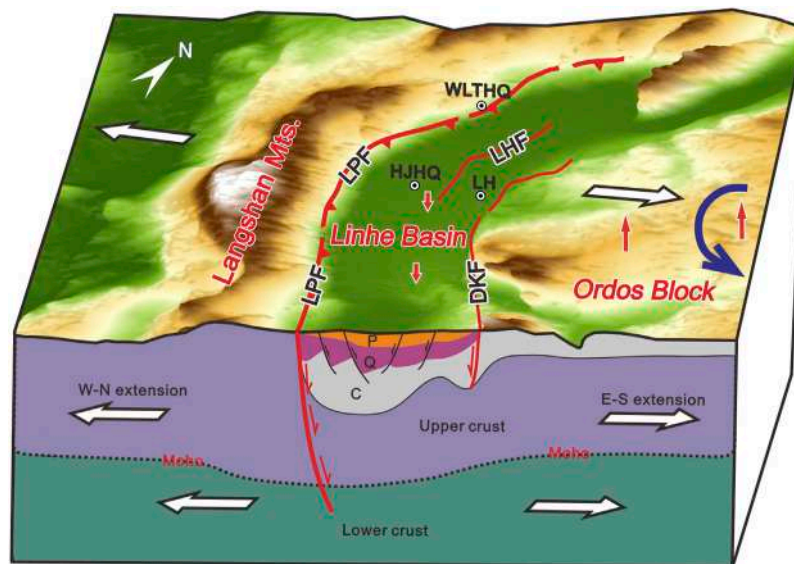


FIGURE 8 | Graphic of the tectonic movement mode in the Langshan Mountain-Linhe Basin area on the northwestern edge of the Ordos Block. The blue arrow represents the counterclockwise rotation of the Ordos Block; Moho quoted from Chen et al. (2020). P, Pleistocene; Q, Quaternary; C, Cenozoic (State Seismological Bureau, 1988)

around Ordos, the LPF may have a large seismic risk (Dong et al., 2018a, 2018b; Liang et al., 2021). The LPF has been located in the transition area between the Pacific extensional tectonic domain and the Tethys compression tectonic domain since the late Mesozoic (Ren et al., 2002; Zhang et al., 2006). It is a pre-existing, large-scale fault system. The study of the activity characteristics and trends of the fault is of great significance to earthquake prevention and disaster reduction in the Linhe Basin and even Hetao Basin (Rao et al., 2016; Dong et al., 2018a, 2018b; Zhang et al., 2020; Liang et al., 2021). Therefore, in addition to using the MT method to obtain the crustal scale electrical structural characteristics of LLA where the LPF is located, this study also used the GPS plane motion data of the study area from 2009 to 2015 (Hao et al., 2019) and the precise leveling vertical motion data from 1970 to 2014 (Hao et al., 2016) recently obtained by the second monitoring center of the China Earthquake Administration. The combination of static deep electrical structure data and dynamic 3D crustal movement field data provides better constraints for the establishment of geodynamic models in the Linhe Basin and the periphery of the Ordos Block, and provides reliable reference data for seismic risk assessment in the LLA area.

Figure 7A shows the GPS velocity distribution field of LLA and its adjacent areas relative to the Eurasian block from 2009 to 2015 (blue plane arrow) and the annual vertical deformation field of LLA and its adjacent areas relative to the Ordos Block obtained from precise leveling data from 1970 to 2014 (color vertical arrow) (data quoted from Hao et al., 2016; 2019). The GPS velocity field within the black box is decomposed into the direction along the MT profile (**Figure 7B**; the north and east are positive). The GPS velocity field shows an overall trend of SE movement in the study area, and the GPS velocity field values

show an increasing trend from the Alxa Block through the LLA to the Ordos Block, increasing from 4 mm/a to 7 mm/a. This shows that the LLA is an evident accelerating zone, indicating that the area is undergoing continuous NW-SE tension. Current research on tectonic stress fields shows that the fault depression system around the Ordos Block, including the structural area of LLA, shows NW-SE horizontal tensile stress (Zhang et al., 2006; Cui et al., 2016), which corresponds to the results obtained in this study. The precise leveling data from 1970 to 2014 indicate that the Linhe Basin is dominated by uplift (colored vertical arrow in **Figure 7A**), and the uplift rate is approximately 0–2 mm/a, which is less than the annual uplift rate of 2–4 mm/a in the hinterland of the Ordos Block (Hao et al., 2016). The Linhe Basin and Ordos Block belong to the same tectonic unit, while the Linhe Basin shows subsidence relative to the Ordos Block, which may be due to the strong tension in LLA. A movement model map of LLA on the northwest edge of the Ordos Block is postulated (**Figure 8**). As the Ordos and Alxa Blocks are currently subjected to the NE pushing of the Tibetan Plateau and the deep subduction and extension of the Pacific plate (Northrup et al., 1995; Zhang et al., 2006), the two blocks continue to carry out evident separation activities, resulting in normal tension fault depression sedimentary activities in LLA, forming a huge fault depression basin. The LPF is the boundary fault between the two blocks and there is concomitantly a strong risk of earthquake preparation.

7 CONCLUSION

Broadband magnetotelluric exploration was carried out across the LLA in the NW-SE direction, and the 3D, deep electrical structure revealed that the LPF in the LLA was an evident

electrical boundary zone on the whole crustal scale and is the main boundary fault of the primary structural block of the Alxa and Ordos Blocks. The MT results also show that Linhe Basin and Ordos Block belong to the same tectonic basement. The Linhe fault (LHF), and Dengkou fault (DKF) belong to the internal faults of the Ordos Block. The upper crust of the Langshan Mountain on the west side of LPF is characterized by high-resistivity, the middle and lower crust have low-resistivity, and the Linhe basin on the east side has a Cenozoic, low-resistivity sedimentary layer of ~5–10 km thick. This indicates that the Linhe Basin is a faulted basin with sedimentary thickness near the Ordos Block, reflecting that the LLA has experienced continuous and strong tension and normal fault depression sedimentary activities since the Cenozoic stage. The current GPS velocity field shows that there is an evident NW–SE accelerating zone in the LLA. The leveling data indicate that the Linhe Basin shows a subsidence trend relative to the Ordos Block, indicating that the area is undergoing continuous NW–SE tension and faulting. It is speculated that there is a risk of earthquake preparation in the LPF.

DATA AVAILABILITY STATEMENT

The raw data supporting the conclusion of this article will be made available by the authors, without undue reservation.

AUTHOR CONTRIBUTIONS

LZ completed the field data acquisition, the data analyses, and wrote the article. QW provided funding for the study. CC assisted in data acquisition. MH provided GPS and leveling data. All

authors contributed to manuscript revision and discussion and approved the submitted version.

FUNDING

This paper is jointly funded by the National Key Research and Development Plan (2017YFC1500100), the Open Fund of the State Key Laboratory of Seismic Dynamics (LED 2019B06, LED 2021B04), and the National Field Scientific Observation and Research Station Project of Continental Rift Dynamics in Taiyuan, Shanxi Province (NORSTY 20-07).

ACKNOWLEDGMENTS

Sun Shufeng, Li Xi'an, Li Fengquan and others provided help in the process of field data acquisition. Researcher Chen Xiaobin of the National Institute of Natural Hazards, Ministry of Emergency Management of China and Dr. Liu Zhongyin of the Institute of Geology, China Earthquake Administration provided MTP and Topeak series software for the 3D inversion calculation and gave detailed guidance; Dr. Xu Jianhong of the Second Monitoring and Application Center, China Earthquake Administration, provided help in structural interpretation.

SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.916044/full#supplementary-material>.

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Late Pleistocene Slip Rates on an Active Normal Fault in the Northwestern Ordos Block, China

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The slip rates of normal faults on the northern margin of the Ordos Block are critical for understanding the deformation of the Ordos Block and adjacent areas. In this study, we obtain the late Pleistocene slip rate of an active normal fault, the Zhuozishan West Piedmont Fault (ZWPF), in the northwestern corner of the Ordos Block. Satellite imagery interpretation and field investigations suggest that the fault displaces late Pleistocene alluvial fans and forms west-facing fault scarps. To quantify the vertical slip rate of the ZWPF, we measured the displacements along the fault scarps using differential GPS and an unmanned aerial vehicle system and dated the displaced geomorphic surfaces using optically stimulated luminescence (OSL) dating methods. The vertical slip rate of the fault is constrained to 0.1 ± 0.02 mm/a, which suggests a horizontal extension rate of 0.05 mm/a for a dip of 65° . This rate is consistent with those on similar structures in the northwestern Ordos Block. Combining this result with published slip rates along other active faults, we infer that the NW-directed extension rate across the northwestern corner of the Ordos Block is $\sim 0.6\text{--}1.3$ mm/a. This rate is comparable with the geodetic strain rate. Given the extension rate, we believe that extensional deformation is mainly concentrated on the western side of the basin.

Keywords: vertical slip rate, normal fault, Zhuozishan west piedmont fault, ordos block, slip partitioning

OPEN ACCESS

Edited by:

Chuanyou Li,
China Earthquake Administration,
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Reviewed by:

Gang Rao,
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 10 April 2022

Accepted: 17 May 2022

Published: 03 June 2022

Citation:

Liu X, Gao Z, Shao Y and Yao Y (2022)
Late Pleistocene Slip Rates on an
Active Normal Fault in the
Northwestern Ordos Block, China.
Front. Earth Sci. 10:916905.
doi: 10.3389/feart.2022.916905

INTRODUCTION

A series of basin systems have formed around the Ordos Block because of the northeastward growth of the Tibetan Plateau and the uplift of the Ordos Block since the Cenozoic; examples include the Hetao Basin to the north, the Jilantai and Yinchuan Basins to the northwest, the Shanxi Graben System to the east and the Weihe Basin to the south (**Figure 1**) (Research Group of Active Fault System around Ordos Massif, 1988). These basin systems are mainly controlled by active normal faults. The vertical slip rate of an active normal fault is an important parameter for understanding the basin evolution and regional slip partitioning (Wallace, 1987; Wesnousky, 2012; Middleton et al., 2016; Liang et al., 2018; Chevalier et al., 2020). The slip rate, together with coseismic displacement, is also used to estimate the average recurrence interval of a strong earthquake and then to estimate the seismic hazard (Wallace, 1970; Anderson, 1979; Molnar, 1979; Field, 2007). Therefore, the slip rate is one of the most important parameters in the study of an active normal fault.

The Jilantai Basin, with the Hetao Basin to the north and the Yinchuan Basin to the south (**Figure 1**), lies in the northwestern corner of the Ordos Block. The Jilantai Basin is bounded on its western side by the normal Bayanwulashan Piedmont Fault (BPF) and on its eastern side by the

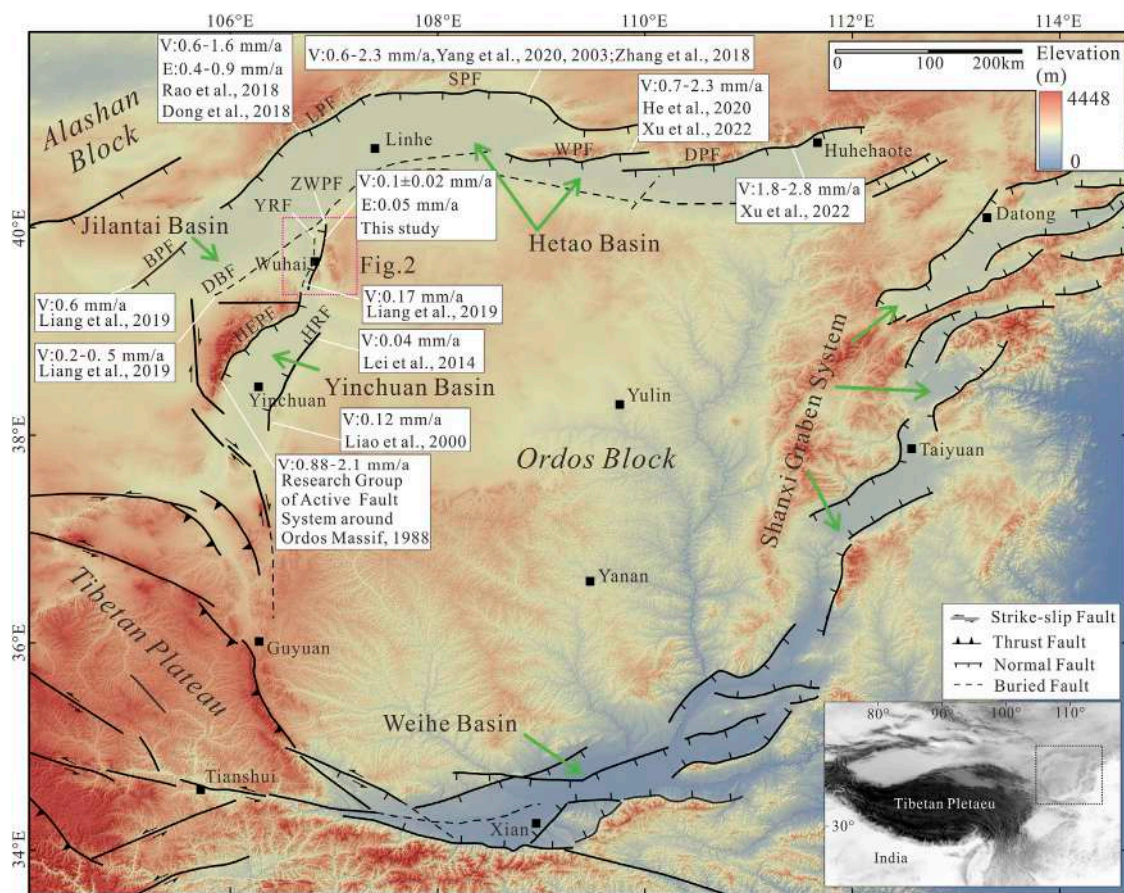


FIGURE 1 | Topographic map of the Ordos Block showing major active faults and the location of the study area. Active faults are modified from Deng (2007) and our own observations. The pink dashed rectangle shows the location of **Figure 2**. Abbreviations of names of active faults: DPF, Daqingshan Piedmont Fault; WPF, Wulashan Piedmont Fault; SPF, Seertengshan Piedmont Fault; LPF, Langshan Piedmont Fault; BPF, Bayanwulashan Piedmont Fault; DBF, Dengkou-Beijing Fault; YRF, Yellow River Fault; ZWPF, Zhuzhishan West Piedmont Fault; HEPF, Helanshan East Piedmont Fault; HRF, Huanghe River Fault. V: vertical slip rate. E: extension rate.

normal Zhuzhishan West Piedmont Fault (ZWPF) (**Figure 1**). For many years, the study of active faults has focused on those around the Hetao Basin and Yinchuan Basin, such as the Daqingshan Piedmont Fault (DPF) (Jiang et al., 2001; Ran et al., 2003; He et al., 2007; He and Ma, 2015), the Wulashan Piedmont Fault (WPF) (Chen, 2002; He et al., 2020), the Seertengshan Piedmont Fault (SPF) (Yang et al., 2002; Chen et al., 2003; Yang et al., 2003; Zhang et al., 2017; Rao et al., 2019), the Langshan Piedmont Fault (LPF) (Rao et al., 2016; Shen et al., 2016; Dong et al., 2018; Rao et al., 2018), and the Helanshan East Piedmont Fault (HEPF) (Deng and Liao, 1996; Lin et al., 2013; Lin et al., 2015). However, the number of studies on the ZWPF is relatively low (Liang et al., 2019).

To determine the slip rate of active faults, the displacements of landforms and corresponding ages of geomorphic surfaces are usually needed (Zhang et al., 2007; Hetzel, 2013; Liu et al., 2017, 2020; Liu et al., 2022). Using optically stimulated luminescence (OSL) dating and displacements of lacustrine layers from boreholes, Liang et al. (2019) reported a vertical slip rate of 0.5–1.0 mm/a for the ZWPF since approximately 70 ka. However, there is no limit on the geomorphic age in the study. In this

contribution, to estimate the Pleistocene slip rate of the ZWPF, we use structure from motion (SfM) techniques to acquire a high-resolution digital elevation model (DEM) to determine vertical displacements, and we apply OSL dating to obtain the abandonment ages of the faulted terraces. Finally, we compare our rate with other results from the northwestern Ordos Block and discuss the roles of the ZWPF in tectonic activity and slip partitioning in the Jilantai Basin.

GEOLOGICAL BACKGROUND

The Jilantai Basin between the Gobi-Alashan and Ordos Blocks is an active faulted basin approximately 140 km long from north to south and 100 km wide from east to west (**Figure 1**). The Wuhai Basin is a subsidiary basin in the eastern of the Jilantai Basin that is approximately 100 km long from north to south and 5–15 km wide from east to west and is bounded on its eastern and western sides by the ZWPF and the Wuhushan Fault (WF) (Liang et al., 2018) (**Figure 2**). Thicknesses of Quaternary deposits within the Wuhai Basin are more than 400 m; among them, late Pleistocene

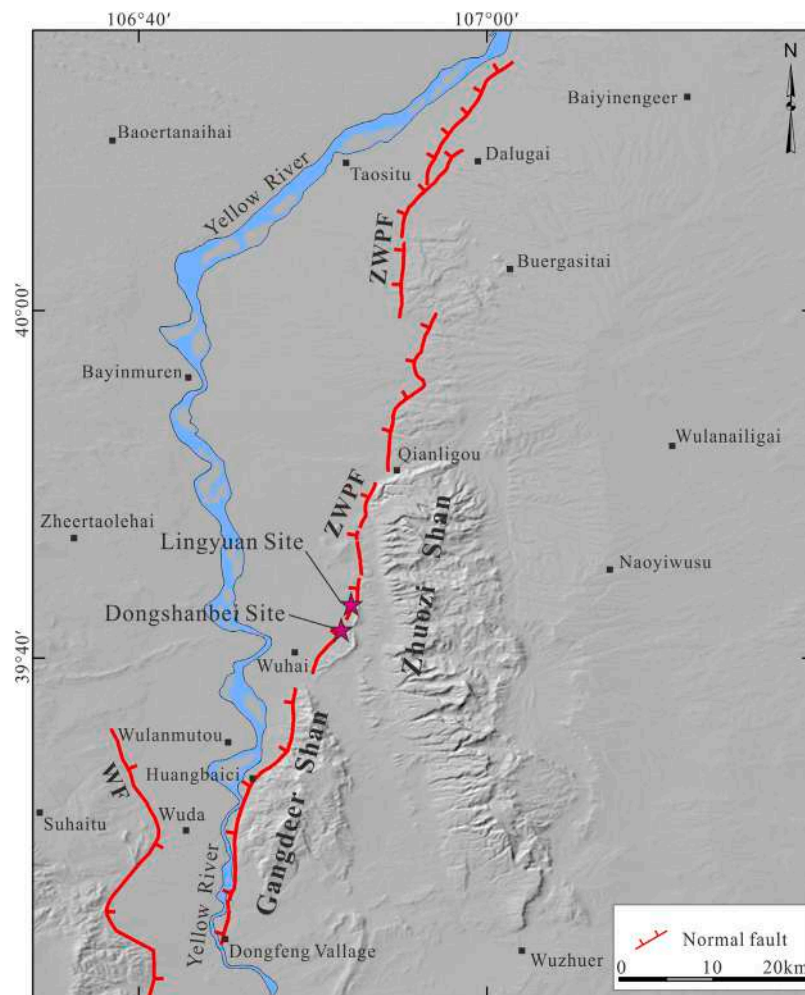


FIGURE 2 | Hill-shaded relief map showing the location of the Zhuozishan Piedmont Fault and its adjacent region. Abbreviations of names of active faults: ZWPf, Zhuozishan Piedmont Fault; WF, Wuhushan Fault.

sections are approximately 90–95 m, and the Holocene sections are approximately 10–12 m (Research Group of Active Fault System around Ordos Massif, 1998; Liang et al., 2019). The Zhuozhi Shan and Gangdeer Shan are located east of the Wuhai Basin (**Figure 2**) and are mainly composed of gneiss, limestone, and shale. The crest of the Zhuozhi Shan has an elevation of ~2,200 m, and the relief decreases from the crest to the Wuhai Basin at elevations of ~1,200 m. Divided by the active normal ZWPf, the ranges can be clearly separated into two regions with different landscape reliefs.

The ZWPf, which runs along the western front of the Zhuozhi Shan and Gangdeer Shan with a length of 90 km, forms the eastern boundary of the Wuhai Basin (**Figure 2**). The ZWPf has experienced strong structural deformation since the Cenozoic, and its movement has changed from a thrust fault to a normal fault (Xing and Wang, 1991). The process of activity has not only created the magnificent landscape of mountains and basins but also controlled the evolution of landforms and topography since the Cenozoic. Especially since the late Quaternary, the ZWPf has

faulted the piedmont alluvial fans and river terraces west of the Zhuozhi Shan, forming obvious west-facing fault scarps. These well-preserved fault scarps provide an opportunity for us to study the slip rate of the fault.

METHODS

To determine the slip rate of the ZWPf, the vertical displacement should be constrained. Through satellite image interpretations, we carried out detailed field investigations and chose the appropriate study site. Based on the SfM technique (Bemis et al., 2014; Micheletti et al., 2015; Bi et al., 2017; Ai et al., 2018), we used photographs from small unmanned aerial vehicles equipped with Real-time kinematic to generate a DEM image with a resolution of centimeter accuracy for the study site. Topographic profiles perpendicular to the fault scarps were derived from the DEM image. Then, the displacement was determined from the vertical distance between the subparallel

TABLE 1 | Result of OSL analysis for the samples on the ZWPF.

Sample	Depth (m)	U-238 (Bq/kg)	Th-232 (Bq/kg)	K-40 (Bq/kg)	Dose Rate (Gy/Ka)	Equivalent dose (Gy)	OSL Age (Ka)
ZZS-OSL-01	0.7	50.9 ± 0.65	44.8 ± 2.6	512.6 ± 5.2	3.78 ± 0.10	86.85 ± 8.12	23.0 ± 3.2
ZZS-OSL-04	0.3	39.8 ± 2.05	40.2 ± 1.8	504.2 ± 8.2	3.45 ± 0.10	72.31 ± 4.59	20.9 ± 2.5
ZZS-OSL-06	0.6	45.61 ± 0.5	47.6 ± 2.6	528.2 ± 9.5	3.77 ± 0.10	95.73 ± 5.09	25.4 ± 2.9
ZZS-OSL-07	1.7	35.91 ± 4.8	36.3 ± 5.4	500.2 ± 2.5	2.99 ± 0.10	384.05 ± 23.8	128.3 ± 15.4
ZZS-OSL-08	1.5	40.55 ± 1.6	45.8 ± 2.6	564.3 ± 3.1	3.66 ± 0.10	41.80 ± 1.71	11.4 ± 1.3
ZZS-OSL-54	0.5	15 ± 1.37	27 ± 1.62	503 ± 9.51	2.88 ± 0.12	57.09 ± 1.00	19.8 ± 0.9
ZZS-OSL-55	1.6	19 ± 1.87	30 ± 1.70	566 ± 9.51	3.23 ± 0.13	41.79 ± 1.69	12.9 ± 0.7
ZZS-OSL-56	2.0	21 ± 0.75	23 ± 1.10	554 ± 9.51	2.73 ± 0.11	146.96 ± 2.91	53.9 ± 2.5
ZZS-OSL-58	0.7	20 ± 1.12	35 ± 1.58	494 ± 6.35	3.17 ± 0.13	140.43 ± 2.60	44.3 ± 1.9
ZZS-OSL-59	0.8	15 ± 3	24 ± 3	528 ± 10	3.02 ± 0.14	207.27 ± 5.53	68.7 ± 3.7
ZZS-OSL-60	0.3	11 ± 1	34 ± 3	509 ± 10	3.14 ± 0.14	72.53 ± 1.32	23.1 ± 1.1
ZZS-OSL-61	0.9	14 ± 2	31 ± 2	535 ± 10	3.16 ± 0.14	30.51 ± 0.64	9.7 ± 0.5

tread lines by fitting straight lines to both the hanging wall and footwall. At the study sites, fault outcrops are also well exposed due to stream incision and indicate that fluvial or alluvial strata have been offset by the fault. Characteristics of stratigraphic sedimentary structures and ages were used to define the activity of the fault.

OSL dating was used to constrain the age of corresponding faulted landform surfaces. Samples collected above fluvial or alluvial gravels can offer a minimum age for the time of river incision or abandonment of the landform surfaces (Pan et al., 2013; Hu et al., 2017). We hammer into the fine grain deposition with a 20 cm-long stainless-steel cube from fresh sediment sections. In the process of sample collection and packaging, we try to avoid the influence of sunlight on samples. The dating was undertaken at the Laboratory of Neotectonic Geochronology of Institute of Disaster Prevention, China Earthquake Administration, by standard procedures for Chinese loess (Forman 1991; Wang, 2006). Organic material and carbonates were removed using 30% H₂O₂ and 37% HCl under subdued red light. 4–11 μm fine-grain quartz fraction and 180–250 μm coarse-grain quartz fraction were separated after neutralization with ammonia and washing with purified water (Aitken, 1998). Then, 30% H₂SiF₆ etching was used repeatedly to remove feldspar and isolate quartz in 4–11 μm fine-grain quartz fractions. The equivalent doses of the fine-grained samples were determined using the single-aliquot regenerative-dose (SAR) dating protocol (Wang, 2006). The determination of environmental dose rate also follows the standard rules (Aitken 1998; Wang et al., 2005). The dating results are listed in Table 1.

RESULTS

Based on the field investigation, we select two sites (Lingyuan site and Dongshanbei site) to constrain the vertical slip rate of the ZWPF (Figure 2). Fault scarps of different heights are well preserved in these two sites, and we can obtain the fault slip rate in different times.

Lingyuan Site

The Lingyuan site is located in the middle segment of the ZWPF (Figure 2). We constructed high-resolution digital topography at

this site by using the SfM technique. Then, a DEM with a resolution of 0.2 m and an associated hill-shaded relief map was acquired, which permitted us to perform a detailed analysis of tectonic landforms (Figure 3A). As revealed by the 0.5 m-interval contour lines, three levels of river terraces (T1–T3) can be identified based on the surface morphologies and different heights above the channel infill (Figure 3B). Aside from the current channel and T1 terrace, fault scarps trending approximately south-north are prominent on the hill-shaded relief map (Figure 3A). Moreover, an outcropping fault indicative of active deformation was also found. Field observations (Figure 4A) and the corresponding sketches (Figure 4B) indicate that a fault offsets a light-yellow fine sand layer (Unit 1), light red sandy gravel layer (Unit 2), and silty sand unit 3. The greyish-yellow loess (Unit 4) is preserved in the hanging wall. Then, the fault and sedimentary units were covered with cyan grey surface soil (Unit 5). We interpret Unit 4 as a scarp-derived deposit that usually accumulates rapidly after the creation of a fault scarp. An inverted triangular filling wedge is formed at the bottom of Unit 4. Thus, the ages of Unit 4 and Unit 3 can give limits on the timing of the most recent earthquake event. We collected and dated two OSL samples (ZZS-OSL-03 and ZZS-OSL-54) at the base of Unit 4 and Unit 3. The results show ages of 20.9 ± 2.5 ka and 19.8 ± 0.9 ka, respectively (Table 1). Accordingly, we think the most recent earthquake event at this locality probably occurred close to ~20 ka.

In the field, fault scarps on the T2 and T3 terraces can be clearly observed (Figure 5A,B). Then, we constructed two topographic profiles across the fault scarps on the DEM to constrain their heights on different terraces (Figure 3B). The displacement can be derived from the vertical distance between the subparallel tread lines on two fitting straight lines above and below the scarp. The heights of the fault scarp on the T2 and T3 terraces are determined to be 2.5 ± 0.2 m and 4.6 ± 0.3 m, respectively (Figure 5C). The vertical displacements increase with the terrace level which indicates an accumulation of fault slip.

To determine the ages of the terraces at the Lingyuan site, we collected three OSL samples from the T1, T2, and T3 terraces (Figure 3B). For the T1 terrace, we collected one sample from the fluvial deposit of fine-grained sand between the gravels at depths of 0.8 m, which has an age of 9.7 ± 0.5 ka (Figure 5D). For the T2

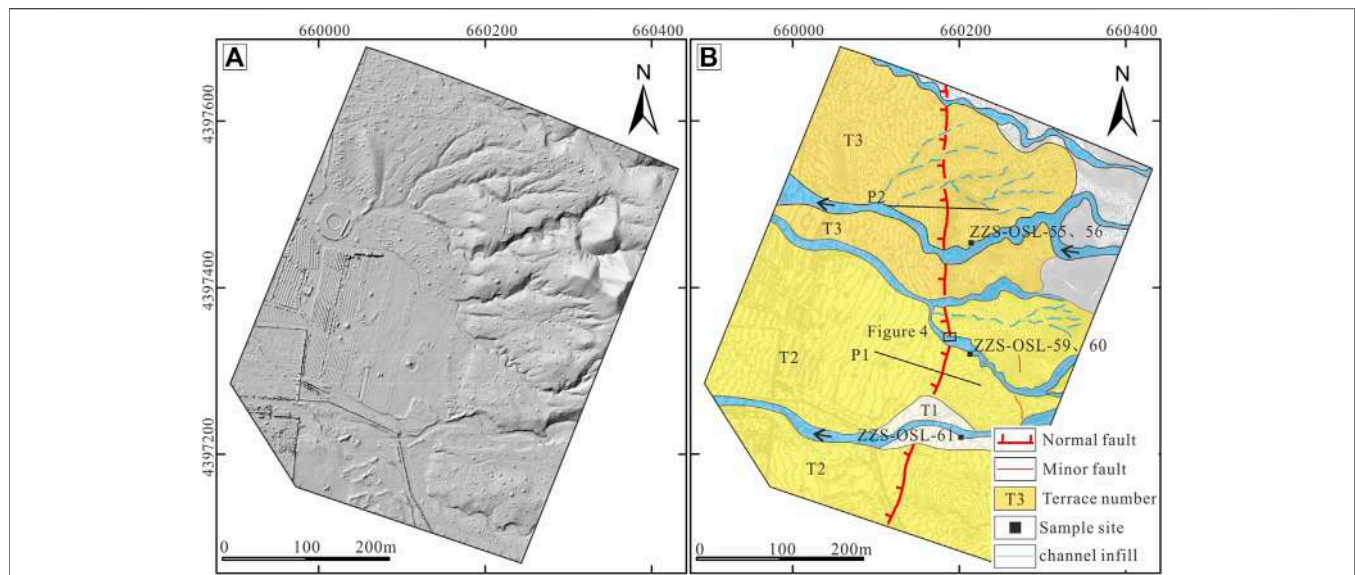


FIGURE 3 | (A) Hill-shaded relief map from high-resolution DEM data illustrating the distribution of fault scarps at the Lingyuan site **(B)** Interpreted geomorphic map with 0.5 m interval contours showing the fluvial terraces, normal fault scarps and sampling positions at the Lingyuan site. Three levels of river terraces are mapped according to their morphologic features and different heights above the channel infill.

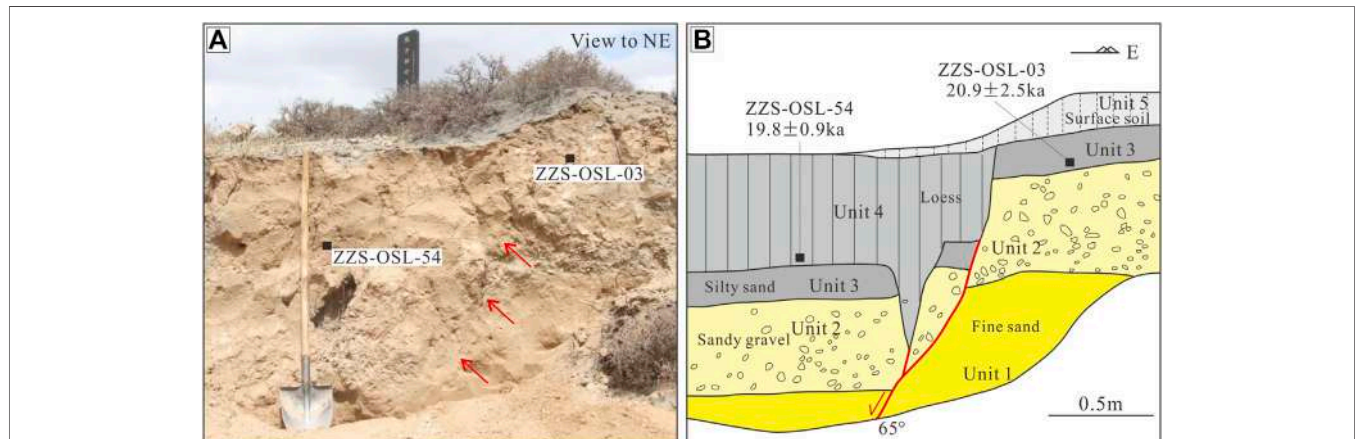


FIGURE 4 | (A) Fault outcrop underneath the scarp. The location of the fault plane is marked with red arrows **(B)** the corresponding sketch at the Lingyuan site. One major normal fault is prominent, displacing sedimentary Units 1–4. Two OSL samples (ZZS-OSL-03 and ZZS-OSL-54) collected from the faulted and covering units were dated, yielding ages of 20.9 ± 2.5 ka and 19.8 ± 0.9 ka, respectively (**Table 1**).

terrace, two samples were collected at depths of 0.3 and 0.8 m, which are dated at 68.7 ± 3.7 ka and 23.1 ± 1.1 ka, respectively (**Figure 5E**). We think the upper sample (ZZS-OSL-59) may be mixed with underexposure material, which leads to older age. Comparing these dates with the ages of the T1 and T3 terraces, we consider 23.1 ± 1.1 ka to better represent the age of the terrace. We also collected two samples for the T3 terrace at depths of 1.6 and 2.0 m, yielding ages of 12.9 ± 0.7 ka and 53.9 ± 2.5 ka, respectively (**Figure 5F**). Considering the ages of the T1 and T2 terraces, we use 53.9 ± 2.5 ka as the age of the T3 terrace. Finally, we determine ages of 9.7 ± 0.5 ka, 23.1 ± 1.1 ka, and 53.9 ± 2.5 ka for terraces T1, T2, and T3 at the Lingyuan site, respectively.

Dongshanbei Site

The Dongshanbei site is located in eastern Wuhai city (**Figure 2**). Several nearly east-west-trending ephemeral streams rising in the Zhuozi Shan have incised into the alluvial fans and formed alluvial terraces with different heights at the mountain pass (**Figure 6A**). Four levels of alluvial terraces are identified by increasing numbers from youngest to oldest (**Figure 6B**). The terrace faces are roughly flat and dip to the west. The heights above the riverbed are approximately 2, 4.5, 9, and 15 m from the T1 to T4 terraces, respectively. The ZWPF offsets the T2, T3, and T4 terraces and forms distinct scarps facing northwest (**Figure 6B**). In the

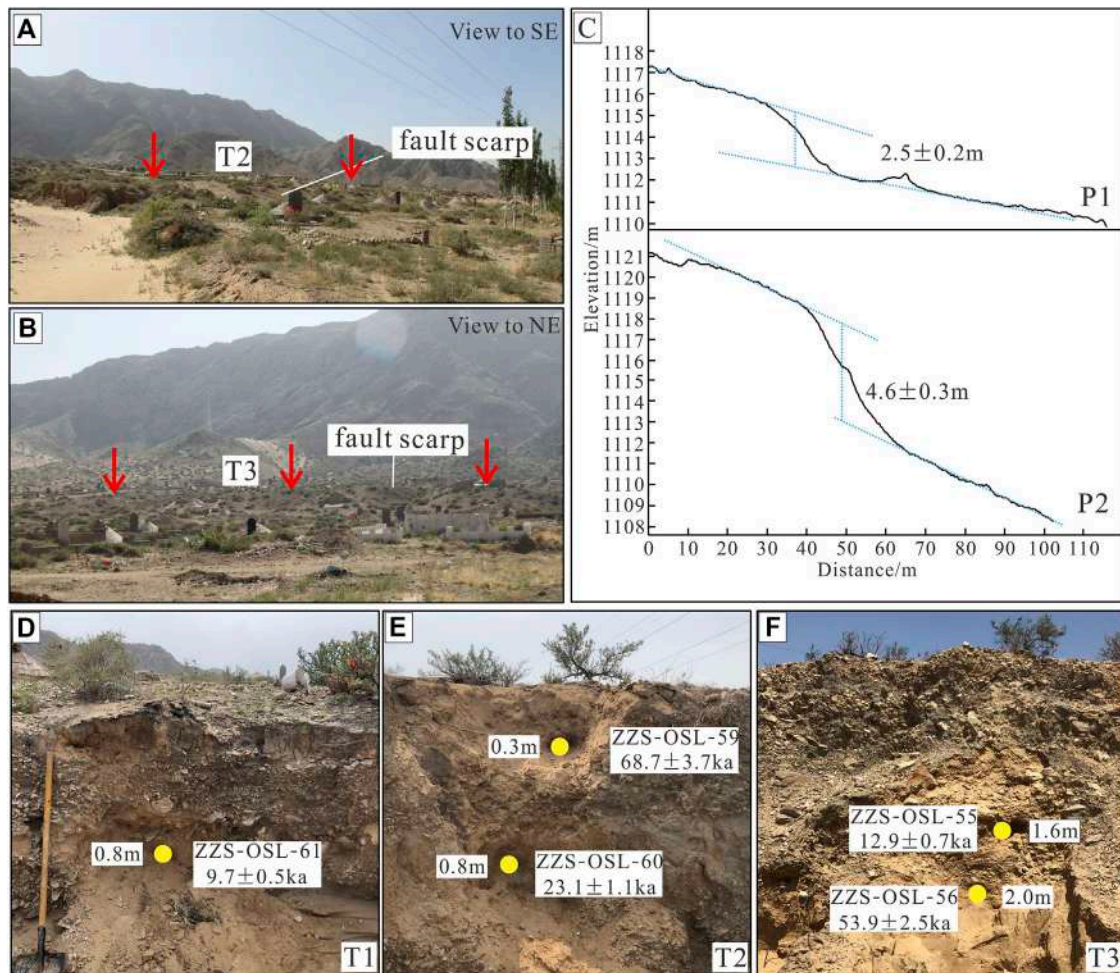


FIGURE 5 | (A) (B) Fault scarps displacing different levels of terraces observed in the field (C). Topographic profiles across the fault scarps on the T2 and T3 terrace surfaces (D) (E), and (F). OSL sample profile and ages for T1, T2, and T3 terraces.

cliff along one stream, we find one outcrop of the fault, showing a normal component fault plane with a dip of approximately 75° (Figure 7). The fault displaces late Pleistocene fluvial gravels (Unit 1). Then, sandy gravel (Unit 2) has accumulated in the footwall and is overlain by loess (Unit 3). We interpret Unit 4 as a colluvial wedge that is usually accumulated rapidly after fault activity. However, it is usually difficult to obtain a suitable sample in a colluvial wedge. The age of Unit 3 can give the reference time of the most recent earthquake event. We collected and dated one OSL sample (ZZS-OSL-08) at the base of Unit 3. The results show an age of 11.4 ± 1.3 ka (Table 1). Accordingly, we think the most recent earthquake event at this locality probably occurred before 11.4 ± 1.3 ka. This may be the same as the latest earthquake event revealed at the Lingyuan site.

Despite the impact of erosion and human activities, fault scarps can still be distinguished on different terraces (Figure 8A–C). The geomorphic surface of the footwall is intact, and the hanging wall is slightly damaged. To constrain

the height of the fault scarps, we carried out topographic surveys across the fault scarp with differential GPS. Three topographic profiles (P1 to P3) were obtained (Figure 6). By fitting straight lines to both the hanging wall and footwall, we measured the heights from P1 to P3 as 2.1 ± 0.2 m, 3.3 ± 0.1 m, and 5.4 ± 1.0 m, respectively (Figure 8D).

We also collected three OSL samples from the T2, T3, and T4 terraces (Figure 6B). For the T1 terrace, the sample was collected from the fluvial deposit of fine-grained sand between the gravels at a depth of 0.6 m, with an age of 25.4 ± 2.9 ka (Figure 8E). For the T2 terrace, the sample was collected at a depth of 1.7 m and was dated at 128.3 ± 15.4 ka (Figure 8F). We also collected one sample for the T4 terrace at 0.7 m depth, yielding an age of 44.3 ± 1.9 ka (Figure 8G). Comparing these dates with the ages of the T3 and T4 terraces, we believe that the date of 128.3 ± 15.4 ka cannot represent the age of the terrace. This may represent early deposition or an underexposed part of the terrace. Finally, we determine ages of 25.4 ± 2.9 ka and 44.3 ± 1.9 ka for terraces T2 and T4 at the Dongshanbei site.

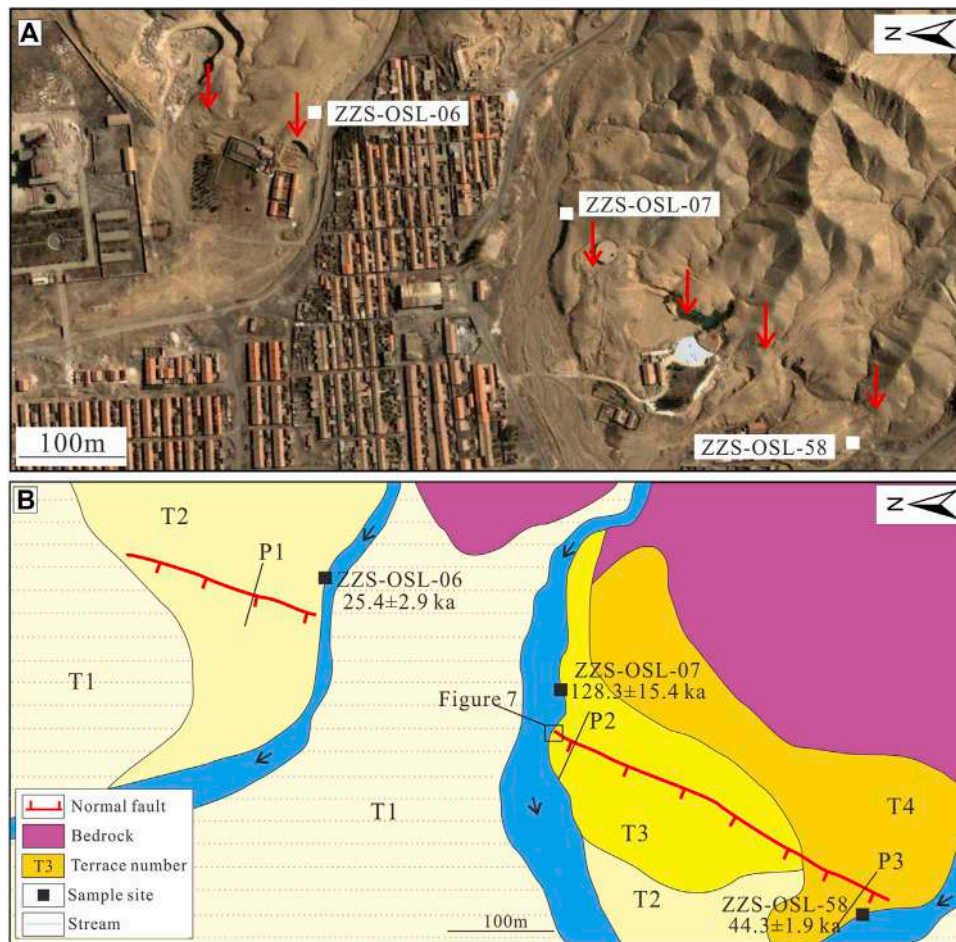


FIGURE 6 | (A) Satellite image from Google Earth. The location of fault scarps is indicated by red arrows **(B)**. Interpreted geomorphic map of the distribution of fluvial terraces, normal fault scarps and sampling positions at the Dongshanbei site. Four levels of river terraces are mapped according to their morphologic features and different heights above the channel infill.

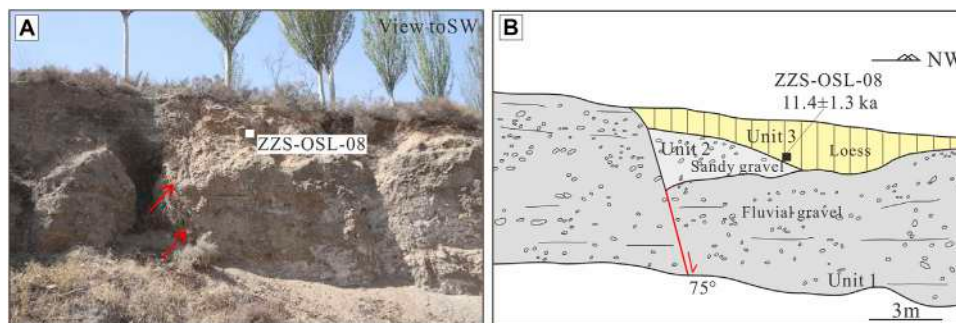


FIGURE 7 | Fault outcrop **(A)** and corresponding sketch **(B)** at the Dongshanbei site. Red arrows indicate the location of fault plane. One major normal fault is prominent, displacing sedimentary Unit 1. One OSL sample (ZZS-OSL-08) collected above the inferred colluvial wedge was dated, yielding an age of 11.4 ± 1.3 ka (Table 1).

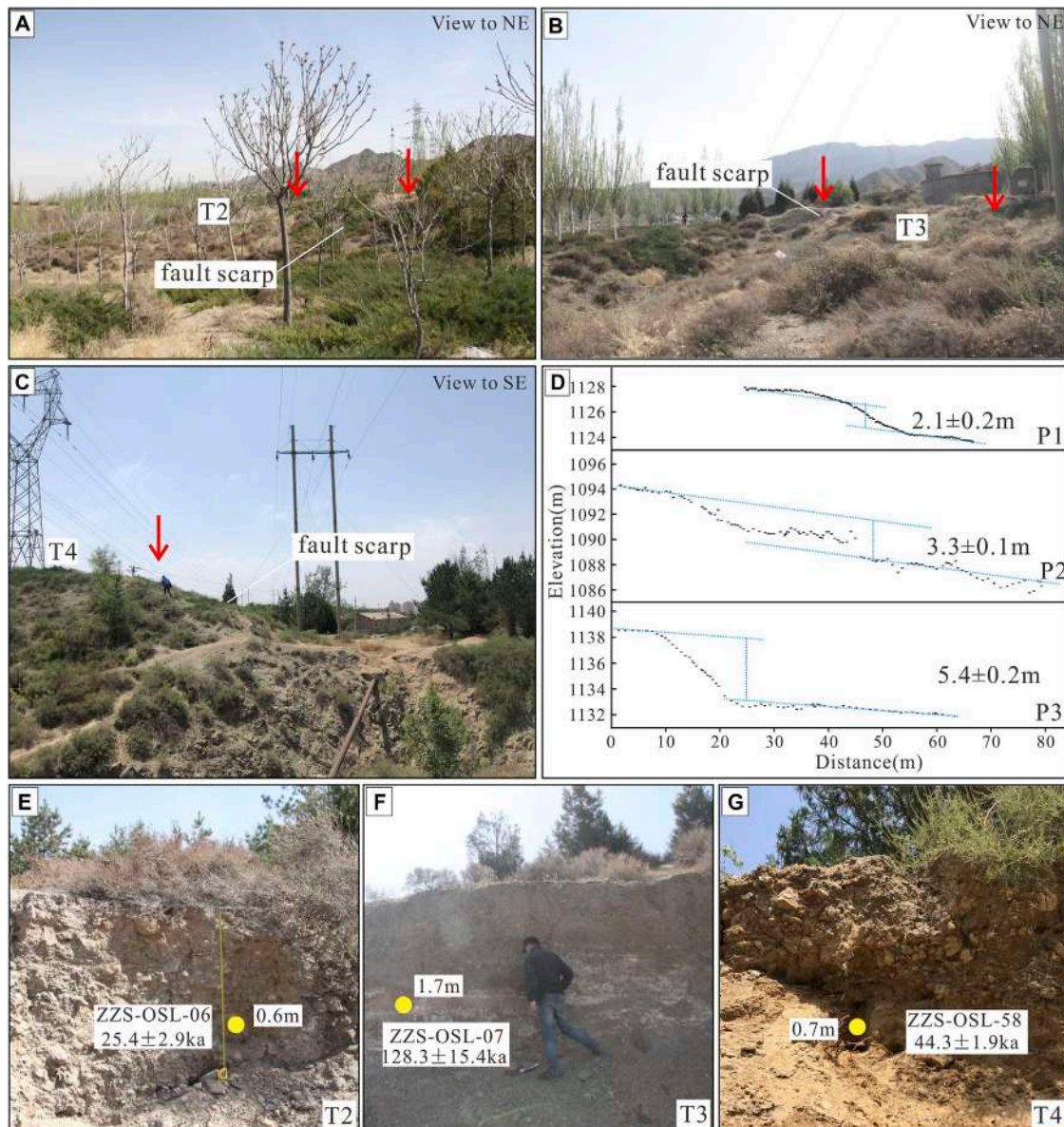


FIGURE 8 | (A) (B), and (C). Fault scarps displacing different levels of terraces observed in the field (D). Topographic profiles across the fault scarps on the T2, T3 and T4 terrace surfaces (E) (F), and (G). OSL sample profiles and ages for T2, T3, and T4 terraces.

Determination of the Slip Rate of the ZWPF

Based on the vertical offsets of the terraces and their corresponding abandonment ages, we determine the vertical component of the slip rate. The uncertainty comes mainly from uncertainties in the ages and the vertical offsets. At the Lingyuan site, an average slip rate of 0.1 ± 0.02 mm/a is determined using vertical displacements (2.5 ± 0.2 m and 4.6 ± 0.3 m) and corresponding ages (23.1 ± 1.1 ka and 53.9 ± 2.5 ka) (Figure 9A). At the Dongshanbei site, we derive an average slip rate of 0.1 ± 0.02 mm/a by using vertical displacements (2.1 ± 0.2 m and 5.4 ± 0.2 m) and corresponding ages (25.4 ± 2.9 ka and 44.3 ± 1.9 ka) (Figure 9B). The two research points are close to each other and show the same vertical slip rate. This also reflects that our results are reliable. Trenches and natural outcrops reveal that the

range of fault dips is $\sim 50^\circ$ – 80° to the west (Zhang, 2014; Liang et al., 2019; Gao, 2020). With an average dip of $\sim 65^\circ$ and a vertical slip rate of 0.1 ± 0.02 mm/a, the extension rate is estimated to be ~ 0.05 mm/a.

DISCUSSION

Comparison With Other Active Faults in the Close Region

The active faults in the northwestern Ordos Block are in a similar tectonic stress field. Therefore, the slip rate of faults should be comparable or regular. There are three basins in the northwestern Ordos Block, the Yinchuan Basin, the Jilantai Basin, and the

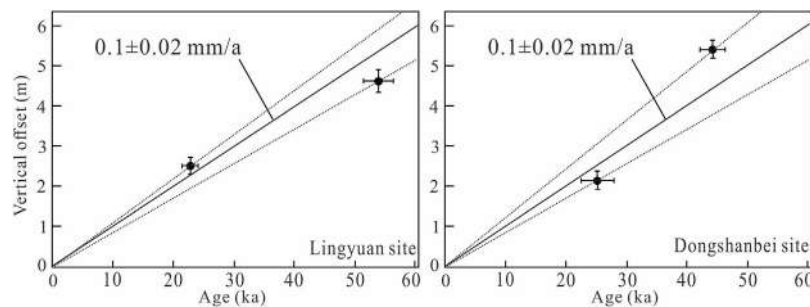


FIGURE 9 | Average vertical slip rates on the ZWPF, which are estimated from vertical offsets recorded by terraces T2 and T3 at the Lingyuan site and T2 and T4 at the Dongshanbei site and their corresponding abandonment ages.

Hetao Basin (**Figure 1**). These basins are mainly controlled by active faults on both sides.

The Yinchuan Basin is bounded on its western side by the normal HEPF and on its eastern side by the normal YRF. For the HEPF, the vertical slip rate has a range of 0.88–2.1 mm/a (Research Group of Active Fault System around Ordos Massif, 1988). The southern segment of the YRF faults surface landform and forms fault scarps. Based on fault scarp measurements and thermoluminescence dating, the vertical slip rate was determined to be 0.12 mm/a (Liao et al., 2000). The northern segment of the YRF is buried. According to borehole displacement measurements and OSL dating, an average vertical slip rate of 0.04 mm/a is obtained (Lei et al., 2014). The normal Bayanwulashan Piedmont Fault (BPF) is the western boundary of the Jilantai Basin. Its vertical slip rate is estimated to be approximately 0.6 mm/a since the late Pleistocene (Liang et al., 2019). At the eastern boundary of the Jilantai Basin and to the west of ZWPF, there is also a buried fault called the Yellow River Fault (YRF). Liang et al. (2018) obtained a vertical slip rate of 0.17 mm/a for the fault by using drill cores and Quaternary dating methods.

The Langshan Piedmont Fault (LPF), Seertengshan Piedmont Fault (SPF), Wulashan Piedmont Fault (WPF), and the Daqingshan Piedmont Fault (DPF) are the western and northern boundary faults of the Hetao Basin. The LPF is located on the northwestern margin of the Hetao Basin. Previous studies have suggested that the vertical slip rate of the LPF is ~0.5–2.2 mm/a (Research Group of Active Fault System around Ordos Massif, 1988; Deng et al., 1999). According to the coseismic displacement and recurrence interval, the vertical slip rate of the LPF is limited to ~0.6–1.6 mm/a (Dong et al., 2018; Rao et al., 2018). For the SPF, many results suggested that the vertical slip rate of the fault is ~0.6–2.3 mm/a (Yang et al., 2002, 2003; Zhang et al., 2017). He et al. (2020) and Xu et al. (2022) reported slip rates of ~0.7–2.3 mm/a for the WPF since the late Pleistocene. For the DPF, the latest results show that the Holocene vertical slip rate of the fault is ~1.8–2.8 mm/a (Xu et al., 2022). However, the activity of the faults on the southern boundary of the Hetao Basin is low, and quantitative research is lacking.

From the comparison of fault slip rates in the northwestern Ordos Block, we found that the slip rates of active faults on the

northern and western boundaries of the basins are high, with an average of more than 1 mm/a, and the values in the southern and eastern boundaries of the basins are low and maybe less than 0.2 mm/a. We infer that this may be related to the subsurface structural architecture. Asymmetric half-graben structures have been formed in these basins, and the subsidence centers are close to the northern and western sides (Research Group of Active Fault System around Ordos Massif, 1988; Lei, 2016; Liang et al., 2018; Rao et al., 2018).

Implications for slip partitioning in the northwestern corner of Ordos Block

The Ordos Block is characterized by counter-clockwise rotation because of the expansion of the Tibetan Plateau and the subduction of the Pacific Plate (Zhang et al., 2002, 2005; Liang et al., 2018). The eastern and western sides of the Ordos Block are the Yinchuan-Jilantai Basin and Shanxi Graben System controlled by dextral normal faults. The southern and northern sides of the Ordos Block are the Weihe Basin and Hetao Basin controlled by sinistral normal faults (Zhang et al., 2002, 2005). The Jilantai Basin and the western part of the Hetao Basin lie in the northwestern corner of the Ordos Block and should be characterized by both extensional and dextral movement. The latest GPS data also reflect these features (Zhao et al., 2017). The extension and dextral slip rates across the northwestern corner of the Ordos Block in the NW direction were deduced to be 1.6 ± 0.4 mm/a and 0.8 ± 0.7 mm/a from GPS velocities, respectively (Zhao et al., 2017). These rates are distributed on a series of active faults across the region.

The extension rate associated with active normal faulting can be assessed in combination with near-surface exposed fault dips (Zhang et al., 1998; Personius et al., 2017). The GPS profile crosses mainly the LPF, the ZWPF, and the Yellow River Fault. The vertical slip rate of the LPF is determined to be ~0.6–1.6 mm/a (Dong et al., 2018; Rao et al., 2018), with a dip of ~60° (Research Group of Active Fault System around Ordos Massif 1988; Rao et al., 2018) and the extension rate is estimated to be ~0.4–0.9 mm/a. In this study, the vertical slip rate of the ZWPF is determined to be 0.1 ± 0.02 mm/a. In the eastern part of the basin, the vertical slip rate of the buried Yellow River fault was determined to be 0.17 mm/a (Liang et al.,

2018). For another buried fault within the basin, namely, the Dengkou-Benjing Fault (DBF), the vertical slip rate is inferred to be 0.2–0.5 mm/a (Liang et al., 2019). The total vertical slip rate in the eastern part of the basin would be ~0.5–0.8 mm/a. If these faults have dips of ~65° similar to the ZWPF, the extension rate would be 0.2–0.4 mm/a. Therefore, the total extension rate across the northwestern corner of the Ordos Block would be ~0.6–1.3 mm/a. This value seems lower than the GPS data (Zhao et al., 2017). If we consider a decrease in fault dip, for example, fault planes flattening with depth and developing listric geometry, the results would become comparable with the GPS data. Comparing the extension rates, we infer that the main extensional deformation occurs on the western side of the basin.

Meanwhile, GPS data also reflect dextral shearing across the northwestern corner of the Ordos Block (Zhao et al., 2017). Unfortunately, we did not find evidence of dextral slip along the ZWPF. Similarly, the LPF on the western side of the basin was demonstrated to show almost pure normal slip (Dong et al., 2018; Rao et al., 2018). The active faults in the northern Ordos Block also do not display the characteristics of sinistral slip reflected by GPS (Zhang et al., 2017; He et al., 2019; Rao et al., 2019; He et al., 2020; Rao et al., 2020). According to the focal mechanism solution of two earthquakes, Rao et al. (2018) inferred that dextral slip movements mostly occur on buried faults within the basin. Normal-faulting earthquakes are mainly concentrated on the margin of the basin (Wen, 2014; Liang et al., 2019). The same situation can also be found in the Yinchuan Basin (Middleton et al., 2016; Liang et al., 2019). If this is a common phenomenon, the result would help us to better understand the slip characteristics of active faults in other basins around the Ordos Block.

CONCLUSION

The ZWPF, along the western margin of the Zhuozhi Shan, forms the eastern boundary of the Jilantai Basin, has grown by normal faulting, and was active during the Pleistocene. The latest earthquake event was dated to ~20 ka. Satellite image interpretations and field investigations show that the fault offsets late Quaternary terraces and alluvial fans and forms permanent west-facing fault scarps. Based on OSL dating, we

obtain the abandonment ages of the faulted terraces. Combining the ages and fault scarp heights obtained from the high-resolution DEM and differential GPS measurements, the vertical slip rate of the ZWPF is determined to be 0.1 ± 0.02 mm/a since the Pleistocene, which suggests an extension rate of 0.05 mm/a with a dip of ~65°. Combining these values with other results, we infer that the NW-directed extension rate across the Jilantai Basin will be ~0.6–1.3 mm/a. The extensional deformation mainly occurred on the western side of the basin.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

XL carried out field work, drew the figures, and wrote the manuscript. GZ, YS, and YY performed the fieldwork, sampling, aerial photogrammetry, and differential GPS measurements. All authors discussed the results and contributed to the manuscript.

FUNDING

This work was supported by the National Key Research and Development Program of China (2017YFC1500100) and the Innovation Team Project of Earthquake Administration of Gansu Province (2019TD0101).

ACKNOWLEDGMENTS

We are grateful to Daoyang Yuan, Wenjun Zheng, Qiyun Lei, and Chuanyong Wu for their discussions and suggestions in the field regarding our preliminary results. We greatly thank the reviewers and editors for their constructive and detailed comments to improve the manuscript.

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Constraining Paleoseismicity of the Wulashan Piedmont Fault on the Northern Margin of the Ordos Block From Fault Scarp Morphology

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OPEN ACCESS

Edited by:

Daoyang Yuan,
School of Earth Sciences, Lanzhou
University, China

Reviewed by:

Junjie Ren,
China Earthquake Administration,
China
Xuhua Shi,
Zhejiang University, China

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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 02 April 2022

Accepted: 25 May 2022

Published: 14 June 2022

Citation:

Bi H, Shi L, Zhang D, Peng H, Yang J,
You Z and Sun X (2022) Constraining
Paleoseismicity of the Wulashan
Piedmont Fault on the Northern Margin
of the Ordos Block From Fault
Scarp Morphology.
Front. Earth Sci. 10:911173.
doi: 10.3389/feart.2022.911173

Fault scarps preserve important information about past earthquakes on a fault, and thus can be applied to investigate the fault slip histories and rupture patterns. In this study, the morphology of fault scarps was used to constrain the paleoseismicity of the Wulashan Piedmont Fault located on the northern margin of the Ordos Block based on high-resolution LiDAR topography. We constructed the vertical displacement distribution of the fault through measuring the heights of a large number of scarp profiles extracted on different geomorphic surfaces along the fault. Through statistical analysis of the dense collection of vertical displacement dataset, a total of seven paleoseismic events were identified which followed a characteristic slip pattern with an average slip of ~1.0 m. We further detected slope breaks in the fault scarp morphology to quantify the number of paleoearthquakes that occurred on the scarps, and discriminated at least five individual surface-breaking events. Both the number and slip of paleoearthquakes recognized from the morphology of fault scarps were in good agreement with previous paleoseismic trenching records. Based on the empirical scaling relationship between moment magnitude and rupture parameters, a moment magnitude of M_w 6.7–7.5 was determined for the paleoearthquakes occurred on the fault. With the fault slip rate derived by previous studies, we estimated an average recurrence interval of 1.3–1.8 kyr for the paleoseismic events, which is very close to the elapsed time since the most recent earthquake, indicating a high potential seismic hazard on the Wulashan Piedmont Fault.

Keywords: fault scarp, vertical displacement, scarp morphology, paleoseismicity, the wulashan piedmont fault

1 INTRODUCTION

Investigating the slip histories and rupture patterns of past earthquakes helps to estimate the size and timing of future earthquakes on the fault, which is of great significance for seismic hazard assessment (Zielke et al., 2015). Historical documents or instrumental datasets provide only short and incomplete records of past earthquakes (McCalpin, 2009), and many devastating earthquakes have been demonstrated to occur on faults with no records of large historical events (Fu et al., 2004; Kaneda et al., 2008; Liu-Zeng et al., 2009). Paleoseismic trenching can provide detailed

information about the time and magnitude of paleoearthquakes, but is greatly limited by geomorphic conditions and few selected excavation sites (Ewiak et al., 2015; Hodge et al., 2020). The occurrence of surface-rupturing earthquakes results in a variety of geomorphic features along the fault, such as fault scarps, and the offsets of rivers, ridges, and terrace risers (Wallace, 1968; Sieh, 1978; Avouac and Peltzer, 1993; Burbank and Anderson, 2009; Scharer et al., 2014; Xue et al., 2021). These geomorphic records can offer us important insights into the rupture histories of past events, including their rupture extent as well as surface slip distribution along the fault strike (McGill and Sieh, 1991; Zielke et al., 2010; Klinger et al., 2011; Zielke et al., 2012; Manighetti et al., 2015; Bi et al., 2018; Wei et al., 2019; Bi et al., 2020; Wei et al., 2021).

Among the geomorphic records of surface-rupturing earthquakes, the fault scarps have been demonstrated to preserve valuable information about past earthquakes (Wallace, 1977; Avouac and Peltzer, 1993; McCaig, 2009). By measuring the heights of fault scarps, we can constrain the vertical displacement distribution of the faulting events (Middleton et al., 2016; Bi et al., 2018; Johnson et al., 2018; Wei et al., 2019; Bi et al., 2020; Pucci et al., 2021; Wei et al., 2021). Moreover, the shape of a scarp can be applied to estimate the age of a single-event scarp based on the scarp degradation model (Wallace, 1977; Bucknam and Anderson, 1979; Nash, 1980; Andrews and Hanks, 1985; Arrowsmith et al., 1998; Kokkalas and Koukouvelas, 2005), and the gradient changes (i.e., slope breaks) in the morphology of a scarp may indicate the number of surface-rupturing events on a fault (Wallace, 1980; Avouac and Peltzer, 1993; Carretier et al., 2002; Ewiak et al., 2015; Wei et al., 2019; Rao et al., 2020; Wei et al., 2021). To analyze the morphological characteristics of the fault scarps, topographic profiles of the scarps are particularly useful. In earlier studies, the topographic profiles were mainly derived through field mapping, which is labor-intensive and time-consuming, thus greatly limiting the number and distribution of the topographic profiles (Avouac and Peltzer, 1993; Carretier et al., 2002; Ewiak et al., 2015). In recent years, high-resolution topographic datasets can be acquired by the Light Detection and Ranging (LiDAR) technology (Hudnut et al., 2002; Cunningham et al., 2006; Arrowsmith and Zielke, 2009; Zielke et al., 2010, 2012). Due to the ability to penetrate vegetation, the LiDAR technology can generate high-resolution “bare-earth” topography of the fault zone, which enables us to analyze and measure the morphological characteristics of the fault scarps in great detail (Johnson et al., 2018; Wei et al., 2019, 2021).

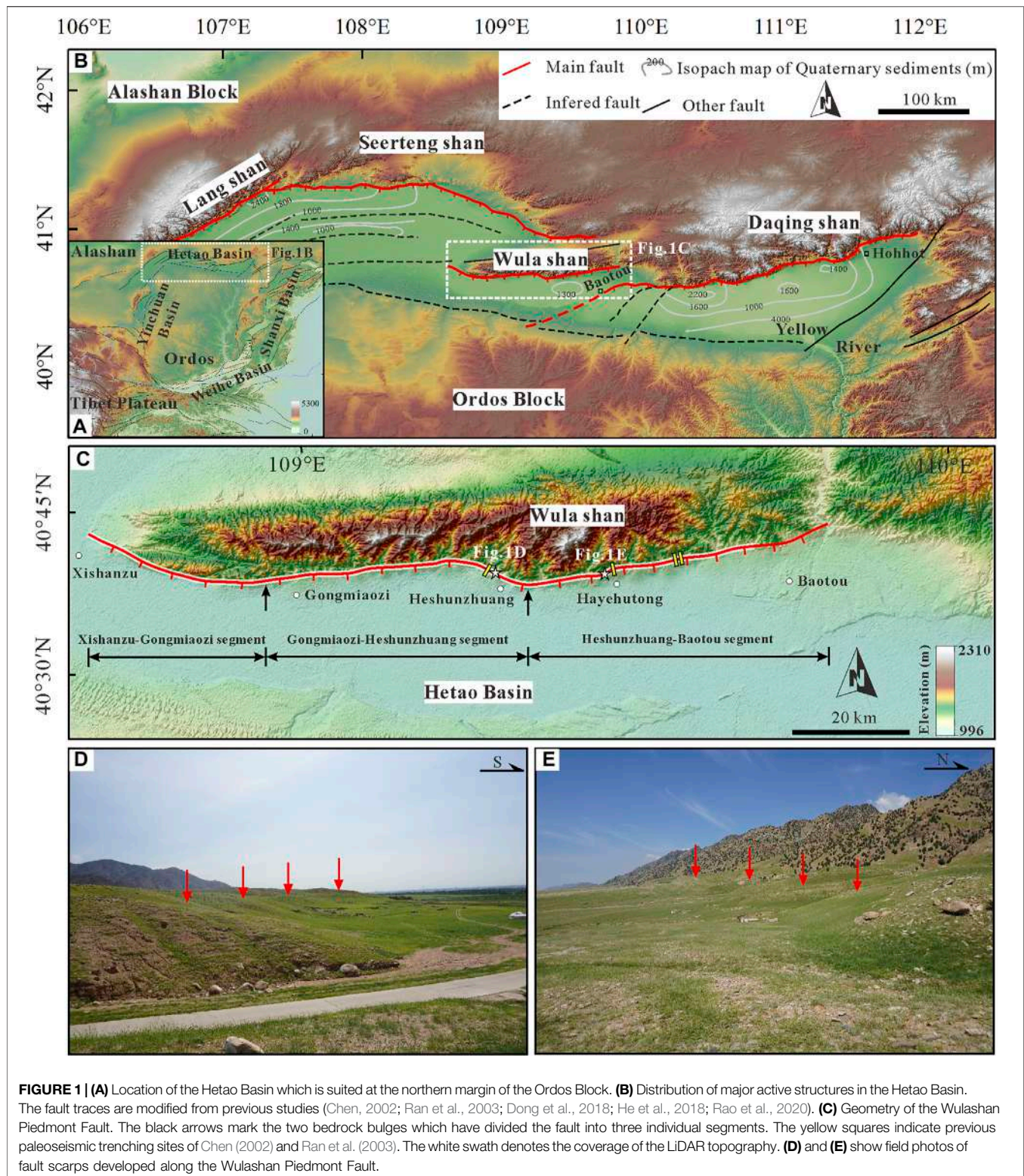
The Hetao Basin is located at the northern margin of the Ordos Block in North China (Ma and Wu, 1987; Zhang et al., 1998). The basin is controlled by a series of active normal faults bounding the northern margin of the basin along the piedmonts of the Lang Shan (Shan means mountain in Chinese), Serteng Shan, Wula Shan, and Daqing Shan from west to east, respectively (Chen, 2002; Ran et al., 2002, 2003; Yang et al., 2002, 2003; Chen et al., 2003; Rao et al., 2016, 2020; Dong et al., 2018; He et al., 2018). Among them, the Wulashan Piedmont Fault is a major active normal fault on the northern margin of the Hetao Basin. Four trenches have been excavated at different sites on the fault to

investigate its rupturing histories (Chen, 2002; Ran et al., 2003). However, limited by the low accuracy of early dating methods and the sparse trenching sites, the paleoseismic results may have large uncertainties. Thus, it is necessary to further investigate the earthquake rupturing histories on this fault to assess the potential seismic hazard in the Hetao Basin area. Long-term activities of the normal faults have produced fault scarps with different heights on the northern region of the Hetao Basin. Since being adjacent to the Wulan Buh and Kubuqi deserts, this region has a semi-arid to arid climate with little precipitation and very sparse vegetation cover. Moreover, human modification and destruction of the landforms is also very limited in this area due to the sparse population (Rao et al., 2020). For the above reasons, fault scarps have been well preserved at the northern margin of the Hetao Basin, making it an ideal place to study the fault slipping histories through morphological analysis of the fault scarps.

In this study, high-resolution LiDAR topography were acquired along the Wulashan Piedmont Fault. On the basis of the LiDAR-derived topography, we tried to investigate the paleoseismicity of the fault from the morphology of fault scarps. A large set of scarp profiles were extracted to reconstruct the vertical displacement distribution along the fault, and well-preserved fault scarps were also selected to quantify the number of paleoearthquakes through identifying the slope breaks in the fault scarp morphology. The slip histories inferred from the morphology of fault scarps were then compared with the findings from previous paleoseismic investigations. Based on the derived rupture parameters and fault slip rate determined by previous studies, we estimated the moment magnitude and recurrence interval of large surface-rupturing events occurred on the Wulashan Piedmont Fault.

2 GEOLOGICAL BACKGROUND

The Ordos Block is a tectonically-stable block suited in North China. Four Cenozoic basins have developed around this block, including the Hetao, Shanxi, Weihe, and Yinchuan Basins (**Figure 1A**) (Ma and Wu, 1987; Zhang et al., 1998). Several devastating historical earthquakes have occurred around the Ordos Block, e.g., the 1303 *M* 8 Hongdong earthquake in the Shanxi Basin (Xu et al., 1993, 2018), the 1556 *M* 8.5 Huaxian earthquake in the Weihe Basin (Zhang et al., 1995; Feng et al., 2020), and the 1739 *M* 8 Yinchuan-Pingluo earthquake in the Yinchuan Basin (Deng and Liao, 1996; Middleton et al., 2016), while no large-magnitude earthquakes have been recorded in the Hetao Basin. However, paleoseismic investigations demonstrated several Pleistocene to Holocene surface-rupturing events in the Hetao Basin. (Chen, 2002; Ran et al., 2002, 2003; Yang et al., 2002, 2003; Chen et al., 2003; Rao et al., 2016; Dong et al., 2018; He et al., 2018). To the north, the basin is bounded by a series of mountain belts, including the Lang Shan, Serteng Shan, Wula Shan, and Daqing Shan, and major active structures have developed along the piedmonts of these mountain ranges (**Figure 1B**).



The Wulashan Piedmont Fault is a major normal fault at the northern margin of the Hetao Basin. The fault starts from Xishanzui in the west, and then goes through Gongmiaozi, Heshunzhuang, and Hayehutong, and finally ends near Baotou

in the east. It dips southward and generally strikes E-W, with a total length of 110 km. Controlled by this fault, the Wula Shan is more than 1,000 m above the ground surface, and the thickness of the Cenozoic sediments is up to 5,100 m in the basin (Chen,

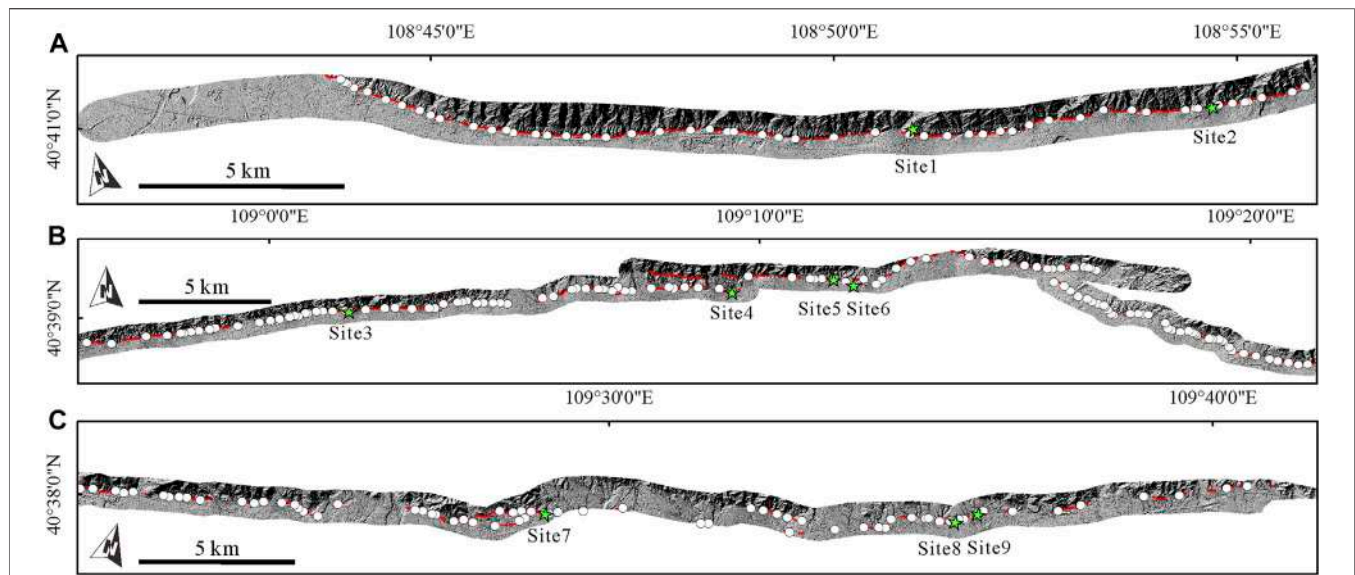


FIGURE 2 | (A), (B), and (C) show the LiDAR topography of the Xishanzui-Gongmiaozi segment, the Gongmiaozi-Heshunzhuang segment, and the Heshunzhuang-Baotou segment of the Wulashan Piedmont Fault, respectively. The red lines denote the mapped fault traces. The white circles mark the locations of all vertical displacement measurements, and the green stars indicate the selected nine study sites where fault scarps have been well preserved.

2002). According to the fault trace geometry, Chen (2002) divided the fault into three segments, i.e., the Xishanzui-Gongmiaozi segment, the Gongmiaozi-Heshunzhuang segment, and the Heshunzhuang-Baotou segment, which are separated by two broad and gentle bedrock bulges (Figure 1C). To investigate the paleoseismicity of the Wulashan Piedmont Fault, a total of four paleoseismic trenches have been excavated at different sites along the fault (Chen, 2002; Ran et al., 2003). Due to the long-term activity of the Wulashan Piedmont Fault, fault scarps with different heights have developed along the fault strike, which have been well preserved owing to the arid climate and limited human destruction in this region (Figures 1D,E).

3 DATA AND METHODS

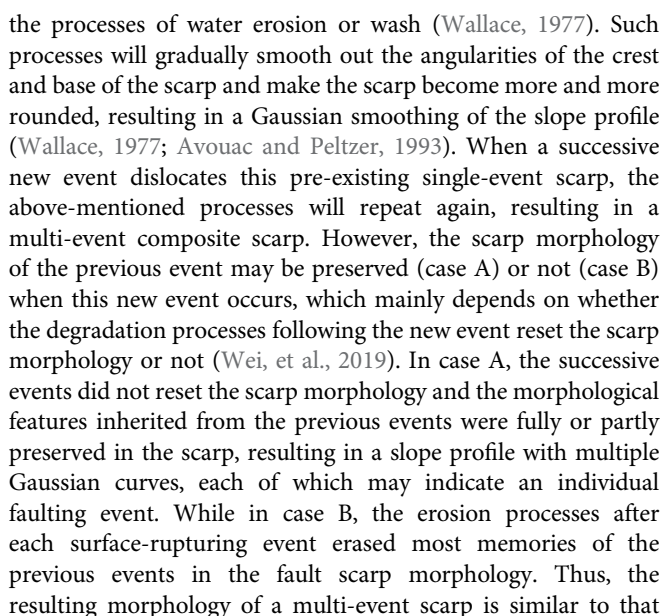
3.1 LiDAR Topography

To accurately analyze and measure the morphology of fault scarps, an airborne LiDAR survey was conducted along the Wulashan Piedmont Fault in August 2019. The topography were collected by a Riegl VUX-1LR airborne LiDAR system carried on a DM-150W fixed-wing unmanned aerial vehicle (UAV). The UAV flew at about 400 m above the ground surface. The scanning angle is $\pm 70^\circ$ and the scanning overlap is set to 50%. The pulse-shot frequency of the LiDAR system is 200 kHz. In order to improve the UAV positioning accuracy during the LiDAR survey, a total of five Trimble GPS base-stations were deployed every ~ 20 km along the fault strike to work with the onboard GPS synergistically. The coverage of the LiDAR topography is about ~ 1 km wide across the fault trace and ~ 90 km long along the fault trace (Figure 1C). The average point density is about 4 points/m², and in some local areas the point density can even reach 10 points/m². To acquire a bare-earth

DEM, the LiDAR points were grouped into different categories. Vegetation points were removed and only the ground points were left to generate the DEM. Due to the arid climate, the vegetation cover is very sparse in this area, so the ground points occupy nearly 90% of the total LiDAR points. By gridding the ground points, a 0.5-m-resolution bare-earth DEM was finally generated for the different segments of the Wulashan Piedmont Fault (Figure 2). To evaluate the precision of the LiDAR topography, a total of 30 check points were measured using a ground Trimble GPS system. A comparison between the LiDAR and ground GPS coordinates of the check points demonstrates that the vertical precision of the LiDAR topography is greater than 10 cm.

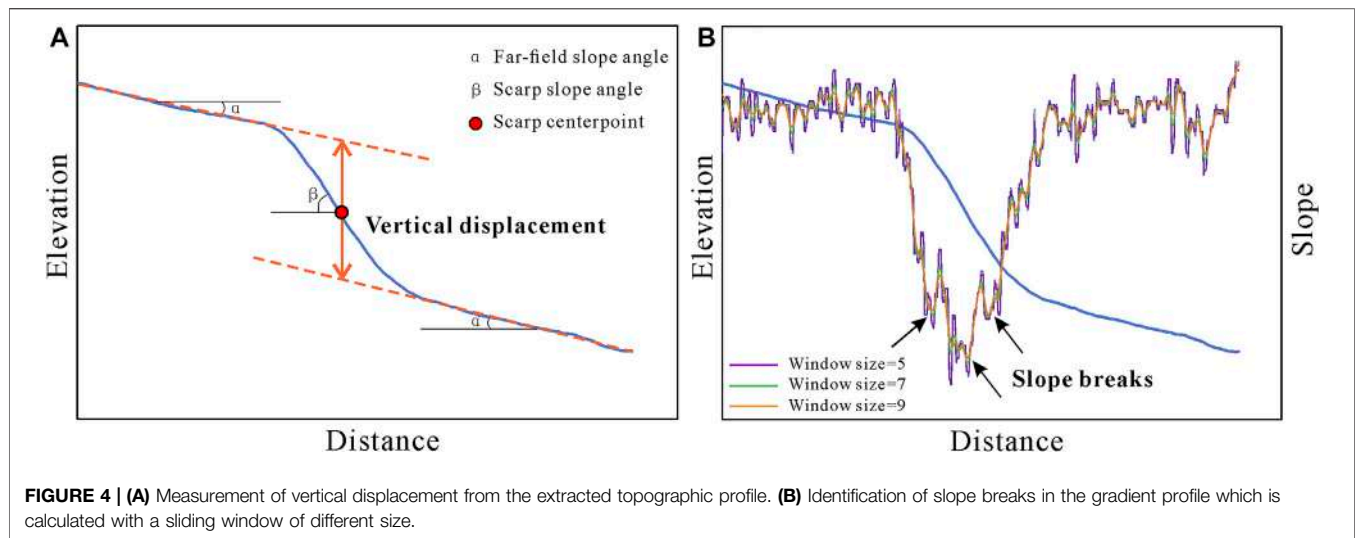
3.2 Fault Scarp Morphology

Before faulting, the original surface generally displays a constant slope angle. Then a surface-rupturing earthquake occurred, leading to the offset of the surface and the formation of a single-event fault scarp. The morphology of the scarp is controlled by many factors, mainly including the faulting mechanics and the properties of the faulted materials (Wallace, 1977). Typically, a single-event fault scarp consists of a free face whose gradient is greater than that of the original surface. Once a fault scarp has been formed, it will suffer from constant degradation process in the following hundreds or even thousands of years which can be divided into two phases. During the initial gravitational phase, the free face will collapse under the effect of gravity and forms a debris wedge at the base of the scarp, resulting in a gravity-controlled face at the repose angle of the material, usually between 30° and 35° (Wallace, 1977; Bucknam and Anderson, 1979; Avouac and Peltzer, 1993). Following this relatively rapid phase, the scarp will experience a much slower diffusive erosion phase mainly due to



3.3 Measuring Height of the Scarp

The high-resolution LiDAR topography provides detailed information of the landforms in the fault zone. On the basis of the LiDAR-DEM and its derivatives, e.g., slope and hillshade maps with various illumination angles, we mapped the fault traces and displaced landforms using the ESRI ArcGIS software. The LiDAR topography enables us to extract a large number of scarp-perpendicular topographic profiles along the fault strike, which has saved us from labor-intensive field mapping. Based on these extracted topographic profiles, we can measure the heights of the scarps to constrain the vertical displacement distribution along the Wulashan Piedmont Fault. To eliminate high-frequency variations caused by small-scale surface texture, swath topographic profiles were extracted perpendicular to the local fault scarp along the fault strike. When extracting the profiles, we



should keep away from the sites that have suffered from serious degradation or erosion process, and vegetation and gullies should also be avoided. The length of the profiles ranges from tens of meters to hundreds of meters to ensure that the far-field slopes of the original upper and lower surfaces can be well captured. Furthermore, if the profile has a large difference in the slope of the up-fault and down-fault surfaces (generally larger than 5°), it should be discarded. Since in this case, the surface on either side of the scarp may have suffered from serious degradation or erosion process, and the extracted topographic profile can no longer well represent the original surface of the fault scarp. To determine the vertical displacement, the heights of the hanging wall and footwall surfaces on each side of the scarp were fitted by two separate lines with a least squares regression. The vertical separation between the two lines at the center point of the scarp was measured as the vertical displacement (**Figure 4A**) (Hanks et al., 1984; Avouac and Peltzer, 1993; Wei et al., 2019). The error of linear regression was considered as the uncertainty of the vertical displacement measurement (Middleton et al., 2016; Bi et al., 2018, 2020).

3.4 Identifying Slope Breaks of the Scarp

According to the morphological evolution models of fault scarps (**Figure 3**), multiple surface ruptures on composite scarps may be identified by changes in scarp gradient, marked by slope breaks in the scarp profile (Wallace, 1980; Avouac and Peltzer, 1993; Carretier et al., 2002; Ewiak et al., 2015; Wei et al., 2019; Hodge et al., 2020). Thus, based on the scarp profiles, we can detect slope breaks in the scarp morphology to constrain the number of paleoearthquakes along the Wulashan Piedmont Fault. On the basis of detailed mapping of the landforms, a total of nine study sites were selected along the fault where well-preserved single or composite fault scarps can be found (the locations of the nine sites are indicated in **Figure 2**). Swath topographic profiles (40 topographic profiles at an interval of 0.5 m) were extracted perpendicular to the local fault scarp for each site. Following many prior studies (Ewiak et al., 2015; Wei

et al., 2019), a sliding window approach was employed to calculate the gradient (i.e., the first derivative) along the extracted topographic profile. In each sliding window, all the elevation data within the window are fitted by a polynomial regression model which is then differentiated to calculate the gradient. Choosing a suitable window size is the key to accurately identifying the gradient changes on a scarp. Slight gradient variations can be captured with smaller window size while larger window size can reduce the noises (**Figure 4B**) (Ewiak et al., 2015). After repeated trials, a window size of 9 was applied in our study. Generally, the slope breaks in the scarp profiles are marked by steep troughs in the calculated gradient profile (**Figure 4B**). For each extracted scarp profile in the swath, we first identified the locations of the crest and base of the scarp, and the segment between them was thus defined as the scarp range. Then, the troughs in the slope profile were searched in this scarp range and the positions (i.e., distance along the fault strike) of all searched troughs were recorded. Through statistical analysis of the recorded positions in all scarp profiles in the swath, we can acquire the probability density distribution of the positions of all searched slope breaks. The number of the peaks of this distribution may thus represent the number of slope breaks, that is, the number of surface-rupturing events recorded by the fault scarp.

4 RESULTS

4.1 Morphological Characteristics of Fault Scarps

The high-resolution LiDAR topography enables us to investigate the morphological characteristics of the fault scarps in great detail. Generally, the fault scarps along the Wulashan Piedmont Fault display a complex morphology produced by more than one surface-rupturing events. Two examples of composite fault scarp profiles are presented in **Figure 5**. It can be observed that the scarps are characterized by the combination

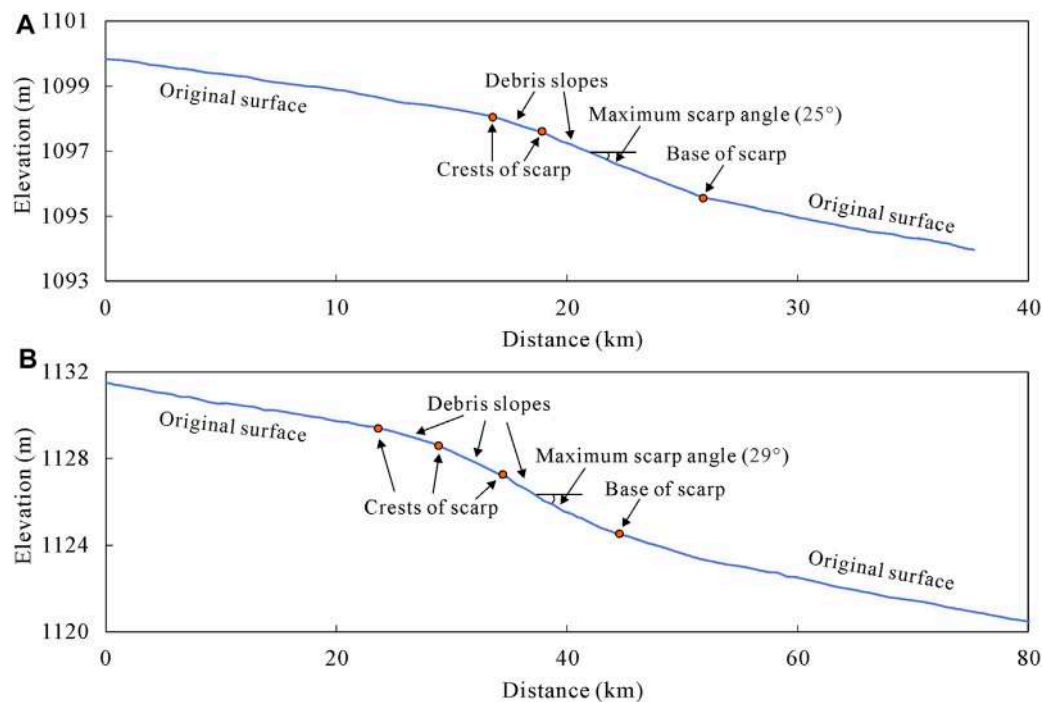


FIGURE 5 | (A) and (B) show two examples of composite fault scarp profiles at the Wulashan Piedmont Fault which are characterized by the combination of several different debris slopes in their morphology.

of several different debris slopes in their morphology which can be used as the indicators for multiple faulting events (Wallace, 1977; Ewiak et al., 2015; Wei et al., 2019). Free faces created by the most recent event have not been observed on the scarp profiles, and the maximum slope angles of the scarps are generally $<30^\circ$ which is smaller than the commonly observed angles of repose ($34^\circ\text{--}37^\circ$) on other faults (Wallace, 1977; Bucknam and Anderson, 1979), indicating that the fault scarps have evolved from the initial gravitational phase into the diffusive erosion phase. This low-angle ($<30^\circ$) scarp slopes are also consistent with previous paleoseismic investigation results, that is the most recent surface-faulting event occurred about $\sim 1,655$ years BP on the Wulashan Piedmont Fault (Chen, 2002; Ran et al., 2003).

4.2 Vertical Displacements of Fault Scarps

Through measuring the heights of fault scarps, we finally obtained a total of 243 vertical displacements along the Wulashan Piedmont Fault. All vertical displacements, distance along the fault, and measurement errors are presented in **Supplementary Table S1** of the Supplementary Material. The vertical displacements are of different amount, ranging from ~ 1 to ~ 70 m along the fault strike. Such large variations in the scarp heights indicate that this dataset includes both single-event coseismic displacements and multi-event cumulative displacements. **Figure 6** presents the distribution of all vertical displacements against their distance along the fault strike. We can see that the number of the displacements generally decreases with the increase of the displacement amount. This is mainly resulted from the steady erosion and degradation processes of the

landforms, making the older geomorphic surfaces generally less well preserved along the fault. To study the activity of the Wulashan Piedmont Fault, Chen (2002) measured the heights of fault scarps and acquired a total of 38 vertical displacements in the field. To evaluate the reliability of the vertical displacements measured in our study, we compared our measurements with the field measurements obtained by Chen (2002) at 28 sites (**Figure 7**). It can be observed that the vertical displacements acquired in our study show good consistency with previous field measurements, presenting a nearly 1:1 fit with a slope of 0.99 and a correlation coefficient of 0.96.

4.3 COPD of Vertical Displacements

Similar to many previous studies, we used a statistical method to analyze the large number of displacement measurements with different uncertainties (McGill and Sieh, 1991; Zielke et al., 2010, 2012; Klinger et al., 2011; Manighetti et al., 2015; Haddon et al., 2016; Bi et al., 2018, 2020). Each displacement is represented by a Gaussian probability distribution function (PDF), with the measured displacement value and its corresponding uncertainty as the mean (μ) and standard deviation (σ) of the Gaussian PDF, respectively. Adding up the PDFs of all displacements forms the cumulative offset probability density (COPD) (McGill and Sieh, 1991) which is always employed to identify individual earthquakes. Since geomorphic surfaces formed in the same period have been displaced by the same number of events, they always appear as clusters of similar displacements, leading to several separated peaks in the COPD plot. The first smallest COPD peak is usually interpreted as the

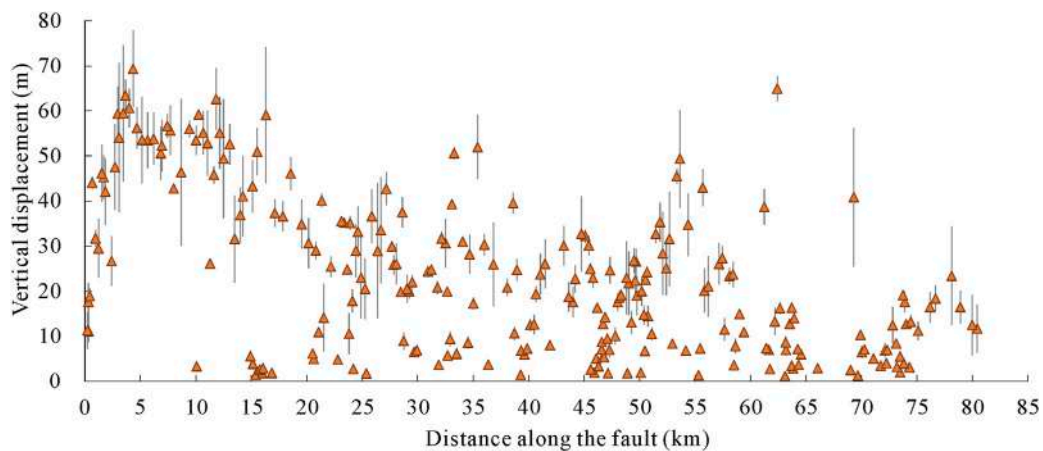


FIGURE 6 | Distribution of all vertical displacement measurements against their distance along the fault strike.

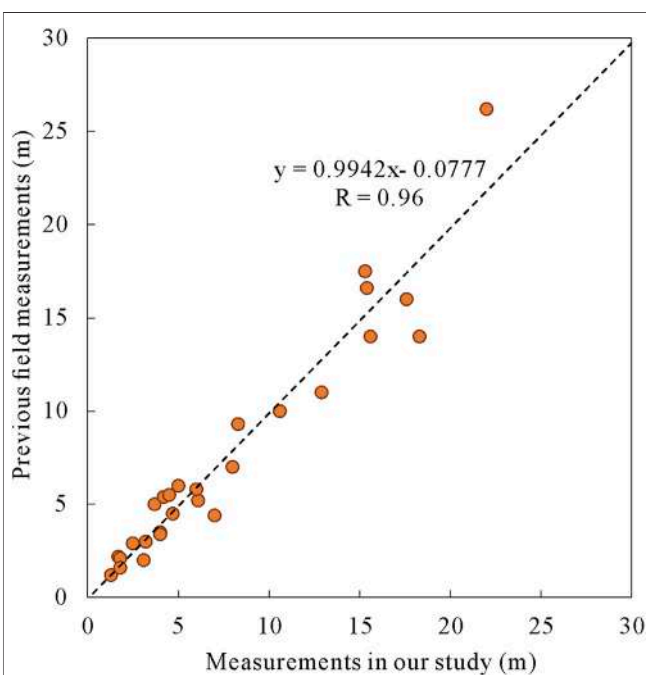


FIGURE 7 | Scatter diagram displaying the comparison between vertical displacement measurements in our study with previous field measurements of Chen (2002).

coseismic slip of the most recent event, while the successive larger peaks may represent the cumulative slip due to two or more preceding earthquakes (McGill and Sieh, 1991; Zielke et al., 2010, 2012; Klinger et al., 2011; Manighetti et al., 2015; Haddon et al., 2016; Bi et al., 2018, 2020).

Considering that the older geomorphic surfaces have experienced a longer period of erosion and degradation processes, the larger displacements will have larger uncertainties. Therefore, only the displacements in the range

of 0–8 m were statistically analyzed in our study. Furthermore, since different segment may have different rupture behavior, the COPD was calculated separately on each individual segment. It should be noted that the Xishanzui-Gongmiaozi segment mainly consists of bedrock fault scarps, and very few small displacements have been preserved on this segment. Thus the COPD plots were derived only for the Gongmiaozi-Heshunzhuang segment (**Figure 8**) and the Heshunzhuang-Baotou segment (**Figure 9**). The COPD curve consists of several peak shapes. The peak values represent the most frequent cumulative displacements of the faulting events, while the width of the peak shape reflects both the measurement uncertainty and the variation of displacement along the fault strike. In our study, a Gaussian PDF was used to fit each peak shape, and the mean (μ) and the half-width (σ) of the Gaussian PDF was considered as the peak value and its corresponding uncertainty. A total of seven prominent peaks can be identified from the COPD curve on the Gongmiaozi-Heshunzhuang segment, which are centered at 1.3 ± 0.3 m, 2.0 ± 0.4 m, 2.7 ± 0.3 m, 3.7 ± 0.3 m, 5.0 ± 0.4 m, 6.1 ± 0.5 m, and 6.9 ± 0.5 m, respectively. The COPD curve also displays seven noticeable peaks on the Heshunzhuang-Baotou segment, at 1.2 ± 0.2 m, 2.0 ± 0.5 m, 2.9 ± 0.4 m, 3.8 ± 0.3 m, 5.0 ± 0.4 m, 6.0 ± 0.5 m, and 7.0 ± 0.5 m, respectively. Zielke et al. (2015) proposed a Monte-Carlo method to determine the mean slip of each event while considering all derived COPD peak values and their corresponding uncertainties. Following this approach, we derived seven successive events with repeated slip of 1.0 ± 0.5 m for both the two segments.

4.4 Slope Breaks of Fault Scarps

A total of nine study sites were selected on different segments of the Wulashan Piedmont Fault where single-event fault scarps or multi-event composite scarps had been well preserved. The locations of the nine sites were indicated in **Figure 2**. Through morphological analysis of the scarp profiles at the nine sites, several slope breaks have been identified that indicate multiple surface-faulting events (**Figures 10–12**). The number of slope breaks ranges from one to five at each site,

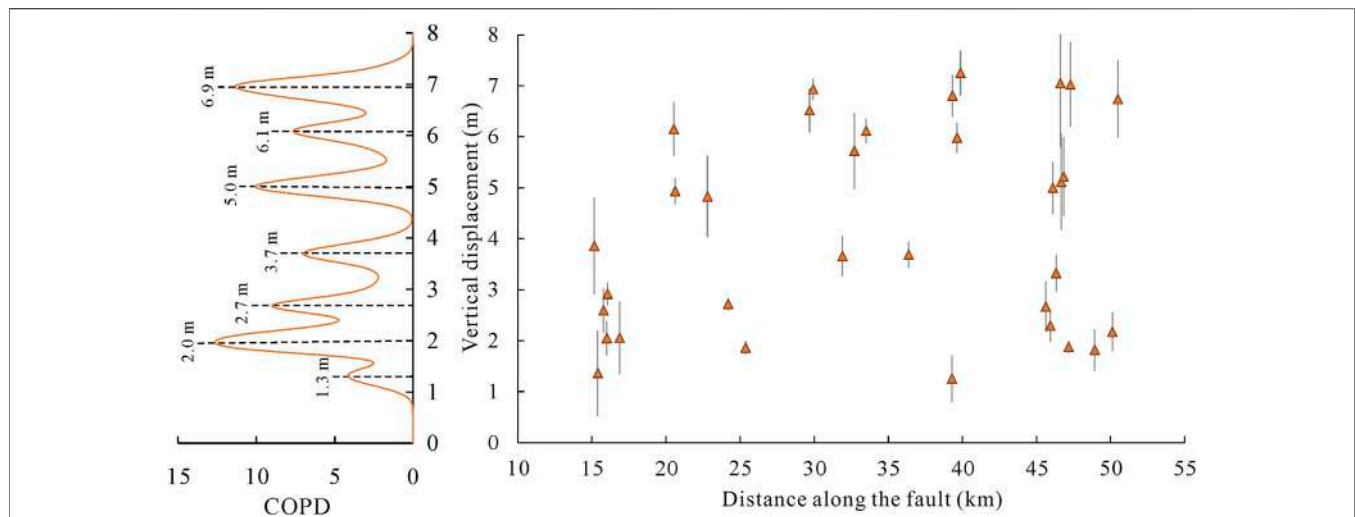


FIGURE 8 | Statistical analysis of the vertical displacements on the Gongmiaozi-Heshunzhuang segment in the range of 0–8 m. Each displacement is represented by a Gaussian PDF. Adding up the PDFs of all displacements forms the COPD curve which generally shows several peak shapes, and each peak may indicate an individual earthquake.

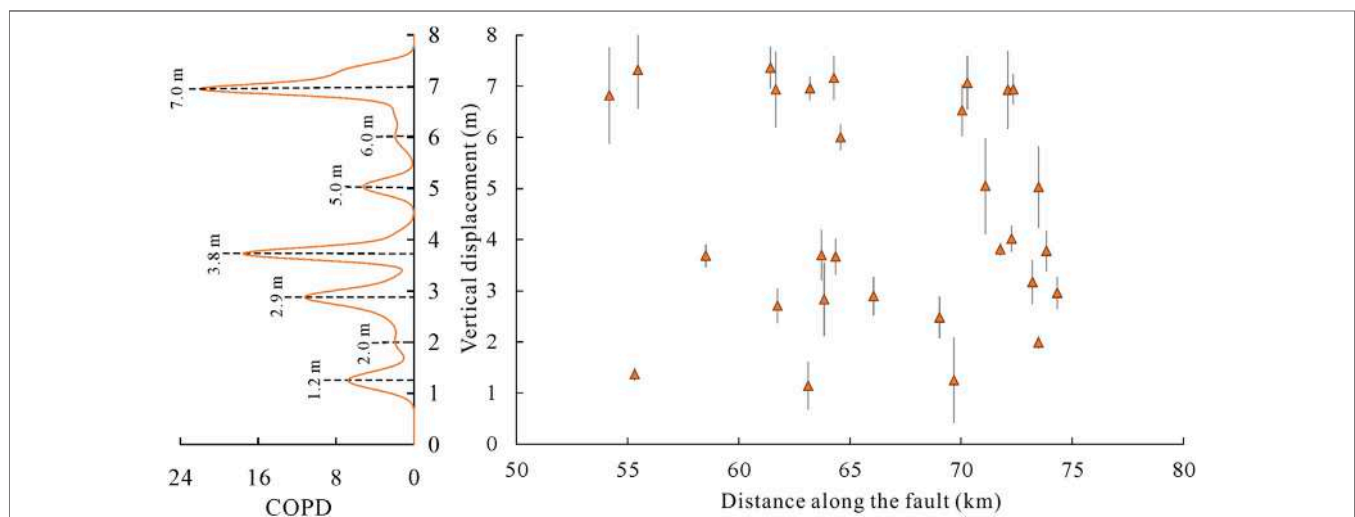
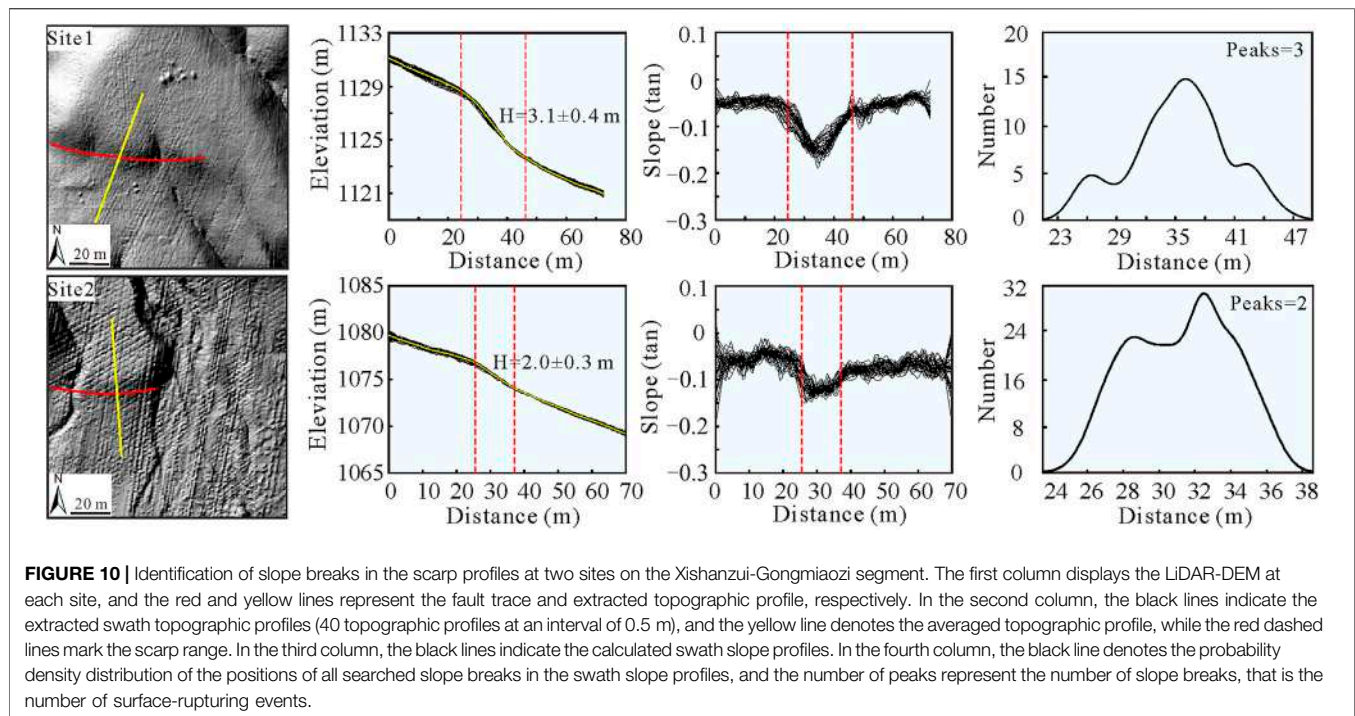


FIGURE 9 | Statistical analysis of the vertical displacements on the Heshunzhuang-Baotou segment in the range of 0–8 m. Each displacement is represented by a Gaussian PDF. Adding up the PDFs of all displacements forms the COPD curve which generally shows several peak shapes, and each peak may indicate an individual earthquake.

depending on the location of the fault scarp. Generally, the number of slope breaks identified from the scarp is positively correlated with the height of the scarp. This is in accordance with expectation, since the scarps with larger heights are generally located on older geomorphic surfaces which have experienced more surface-rupturing events. By comprehensively analyzing the slope breaks at the nine sites, we found that the average heights of the fault scarps with one to five breaks are 1.2, 2.0, 3.1, 5.5, and 6.5 m, respectively, which is generally in good agreement with the peak values of the COPD plot. Furthermore, among the nine study sites, two sites (sites 7 and 8) are suited very close to paleoseismic trenches excavated by previous studies (Chen, 2002;

Ran et al., 2003). Thus, the number of events recognized from the fault scarp morphology can be compared with previous paleoseismic trenching records. The site 7 is located beside the Hayehutong trench which has revealed three paleoearthquakes. However, Chen (2002) reported that the paleoseismic events recorded by this trench were incomplete, since the total displacement of the three events exposed from the trench is much smaller than the height of the fault scarp where the trench has been excavated. A total of five slope breaks have been detected from the scarp profiles at site 7, indicating at least five surface-rupturing events, which can thus be a good complement to the paleoseismic trenching records. Moreover, the site 8 is suited on



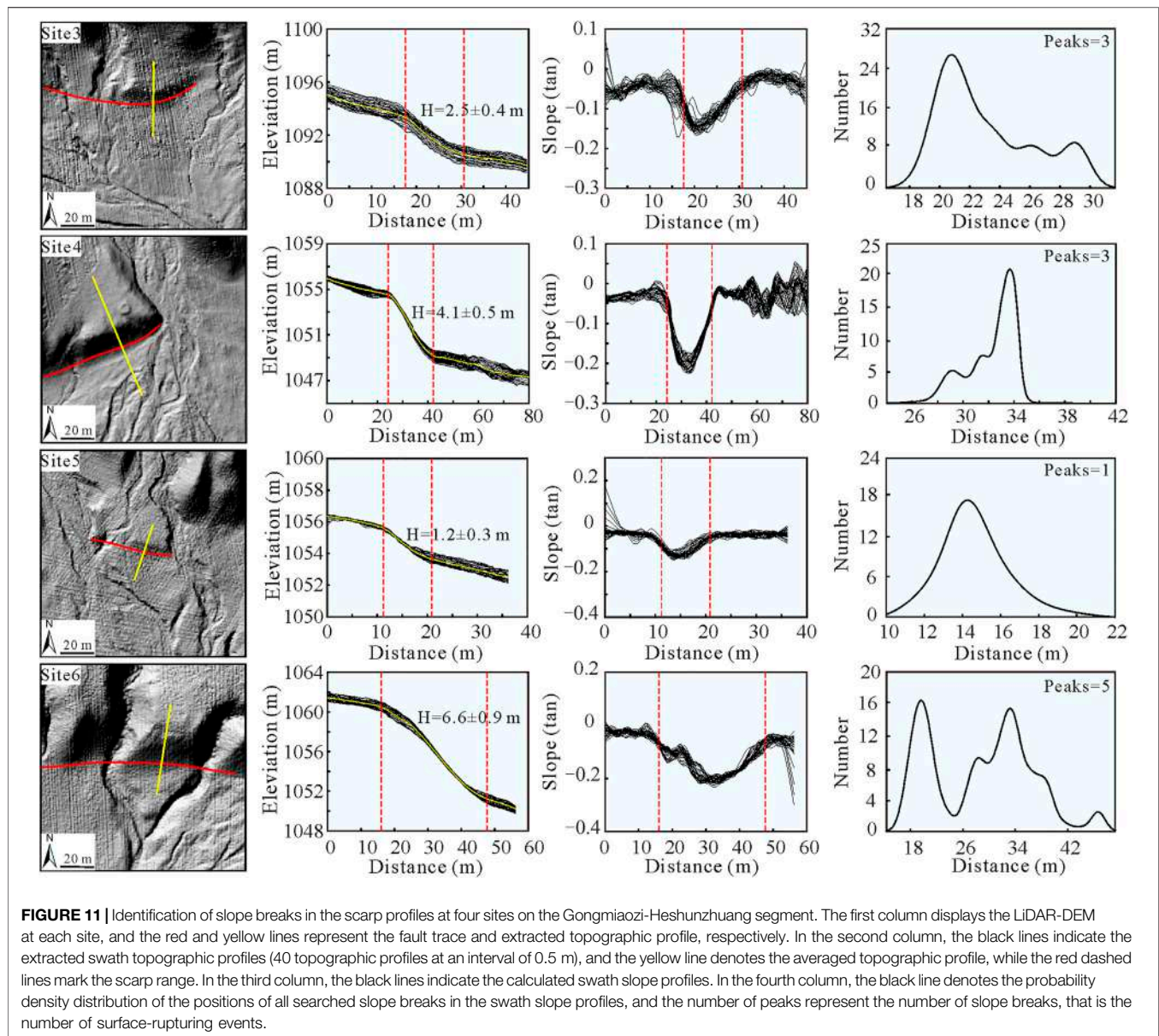
the left side of the Hademenjinkuang trench which exposed a total of four paleoseismic events. The scarp profiles at site 8 also show four obvious slope breaks, which is in good consistence with the four events recorded in the paleoseismic trench.

5 DISCUSSION

5.1 Paleoseismicity of the Wulashan Piedmont Fault

By constraining the paleoseismicity of a fault, we are able to gain important insights into the rupture histories and recurrence patterns of past events occurred on the fault, which provides valuable information on the future behavior of the fault (Zielke et al., 2015). Recently, a few studies have investigated the fault rupture histories through identifying the COPD peaks of a dense collection of displacement dataset along the fault (Zielke et al., 2010, 2012; Klinger et al., 2011; Manighetti et al., 2015; Haddon et al., 2016; Bi et al., 2018, 2020). Similar to these prior works, we reconstructed the vertical displacement distribution of the Wulashan Piedmont Fault through measuring the heights of a large number of fault scarp profiles extracted on different geomorphic surfaces along the fault strike. By statistical analysis of the vertical displacements on the Gongmiaozi-Heshunzhuang segment and the Heshunzhuang-Baotou segment respectively, we found that the COPD plot presents seven pronounced peaks on both segments, which are separated by a similar slip increment of ~ 1 m. This slip increment is also very close to the first smallest peak value which is always interpreted as the coseismic slip of the most recent earthquake. Therefore, we infer that at least seven paleoearthquakes have occurred on the Wulashan Piedmont

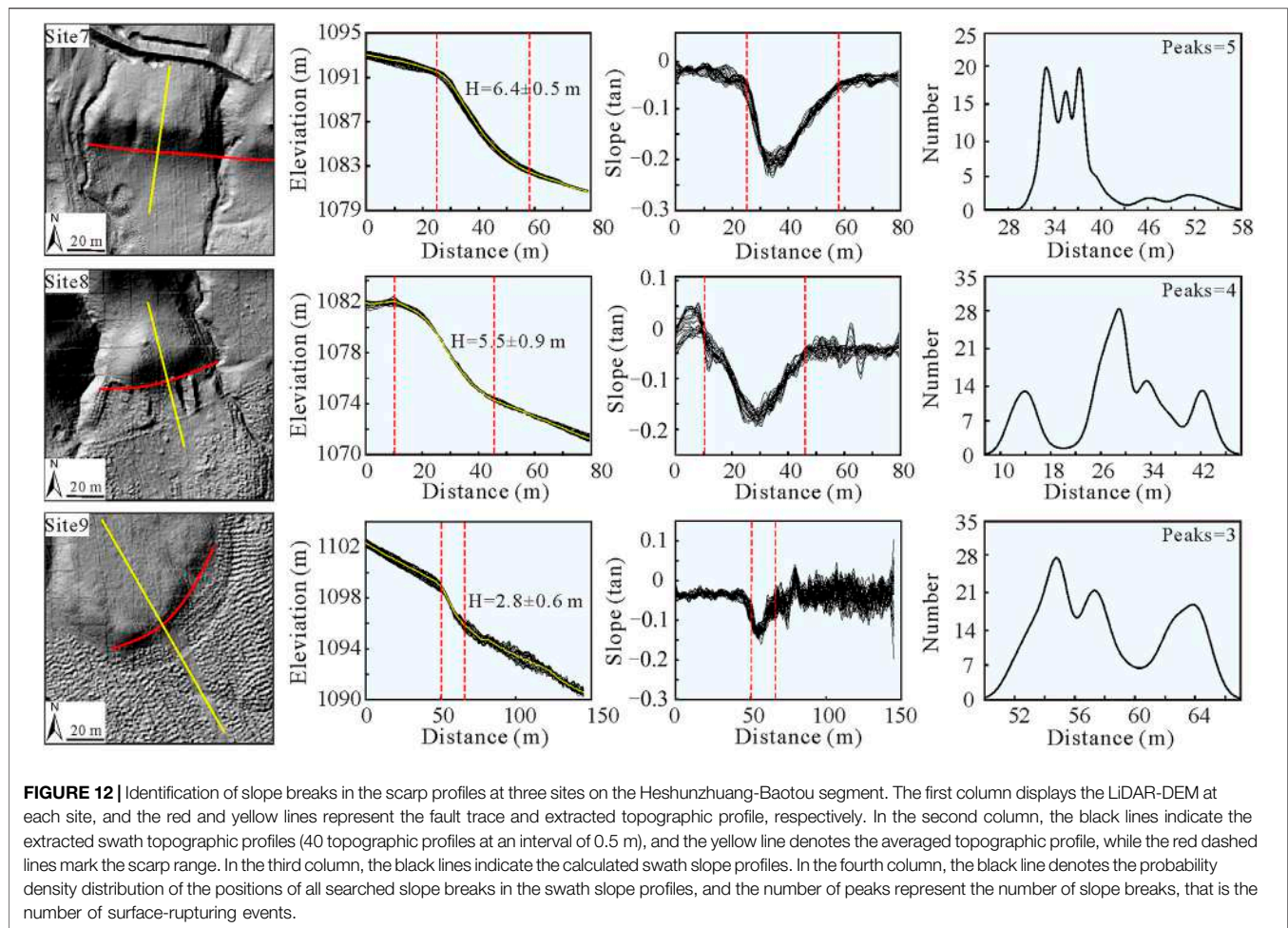
Fault, following a characteristic slip pattern. Since the COPD peak values on both segments are very similar, the seven events may have broken the two segments together, producing a cascade rupture between the two segments. By applying the Monte Carlo method proposed by Zielke et al. (2015), we derived the average slip of the seven paleoseismic events to be 1.0 ± 0.5 m. This derived average slip is in good agreement with the coseismic slip of 1.2–2.4 m determined by previous paleoseismic investigation (Chen, 2002). Besides the heights of the fault scarp, the scarp shape also preserves valuable information about the history of a fault, and has been employed to identify individual earthquakes through detecting the gradient changes (slope breaks) in the scarp profiles on many faults (Wallace, 1980; Avouac and Peltzer, 1993; Carretier et al., 2002; Ewiak et al., 2015; Wei et al., 2019, 2021; Rao et al., 2020). Through morphological analysis of the scarp profiles extracted at nine sites along the Wulashan Piedmont Fault, we found that the scarp with a height of ~ 1.2 m displays only one slope break, which may represent the coseismic slip produced by the most recent earthquake. While five slope breaks have been identified from the fault scarps with a height of ~ 6.5 m, indicating at least five faulting events. We further comprehensively compared the slope breaks at the nine sites and found that the fault scarps with one to five breaks have an average height of 1.2, 2.0, 3.1, 5.5, and 6.5 m, respectively, generally in good consistence with the peak values of the COPD plot. These derived slope breaks suggest that at least five surface-rupturing events have occurred on the Wulashan Piedmont Fault. There are two events that have not been recorded by the fault scarp morphology, which however have been recognized from the COPD plot of the vertical displacements of fault scarps. Moreover, previous paleoseismic excavations also demonstrated that at least six paleoearthquakes had happened on the Wulashan Piedmont Fault since the



Holocene, and these earthquakes have followed a quasi-periodic recurrence model with similar recurrence intervals (Ran et al., 2003), which is in good agreement with the paleoseismicity constrained from the morphology of fault scarps. Based on these above results, we demonstrate that both the vertical displacements of fault scarps and the slope breaks in the scarp morphology can be applied to quantify the number and slip of paleoseismic events. However, we should emphasize the importance of integrating the morphological analysis results of fault scarps with the findings from the paleoseismic trenches. They can provide good validation and supplement for each other to acquire more complete and reliable fault rupturing histories, especially on faults with complex fault geometries or slip histories (Ewiak et al., 2015).

5.2 Seismic Risk on the Wulashan Piedmont Fault

Knowledge of the magnitude and recurrence interval of past earthquakes can help us estimate the probability and severity of future potential events on the fault, thus greatly benefiting seismic hazard assessment (McCalpin, 2009). From the COPD peaks of the vertical displacements measured from the fault scarps along the fault strike, we determine that at least seven paleoearthquakes have ruptured the Wulashan Piedmont Fault, with an average slip of 1.0 ± 0.5 m. Supposing an average fault dip of $\sim 50^\circ$ acquired from field observations (Chen, 2002) and considering the uncertainty of the average slip, the average slip of the paleoseismic events on the fault plane at the surface was estimated to be 0.7–2.0 m. Through analyzing the source



parameters of a large number of historical earthquakes worldwide, Wells and Coppersmith (1994) established an empirical relationship between the moment magnitude M_w and the average displacement (AD) as following: $M_w = 6.78 + 0.65 \log(AD)$. According to this formula, a moment magnitude of M_w 6.7–7.0 was derived for the paleoearthquakes occurred on the Wulashan Piedmont Fault. Considering the similar COPD peak values on the Gongmiaozi-Heshunzhuang segment and the Heshunzhuang-Baotou segment, it is deduced that the seven events may have broken the two segments simultaneously, producing a cascade rupture between the two segments. Due to the limited number of small displacements on the west Xishanzui-Gongmiaozi segment, the displacements were not statistically analyzed on this segment. However, we cannot rule out the possibility that all the three segments have failed together in the seven large paleoearthquakes, leading to a rupture of the whole length of the fault. Thus, the rupture length is estimated to be about 70–110 km. With the scaling relationship between the moment magnitude M_w and the surface rupture length (SRL) (Wells and Coppersmith, 1994): $M_w = 4.86 + 1.32 \log(SRL)$, a moment magnitude of M_w 7.3–7.5 was determined for these paleoseismic events. We also calculated the M_w using an

updated surface rupture length regression developed more recently by Wesnousky (2008), that is, $M_w = 6.12 + 0.47 \log(SRL)$. Based on this formula, a moment magnitude of M_w 7.0–7.1 was yielded. All the above estimates demonstrate that the paleoearthquakes occurred on the Wulashan Piedmont Fault have a moment magnitude in the range of M_w 6.7–7.5. Assuming that the accumulated strain on a fault was released only through repeated seismic slip, the average recurrence interval of paleoearthquakes occurred on the fault can be roughly estimated based on the coseismic slip and the fault slip rate. Chen (2002) has estimated the vertical slip rate of the Wulashan Piedmont Fault which is about 0.55–0.77 mm/yr since Holocene. From the COPD peaks, we have estimated the coseismic slip of the paleoseismic events to be ~1.0 m. Thus, the average recurrence interval of paleoearthquakes that have occurred on the Wulashan Piedmont Fault was estimated to be 1.3–1.8 kyr, which is comparable with the average recurrence interval of ~1,346 years derived from previous paleoseismic investigations (Chen, 2002; Ran et al., 2003). Moreover, the paleoseismic trenching results revealed that the most recent event occurred at about ~1,655 years BP on the fault (Chen, 2002; Ran et al., 2003). Thus, the elapsed time since the most recent event is very close to our estimated earthquake recurrence

interval, indicating a high potential seismic risk on the Wulashan Piedmont Fault.

6 CONCLUSION

This study constrained the paleoseismicity of the Wulashan Piedmont Fault suited at the northern margin of the Ordos Block through morphological analysis of the fault scarps. Based on high-resolution LiDAR topography, a large number of vertical displacements were measured from the fault scarps to constrain the vertical displacement distribution along the fault. Well-preserved fault scarps were also selected to quantify the number of surface-rupturing events through detecting the number of slope breaks in the fault scarp morphology. Both the COPD peaks of the vertical displacements and the slope breaks in the scarp profiles demonstrate that at least seven paleoseismic events have ruptured the Wulashan Piedmont Fault. These events displayed a very similar slip of ~ 1.0 m, and thus may have followed a characteristic slip pattern. Through comparison with previous paleoseismic investigations, it was found that the events identified from the morphological analysis of fault scarps are in good consistence with those revealed from paleoseismic trenches. By applying empirical scaling relationships between moment magnitude and rupture parameters, we estimated a possible moment magnitude of M_w 6.7–7.5 for the paleoearthquakes occurred on the fault. Based on the fault slip rate determined in previous studies and the coseismic slip derived from the fault scarp morphology, we estimated the average recurrence interval for large earthquakes to be 1.3–1.8 kyr on the fault. Considering that the elapsed time since the most recent event is very close to the estimated average recurrence interval, there may be a high potential seismic risk on the Wulashan Piedmont Fault.

This study has demonstrated the applicability of the fault scarp serving as a geomorphic marker to quantify both the number and slip of past earthquakes on a fault based on high-resolution topography, which can provide a good complement to the paleoseismic trenching records. However, it should be noted that since this method relies on the preservation of subtle morphological features inherited from previous events, it is more applicable in arid and semi-arid areas with low erosion rates. If a composite fault scarp has suffered sufficient erosion after a seismic event, it would gradually evolve into a single-event

fault scarp shape. In this case, discriminating individual surface-breaking events from the scarp morphology would be difficult. Furthermore, though the fault scarp morphology can be applied to constrain the rupturing histories of a fault, the morphological records should be compared with the events recorded in paleoseismic trenches to avoid overrepresentation or underrepresentation of the paleoearthquakes. Combining both the morphological records of fault scarps and paleoseismic trenching results can lead to a more comprehensive understanding of the fault slip histories and rupture patterns, and thus a better assessment of potential seismic hazard on the fault.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

HB is responsible for making the research plan and writing the manuscript. LS is responsible for processing data and doing experiments. HP, JY, ZY, and XS are responsible for field investigation. DZ is responsible for professional guidance.

FUNDING

This research was funded by the National Key Research and Development Program of China (2017YFC1500101), the National Natural Science Foundation of China (41972228, 42174062), the Guangdong Province Introduced Innovative R&D Team of Geological Processes and Natural Disasters around the South China Sea (2016ZT06N331).

SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.911173/full#supplementary-material>

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Conflict of Interest: The reviewer (JR) declared a shared affiliation with the authors (HB, ZY) to the handling editor at the time of review.

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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New Insights Into the Paleoseismic History of the Mae Hong Son Fault, Northern Thailand

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OPEN ACCESS

Edited by:

Wenjun Zheng,
Sun Yat-sen University, China

Reviewed by:

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University of Pavia, Italy
Xiaodong Yang,
South China Sea Institute of
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Specialty section:

This article was submitted to
Structural Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

Received: 15 April 2022

Accepted: 10 June 2022

Published: 12 July 2022

Citation:

Chansom C, Jitmahantakul S,
Owen LA, Wiwegwin W and
Charusiri P (2022) New Insights Into
the Paleoseismic History of the Mae
Hong Son Fault, Northern Thailand.
Front. Earth Sci. 10:921049.
doi: 10.3389/feart.2022.921049

The Mae Hong Son Fault (MHSF) is a north-trending active fault in northern Thailand. The largest earthquake ever recorded in Thailand occurred in February 1975 with a magnitude of 5.6 and was associated with the southern end of the MHSF. Paleoearthquake magnitudes, recurrence intervals, and slip rates for the MHSF are evaluated using the morphological characteristics of the MHSF aided with a 12.5-m-resolution digital elevation model (DEM) and using fault trenching. Morphotectonic analysis, including studies of offset streams, linear valleys, triangular facets, and fault scarps, helps illustrate dextral fault movements within the MHSF zone. Two separated N–S trending basins, the Mae Hong Son to the north and the Mae Sariang to the south, are present along the MHSF. Between these basins, fault displacements decrease toward the Khun Yuam area. Surface rupture length investigation from fault segments in both basins indicates maximum credible earthquake magnitudes between 5.8 and 6.3. Fault trenching and road-cut studies show that nine earthquakes occurred along the MHSF over the past ~43 ka. Optically stimulated luminescence (OSL) dating help define the timing of the earthquakes to ~43, ~38, ~33, ~28, ~23, ~18, ~13, ~8, and ~3 ka. The recurrence interval of earthquakes on the Mae Hong Son Fault is ~5,000 years and the fault has a slip rate of ~0.04–0.15 mm/a.

Keywords: active fault, paleoearthquake, Mae Hong Son Fault, Mae Hong Son Basin, Mae Sariang Basin, OSL dating

1 INTRODUCTION

Thailand has experienced many M 3–6 and a few M ≥ 6 earthquakes since at least 624 BC based on historical and instrumental records (Charusiri et al., 2007; Wiwegwin et al., 2020; USGS, 2021; Thai Meteorological Department, 2022). These moderate to strong earthquakes have been detected along major fault traces in northern Thailand (Figure 1). Even though this region is located far away from the present-day plate boundary of Southeast Asia, the Andaman–Sumatra subduction zone (Subarya et al., 2006; McCaffrey, 2009; Roy et al., 2011), paleoseismological investigations reveal that Thailand is to some extent controlled by active indent–linked strike-slip faults due to the plate boundary (Fenton et al., 2003; Pailoplee et al., 2009; Wiwegwin et al., 2014; Pailoplee and Charusiri, 2017). The largest (Mw 6.2) instrumentally recorded earthquake in Thailand occurred on 5 May 2014, causing the greatest amount of damage in Thailand's history. The epicenter was associated with the north-trending Mae Lao Fault in Chiang Rai Province,

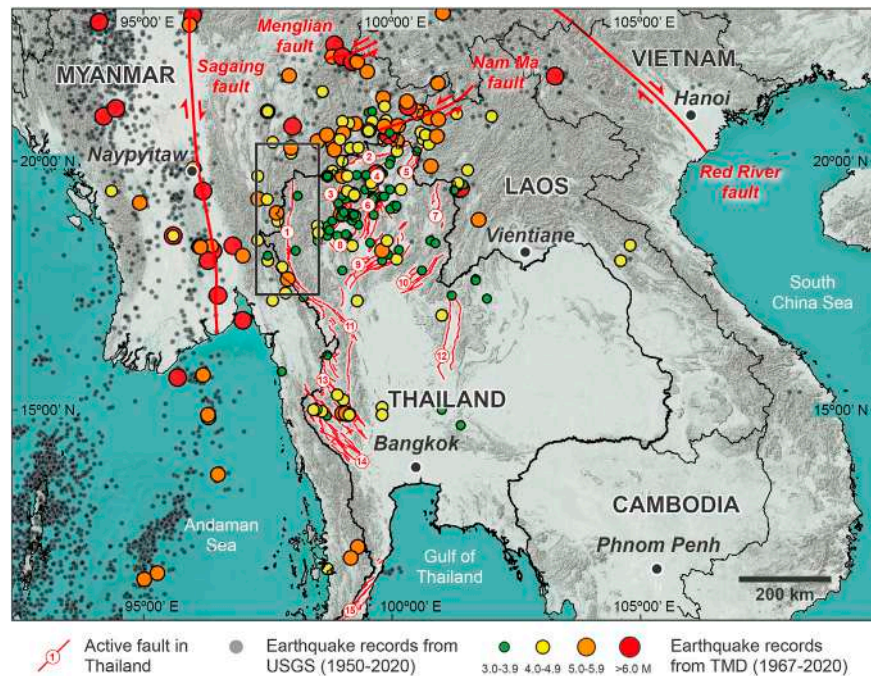


FIGURE 1 | Major active faults in Thailand and adjacent regions (modified from Department of Mineral Resources (2019) and Wang et al. (2014)) and epicentral distribution from 1967 to 2021 (USGS, 2021; Thai Meteorological Department, 2022). Fault name no. 1: Mae Hong Son Fault; 2: Mae Chan Fault; 3: Wiang Heang Fault; 4: Mae Lao Fault; 5: Mae Ing Fault; 6: Phayao Fault; 7: Pua Fault; 8: Mae Tha Fault; 9: Thoen Fault; 10: Uttaradit Fault; 11: Moei Fault; 12: Petchabun Fault; 13: Sri Sawat Fault; 14: Three Pagodas Fault; 15: Ranog Fault. The rectangle shows the location of the study area.

~27 km from Chiang Rai City. More than 50,000 people felt this earthquake, including people living in Bangkok, and more than 15,000 buildings were damaged, resulting in two deaths and 107 injuries (DDPM, 2014).

In 1975, an Mw 5.6 earthquake near the southern part of the north-trending Mae Hong Son Fault (MHSF) caused minor damage to Mae Hong Son and surrounding areas (**Figure 2**). With an intensity of VI, the earthquake was felt by people in northern and central parts of Thailand and central Myanmar (Prachaub, 1990). The MHSF is a crustal boundary fault between the Inthanon zone and the Sibumasu block (Hisada et al., 2004). The trend of the MHSF is almost parallel and close to the Sagaing Fault (~180 km to the west of Mae Hong Son Province), which is the most prominent active fault in Myanmar. Based on earthquake data and significant morphotectonic landforms, the MHSF is considered one of only a few active faults in Thailand (Charusiri et al., 2007; Department of Mineral Resources, 2019). Paleoseismic studies by Wiwegwin et al. (2014) show that there have been at least eight earthquakes within the Mae Hong Son area during the past 78 ka, with a likely earthquake recurrence interval of 10,000 years. They also suggested that there is a low probability of a large earthquake on the MHSF. The analysis of the seismic hazard by probabilistic seismic hazard analysis (PSHA) by Pailoplee and Charusiri (2016) suggested that the Mae Hong Son Province has the second-highest seismic hazard level in Thailand, with a 22% probability of exceedance (POE) of an earthquake with an intensity of VII occurring in the next 50 years.

Our study, therefore, aimed to provide a better understanding of the nature of seismicity in the Mae Hong Son area to aid in

earthquake hazard assessment. We used paleoseismic studies of fault segments along the MHSF and examined the tectonic geomorphology using high-resolution digital elevation models, Landsat-7 images, and satellite images available on Google Earth. In addition, we defined the fault slip rate, date past earthquakes, and determine earthquake recurrence intervals along the MHSF. The study area covers ~25,500 km² and is bounded by latitude 17°38'N–19°48'N and longitude 97°20'E–98°39'E in Mae Hong Son Province. Fault trenching and tectonic geomorphology were carried out across the three selected fault segments along the MHSF, which we name Ban Yod, Khun Yuam, and Mae Tha Lu. The Ban Yod fault segment is a ~4.2-km-long NNE-trending right-lateral strike-slip fault, at the northern end of the MHSF (**Figure 3**). The Khun Yuam fault segment is in the central part of the MHSF zone and is a ~3.5-km-long N- to NNE-trending right-lateral strike-slip fault that cuts across Khun Yuam District (**Figures 2, 3**). The Mae Tha Lu fault segment at the southern end of the MHSF is a ~8.8-km-long N- to NNW-trending right-lateral strike-slip fault (**Figure 4**). We, then, compared our paleoseismic findings with previous studies.

2 NEOTECTONIC SETTING AND PRESENT-DAY SEISMICITY

Seismicity in the Myanmar–Laos–Northern Thailand region is associated with Indian–Eurasian continental–continental collision (Peltzer and Tapponnier, 1988; Wang et al., 2014). A NE-to NNW-trending strike-slip fault network is dominant and has developed

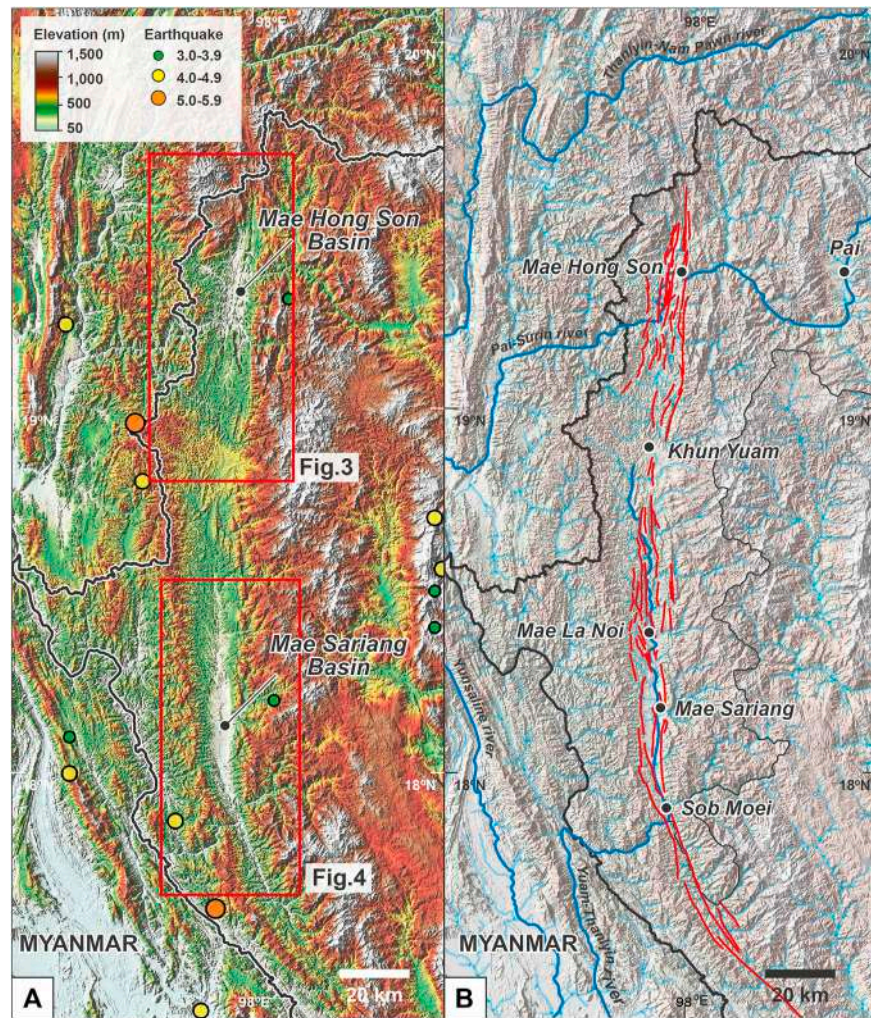


FIGURE 2 | Study area shows: **(A)** enhanced ALOS DEM data highlighting the Mae Hong Son Basin in the north and the Mae Sariang Basin in the south of the Mae Hong Son Province; and **(B)** hillshade ALOS DEM and stream maps highlight the north trending fault segments throughout Mae Hong Son Province. The red rectangles in part **(A)** highlight the geomorphic study areas shown in other figures.

since the Late Cretaceous (Morley, 2004). Some of these fault trends strongly followed pre-collision fabrics within the West Burma and Sibumasu terranes (Morley et al., 2007; Searle and Morley, 2011). As India moved northward into Asia, several of the major faults, such as the Red River, Moei (also known as Mae Ping), and Three Pagodas faults (**Figure 1**), have rotated clockwise and moved laterally. A regional reversal of slip from left to right lateral on these faults is placed around the Late Oligocene (Allen et al., 1984; Lacassin et al., 1997; Lacassin et al., 1998).

Historical and instrumental recorded seismicity shows that at least 20 destructive earthquakes with $M > 6$ have occurred in this region (Ekström et al., 2012; Shi et al., 2018). Most earthquakes are clustered along the sinistral and dextral strike-slip faults between the Sagaing (bounds on the west) and the Red River (bounds on the east) faults (**Figure 1**). Recent significant earthquakes on these faults include the 1988 Mw 7.0 Lancang–Gengma earthquake at the north-western end of the Lancang Fault in south China (Chen and Wu,

1989), the 1995 Mw 6.8 Menglian earthquake at the southern of the Menglian Fault in eastern Myanmar (Ji et al., 2017), the 2011 Mw 6.8 Tarlay earthquake at the western end of the Nam Ma Fault in eastern Myanmar (Tun et al., 2014), and the 2014 Mw 6.1 Mae Lao earthquake along the Mae Lao Fault in northern Thailand (Pananont et al., 2017). In addition, small and shallow earthquakes in the Myanmar–Laos–Northern Thailand region demonstrate widespread deformation across the fault systems between the Sagaing and the Red River faults, suggesting that these faults are active today.

3 MORPHOTECTONIC LANDFORM INTERPRETATION

High-resolution terrain corrected ALOS PALSAR dataset 2007 with a pixel size of 12.5 m, Landsat-7 images satellite from

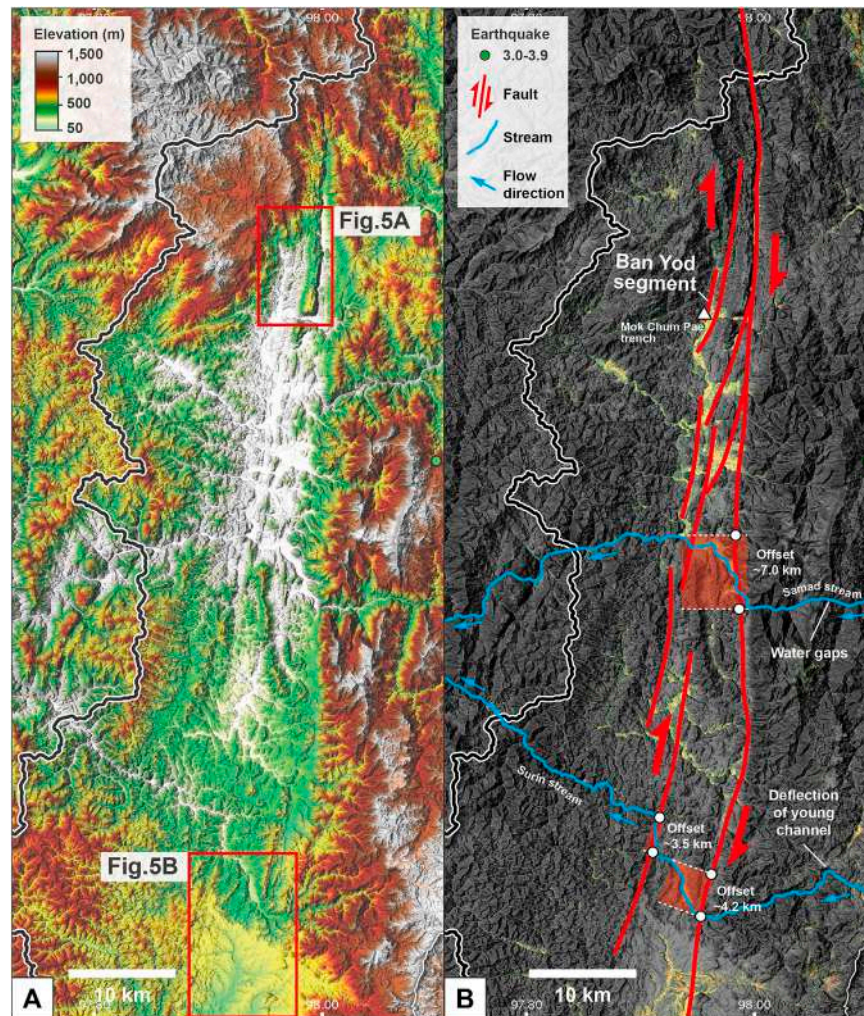


FIGURE 3 | Mae Hong Son Basin on the north side of Mae Hong Son Province is shown as an **(A)** enhanced ALOS DEM, and **(B)** detailed interpretation at the Samad stream displays offset stream with maximum offset of ~7.0 km suggesting the right-lateral strike-slip fault movement and at the Surin stream shows offset stream of ~3.5 and ~4.2 km (maximum offset ~7.7 km) also suggesting the right lateral strike-slip fault movement. Locations are shown in **Figure 2A**.

National Aeronautics and Space Administration (NASA), and most recent high-resolution images available in Google Earth date to 2021 were analyzed for identifying the morphotectonic landforms implying the possible active fault segments in the MHSF (**Figure 2**). Triangular facets, fault scarps, linear valleys, hot springs, offset streams, and shutter ridges were mapped.

Several offset streams in the study area occur along the MHSF (**Figures 3, 4**). These include the Samad stream at the Ban Pha Bong segment, Surin stream at the Ban Mae Surin segment, and Mae Han and Mae Salab streams in the Mae Sariang Basin with maximum displacements of 7.0, 7.7, 4.0, and 4.6 km, respectively. The lengths of the shutter ridges in the study area range from 150 to 300 m. Good examples are Ban Khun Yuam (150-m-long) in the middle part of the MHSF (**Figure 5B**) and Ban Mae Tha Lu (250-m-long) in the southern part of the MHSF (**Figure 5C**). Linear valleys are also present along the MHSF, such as at Ban Yod (5-m-long) in the north (**Figure 5A**) and Khun Yuam in the

middle stretches of the MHSF (**Figure 5B**). Triangular facets have height ranges from 1 to 5 m and their base varies from 2 to 10 m. Several triangular facets are present along the MHSF at Ban Yod, Khun Yuam, and Ban Mae Tha Lu (**Figures 5A–C**). Several fault scarps are present along the MHSF, particularly at Ban Yod and Khun Yuam fault segments.

4 PALEOEARTHQUAKE STUDY

4.1 Trench Excavations and Stratigraphy

4.1.1 Ban Yod Trench

The Ban Yod Fault trench is located (19°26'29"N and 97°57'41"E; **Figure 5A**) within the Ban Yod segment where small offset streams and fault scarps were observed. The trench was 19-m-long, 3-m-wide, and 3-m-deep. Seven depositional units (A–G) were exposed in both sidewalls (**Figure 6; Table 1**).

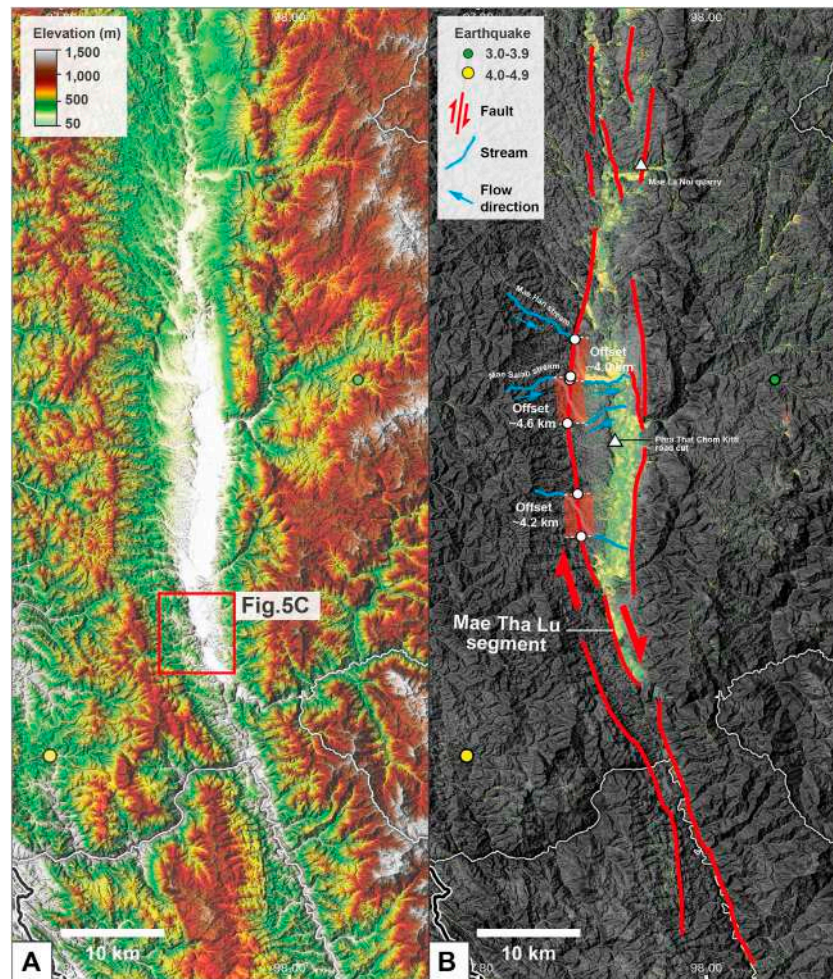


FIGURE 4 | Mae Sariang Basin on the south of the Mae Hong Son Province shows (A) enhanced ALOS DEM, and (B) detailed interpretation at the Mae Han stream displays offset stream with maximum offset of ~4.0 km suggesting the right lateral strike-slip fault movement and at the Mae Salab stream shows the offset stream of ~4.6 km suggesting the dextral movement. Locations are shown in **Figure 2A**.

Unit A is the oldest unit and is a 0.5–1.25-m-thick colluvium deposit composed of brownish-gray gravel with largely moderately sorted clast-supported subangular to rounded pebbles to cobbles with a matrix of fine sand, silt, and clay. The clasts are mainly sandstone, shale, and quartz. Unit B is a ~0.6-m-thick colluvial deposit consisting of moderately sorted clast-supported subangular to subrounded boulders, pebbles, sand, silt, and clay, mainly of sandstone and shale. Unit C is a ~0.1-m-thick fluvio-lacustrine deposit that contains a few gravels within a dark gray clayey silt matrix. The gravels are usually subangular sandstone fragments. This unit contains a lot of charcoal fragments. Unit D (av. 0.60-m-thick) is an alluvial unit with light gray clay with a fine sand layer. This unit shows laminations.

Unit E is a ~0.40-m-thick fluvial deposit of clast-supported gravel, sand, silt, and clay. Most clasts are subangular and composed mainly of sandstone, shale, and quartz. Unit F is a ~0.5-m-thick colluvial deposit within a channel and is composed of clast-supported subrounded to rounded gravel unit mainly

consisting of sandstone, shale, and quartz. At least three graded bedding are present, and the upper part of this unit contains dark clay. The succession is capped with dark topsoil which is unit G that has some sand and clay lens. Units A and C are displaced by an east-dipping dip-slip fault. Units E, F, G, and H were deposited and covered by unit I and topsoil Unit D.

4.1.2 Khun Yuam Trench

The Khun Yuam Fault trench is located at a frontal foothill (18°48'43"N and 97°56'33"E; **Figure 5B**), near Khun Yuam District. The 18-m-long, 3-m-wide, and 3.5-m-deep trench exposed 12 depositional units (A–L) in both of its walls (**Figure 7**; **Table 1**).

Unit A is a ~0.4-m-thick alluvial deposit with light brownish gray, mainly clay with a few gravels. Unit B is a ~0.2-m-thick alluvial/colluvial deposit of gravel, sand, and clay. The gravel is pebble size (1.5 cm). Unit C is a ~0.1-m-thick colluvial/alluvial deposit that contains light brown to yellowish-brown sand, silt, and clay, and some gray to purple. The sediments in this unit are

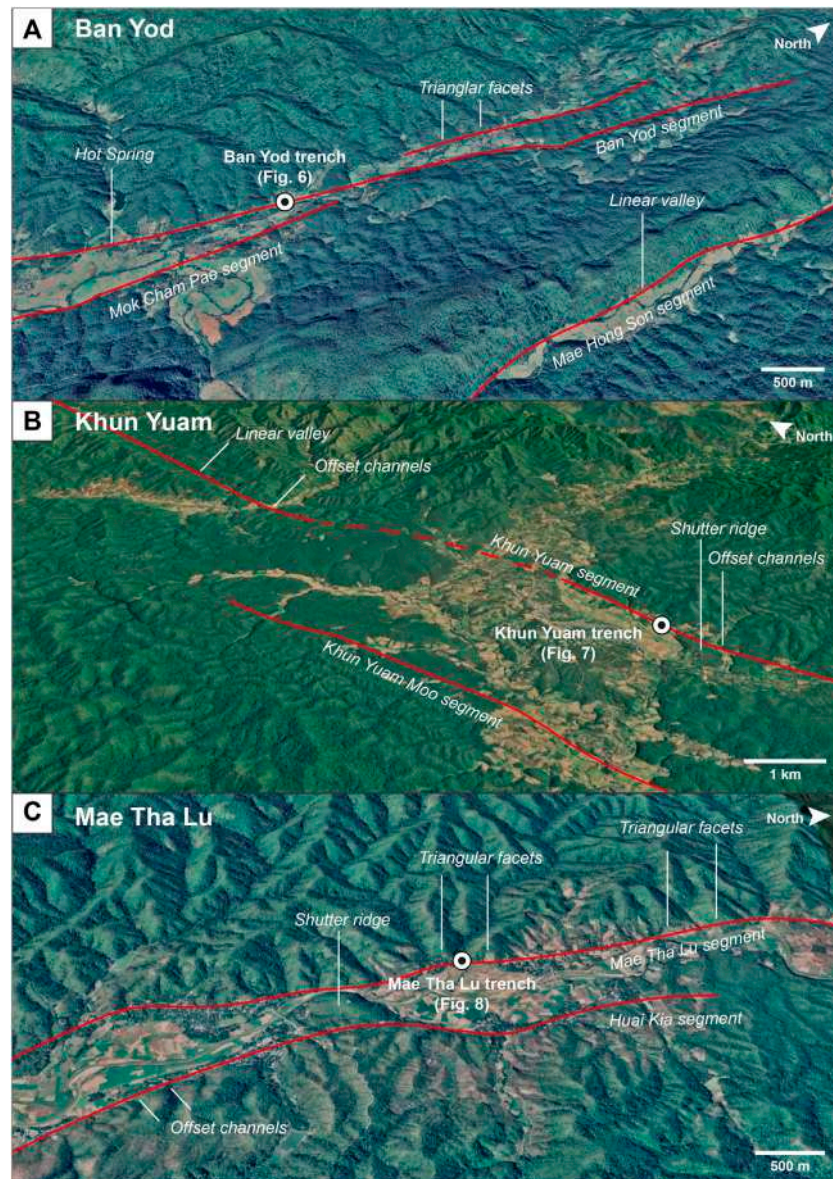


FIGURE 5 | Google Earth images show the location of the fault trenches and detailed tectonic geomorphic evidence at **(A)** the Ban Yod, **(B)** Khun Yuam **(B)**, and **(C)** Mae Tha Lu.

mainly clay with sand and sand lens. Unit D is a ~0.5-m-thick colluvial deposit that is composed of moderate to poorly sorted gravel, sand, silt, and clay. The clasts are subangular to subrounded pebbles to boulders.

Unit E is a ~0.5-m-thick fluvial channel deposit composed of sand, silt, and clay. The grain size is generally pebble of quartz and sandstone. At least three graded beds are present. Unit F is a ~0.2-m-thick alluvial deposit that contains orangish-brown sand, silt, and clay. Unit G is a ~0.6-m-thick alluvial channel deposit that contains yellowish-brown sand, silt, and clay with some gravel. The gravel size ranges from pebble to cobble. The lamination and chaotic structure could be found in the unit. Unit H is a ~0.2-m-

thick alluvial channel deposit that contains dark brown to brownish-gray sand and gravel. The gravel size ranges from pebble to cobble and the gravel shape is subangular to subrounded. The sediments in this unit are moderately well sorted.

Unit I is a ~0.25-m-thick alluvial that contains yellowish-brown silt, sand, and gravel. The sediments in this unit are mainly silt and sand with scattered gravel. The gravel size ranges from pebble to cobble and the gravel shape is angular to subrounded. The sediments in this unit are moderately well sorted. Unit J is a ~0.2-m-thick colluvial deposit that is composed of dark brown gravel, sand, silt, and clay. The gravel is matrix-supported and

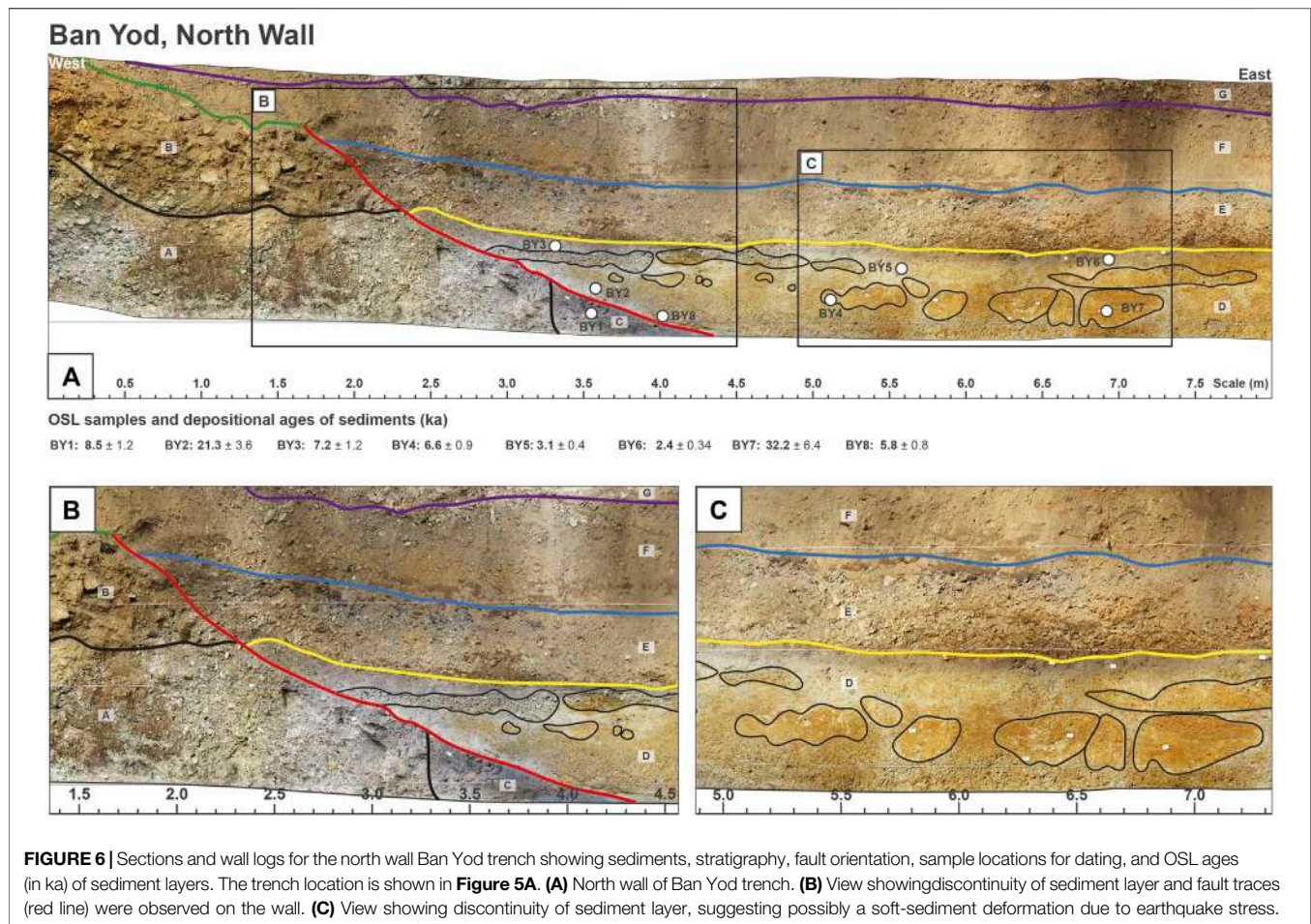


FIGURE 6 | Sections and wall logs for the north wall Ban Yod trench showing sediments, stratigraphy, fault orientation, sample locations for dating, and OSL ages (in ka) of sediment layers. The trench location is shown in **Figure 5A**. **(A)** North wall of Ban Yod trench. **(B)** View showing discontinuity of sediment layer and fault traces (red line) were observed on the wall. **(C)** View showing discontinuity of sediment layer, suggesting possibly a soft-sediment deformation due to earthquake stress.

mainly composed of poorly to moderately sorted angular to subrounded pebbles to cobbles up to 12 cm in diameter, mainly sandstone, shale, quartz, and granite.

Unit K is a ~0.35-m-thick alluvial deposit clast of weathered rock of sand and gravel, matrix: fine sand, clay. The succession is capped with soil that has laminated fine to very coarse sand. Sediment structures in unit F and unit G are soft-sediment deformation, they are mostly ball-and-pillow structures from unconsolidated sediment with low shear resistance because of tectonic or sedimentary processes (Allen, 1982). The soft-sediment deformation structures in the Khun Yuam trench are interpreted to be produced by seismic activity. Units A, B, C, and D are displaced by an oblique strike-slip fault (**Figure 7**).

4.1.3 Mae Tha Lu Trench

The Mae Tha Lu Fault trench is located ($17^{\circ}58'45''\text{N}$ and $97^{\circ}55'34''\text{E}$; **Figure 5C**) at a frontal foothill, near Mae Sariang District. The 14-m-long, 3-m-wide, and 3.5-m-deep trench exposed basement rock and nine depositional units (A–H) and topsoil in both walls of the trench (**Figure 8**; **Table 1**).

Unit A is a ~1.5-m-thick colluvial deposit that contains reddish-brown gravel and sand. These clasts are clast-supported and are subangular to subrounded sandstone and

shale. The sediment ranges from boulders to cobbles with fining upward succession. Unit B is a ~0.4-m-thick colluvial deposit that contains clay with gravel, almost fine grain, and gravel can be found in the west of the trench, matrix-supported, lateritic texture, more Fe and Mn concretion (black spot). Unit C is a ~0.25-m-thick colluvial deposit consisting of gravel, sand, silt, and gravel. The gravels ranging in size from pebble to cobble are fine-grained sandstone and shale, well-sorted to moderate-sorted (with more clay in the lower part). Unit D is a ~0.5-m-thick alluvial deposit composed of clay, silt with Fe or Mn concretion, and little gravel. Gravels mostly are quartz, with a reddish-brown lateritic texture.

Unit E is a ~0.3-m-thick alluvial deposit consisting mainly of sand and clay, with a few gravels present. Unit F is a ~0.6-m-thick alluvial deposit composed of matrix-supported, angular to subrounded gravel, sand, silt, and clay. Unit G is a ~0.5-m-thick colluvial deposit composed of, matrix-supported, angular to subrounded clasts are quartz and sandstone. Unit H is a ~0.2-m-thick alluvial deposit composed of dark brown sand, silt, and clay. The succession is capped with a ~0.3-m-thick dark brown soil unit, composed mainly of sand, silt, and clay with some roots. Unit A is a colluvial wedge associated with a 70° -dipping and ENE (062°) trending oblique strike-slip fault, with some gravels near

TABLE 1 | Description of stratigraphic units in trenches at Ban Yod, Khun Yuam, and Mae Tha Lu, as presented in **Figures 6–8**.

Ban Yod trench	
Unit	Description
G	Dark sandstone and clay lens; topsoil
F	Dark brown, clast-supported sandy gravel. Clasts are subrounded to rounded. The upper part of this unit contains dark clay; fluvial deposit
E	Brown, clast-supported sandy gravel. Most clasts are subangular; fluvial deposit
D	Light gray, matrix-supported silty sand, lamination, gravel lens, dark clay layer on the top, fining upward; alluvial deposit
C	Dark gray, matrix-supported clayey silt with scattered gravels. The gravels are subangular sandstone fragments; fluvio-lacustrine deposit
B	Brownish-gray, clast-supported with pebbly to boulder gravel. Clasts are mainly of subangular to subrounded sandstone and shale, moderately sorted (gravel size); colluvial deposit
A	Brownish-gray, clast-supported with pebbly to cobbly gravel. Most clasts are subangular to rounded with moderately sorted, fining upward; colluvial deposit
Khun Yuam trench	
L	Dark brown and gray, matrix-supported with silty sand, found roots; topsoil
K	Light brown, matrix-supported clayey sand with scattered weather rocks and gravel; alluvial deposit
J	Brownish-gray, matrix-supported gravelly sand with scattered pebbly to cobbly gravel. Most clasts are angular to subangular of shale, quartz and granite, poorly sorted; colluvial deposit
I	Yellowish-brown, matrix-supported silty sand with scattered pebbly gravel. Most clasts are angular to subrounded, moderately sorted; alluvial deposit
H	Dark brown to brownish-gray, clast-supported sandy gravel. Most clasts are subangular to subrounded. The gravel size ranges from pebble to cobble, moderately sorted; alluvial deposit (channel sediment)
G	Yellowish-brown, matrix-supported sandy silt with scattered pebbly to cobbly gravels, lamination and chaotic structure; alluvial deposit
F	Light brown, matrix-supported clayey sand with scattered pebbly gravel; alluvial deposit
E	Light brown to dark brown, sandy silt with pebbly gravel. Graded bedding at least three sequences; fluvial deposit (channel sediment)
D	Yellowish-brown, gravelly sand with scattered pebbly to boulderly gravel. Most clasts are subangular to subrounded, moderate to poorly sorted; colluvial deposit
C	Yellowish-brown with some gray to purple, silty clay with sand lens; alluvial deposit
B	Light brown, matrix-supported gravelly sand with scattered gravel. Most clasts are pebble, wedge shape; alluvial deposit
A	Light brownish-gray, silty clay with scattered gravel; alluvial deposit
Mae Tha Lu trench	
I	Dark brown, mainly sand, silt, and clay with some roots; topsoil
H	Dark brown, fine-grained sand, silt, and clay with scattered pebbles; alluvial deposit
G	Brown, matrix-supported with silt and sand. Clasts are angular to subrounded. Some roots can be observed here; colluvial deposit
F	Dark reddish-brown, matrix-supported sand, silt, and clay with scattered pebbles. Clasts are angular to subrounded; alluvial deposit
E	Reddish-brown, matrix-supported sand and silt with scattered pebbles; alluvial deposit
D	Reddish-brown, matrix-supported silty sand with scattered pebbles. Fe and Mn concretion with lateritic texture; alluvial deposit
C	Reddish-brown, clast-supported pebbly to cobbly gravel and sandy gravel, well- to moderate-sorted; colluvial deposit
B	Brown, matrix-supported silty sand with scattered pebbles. Fe and Mn concretion with lateritic texture, continually deposit from A3; colluvial deposit
A	Reddish-brown, clast supported pebbly and cobbly gravel with subangular to subrounded gravels. Boulders of conglomerate with bedding of gravel; colluvial deposit

the fault plane aligned at a high angle to bedding. Surface morphology such as triangular facets that are commonly found in this area supports dominant normal faulting.

4.2 Dating Results and Paleoearthquake Events

All 22 OSL samples were collected from sandy clay and silty sand units from the Ban Yod, Khun Yuam, and Mae Tha Lu

trenches (**Table 2**). The samples were analyzed in the Geochronology Laboratories at the University of Cincinnati. **Table 2** provides radioisotope, water content, cosmic dose rate, DR and DE values, and OSL ages of each sample, with explanations of the methods used to calculate dose rates and age calculation uncertainties. Dose rate calculations follow the details highlighted in the table's footnotes and are confirmed using the Dose Rate and Age Calculator (DRAC) of Durcan et al. (2015). The DR for all samples has

Khun Yuam, Noth Wall

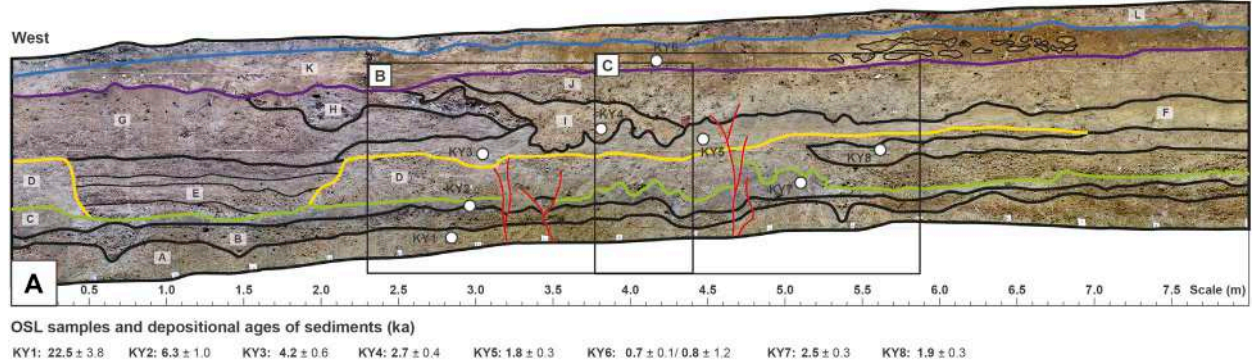


FIGURE 7 | Sections and wall logs on the north wall at Khun Yuam trench showing sediment stratigraphy, fault orientation, sample locations for dating, and OSL ages (in ka) of sediment layers. The trench location is shown in **Figure 5B**. (A) North wall of Khun Yuam trench (B) and (C) closed-up photographs showing discontinuity of sediment layer and fault traces (red line) were observed in the wall.

Mae Tha Lu, Noth Wall

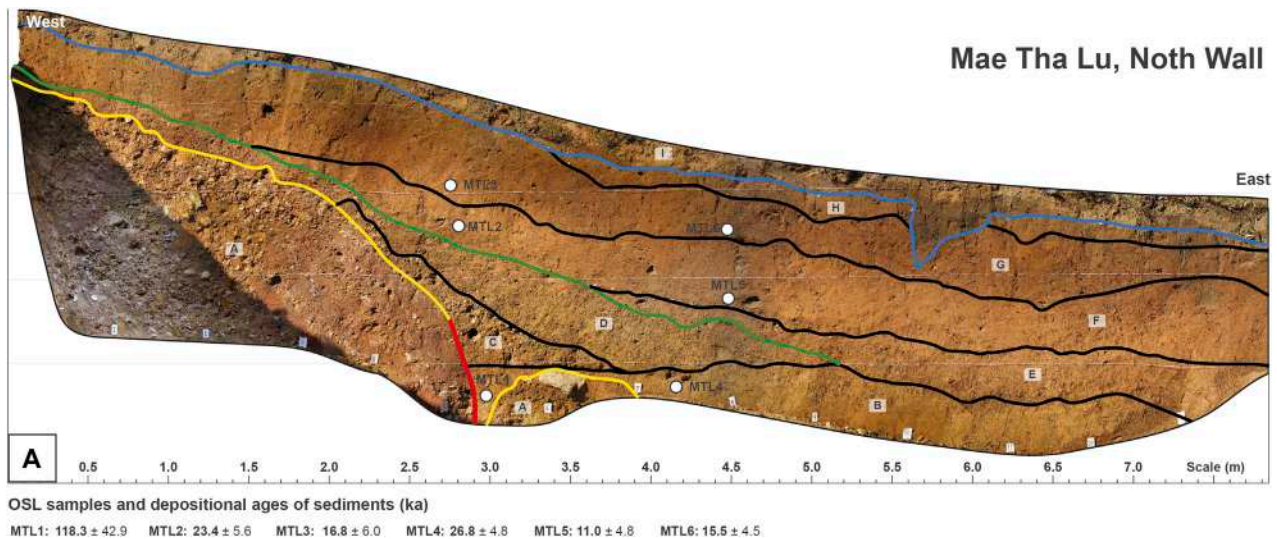


FIGURE 8 | Paleoseismic trench section and wall log section on the north wall at Mae Tha Lu trench showing sediment stratigraphy, fault orientation, sample locations for dating, and OSL ages (in ka) of sediment layers. The trench location is shown in **Figure 5C**.

TABLE 2 | OSL dating results of extracted sediments from the study area, Mae Hon Son Province, northern Thailand, containing radioisotope concentrations, moisture content total dose rate, equivalent dose estimates, and absolute ages.

Trench	Sample name	Unit	U ^a (ppm)	Th ^a (ppm)	K ^a (%)	Rb ^a (ppm)	Depth (m)	Latitude (°S)	Longitude (°W)	Altitude (m asl)	Cosmic dose rate ^{b,c} (Gy/ka)	Total equivalent dose rate ^{b,c,d} (Gy/ka)	Number of aliquots that past all test and uses in age calculations ^e	Dispersion (%)	DE arithmetic mean ^f (Gy)	Arithmetic mean age ^f (ka)	DE weighted mean ^h (Gy)	Weighted mean age ^h (ka)
BY	BY1	C	1.43	9.08	0.91	122.70	3.0	19.4416	97.9613	236	0.14±0.02	1.47±0.10	6	7	14.52±0.49	9.9±4.8	12.5±0.13	8.5±1.2
BY	BY2	D	1.53	7.73	0.68	107.50	2.5	19.4416	97.9613	236	0.15±0.02	1.29±0.09	6	8	32.19±1.25	25±14.2	27.48±0.32	21.3±3.6
BY	BY3	D	1.05	4.25	0.32	96.10	2.0	19.4416	97.9613	236	0.16±0.02	0.68±0.04	6	4	5.62±0.12	8.2±3.1	4.88±0.05	7.2±1.2
BY	BY4	D	0.70	8.52	1.45	75.30	2.5	19.4416	97.9613	236	0.15±0.02	1.63±0.12	24	18	12.82±0.48	7.9±4.0	10.69±0.11	6.6±0.9
BY	BY5	D	0.74	8.27	1.36	81.80	2.5	19.4416	97.9613	236	0.15±0.02	1.67±0.13	15	13	6.08±0.21	3.7±1.7	5.11±0.06	3.1±0.4
BY	BY6	D	0.18	7.30	1.39	90.80	2.3	19.4416	97.9613	236	0.15±0.02	1.43±0.11	17	15	4.03±0.17	2.8±1.6	3.37±0.03	2.4±0.3
BY	BY7	D	1.35	5.97	0.49	75.30	2.3	19.4416	97.9613	236	0.15±0.02	0.86±0.05	6	3	31.73±0.58	37.1±11.9	27.51±0.31	32.2±6.4
BY	BY8	D	0.50	8.31	1.40	102.30	2.8	19.4416	97.9613	236	0.14±0.02	1.55±0.12	24	11	10.53±0.26	6.8±2.3	9.02±0.09	5.8±0.8
KY	KY1	A	7.14	32.12	3.46	196.90	3.0	18.8120	97.9426	547	0.15±0.02	4.24±0.27	8	7	111.58±3.21	26.3±11.9	95.41±1.03	22.5±3.8
KY	KY2	C	6.73	27.98	2.08	139.40	2.5	18.8120	97.9426	547	0.16±0.02	3.37±0.21	6	7	24.80±0.79	7.3±3.7	21.38±0.21	6.3±1.0
KY	KY3	F	6.25	30.20	2.02	130.90	1.8	18.8120	97.9426	547	0.17±0.02	3.44±0.22	7	9	16.99±0.69	4.4±3.2	14.51±0.13	4.2±0.6
KY	KY4	I	5.52	20.97	3.52	193.20	1.5	18.8120	97.9426	547	0.18±0.02	4.17±0.30	6	7	13.18±0.42	3.2±1.4	11.37±0.11	2.7±0.4
KY	KY5	F	3.59	41.63	1.53	121.90	1.8	18.8120	97.9426	547	0.17±0.02	3.47±0.24	21	17	7.34±0.33	2.1±1.4	6.10±0.06	1.8±0.3
KY	KY6	K	3.23	38.07	4.44	183.30	0.5	18.8120	97.9426	547	0.21±0.02	4.78±0.33	20	45	5.16±0.76	1.1±2.3	3.41±0.04	0.7±0.1/ 0.8±1.2 ⁱ
KY	KY7	C	3.25	40.78	1.95	129.40	2.5	18.8120	97.9426	547	0.16±0.02	3.68±0.26	23	14	11.15±0.37	3.0±1.5	9.36±0.08	2.5±0.3
KY	KY8	D	3.10	30.71	1.30	121.10	2.0	18.8120	97.9426	547	0.17±0.02	2.79±0.19	21	8	6.32±0.14	2.3±0.7	5.43±0.05	1.9±0.3
MTL	MTL1	B	1.24	9.31	1.73	26.81	2.8	17.9793	97.9176	178	0.14±0.02	1.98±0.15	6	7	267.46±12.01	135.3±80.0	233.98±6.43	118.3±42.9
MTL	MTL2	F	1.80	15.55	1.93	25.61	1.3	17.9793	97.9176	178	0.17±0.02	2.93±0.24	6	3	78.93±1.86	26.9±7.8	68.63±1.34	23.4±5.6
MTL	MTL3	G	1.99	13.88	1.54	26.19	1.0	17.9793	97.9176	178	0.18±0.02	2.22±0.16	6	8	42.74±1.96	19.3±12.4	37.30±0.94	16.8±6.0
MTL	MTL4	B	1.09	19.03	3.11	25.37	2.5	17.9793	97.9176	178	0.15±0.02	2.97±0.22	20	9	92.86±2.27	31.2±10.4	79.70±1.05	26.8±4.8
MTL	MTL5	F	1.48	18.41	2.76	25.86	1.1	17.9793	97.9176	178	0.18±0.02	3.88±0.34	7	8	50.92±2.61	13.1±7.7	42.67±1.63	11.0±4.8
MTL	MTL6	G	1.80	13.84	1.44	26.45	1.4	17.9793	97.9176	178	0.17±0.02	2.23±0.16	6	3	39.63±1.09	17.8±6.6	34.41±0.74	15.5±4.5

^aElemental concentrations from gamma-ray spectrometry of whole sediment measured at Kasetsart University, Bangkok, Thailand.

^bEstimated fractional day water content for whole sediment is taken as 10% and with an uncertainty of ± 5%.

^cEstimated contribution to dose-rate from cosmic rays calculated according to Prescott and Hutton (1994). Uncertainty taken as ±10%.

^dTotal dose-rate from beta, gamma, and cosmic components. Cosmic dose rate was determined using Prescott and Hutton (1994). Beta attenuation factors for U, Th and K compositions incorporating grain size factors from Mejdahl (1979). Beta attenuation factor for Rb is taken as 0.75 (cf. Adamiec and Aitken, 1998). Factors utilized to convert elemental concentrations to beta and gamma dose-rates from Adamiec and Aitken, 1998 and beta and gamma components attenuated for moisture content. Dose rates calculation was confirmed using the Dose Rate and Age Calculator (DRAC) of Duncan et al. (2015).

^eNumber of replicated equivalent dose (D_E) estimates used to calculate D_E. These are based on recuperation error of < 10%. The number in the parentheses is the total measurements made including failed runs with unusable data. The number in square parentheses is the percentage of aliquots used for minimum age 2-mixing model.

^fAverage equivalent dose (D_E) determined from replicated single-aliquot regenerative-dose (SAR; Murray and Wintle, 2000) runs. The uncertainty is the standard error and includes an uncertainty from beta source estimated of ±2.5%.

^hWeighted average equivalent dose (D_E) determined from replicated single-aliquot regenerative-dose (SAR; Murray and Wintle, 2000) runs. The uncertainty is the standard error and includes an uncertainty from beta source estimated of ±2.5%.

ⁱAge based on minimum population in 2-mixing model using the program of Vermeesch (2009). 0.8 ± 1.2 age is based on minimal age from two mixing model.

^jUncertainty incorporate all random and systematic errors, including dose rates errors and uncertainty for the D_E.

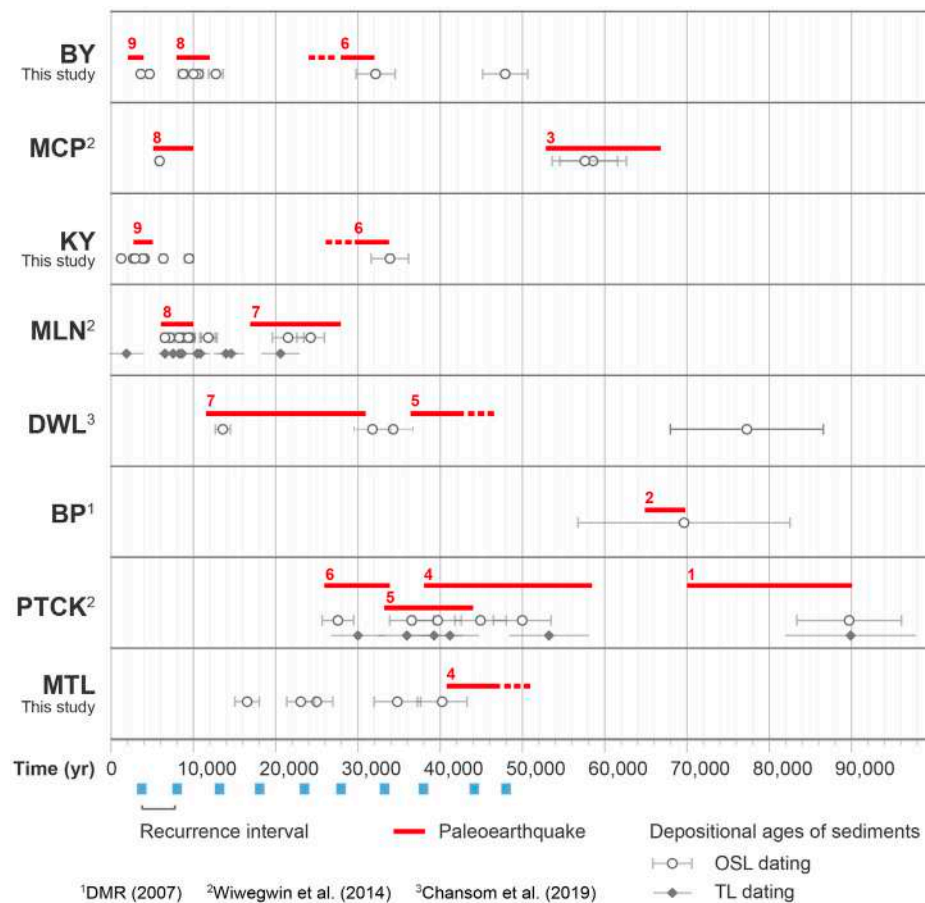


FIGURE 9 | Diagram shows the depositional ages of young sediments in the trenches from the OSL ages (in years BP) and interpreted paleoearthquake events in individual trenches, MHSF. Fault segments: Ban Yod (BY), Mork Cham Pae (MCP), Khun Yuam (KY), Mae La Noi (MLN), Doi Wiang Luang (DWL), Ban Pae (BP), Phra Tat Chom Kitti (PTCK), Mae Tha Lu (MTL).

TABLE 3 | Rates of right-lateral strike-slip faulting along the Mae Hong Son Fault.

Lateral slip rate: Ban Yod segment					
Measure	Mean	Mode	Median	68.27% Interval	95.45% Interval
Offset (m)	4	4	4	+0.1/−0.1	+0.2/−0.2
Age (ka)	3	3	3	+1/−1	+0.2/−0.2
Slip rate (mm/yr)	1.5	1.1	1.3	+0.7/−0.3	+2.4/−0.5
Lateral slip rate: Mae Tha Lu segment					
Offset (m)	6.8	10	6.6	+3.3/−2.6	+3.4/−2.8
Age (ka)	48	48	48.1	+10/−10.1	+20/−20
Slip rate (mm/yr)	0.1	0.1	0.1	+0.1/−0.1	+0.2/−0.1

a value that varies from 0.64 to 4.78 Gy/ka within the normal range for terrestrial sediments. The water contents of samples are between 16 and 69%.

The spread of DE for the most samples with low dispersion (< 20%), we use the weighted mean value for the DE values (Table 2). However, there is one sample (sample name: KY6)

where the spread of DE was relatively large for this sample (dispersion > 20%), suggesting possible partial bleaching. This can result in an overestimate of age. For this sample, we assumed a 2-mixing model using the Radial Plotter (Vermeesch, 2009) and calculated the age based on the younger population. The OSL dates of the sediment are between ~0.7 and 118 ka.

Eight samples were collected for OSL dating (BY1–8) in the Ban Yod trench (**Figure 6**). The OSL ages indicated that the sediments in the trench wall were deposited since 32.2 ± 6.4 ka. Based on the stratigraphic unit and OSL ages, three earthquake events occurred at ~30, 8, and 3 ka.

Eight samples (KY1–8) were collected for OSL dating from the Khun Yuam trench (**Figure 7**). The OSL ages indicated that the sediments in the trench wall were deposited since 22.5 ± 3.8 ka and earthquakes occurred at ~30 and 3 ka.

Six samples were collected for OSL dating (MTL1–6) from the Mae Tha Lu trench (**Figure 8**). The OSL ages indicated that the sediments in the trench wall were deposited since 118 ± 43 ka and an earthquake occurred older than the oldest sediment age (unit B) at 118 ± 43 ka as the unit B deposited after the observed fault cut the unit A.

5 DISCUSSION

5.1 Paleoearthquake Events and Recurrence Interval

The alignment of geochronological data from this study and previous works (Department of Mineral Resources, 2007; Wiwegwin et al., 2014; Chansom et al., 2019) suggests that there have been nine earthquakes occurred by the MHSF. All earthquake events with dating results are described later in chronological order. It should be noted that our results are included since the fourth earthquake (**Figure 9**).

The first earthquake is related to the offset sedimentary layers on the Phra That Chom Kitti segment (**Figure 9**), as reported earlier by Wiwegwin et al. (2014). The fault cuts a sedimentary unit that was OSL and TL dated by Wiwegwin et al. (2014) to 89.7 ± 6.5 ka, and the movement must be younger than this unit. Based on the ages by Wiwegwin et al. (2014), the timing of the event is roughly constrained as having occurred at ~78 ka.

The second earthquake is intimately related to a movement on the Ban Pae segment according to Department of Mineral Resources (2007). The fault cuts through a sedimentary unit that was deposited at 69.7 ± 1.3 ka based on the OSL age (**Figure 9**; Department of Mineral Resources, 2007). Therefore, the timing of the earthquake event was reinterpreted as having occurred at ~68 ka (Wiwegwin et al., 2014).

The third earthquake is demonstrated by the movement of the Mok Chum Pae segment. The fault cuts sediment layers deposited during the period of 58.7–57.7 ka based on the OSL age of Wiwegwin et al. (2014). They interpreted the timing of the event that might have occurred at ~58 ka.

The fourth earthquake is determined by the movement on the Phra That Chom Kitti segment (Wiwegwin et al., 2014). The fault cuts the lower part of sediment layers deposited during the period of 53.3 ± 4.8 to 41.2 ± 3.5 ka (not cutting the upper part). This event was also evident on the Mae Tha Lu segment and is related to a movement of the fault cutting unit A prior to the sediment layers deposited at 26.3 ± 4.8 ka (unit B). Thus, the timing of the earthquake event might have occurred during the period of ~53.0–41.2 ka.

The fifth earthquake is evident from the offset recognized on the Doi Wiang Luang segment (Chansom et al., 2019). The fault cuts through a sedimentary unit (unit B) that was deposited at 77.3 ± 9.4 ka and cuts through units B–E. The movement must be younger than unit F which was deposited at 34.3 ± 2.4 ka. Thus, based on the OSL ages, it is possible that this fault was active at ~34.3 ka. This paleoearthquake event was also recognized in the Phra That Chom Kitti segment that was reported by Wiwegwin et al. (2014).

The sixth earthquake event is visible from the offset sedimentary layers on the Phra That Chom Kitti segments (Wiwegwin et al., 2014). The fault cuts the lower part of the sediment deposit at 27.6 ± 2.0 ka. Therefore, based on OSL age data of this study and Wiwegwin et al. (2014), it is likely that this fault might occur at ~30 ka.

The seventh earthquake is related to the offset sedimentary layers on the Mae La Noi segment and Doi Wiang Luang segment as reported by Wiwegwin et al. (2014) and Chansom et al. (2019), respectively. Based on the OSL ages of Wiwegwin et al. (2014) and Chansom et al. (2019), it is possible that this fault was active before ~22 ka.

The eighth earthquake occurred at ~8.0 ka. The event is related to the offset sedimentary layers on the Ban Yod segment, Mae La Noi segment, and Mok Chum Pae segment (Wiwegwin et al., 2014).

The last earthquake occurred at ~3.0 ka, as confirmed by our OSL dating at the Khun Yuam and Ban Yod segments. However, evidence for this youngest faulting event is clear in the trench.

From the dating of these past earthquakes, the recurrence interval of the MHSF is approximately ~5 ka (**Figure 9**). This refines the previous studies such as Wiwegwin et al. (2014) that estimated the recurrence interval of the MHSF to be ~10 ka.

5.2 Determination on Slip Rates

This study determined both lateral and vertical slip rates from the probability distribution of the displacement of the fault and the age of the earthquakes using the MATLAB code of Zechar and Frankel (2009) to calculate the slip rate and propagate the uncertainties in the displacements and age earthquakes as found in the trench. For the lateral slip rate determination, we used five offset streams at the southern terminus of the Ban Yod segment of the MHSF, with an offset of 4 ± 0.1 m and a latest earthquake event age of 3 ± 1 ka to yield a slip rate of $1.3 + 0.7/-0.3$ mm/a. (**Table 3**). Four offset streams were found at the Mae Tha Lu segment with an offset of $6.6 + 3.3/-2.6$ m and an OSL age of 48 ± 10 ka, yielding a slip rate of 0.1 ± 0.1 mm/a.

For the vertical slip rate, using the vertical offsets at the Khun Yuam trench is ~20 cm and the Doi Wiang Luang road cut exposure is ~75 cm and the age of a recurrence interval of the MHSF is ~5,000 year. This yields the vertical slip rates on the Khun Yuam segment and the Doi Wiang Luang segment of 0.04 and 0.15 mm/a, respectively. The average vertical offset from this study is constant with the average vertical slip rate of Wiwegwin et al. (2014) who estimated that the slip rate is ~0.03–0.13 mm/a on the Mae La Noi segment no.1 and the Phra That Chom Kitti segment. In addition, the estimated slip rate reported by

Department of Mineral Resources (2007) is ~ 0.0028 mm/a on the Ban Pae segment no. 31, this slip rate is less than Wiwegwin et al. (2014), and this study.

Although the MHSF is almost parallel to and has the same sense of movement as the Sagaing Fault, the slip rate of the MHSF seems to be much smaller than that of the Sagaing Fault (18 mm/a). This is on the ground that the MHSF is an inland fault that is far away from the active plate boundary whereas the Sagaing Fault represents the plate boundary between the Western Burma block and the Sibumasu block (Aung, 2005; Searle and Morley, 2011). On the other hand, the Mae Chan Fault, which lies within the complex collisional zone between the Indian and the Eurasian plates, has a slip rate of 1–2 mm/a (Weldon et al., 2016), higher than that of the MHSF. We interpreted that this is mainly due to the intraplate clockwise rotation of the mainland SE Asia block becoming larger than that of the northern part of the Shan-Thai (Sibumasu) block located more southward. The left-lateral Mae Chan Fault and other associated Nam Ma and Meng Xing faults may have accommodated and related the dextral movement of the MHSF.

5.3 Maximum Credible Earthquake

Maximum credible earthquakes (MCEs) are estimated using the fault rupture length at the surface (SRL) based on the equation proposed by Wells and Coppersmith (1994). The SRL used for the MCE calculation is the length of the longest fault segment in individual fault segments which were interpreted in Chansom (2019). The fault segments belonging to the MHSF have a wide range from 3.03 to 28.55 km. Therefore, these observations demonstrate that the dominant mode of slip across the MHSF is unstable sliding in the credible earthquakes with a maximum of Mw5.6–Mw6.8. The MCEs of the MHSF seem to be lower than those of the Mae Chan Fault–Jing Hong Fault [$\sim$ Mw7.1, Xie and Tsai (1983)], and the Sagaing Fault system ($\sim$ Mw 7.3; Wang et al., 2014). We consider that the MHSF consists of chopped (or broken) fault segments than those of the two fault systems. The broken geometry of the MHSF is probably large to the fault small accumulation of the slip (Wesnousky, 1988; Stirling et al., 1996; De Jossineau and Aydin, 2009). This broken fault is quite similar to the Sumatra Fault (Wang et al., 2018).

5.4 Influence of Pre-Existing Fabrics on Fault Kinematics and Basin Geometry

The MHSF is located on the boundary between the Inthanon zone and the Sibumasu block (Hisada et al., 2004), where a narrow N-trending weak zone of the Mae Sariatang Trough was closed and the Inthanon zone thrust over the eastern margin of the Sibumasu block (see review in Morley, 2018). The continuous collision between the Eurasian and Indian plates during the Oligocene may have caused the reactivation of this pre-existing fabric under a N–S to NNW–SSE plate motion (Shi et al., 2018). According to ALOS DEM data, two separated N–S trending basins are present along the MHSF (Figure 2). Their basin geometries reflect the strong influence of the pre-existing fabrics under dextral strike-slip-

dominated deformation. The MHSB is located to the north, it is ~ 36 km long and 7 km wide (l/w ratio = 5.16) with its lowest elevation at 180 m above the mean sea level (amsl). The basin's depocenters widen southward, separated by intrabasinal high. Many sets of offset streams and linear valleys suggest that faults within the basin are potentially tectonically active. Although Pailoplee and Charusiri (2016) suggested that large earthquakes are not likely in this region.

The Mae Sariatang Basin (MSB) is located to the south with the lowest elevation at 150 m amsl. The basin is a spindle-shaped pull-apart, some 65 km long and 9 km wide (l/w ratio 7.23). The MSB represents pure strike-slip motion under a slip vector parallel to the underlapping NNW–N-trend fault segments, for example, the Mae Sariatang fault splay of the Mae Ping fault zone (Smith et al., 2007), whereas the MHSB is a pull-apart basin formed under an oblique relative motion to the basin axis. The MHSB and MSB are structurally separated by Khun Yuam High (Figure 2) where vertical displacement decreases. There is no thoroughgoing fault formed between the two basins indicating a low amount (a few km) of dextral strike-slip displacement of the MHSF. Outcrop study along the Mae Ping fault zone suggests minor dextral deformation probably in the order of kilometers (Smith et al., 2007). Therefore, the MHSF is currently undergoing episodic, low-strain-rate dextral motion, with crustal deformation resulting in the formation of MHSB and MSB.

6 CONCLUSION

We draw the following conclusions:

- 1) The MHSF is mainly a right-lateral strike-slip fault trending north with some northwest and northeast direction traces and landforms along the fault including offset streams, triangular facets, shuttle ridges, linear valleys, and fault scarps.
- 2) The MHSF has produced at least nine earthquakes over the past ~ 43.0 ka in the Holocene, yielding the recurrence interval of ~ 5.0 ka. The maximum credible earthquake of the MHSF range from Mw 5.6 to 6.8, estimated by the surface rupture length of about ~ 3.0 – 28.5 km, which results in the slip rate of ~ 0.04 – 0.15 mm/a.
- 3) The MHSF has a low amount (a few km) of dextral strike-slip displacement. The north-trending MHSB and MSB in the northern and southern parts of the MHSF are structurally separated by Khun Yuam High with no geomorphic evidence of thoroughgoing large-displacement strike-slip fault segment between the basins.
- 4) The broader implications of this study include the low rates of slip on the MHSF and scattered small earthquakes may be common yet difficult to identify with seismologic observations. Therefore, predictions of the anomalous seismogenic behavior of MHSF are not consistent, which suggests that it is neither weak and creeping, nor does it generate anomalously large earthquakes with exceptionally long recurrence intervals. Rather, our results suggest

that the seismogenic behavior of the MHSF is very similar to other more optimally oriented faults in northern Thailand.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/supplementary material; further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

CC, WW, and PC contributed to the fieldwork and paleoearthquake study. CC and LO contributed to earthquake dating. SJ performed structural interpretation, basin analysis, and produced figures of the manuscript. CC, SJ, and PC wrote the first

draft of the manuscript. All authors contributed to manuscript revision, read, and approved the submitted version.

FUNDING

This research was supported by grants from the Graduate School collaborate with the Faculty of Science, Chulalongkorn University.

ACKNOWLEDGMENTS

We are grateful to Dr. Paula M. Figueiredo and Dr. Peerasit Surakietchai who aided with OSL dating. We also thank Suwith Kosuwan and Tassana Jadeanan who aided with the trenching and discussions. We appreciated helpful comments and suggestions from an reviewer that led to improvements in this study.

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 09 June 2022

ACCEPTED 01 July 2022

PUBLISHED 22 July 2022

CITATION

Song S, Li Y and Hao M (2022), Active
crustal deformation model of the
Fen–Wei rift zone, North China:
Integration of geologic, geodetic, and
stress direction datasets.
Front. Earth Sci. 10:964800.
doi: 10.3389/feart.2022.964800

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Active crustal deformation model of the Fen–Wei rift zone, North China: Integration of geologic, geodetic, and stress direction datasets

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The Fen–Wei rift zone (FWRZ) of North China is an important zone of active crustal deformation representing a transition from extrusion tectonics related to the Tibetan Plateau to subduction tectonics related to the potential far-field influence of the west Pacific plate. In this study, we determined the kinematic constraints of active crustal deformation in the FWRZ, which are fundamental for forecasting seismicity. NeoKinema, a kinematic finite-element model, was employed to estimate the long-term fault slip rates, distributed crustal deformation field, and on- and off-fault strain-rate fields in the FWRZ by fitting updated geological fault slip rate, geodetic GPS velocity, and principal compressive stress direction datasets. Our results show that the FWRZ is a characteristic low-strain kinematic setting, with most active faults exhibiting slip rates of less than 1 mm/a. The total sinistral shear rate from the southern Ordos block to the Qinling Mountains is approximately 1 mm/a, indicating limited tectonic extrusion along the EW-trending Qinling Mountains. Additionally, the central Shanxi rift exhibits prominent dextral shear of ~0.5 mm/a that decreases toward its north and south ends, corresponding to crustal extension of 1.1–1.2 mm/a in the Datong and Yuncheng basins, respectively. However, this significant crustal extension cannot be solely attributed to terminal effects caused by dextral shear in the central Shanxi rift. A comparison between predicted seismicity and historical earthquake records reveals some remarkable seismic gaps, particularly in the Datong, Hancheng, and Yuncheng basins, indicating higher seismic potential in these locations. This study provides insights into the long-term crustal deformation processes and regional seismic potential of the FWRZ.

KEYWORDS

Fen–Wei rift zone, long-term fault slip rate, strain partitioning, seismicity forecast, neokinema

1 Introduction

According to the active-block theory of tectonics in continental interiors (Deng et al., 2003; Zhang et al., 2003), active crustal deformation is the product of long-term (on a timescale of 10^4 years) tectonic loading since the Late Quaternary. This not only includes discontinuous deformation on the boundaries of active blocks (terrane) recorded by river/land offsets but also the motion and internal deformation of the terrane itself (Deng et al., 2003; Zhang et al., 2003). Determining the detailed kinematic constraints of active crustal deformation is

key for understanding continental dynamics (Zhang et al., 2004; Bird, 2009) and predicting the long-term risk of strong earthquakes, which can then contribute to providing quantitative seismic risk assessments (Bird and Kagan, 2004).

The Fen-Wei rift zone (FWRZ) in North China, which extends for approximately 1,200 km, is a famous Cenozoic intra-continental rift zone within the Asian continent (Figure 1). The FWRZ lies between two latitudinal active tectonic belts: the Qinling Mountains to the south and the Yanshan Mountains to the north (Figure 1). The former is recognized as the important pathway of “extrusion tectonics”

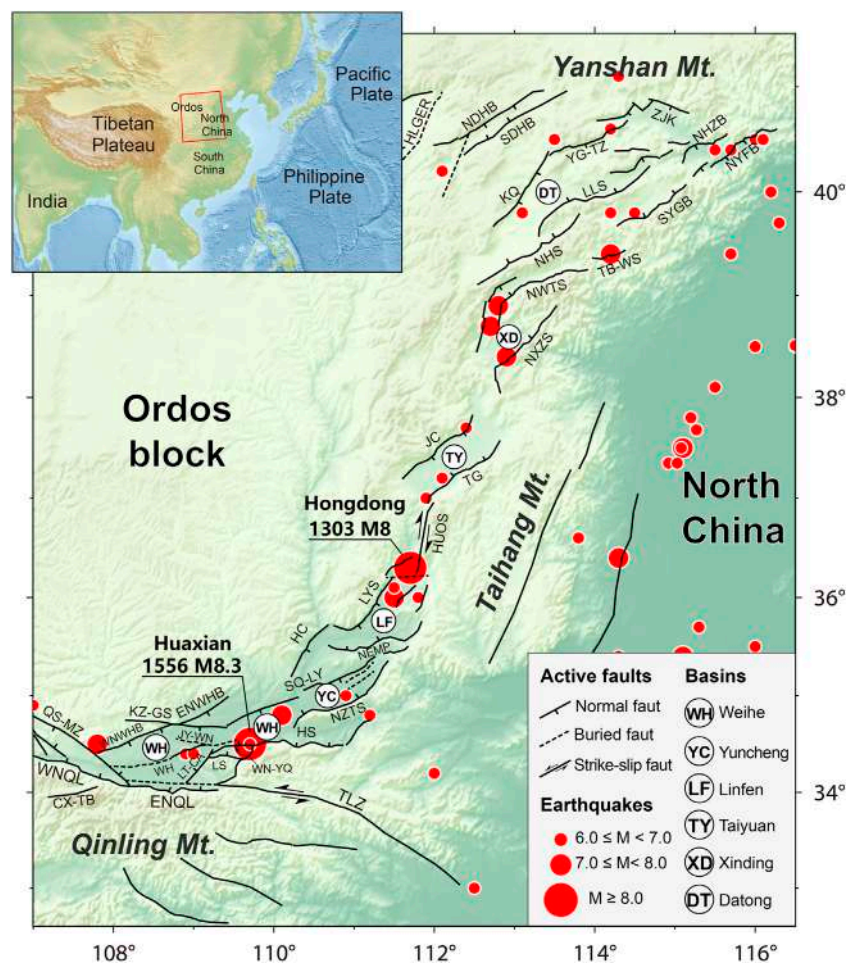


FIGURE 1

Distribution of active faults and strong earthquakes in the FWRZ and its adjacent regions. The fault data mainly refer to Deng et al. (2002) and Xu et al. (2016). And the seismic data is cited from the National Earthquake Data Center (<http://data.earthquake.cn>). Fault abbreviations are CX-TB, the Chengxian-Taibai Fault; QS-MZ, the Qishan-Mazhao Fault; WNQL, the Western segment of North Qinling Fault; ENQL, the Eastern segment of North Qinling Fault; HS, the Huashan Fault; NZTS, the North Zhongtiaoshan Fault; HC, the Hancheng Fault; TLZ, the Tieluzi Fault; LT-CA, the Lintong-Changan Fault; WH, the Weihe Fault; WN-YQ, the Weinan-Yuanqian Fault; JY-WN, the Jingyang-Weinan Fault; LS, the Lishan Fault; NEMP, the North Fault of Emei Platform; SQ-LY, the Shuangquan-Linyi Fault; KZ-GS, the Kouzhen-Guanshan Fault; WNWHB, the Western segment of North Weihe Basin Fault; ENWHB, the Eastern segment of North Weihe Basin Fault; LYS, the Luoyunshan Fault; HUOS, the Huoshan Fault; JC, the Jiaocheng Fault; TG, the Taigu Fault; NXZS, the North Xizhoushan Fault; NWTS, the North Wutaishan Fault; NHS, the North Hengshan Fault; LLS, the Liulengshan Fault; KQ, The Kouquan Fault; YG-TZ, the Yanggao-Tianzhen Fault; ZJK, the Zhangjiakou Fault; NHZB, the North Huaizhuo Basin Fault; NYFB, the North Yangfan Basin Fault; SYGB, the South Yuguang Basin Fault; TB-WS, the Taibai-Weishan Fault; NDHB, the North Daihai Basin Fault; SDHB, the South Daihai Basin Fault; DQS, the Daqingshan Fault; HLGER, the Helinge'er Fault.

induced by eastward expansion of the Tibetan Plateau (Zhang et al., 1995; Tapponnier et al., 2001). Conversely, the latter is part of the roughly EW-trending Zhang-Bo seismic zone caused by the far-field effects of western Pacific subduction (Zhang et al., 2018; Hao et al., 2020). Thus, constraining active crustal deformation in the FWRZ can provide insights into the transition between these two tectonic domains. Moreover, previous studies have shown that the FWRZ corresponds to transtensional tectonics during the Late Quaternary (Xu et al., 1993; Deng et al., 1999). However, the spatial pattern of strain partitioning has not been clarified, resulting in substantial debate over the dynamic tectonic mechanism (Xu et al., 1992; Deng et al., 1999; Zuza and Yin, 2016; Qiu and Qiao, 2017; Liu et al., 2021). Furthermore, intense continuous tectonic loading since the Late Quaternary has induced frequent and intense seismicity in the FWRZ (Figure 1), indicating the potential for strong earthquakes in this region (Work Group of M7, 2012). Thus, constraining active crustal deformation in the FWRZ can help elucidate several important issues.

Typical kinematic data used to constrain crustal deformation include fault slip rates, GPS velocity fields, and principal stress directions. Although the geological fault slip rate represents the average rate of long-term crustal deformation across an active fault, measurement errors related to river/land offsets and chronology increase the uncertainty of estimates. Moreover, it is difficult to quantitatively describe block (terrane) motion and internal deformation using only traditional geological methods. Alternatively, geodetic GPS data can provide high-precision constraints of crustal deformation with flexible spatial scales and can be a useful supplement in active tectonic studies in the FWRZ (Wang and Shen, 2020; Hao et al., 2021). However, it is uncertain whether present-day crustal deformation revealed by geodetic GPS (on the timescale of several decades or within one seismic cycle) can represent long-term crustal deformation processes (on a timescale of 10^4 years or several seismic cycles). Furthermore, geodetic GPS limited in its ability to precisely determine the activity for a single fault with a low slip rate (<1 mm/a) because low slip rates may be concealed by the GPS observational error (Mazzotti and Gueydan, 2018). This is pertinent to the FWRZ because active tectonic surveys have revealed a large number of active faults with a slip rate of <1 mm/a (Deng et al., 1994; Xu, 2002; Xu et al., 2011; Luo et al., 2020).

Previous studies have shown that the interseismic GPS velocity field is approximately equivalent to the long-term crustal deformation (Zhang et al., 2004; Liu and Bird, 2008; Carafa et al., 2020; Gan et al., 2021). Furthermore, Middleton et al. (2017) demonstrated that the long-term crustal extension rate across the northern Shanxi rift constrained by geological fault slip rates is consistent with regional GPS data. Thus, joint fitting of the various kinematic constraints on crustal deformation in the FWRZ, such as geological and

geodetic datasets, could provide an optimal solution of active crustal deformation that includes the long-term fault slip rate, distributed crustal deformation field, and strain-rate field (Bird, 2009). Such a method can exploit the advantages of different datasets in revealing long-term crustal deformation. That is, geological fault slip rates can compensate for the inability of GPS to detect faults with a low slip rate (<1 mm/a), whereas GPS compensates for the inability of geological data to precisely detect crustal deformation on a regional scale (Khodaverdian et al., 2016; Yang et al., 2021). Moreover, this joint fitting method has been successfully applied to continental active tectonic zones, including low-strain active rifting belts in central Italy, which have a similar active tectonic background to the FWRZ (Carafa et al., 2020).

Therefore, in this study, we employed NeoKinema (Bird and Liu, 2007), a kinematic finite-element model, to integrate various kinematic datasets of the FWRZ, including updated geological fault slip rates, geodetic GPS velocity fields, and principal compressive stress directions. These data were then used to construct a detailed active crustal deformation model of the FWRZ, which comprises long-term slip rates of the most active faults, the long-term crustal deformation field, and strain-rate fields (including on-fault and off-fault fields). The following specific questions were addressed in this study: 1) What are the overall characteristics of long-term fault activity in the FWRZ? 2) Does significant tectonic extrusion exist along the EW-trending Qinling Mountains? 3) What is the distribution of shear and extensional strain along the Shanxi rift? 4) Which areas exhibit strong earthquake potential in the FWRZ?

2 Active tectonic background

The NE–NEE trending FWRZ comprises a series of linear rifted basins with a left-stepping en-echelon distribution (Figure 1). The results of deep geophysical exploration, such as the uplifted Moho surface and high heat flow, indicate that the FWRB exhibits clear characteristics of intra-continental rifting (Jing et al., 2005; Tang et al., 2010; Li et al., 2014). The FWRZ predominantly consists of the Weihe and Shanxi rifts (Figure 1). The former is located between the southern margin of the Ordos block and the Qinling Mountains and is connected to the east with the Linfen Basin of the Shanxi rift (Figure 1). Previous studies have illustrated that the North Qinling Fault controls activity along the Weihe rift (Jing et al., 2005; Li et al., 2016). The North Qinling Fault has shown strong normal dip-slip activity since the Late Quaternary, with stronger faulting on its eastern segment than on its western segment (Figure 1). To the east, the Huashan Fault is considered to represent the seismogenic structure of the 1,556 Huaxian $M_{8\frac{1}{4}}$ earthquake (Shaanxi Earthquake Agency, 1996). Although a series of normal dip-

slip faults have also developed in the northern margin of the Weihe rift, their activity is generally weaker than that of the North Qinling Fault (Xu et al., 1988). According to left-lateral river/land offsets along the Tieluzi Fault in the eastern Qinling Mountains (Figure 1), the Qinling Mountains have been interpreted as a pathway of eastward lateral extrusion of the Tibetan Plateau (Zhang et al., 1995; Tapponnier et al., 2001). However, previous estimates of the left-lateral strike-slip rate on the Tieluzi Fault show notable discrepancies (Zhang et al., 1995; Zhou et al., 2001). For example, Yang et al. (2005) reported that the left-lateral strike-slip rate of the Tieluzi Fault varied from 0.5 to 1.25 mm/a in the Late Pleistocene.

The NE-trending Shanxi rift marks the eastern boundary of the Ordos block, which consists of the Yuncheng Basin, the Linfen Basin, the Taiyuan Basin, the Xinding Basin, and the Datong Basin (Research Group on Active Fault System around the Ordos, 1988; Wang, 1996) (Figure 1). Xu et al. (1993) emphasized that transtensional tectonics since the Late Quaternary explains the dynamic deformation background of fault activity and the frequent occurrence of strong earthquakes in the Shanxi rift (Figure 1). This kinematic model shows that the central part of the Shanxi rift is dominated by right-lateral shear, which contributes to normal faulting and crustal extension in the north and south ends of the Shanxi rift, respectively (Figure 1). According to historical records, seven destructive earthquakes with $M \geq 7$ have occurred in the Shanxi rift, including the Hongdong 1303 M 8.0 earthquake, which caused the death of 20 million people (Xu and Deng, 1990; Song, 2011; Xu et al., 2018). Previously reported normal dip-slip rates for the most active faults in the Shanxi rift are generally <1 mm/a for the Late Quaternary or Holocene (Deng et al., 1994; Dou and Yu, 1996; Jing et al., 2005; Xu et al., 2013; Middleton et al., 2017; Sun, 2018; Luo et al., 2020), and right-lateral strike-slip rates do not typically exceed 2 mm/a (Xu et al., 2014; Sun, 2018). Furthermore, the estimated horizontal extensional rate in the NW-trending Datong Basin (less than 1–2 mm/a, Zhao et al., 2017) is approximately compatible with the long-term extension rate determined by geological data (Middleton et al., 2017).

3 Kinematic model

3.1 Model theory

In this study, we used NeoKinema, a kinematics finite-element program (Bird and Liu, 2007; Carafa et al., 2020) to jointly fit the most recent kinematic datasets (geological slip rate, geodetic velocity, and stress direction) using a weighted least-squares method and then determine the optimal long-term fault slip rates and horizontal deformation field in the FWRZ and its surrounding areas. The objective function of the NeoKinema model is expressed as

$$\Pi = -\left(\bar{\vec{p}} - \bar{\vec{r}}\right)^T \left[\bar{\vec{C}}_{GPS}^{-1}\right] \left(\bar{\vec{p}} - \bar{\vec{r}}\right) - \frac{1}{L_0} \sum_{m=1}^M \int_{length} \frac{(\vec{p}_m - \vec{r}_m)^2}{\sigma_m^2} dl - \frac{1}{A_0} \sum_{n=1}^3 \iint_{area} \frac{(\vec{p}_n - \vec{r}_n)^2}{\sigma_n^2} da \quad (1)$$

In the first term, $\bar{\vec{p}}$ represents the predicted geodetic velocities, and $\bar{\vec{r}}$ represents the observed geodetic velocities and its covariance matrix $\bar{\vec{C}}_{GPS}$. The second term involves the geological long-term slip rates, $\vec{r}_m$, with corresponding errors, σ_m ; and $\vec{p}_m$ represents the predicted offset rates (m indicates the number of slip rates). The third term concerns the areas and directions of the distributed permanent deformation rate tensors of unfaulted (continuum) elements in the model. $\vec{p}_n$ represents the strain rate of the model.

In NeoKinema, some constraints (geological slip rates) are considered to apply along the fault traces, whereas other constraints (minimization of strain rate and isotropy) are applied across the unfaulted (continuum) elements. In the equation, n represents three types of constraints: $n = 1$ is the microplate constraint, whereby the target strain rate in the unfaulted continuum is assigned to be zero, with a statistical uncertainty. For regions where unknown faults might be buried or overlooked, a larger standard deviation can be attached to this zero target rate. Additionally, $n = 2$ and $n = 3$ are isotropic constraints, with the assumption that the principal strain rate direction should agree with the principal direction of stress in the unfaulted continuum elements. For elements with few GPS observations, the principal strain-rate direction is constrained by the interpolated principal stress direction. For elements with deviations between the principal strain rate direction and principal stress direction, iterative optimization is performed to obtain the optimal solution by setting a purely arbitrary uncertainty to the strain rate. The detailed algorithm and formula derivation for NeoKinema can be found in previous literature (Liu and Bird, 2008; Carafa et al., 2020).

Because the long-term predicted fault slip rates and deformation field are derived from the joint inversion of three independent observation systems (geological fault slip rate, geodetic GPS velocity, and horizontal principal compressive stress direction), NeoKinema embeds three different weighting parameters (L_0 , A_0 , and μ) to obtain the best fitting model in the acceptable solution space. Among them, L_0 is the reference length, which represents the unit length weight of the fault trace relating to the long-term slip rate, and A_0 is the reference area, which represents the unit area weight under the micro-block structure and isotropic constraints. The fitting quality of the model is governed by different dimensional tuning parameter combinations of L_0 and A_0 . In NeoKinema, it is assumed that the target value of the strain rate in the

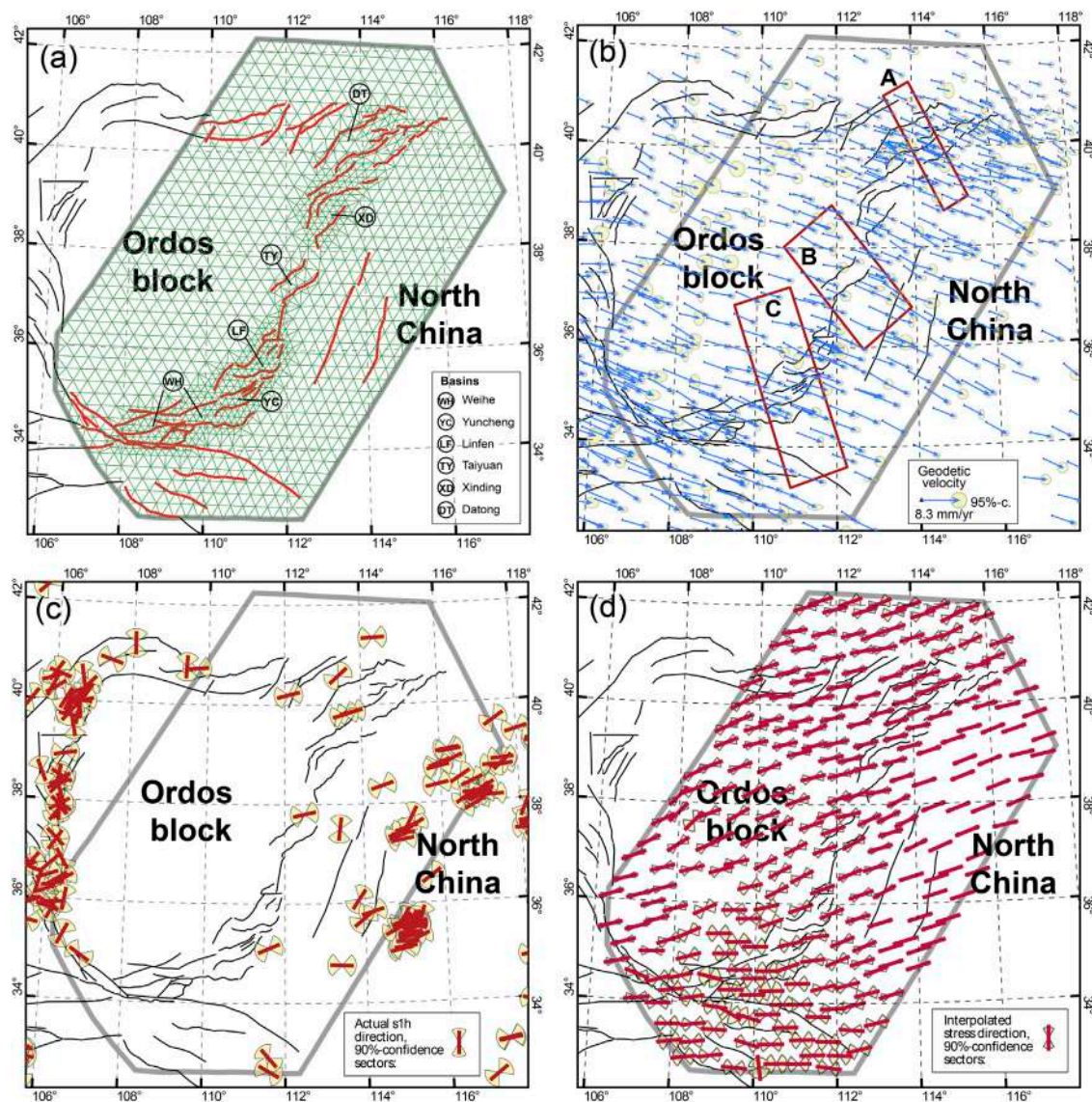


FIGURE 2

(A) Traces of active faults and finite-element grids used in this study. The solid red lines signify the active faults involved in the model. Thin green lines represent the finite-element grid. The model includes 1,069 nodes and 540 triangular elements. The length of the grid side is between 20 and 100 km. (B) GPS benchmark velocity field. The blue vectors represent GPS velocity constraints from Hao et al., 2021. The error ellipse represents 95% confidence. All velocities are with respect to the stable Eurasia plate. The red rectangle boxes with capital letters show the locations of GPS profiles which will be discussed in the next text. (C) The directions of the principal compressive horizontal principal stress from the World Stress Map (WSM database release 2016) (Heidbach et al., 2009; <http://www.world-stress-map.org/>). (D) Interpolated directions of the principal compressive horizontal principal stress by NeoKinema. For legibility, only one-quarter of the interpolated directions are shown. The heavy grey polygonal area signifies the simulation scope. The solid black lines represent the major active faults in the area.

unfaulted continuum elements is zero. However, the possibility of buried or undiscovered active faults in these regions cannot be ruled out. Moreover, no strain rates are known in the first calculation. Therefore, a small strain-rate uncertainty, μ , must be assigned for reconciliation. A detailed description of the selection of the two tuning parameters (L_0 and A_0) and the uncertainty (μ) is presented below.

3.2 Input data: Kinematic modeling constraints

3.2.1 Active fault traces

The active fault database of the FWRZ (Deng et al., 2002; Xu et al., 2016) was employed to collect the latest constraints of active faults and construct the fault geometry, as shown in

TABLE 1 Geological fault slip rates used in the model.

Abbreviation ^a	Geological slip rate ^b (mm/a)	References
WNQL	0.11–0.45 (N)	Shaanxi Earthquake Agency (1996)
ENQL	0.5–0.8 (N)	Shaanxi Earthquake Agency (1996)
HS	1.1–1.6 (N)	Xu et al. (2017)
WH	0.16–0.4 (N)	Tian et al. (2014); Tian et al. (2016)
WN-YQ	0.29 (N)	Li et al. (2019)
LS	0.1–0.3 (N)	Xu et al. (2019)
NZTS	0.75 ± 0.05 (N)	Si et al. (2014)
WNWHB	~0.2 (N)	Xu et al. (1988)
ENWHB	~0.4 (N)	Xu et al. (1988)
KZ-GS	>0.19 (N)	Yang et al. (2021)
SQ-LY	0.54 (N)	Xu et al. (1988)
HC	0.6 (N)	Xu et al. (1988)
LYS	0.47–0.88 (N)	Sun (2018)
NEMP	≥0.1–0.36 (N)	Xu et al. (2014)
HUOS	0.76–1.49 (N)	Xu et al. (2013)
TG	0.16–0.19 (N)	Jing et al. (2016)
JC	0.86–1.5 (N)	Ma et al. (1999)
NXZS	1.0–1.5 (N)	Dou and Yu (1996)
NWTS	1.0–1.2 (N)	Liu et al. (1991)
NHS	0.78–1.5 (N)	Wang (1996)
LLS	0.18–0.63 (N)	Sun (2018)
TB-WS	0.34 (N)	Xu (2002)
SYGB	0.75 (N)	Xu (2002)
NYFB	0.2–0.58 (N)	Fang et al. (1993)
NHZB	0.2–0.55 (N)	Fang et al. (1993)
KQ	0.17–0.53 (N)	Xu et al. (2011)
YG-TZ	0.12–0.2 (N)	Luo et al. (2020)
DQS	0.37–1.7 (N)	Wu et al. (1996)
NDHB	0.17–0.28 (N)	Bi (2012)
SDHB	0.17–0.28 (N)	Bi (2012)
TLZ	0.5–1.25 (L)	Yang et al. (2005)

^aAbbreviations of faults are given in Figures 1, 7.

^bL represents Left-lateral horizontal strike-slip rate, N represents normal vertical slip rate.

Figure 2A. The resulting model involves 80 active faults, all of which have clear surface traces and dip angles. To construct a fault geometry that most closely resembles the realistic distribution of faults and highlight the major active faults, the fault geometry included as many active faults as possible. However, some secondary faults were excluded according to the following considerations: 1) secondary faults without a historical record of strong earthquakes and 2) faults lacking explicit morphological evidence of strong earthquakes in the Late Quaternary.

Table 1 lists the Late Quaternary geological slip rates of the 32 faults used as the model input in this study. The fault slip rates, together with their standard deviations (STD), were taken as prior constraints in NeoKinema. We

determined the STD of the fault slip rates according to the following conventions (Liu and Bird, 2008; Khodaverdian et al., 2016): 1) if a range of geological slip rates was available, the median value was used as an *a priori* constraint, and the width of the range was assigned to be equal to ±2 mm/a; 2) if a unique value was given rather than a range, its corresponding STD was assumed to be one half of the available slip rates; 3) if no slip rate was available, a zero target rate was assigned with a large STD of approximately 5 mm/a. According to the above principles, fault slip rates with large uncertainties were assigned less weight than other kinematic constraints (geodetic velocities and stress direction) when predicting the slip rates.

3.2.2 GPS velocity field

In this study, we used the GPS horizontal velocities of Hao *et al.* (2021), which were mainly collected from the following: 1) 1999–2019 observation data from the Crustal Motion Observation Network of China (CMONOC-I/II); 2) three-epoch campaign observations implemented during 1999–2019 from the National GPS Geodetic Control Network of China; and 3) 2006–2015 observation data from the regional GPS network in the FWRZ. During data processing, the co-seismic and post-earthquake deformation of the 2008 Wenchuan *M*_s 8.0 earthquake and the 2011 Japan Tohoku-Oki *M*_s 9.0 earthquake were effectively excluded to obtain the latest regional long-term horizontal velocity field with the highest density of GPS stations and longest time span. For details on the data processing method, refer to Hao *et al.* (2021).

To improve the calculation accuracy, GPS velocities with large errors and obvious outliers were excluded from this study. Considering the uncertainty of fault traces, all GPS sites located within 2 km of the fault trace with slip rates of over 1 mm/a were deleted in the preprocessing program of NeoKinema (Bird and Liu, 2007). After careful station selection, 468 GPS benchmark velocities were used in the simulation (Figure 2B). Moreover, the velocity predicted by NeoKinema represents the long-term (steady-state) crustal deformation. Therefore, GPS velocities near the faults, which are affected by interseismic fault locking, were corrected by the negative dislocation model for elastic deformation, with a uniform locking depth of 1–15 km (Wang *et al.*, 2011). The velocities derived by the model after eliminating elastic deformation caused by interseismic fault locking are shown in Supplementary Figure S1.

3.2.3 Principal compressive stress directions

The principal compressive stress direction data, which were primarily obtained from focal mechanism solutions, indicate crustal deformation characteristics with a seismic period of up to 10³ years. Therefore, adding the principal compressive stress direction undoubtedly provides a beneficial constraint for obtaining the long-term crustal deformation. We used principal compressive stress direction data from the World Stress Map Project (Heidbach *et al.*, 2009; <http://www.world-stress-map.org/>). Specifically, two types of highly reliable principal compressive stress direction data were employed in the model: data derived from the focal mechanism solution and borehole breakout data. A total of 541 original principal compressive stress field directions (Figure 2C) were collected.

As mentioned above, NeoKinema assumes that the principal strain rate directions should be consistent with the principal stress directions in unfaulted continuum elements. These discretely observed stress directions were interpolated to all unfaulted continuum elements using a clustering algorithm (Bird and Li, 1996). The interpolated principal compressive stress direction then serves as the relaxed constraint of the

principal direction of the strain rate in the unfaulted continuum elements which has limited impact on the solutions (Figure 2D).

3.3 Multi-parameter optimization

In NeoKinema, the joint fitting quality is controlled by adjusting three tuning parameters (strain rate uncertainty, μ , reference length, L_0 , and reference area, A_0) for the optimal solutions among the multi-source constraints. Therefore, a systematic grid search in the parameter space with a total of 512 sets of tests was constructed to determine the optimal parameter combination value. The combination of eight μ values (3.90×10^{-18} – $1.50 \times 10^{-16} \text{ s}^{-1}$), eight L_0 values (1.70×10^3 – $8.80 \times 10^4 \text{ m}$) and eight A_0 (3.00×10^8 – $1.00 \times 10^{10} \text{ m}^2$) values was used as the input parameters. Therefore, for each μ , there were 64 parameter combinations of both L_0 and A_0 .

Referring to the method of Liu and Bird (2008), we calculated the L1-norm and L2-norm [root mean square (RMS) of misfit errors] of the continuum-stiffness, principal compressive stress direction, fault slip rate, and geodetic data. The results show that the misfit errors of different constraints, including the fault slip rates, geodetic velocities, and stress direction data, tend to decrease with μ , indicating that the flexibility of the finite-element grid increases with μ . Therefore, a large value of μ can hinder convergence of the NeoKinema model. Moreover, the *prior* μ should be consistent with the *posterior* μ^* in an ideal model (Liu and Bird, 2008; Bird, 2009). Figure 3 shows the calculated L2-norm errors and strain rates (mean of absolute continuum strain rates and their RMS values) with the constraints for different μ values. Each μ (solid circle) corresponds to a different (L_0 , A_0) combination (64 in total) with a mean absolute continuum strain rate and RMS (two open circles connected by a horizontal line in Figure 3). The mean absolute continuum strain rate and its RMS represent the lower and upper bounds, respectively, of the expected continuum strain rates. An appropriate *prior* μ should fall within this range. According to Figure 3, $\mu = 3.2 \times 10^{-17}$ meets the above criteria.

Thus, after fixing μ at 3.2×10^{-17} , we further searched for the optimal combination of L_0 and A_0 . Figure 4A illustrates the relationship between L_0 and the corresponding L2 norm within the variation range of L_0 after setting different A_0 values. When L_0 is larger than $1.65 \text{ m} \times 10^4 \text{ m}$, the fitting results of the continuum-stiffness, stress direction, and geodetic data were not notably improved; however, the fitting quality of the predicted quality of the fault slip rate was reduced. Therefore, L_0 cannot be too large. Figure 4B shows the relationship between A_0 and the corresponding L2 norm within the variation range of A_0 for different L_0 values. When A_0 exceeds $2.2 \times 10^9 \text{ m}^2$, the misfit error of the geodetic data and fault slip rate were reduced, whereas the misfit error of the continuum-stiffness and principal compressive stress direction data were increased. Combining the

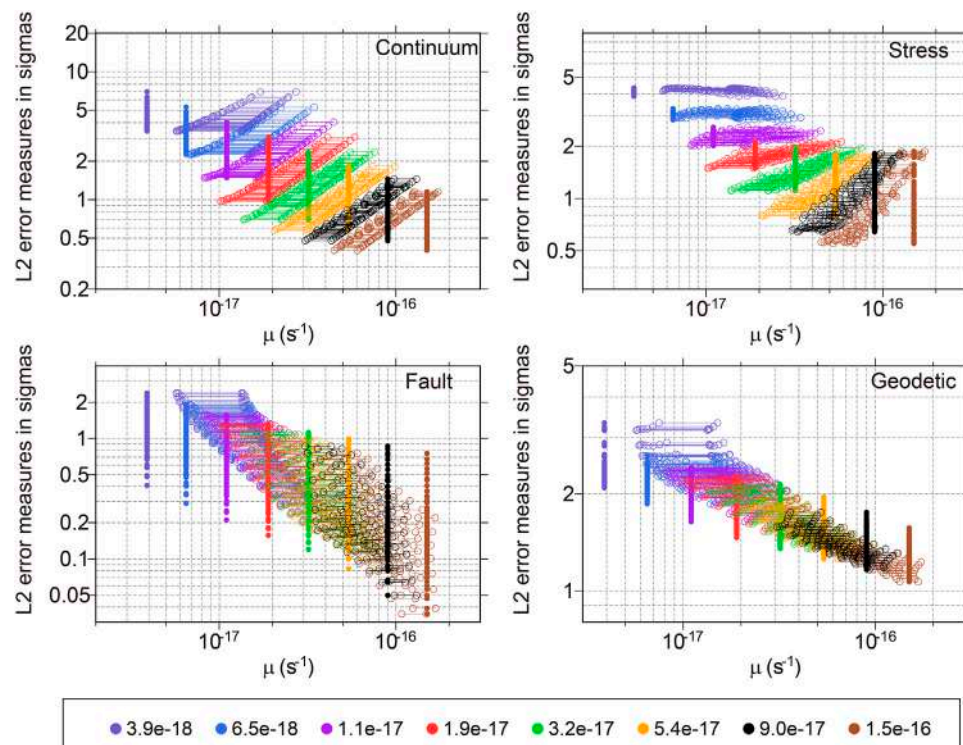


FIGURE 3

Different μ as *a priori* versus L2-norm errors (RMS prediction errors in units of standard deviations) of continuum-stiffness, stress direction, geological slip rate and geodetic velocity. (For each μ , L2 misfit errors of 64 models with various combinations of L_0 and A_0 are plotted as solid dots. The mean of absolute continuum deformation rates and their RMS associated with each μ versus L2 misfit errors of these models are plotted as open circles, connected by horizontal lines. Different colors represent different μ and the associated continuum deformation rates. For self-consistency, μ should fall within the ranges between the mean of absolute continuum deformation rates and their RMS.).

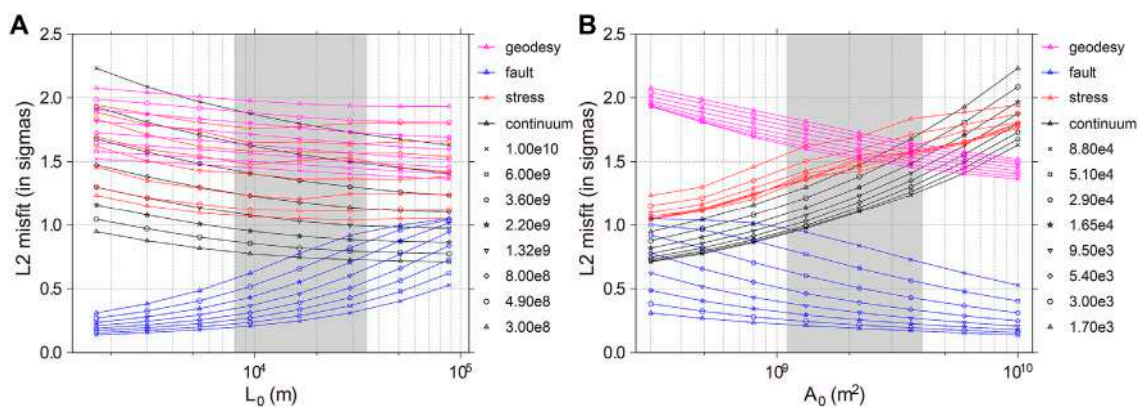
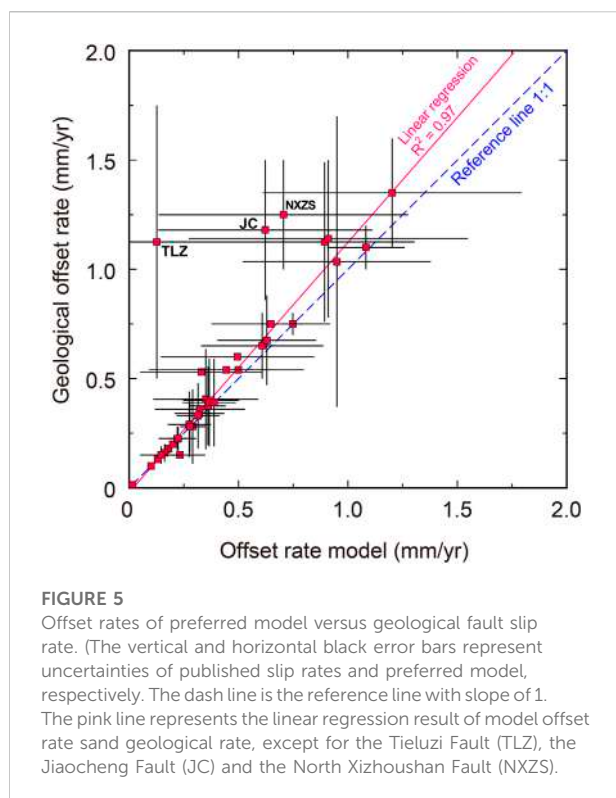


FIGURE 4

The L2-norm misfit errors of continuum strain rate, stress direction, fault slip rate and geodetic velocity under the model with different combinations of L_0 and A_0 . (A) Reference length L_0 versus the L2-norm misfit errors; (B) Reference length L_0 versus the L2-norm misfit errors. Continuum strain-rate, stress direction, fault slip rate and geodetic velocity errors are represented by symbols and connected lines in black, red, blue and magenta, respectively. Different symbols represent misfit errors versus L_0 at different A_0 . Grey shadow shows the preferred range of L_0 .



results in Figures 4A,B revealed that, if μ is fixed at $3.2 \times 10^{-16} \text{ s}^{-1}$, the combination of $L_0 = 1.65 \times 10^4 \text{ m}$ and $A_0 = 2.2 \times 10^9 \text{ m}^2$ leads to reasonable misfit errors. Thus, the L2 norm (normalized RMS error) of geological fault slip rates, GPS velocities, and interpolated principal compressive stress directions were restricted to STDs of 0.384, 1.608, and 1.372, respectively.

Consequently, the fixed μ ($\mu = 2.5 \times 10^{-16} \text{ s}^{-1}$), L_0 , and A_0 ($L_0 = 1.65 \times 10^4 \text{ m}$, $A_0 = 2.20 \times 10^9 \text{ m}^2$) values were used as the optimal parameter combination to obtain the optimal model. Figure 5 shows the predicted and published long-term fault slip rates, and details of this comparison are presented in Table 2. The model-derived long-term velocities at GPS benchmarks and the most-compressive strain-rate axes (i.e., the horizontal principal compressive stress directions) from our optimal model are shown in Supplementary Figures S2, S3, respectively. In Figure 5, the relationship between the model offset rates and geological fault slip rates falls along the reference line with a slope of one. We then fitted a linear regression for the relationship between the offset rates of the preferred model and the geological result, except for the Tieluzi Fault, Jiaocheng Fault, and North Xizhoushan Fault (with discrepancy above 1 mm/a). The coefficient of determination (R^2), which is a statistical measure in regression models, quantifies the goodness of fit. R^2 ranges between 0 and 1, with values closer to one indicating a stronger correlation. Thus, the R^2 value of 0.97 obtained in our regression model indicated a very strong correlation between the offset rates of

the preferred model and the geological result. It is worth noting that the left-lateral geological strike-slip rate of the Tieluzi Fault is 0.5–1.25 mm/a (Yang et al., 2005), which is significantly larger than the model result of $0.13 \pm 0.11 \text{ mm/a}$. In addition, the model slip rates of the Jiaocheng Fault and North Xizhoushan Fault bounding the Shanxi rift were lower than the geological constraints (Figure 5; Table 2). Furthermore, the model slip rates exhibited a large fitting error, reflecting a significant discrepancy between prior geological constraints and other constraints (especially the GPS velocity). This discrepancy does not exclude the influence of GPS observation uncertainties and the distribution of GPS stations but indicates a potential for overestimating the geological slip rate through uncertainty of the geochronological constraints. The discussion above also illustrates the importance of the joint inversion of geological and geodetic measurements.

4 Results

4.1 Long-term fault slip rate

As shown in Figure 6 and Table 2, the optimal model provides detailed fault slip rates for the most active faults in the FWRZ (Figure 6; Table 2). The predicted slip rates on most active faults are less than 1 mm/a, generally implying a low-strain kinematic background in the Late Quaternary. For the Weihe rift (Figure 1), the predicted normal dip-slip rate on the western segment of North Qinling Fault is $0.29 \pm 0.06 \text{ mm/a}$, which is lower than that of the eastern segment of the North Qinling Fault ($0.61 \pm 0.28 \text{ mm/a}$) and the Huashan Fault ($1.20 \pm 0.59 \text{ mm/a}$) (Figure 6A). Moreover, the normal dip-slip rate on the western segment of North Weihe Basin Fault ($0.19 \pm 0.01 \text{ mm/a}$) is slightly lower than that on the eastern segment of North Weihe Basin Fault ($0.37 \pm 0.12 \text{ mm/a}$) (Figure 6A). Thus, normal faulting activities along the south and north margins of the Weihe rift tend to increase in an eastward direction (Li et al., 2016).

Notably, the left-lateral strike-slip rate on the Tieluzi Fault is only $0.13 \pm 0.11 \text{ mm/a}$ (Figure 6B), which is substantially lower than the estimates derived from geological offsets in the Late Quaternary (0.5–1.25 mm/a) (Yang et al., 2005). Although the modeled long-term velocity field supports the existence of left-lateral strike-slip motion in the EW direction between the Ordos block and the Qinling Mountains (Figure 7), the left-lateral shearing zone is diffuse and not localized on individual faults. Therefore, further studies are required to determine whether this discrepancy is caused by time-varying fault motion or by observational errors in geodetic and/or geological measurements of dislocation and/or chronology.

Similar to the Weihe rift, the normal dip-slip rate on most faults in the Shanxi rift are assigned geological constraints

TABLE 2 Comparison of predicted fault slip rates with published studies.

Fault	Fault slip rate (mm/a)		References
Abbreviation ^a	Fitting slip rate ^b	Slip rate ^c	
WNQL	0.29 ± 0.06 (N)	0.11–0.45 (N)	Shaanxi Earthquake Agency (1996)
ENQL	0.61 ± 0.28 (N)	0.5–0.8 (N)	Shaanxi Earthquake Agency (1996)
HS	1.20 ± 0.59 (N)	1.1–1.6 (N)	Xu et al. (2017)
WN-YQ	0.28 ± 0.09 (N)	0.29 (N)	Li et al. (2019)
LS	0.32 ± 0.10 (N)	0.3 (N)	Xu et al. (2019)
WH	0.28 ± 0.02 (N)	0.16–0.4 (N)	Tian et al. (2014); Tian et al. (2016)
SQ-LY	0.49 ± 0.15 (N)	0.54 (N)	Xu et al. (1988)
LT-CA	0.015 ± 0.01 (N)	0.015 (N)	Li et al. (2012)
KZ-GS	0.13 ± 0.03 (N)	>0.19 (N)	Yang et al. (2021)
WNWHB	0.19 ± 0.01 (N)	~0.2 (N)	Xu et al. (1988)
ENWHB	0.37 ± 0.12 (N)	~0.4 (N)	Xu et al. (1988)
HC	0.50 ± 0.35 (N)	0.6 (N)	Xu et al. (1988)
ZTSB	0.75 ± 0.05 (N)	0.75 ± 0.05 (N)	Si et al. (2014)
TLZ	0.13 ± 0.11 (L)	0.5–1.25 (L)	Yang et al. (2005)
LYS	0.63 ± 0.23 (N)	0.47–0.88 (N)	Sun (2018)
NEMP	0.32 ± 0.20 (N)	≥0.36 (N)	Xu et al. (2014)
HUOS	0.23 ± 0.12(R)/0.89 ± 0.41 (N)	0.15 (R)/0.76–1.49 (N)	Xu et al. (2013)
TG	0.18 ± 0.01 (N)	0.16–0.19 (N)	Jing et al. (2016)
JC	0.62 ± 0.49 (N)	0.86–1.5 (N)	Ma et al. (1999)
NXZS	0.43 ± 0.21 (R)/0.71 ± 0.57 (N)	1.0–1.5 (N)	Dou and Yu (1996)
NWTS	1.08 ± 0.17 (N)	1.0–1.2 (N)	Liu et al. (1991)
NHS	0.91 ± 0.63 (N)	0.78–1.5 (N)	Wang (1996)
LLS	0.35 ± 0.24 (N)	0.43–0.75 ^a /0.18–0.63 ^b (N)	^a Deng et al. 1994; ^b Sun 2018
KQ	0.33 ± 0.28 (N)	0.17–0.53 (N)	Xu et al. (2011)
YG-TZ	0.35 ± 0.24 (N)	0.12–0.2 (N)	Luo et al. (2020)
SYGB	0.65 ± 0.27 (N)	0.75 (N)	Xu (2002)
NHZB	0.36 ± 0.08 (N)	0.2–0.55 (N)	Fang et al. (1993)
TB-WS	0.32 ± 0.11 (N)	0.34 (N)	Xu (2002)
NYFB	0.37 ± 0.12 (N)	0.2–0.58 (N)	Fang et al. (1993)
DQS	0.95 ± 0.43 (N)	0.37–1.7 (N)	Wu et al. (1996)
NDHB	0.23 ± 0.01 (N)	0.17–0.28 (N)	Bi (2012)
SDHB	0.23 ± 0.01 (N)	0.17–0.28 (N)	Bi (2012)

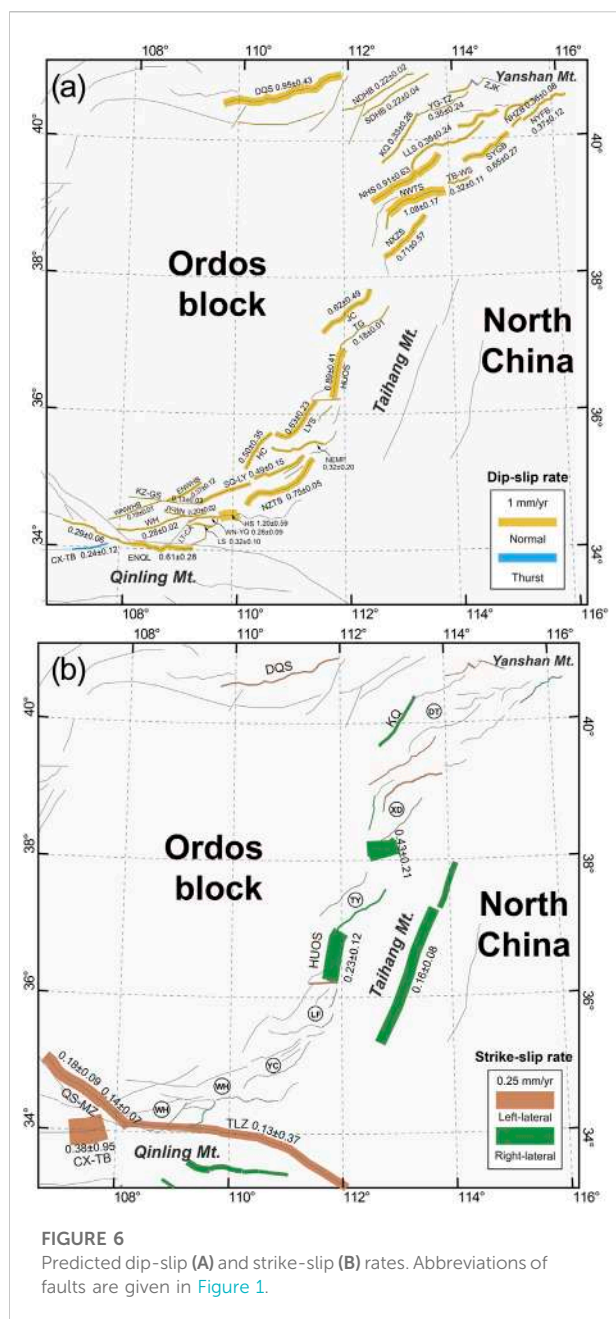
^aAbbreviations of fault are given in Figures 1, 7.

^bR and L in brackets denote right-lateral and left-lateral strike-slip, respectively.

^cThe superscript before the reference corresponds to the superscript in the fault slip rate column.

(Figure 1; Table 1). The overall low post-fit uncertainties of predicted slip rates (Table 2) illustrate relatively significant consistency between the geological results and other kinematic constraints (GPS and principal stress direction). Our predicted fault slip rates on the Shanxi rift reveal that rifting/crustal extension at its southern and northern ends are considerably higher than those in the central part of the rift (Figure 6A). Previous studies have highlighted the dextral strike-slip nature of N-S-trending active faults in the middle part of the Shanxi rift but disagreed on the magnitude of the slip rate (Xu et al., 1993). For cross-validation, we did not assign any prior constraint in our

model. Therefore, the derived strike-slip rates in the middle part of the Shanxi rift are the results constrained with GPS and principal stress direction datasets. The results of the optimal model (Figure 6B) confirm that NNE-trending active faults exhibit relatively right-lateral strike-slip motion in the middle part of the Shanxi rift. The predicted right-lateral strike-slip rates of the Huoshan Fault and North Xizhoushan Fault are 0.23 ± 0.12 mm/a and 0.43 ± 0.21 mm/a, respectively, which are very low compared to their normal dip-slip components (accounting for approximately 26% of the dip-slip) and in good agreement with a recent geological study (Xu et al., 2013).



4.2 Velocity and strain-rate fields

The predicted long-term average horizontal crustal motion velocities and strain rates are shown in Figure 7 and Figure 8, respectively. According to Figure 7, the long-term horizontal velocity is characterized by movement in the SEE-SE direction, and the velocities gradually decrease from the north to south of the entire model domain with respect to the stable Eurasia plate. The magnitude of the velocity vector in the Ordos block varies smoothly, with a velocity gradient of only approximately 1–1.5 mm/a over more than 400 km in the N–S direction

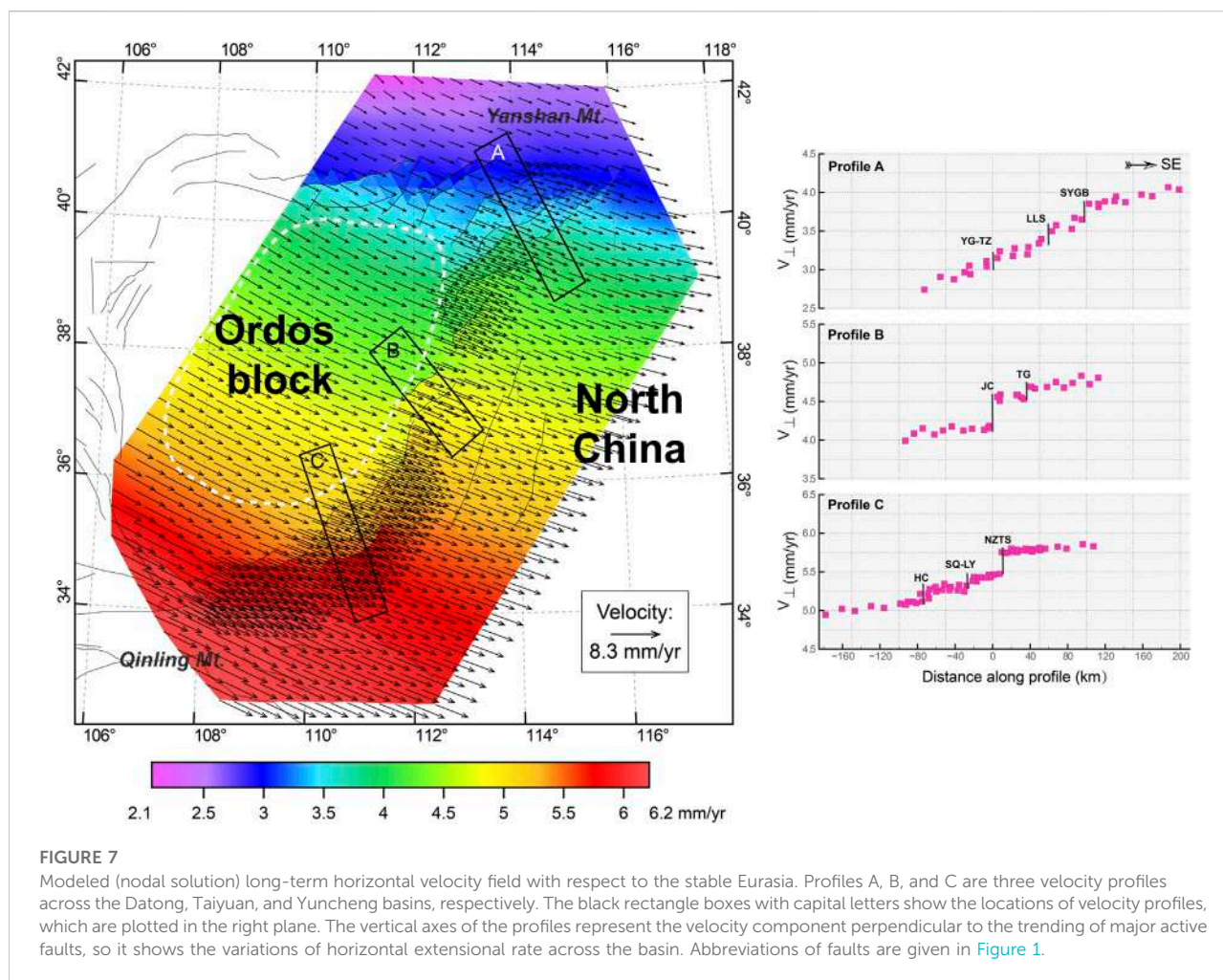
(area outlined with a white dotted line in Figure 7). This result is consistent with the rigidity and stability of the Ordos block and compatible with its lower strain rate ($1 \times 10^{-17.0} \text{ s}^{-1}$ — $1 \times 10^{-16.6} \text{ s}^{-1}$ / $3.1536 \times 10^{-10} \text{ a}^{-1}$ — $3.1536 \times 10^{-9.6} \text{ a}^{-1}$) obtained by our joint inversion. In contrast, the velocity gradient of the FWRZ, which is located between the Ordos block and North China, varies significantly and exhibits horizontal extensional motion perpendicular to the strikes of the primary active faults. In particular, the northern and southern parts are consistent with the development of many active faults with a large strain rate (approximately $1 \times 10^{-14.9} \text{ s}^{-1}$ / $3.1536 \times 10^{-7.9} \text{ a}^{-1}$). As shown in Figure 7, the horizontal extensional rate across the Datong Basin is 1.0–1.2 mm/a (Profile A, Figure 7), which is larger than extensional rate of $\sim 0.8 \text{ mm/a}$ across the Yuncheng Basin (Profile C, Figure 7). For the middle part of the FWRZ, the extensional rate across the Taiyuan Basin is lower than that across the Datong and Yuncheng basins, and extensional deformation is predominantly localized at the Jiaocheng Fault, which borders the Taiyuan Basin to its west (Profile B, Figure 7). Furthermore, a left-lateral velocity gradient of $\sim 1 \text{ mm/a}$ exists between the southern margin of the Ordos Block and the Qinling Mountains across the Weihe rift. However, this differential movement is diffusely distributed over 100–150 km in the N–S direction. As such, it is hard to identify whether the sinistral strike-slip rate on the Tieluzi Fault is greater than 1 mm/a.

5 Discussion

In this study, we integrated geological, geodetic GPS, and principal compressive stress direction datasets to obtain refined long-term (steady-state) fault slip rates, the long-term distributed horizontal velocity field, and the strain-rate field in the FWRZ. Although our results represent a trade-off among various kinematic constraints controlled by a set of weights, the posterior variances of multiple datasets approximate their respective variances, indicating that the different datasets are mutually consistent. The long-term crustal deformation results (timescale of 10^4 years or over several seismic cycles) are consistent with the regional crustal deformation revealed by present-day interseismic geodetic GPS (timescale of decades or within one seismic cycle). This kinematic model of long-term deformation provides primary constraints for seismic risk assessments and research on active regional tectonics in the FWRZ.

5.1 Reference model

Previous studies have shown that the interseismic GPS crustal deformation field is approximately consistent with the long-term crustal deformation (Zhang et al., 2004; Liu and Bird,

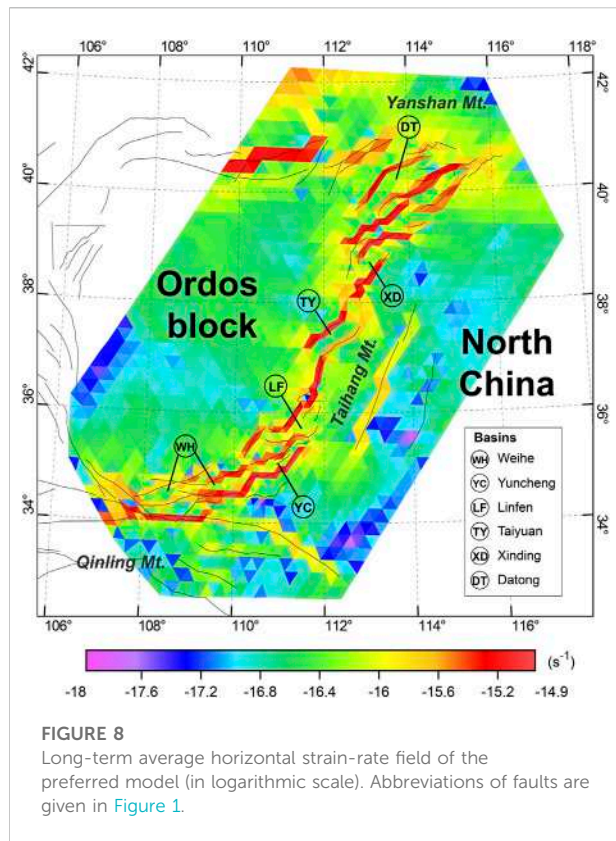


2008; Middleton et al., 2017; Gan et al., 2021), which forms the basis of our joint fitting inversion. However, the consistency between the regional crustal deformation characteristics revealed by interseismic GPS data and the geological fault slip rates of the FWRZ should still be carefully checked.

Therefore, we constructed a reference model without any GPS priorities but constrained by the fault geological slip rates and principal compressive stress directions. Thus, the fitting results of this model represent the reference long-term crustal deformation. Furthermore, the same systematic optimization method shown in [Section 3.3](#) was used to determine the optimal parameter combination ($\mu = 3.2 \times 10^{-17}$, $L_0 = 1.65 \times 10^4$ m, $A_0 = 2.20 \times 10^9$ m²) ([Figure 9](#)). After fixing these optimal parameters, we obtained the fitting results and compared the modeled velocity field of the reference model ([Supplementary Figure S4](#)) with the original GPS results ([Figure 2B](#)). Three different regions of the FWRZ, the Datong Basin, the Taiyuan Basin, and the Yuncheng Basin, were selected to represent three velocity profiles with the same locations ([Figure 2B](#); [Supplementary Figure S4](#)).

We projected the velocity components within the corresponding profile domains perpendicular to the relative fault strikes and approximately perpendicular to the predicted horizontal principal stress directions (most-compressive strain-rate axes) of the preferred model ([Supplementary Figure S3](#)). Eventually, we derived the velocity profiles shown in [Figure 10](#). The horizontal velocities derived from the reference model show extension of 1.0–1.2 mm/a across the Datong Basin ([Figure 10A](#)), which is consistent with the far-field extensional rate of 1.0–1.2 mm/a revealed by GPS observations ([Figure 10B](#)), as well as our optimal model of joint fitting inversion described in [Section 3](#) ([Figure 7](#)). Similarly, the results for Yuncheng Basin show that the long-term crustal deformation derived from the reference model ([Figure 10E](#)) is compatible with the results revealed by interseismic GPS measurements ([Figure 10F](#)).

Previous studies have shown that significant normal dip faulting and seismic activity occurred since the Late Quaternary to the Holocene in the Taiyuan Basin ([Ma et al., 1999](#); [Xie et al., 2004, 2008](#); [Jing et al., 2016](#)). Notably, as shown in [Figure 10D](#), it is difficult to determine the significant extensional signal across the Taiyuan

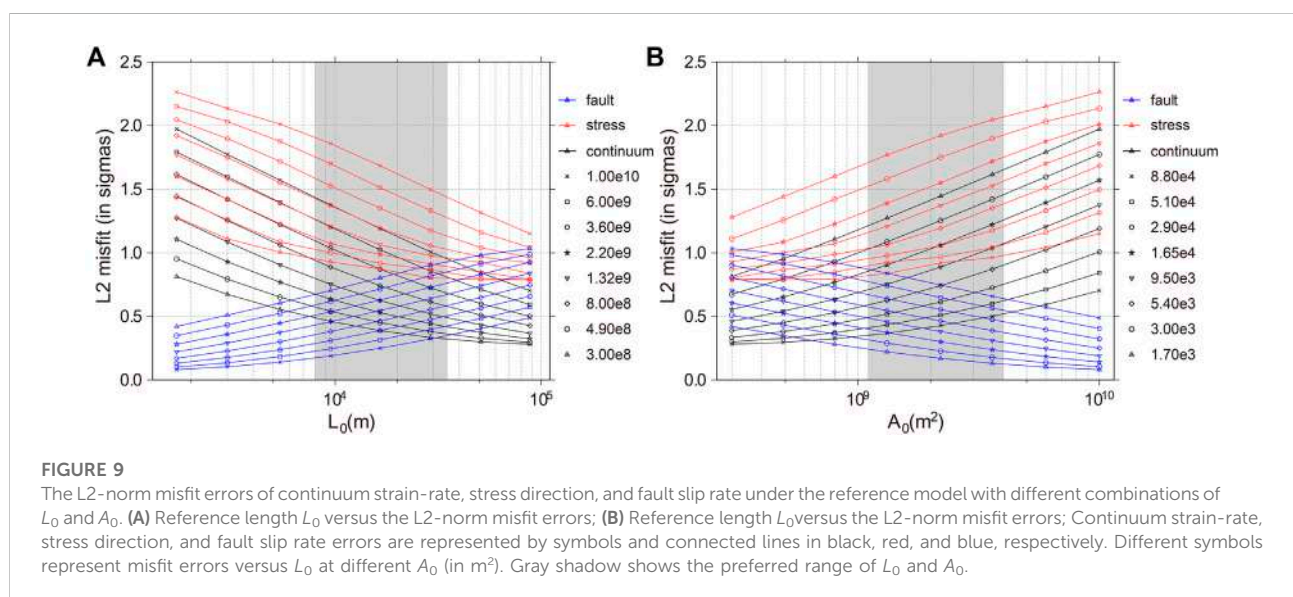


Basin using only present-day GPS observations. However, obvious extensional movement is revealed in the corresponding horizontal velocity profile (Figure 10C) derived from the reference model. Therefore, it is necessary to apply joint inversion objectively to recover the long-term deformation field.

The above comparison suggests that the long-term and short-term interseismic crustal deformation results for the FWRZ are approximately compatible. It also indicates that, in a low-strain region such as the FWRZ, geodetic GPS can accurately define the total far-field rate across some active tectonic zones and constrain blocks (terrane) far from the active boundaries. However, where GPS is not sensitive to very slight motion on individual faults, the geological slip rate can compensate for the limitations of GPS and recover more objective long-term deformation and strain-rate fields.

5.2 Seismic hazard forecasting

The FWRZ is famous for strong earthquakes within mainland China. According to historical records, there have been numerous earthquakes with $M \geq 7$, including the Huaxian 1556 M 8.3 earthquake, which caused the most deaths from any earthquake in human history. According to the long-term (steady-state) fault slip rate (Figure 5) and the long-term strain-rate field (Figure 7) inferred from the optimal model, we employed the tectonic-dependent SHIFT (Seismic Hazard Inferred from Tectonic) model to calculate long-term seismicity in the FWRZ. The calculation included two steps: 1) calculation of the seismic moment rate of each deformation unit using the long-term strain rates coupled with the seismogenic layer thickness (cz); 2) forecast the seismicity from the seismic moment rate with an assumed magnitude–frequency distribution and seismicity coefficients (corner magnitude m_c and asymptotic spectral slope β) (Bird and Kagan, 2004). As the above parameters were obtained using focal mechanism solutions of different tectonic units worldwide, the seismic model is a tectonic-dependent model.



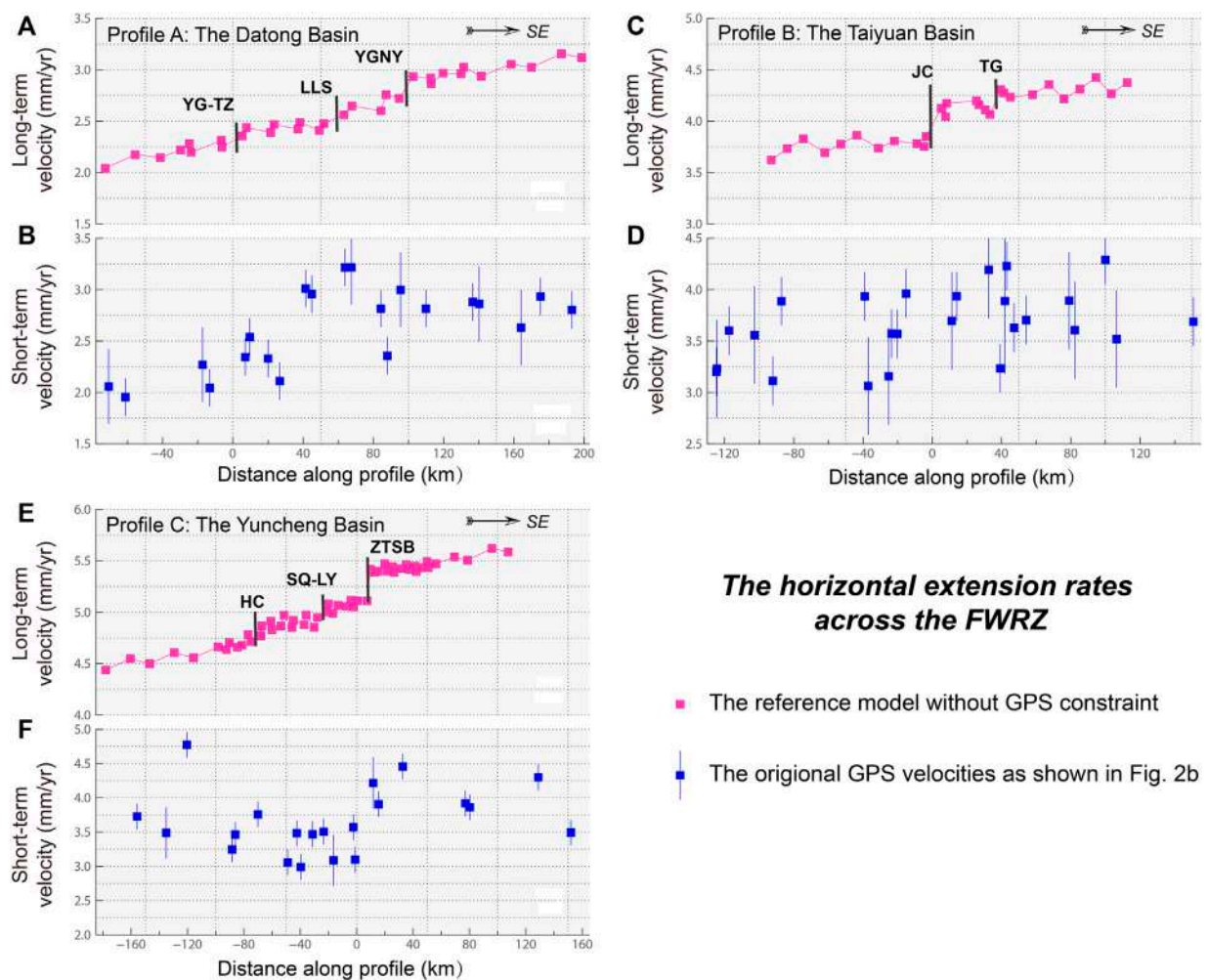


FIGURE 10

The long-term horizontal velocity profiles (pink square dots, see Figure 7 for the profile location) from our reference model and GPS (short-term) velocity profiles (blue square dots, see Figure 2B for the profile location) perpendicular to the striking direction of primary active faults. (A, B): the long-term and short-term velocity profile of the Datong Basin; (C, D): the long-term and short-term velocity profile of the Taiyuan Basin; (E, F): the long-term and short-term velocity profile of the Yuncheng Basin; Abbreviations of faults are given in Figure 1.

In this study, the magnitude threshold of long-term shallow earthquakes (depth <70 km) was specified as $M 6$ according to the following considerations: 1) $M 6$ is greater than the lower boundary of the minimum magnitude predicted by the SHIFT model (Bird and Kagan, 2004), which can ensure a high accuracy of calculation; 2) although some studies have proposed that the minimum magnitude of completeness for the 500-year historical earthquake catalog in this region is $M 5$ (Xu and Gao, 2014), for a large number of non-instrumentally recorded earthquakes, the magnitude and location are relatively high. To further determine the magnitude threshold, we plotted the relationship between the historical earthquake magnitude ($M 5$ – $M 7$) and cumulative number of earthquakes ($> M 4$) (Figure 11A). In contrast to the fitting line, Figure 11A shows a remarkably steep slope between

$M 5$ and $M 6$, implying that the seismic catalog of $M 5$ – $M 6$ is incomplete. The histogram (Figure 11B) also shows a loss of earthquake records between $M 5$ and $M 6$ (Figure 11). Finally, $M 6$ was determined using the magnitude threshold of seismic activity, and the SHIFT model was used to predict regional long-term seismic activity in the FWRZ and adjacent areas.

As a developed area of ancient civilization, the FWRZ has sufficient historical earthquake records. Studies have shown that most areas have relatively complete earthquake catalogs of destructive earthquakes of $M \geq 5$ since 1,500 (Xu and Gao, 2014), which facilitates the analysis of regional strong earthquake potential through a comparison with seismicity data calculated by the SHIFT model. Based on the regional long-term seismicity calculation (Figure 12), we further divided the FWRZ into

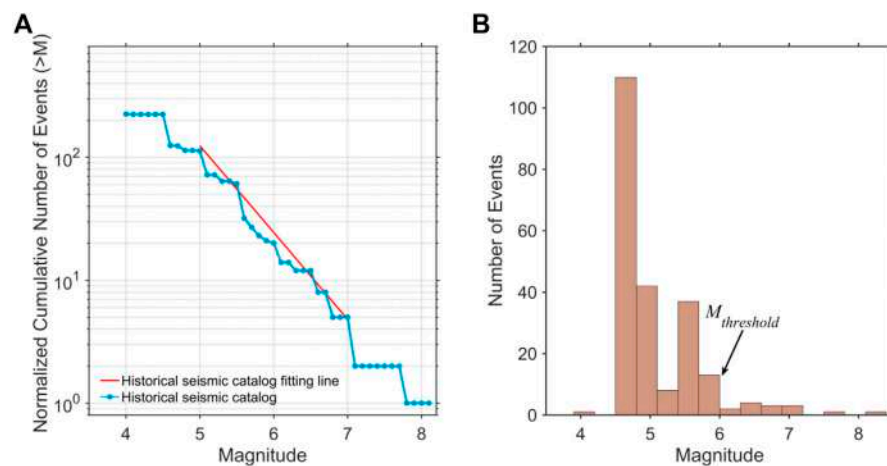


FIGURE 11

(A) The G-R magnitude frequency relation of historical earthquakes in the FWRZ. (B) Frequency as a function of magnitude corresponding to the historical earthquake catalog in (A). $M_{threshold}$ represents the magnitude threshold of long-term shallow earthquakes for calculating the long-term seismicity in the FWRZ. The historical earthquakes data are after the National Earthquake Data Center (<http://data.earthquake.cn>).

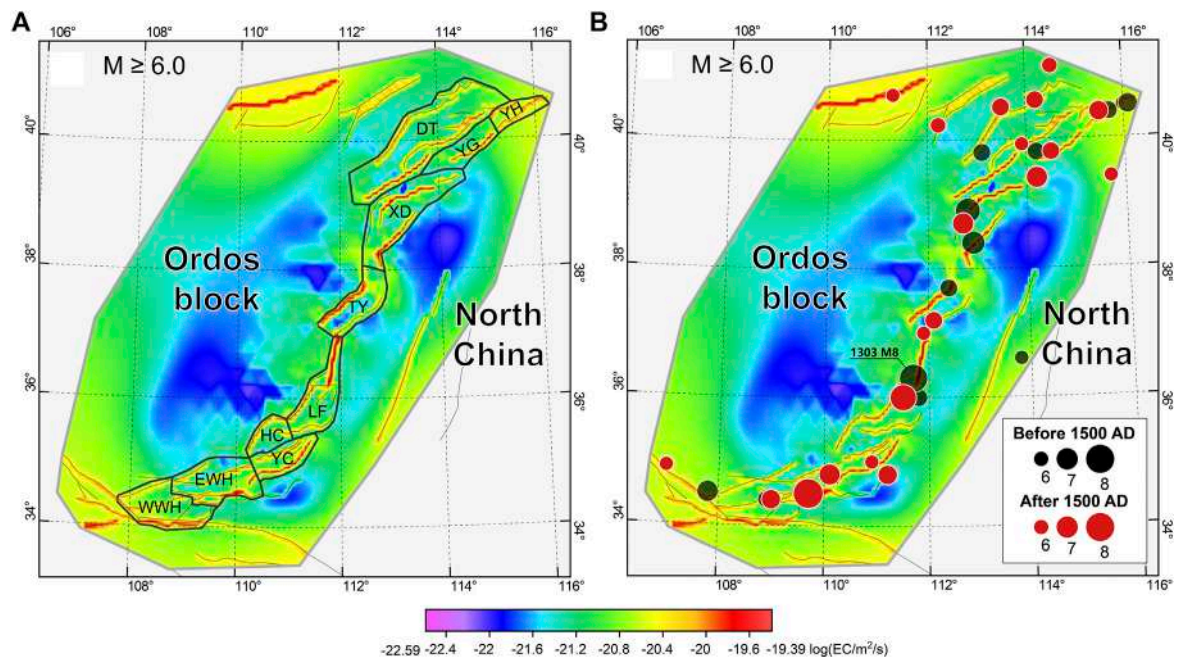


FIGURE 12

(A) The forecast of shallow seismicity rate (in epicenters per square meter per second) in the FWRZ for a threshold M 6.0, evaluated on a $0.0625^\circ \times 0.0625^\circ$ grid. The black polygonal areas represent the partition boundaries used to count seismicity. Abbreviations are: WWH, the Western Weihe Basin; EWH, the Eastern Weihe Basin; YC, the Yuncheng Basin; HC, the Hancheng Basin; LF, the Linfen Basin; TY, the Taiyuan Basin; XD, the Xinding Basin; DT, the Datong Basin; YG, the Yuguang Basin; YH, the Yanhuai Basin. (B) Consistent with (A), the red and black dots represent the epicenter distribution of historical earthquake occurred in the FWRZ before and after 1,550, respectively.

10 sub-seismic activity zones (Figure 12A; Supplementary Figure S5) according to previous seismic structure partitioning (Li et al., 2020) and compared the calculated seismicity with the actual earthquake records of $M \geq 6$ (Figure 12B; Supplementary Table S1) from 1,500 to 2022 (522 years). According to the statistical results, calculated earthquake activity of $M \geq 6$ for the Datong Basin and adjacent areas in the northern FWRZ is $3.96/522 \text{ a}^{-1}$, whereas the actual record is $2/522 \text{ a}^{-1}$. For the western FWRZ, the calculated seismicity is $2.27/522 \text{ a}^{-1}$, compared to the actual record of $1/522 \text{ a}^{-1}$. Thus, the relatively significant seismicity deficit in the Datong Basin indicates a background of high seismic risk. In addition, the Yuncheng and Hancheng basins, which are closely related both spatially and tectonically, also indicate a seismicity deficit. This seismicity deficit also characterizes the Linfen Basin; however, the M 8.0 earthquake in 1,303 (Figure 12B) occurred close to the lower boundary year (1,500) used for the calculation. Therefore, the strong seismic risk in the Linfen Basin should be lower than that in the aforementioned regions and is worthy of further study.

5.3 Tectonic implications

Previous studies have indicated that the Shanxi rift, as the main body of the FWRZ (Figure 1), exhibits the characteristic strain partitioning pattern of transtensional tectonics. At its southern (the Yuncheng Basin) and northern (the Datong Basin) ends, horizontal extension is predominantly controlled by fault-tip transtension of the right-lateral shearing localized on the Huoshan Fault and the North Xizhoushan Fault in the central Shanxi rift (Xu and Ma, 1992). The results of our optimal model are generally consistent with this spatial pattern of strain partitioning (Figures 6, 7). However, the predicted right-lateral shearing of $\sim 0.5 \text{ mm/a}$ on the N–S-trending faults located in the central Shanxi rift is incompatible with the high extensional rates of $1\text{--}1.2 \text{ mm/a}$ in the southern and northern Shanxi rift. Thus, we cannot rule out the hypothesis that the deep Earth participates in crustal extension through material exchange in the vertical direction (Deng et al., 1999; Zhu et al., 2011; Lei, 2012; Zhang et al., 2016). It is possible that the far-field effect of Pacific subduction might be superimposed on the horizontal crustal extension (Hao et al., 2021).

Our predicted results do not support significant extrusion of the Tibetan Plateau along the E–W-trending Qinling Mountains (Zhang et al., 1995; Tapponnier et al., 2001). Specifically, the estimated sinistral strike-slip rate of the Tieluzi Fault within the eastern part of the Qinling Mountains is very low ($\sim 0.2 \text{ mm/a}$) (Figure 6). Although this result could be attributed to the sparse GPS stations or observation errors, the total sinistral shear rate across the southern Ordos block and Qinling Mountains determined from the predicted long-term velocity field is approximately 1 mm/a (Figure 7). Furthermore, a recent study revealed that eastward crustal extrusion of the Tibetan Plateau is resisted by the rigid Ordos block and transformed into thrust faulting and crustal folding in the southwestern corner of the

Ordos block (Zheng et al., 2013). The remaining sinistral shear strain is $\sim 1 \text{ mm/a}$, which is accommodated by the NW-trending Qishan–Mazhao Fault in the southwestern end of the Weihe rift (Figure 1) (Li et al., 2018). Thus, it is inferred that sinistral shear induced by “extrusion tectonics” decreases in an eastward direction because of accommodation of depression activity in the Weihe rift, which appears to be consistent with our results.

6 Conclusion

In this study, we constructed a detailed active crustal deformation model of the FWRZ, which consists of the long-term slip rates of most active faults, the long-term crustal deformation field, and the strain-rate field (including on-fault and off-fault fields). This model was constructed by joint fitting of the most recent geological fault slip rate, GPS velocity field, and principal compressive stress direction datasets. The major conclusions of this study are as follows:

- 1) The majority of predicted long-term fault slip rates in the FWRZ are less than 1 mm/a , which implies a low-strain kinematic background. Furthermore, normal dip-slip faulting in the Weihe rift, which represents the south part of the FWRZ, gradually increases in an eastward direction. The predicted sinistral strike-slip rate of the Tieluzi Fault is very low ($\sim 0.2 \text{ mm/a}$), indicating limited tectonic extrusion along the E–W-trending Qinling Mountains. Indeed, the total sinistral shear rate from the southern Ordos block to the Qinling Mountains is only approximately 1 mm/a .
- 2) Although the FWRZ is dominated by normal dip-slip faulting, dextral shear of $\sim 0.5 \text{ mm/a}$ exists on the N–S-trending faults in the central Shanxi rift, which decreases from the central Shanxi rift to its northern and southern ends. Conversely, normal dip-slip faulting is enhanced in these regions, with relatively large crustal extensional rates of $1.1\text{--}1.2 \text{ mm/a}$ across the Datong and Yuncheng basins. However, this significant crustal extension cannot be solely attributed to the terminal effects caused by limited dextral shear in the central Shanxi rift.
- 3) A comparison of historical earthquake records and predicted seismicity based on calculated long-term strain rates reveals some prominent areas of seismicity deficit, where the predicted seismicity is much higher than that in the record, including the Datong, Hancheng, and Yuncheng basins, indicating there higher seismic potential and being worthy of further study.

Data availability statement

The raw data supporting the conclusion of this article will be made available by the authors, without undue reservation.

Author contributions

SS performed the experiment and contributed significantly to manuscript preparation; YL contributed to the conception of the study, performed the data analyses, and wrote the manuscript. MH helped perform the analysis and provided constructive discussions.

Funding

This research was jointly supported by the National Key Research and Development Program of China (2017YFC1500100), the National Natural Science Foundation of China (42072243, 41874017), and the Continental Rift Valley Dynamics State Observatory of Taiyuan (NORSTY 2021-05).

Acknowledgments

We thank Peter Bird for providing the program NeoKinema. The figures in this paper were created using GMT tools (Wessel and Smith, 1995).

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Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.964800/full#supplementary-material>

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OPEN ACCESS

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SPECIALTY SECTION
This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 27 April 2022
ACCEPTED 08 July 2022
PUBLISHED 08 August 2022

CITATION
Liu L, Zhuang W, Ji L, Zhu L and Jiang F
(2022), Fault locking of the
Qilian–Haiyuan fault zone before the
2022 Menyuan $M_s6.9$ earthquake and its
seismic hazards in the future.
Front. Earth Sci. 10:929597.
doi: 10.3389/feart.2022.929597

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Fault locking of the Qilian–Haiyuan fault zone before the 2022 Menyuan $M_s6.9$ earthquake and its seismic hazards in the future

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By using GPS-derived velocities of 2015–2021 and a negative dislocation program, we inverted the locking degree and slip rate deficit in the Qilian–Haiyuan fault zone, and combined with the distribution of small earthquakes in the fault, we studied the characteristics before the 2022 Menyuan $M_s6.9$ earthquake and analyzed the future seismic hazards of each segment within this fault zone. The regional crustal deformation pattern is discussed with regard to the fault slip rate and regional strain rate field. The preliminary results show that before the earthquake, the seismogenic fault was strong locked, with a high locking depth, the slip rate deficit was large, and the distribution of small earthquakes was relatively few, these characteristics are closely related to the occurrence of strong earthquakes, according to the aftershock relocation results, further, it is believed that the earthquake may link the Lenglongling and Tuolaishan faults into a large strike-slip fault. The Jinqianghe fault, the Lenglongling fault, and the eastern segment of the Tuolaishan fault are strongly locked, with high locking depth and large slip rate deficit, combined with the occurrence of small earthquakes and the locking degree before the 2022 Menyuan $M_s6.9$ earthquake, indicate that the eastern segment of the Tuolaishan fault is highly likely to have strong earthquakes in the future, which requires further attention. In addition, the strike-slip rate of the Qilian–Haiyuan fault zone is mainly between 3.9 and 4.3 mm/yr, the overall movement of the fault is consistent, and the compressional rate gradually decreases from 2.9 mm/yr in the western segment to 1 mm/yr in the eastern segment; the fault compressional rate may be related to the crustal shortening (formation basin and uplift mountain). Therefore, the present-day crustal deformation in the northeastern margin of the Tibetan Plateau is mainly distributed in the shortened region of the crust on the Qilian Shan area and left-lateral strike-slip localized on the Qilian–Haiyuan fault zone.

KEYWORDS

Qilian–Haiyuan fault zone, Menyuan $M_s6.9$ earthquake, fault locking, seismic hazard, crustal deformation

1 Introduction

The northeastern margin of the Tibetan Plateau belongs to the Qaidam–Qilian active block (Zhang et al., 2003), which is the front part of the expansion of the Tibetan Plateau to the interior of the continent. It absorbs and modulates the NNE compression and convergence of the Indian plate to the Eurasian plate and has strong tectonic activity, causing the occurrence of relatively frequent moderate and strong earthquakes (Dewey and Burke, 1973; Métivier et al., 1998; Zhang et al., 2004; Molnar and Stock, 2009). The Qilian–Haiyuan fault zone is a large strike-slip fault zone in the northeastern margin of the Tibetan Plateau, this fault not only controls the geometry and tectonic pattern of the northeastern margin of the Tibetan Plateau, but also plays an important role in regulating the eastward movement of crustal material in the northeastern margin relative to the Alashan block. The fault zone is mainly left-lateral strike-slip, with a total length of approximately 1,000 km, and is mainly composed of the Tuolaishan, Lenglongling, Jinqianghe,

Maomaoshan, Laohushan, and Haiyuan faults from west to east (Tapponnier and Molnar, 1976; Peltzer and Tapponnier, 1988; Gaudemer et al., 1995; Tapponnier et al., 2001; Zheng et al., 2013; Daout et al., 2016). Currently, faults are highly seismically active. According to historical records, there have been several moderate and strong earthquakes in the Qilian–Haiyuan fault zone since 1900, such as the 1920 Haiyuan $M_{8.5}$, 1986 Menyuan $M_{6.4}$, 1990 Tianzhu $M_{6.2}$, and 2016 Menyuan $M_{6.4}$ earthquakes. In addition, the Menyuan $M_{6.9}$ earthquake (8 January 2022) also occurred in the Tuolaishan and Lenglongling faults (Li et al., 2022; Yang et al., 2022). According to the Global Centroid Moment Tensor (GCMT) inversion results, the epicenter was located at 37.80°N and 101.31°E, with a focal depth of 14.8 km (<https://www.globalcmt.org/CMTsearch.html>), and the focal mechanism was a high-angle left-lateral strike-slip fault. This shows that the Qilian–Haiyuan fault zone controlled a series of historical earthquakes and is a natural testing ground for studying the mechanism of earthquake initiation (Figure 1).

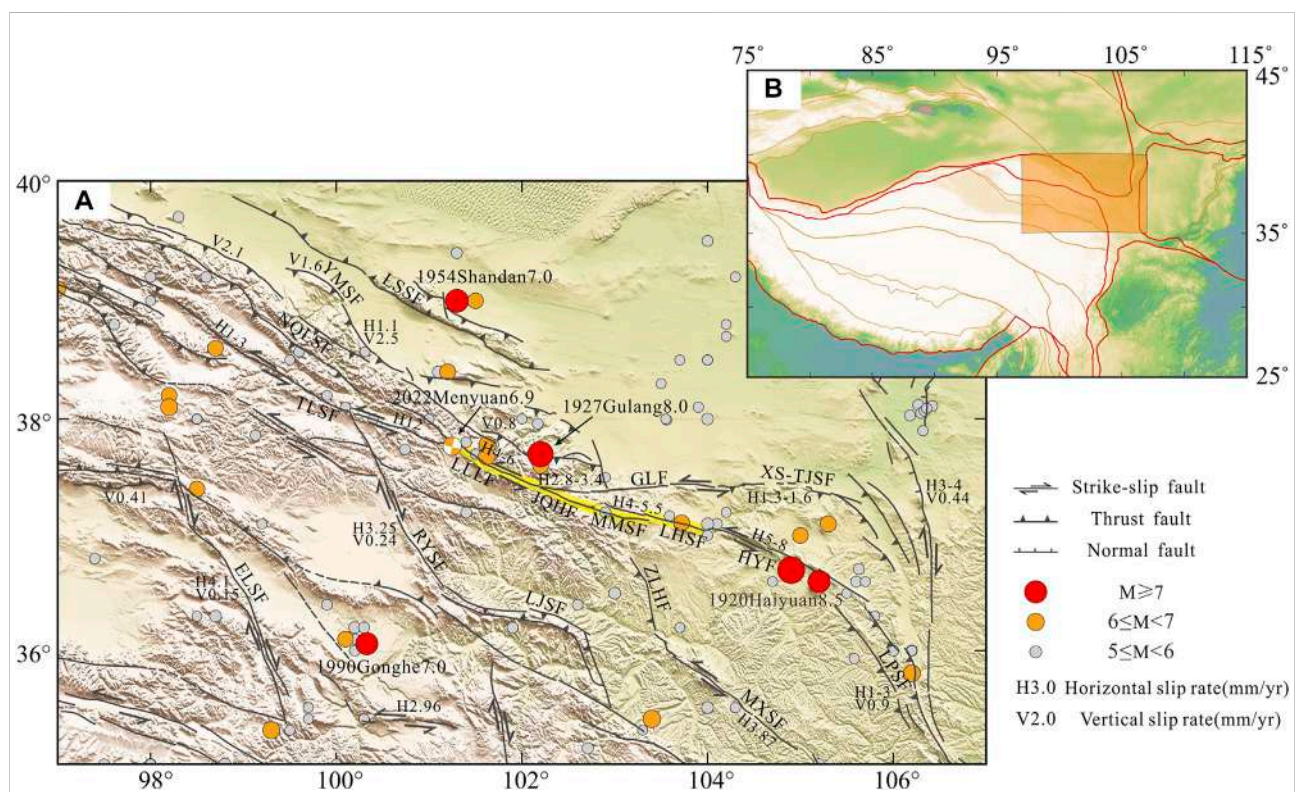


FIGURE 1

(A) Active faults and earthquakes in the northeastern margin of the Tibetan Plateau. Black lines represent fault lines mapped by (Xu et al., 2017). Gray, orange, and red circles indicate the locations of $M_{5 \sim 6}$, $M_{6 \sim 7}$, and $\geq M_7$ earthquakes that have occurred since 1900, respectively. The focal mechanism solutions were obtained from Global CMT Catalog (<http://www.globalcmt.org/CMTsearch.html>). Yellow line follows the Tianzhu seismic gap (Gaudemer et al., 1995). The Geologic slip rate from Yuan et al. (2004). LSSF: Longshoushan fault; YMSF: Yumushan fault; NQLSF: Northern Qilian Shan fault; TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQHF: Jinqianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYF: Haiyuan fault; GLF: Gulang fault; XS-TJSF: Xiangshan–Tianjingshan fault; LPSF: Liupanshan fault; ELSF: Elashan fault; RYSF: Riyueshan fault; LJSF: Lajishan fault; ZLHF: Zhuanglanghe fault; MXSF: Maxianshan fault (B) Location of the study area in relation to the Tibetan Plateau. The orange box is the area shown in (A). The orange lines are active block boundaries mapped by Zhang et al. (2003).

Earthquake occurrences are closely related to fault activity. Strong and large earthquakes often occur in fault-locked segments or asperities with high strain accumulation in the active fault zone (Wiemer and Wyss, 1997; Wyss and Wiemer, 2000). The inversion of the fault locking degree and slip rate deficit based on geodesic observation data can facilitate the identification of fault asperities and study fault seismogenic capacity. For example, after the Wenchuan $M_s8.0$ earthquake in 2008, the locking degree of the Longmen Shan fault was already in high at the source before the earthquake, but it was adjusted to non-locking after the earthquake (Zhao et al., 2012, 2020). After the Chile $M_w8.8$ earthquake in 2010, Madariaga et al. (2010) predicted that a large earthquake would occur in this region 10 years ago, mainly based on the strong locking of the fault and rupture gap of the strong earthquake (Moreno et al., 2010). The strong locking of the Japan Trench fault before the $M_w9.0$ earthquake of 2011 in Japan, based on GPS data was essentially consistent with the coseismic dislocation concentration area of this earthquake in space (Hashimoto et al., 2009; Loveless and Meade, 2011). The source of the 2015 $M_w7.8$ earthquake in Nepal was located at the edge of the strong locked fault, and between the rupture zone of the Karnali $M8.2$ earthquake in 1,505 and the Bihar $M8.1$ earthquake in 1934 (Zhao et al., 2017). Therefore, the locking degree and slip rate deficit characteristics of faults are important for predicting the seismic hazards of the fault zone (McCaffrey, 2005; McCaffrey et al., 2007; Cavalié et al., 2008; Jolivet et al., 2013).

Thus far, extensive research on the seismic hazards of the Qilian–Haiyuan fault zone has been performed. Using field investigations and remote-sensing images, Gaudemer et al. (1995) found that there was a 220 km long earthquake hazard zone–“Tianzhu seismic gap” on the Qilian–Haiyuan fault zone. Using field geological investigations, Guo (2019) considered that the 160 km long zones of the eastern segment of the Jinqianghe, Maomaoshan, and Laohushan faults may experience large earthquakes in the future. Hao et al. (2017) used GPS data to invert the locking degree of the Haiyuan–Liupanshan fault and analyzed the areas where strong earthquakes may occur in the future. Moreover, Li et al. (2014) and Li et al. (2016) performed inversion of the locking of the Qilian–Haiyuan–Liupanshan fault zone and believed that the Jinqianghe–Maomaoshan fault had a high seismic hazard. Therefore, the Qilian–Haiyuan fault zone is still the focus of strong earthquakes, and the Menyuan $M_s6.9$ earthquake in 2022 has once again made us realize the strong seismicity of the Qilian–Haiyuan fault zone. Are there any abnormal characteristics before the earthquake? How will this affect the seismic hazard of faults in the future? These problems are of great significance for future seismic potential assessments of fault zones and require further analyses.

In this paper, we examined the Qilian–Haiyuan fault zone as the research object, utilizing the GPS velocity field results from 2015 to 2021, the strain rate field of the region and slip rates of major faults were obtained. Further, the locking degree and slip rate deficit of the fault zone were inverted, and combined with the regional relocation of small earthquakes and the relocation of the aftershocks of the 2022 Menyuan $M_s6.9$ earthquake, the present-day activity and characteristics of small earthquakes occurring within this fault zone were comprehensively analyzed. Subsequently, according to the characteristics of this fault zone before the Menyuan $M_s6.9$ earthquake, the segments of the fault where strong earthquakes may occur in the future are determined. Finally, the regional crustal deformation pattern is discussed with regard to the strain rate field and fault slip rate, for the regions for which strong seismic hazard prediction and the tectonic transformation pattern of the northeastern margin of the Tibetan Plateau, to provide appropriate references and constraints.

2 Faults slip rate and strain rate field

2.1 GPS data processing

We collected GNSS observation data from the Crustal Movement Observation Network of China (CMONOC I/II) from 2015 to 2021. We adopted a strategy to uniformly process/reprocess daily GPS data using the GAMIT software (Herring et al., 2015) to ensure the quality and homogeneity of the solutions. The main methods used are as follows: The baseline calculation of the GNSS observation data was carried out using a single day as one period. In addition to allowing the slight adjustment of satellite orbit and Earth rotation parameters, we used the gravity field model EGM08, optical pressure model BERNE, Earth radiation model TUME1, antenna inference model ANTBK, geomagnetic field model IGRF13, and high-order ionospheric correction model GMAP to estimate and correct the relevant parameters in data processing.

To avoid the differences in calculation results caused by the difference between the model and frame, the same model and method were used to process the data from approximately 70 IGS stations and reference stations selected globally with a uniform distribution and obtain the baseline relaxation solution. HTOGLB was then used to convert the single-day relaxation solution files into GLOBK-approved single-day baseline relaxation solutions. Furthermore, GLRED was used to calculate the coordinate time series under the ITRF 2014 (Altamimi et al., 2017). The repeatability of the day coordinates was then checked. Finally, the seasonal variation of annual and half-yearly cycles, the step caused by instrument replacement, and other nonstructural factors were removed by model fitting. Next, the GNSS velocity field under the

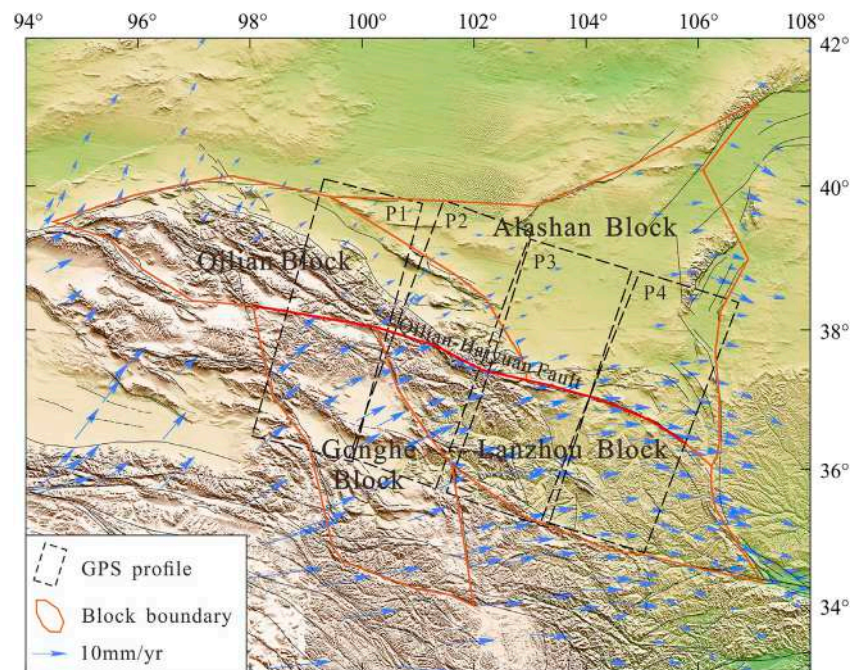


FIGURE 2

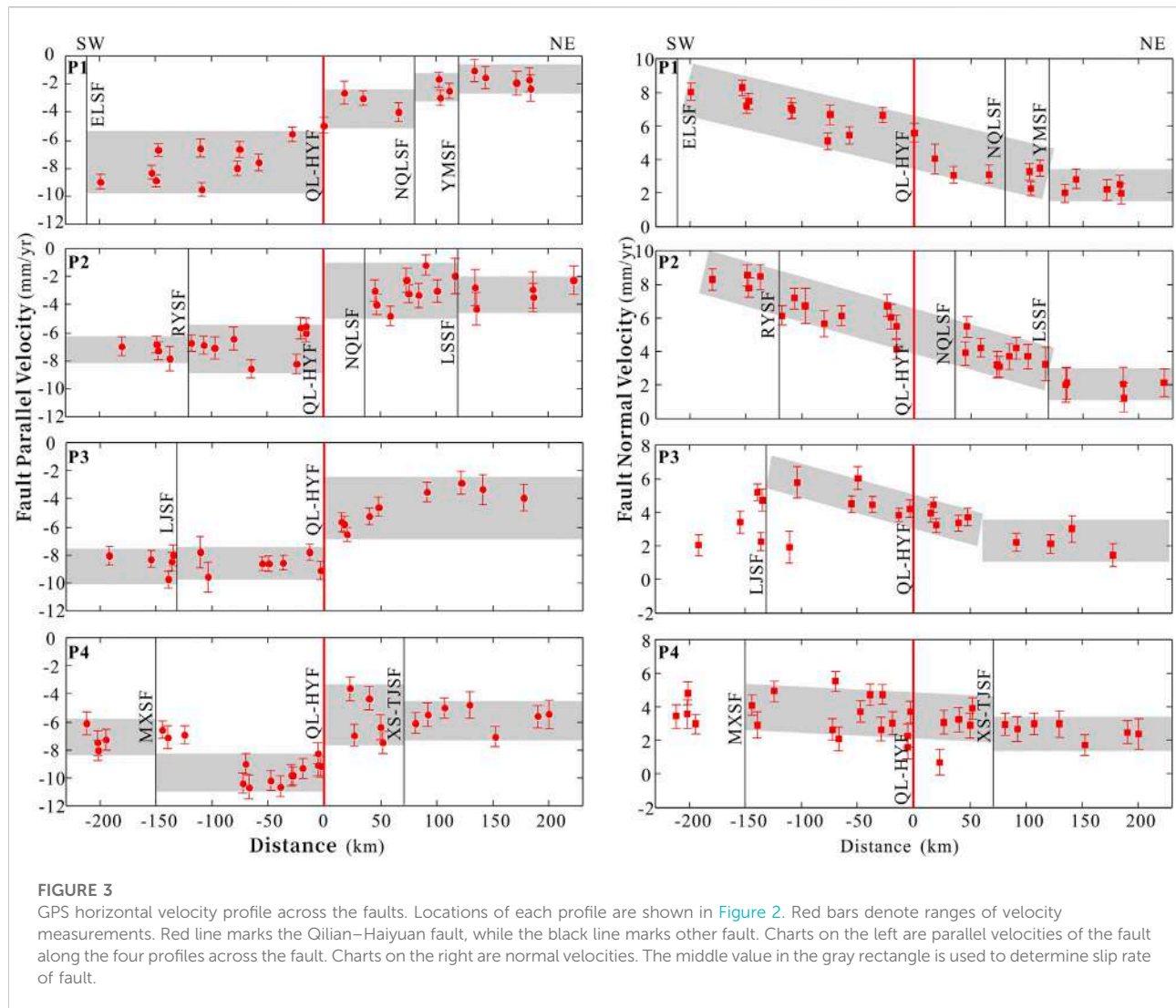
GPS horizontal velocity fields and block boundaries. The red line indicates the Qilian–Haiyuan fault zone. The orange lines indicate active block boundaries. Blue arrows represent the GPS horizontal velocities with respect to stable Eurasia plate. Black dotted rectangles show the locations of GPS horizontal velocity profiles in [Figure 3](#).

ITRF2014 framework was estimated from the clean time series of many years; using the Euler rotation vector between the Eurasian frame (provided by [Altamimi et al. \(2017\)](#)) and ITRF 2014, the GNSS horizontal motion velocity field into the stable Eurasian reference frame was calculated. At the same time, the measurement stations in the GNSS velocity field that deviated significantly from the direction and size of the research region were removed ([Figure 2](#)).

2.2 GPS velocity profiles

The GPS velocity profile of the fault can directly reflect the differential motion of the blocks on both sides and the strain accumulation state of the fault ([Wang and Shen, 2020](#); [Song et al., 2022](#)). Therefore, to study the current slip rate of the Qilian–Haiyuan fault zone and the regional differential motion characteristics, we constructed four profiles from west to east across the Qilian–Haiyuan fault zone (profiles P1, P2, P3, and P4 in [Figure 2](#)). The GPS horizontal velocity field was projected along the fault strike direction and normal fault strike direction, the strike-slip rate or extensional/compressional rate of the fault was estimated from the difference between the average velocities of two stations on both sides of the fault, and the error propagation law was used

to calculate the rate error ([Figure 3](#); parallel fault rates on the left and normal fault rates on the right). Profile P1 mainly includes the Elashan, Tuolaishan, the Northern Qilian Shan, and Yumushan faults. It can be seen that the Tuolaishan fault is given priority to with left-lateral strike-slip, and compressional movement, has a strike-slip rate of 4.0 mm/yr and a compressional rate of 2.9 mm/yr. The Northern Qilian Shan and Yumushan faults also have the left-lateral strike-slip and compressional movement, however, their fault activity is weak and the slip rate is about 1 mm/yr. Profile P2 mainly includes the Riyueshan, Lenglongling, the Northern Qilian Shan, and Longshoushan faults. The Lenglongling fault has a left-lateral strike-slip rate of 3.9 mm/yr and a compressional rate of 2.2 mm/yr. The Riyueshan fault has no obvious strike-slip and a compressional rate of 2.2 mm/yr, while the Longshoushan fault mainly shows thrusting deformation and has a compressional rate of 1.9 mm/yr. Profile P3 mainly includes the Lajishan and the Jinqianghe–Maomaoshan–Laohushan faults. The Jinqianghe–Maomaoshan–Laohushan fault has a strike-slip rate of 3.9 mm/yr and a compressional rate of approximately 1.3 mm/yr. The Lajishan fault has weak strike-slip and extensional effects. Profile P4 includes the Maxianshan, Haiyuan, and Xiangshan–Tianjingshan faults, which have a left-lateral strike-slip rate of 4.3 mm/yr and a compressional



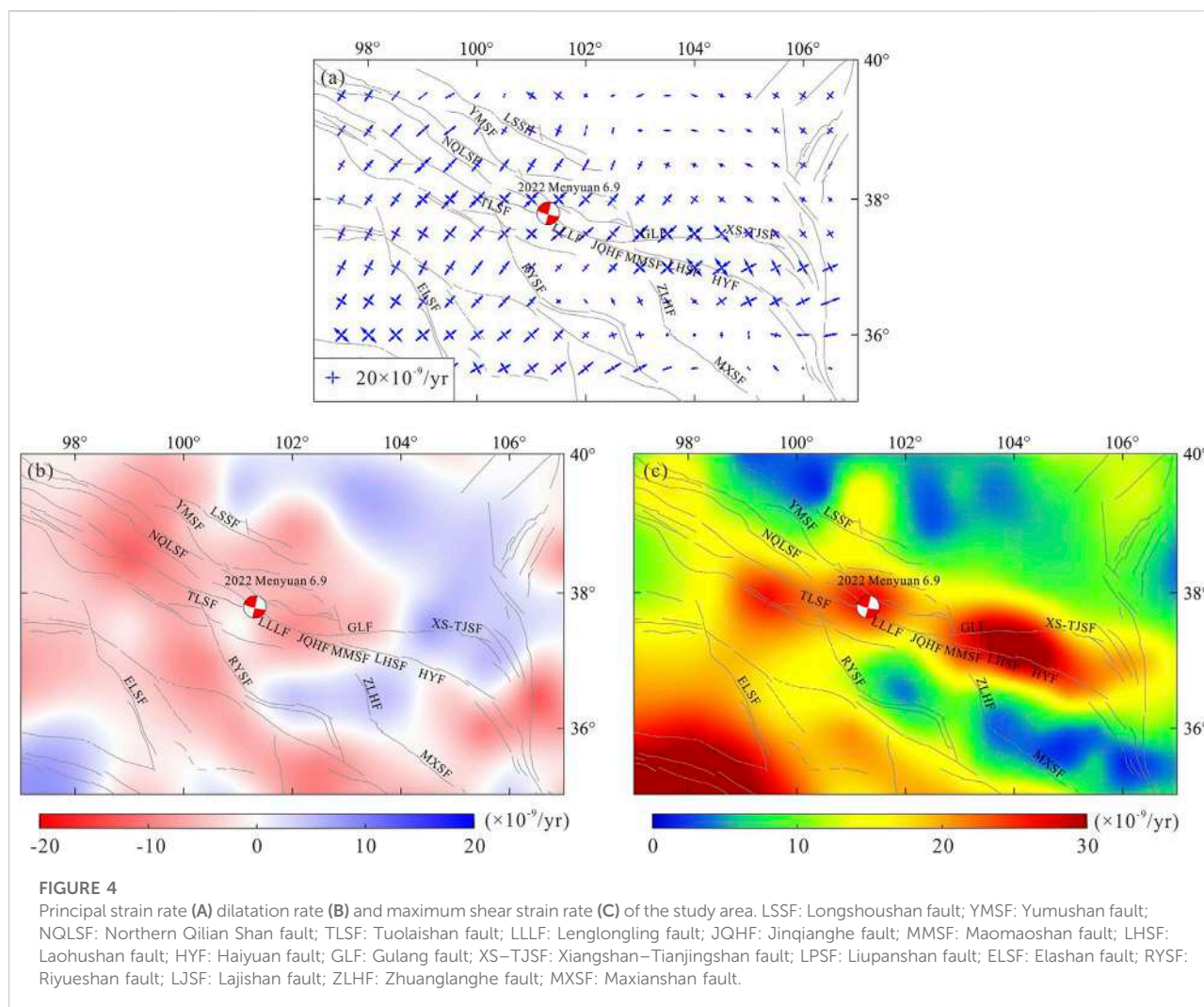
rate of approximately 1 mm/yr. However, the Maxianshan and Xiangshan-Tianjingshan faults are not active at present and have weak strike-slip and compression rates.

In the whole region, it can be seen that profiles P1 and P2 cross the whole Qilian Shan region, and the profiles from SW to NE exhibit left-lateral strike-slip with thrusting deformation. The velocity changes along parallel faults were mainly concentrated on both sides of the faults, and the velocity changes along normal faults were linear gradients. The total crustal shortening rate is 6–7 mm/yr. This shortening is distributed throughout the Qilian Shan across a width of 200–250 km, the linear velocity gradient suggests relative homogeneous convergence across the whole mountains rather than major slip on a particular faults or structures, which supports the “continuous deformation model” of the Tibetan Plateau (England and Houseman, 1986). In profiles P3 and P4, the region mainly showed a

left-lateral strike-slip, and compression shortening was not evident. The velocity step along the parallel fault direction is concentrated on the Qilian-Haiyuan fault, with a width of 50 km. The fault normal velocity is weakened in profile P3, the crustal shortening rate is reduced to 3 mm/yr and the width is reduced to 150 km. The faults in profile P4 are almost non-compression, and the crustal shortening rate of the entire profile is only 1–2 mm/yr. In addition, it can be seen from the four profiles that the parallel and normal fault velocities of the GPS in the Alashan block are basically unchanged.

2.3 Strain rate field

The spatial inconsistency of the GPS horizontal velocity field is a direct reflection of crustal deformation, and the



strain-rate field is an important index for describing regional deformation. This index is not affected by the reference frame and can reflect the regional deformation characteristics at different resolutions (Yang et al., 2002; Meng et al., 2009). To visualize the crustal deformation of the studied region, we calculated the strain rate field using the continuous function method developed by Shen et al. (2015). This method interpolates a set of discrete GPS velocities and introduces a spatial weighting function with a smoothing distance optimally determined according to the *in situ* data quality. It is robust because it does not depend on certain assumptions regarding the data. As shown in Figure 4A, the results of the principal strain rate show that the crust rotates clockwise along the northeastern margin of the Tibetan Plateau, and the orientation of the principal compressional strain rate is mainly NE-NEE. The principal strain rate is larger along the Qilian-Haiyuan fault zone, and the extensional strain increases gradually from west to east, while the principal

strain rate in the Alashan block is smaller. The dilatation rate (Figure 4B) shows that it is mainly compressional in the Qilian Shan, changes to extension east of the Maomaoshan fault, and then changes to compression in the surrounding area of Liupanshan. There was also a weak extension in the Alashan block. The maximum shear strain rate (Figure 4C) shows that there is an area with a high shear strain along the Qilian-Haiyuan fault zone, which is consistent with the conclusion that the fault is mainly strike-slip. The strain rate field corresponded well with the GPS profile. The strike-slip movement is evident in the Qilian-Haiyuan fault zone, and the compression movement gradually weakens on both sides of the fault zone from west to east. In addition, before the 2022 Menyuan $M_6.9$ earthquake, the dilatation rate at the seismogenic location was small, whereas the maximum shear strain rate was high ($24 \times 10^{-9}/\text{yr}$), which was consistent with the left-lateral strike-slip being the focal mechanism of this earthquake.

3 Fault locking degree and slip rate deficit on the Qilian–Haiyuan fault zone

3.1 Block Model

We used the negative dislocation inversion model, which assumes that GPS velocities are a combined effect of rotation, interseismic elastic strain from bounding fault locking, and horizontal strain within blocks (Eq. 1). The block rotation Euler pole, fault slip rate, and block boundary fault coupling coefficients (Φ) are retrieved by grid search and simulated annealing using a downhill simplex method. The model principle, inversion process, and parameter control have been introduced in a previous study (McCaffrey, 2009). The modeling was implemented using the open-source TDEFNODE software (McCaffrey, 2009). The quality of model parameter fitting is represented by Eq. 2, when χ_n^2 approaches 1, the model is considered to represent the best-fitting observation data (McCaffrey, 2002).

$$\bar{V}_{sf} = \bar{V}_{br} + \bar{V}_{is} + \bar{V}_{fs} \quad (1)$$

Herein, $\bar{V}_{sf}$ is the measured surface velocity, $\bar{V}_{br}$ is the velocity caused by rotation of the blocks, $\bar{V}_{is}$ is the velocity caused by the internal strain of the blocks, and $\bar{V}_{fs}$ is the velocity caused by the negative dislocation effect of fault locking.

$$\chi_n^2 = \left[\sum_{i=1}^n (r_i / f \sigma_i)^2 \right] / dof \quad (2)$$

Herein, r_i is the residual, σ_i is the standard deviation, f is the weight factor, and dof is the degrees of freedom (number of observations minus number of free parameters).

Our model was constructed as follows: The northeastern margin of the Tibetan Plateau is divided into four tectonic blocks based on the GPS velocities and geological and seismological evidence for active faulting (Zhang et al., 2003; Wang et al., 2011; Li et al., 2015): the Alashan, Lanzhou, Gonghe, and Qilian blocks (The orange lines in Figure 2). In our inversion, the Alashan block acted as rigid block (Zhang et al., 2005; Wang et al., 2009), while the other blocks underwent internal deformation. We removed the sites that were obviously different with regard to movement trend and size inside and around the block, and 153 GPS sites were selected to participate in the inversion. During the inversion, the simplified Qilian–Haiyuan fault zone was a single continuous fault with NWW strike and SSW dip and dip angles of 80°, 75°, and 70° from west to east. The fault plane was composed of horizontal and deep nodes. Along the strike of the Qilian–Haiyuan fault zone, the distance between the nodes was approximately 30 km, with a total of 26 nodes. Seven nodes were set in the depth direction and the distances were 0.1, 5, 10, 15, 20, 25, and 30 km. We also assume that the fault is fully locked at the surface ($\Phi = 1$) and freely slips below a depth of 30 km ($\Phi = 0$), with the Φ decreasing down-dip from

1 to 0 between 0.1 and 30 km at depth (McCaffrey, 2002; Wang et al., 2003).

3.2 Fault locking on the Qilian–Haiyuan fault zone

We present the fault locking (Figure 5) of the Qilian–Haiyuan fault zone for our best-fitting model on the basis of the parameter setting and inversion strategies described above. It can be seen that the Jinqianghe fault, the Lenglongling fault, and the eastern segment of the Tuolaishan fault are strongly locked; the locking depth of the Jinqianghe fault and the western segment of the Lenglongling fault is about 25 km. The 2022 Menyuan $M_s 6.9$ earthquake occurred in the strongly locked area of the western segment of the Lenglongling Fault. The locking depth of the eastern segment of the Lenglongling fault is only 6 km, and the locking depth of the eastern segment of the Tuolaishan fault is 12–24 km. There is weak locked in the western segment of the Tuolaishan fault, with a locking depth of 6–15 km. However, the middle segments of the Tuolaishan, Maomaoshan, Laohushan, and Haiyuan faults are unlocked or weakly locked. They are weakly locked at the shallow part and show creep sliding below a depth of 5 km.

3.3 Slip rate deficit on the Qilian–Haiyuan fault zone

When the fault is locked, the slip rate deficit accumulates near the fault in the form of strain energy (McCaffrey et al., 2007). The slip rate deficit can indicate how fast the slip of the two sides of the fault is transformed into strain energy, which is of great significance to the judgment of fault-associated seismic hazards (Jolivet et al., 2013). The distribution of the slip rate deficit of the Qilian–Haiyuan fault zone (Figure 6) is consistent with the distribution of the fault locking, and the slip rate deficit gradually decreases from the shallow part to the deep part of the fault. In the eastern segment of the Tuolaishan fault, the slip rate deficit is 6–6.5 mm/yr and the depth is 12–24 km. The slip rate deficit of the western segment of the Lenglongling fault is 6 mm/yr and the depth is 25 km, while the slip rate deficit of the eastern segment is 5.5 mm/yr and the depth is only 6 km. The slip rate deficit of the Jinqianghe fault is 5–5.5 mm/yr and the depth is 25 km. However, in the middle and western segment of the Tuolaishan fault, the slip rate deficit is mainly 3–4 mm/yr, and only the depth of the middle-west segment reaches 15 km, while the rest of the fault is concentrated in the shallow surface of 5 km. The slip rate deficit of the Maomaoshan fault gradually decreases from 3 mm/yr in the shallow region to 1.7 mm/yr in the deep region. The slip rate deficit of the Laohushan and Haiyuan faults is about 3 mm/yr, only distributed in the shallow surface of 4 km, gradually changing to 0 mm/yr in the deep region.

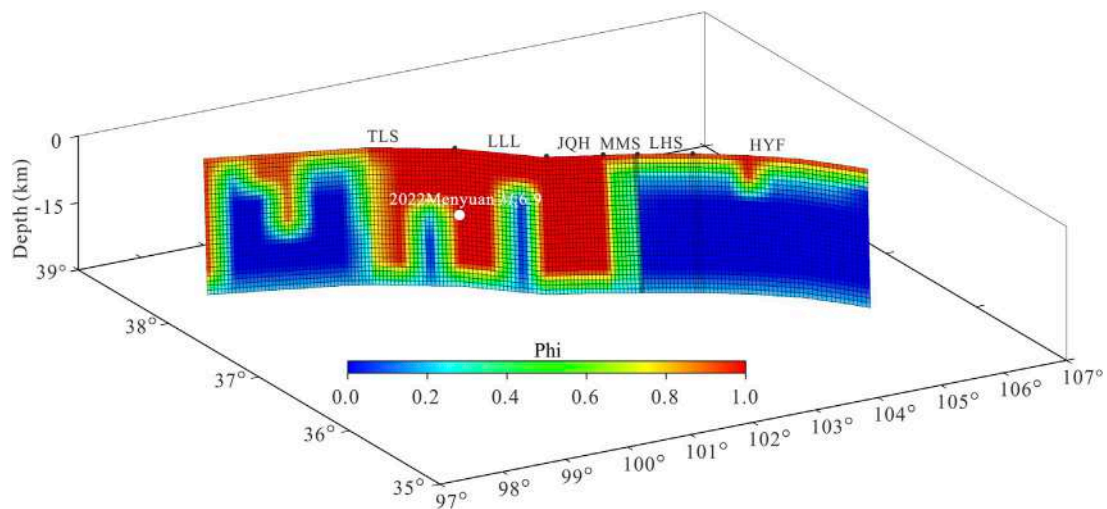


FIGURE 5

Locking degree of the Qilian–Haiyuan fault zone. Hypocenter parameters were obtained from Global CMT Catalog (<http://www.globalcmt.org/CMTsearch.html>). TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQH: Jinjianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYF: Haiyuan fault.

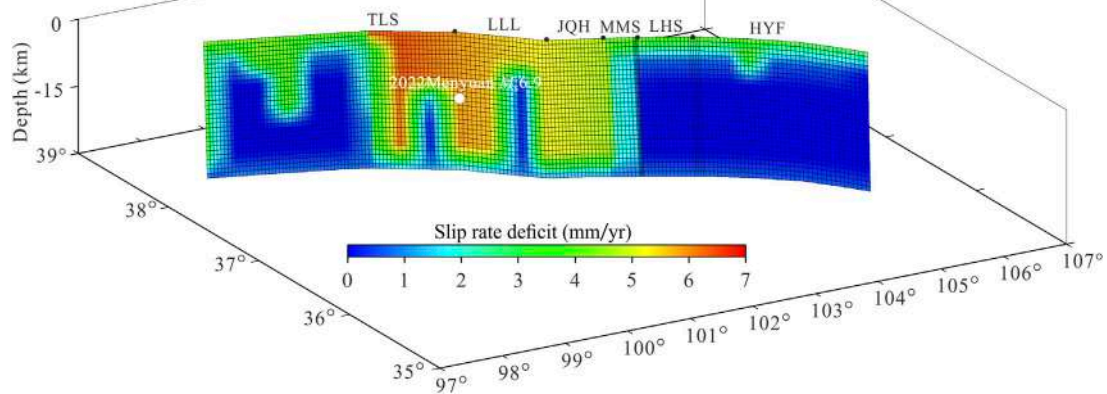


FIGURE 6

Slip rate deficit of the Qilian–Haiyuan fault zone. Hypocenter parameters were obtained from Global CMT Catalog (<http://www.globalcmt.org/CMTsearch.html>). TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQH: Jinjianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYF: Haiyuan fault.

4 Distribution of small earthquakes in the Qilian–Haiyuan fault zone

To study the relationship between the locking degree and small earthquakes of the Qilian–Haiyuan fault zone, based on the relocation catalog of small earthquakes with $M_L 1.5$ and above

during 2009–2019 provided by Mr. Long Feng of the Sichuan earthquake agency, the earthquakes occurring within 15 km on both sides of the fault strike were selected and projected onto the fault plane along the fault strike (Figure 7); it can be seen that the distribution of earthquakes on the Qilian–Haiyuan fault zone is dense and uneven in each segment of the fault. The locking depth

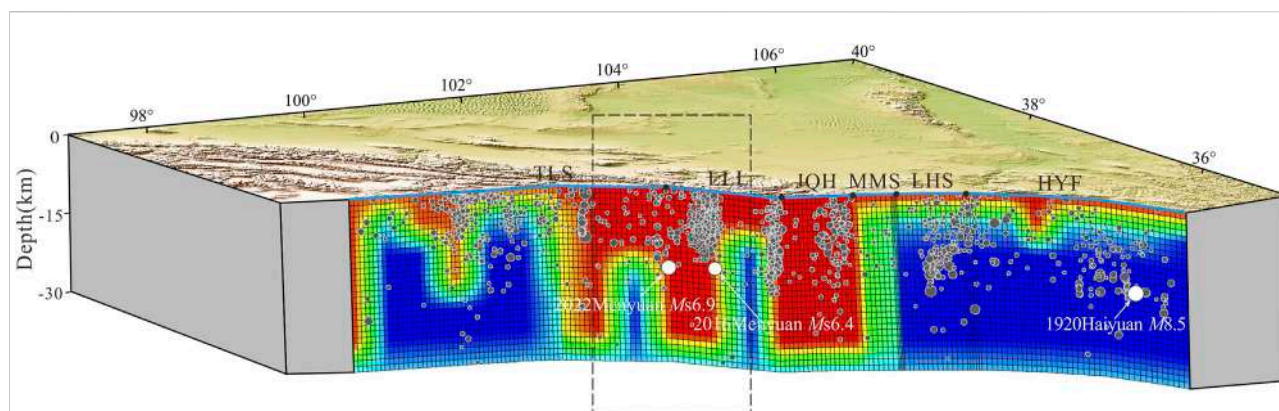


FIGURE 7

Locking degree and distribution of small earthquakes in the Qilian–Haiyuan fault zone. The white circles indicate the locations of the 1920 Haiyuan M8.5, 2016 Menyuan $M_s6.4$, and 2022 Menyuan $M_s6.9$ earthquakes. The gray circles are the relocation catalog of small earthquakes during 2009–2019. Black dotted rectangle shows the profile of small earthquakes in Figure 9. TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQHF: Jinqianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYF: Haiyuan fault.

is consistent with the depth of small earthquakes, which indicates that the fault is locked at a depth of 25 km (Schmittbuhl et al., 2015; Li et al., 2020). In the middle and western segments of the Tuolaishan fault, the fault is weakly locked, the locking depth is shallow, and there are more small earthquakes, whereas in the eastern segment, the fault is strongly locked, the locking depth is deep, and the distribution of small earthquakes is lesser. The Lenglongling fault is strongly locked; the locking depth of its western segment is larger than that of its eastern segment, small earthquakes are concentrated in the middle-west segment of the fault, and the depth of this fault is more than 14 km. The 2016 Menyuan $M_s6.4$ earthquake occurred in this area, the occurrence of intensive small earthquakes here is related to the energy release of the 2016 Menyuan $M_s6.4$ earthquake. The reason why the epicenter of this earthquake is in the strong locked area may be that because the GPS data (2015–2021) contain information regarding the 2016 Menyuan $M_s6.4$ earthquake, they also reflect the characteristics of strong locked in this region before the earthquake. There are fewer earthquakes at the intersection of the Lenglongling fault and the Tuolaishan fault, and the 2022 Menyuan $M_s6.9$ earthquake occurred in this area, the two Menyuan earthquakes were located on the east and west sides of the strongly locked segment of the Lenglongling fault, which is well explained by the occurrence of strong earthquakes in the strongly locked segment of this fault. The impact of this earthquake on faults will be discussed in the next section. The Jinqianghe fault shows strong locked and a large locking depth. There are many small earthquakes in the eastern segment of this fault, at the intersection of the Jinqianghe and Lenglongling faults; these earthquakes are concentrated and distributed in the band at a depth of 18 km. This may be due to the strong tectonic activity at the intersection of the Gulang fault and the Qilian–Haiyuan fault,

which makes the occurrence of earthquakes more likely (Gao, 2018; Liu et al., 2020). The locking degree of the Maomaoshan fault is approximately 0.5, the earthquake distribution is low. Further, the Laohushan fault is weakly locked and shows a shallow the locking depth, there are many small earthquakes with a depth of less than 18 km. The large number of earthquakes corresponds to the conclusion of a study by Jolivet et al. (2012), which reported that there is a creep segment in the Laohushan fault; The Haiyuan fault is weakly locked and shows a shallow locking depth. The frequency of the occurrence of small earthquakes was lower in the western segment and higher in the eastern segment, and the depth of the small earthquakes was approximately 18 km, which is consistent with the epicenter depth of the 1920 Haiyuan M8.5 earthquake (Zhan et al., 2004; Han et al., 2008). Since the 1920 Haiyuan M8.5 earthquake, which occurred in the southeastern segment of the Haiyuan fault, this fault may have been affected by the fault adjustment-associated movement after the earthquake, due to which many small earthquakes still occur in this region.

5 Discussion

5.1 Reliability analysis of results

The χ_n^2 is the ratio of the variance of posterior unit weight to the variance of the anterior unit weight. The inversion result was better when the value was approximately 1. According to Eq. 2, it can be seen that the size of χ_n^2 is related to the weight factor f , which ranges from 1 to 5 for the GPS horizontal velocity field data error (Mao et al., 1999). In the inversion, the value of χ_n^2 can be adjusted as close to 1 as possible by gradually changing the size of f , via a large number of test

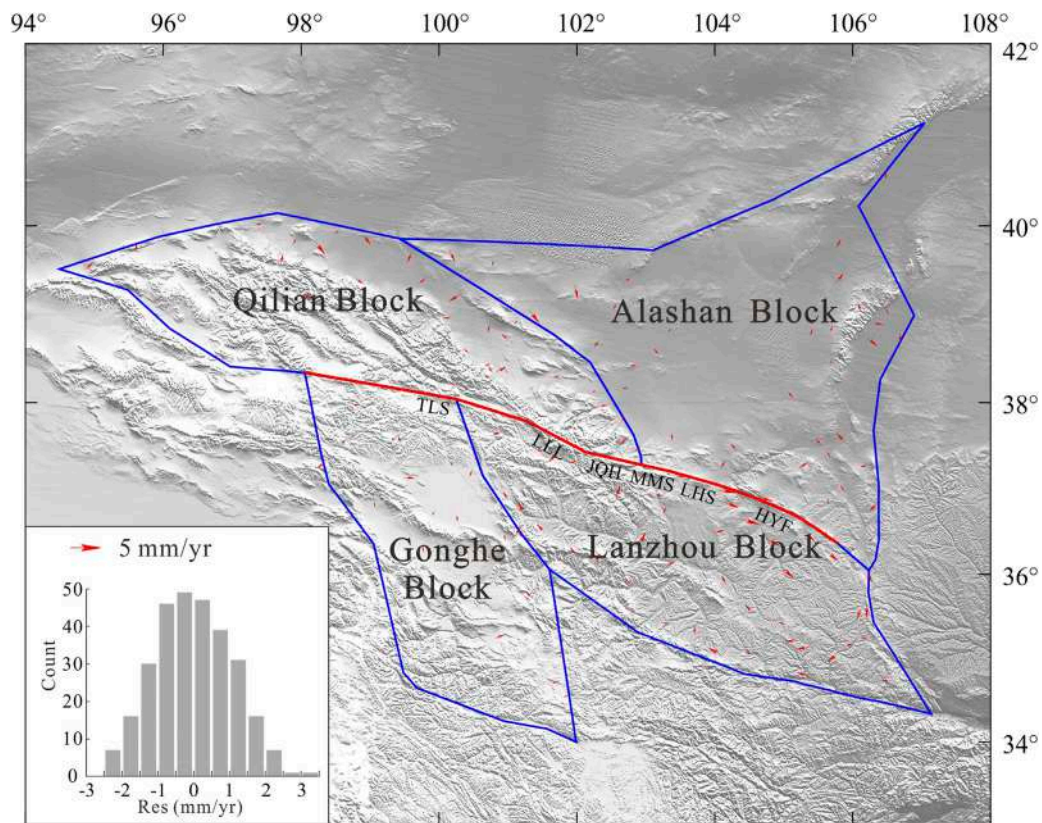


FIGURE 8

GPS velocity residuals associated with the block kinematic model and their statistical histogram. The red line is the Qilian–Haiyuan fault zone.

The blue lines are active block boundaries. Red arrows represent the GPS velocities residuals. TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQHF: Jinqianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYF: Haiyuan fault.

calculations, we found that when the weight factor $f = 3.9$, $\chi_n^2 = 1.06$ and the WRMS is 1.07 mm/yr, which is considered to represent the best-fitting model. At the same time, the inversion calculation of $f = 3.7$ and $f = 4.1$ shows that the locking degree of the fault is basically consistent with the best-fitting model, indicating that the inversion results have good stability. To better illustrate the model selection and fitting effects, we drew the residual distribution of the best-fitting model (Figure 8). It can be seen from the figure that the direction of the residual distribution is random, and only a few sites have large residual GPS velocity values, whereas most stations inside the block and near the fault have small velocity residual values (<2 mm/yr). In addition, statistical analysis of the model-fitting residual distribution and internal strain residual of the block is also an important criterion for judging the quality of model inversion (Li et al., 2016). The statistical results of the velocity residual values conform to the Gaussian normal distribution; thus, the model is considered to be effective and the fitting results are good. Overall, the current model explains the GPS data and describes the seismic behavior of the northeastern margin of the Tibetan Plateau.

5.2 Seismic hazard analyses along the Qilian-Haiyuan fault zone

When an active fault is locked, it continuously accumulates energy under the action of stress. When it exceeds its stability limit, an earthquake occurs and releases energy, which then enters the next seismic cycle (Wallace et al., 2004; Zhang et al., 2013a; Qiu and Qiao, 2017). To analyze the seismic hazards of the Qilian–Haiyuan fault zone in the future, we studied the seismic hazards of each segment of the fault on the basis of the data of the fault locking degree and occurrence of small earthquakes.

In the eastern segment of the Tuolaishan fault, the intersection of the Tuolaishan fault and the Lenglongling fault, and the western segment of the Jinqianghe fault, the locking degree is strong, the locking depth and slip rate deficit (>5 mm/yr) are large, and the distribution of small earthquakes is low, indicating that the energy accumulation of these fault segments is fast, and the seismic hazard is high in the future. The eastern segment of the Lenglongling fault has strong locking and shows a low frequency of small earthquakes, but its locking

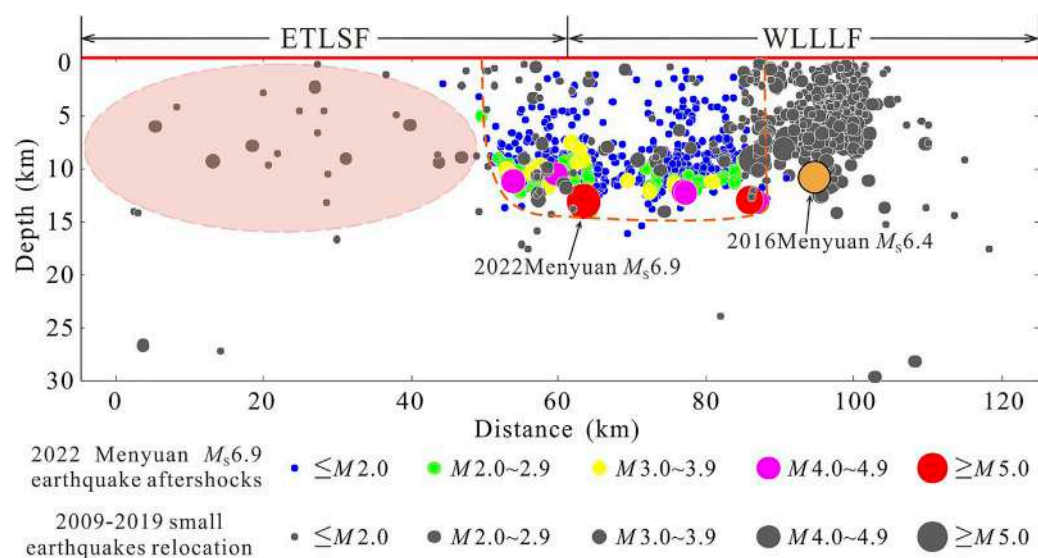


FIGURE 9

Profile of the 2009–2019 small earthquakes relocation (colored circles) and 2022 Menyuan $M_{5.6.9}$ earthquake aftershocks (gray circles). Location of profile is shown in Figure 7. The pink ellipse corresponds to area of high risk for future earthquake. ETLFS: eastern segment of the Tuolaishan fault; WLLLF: western segment of the Lenglongling fault.

depth is low (only 6 km); the degree of locking in this segment may be increasing, and the seismic hazard here is moderate. The shallow region of the Maomaoshan fault has a high degree of locking, the locking degree is only 0.5 at the deep region of this fault. The distribution of small earthquakes was less, the current energy accumulation rate of the fault was relatively slow, the slip rate deficit was only 2–3 mm/yr, and the seismic hazard was moderate. The middle and western segments of the Tuolaishan, Laohushan, and Haiyuan faults are weakly locked, with a small locking depth, low slip rate deficit, and high frequency of earthquakes, indicating that these fault segments show frequent seismic activity, a weak energy accumulation level, and a low seismic hazard.

To analyze the impact of the 2022 Menyuan $M_{5.6.9}$ earthquake on the future seismicity of the fault, we collected the relocation results of aftershocks within 9 days after the earthquake (Fan et al., 2022), and the regions located within 60 km on each side of the earthquake epicenter (black dotted box in Figure 7) was selected to plot the profile of the 2009–2019 small earthquakes relocation and 2022 Menyuan $M_{5.6.9}$ earthquake aftershocks (Figure 9). It can be seen that the aftershocks are mainly distributed between a depth of 5–15 km, and the aftershocks occurred mainly between the western segment of the Lenglongling fault and the eastern segment of the Tuolaishan fault, with a length of approximately 40 km, consistent with other research results (Xu et al., 2022; Yang et al., 2022), which is longer than the surface rupture zone (20 km) observed in a field geological survey. Combined with the focal mechanism data of the

2022 Menyuan $M_{5.6.9}$ earthquake and the distribution of aftershocks, we assume that the Lenglongling and Tuolaishan faults will be connected to a large strike-slip fault under the influence of this earthquake. Based on the InSAR coseismic deformation field and optical remote sensing interpretation, it was also found that the surface rupture caused by the earthquake extended to the Tuolaishan fault (Li et al., 2022; Yang et al., 2022). In addition, aftershocks were distributed on the Lenglongling fault and stopped at the location of the 2016 Menyuan $M_{5.6.4}$ earthquake, indicating that the middle segment of the Lenglongling fault, which has experienced an earthquake, is still in the post-earthquake adjustment phase, and the occurrence of aftershocks cannot continue without the accumulation of stress. Compared with the relocation distribution of small earthquakes from 2009 to 2019, the aftershocks of this earthquake filled the gap of small earthquakes in this segment and continued to migrate westward along the 2016 Menyuan $M_{5.6.4}$ earthquake. It is found that there is a seismic gap with a length of about 50 km in the eastern segment of the Tuolaishan fault to the west of the aftershock distribution. If the Tuolaishan and Lenglongling faults are connected, the next strong earthquake may migrate westward from the Lenglongling fault to the Tuolaishan fault, this segment is strongly locked and shows a high slip rate deficit, similar to the characteristics observed before the 2022 Menyuan $M_{5.6.9}$ earthquake. Moreover, Xu et al. (2022) hypothesized that there is still some stress accumulation in this area, and thus, a risk of strong earthquakes. Therefore, more attention needs to be paid to the eastern segment of the Tuolaishan fault.

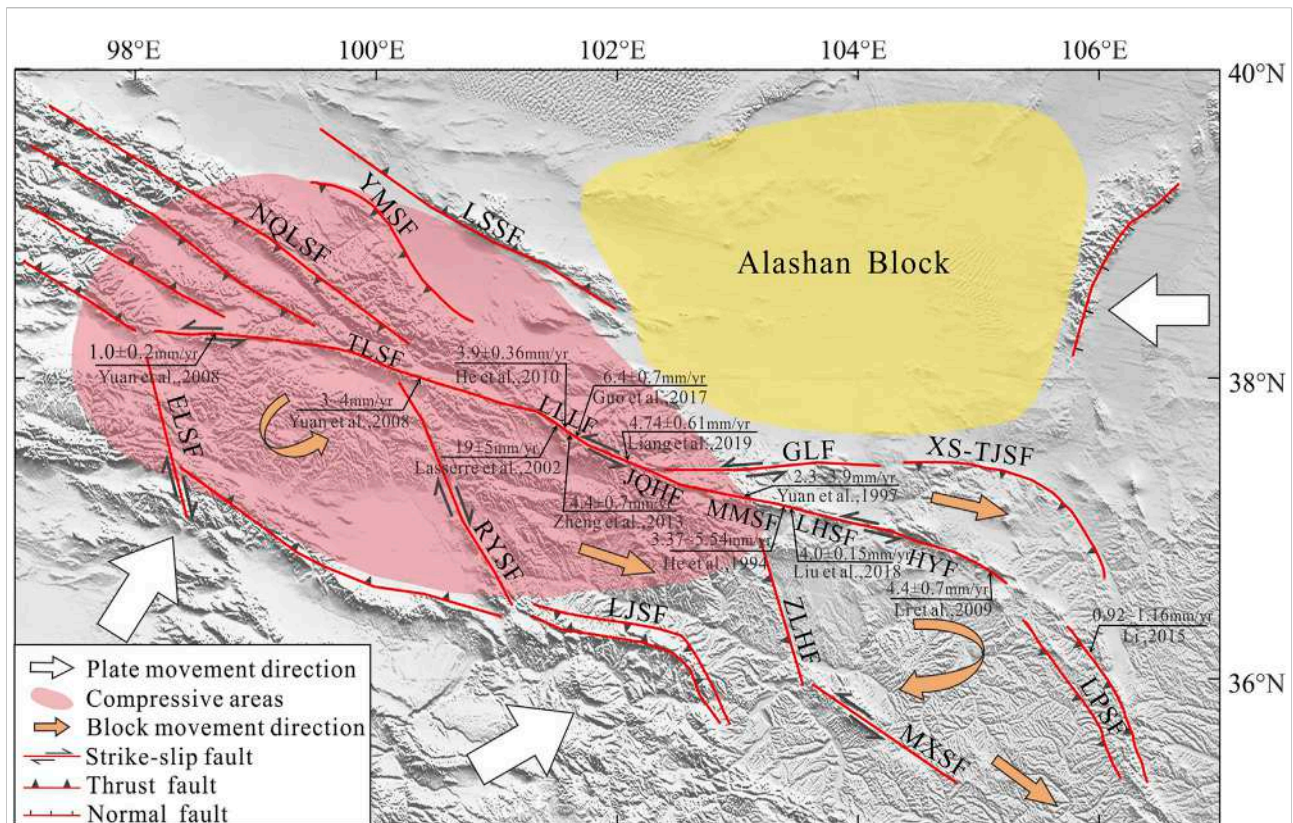


FIGURE 10

Tectonic transformation pattern in the northeastern margin of the Tibetan Plateau (modified from Yuan et al. (2004) and Zheng et al. (2013)).

Red lines represent fault lines. White arrows represent the plate movement direction. Orange arrows represent the block movement direction. Pink shadow represents the compressive area. Yellow shadow represents the blocking area. LSSF: Longshoushan fault; YMSF: Yumushan fault; NQLSF: Northern Qilian Shan fault; TLSF: Tuolaishan fault; LLLF: Lenglongling fault; JQHF: Jinjianghe fault; MMSF: Maomaoshan fault; LHSF: Laohushan fault; HYSF: Haiyuan fault; GLF: Gulang fault; XS-TJSF: Xiangshan-Tianjingshan fault; LPSF: Liupanshan fault; ELSF: Elashan fault; RYSF: Riyueshan fault; LJSF: Lajishan fault; ZLHF: Zhuanglanghe fault; MXSF: Maxianshan fault.

5.3 Regional crustal deformation model

The fault slip rate can not only reflect the most recent behavior of fault movement, but also reflect the main features of regional dynamics (Zhang et al., 2004; Gan et al., 2007; Zhang et al., 2013b). Several previous studies regarding the strike-slip rate of the Qilian-Haiyuan fault zone have been performed thus far; the geological slip rates reported in these studies vary greatly, showing a maximum value of 19 ± 5 mm/yr (Lasserre et al., 1999). The reliability of these results is controversial (Zheng et al., 2013; Guo et al., 2017), however, most studies have reported that the geological slip rate ranges from 4 to 6 mm/yr (He et al., 1994, 2010; Yuan et al., 1997, 2008; Li et al., 2009; Guo et al., 2017; Liu et al., 2018; Liang et al., 2019). The results obtained using geodetic data show that the slip rate ranges from 2 to 8 mm/yr (Thatcher, 2007; Cavalié et al., 2008; Duvall and Clark, 2010; Loveless and Meade, 2010; Zheng et al., 2013; Li et al., 2016). However, the fault slip rate obtained in this study is mainly between 3.9 and 4.3 mm/yr,

which is basically consistent with, or slightly smaller than, the strike-slip rates reported in most geological and geodetic studies. It can be considered that the present movement of the Qilian-Haiyuan fault zone is basically consistent with the characteristics of geological long-term movement, the fault activity has a good inheritance, and the main segment of the fault maintains a strike-slip rate of 4 mm/yr, indicating that the overall motion of the fault is consistent and corresponds to the concentration of the maximum shear strain rate in the Qilian-Haiyuan fault zone, while the strike-slip rate decreases rapidly at both ends of the fault (Yuan et al., 2008; Li, 2015). Furthermore, the motion characteristics of this strike-slip fault are similar to those reported in previous studies (Dawers et al., 1993; Duvall and Clark, 2010; Harkins et al., 2010; Zheng et al., 2013). Many studies have discussed strike-slip rates on the Qilian-Haiyuan fault, but only a few studies have summarized the vertical components of slip rates (and shortening rates). In accordance with the GPS results discussed above, the compressional rate of the Qilian-Haiyuan fault zone

decreases from 2.9 mm/yr in the western segment to about 1 mm/yr in the eastern segment, these results are consistent with the dilatation rates observed. At the same time, there was good correspondence with the topography observed. The west side is dominated by high-altitude mountains due to strong compression, whereas the east side is dominated by low-altitude basins due to weak compression. This indicates that the fault compressional rate may be related to crustal shortening (formation basin and uplifted mountain).

Based on the above characteristics, we discuss the regional current crustal deformation patterns. Under the dual action of the NE pushing of the Tibetan Plateau and the blocking of the Alashan block, strong tectonic deformation occurred in the northeastern margin of the Tibetan Plateau, and the direction of crustal movement was clockwise (Yuan et al., 2004). The Qilian Shan area is mainly subjected to strong compression, which results in the bending and rupture of strata and the formation of the Qilian fold system. There is obvious crustal shortening movement from the Elashan fault to the Yumushan fault, this is also affected by the interactions of block rotation and faults with different properties in the region and the transformation of the block rotations (Duvall and Clark, 2010; Ge et al., 2013) from the Qilian Shan to the east along the northeastern margin of the Tibetan Plateau. The crustal shortening rate gradually decreases, and the range of compression deformation also decreases. There is no obvious crustal shortening near the Haiyuan fault, which is a good explanation for the NE pushing being mainly distributed and absorbed by thrust faults, crustal thickening, and locally high marginal mountains. The decrease in the slip rate suggests that the boundary fault does not transfer a significant portion of the convergence between India and Asia out of India's path into Eurasia but merely redistributes the crustal thickening. The regional left-lateral strike-slip rate is mainly concentrated in the Qilian–Haiyuan fault zone, the strike-slip rate of each segment is basically the same, while the strike-slip movement of the other faults is not strong (the strike-slip rates for most of the faults are less than 1 mm/yr), indicating that the present-day crustal deformation in the northeastern margin of the Tibetan Plateau is mainly distributed in the shortened region of the crust on the Qilian Shan area and left-lateral strike-slip localized on the Qilian–Haiyuan fault zone (Figure 10).

6 Conclusion

In the paper, we used the GPS velocity field results from 2015 to 2021 and a negative dislocation program to invert the fault locking degree and the slip rate deficit, we obtained data regarding the current slip rate and strain rate field in the Qilian–Haiyuan fault zone. Combined with the results of the analysis of the relocation of small earthquakes, the characteristics of this fault zone before the 2022 Menyuan $M_s6.9$ earthquake were studied, the future seismic hazards of each segment of this

fault zone were analyzed, and the regional crustal deformation pattern is discussed. The results are as follows:

- (1) Before the 2022 Menyuan $M_s6.9$ earthquake, the dilatation rate at the seismogenic location was small, while the maximum shear strain rate was high ($24 \times 10^{-9}/\text{yr}$). The seismogenic fault was strongly locked, the locking depth was deep, the slip rate deficit was large, and the distribution of small earthquakes was relatively few, which indicates that these characteristics are closely related to the occurrence of strong earthquakes. In addition, according to the aftershock relocation results, it is believed that the earthquake may link the Lenglongling and Tuolaishan faults into a large strike-slip fault.
- (2) The eastern segment of the Tuolaishan fault is strongly locked, with high locking depth and large slip rate deficit, and the distribution of small earthquakes is few, combined with the small earthquakes and the locking degree before the 2022 Menyuan $M_s6.9$ earthquake, indicate that the eastern segment of the Tuolaishan fault is highly likely to have strong earthquake in the future, which requires further attention.
- (3) The strike-slip rate of the Qilian–Haiyuan fault zone is mainly between 3.9 and 4.3 mm/yr, the overall movement of the fault is consistent, and the compressional rate gradually decreases from 2.9 mm/yr in the western segment to 1 mm/yr in the eastern segment, the compressional rate of the fault may be related to crustal shortening (formation basin and uplift mountain). Therefore, the present-day crustal deformation in the northeastern margin of the Tibetan Plateau is mainly distributed in the shortened region of the crust on the Qilian Shan and the left-lateral strike-slip localized on the Qilian–Haiyuan fault zone.

Data availability statement

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author.

Author contributions

LL: Conceived this research, Writing - original draft; editing, WZ: GPS data processing LJ: Guide and supervision. LZ: Methodology, Block Model FJ: Strain Rate Field. All authors reviewed and supervised the manuscript and approved it for publication.

Acknowledgments

We are grateful to McCaffrey for permitting us to use of the TDEFNODE software. We would like to thank professor Fang Lihua and Long Feng for providing relocation results of aftershocks and

small earthquakes. Some figures and maps were prepared using Generic Mapping Tools software (Wessel et al., 2013).

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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OPEN ACCESS

EDITED BY

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 18 May 2022

ACCEPTED 08 July 2022

PUBLISHED 12 August 2022

CITATION

Liu K, Li Y, Nan Y, Liu B and Wang W
(2022), Detailed shallow structure of the
seismogenic fault of the
1976 M_s7.8 Tangshan
earthquake, China.
Front. Earth Sci. 10:946972.
doi: 10.3389/feart.2022.946972

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Detailed shallow structure of the seismogenic fault of the 1976 M_s7.8 Tangshan earthquake, China

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In 1976, an M_s7.8 strong earthquake occurred in the Tangshan region of
Northeast China within an ancient craton. In this study, we conducted
shallow seismic exploration, drilling-based exploration, and exploratory
trench excavation of ground fissures in the Tangshan Fault Zone and
obtained a high-resolution shallow seismic profile. Through analysis of its
lithology and sedimentary cycle, we constructed a composite drilling profile
across the faults. Coupled with the shallow fault combination patterns identified
from the exploratory trenches, the profile reveals that the Guye-Nanhu Fault
was the shallow response fault in the seismogenic structure of the Tangshan
earthquake. This fault is a strike-slip fault with a positive flower structure; the
reverse fault branches become progressively steeper with depth and converge
downward toward the vertical main strike-slip fault. A high-angle thrust fault is
located to the west of the main strike-slip fault, and a series of small-scale
normal faults appear in front of this fault, owing to local extension. The tectonic
form revealed by the deep seismic reflection profiles further supports the
superficial tectonic model. The near-NEE orientation of the stress field in
North China and the lower strike-slip movement component of the Tanlu
Fault Zone facilitate faulting in the Tangshan–Hejian–Xingtai Fault Zone, which
lies diagonally in the middle rectangular area of the North China Plain faulted
basin. The detailed structural model of the seismogenic fault obtained by
various detection methods is of great significance for understanding the
seismogenic mechanism of the Tangshan earthquake.

KEYWORDS

Tangshan earthquake, shallow seismic exploration, composite drilling profile,
exploratory trench, deep seismic reflection profile, flower structure

Introduction

On 28 July 1976, an $M_s 7.8$ strong earthquake struck Tangshan city in the North China Plain region, killing 242,000 people and destroying the entire city. As this high-magnitude earthquake occurred in the North China craton, its seismogenic structure has attracted substantial research attention.

For example, by investigating the homogenous distribution of fissures resulting from the Tangshan earthquake, it was initially proposed that the seismogenic fault was located in the Tangshan Fault Zone and corresponded to the NW-trending No. 5 seam of the Tangshan Mine, with a dip angle of $70\text{--}80^\circ$ (Wang et al., 1981). However, the results of shallow artificial seismic exploration indicated that the seismogenic fault was an NW-trending, high-dip, right-lateral strike-slip fault with a normal component, causing dislocation of the Holocene and Pleistocene strata (Hao and You, 2001). Conversely, it has been suggested that the seismogenic fault is a high-angle, west-dipping, thrust fault with a strike-slip component and an offset of 15 m that accrued during the Late Pleistocene (You et al., 2002). Furthermore, in a study of ground subsidence in a strong earthquake area, Qiu et al. (2005) discovered a large fault related to the Tangshan earthquake called the Fuzhuang-Xihe Fault, with a length of approximately 90 km and an offset of 3 m at the surface, characterized as a shovel-shaped normal fault.

In recent years, shallow artificial seismic exploration, drilling-based exploration, and exploratory trench excavation conducted near the remains of Niumaku and Tangshan Asylum have revealed that the Tangshan Fault Zone exhibits features of reverse and strike-slip movements at shallow depths (Liu, 2011). A composite drilling columnar profile at the earthquake relic site of Tangshan No. 10 Middle School revealed a near-vertical fault beneath the surface cracks, with a dip angle $>83^\circ$ and a middle Pleistocene vertical throw of 40 m (Liu et al., 2013). Moreover, Guo et al. (2011) analyzed a composite drilling profile and an exploratory trench in Sunjialou, Tangshan, and proposed that the seismogenic fault of the Tangshan earthquake was a normal fault (Guo et al., 2011, Guo et al., 2017; Guo and Zhao, 2019).

After the earthquake, numerous studies employed deep detection methods in the Tangshan earthquake area, from which crustal models and crustal thickness distributions were acquired (Zeng et al., 1985; Zhang et al., 1994). For example, a low-velocity layer was discovered in the lower part of the Tangshan fault block, which rapidly pinches out on both sides (Long and Zelt, 1991; Zhang et al., 1994; Lai et al., 1998; An et al., 2009). The epicenter of the main shock was located at the intersection of the NE-trending low-velocity belt and the NW-trending high-velocity belt (Lai et al., 1998). Moreover, Shao et al. (1986) proposed that the Tangshan earthquake was located in the transition zone of upper mantle differential movement. Huang and Zhao (2006) obtained a 3D crustal P-wave velocity structure

model of the Beijing–Tianjin–Tangshan area using accurate P-wave arrival time data from the dense digital seismic network of the capital region and found that high-conductivity and low-velocity anomalies under the Tangshan earthquake source area were related to the existence of fluids. Furthermore, according to data from a mobile seismic observation array and the receiver function inversion method, Liu et al. (2007) found that the crust–mantle interface in the Tangshan region has experienced clear fault-block uplift. Additionally, a 2D deep seismic reflection profile revealed the existence of a large flower structure system in the deep part of the Tangshan Fault Zone (Liu, 2011; Liu et al., 2011).

Li et al. (2017) argued that the presence of low-velocity bodies in the crust makes it easy for tectonic stresses to accumulate and thus trigger large earthquakes. Applying high-resolution seismic imaging of P and S waves, Li et al. (2018) proposed that the migration of magma from the mantle to the lower crust may be an important factor controlling strong earthquakes within the North China craton. In recent years, experimental work in dense urban areas using the horizontal-to-vertical spectral ratio (HVSr) method with a dense seismic array as a low-cost and effective survey method has revealed that the Tangshan Fault Zone has been damaged and modified by strong seismic activity since the Quaternary (Bao et al., 2018, Bao et al., 2019, 2021). Mearns and Sornette (2021) proposed “a transfer fault complex” to explain the dynamics and faulting mechanisms of the Tangshan earthquake. In this study, shallow seismic exploration, composite drilling profile, and exploratory trenches have been used to investigate the surface rupture zone of the 1976 Tangshan earthquake. Furthermore, the detailed structural form of the shallow structure of the seismogenic fault has been obtained to better understand the seismogenic structure of the Tangshan earthquake.

Geological background

The research area is located at the boundary region of two geological units: the Yanshan uplift and the North China Plain faulted basin. Since the Mesozoic and Cenozoic, the Yanshan Mountains have been rising continuously, whereas the Bohai Bay Basin has experienced strong subsidence (Guo et al., 1977). The research area has been in a state of tectonic uplift for a long time, with Paleogene and Neogene strata now missing from the stratigraphic record (Figure 1). During the Quaternary period, the crust differentiated from north to south, and rivers became widespread. Except for a few hills with exposed bedrock scattered in the northeastern mountainous area and Tangshan city, most of the region is covered by Quaternary alluvial deposits. Regional strata and drilling profile data show that the sediments are predominantly sandy clay and sandy gravel, with highly variable thickness, which are controlled by paleogeographic evolution. Generally, the thickness of the Quaternary deposits

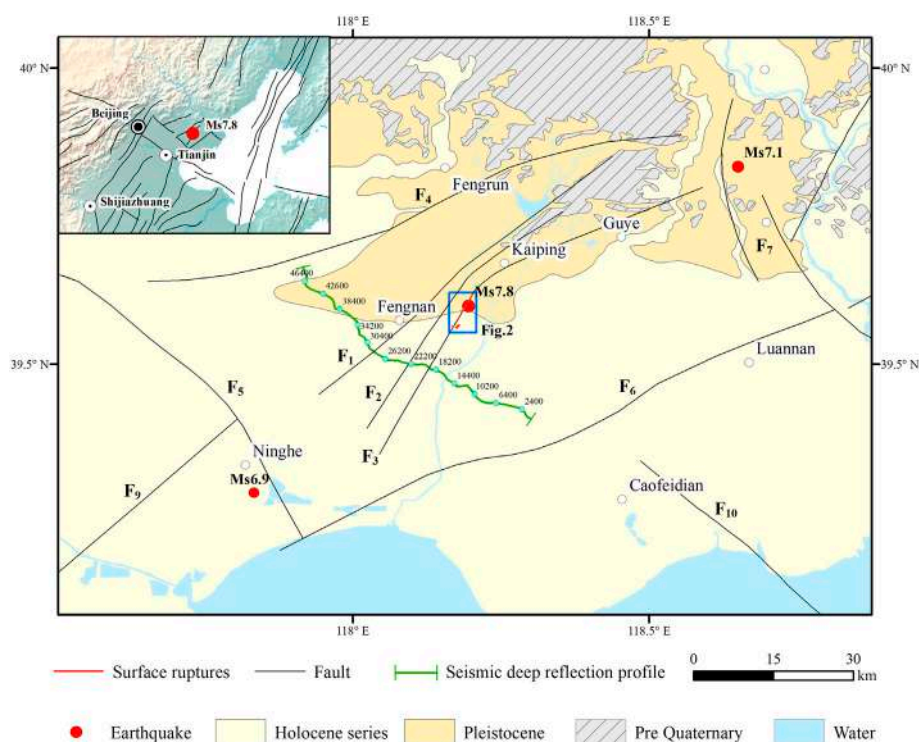


FIGURE 1

Map showing the fault distribution in the Tangshan research area in China. F₁: Douhe Fault; F₂: Weishan-Fengnan Fault; F₃: Guye-Nanhu Fault; F₄: Fengtai-Yejituo Fault; F₅: Jiyun River Fault; F₆: Ninghe-Changli Fault; F₇: Luanxian-Laoting Fault; F₈: Lulong Fault; F₉: Cangdong Fault; F₁₀: Bogezhuang Fault.

decreases from east to west but gradually increases from north to south on both sides of the Tangshan Fault.

Several active faults exist in the area. The NEE-trending Ninghe-Changli, Fengtai-Yejituo, NW-trending Luanxian-Laoting, and Jiyun River faults enclose the research area in a diamond-shaped block. The Tangshan Fault Zone forms a diagonal line across the diamond-shaped block and consists of three nearly parallel faults: the Douhe Fault (F₁), Weishan-Fengnan Fault (F₂), and Guye-Nanhu Fault (F₃) (Figure 1).

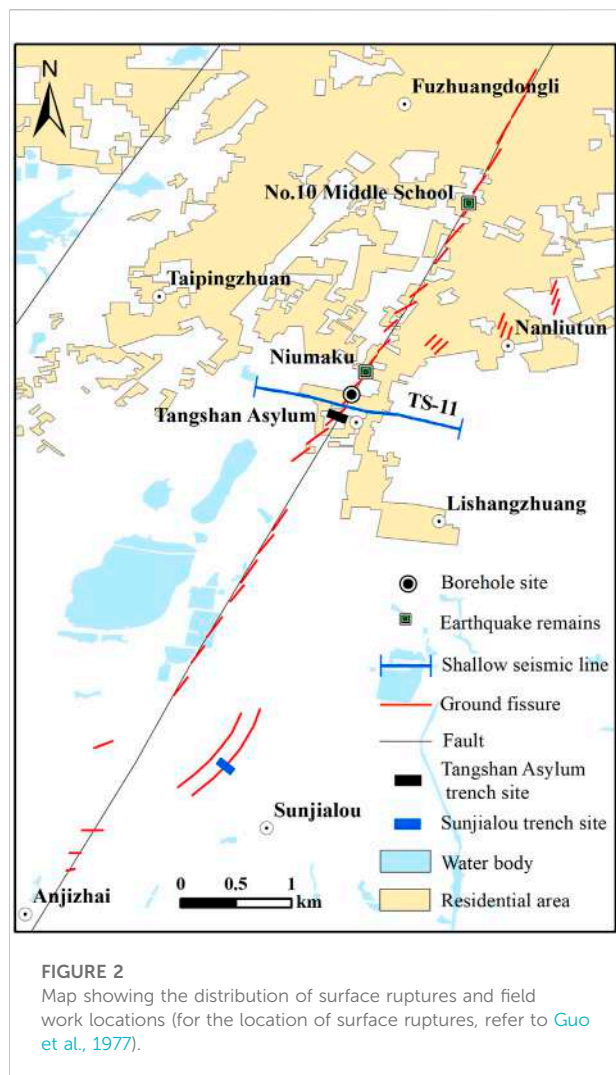
Characteristics of the shallow structure of the Tangshan earthquake seismogenic fault

Ground fissure zone

In 1976, after the Tangshan earthquake, a 10-km-long ground fissure zone (Figure 2) appeared along the Guye-Nanhu Fault (F₃), with an overall trend of N30°E and a width of ~30 m, exhibiting right-lateral strike-slip (Guo et al., 1977; Wang et al., 1981). Earthquake traces such as tensile cracks and

extruded mole tracks have been found on the ground surface (Guo et al., 1977). The surface projection of the epicenter of the M_s7.8 earthquake is located above this surface rupture zone. However, the tectonic relationship between the surface rupture zone and the fault has long been debated, as have the shallow distribution characteristics of the seismogenic fault.

Previous research has conducted exploratory trench excavation across surface ruptures in the Niumaku area (Figures 3C,D) and found other ruptures resulting from an earlier earthquake (Wang et al., 1978). Moreover, exploratory trench excavation conducted at the site of the former Tangshan No. 10 Middle School (Figures 3A,B) revealed that the 1976 surface ruptures disappear at a depth of 2.3 m and do not directly penetrate the fault (Wang et al., 1978). Similarly, based on trenching near Niumaku, Wang and Li (1984) found no faults but three groups of surface ruptures at different depths. Thus, previous trenching results did not reveal the existence of seismogenic faults, especially because the depth of the trenches was too shallow to reveal fault traces at depth. The composite drilling profile at the site of the former Tangshan No. 10 Middle School revealed a near-vertical fault beneath the surface cracks, with a dip angle >83°, and a Middle Pleistocene perpendicular fault with an offset of 40 m (Liu et al., 2013).



In recent years, trenches have been excavated beneath surface cracks in the Sunjialou area (blue rectangular area in Figure 2) (Guo et al., 2011, Guo et al., 2017; Guo and Zhao, 2019). Here, the cracks are unique, with a subsidence zone observed on the southeastern side of the right-lateral strike-slip, en-echelon surface cracks, and ~700 m long and 100 m wide (Figure 2). Guo et al. (2011) analyzed the drilling profile and trench in Sunjialou, Tangshan, and proposed that the seismogenic fault of the Tangshan earthquake was a normal fault. Extending the Sunjialou rupture to the SW connects it with one part of the Fuzhuang-Xihe subsidence zone, which also belongs to the surface rupture zone of the Tangshan earthquake, resulting in a total length >47 km (Guo et al., 2011, Guo et al., 2017; Guo and Zhao, 2019).

Three shallow artificial seismic profiles completed by our research team conducted at the location of the exploratory trench in Sunjialou revealed that the fault is located under an en-echelon-shaped surface rupture zone on the west side, with

no fault present in the lower part of the trench. These results are covered in a separate article and are not discussed further here.

Shallow artificial seismic detection

A 1.6-km-long shallow seismic survey line was laid on Nanhu Avenue, 200 m south of the earthquake relic site of Niumaku (Ts-11 in Figure 2), with a wave detection distance of 3 m and a common depth point distance of 1.5 m. A near-vertical main fault (F_3) at stake No. 713 and common depth point 1,426 is interpreted as the upper breakpoint of the fault, with latitude and longitude coordinates of 39.591°N and 118.188°E.

The strongly reflected wave groups on the east side of the fault (T_1 and T_Q) are stable (Figure 4), with good continuity and level occurrence, and a decrease in depth near the fault. The T_Q wave group on the western side of the fault is also stable with good continuity. The T_1 wave group exhibits relatively poor continuity and decreases in depth at the side close to the main fault, indicating an anticline, which was likely caused by horizontal extrusion. As the fault is heavily influenced by horizontal tectonic extrusion, the occurrence of the two wave groups on both sides of the fault generally indicates an anticline, and the overall appearance of the two wave groups is almost vertical. The overall positions of T_1 , T_Q , and T_N , as well as the top surface of the bedrock on the east side of the fault, are lower than those on the west side. In addition, the wave groups of the fault are offset downward. Therefore, we infer the existence of multiple branch faults that diverge upward and converge downward to the main fault.

Comparison of composite drilling profiles

Based on the location of the upper breakpoint interpreted by the shallow seismic exploration results, six field sites were selected for drilling-based exploration on the southern side of Nanhu Avenue and west of Tangshan Asylum (Figure 2). Through a comparison of key strata and an analysis of sedimentary cycles, and considering data from thermoluminescence dating, six obvious fault displacements are identified, which are 0.9, 2.1, 4.4, 5.2, 6.1, and 7.7 m from the top to bottom (Figure 5). At this site, an assemblage of gray gravel-bearing medium sand is developed at a depth of 30 m below ground. The gravel content is approximately 50%, and the gravel diameter is 1–5 cm, with good psephicity. The gravel components are predominantly limestone and quartzite and form a typical key bed with an offset of 7.7 m (Figure 5).

Thermoluminescence dating and regional data suggest that the Quaternary deposits at the site mainly developed in the Middle Pleistocene Yangliuqing Formation (Qp^2Y), the Upper

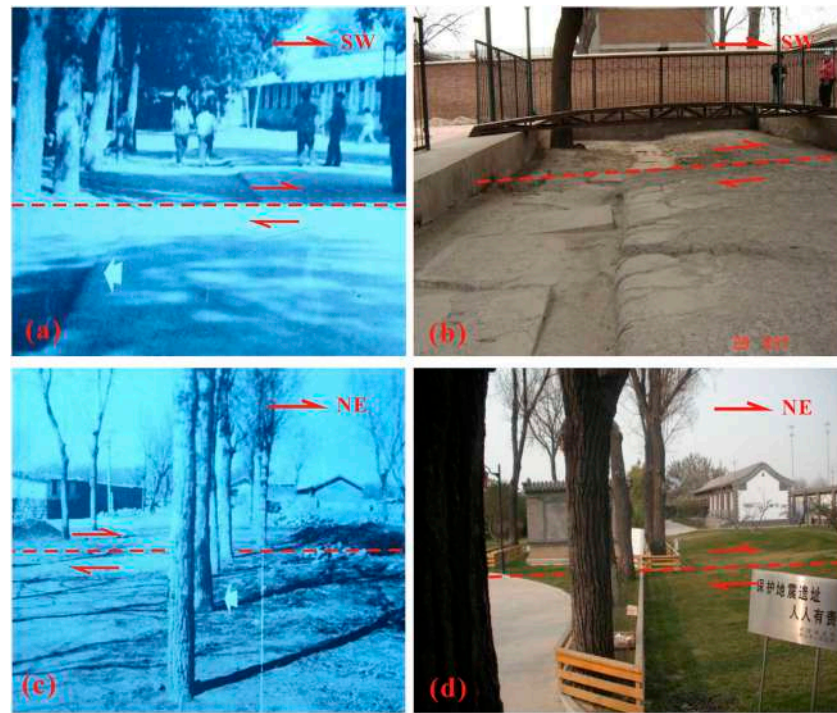


FIGURE 3

Photographs of surface ruptures. (A) Offset footpath with surface ruptures at the Tangshan No. 10 Middle School site, photographed after the earthquake (Fang et al., 1981); (B) earthquake relic site of Tangshan No. 10 Middle School; (C) offset trees in a yard with surface ruptures in Niumaku, photographed after the earthquake (Fang et al., 1981); (D) earthquake relic site at Niumaku.

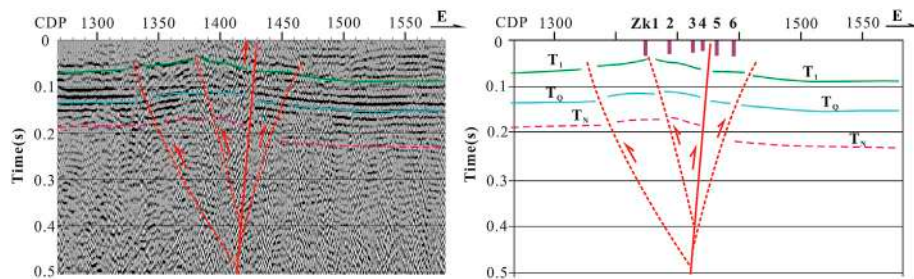


FIGURE 4

Ts-11 shallow seismic survey lines. T_1 : Quaternary strong reflection interface; T_0 : bottom boundary of the Quaternary; T_n : unconformity surface at the base of the Neogene formation; CDP: common depth point.

Pleistocene Ouzhuang Formation (Qp^3O), and Holocene deposits, which are thinner and substantially influenced by human activity. It is presumed that the Ouzhuang Formation (Qp^3O) is in conformable contact with the underlying strata. The sediments are mainly brownish-yellow clay, silty clay, yellowish-brown silty sand, medium-fine sand, and medium-coarse sand and can be divided into two cycles with clear boundaries and fining-upward sequences. The upper part of the cycle is

dominated by clay, the middle part by silty clay, and the lower part by fine and medium sand.

The sediments of the Yangliuqing Formation (Qp^2Y) are mainly brownish-yellow clay, silty clay, yellowish-brown to yellow silty sand, muddy silt, and grayish-yellow medium-coarse sand with a gravel layer in the lower part. Based on the sedimentary characteristics, the sediments can be divided into three cycles with clear boundaries and upward-fining sequences.

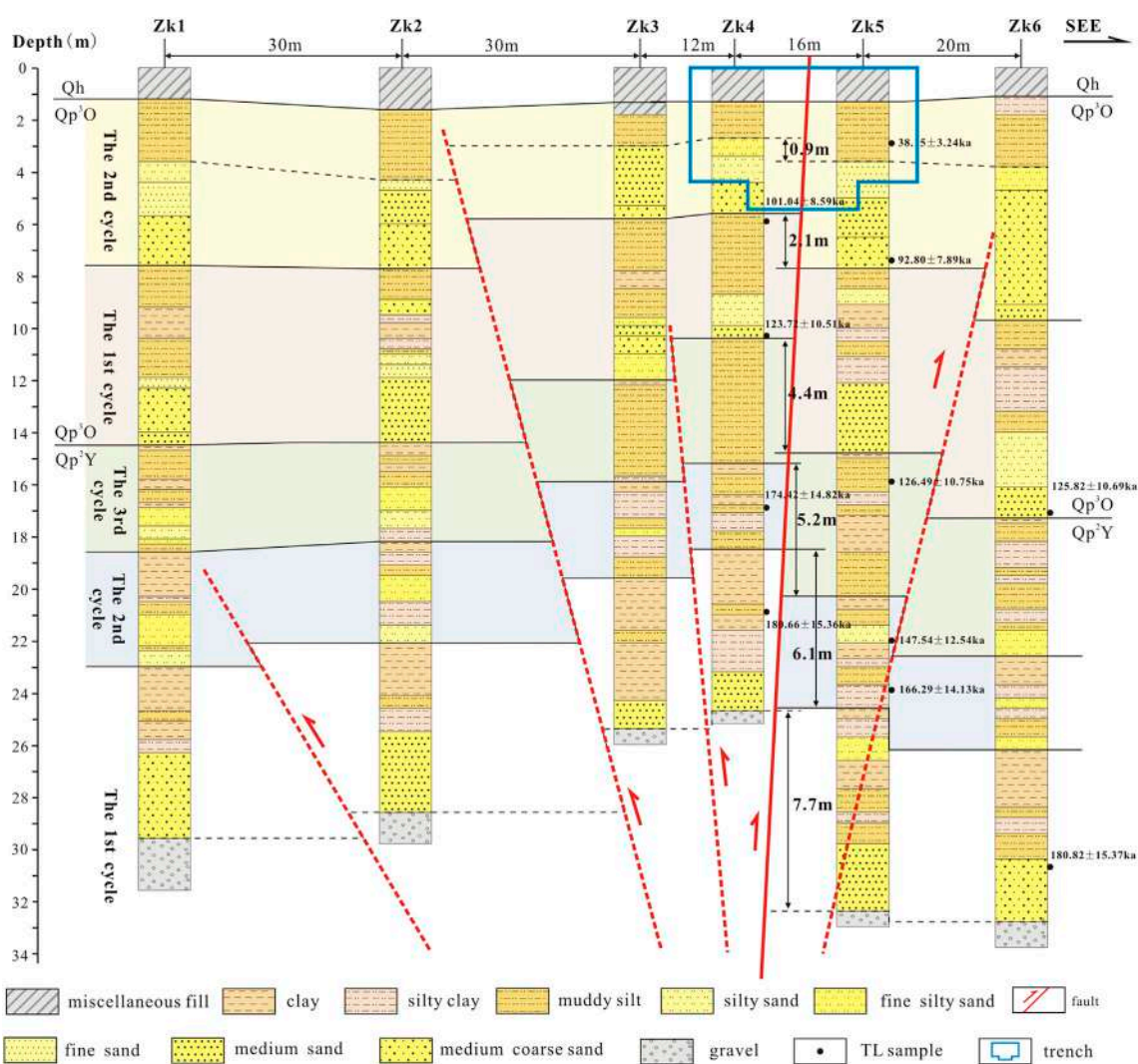


FIGURE 5
Composite drilling columnar comparison diagram at the Tangshan Asylum site.

The upper part is dominated by clay, the middle part is dominated by silty clay and muddy silt, and the lower part is dominated by fine and medium-coarse sand or gravel. The first cycle is a gravel–clay structure in its lower part, with a gravel layer at the bottom, an assemblage of dark gray fine sand in the middle, and a drop of 7.7 m at the top surface of the gravel layer. The second cycle is a structure ranging from fine silty sand to clay layers with highly variable thickness; the bottom of the east side is fine silty sand, whereas the bottom of the west side is muddy silt; the thickness of the overall breakpoint on the west side is thinner than that on the east side, and there is a 6.1 m offset between the west and east sides. The third cycle is a structure ranging from a silty sand layer to a clay layer. The average thickness of the sand layers is almost the same; however, the sand layer on the west side

of the breakpoint is slightly thinner than that on the east side, and there is an offset of 5.2 m between the west and east sides.

The strata geometry and fault distribution revealed by the composite drilling profile agree with those of the shallow artificial seismic profile. The strata layers are almost horizontal but bend near the main fault. The NW side rises near the fault, whereas the SE side subsides toward the fault, resulting in a relatively significant drop. By comparing the sedimentary cycles and key strata, the composite drilling profile reveals that the main fault is located between ZK4 and ZK5. There are several fault branches on both sides of the main fault, which diverge upward and converge downward to the main fault. The overall motion of the faults is characterized by a thrust component, consistent with the shallow artificial seismic detection results.



FIGURE 6
Panoramic view of the exploratory trench at the Tangshan Asylum site.

Fault structure revealed by exploratory trench excavation

Trench characteristics

Based on the extension direction of the surface rupture zone in the Niumaku earthquake relic area and the results of shallow seismic exploration and composite drilling, the precise location of the surface rupture was obtained. Thus, an exploratory trench was excavated 300 m south of the Niumaku crack zone on the west side of Tangshan Asylum. The trench was approximately 25 m long, 13 m wide at the top, and 10 m wide at the bottom. It is a two-stage ladder-like structure, with the first stage being 5 m deep and the second stage being 1.4 m deep, with a volume of approximately 1,500 m³ to the shallow water level (Figure 6).

The trench excavated at the Tangshan Asylum site is a large assemblage of marine sedimentary sand layers located 2–3 m below the surface, in which the distribution of faults is preserved (Figure 7). In general, although the large assemblage of loose sand layers is highly unfavorable for the construction, cleaning, and preservation of the trench, it creates unique conditions for studying the characteristics of the Tangshan Fault Zone structure, which is dominated by high-angle strike-slip movement. One of the main characteristics of the fault zone is loose marine sedimentary sand layers with small grain sizes and well-developed bedding. There are multiple types of bedding, such as horizontal bedding, massive bedding, and planar cross-bedding, which can completely record and clearly identify small fault movements. This is an advantage that is not often available in trenches elsewhere. Another major characteristic is that mud was drawn into the main faults with relatively large-scale slippage during the rupture process. As the loose marine sedimentary

sand layers around the faults are gradually exposed, these main faults become especially prominent (Figure 8B).

Multiple fault branches were revealed by the trench, which provide more structural information. The trench was carefully cleaned and cataloged, and each fault branch was numbered and classified (Figure 7) to best reflect the true traces of the faults. The upper part of the trench was divided into the following five stratigraphic units from top to bottom (Figure 7):

- ① Artificial fill, containing coal cinder, bricks, and other construction waste.
- ② Dark gray cultivated soil, with two groups of ground cracks of different timings, developed in the upper and lower layers, representing two separate earthquake events (Liu et al., 2013).
- ③ Yellowish-gray silty clay, whose sand content gradually increases from top to bottom and gradually changes to yellowish-brown. A large number of cracks are observed in the upper part. This layer contains some sole-shaped black blocks in the south wall, with some white calcareous nodules in the middle that absorb black substances such as carbon and manganese (Figure 9).
- ④ A cinerous mud layer, with a uniform composition, shows strong water absorption in the lower part. This layer is divided into upper and lower groups in the north wall of the trench. Despite having the same composition, the upper group is greenish, whereas the lower group is cyan. In the depression at the eastern end of the south wall, there are several assemblages of thin, muddy layers whose color gradually changes from green to cyan; these can be divided into four assemblages of thin layers.
- ⑤ A large assemblage of fine silty sand layers with various types of marine sediment structures, such as cross-bedding, parallel bedding, and horizontal bedding. The sandstone body is grayish-yellow, dry, and loose.

The lower part of the trench is a large assemblage of fine silty sand layers (layer ⑤), which can be subdivided into the following five stratigraphic units from top to bottom (Figure 7):

- ⑤-1: A grayish-yellow silty sand layer characterized by horizontal bedding.
- ⑤-2: A grayish-yellow sandy gravel layer characterized by planar cross-bedding; gravel is arranged in a specific direction, with a diameter of 0.3–1.0 cm.
- ⑤-3: A yellowish-brown fine silty sand layer, with no clear bedding, showing good consolidation. In the south wall of the trench, a layer of gravel is present at the bottom of this layer, arranged in a specific direction, with a diameter of 0.5–5.0 cm.
- ⑤-4: A grayish-yellow silty sand layer characterized by inclined bedding containing a gravel layer, with gravels arranged in a specific direction and a gravel diameter of 0.3–1.0 cm. Occasionally, large gravels with diameters of 2–5 cm were observed in the southeast wall. The south wall

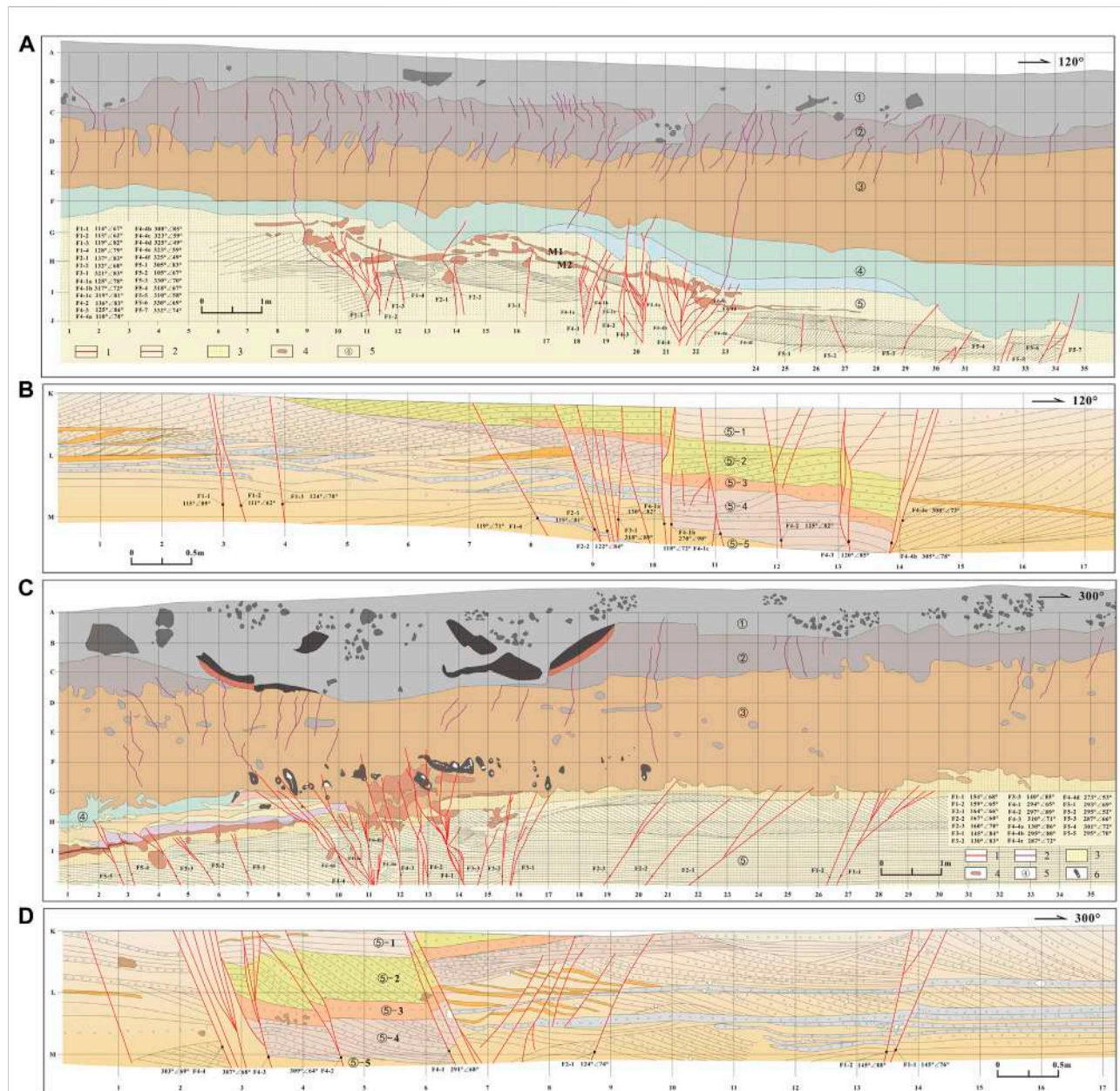


FIGURE 7

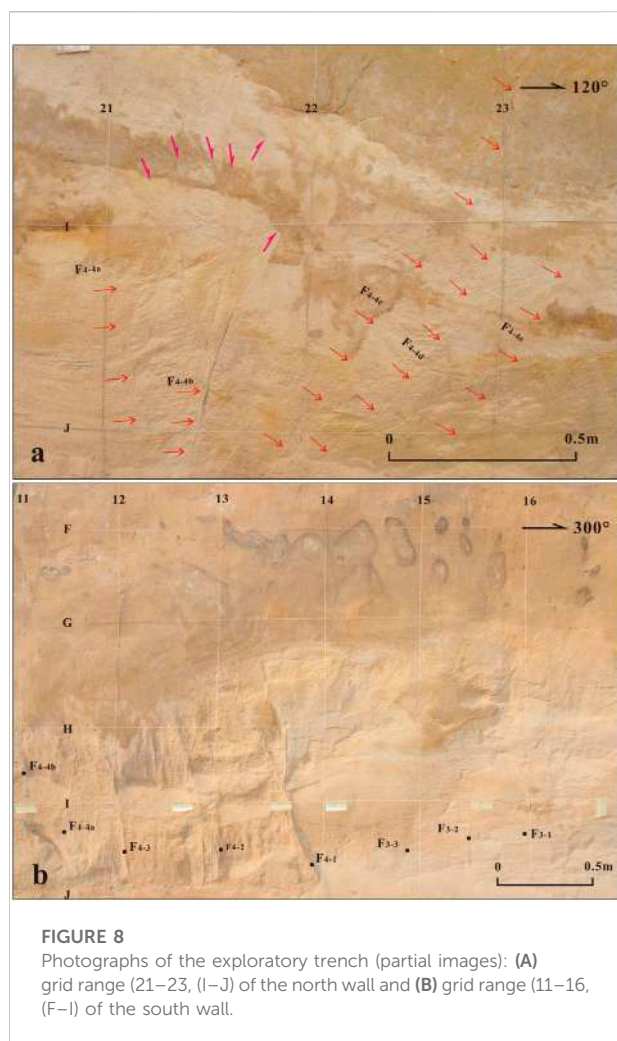
Sections of the exploratory trench at the Tangshan Asylum site. (A) Section of the upper part of the north wall; (B) section of the lower part of the north wall; (C) section of the upper part of the south wall; (D) section of the lower part of the south wall. 1: fault; 2: cracks; 3: large assemblage of fine silty sand layers; 4: brown boulder clay; 5: strata layer number; 6: sole-shaped black block with white calcareous nodules.

of the trench clearly shows that the fault offsets a piece of boulder clay with a diameter of ~10 cm (Figure 7D).

⑤-5: A grayish-yellow silty sand layer dominated by planar cross-bedding, with horizontal bedding and wedge-shaped bedding developed in between. In the southeast wall of the fault, there are several yellowish-brown thin, muddy silt layers that are offset by the fault zone, indicative of a normal fault.

Fault structural features

A large number of cracks are observed near the surface of the trench, divided into two layers, consistent with previous studies (Wang et al., 1978, Wang et al., 1984). The upper layer of the SE-trending cracks is covered by miscellaneous surface fill, presumably caused by the 1976 Tangshan earthquake.



Two assemblages (M_1 and M_2) of grayish-brown thin mud layers occur in the north wall (Figure 7A; Figure 10B) and are repeatedly offset by the faults. Conversely, only one assemblage of grayish-brown thin mud layers has developed at the east end of the south wall (M), which is offset by the fault into multiple small mud blocks within grids 10–14 (Figure 7C).

The fault groups revealed by the trench diverge upward and can be divided into five groups, F_1 to F_5 , each of which can be further subdivided. Among them, the F_4 fault group has the densest fault branches, concentrated in the center of the trench. The F_4 fault group in the north wall is concentrated in the upper grids 18–24 and the lower grids 10–15; in the south wall, the F_4 fault group is concentrated in the upper grids 9–15 and the lower grids 2–7 (Figure 7). The fault group F_4 corresponds to the main fault and can be subdivided into F_{4-1} to F_{4-6} in the upper part of the north wall, among which the fault F_{4-4} is relatively complex, diverging upward and converging downward into a fault; F_{4-4} can be further subdivided into F_{4-4a} , F_{4-4b} , F_{4-4c} , F_{4-4d} , and F_{4-4e} (Figure 8A). There is clear offset and traction deformation in the

strata near the fault. In the upper part of the south wall, the fault group F_4 is also subdivided into F_{4-1} , F_{4-2} , F_{4-3} , and F_{4-4} , among which the fault F_{4-4} is composed of multiple small faults that intersect and converge. Here, F_{4-4} is further subdivided into F_{4-4a} , F_{4-4b} , F_{4-4c} , and F_{4-4d} (Figure 8B).

The number of fault branches is equivalent in the upper and lower parts, as well as in the north and south walls of the trench. Figure 10 shows photographs of the numbered faults in the upper and lower parts of the trench. We selected grids 8–11 on the north wall in the lower part of the deep trench and excavated a horizontal trench on its upper platform to reveal the horizontal extension direction of the fault group and determine the corresponding relationship between the lower fault in the small trench and the upper fault. This analysis confirmed fault branch numbering.

The trench reveals a large number of fault branches and cracks, which exhibit good regularity and diverge upward. The strata at the same level are higher in the west and lower in the east, with nearly vertical faults in the center, exhibiting strong strike-slip features (Figures 9B,E). The central faults gradually undergo a transition westward into fault branches dominated by normal fault components (Figures 9A,F) and eastward into fault branches dominated by thrust components (Figures 9C,D). The amount of perpendicular offset between the fault branches at both ends is <5 cm. The closer to the central main fault, the clearer the fault plane and the larger the dislocation. The sedimentary sand layers on both sides of the central main fault are completely different, showing strong characteristics of the strike-slip movement. The fault distribution characteristics revealed by the trench at the Tangshan Asylum site are similar to a flower structure and are the most significant structural features in the shallow part of the Tangshan Fault Zone.

Discussion

Kinematic characteristics of the seismogenic fault

The movement characteristics of the seismogenic fault of the Tangshan earthquake have long been debated, with previous research suggesting a right-lateral strike-slip fault (Qiu, 1976; Zhang et al., 1980), a right-lateral strike-slip fault with a normal component (Hao and You, 2001), a high-angle thrust fault (You et al., 2002; Liu et al., 2013), or a normal fault (Qiu et al., 2005; Guo et al., 2011; Guo et al., 2017). All of these interpretations were obtained from a single detection method, which is highly limiting for an earthquake as complex as the Tangshan event. Therefore, it is necessary to integrate multiple detection methods for a more thorough analysis. This study combined analyses of surface ruptures, exploratory trenches, composite drilling columnar profiles, shallow seismic profiles, and focal

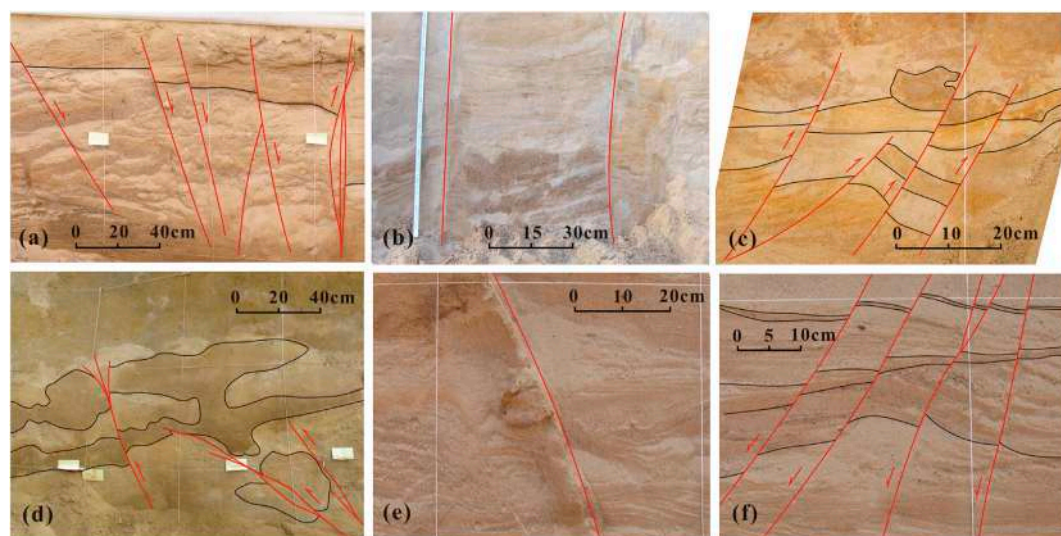


FIGURE 9

Fault patterns in the trench. (A) Normal faults in the western part of the north wall (specific location shown in grids 8–10 in Figure 7B); (B) strike-slip faults in the middle part of the north wall (specific location shown in grids 16–19 in Figure 7A); (C) thrust faults in the eastern part of the north wall (specific location shown in grids 34–35 in Figure 7A); (D) thrust faults in the eastern part of the south wall (specific location shown in grids 2–5 in Figure 7C); (E) strike-slip faults in the middle part of the south wall (specific location shown in grids 6–7 in Figure 7D); (F) thrust faults in the eastern part of the south wall (specific location shown in grids 7–8 in Figure 7C).

mechanism solutions to comprehensively evaluate the kinematic characteristics of the Tangshan earthquake seismogenic fault.

After the $M_s7.8$ Tangshan earthquake in 1976, a 10-km-long surface rupture zone (Figure 2) appeared in the meizoseismal area along the Tangshan-Guye Fault (F_3) (Guo et al., 1977). The two sides of the rupture exhibit horizontal deformation in a right-lateral manner. The rupture zone is composed of no more than 20 deformed cracks in an en-echelon structure in the reverse direction. The maximum horizontal right-lateral offset is 2.3 m. The NW side rises relative to the center, whereas the SE side subsides by 0.2–0.7 m.

Previous research on the surface rupture zone (Guo et al., 1977) coupled with the large-scale “ $M_s7.8$ Tangshan Earthquake Seismogenic Structure Map (1:1,000)” obtained from the Hebei Earthquake Agency reveals that the surface rupture zone conforms to the Riedel shear model. The strike of the surface rupture zone is $NE30^\circ$, with a relatively well-developed R shear that moves in a direction of $45\text{--}60^\circ$. Moreover, a small number of push-ups with a long axis direction of 150° also exist. Riedel shears are closely related to the earthquake rupture mechanism (Tchalenko, 1970; Tchalenko and Ambraseys, 1970; Aydin and Page, 1984). Laboratory experiments and field-based seismic surface rupture investigations have shown that Riedel shears are the main type of initial rupture for seismic faults, especially regarding the surface rupture of large strike-slip earthquakes (Tchalenko, 1970; Tchalenko and Ambraseys, 1970; Deng and Zhang, 1984; Bergerat et al., 2003; Lin et al., 2004; Lin et al., 2011; Rao et al., 2011; Li et al., 2015; Liu et al., 2021). This also explains

why the series of cracks observed in the trench do not break through the surface but are instead typical surface ruptures characterizing strike-slip earthquakes.

After the trench excavation at the Tangshan Asylum site, a parallelepiped was identified in the marine fine silty sand layer in the north wall (Figure 11). This parallelepiped is brownish-yellow, cemented by silt and mud, hard, and regular in shape, located in the gray-white marine fine silty sand layer 4 m below the ground surface, where the sand body has clear bedding and a uniform texture and is not affected by human activities (Figures 11A,B). This feature is located in the fault zone and is presumably caused by long-term creep along the fault. The boulder clay has a standard positive torsion geometrical shape, and the response fault exhibits right-lateral strike-slip motion (Figures 11C–E).

At the earthquake relic site of the Tangshan No. 10 Middle School, seven holes were drilled close together across the surface rupture over 22 m, with the deepest hole being 106 m. These revealed that the main fault has a dip angle $>83^\circ$ (Liu et al., 2013). The shallow seismic profile (TS-11) reveals multiple fault branches and the high dip angle of the main fault. The trench reveals that the main fault group (F_4) has a high dip angle, most fault branches have dip angles $>80^\circ$ (Figure 7), and the faults diverge upward and converge downward toward the main fault, which is almost vertical, with clear strike-slip and vertical offsets (Figure 9).

Based on initial motion P-wave data for the 1976 $M_s7.8$ Tangshan earthquake collected from seismic stations around the world, the calculated focal mechanism

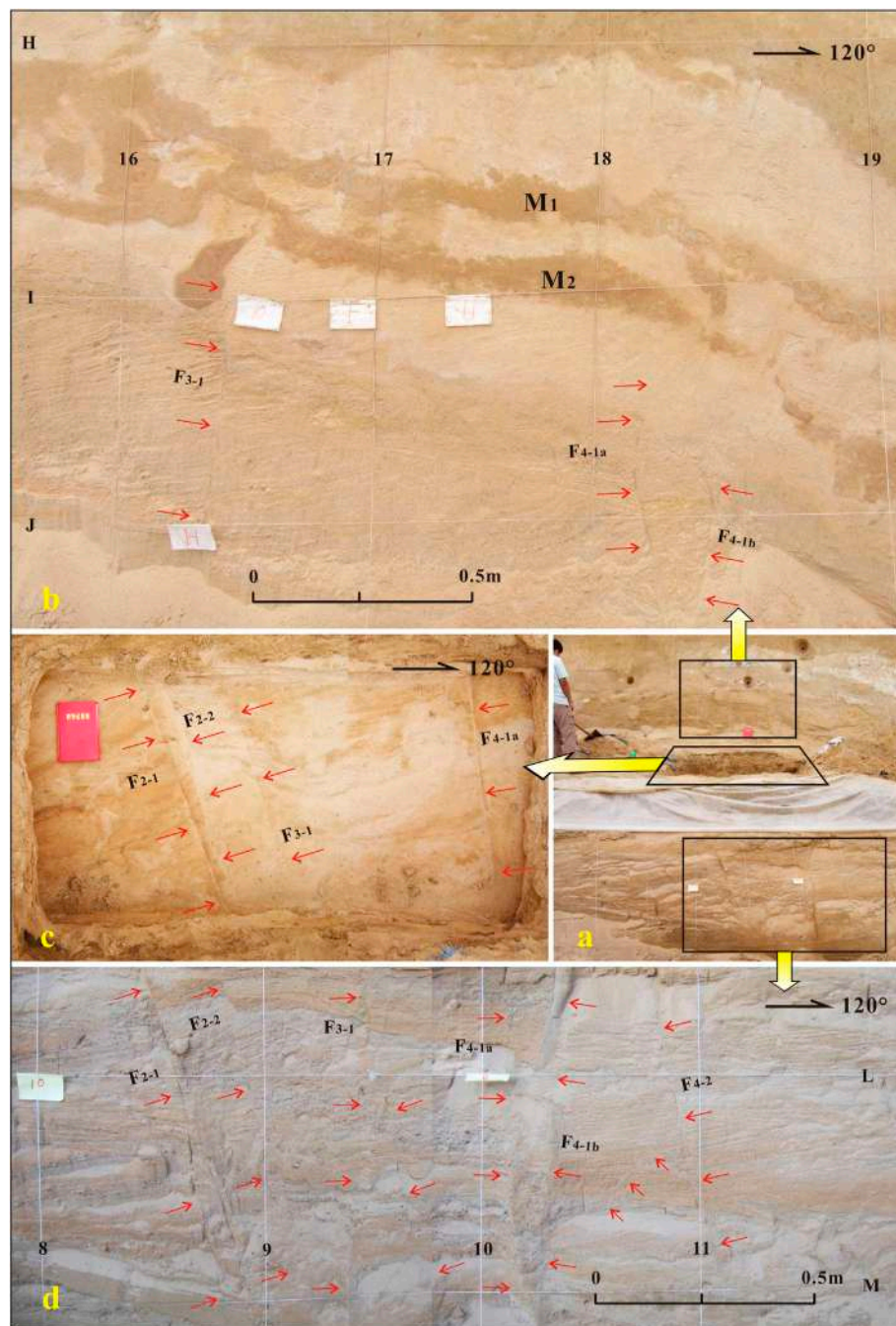


FIGURE 10

Fault relationships revealed in the upper part of the north wall of the exploratory trench and deep trench. Photographs showing (A) the north wall of the exploratory trench and deep trench, (B) part of the north wall, (C) the exploratory trench in the horizontal direction, and (D) part of the north wall of the deep trench.

solutions are similar. The Tangshan earthquake manifested as a high-dip, right-lateral, strike-slip earthquake (Qiu, 1976; Butler et al., 1979; Zhang et al., 1980; Nabelek et al., 1987; Shedlock et al., 1987; Huang and Yeh, 1997). Bulter et al. (1979) also analyzed the far-field body waves and surface waves of the

earthquake, and through fitting inversion of synthetic seismic waveforms and observational records, they suggested that the main shock of the Tangshan earthquake was a complex faulting sequence composed of several events, including strike-slip and thrust events.

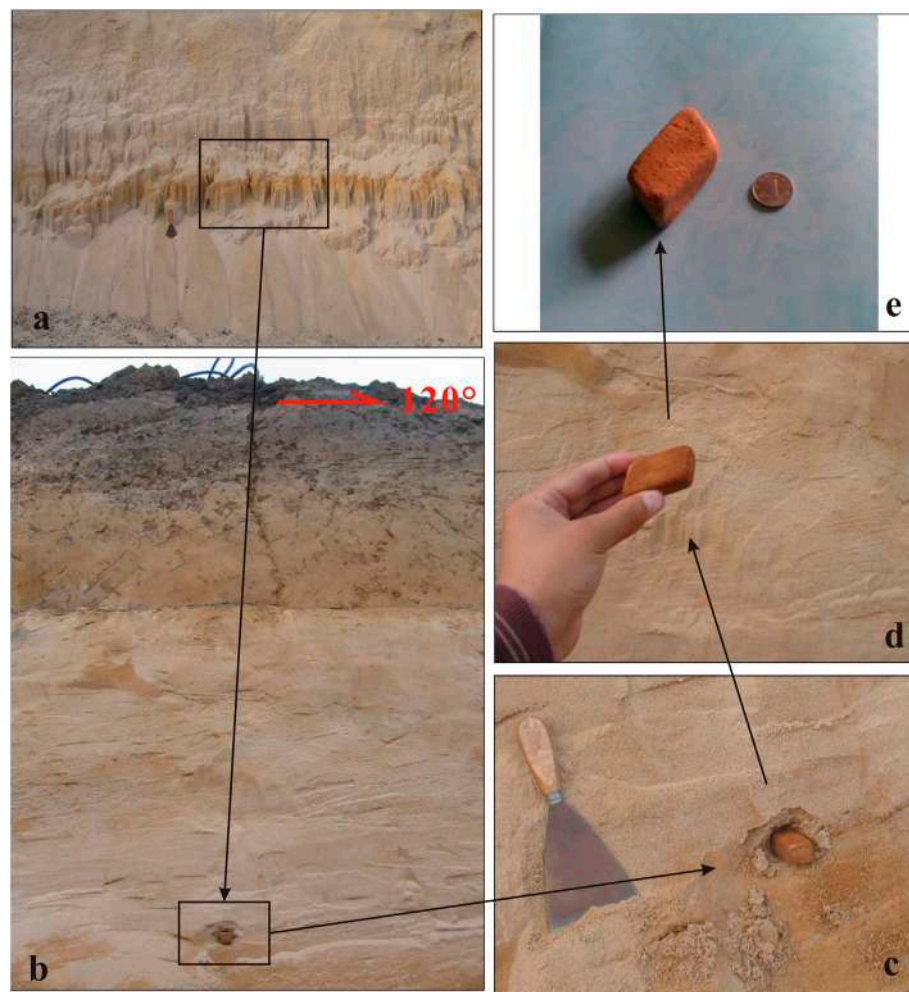


FIGURE 11

Photographs of parallelepiped boulder clay samples in the north wall of the exploratory trench. (A) Sand layer before cleaning; (B) sand layer after cleaning; (C) sample *in situ*; (D) specific sample shape; (E) surface of the preserved boulder clay with longitudinal cracks.

Based on these results, we suggest that the Guye-Nanhu Fault (F_3), which represents the shallow response to the seismogenic structure of the Tangshan earthquake, is characterized by right-lateral strike-slip movement with a thrust component and an almost vertical fault plane.

Deep and shallow structural model of the seismogenic fault

The en-echelon-shaped surface rupture zone is characterized by right-lateral strike-slip motion. The shallow seismic profile (TS-11) and the composite drilling columnar profile obtained at the Tangshan Asylum site reveal that there is an almost vertical main fault in the lower part of the surface rupture zone, where there is a clear perpendicular offset. Furthermore, multiple fault

branches diverge upward and converge downward to the main fault, conforming to a typical positive flower structural model.

The trench excavated at the Tangshan Asylum site reveals the specific shape of the faults, which are characterized by a combination of fault branches that tend to diverge outward like a flower. The normal fault branches also become progressively steeper with depth and converge downward to the vertical main strike-slip fault with a deep-cut basement. A vertical dislocation occurs between the two sides of the fault zone, where the west side rises but the east side subsides, relative to the fault zone. The central faults are dominated by strike-slip motion. The fault branches to the west are predominantly normal faults, whereas those to the east are reverse faults, and those close to the central faults are strike-slip faults. The farther the faults are to the west and east, the gentler the dip angle is and the clearer the normal and reverse faulting components are. The profile is

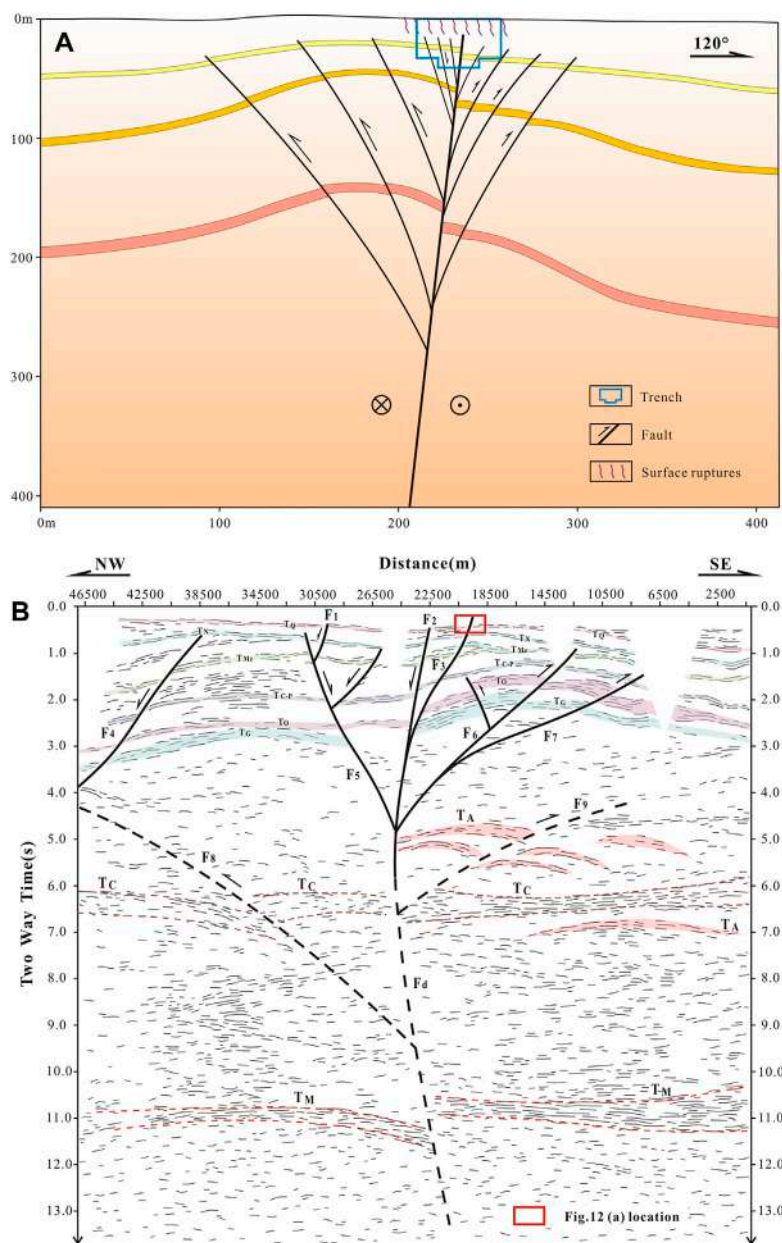


FIGURE 12

Schematic representation of the seismogenic fault of the Tangshan earthquake: (A) shallow part of the seismogenic fault and (B) deep seismic reflection profile across the Tangshan Fault Zone in the south (adapted from Liu et al., 2011).

characterized by a combination of fault branches diverging outward like a flower. Again, the fault branches become progressively steeper with depth and converge downward to the vertical main strike-slip fault. The trench excavated at the Tangshan asylum site reveals a unique structural model, which is confirmed by the composite drilling columnar profile and the shallow seismic exploration results. Figure 12A shows a model of the shallow structure of the seismogenic fault of the Tangshan earthquake, which manifests as a positive flower structure. A

high-angle thrust fault is located to the west of the main fault, and a series of tensile normal faults appear in front of the high-angle thrust faults, owing to surface tension and other factors, for example, the normal faults in the western section of the exploratory trench.

A 49.4-km-long deep seismic reflection profile with a line of measurement perpendicular to the Tangshan Fault Zone was completed by our research team (Figure 1). T_Q , the reflection wave on the profile, is interpreted as bottom boundary reflections

of the Quaternary overburden, T_N as the bottom boundary of the Neogene stratum, T_{Mz} as the bottom boundary of the Mesozoic stratum, and T_O and T_{C-P} as reflections of the Paleozoic Ordovician, Carboniferous, and Permian strata, respectively; the reflection wave group T_C is the reflection from the upper and lower crustal partition interface, while T_M is the reflection from the crust–mantle transition zone; T_A is interpreted as seismic multi-arch structures.

The deep crustal fault, F_d , revealed by the deep seismic reflection profile is located below line stake No. 22500 (Figure 12B). There are obvious transverse discontinuities in the upper and lower crustal partition interface reflection T_C and the crust–mantle transition zone reflection T_M near the deep fault, and the interface occurrence and wave characteristics of the lower crustal reflection on both sides are also clearly different. Thus, F_d is shown to be a deep fault between the upper and lower crustal partition and the Mohorovicic discontinuity. The existence of this deep fault provides conditions for the upwelling of thermal materials in the upper mantle and intracrustal tectonic deformation.

The T_C interface on the west side of deep fault F_d presents a subducting underthrust feature; therefore, we suppose that there is an F_8 fault. There is a clear overlap of arcuate tectonic features at the bottom of the upper crust, suggesting the existence of an F_9 fault. These two faults merge downward over the deep fault F_d .

Faults F_1 – F_3 , as revealed by the deep seismic reflection profile, correspond to three branching faults of the Tangshan Fault Zone. F_5 – F_7 extend downward at different dips in the profile and converge at a depth of approximately 10 km to form an almost upright fault inserted into the upper crustal floor, near the upper and lower crustal partition interface, which may be associated with the deep fault of the Mohorovicic discontinuity. The shallow seismic exploration, drilling, and excavated trench area is located near the epicenter, while the 2D deep seismic reflection line is 13.1 km from the epicenter. Therefore, there may be inconsistencies in the deep and shallow survey results in terms of structural details, but the general form should be consistent.

A deep crustal fault provides the deep power source for the Tangshan earthquake. The deep tectonic form revealed by the deep seismic reflection profiles further supports the superficial tectonic model. This complex system of deep and shallow faults constitutes the tectonic background of the Tangshan earthquake. In summary, there is a main upright strike-slip fault, deep in the basement beneath the Tangshan Fault Zone, which faults the Mohorovicic surface. The branching faults are gentle on the top and steep on the bottom and converge downward to the main fault, displaying different occurrence characteristics owing to the influence of the stress field in different geological periods. The Tangshan Fault Zone is characterized by right strike-slip movement under the near-NEE stable stress field (Wan, 2007) during the Neotectonic period (since 0.78 Ma). A large number

of detailed survey works reveal that the faults are characterized by a flower-like combination of branching reverse faults that spread outward at the shallow section, with the branching reverse faults gentle on the top and steep on the bottom, protruding inward and converging downward to the main strike-slip faults. These features form a positive flower-like tectonic structure overall (Figure 12A).

Newly generated fault zone in North China

The North China Plain fault basin is bounded by the Yanshan uplift, the Taihang uplift, the Luxi uplift, and the Tanlu Fault Zone (the Ludong and Liaodong uplifts). The NNE-oriented Tanlu Fault Zone is a large-scale and structurally complex fault zone in eastern China, cutting through the lithosphere for over 3,500 km (Wan, 1996) and dominating the eastern boundary of the North China Plain faulted basin together with the Ludong and Liaodong uplifts (Figure 13).

Between 1966 and 1976, the Xingtai $M_s7.2$, Hejian $M_s6.3$, and Tangshan $M_s7.8$ earthquakes occurred in the North China Plain region in sequence from north to south. Previous work has indicated the existence of a newly generated fault zone that began to develop in the Neogene based on the linear alignment of seismotectonics (Xu et al., 1996). In this study, we focused on exploring the mechanism of the formation of the Tangshan–Hejian–Xingtai Neotectonic zone from the perspective of the dynamics of the Tangshan earthquake.

Angular unconformities between structures of the Early and Middle Pleistocene are widespread in mainland China, and the characteristics of tectonic activity since the Middle Pleistocene are generally consistent. The horizontal principal stress during the Neotectonic period (since 0.78 Ma) is in the near-NEE direction in northern China (Wan, 2007). In this tectonic setting, most of the pre-existing high-dipping faults in the upper crust of mainland China happen to intersect the maximum principal compressive stress direction almost perpendicularly or at large angles; therefore, the fault surfaces are extruded relatively tightly. Their activity often shows a certain left or right strike-slip characteristic because of the different angles between the direction of maximum principal stress and the fault plane. The northern section of the Tanlu Fault Zone is extruded in the NNE direction, while the southern section is extruded in the near east–west direction, presenting a reverse fault with a slight right strike-slip (Wan 1996).

The near-NEE-oriented stress field direction in North China and the lower strike-slip component of the Tanlu Fault Zone result in stress accumulation within the North China Plain faulted basin. The middle of this basin is approximately rectangular (Figure 13, blue dashed box), and the Tangshan–Hejian–Xingtai Fault Zone is located exactly at the diagonal position of the rectangle, which means that faults are easily developed.

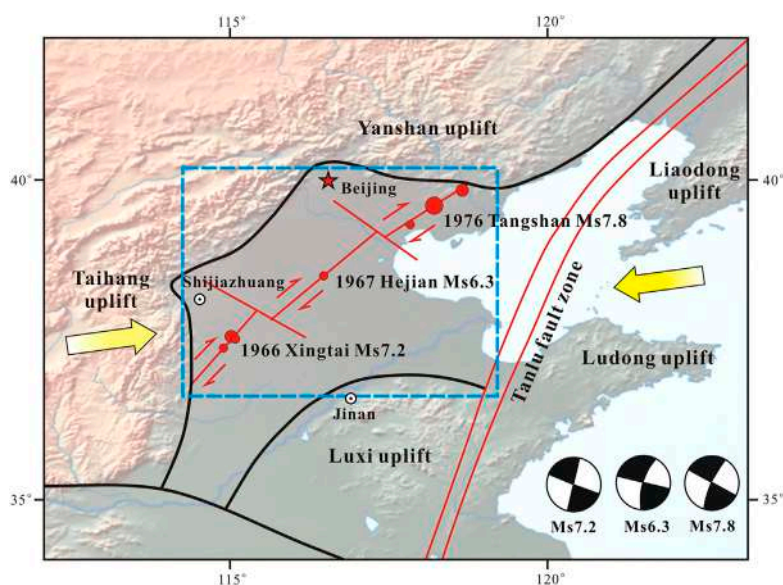


FIGURE 13

Tectonic map of the North China Plain faulted basin (focal mechanism solutions according to Xu et al., 1996; stress field according to Wan, 2007).

Based on the data from small seismic positioning, petroleum seismic profiles, and deep seismic reflection profiles, the Tangshan–Hejian–Xingtai Fault Zone is speculated to be a newly generated seismotectonic zone with a length of 500 km, which is oblique and shows a gradual change in fault orientation from near-NNE to near-NE from south to north, similar to the change in focal mechanism solutions (Figure 13).

Deep physical survey data reveal the existence of a buried fault from the focal location to the Mohorovicic discontinuity below the epicenter of the 1976 Xingtai earthquake, which is a high-dip supracrustal fault (Wang et al., 1993) similar to the deep fault of the Tangshan earthquake. Both are large ultracrustal deep faults.

The Tangshan Fault Zone is the largest in tectonic scale, and several paleoseismic events, characterized by alternating periods of relatively intensive earthquakes and quiet periods (Liu et al., 2013), have occurred along this section since the Middle Pleistocene. Therefore, the Tangshan Fault Zone can be presumed as the first active section. No tectonically significant surface fault zones have been found in the Xingtai earthquake, and there are only a few paleoseismic studies. Xu et al. (2000) suggested that the Xingtai earthquake fault was a new fault in the process of upward extension. The available research on the Hejian earthquake is least among these earthquakes. Throughout history, no $\geq M_s7$ earthquakes have occurred in the Hejian area, although it is speculated to be a seismic hazard section in the future. The inconsistency of the seismic activity characteristics of different sections also indicates that the fault zone is gradually penetrating and has a clear Neogenic nature.

The frequent occurrence of aftershocks during the 55 years after the Xingtai earthquake and the 45 years after the Tangshan earthquake further suggests that this is a gradually penetrating Neogenic fault zone. Therefore, this fault zone is assumed to have experienced overall faulting from south to north during the 10-year period from 1966 to 1976. The Tangshan and Xingtai sections have already experienced earthquakes of magnitude 7 or greater; therefore, the risk of earthquakes of magnitude 7 or greater occurring in the Hejian section is high. Yin et al. (2015) comprehensively examined historical earthquake records and paleoearthquake research achievements in the North China basin and suggested that there is an earthquake gap between Hejian and Tianjin along the “Tangshan–Hejian–Cixian” right strike-slip fault zone, which is consistent with our conclusions.

Conclusion

- (1) Based on analyses of surface ruptures, exploratory trenches, composite drilling columnar profiles, shallow seismic profiles, and focal mechanism solutions, our study indicates that the NNE-trending Guye-Nanhu Fault represents the shallow response to the seismogenic structure of the Tangshan earthquake. This fault has an almost vertical dip and NW-trending main fault direction with characteristics of thrust movement.
- (2) The trench excavated at the Tangshan Asylum site reveals a large number of fault branches and cracks under the surface

rupture, which are highly regular and diverge upward. The strata at the same layer are higher in the west and lower in the east, with nearly vertical faults in the center, exhibiting strong strike-slip features. The central faults gradually transition westward into fault branches dominated by normal faulting components, transition eastward into fault branches dominated by thrust components, and converge downward to the near-vertical main strike-slip fault. There is a vertical offset between the two sides of the fault zone, with the west side rising and the east side subsiding. The central faults are dominated by strike-slip motion. The fault branches on the west are predominantly normal faults, those on the east are reverse faults, and those close to the central faults are strike-slip.

- (3) Shallow seismic exploration profile, composite drilling columnar profile, and exploratory trench data reveal that the shallow structure of the seismogenic fault of the 1976 Tangshan earthquake conforms to a positive flower structure model, where the profile is characterized by a combination of reverse fault branches diverging outward like a flower. The reverse fault branches become progressively steeper with depth, bulge inward, and converge downward to the vertical main strike-slip fault, typically forming a positive horst-shaped structural pattern. A high-angle thrust fault is located to the west of the main strike-slip fault, and a series of tensile normal faults appear in front of the high-angle thrust fault, owing to local tension and other factors; this corresponds to the small normal fault group in the western section of the trench. The tectonic form revealed by the deep seismic reflection profiles further supports the superficial tectonic model.
- (4) Based on the linear alignment of regional seismic tectonics, the newly generated Tangshan–Hejian–Xingtai Fault Zone is speculated to exist, with a length of 500 km and an oblique alignment. The near-NEE orientation of the stress field in North China and the lower strike-slip movement component of the Tanlu Fault Zone have led to easy faulting in the Tangshan–Hejian–Xingtai Fault Zone, which lies diagonally in the middle rectangular area of the North China Plain faulted basin. The inconsistency of the seismic activity in different sections also indicates that the fault is gradually penetrating and has a distinct Neogenic nature. Earthquakes of magnitude 7 or greater have already occurred in the Tangshan and Xingtai sections; therefore, there is a relatively high risk of earthquakes of magnitude 7 or greater occurring in the Hejian section.

Data availability statement

The raw data supporting the conclusions of this article will be made available by the authors, without undue reservation.

Author contributions

KL and YL contributed to the conception and design of the study. KL completed the trenching and drilling work and wrote the first draft of the manuscript. BL completed seismic deep reflection interpretation. WW completed shallow seismic exploration. YN completed the drawing. YL, YN, and WW wrote sections of the manuscript. All authors contributed to manuscript revision and read and approved the submitted version.

Funding

This research was financially supported by the National Natural Science Foundation of China (Grant No. 41631073) and the Opening Foundation of Hebei Key Laboratory of Earthquake Dynamics (Grant No. FZ212101).

Acknowledgments

The authors thank Guosheng Qu and Jianqiang Chen for their help with field work.

Conflict of interest

WW was employed by Xi'an Research Institute Co., Ltd., China Coal Technology and Engineering Group Corp.

The remaining authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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SPECIALTY SECTION

This article was submitted to
Geohazards and Georisks,
a section of the journal
Frontiers in Earth Science

RECEIVED 12 June 2022

ACCEPTED 05 July 2022

PUBLISHED 15 August 2022

CITATION

Zhou Z, Pei J, Li J, Cai Y and Hou L
(2022), High-resolution
magnetostratigraphic records of the
pliocene sedimentary successions in
Yengisar section, NW China, and its
tectonic implications.
Front. Earth Sci. 10:967346.
doi: 10.3389/feart.2022.967346

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High-resolution magnetostratigraphic records of the pliocene sedimentary successions in Yengisar section, NW China, and its tectonic implications

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The Neogene strata upward-coarsening sandstone and conglomerate sequences at the periphery of the Northeastern Pamir record the intense uplift of the paleosurface of the building mountains. To further improve our knowledge of source-sink processes, a detailed magnetostratigraphic investigation was carried out along the Yengisar section, which is located at the southwestern margin of the Tarim Basin. The new high-resolution magnetostratigraphic data revealed that the Artux Formation was deposited from 4.9 Ma to 1.9 Ma with three sedimentation rates changes. The variations in sedimentation rate may be due to the pulsating exhumation of the Western Kunlun Mountain, caused by the northward motion of the Pamir salient. By integrating the evidences from the seismic reflection profile and other magnetostratigraphic investigations in this region, the basal age of the Xiyu Formation and the early stages of growth strata deposition were estimated at ~1.9 and 1.45 Ma, respectively. Based on the data, we propose that the progradation of the Xiyu Formation and the migration of the deformation front are the two independent responses of the sink region to the uplift events of the source region.

KEYWORDS

magnetostratigraphy, Pliocene, Pamir salient, sedimentation rate, Southwestern Tarim

Introduction

The Indo-Eurasia continental collision resulted in the deformation of the Western syntax of the Tibetan Plateau (Pamir Plateau), which is considered as one of the most remarkable regions in the globe (Burtman and Molnar, 1993; Zhao et al., 2008a; Cowgill, 2010; Chen et al., 2014) (Figure 1). In fact, the collision is responsible for the formation of the Paratethys Sea, the northern branch of the Neo-Tethys Ocean, which retreated to the Mediterranean during the Cenozoic (Bosboom et al., 2011; Sun and Jiang, 2013; Bosboom et al., 2014; Sun et al., 2016; Li et al., 2017). Furthermore, the Cenozoic tectonic processes have shaped the present surface topography, which is considered as a gigantic orogeny-basin coupling system (Kunlun Mountain, Tian Shan, and Tarim Basin) (Cao et al., 2018; Li et al., 2019). The Tarim Basin preserves more than 10 km thick of Cenozoic strata, which provide an excellent chance to testify the tectonic evolution of the plateau uplift based on high-resolution age constraints. Thus, dating the depositional age of the Cenozoic sedimentary successions is crucial to understanding the coupling mechanism between the mountain building, sediment deposition, and aridification history.

In the Southwestern Tarim Basin, substantial sections containing the Cenozoic strata at the piedmont of the Tian Shan and the Kunlun Mountain are incised by the drainage system shed from the growing plateaus or highlands (Figures 1, 2A). In recent years, considerable attention has been devoted to these well-exposed outcrops to decipher the paleogeographic evolution (e.g. Zheng et al., 2000; Chen et al., 2002; Chen et al., 2007a; Chen et al., 2007b; Huang et al., 2006; Charreau et al., 2006; Heermance et al., 2007; Pei et al., 2008; Pei et al., 2011a; Pei et al., 2011b; Bosboom et al., 2011; Bosboom et al.,

2014; Sun and Jiang, 2013; Tang et al., 2015; Yang et al., 2015; Chen et al., 2015; Qiao et al., 2016; Qiao et al., 2017; Zhou et al., 2016; Liu et al., 2017). Magnetostratigraphy is the most widely applied and most powerful method, due to a lack of diagnostic fossil and volcanic layers in continental deposits. Previous studies have focused mainly on the following three important scientific problems: 1) When did the Para-Tethys Sea finally retreat from the Tarim Basin (Guo et al., 2002; Bosboom et al., 2011; Bosboom et al., 2014; Sun and Jiang, 2013; Sun and Jiang, 2016; Yang et al., 2015; Zhang et al., 2001); 2) How did the accumulation of the Tarim Basin response to the global climate changes (Sun and Liu, 2006; Charreau et al., 2009; Pei et al., 2011a; Zheng et al., 2015a); 3) the uplift processes of the Western tectonic syntax and the Tian Shan (Zheng et al., 2000; Chen et al., 2002; Huang et al., 2006; Pei et al., 2011b; Chen et al., 2015; Qiao et al., 2016). The stratigraphic attribute and the origin of the Late Cenozoic deposits consisting of the Artux Formation and the Xiyu Formation (Xiyu Conglomerate) had been widely studied in the past decades. Most researchers have considered that the upward-coarsening sediments recorded an abrupt change in paleoslope, as a result of immense surface uplift in the mountains (Huang 1957; Li et al., 1979; Zheng et al., 2000), while others believe it was triggered by dramatic climate changes at ~2–4 Ma ago (Molnar et al., 1993; Zhang et al., 2001; Wang et al., 2003; Charreau et al., 2009).

To better decipher the spatial characteristics and deformation behavior of the uplift-related molasse deposits, here we conduct a new magnetostratigraphic study in detail in the Yengisar section located at the southwestern margin of the Tarim Basin. Moreover, we discussed the sedimentation rate changes of the Pliocene successions and their potential causes. Combining with other magnetostratigraphic data and

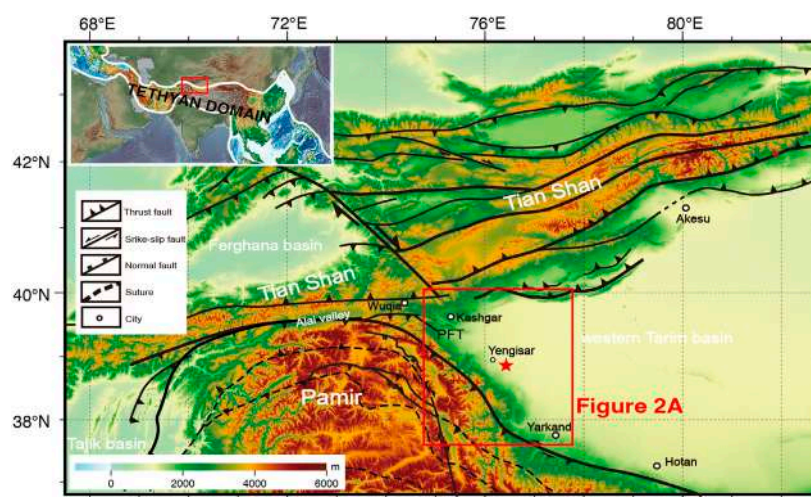


FIGURE 1

Simplified tectonic map of the western Tarim Basin and its surrounding orogens (after Xu et al., 2011; Li et al., 2017).

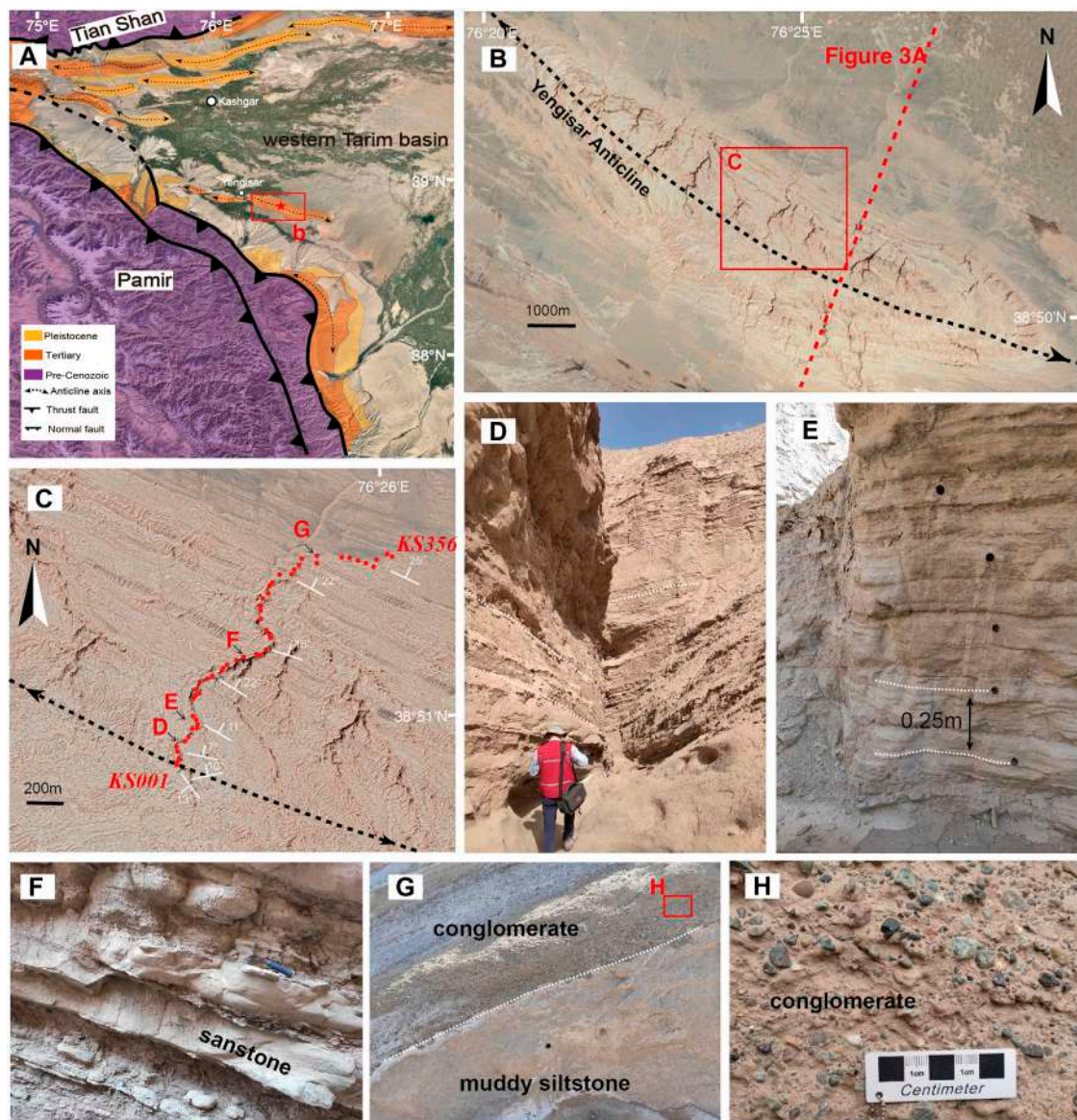


FIGURE 2

(A) Geological sketch map of the study area with highlighting on the Cenozoic strata distribution (after Pan et al., 2009). The red aster denotes the location of the studied section. (B) and (C) show the detailed sampling sites along the canyon across the northeast limb of the Yengisar anticline. (D), (E), (F), (G), and (H) show the stratigraphy, lithology and sampling snapshot in the fieldwork, the spots for photography are marked in (C).

geophysical evidence, we further summarized the attribution of the variational basal age of the Xiyu Formation.

Regional geological setting and stratigraphy

The Western Tarim Basin is a depression surrounded by the Tian Shan to the north and the West Kunlun Mountain to

the southwest. The long and the short distance effects of the Indo-Eurasia collision during the Cenozoic caused a strong uplift, shortening, and strike-slip deformation in the surrounding areas of the Western Tarim Basin (Figures 1, 2A) (Xu et al., 2011; Chen et al., 2014). As part of the large-scale Meso-Cenozoic sedimentary basin in the West of China, the Western Tarim Basin preserves over 10 km thick strata, which becomes thinner in the east (Chen et al., 2001; Li et al., 2019).

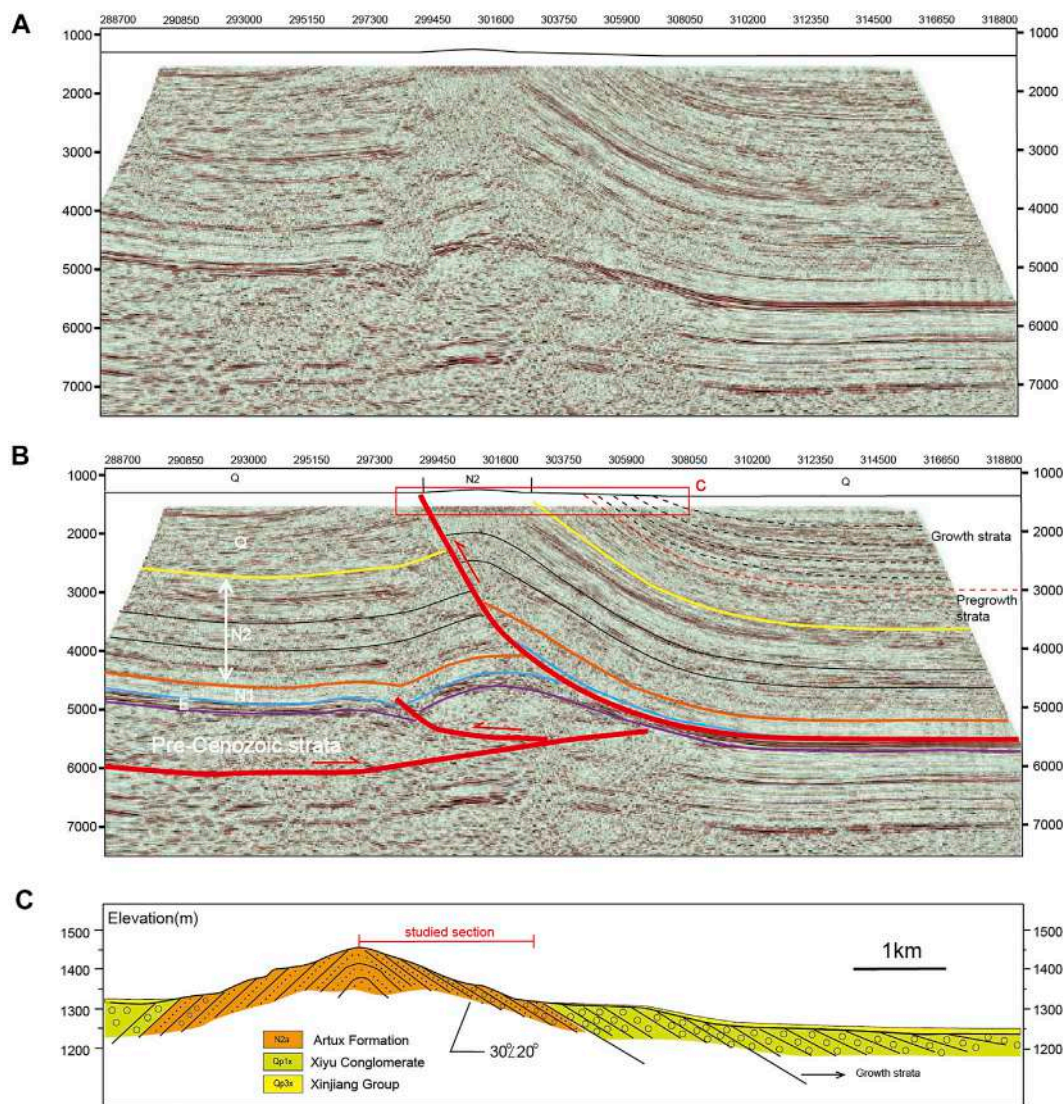


FIGURE 3

(A) and (B), Seismic reflection profile and its interpretation (by Gao, 2012). The location of the survey line was displayed in Figure 2B. The profile shows that the boundary (yellow solid line) between the Artux Formation and the Xiyu Conglomerate is near the top of the studied successions. The basal layer (red dot line) of the growth strata (black dot lines) is deposited within the Xiyu Conglomerate. (C), The schematic diagram for the sampling section as demonstrated in (B).

The lithostratigraphy of the Cenozoic sequences in the Western Tarim Basin along the western Kunlun-Pamir was divided into the Paleogene Kashi, the Miocene Wuqia groups, the Artux Formation, and the Xiyu Formation. Indeed, the Kashi Group is composed of the Aertashi, Qimugen, Gaijitage, Kalatar, Wulagen, and Bashibulake formations in ascending order, recording series of marine transgressions and regressions, while the Wuqia group, including the Keziluyi, Anjuan, and Pakabulake formations, was deposited in a terrestrial environment (BGMRXJ, 1999; Guo et al., 2002; Bosboom et al., 2014;

Chen et al., 2015; Sun et al., 2016; Li et al., 2017; Zhang et al., 2001).

In this study, we focused on the Artux Formation in the Yengisar section during the Pliocene epoch (Figure 2A). The Artux Formation consists of fluvial, yellow-gray to tan mudstone, siltstone, and fine- to medium-grained sandstone which generally coarsen upward (Zheng et al., 2002; Pei et al., 2008). The lithology of the upper part of the Artux Formation is marked by an abrupt coarsening and darkening (Figure 2). On the other hand, the Xiyu Formation overlies conformably the Artux Formation and is dominated by massive, thickly bedded, and

pebble-to-cobble conglomerates, typical of the channel and debris-flow deposits of alluvial fans and gravel-bed braided rivers, locally interbedded with minor lenticular sandstone and mudstone beds in its lower part (Chen et al., 2000; Chen et al., 2007a; Zhang et al., 2001).

Materials and methods

The sampled section is exposed in an eroded canyon. It is located in the middle of the northeast flank of the Yengisar anticline, about 25 km southeast of the Yengisar county (Figure 2). In the Yengisar, the strike fold anticline axis is N20°W. In fact, the sampled section is almost perpendicular to the fold axis, with a horizontal width of about 2 km (Figures 2B,C). The dipping strata direction is Northeast, with an angle ranging from 10° to 25°. According to the field measurement result in this study, the true thickness of the section is about 520 m. In addition, based on the abutting seismic profile of the section (Gao, 2012), the upper part of the Artux Formation is exposed in the canyon (Figures 2, 3), because it was covered by the diluvial fan sediments of the Xinjiang Group during the late Quaternary. Nevertheless, no contact between the Artux Formation and the Xiyu Formation at the top of the section was observed. The transitional contact should be near the top of the sampled section, to identify the basal layer of the growth strata developed in the Xiyu Formation (Figure 3B).

In total, 356 oriented core samples were collected from the bottom to the top of the sedimentary sequence using a portable gasoline-powered drill (Figures 2C, 3C), at a sampling interval of 0.25 m (if the petrology is available) (Figure 2E). Indeed, an interval of multiple meters was considered where the sequence is composed mostly of coarse sandstone and granule conglomerate (Figures 2G,H; Supplementary Figure S1). All the core samples were oriented first by magnetic compass after a local declination correction, determined from the International Geomagnetic Reference Field (IGRF), and then cut into cylindrical specimens using a double-bladed saw in the laboratory.

The rock magnetic experiments for representative specimens were conducted at the Key Laboratory of Paleomagnetism and Tectonic Reconstruction of the Ministry of Natural Resources, Institute of Geomechanics, Beijing. The temperature-dependence of magnetic susceptibility (κ -T curve) was carried out using the Kappabridge KLY-4, equipped with a CS-4 high-temperature furnace (Agico Ltd., Brno), by heating from room temperature to 700°C, followed by cooling back to room temperature. In addition, the first-order reversal curves (FORCs) and magnetic hysteresis loop were obtained using a Princeton Vibrating Sample Magnetometer (MicroMag 8600 VSM). The FORCs

diagrams were illustrated by FORCinel (v 3.06) using a smoothing factor value of 3 (Harrison et al., 2018).

In total, 331 specimens collected from the Yengisar section were subjected to stepwise thermal demagnetization (TD) experiment using a thermal demagnetizer (TD-48) in a magnetically shielded room, with a residual magnetic field of less than 300 nT. After each step of TD, remanent magnetizations were measured using a 2G-755 cryogenic magnetometer at the Key Laboratory of Paleomagnetism and Tectonic Reconstruction of the Ministry of Natural Resources, Institute of Geomechanics, Beijing. For all specimen, the demagnetization was carried out first within a temperature interval of 40°C or 100°C to 500°C, and subsequently in steps of 20°C or 10°C to 680°C.

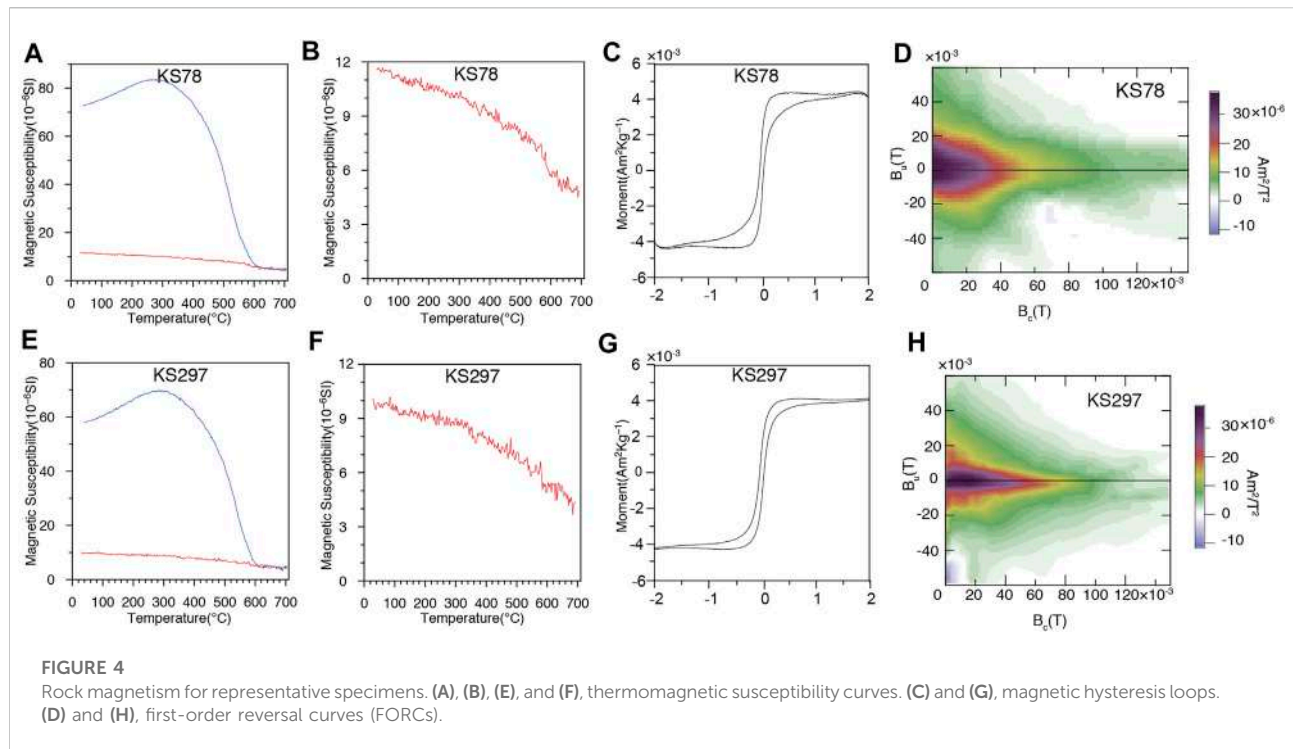
The demagnetization results were analyzed using principal component analysis (PCA) (Kirschvink, 1980) and Fisher statistic methods (Fisher, 1953). Furthermore, paleomagnetic data were analyzed using the online software paleomagnetism.org (Koymans et al., 2020).

The anisotropy of magnetic susceptibility (AMS) of 326 specimens was measured using the KLY-4 Kappabridge susceptibility meter at the Key Laboratory of Paleomagnetism and Tectonic Reconstruction of the Ministry of Natural Resources, Institute of Geomechanics, Beijing.

Results

Rock magnetic experimental results

Two representative samples were selected from the section for rock magnetic experiments, to obtain the behavior of magnetic materials. As shown in the κ -T curve diagrams (Figures 4A,B,E,F), the heating curves of the KS78 and KS297 have gradually decayed to 680°C, indicating that hematite is the dominant magnetic carriers in the collected samples. However, the cooling curves were remarkably higher than the heating curves, due to the formation of new magnetic minerals during the heating process. The magnetic hysteresis loop revealed that the representative samples are fully saturated at an applied field of 2 T, with a low coercivity of about 10–15 mT (Figures 4C,G). The shape of hysteresis loop show that samples have obvious characteristics of wasp-waisted, which may indicate that the magnetic minerals in the sediments are mainly the mixture of single-domain (SD) and superparamagnetic (SP) particle sizes or the combination of different kinds of magnetic minerals (Tauxe et al., 1996). Moreover, the FORC diagrams obtained show that samples were dominated by SD and pseudo-single domain (PSD) particles (Figures 4D,H). Rock magnetic results suggest that SD hematite is the primary magnetic carrier mineral in the sedimentary rocks of the Yengisar section.



Demagnetization

After the removal of viscous remanent magnetization at a temperature generally below 200°C , one stable magnetization component was extracted from the 322 samples collected from the upper part of the Artux Formation. Demagnetization behaviors of the remanent magnetization showed a linear decay to the origin at about 680°C (Figure 5). Indeed, nine specimens showed noisy thermal demagnetization behaviors, making isolation of the characteristic remanent magnetization (ChRM) difficult (Figure 5D). On the other hand, 322 ChRMs were of dual polarity and have passed the reversal test (Figures 6A,B). All the negative ChRMs are flipped to positive. Then 48 ChRM directions was rejected after a 45-cutoff and the remains reported a mean direction value of $D_g = 330.6^{\circ}$, $I_g = 61.1^{\circ}$ ($k = 17.6$, $\alpha_{95} = 2.2^{\circ}$) before and $D_s = 355^{\circ}$, $I_s = 46.6^{\circ}$ ($k = 16.6$, $\alpha_{95} = 2.1^{\circ}$) after tilt correction (Figures 6C–F).

Magnetostratigraphy

To define magnetic polarity zones (Figure 7), all the ChRM directions were used to calculate the virtual geomagnetic pole (VGP) and were plotted against stratigraphic levels.

The obtained result showed four pairs of normal and reverse polarity zones (N1–N4, R1–R4) with four tentative narrow intervals (inferred by less than three continuous specimens)

(Figure 7). Based on the seismic reflection profile, it is inferred that about 2 km thick of Artux Formation below the Xiyu Formation was developed in this region, while the studied section contains 520 m thick strata of the upper part of the formation, which have continuous contact with the overlying layers (Figure 3).

According to several authors, the basal age of the Artux Formation and the upper limit age are time-transgressive (Zheng et al., 2000; Chen et al., 2002; Chen et al., 2007b; Heermance et al., 2007; Chen et al., 2015; Qiao et al., 2016; Qiao et al., 2017). Furthermore, the chronology of the sedimentary package in the same section remains different. Zheng et al. (2000) reported that the Artux Formation, exposed in the Kekeya section at the southwestern edge of the Tarim Basin, is dominated by a long-reversed polarity zone, and assumed to correlate with the Gibert reversed chron. Thus, the Artux Formation has a depositional age ranging from 4.6 to 3.5 Ma (Figure 2A). Volcanic tuff intercalated in the Xiyu Formation was recently identified in the Kekeya section. Based on new radioisotopic data, the stratigraphic age of the Artux Formation ranges from ~ 22.6 to 15 Ma (Zheng et al., 2015a; Zheng et al., 2015b; Sun et al., 2015). Since the depositional onset of this coarse clastic episode is diachronous, the Artux Formation was considered as a lithostratigraphic unit, rather than a chronostratigraphic marker (BGMXJ, 1999). Based on the traditional opinion adopted, the Artux Formation is believed to be deposited approximately during the Pliocene, rather than the early Miocene, due to the micropaleontologic

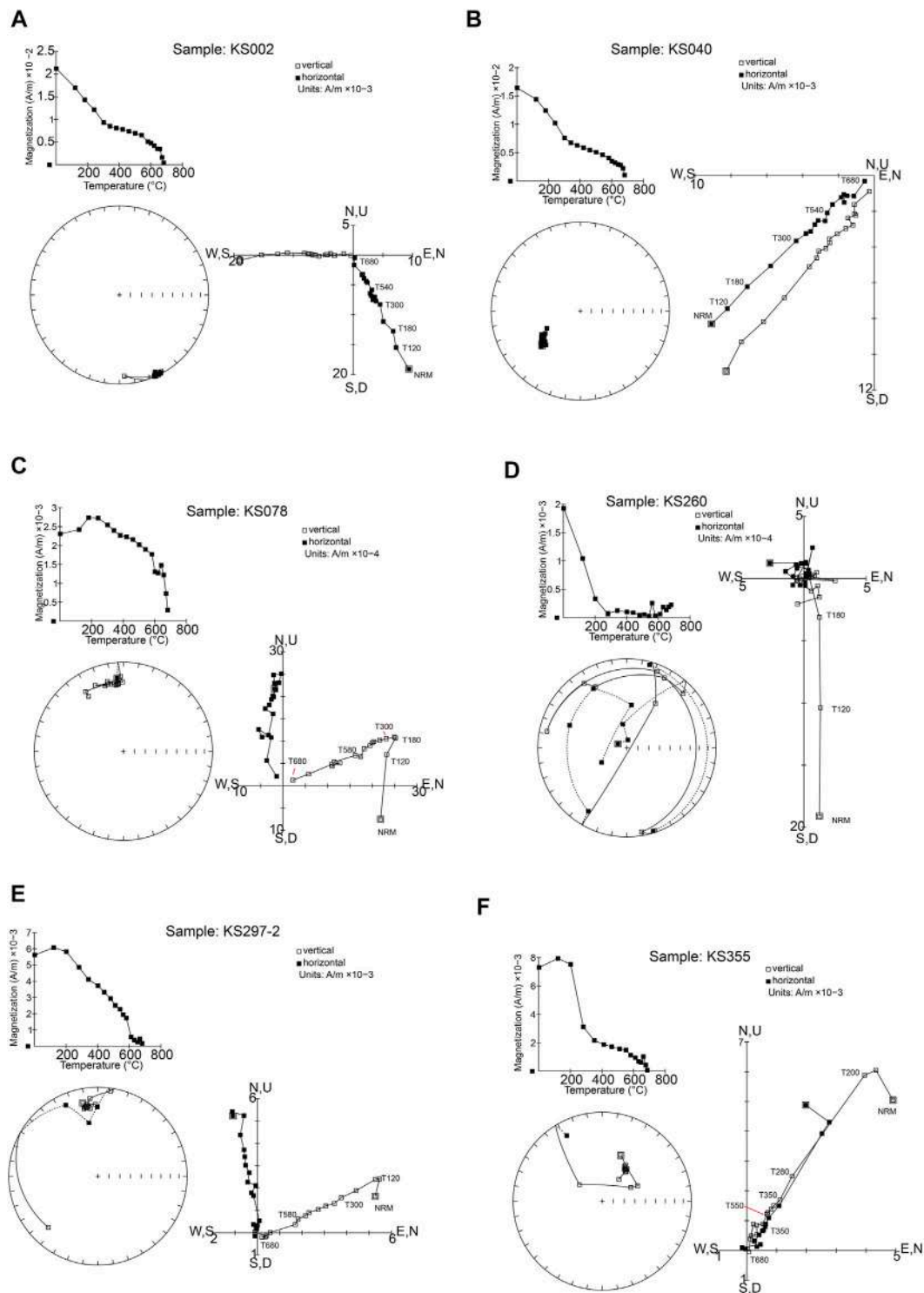


FIGURE 5

Zijderveld vector diagrams and Equal-area plots for representative specimens. (A) Sample: KS002, (B) Sample: KS040, (C) Sample: KS078, (D) Sample: KS260, (E) Sample: KS297-2, (F) Sample: KS355.

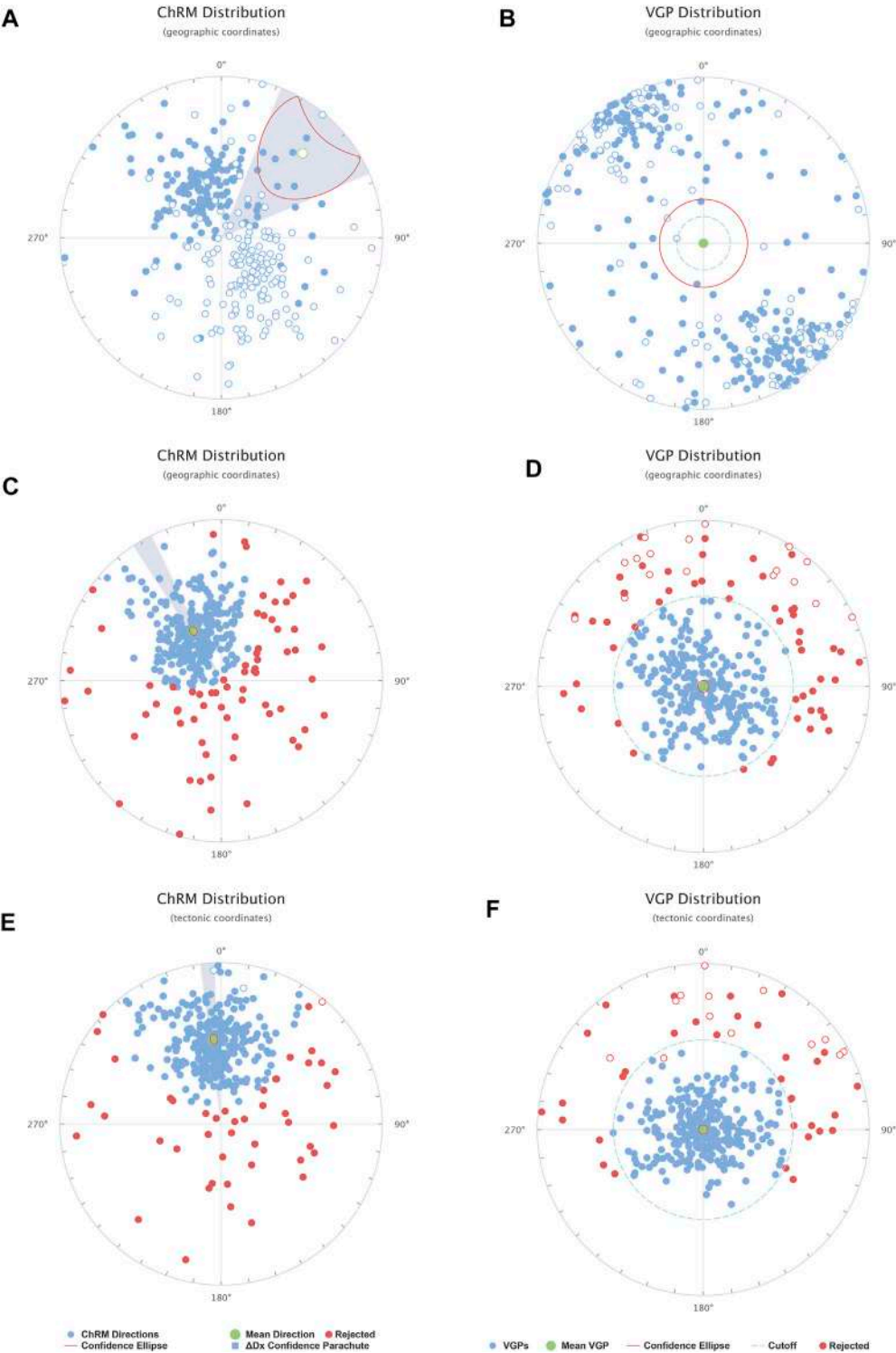


FIGURE 6
In-situ [geographic coordinates, (A, B, C, D)] and tilt-corrected [tectonic coordinates, (E, F)] Equal-area stereographic projection of ChRMs and VGPs from the Yengisar section. (A) and (B) show the raw data of dual polarity. (C), (D), (E) and (F) show the pattern after 45°-cutoff of all ChRMs contain positive and flipped directions.

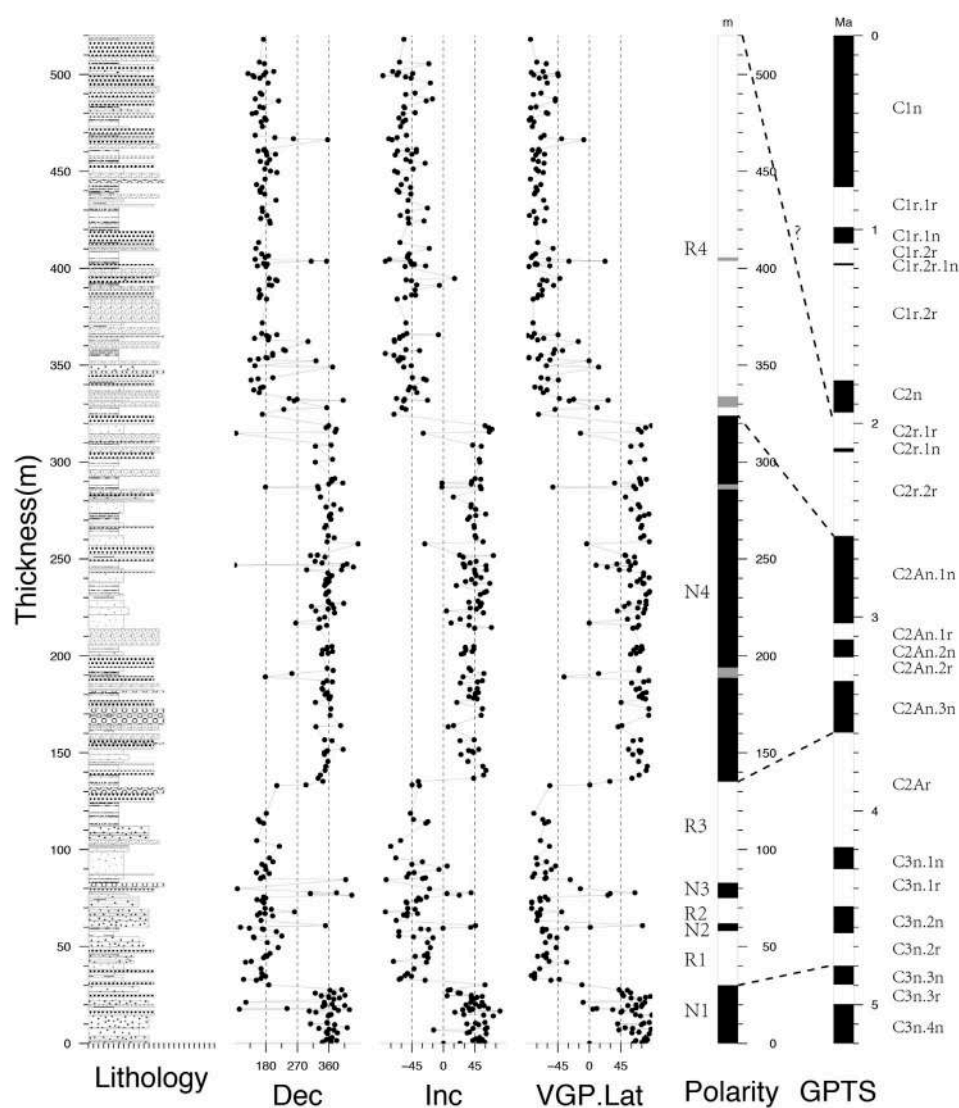


FIGURE 7

Lithology and magnetostratigraphic results from the Yengisar section in the western Tarim Basin. The characteristic remanence declination and inclination and VGP latitude are plotted as a function of stratigraphic level and the correlation with the geomagnetic polarity time scale (Gradstein et al., 2012).

evidence. Indeed, this formation contained Ostracoda and Charophyta assemblages, of which *Eucypris notabilis*, *Advenocypris decuria*, and *Candona neglecta* were found in the Pliocene strata in Russian, while *Cyprideis punctillata* was observed in the Pliocene strata series in the Qaidam basin (BGMRXJ, 1999). The micropaleontologic evidence for regional correlation with sedimentary sequence indicated that the lithostratigraphic ages of the Artux Formation cannot be much older than 5.3 Ma.

With these constraints, the magnetostratigraphy for the terrestrial sequence of the studied section can be reconstructed. The complete pattern of magnetic polarity zones presented in Figure 7, was correlated with the GPTS 2012 (Gradstein et al.,

2012). In fact, the middle part of the polarity zones (N4) can be correlated to C2An. Moreover, the magnetic polarity zones from N1 to R3 were correlated with chrons C3n2r through C2Ar. While the long-reversed intervals R4 were correlated to C2r.

According to the correlation analysis, the depositional age of the contiguous sedimentary package in the Yengisar anticline section varies from 4.9 to 1.9 Ma.

Anisotropy of magnetic susceptibility

The AMS can measure the rock fabrics, potentially governed by depositional conditions or post-depositional sedimentary

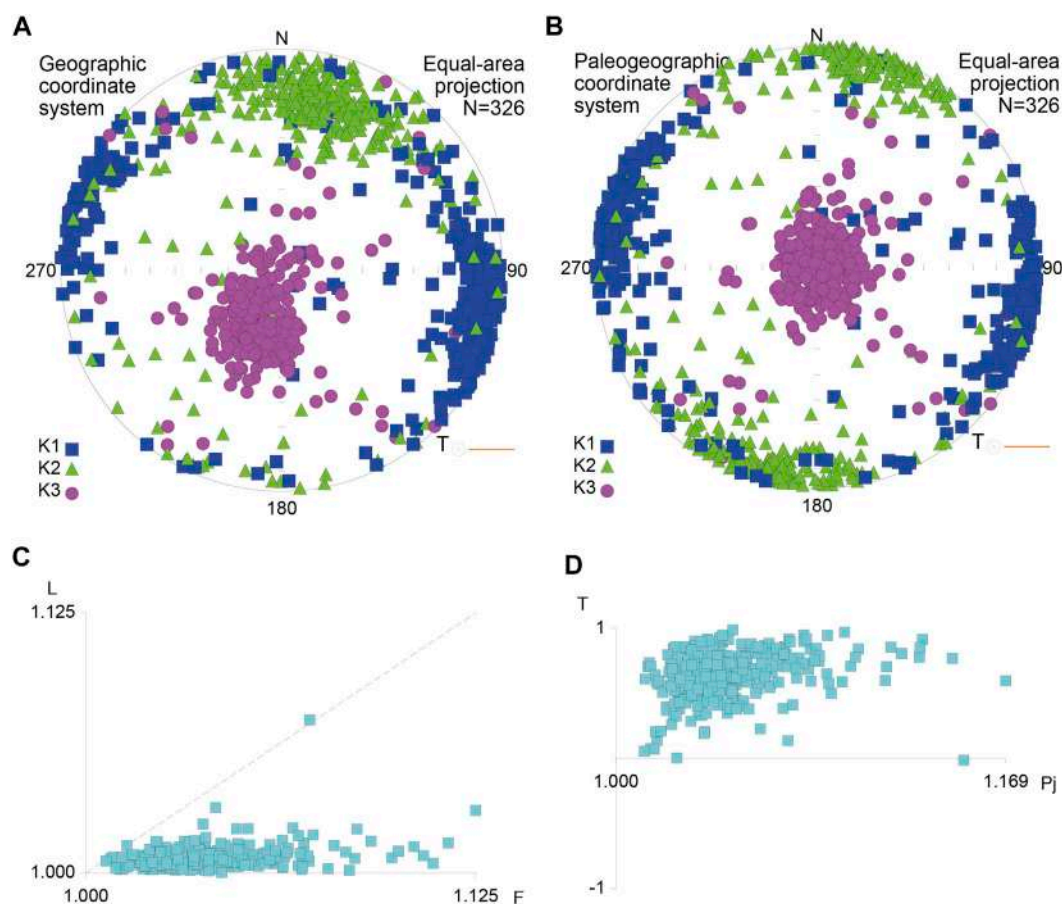


FIGURE 8

Anisotropy of magnetic susceptibility (AMS) results from the Yengisar section. Stereographic projections (lower hemispheres) of AMS orientations before (A) and after (B) tilt correction. (C), F (foliation)-L (lineation) diagram. (D), P_j -T (corrected AMS degree versus AMS shape parameter) plot.

process. This technique can also record the imprint of later tectonism (Borradaile and Henry, 1997; Borradaile and Jackson, 2004).

The AMS ellipsoids of the Artux Formation from the Yengisar anticline are typical primary sedimentary fabrics (Figure 8). This kind of magnetic fabric is controlled by depositional and diagenetic processes and is characterized by oblate AMS ellipsoids with maximum and intermediate susceptibility axes (K_1 and K_2 , respectively) within the bedding plane, and minimum susceptibility axis (K_3) perpendicular to the bedding plane (Figure 8B). All specimens plot in the oblate region ($K_1 \approx K_2 > K_3$) on F (foliation)-L (lineation) and P_j (corrected anisotropy)-T (shape parameter) diagram (Figures 8C,D).

The AMS parameters were filtered individually, and weights were given based on Gaussian function, with 20 m intervals in stratigraphic levels (Figure 9). The filtered data identified the trend of signal changes. As seen in Figure 9, the bulk magnetic susceptibility

(K_m) values range from 200 to 400 μSI , with a gradual decrease from the base level of this section (0 m, ~ 4.6 Ma) to the middle layers at 250 m (~ 2.9 Ma). After an abrupt increase in the K_m value, a second gradual decreasing trend to the 430 m level (~ 2.1 Ma) was observed (Figure 9). The overall low values of bulk susceptibility indicate that paramagnetic and antiferromagnetic minerals such as clays and hematite, predominantly control both the susceptibility and anisotropy of these studied sediments (Tarling and Hrouda, 1993), which is consistent with the rock magnetic results.

The P_j values range from 1.012 to 1.169, with a mean of 1.049. Indeed, the low P_j values and the relatively constant high level of spherical shapes (T) indicate that the magnetic anisotropy is the result of the compaction effect. The corrected anisotropy (P_j) values showed a similar trend as compared to the K_m parameter (Figure 9), resulting in a weak positive correlation ($r = 0.43$). On the other hand, a weak negative correlation ($r = -0.33$) was observed between the spherical shapes and the declinations of K_1 (Figure 9). This finding can be explained by the fact that the

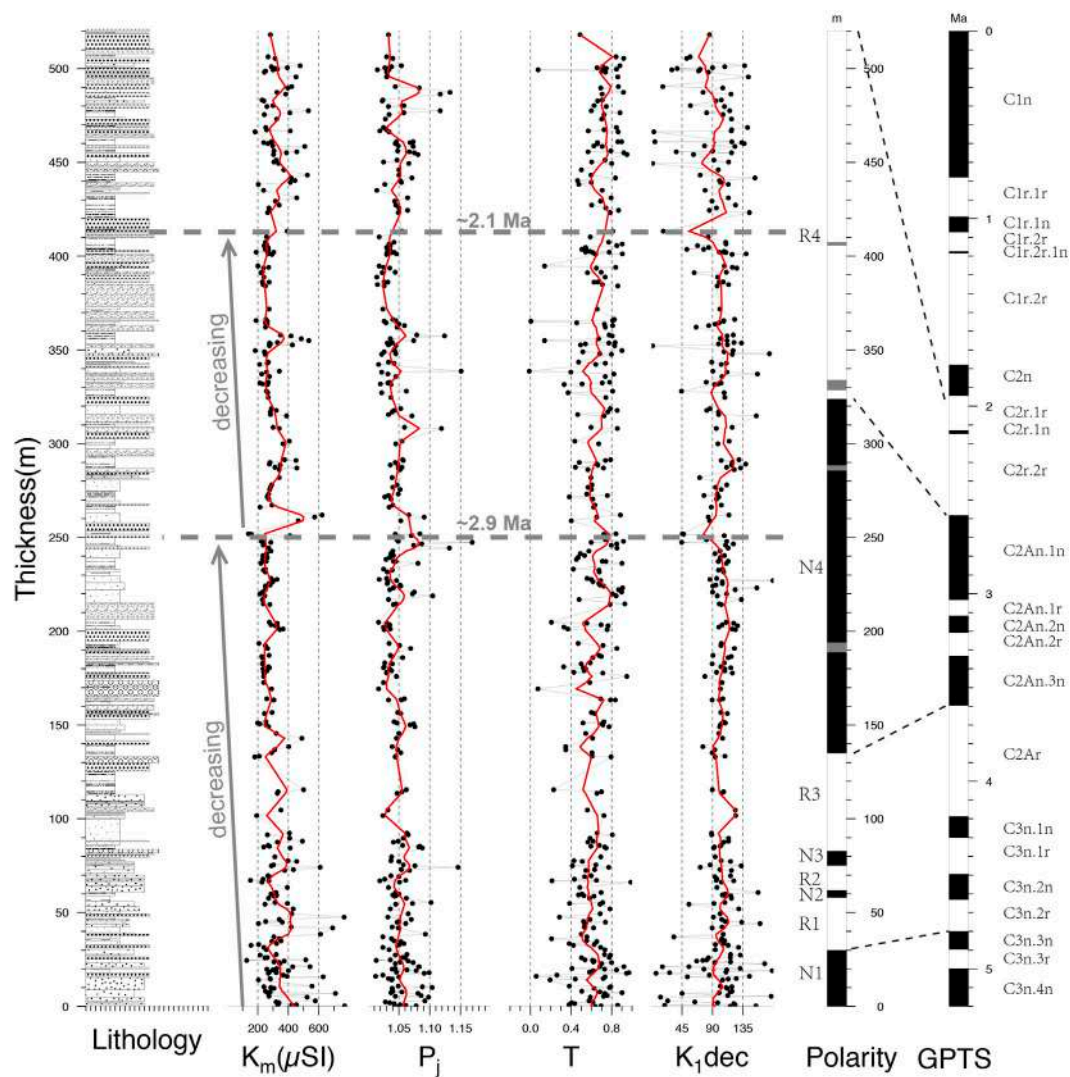


FIGURE 9

Magnetic susceptibility (K_m), the Degree of Anisotropy (P_j), the Shape Parameter (T) and the direction of the maximum susceptibility axes (K_1) as a function of depth in the Yengisar section. The red curves are obtained by data filtering, weights are given by Gaussian function with 20 m intervals in stratigraphic levels.

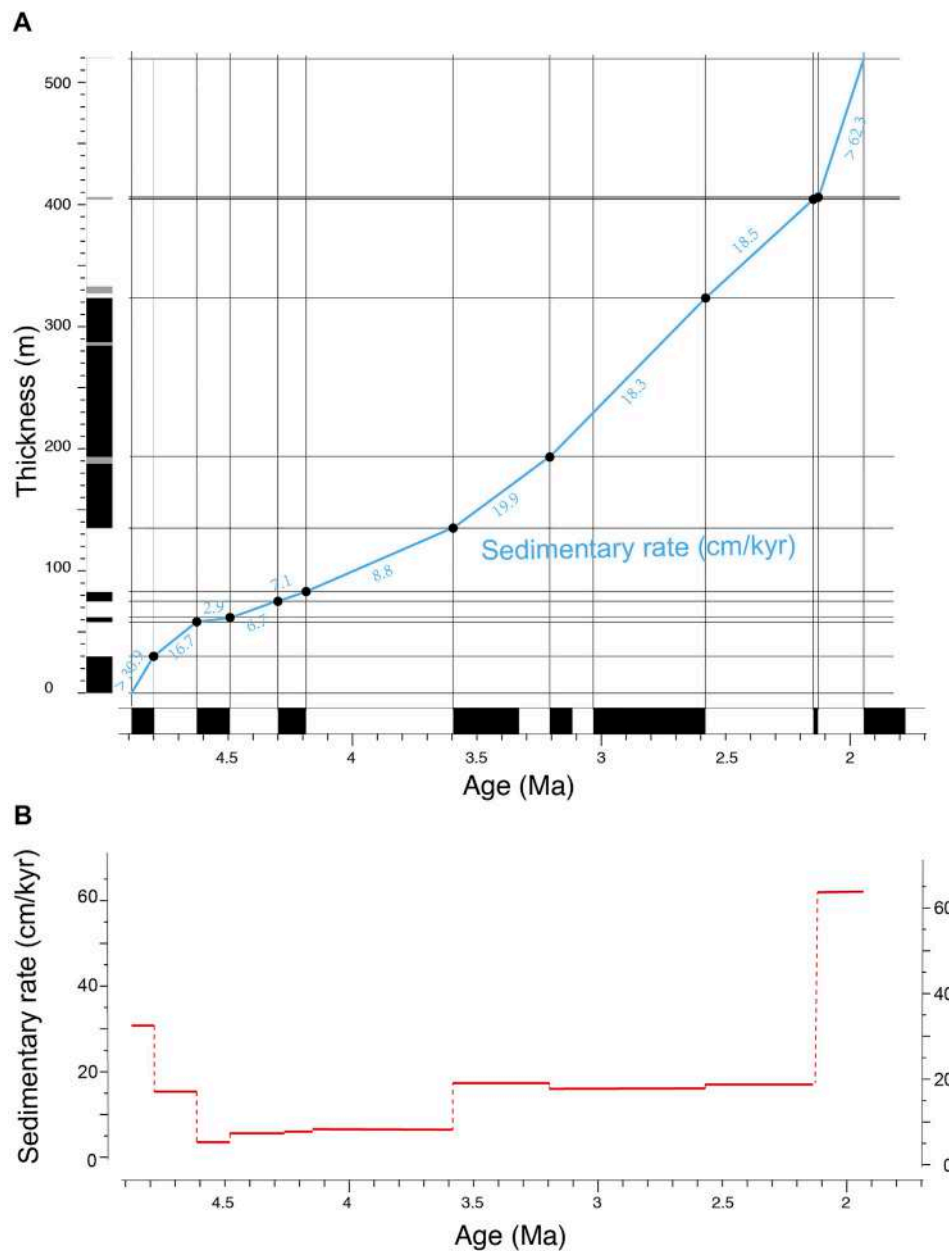
shapes of the AMS ellipsoids are generally influenced by the combination of depositional compaction and paleocurrent or compressional stress field.

Discussion

Changes in sedimentation rate and its triggers

The sedimentation rate can be determined by magnetostratigraphic ages and stratigraphic thicknesses (Figure 10A). The results revealed that the sedimentary

rates were not corrected for post-depositional compaction, due to the unknown history of compaction. In addition, the Tarim Basin is internally filled, and the total sediment flux directly reflects the denudation rates over the source region. Therefore, rates represented here should be minima that show how sedimentation rate change at specific localities relative to the topographic evolution of the range (Metivier and Gaudemer, 1997; Charreau et al., 2006). As reported above, the directions of the paleo-drainages were E-W to SEE-NWW, as indicated by the K_1 directions (Figure 9). In addition, the Pliocene alluvial occurs mostly at the outlets of modern rivers, were responsible for the Artux Formation (Figure 11). These two pieces of evidence indicate that the

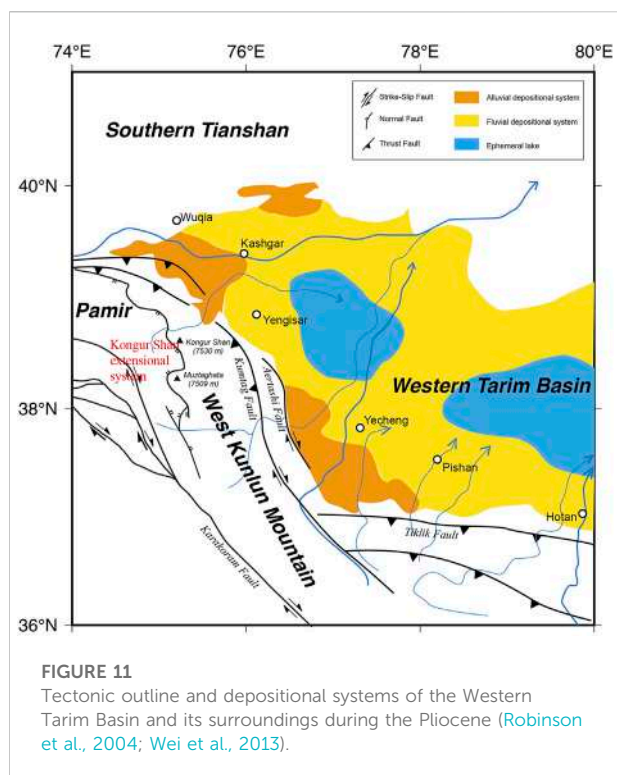
**FIGURE 10**

The stratigraphic height as a function of magnetostratigraphic age in the Yengisar succession of the western Tarim Basin showing sedimentation rates in the studied section. **(A)** Plot of the sedimentary thickness versus time; **(B)** Plot of sedimentary rate variation over time.

provenance of the Pliocene sedimentary successions in Yengisar anticline is the West Kunlun Mountain rather than the South Tian Shan.

The magnetostratigraphic age plot within the composite section (Figure 10A) showed a gradual decrease in sedimentation rate, from >30.9 cm/kyr to 2.9 cm/kyr during the 4.9–4.6 Ma period, representing about 10-fold change.

This may be due to the presence of weakened topographic uplift in the provenances and the subsequent equilibrium patterns between erosion and deposition in the same regions. The sedimentation rate period (2.9 cm/kyr) continues about 0.1 Myr (Figure 10A). Abrupt increases in sedimentation rate ranging from ~8 cm/kyr to ~19 cm/kyr and from ~19 cm/kyr to >62.3 cm/kyr were observed at 3.6 Ma and 2.1 Ma, respectively. By



considering the ascending order of the frequent and thicker massive coarse sandstone layers, it is argued that these two changes in sedimentation rate explain the move away from the denudation region, which was more likely induced by the strong basinward thrusting of the orogenic belts adjacent to the western Tarim Basin that has occurred since the Pliocene (Qu et al., 2004; Wei et al., 2013).

The first decrease (at ~4.6 Ma) and the two subsequent increases (at 3.6 Ma and 2.1 Ma) in the sedimentation rate indicate that the Western Kunlun Mountain was characterized by pulse and strength uplift events during the Pliocene epoch (Figure 10B).

However, distinguishing between tectonic and climatic forces on sedimentation rate changes is not straightforward (Molnar and England, 1990). According to Zhang et al. (2001) and Wang et al. (2003), the significant increases in grain sizes of sediments and sedimentation rate were governed by climate changes during the Pliocene period. Based on the tectonically driven hypothesis, the tectonic uplift events were among the first-order control factors. An abrupt and frequent climate change since ~5 Ma was considered responsible for the disequilibrium of the landscape's erosion (Zhang et al., 2001; Westerhold et al., 2020). If the denudation region was in a weakened uplift activity a new equilibrium configuration is expected to be reestablished after a higher sedimentation time. Since the Pliocene, the rates observed from the studied section were not consistent with this plausible prediction. In contrast,

three abrupt changes (~4.6, 3.6, and 2.1 Ma) were detected through the high-resolution magnetostratigraphic data (Figure 10). More likely, the basinward propagation of the provenances was induced by the pulsating exhumation of the ranges. Moreover, the deceleration of dextral slip between Pamir and Tarim was accommodated in the Eastwestern Kongur extensional system (Figure 11) (Arnaud et al., 1993; Brunel et al., 1994; Robinson et al., 2004; Sobel et al., 2011). Plentiful detrital zircons ages are synchronous with major fluvial sedimentation phases, at the outlets of the northeast-flowing rivers in the Tarim Basin, suggesting rapid Pliocene–Quaternary exhumation of the extensional system (Cao et al., 2013; Cao et al., 2018).

Constraints on the tectonic-sedimentary process during the Pliocene

The pre-Pliocene deposits of the Western Kunlun Mountain front record the transition in the depositional environment from the shallow marine to the inland sea, then to a foreland basin in response to the northward impact of the Pamir salient on the basin (Zhang et al., 2001; Bosboom et al., 2011; Sun and Jiang, 2013; Bosboom et al., 2014; Li et al., 2017; Li et al., 2019). Since the Pliocene, the Artux Formation and the Xiyu Formation have been deposited along the Western Tarim Basin margins, and have been considered as the molasse in the foreland basin. These upward coarsening successions provide important information on the growth processes of the surrounding ranges (Zheng et al., 2000). In some sections, eolian sandstone was also identified within these coarse sediments, providing crucial evidence for the inland desertification origin (Sun and Liu, 2006; Zheng et al., 2015a; Sun et al., 2015). Thus, the Artux Formation was most closely related to the present geologic environment and the modern river systems (Figure 11).

During the Pliocene, the depositional systems from source to sink might have changed over time, as a result of the northward encroaching and the intense basinward thrusting of the orogenic belts. The strata of the Yengisar section show the shift upwardly from distal to middle fan systems during the depositional periods of the Artux Formation (Supplementary Figure S1). Subsequently, the alluvial fan gradually propagated basinward until the termination of the Xiyu Formation. The seismic section profile indicated that the basal age of the Xiyu conglomerate and the commencement of growth strata occurred in the early Pleistocene (1.9 Ma) and the late Pleistocene (1.45 Ma), respectively (Figure 3). These ages were estimated by the ratio of 1.8, obtained using the ratio of the Artux Formation to the Xiyu Formation sedimentation rate from the Kekeya section (Zheng et al., 2002). The Yengisar anticline showed a fold deformation at approximately 1.45 Ma, corresponding to the depositional

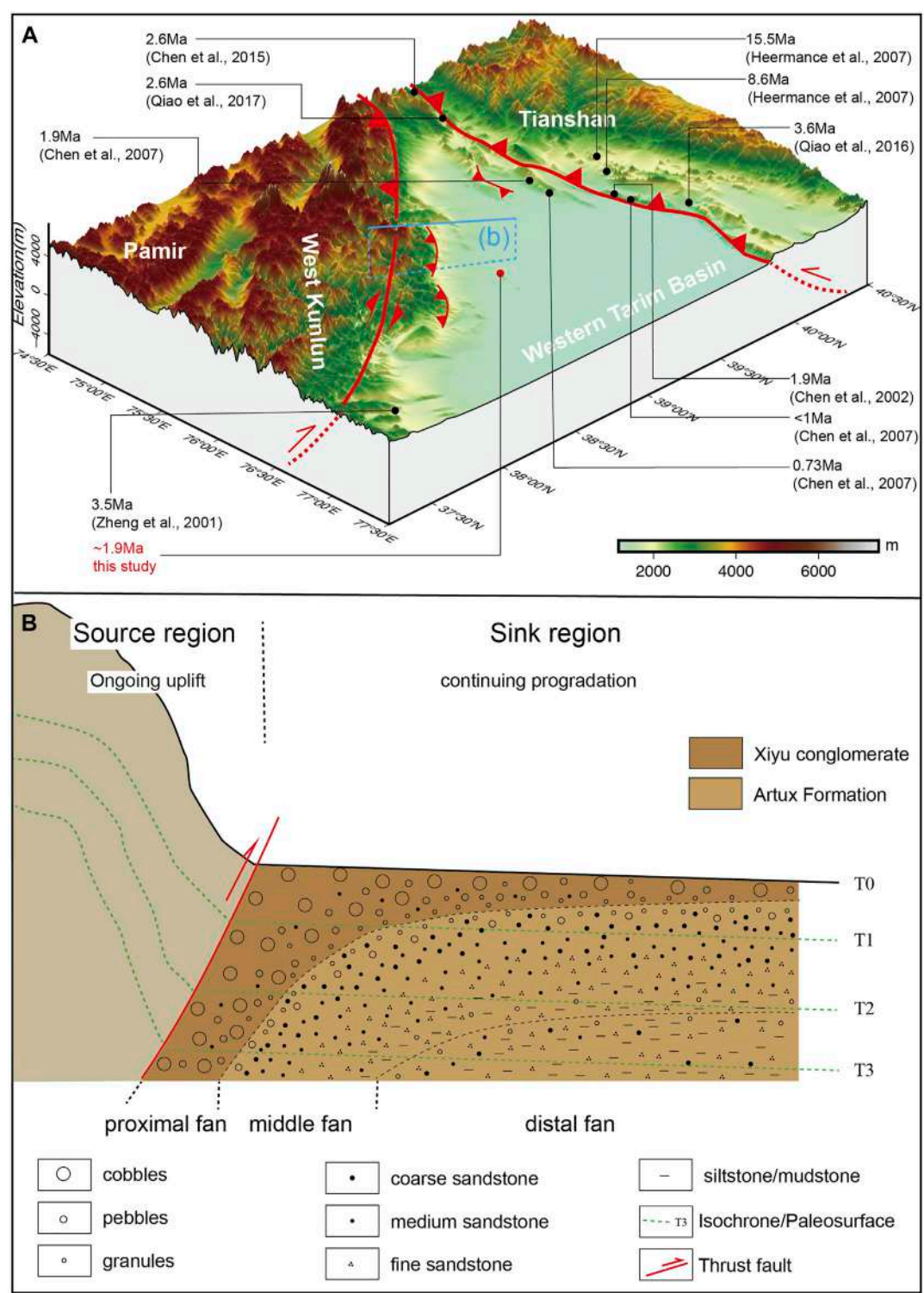


FIGURE 12
(A), Basal ages of the Xiyu Formation in the edge of western Tarim Basin. (B), Conceptual model of the progradational alluvial fan of upward coarsening deposits in the piedmont of an ongoing uplift mountain.

age of the Xiyu Formation (Pan et al., 2009; Fu et al., 2010; Li et al., 2010; Liu et al., 2011).

New perspective on progradation of the Artux Formation and Xiyu Formation

Based on detailed magnetostratigraphy of the upper Cenozoic strata from seven sections in the Southwestern Tian Shan, Chen et al. (2007a) suggested that the basal age of the Xiyu Formation ranges from 15.5 Ma at the northernmost part of the foreland, to 8.6 Ma in the central part of the foreland (Heermance et al., 2007), and from 1.9 Ma, to 0.73 Ma along the southern deformation front of the foreland basin (Figure 12A).

Recently, Qiao et al. (2016) published a new magnetostratigraphic data of the Neogene successions in Kashi area and suggested that the basal age of the Xiyu Formation is 3.6 Ma. On the other hand, in the northeastern part of the West Kunlun Mountain, detailed magnetostratigraphy of 4.5 km Kekeya section near Yecheng showed that the onset of the Xiyu Formation was commenced at 3.5 Ma (updated to ~15 Ma by new chronological data, with refutation) (Zheng et al., 2000, 2015a, 2015b; Sun et al., 2015), whereas the age of the basal conglomerate in the Yengisar section is ~1.9 Ma in this study. All these data suggest that the Xiyu Formation is highly time-transgressive and cannot be considered a chronostratigraphic rock unit (Chen et al., 2007b; Heermance et al., 2007; Qiao et al., 2016).

There are two competing hypotheses of the sedimentary transition recorded in the Xiyu Formation. The first emphasizes that the increases in grain sizes of sediments were caused by frequent and abrupt climate changes during the post-Late Pliocene (Molnar et al., 1993; Zhang et al., 2001; Wang et al., 2003). While another hypothesis argues that the massive conglomerates are the sedimentary response of an intense exhumation of the orogeny related to continuous India-Eurasia convergence (Huang, 1957; Li et al., 1979; Zheng et al., 2000; Chen et al., 2001; Huang et al., 2006; Zhao, 2019).

The conformable contact showed the lack of significant tectonic movement when the Artux Formation deposition translated to the Xiyu Formation in some outcrops. This is one of the main pieces of evidence for advocates of climate hypotheses to disapprove tectonic-driven theory (Zhang, 2004; Zhao et al., 2008b). However, from the source-to-sink perspective, the conformable contact showed that the tectonic environment in the sink region was relatively equable. The abrupt increase in sediment particle size in the sink region is due to the strong uplift in the source region (Figure 12B). In fact, the initial depositional age of the Xiyu Formation ranges from

15.5 Ma to less than 1 Ma, which does not reconcile the climate origin models, due to the synchronous climate changes.

It seems reasonable that the spatiotemporal variation in tectonic uplift events explains the time-transgressive of the massive conglomerates. However, explaining the diachronous successions in the local area (southwestern piedmont of the Tian Shan) remains unclear. Indeed, Chen et al. (2007a) proposed the progradation model to address this issue. They argued that the extension of the thrust fold belt toward the interior of the basin was well consistent with the basal age changes of the conglomerate, thus, the propagation of the conglomerate is directly controlled by local tectonic deformation. Is there a straightforward causal linkage between basinward progradation of conglomerate and pulsed migration of deformation?

Certainly, the increase in gravel progradation rates appears linked to higher deformation rates across the piedmont anticlines in the Southwestern Tian Shan since 5 Ma (Chen et al., 2007b; Heermance et al., 2007). There is no consensus on the sequential onset between the bottom of the Xiyu Formation and the growth strata that represent tectonic deformation in the piedmont of the Southwestern Tian Shan and the northeastern part of West Kunlun Mountain (Figures 1, 2). Unlike the Southwestern Tian Shan, in front of the West Kunlun Mountain, the growth strata in the Kekeya section were developed mainly in the Artux Formation (Zheng et al., 2002). Therefore, no causal links exist between the diachrony of the Xiyu Formation and the migration of the local tectonic deformations.

We put forward a new perspective that both the conglomerate progradation and migration of the deformation front are the sink region responses to the uplift events of the source region, in sedimentology and structural geology, respectively.

The Artux Formation and the Xiyu Formation were originally considered to be a suite of typical molasse deposits. According to Zheng et al. (2002), Chen et al. (2007a), Zhao, (2019), the Artux Formation is typical in braided river facies, and some depositional segments are mid-fan to distal fan sub-facies, filled in an alluvial-fan system draining from the West in the Kashi subbasin (Wei et al., 2013). On the other hand, the lower part of the Xiyu Formation comprises middle and distal alluvial fan facies, while the upper part is mainly proximal fan sub-facies (Figure 12B). Indeed, facies analysis showed that both formations are typical progradation alluvial fan models. At a fixed fan cross-section, the sediments coarsen upwards over time due to the onset of the orogenic front. The progressive migration of the alluvial fan basinward resulted from the uplift events of the orogen. The development of the prograding alluvial fan strata is different from that of the interior of a large-scale stable basin. In the former conditions, its stratigraphic layers have an obvious time-transgressive feature (Figure 12).

On the other hand, induced by the incremental altitude difference, the lateral extrusion gradually propagates from the bottom of the mountain to the basin, accompanied by the immense uplift.

Based on the results of remote sensing techniques, field investigations, tectonic geomorphology as well as active tectonic in the edge of Southwestern Tarim Basin, reported by several researchers (Pan et al., 2009; Fu et al., 2010; Li et al., 2010; Liu et al., 2011), the frontal thrust belt of the mountains are attributed to the boundary fault of the plateau in the form of thrusting extension in Pliocene and Pleistocene (Qu et al., 2005). In other words, the syntectonic growth strata have deposited on the flanks of one anticline during the tectonic deformation stage, which belongs to the structural geologic response to the continuous uplift of the surrounding mountains.

Conclusion

In this study, a detailed magnetostratigraphic investigation in the Yengisar section located at the southwestern part of the Tarim Basin was conducted. This investigation provided a high-resolution chronology for the Pliocene sedimentary successions. In fact, magnetic polarity zones correlation suggests that the sampled successions were deposited from the ~4.9 Ma to 1.9 Ma. Combining the seismic reflection profile and previous magnetostratigraphic data in this region, the Xiyu Formation and the growth strata began deposited at about 1.9 and 1.45 Ma, respectively.

On the other hand, the observed sedimentation rate revealed a gradual decrease, from >30.9 cm/kyr at 4.9 Ma to 2.9 cm/kyr at 4.6 Ma. This 10-times decrease in sedimentation rate was induced by the weakening of the topographic uplift of the Western Kunlun Mountain. In addition, abrupt increases were found in the sedimentation rate, from ~8 to ~19 cm/kyr at 3.6 Ma and from ~19 cm/kyr to >62.3 cm/kyr at 2.1 Ma. These increases in sedimentation rate are due to the propagations of the strong basinward thrusting triggered by the pulsating growth of the Western Kunlun Mountain.

The basal age of the Xiyu Formation varies from Middle Miocene to Late Pleistocene. Diachrony of the Late Cenozoic upward-coarsening deposits is the result of the alluvial fan propagation induced by continued uplift and increased exhumation of the mountains. We favor the tectonic-theory for the origin of massive conglomerate and argue that the time-transgressive of the basal layer is an intrinsic feature of alluvial fan propagation rather than the migration of the local deformation front.

Data availability statement

The datasets presented in this study can be found in online repositories. The names of the repository/repositories and

accession number(s) can be found below: https://github.com/zifeizz/Data_PMD.

Author contributions

ZZ: Conceptualization, methodology, investigation, formal analysis, writing—original draft; JP: Conceptualization, funding acquisition, resources, supervision, writing—review, and editing; JL: Resources, investigation, review, and editing; YC: Investigation, data preparation, review, and editing; LH: Investigation, visualization, review, and editing.

Funding

This work was financially supported by the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901), the National Natural Science Foundation of China (Grant 42002243, 41672202).

Acknowledgments

We thank the Editor and reviewers for providing constructive comments and suggestions leading to the improvement of the manuscript.

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.967346/full#supplementary-material>

SUPPLEMENTARY FIGURE S1

Detailed stratigraphic level for sampling.

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 02 June 2022

ACCEPTED 20 July 2022

PUBLISHED 23 August 2022

CITATION

Zhou Q, Shen X, Huang H, Cheng S and
Zhang J (2022), Reworking of ancient
tectonic amalgamation belt beneath the
central north of North China Craton
revealed by dense seismic observations.
Front. Earth Sci. 10:960358.
doi: 10.3389/feart.2022.960358

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Reworking of ancient tectonic amalgamation belt beneath the central north of North China Craton revealed by dense seismic observations

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The North China Craton (NCC) is one of the oldest cratons in the world, and its internal tectonic belt is often used to investigate the earth's tectonic evolution events. During the late Mesozoic and Cenozoic, the western Pacific subduction zone caused the restructuring of NCC by damaging the craton beneath eastern NCC, resulting in the distinct lateral differences between western and eastern NCC, which ultimately formed the current NCC. Furthermore, the subsequent tectonic events activated the ancient tectonic weak zones, and their traces are imprinted in the deep earth. Here, we investigated the crust structures with a high-density seismic array beneath the splice position of the eastern margin of the Khondalite Belt and the northern part of the central orogenic belt in NCC. The array included 140 short-period seismographs spaced at 2–3 km intervals, which recorded teleseismic three-component waveforms over a one-month period. P-wave receiver functions calculated from 25 teleseismic events provided an image of the crustal structure. The weak Moho and Moho offset under the study area are visible in the migration image of receiver functions. The geological investigations and the rock outcrops were combined to establish the strong coupling relationship between the present surface fault-depression system and deep structures. The deep material circulation, which governs the surface extension of the basin-range structure, is controlled by the deep material circulation which is ultimately derived from the continuous subduction of the western Pacific. The study's findings indicate that the ancient amalgamative belt might have transformed into a weak zone easily susceptible to modification by plate tectonic movements.

KEYWORDS

North China Craton, dense seismic observations, receiver function, weak zone, reworking

1 Introduction

Several tectonic events in the North China Craton (NCC) since the Archean have been documented, and the available geological evidence explains the various phases of NCC. Three major tectonic events that occurred during the late Paleoproterozoic were reported by many scholars (Kusky et al., 2016; Li S et al., 2015; Zhai and Santosh, 2011; Zhao et al., 2005, 2012, 2013). The western NCC (WNCC) was formed by the amalgamation of the Yinshan block and the Ordos block at ~1.95 Ga, while the eastern NCC (ENCC) was formed by the collision between Longgang and Nangrim blocks at ~1.90 Ga. Furthermore, WNCC and ENCC amalgamated as a result of a tectonic event at ~1.85 Ga to form the uniform basement of NCC along a central orogenic belt (Trans-North China Craton, TNCO) (Zhao et al., 2001; 2005). The NCC remained stable for a long time until the significant Mesozoic lithospheric thinning. Numerous ancient tectonic structures have been transformed due to the collision between India and Eurasia plates and the subduction of the Pacific plate, which resulted in the elevation of the Tibetan Plateau (Zhu et al., 2011; 2012). This indicates the vulnerability of the amalgamation belts to the subsequent tectonic disturbances. This Paleoproterozoic evolution process has also resulted in a unique tectonic background at the boundary between the eastern side of the

Khondalite belt and the northern part of TNCO (Figure 1A). The central tectonic belt and Khondalite rock belt are connected along the Huai'an-Zhangjiakou area in Hebei province, China. The Shanxi-Hebei-Inner Mongolia basin tectonic belt, comprising a series of parallel Holocene faults and basins, characterizes the current surface structures and activities that are relatively distinctive (Figure 1B&C). The tectonic environment in the area is responsible for the relatively high tectonic activity (Li Z et al., 2015), as evidenced by the earthquake with ML = 6.2 triggered in Zhangbei in 1998 (Yang et al., 2002). The potential of this area for the occurrence of strong earthquakes is also evidenced by the GNSS observation data (Wu et al., 2021). The Huai'an-Zhangjiakou is a highly significant area of TNCO due to the area's paleotectonic and neotectonic activities. From northwest to southeast, the area comprises Yinshan block, Khondalite series, Huai'an complex, and Hengshan complex which are the subdivisions of metamorphic outburst classification (Wang et al., 2016). The temperature and pressure environment at a depth of 35–40 km at the bottom of the Paleoproterozoic lower crust is evidenced by the high pressure in the mafic granulites. The subduction–collision–exhumation events of Paleoproterozoic orogenic tectonic development transported these high-pressure granulites and eclogites into the core crust (Zhang et al., 2019). Geophysical methods can effectively understand the current deep structures and the

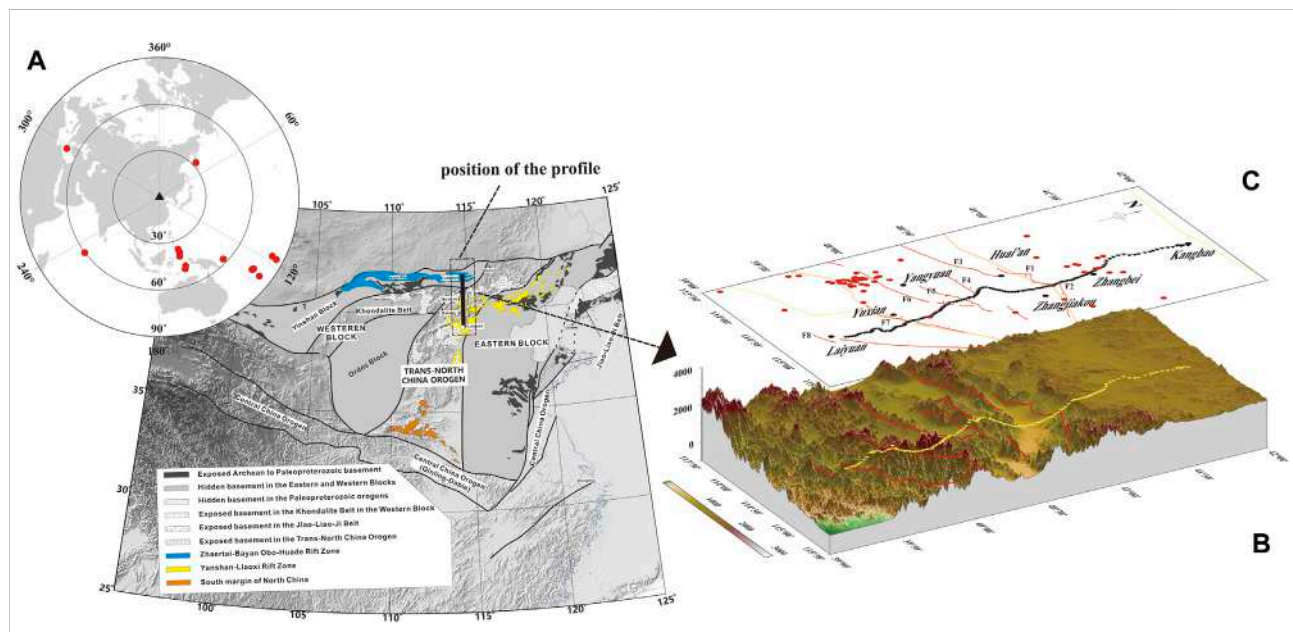


FIGURE 1

Geological background and the station map. (A) Integration of the North China Craton and the position of the profile (Zhao et al., 2012) and teleseismic events and station distribution. Red circles represent the events recorded, and the black triangle represents the center of the array; (B) spatial relationship between the topography, stations, Cenozoic fault distribution, and dense seismic array in the study area; (C) plane relationship between platform array and faults. The fault names are: F1-Huai'an-Wanquan Basin northern margin fault; F2-Zhangjiakou fault; F3-Tianzhen-Yanggao basin northern margin fault; F4-Huai'an Basin southern margin fault; F5-Yangyuan basin northern margin fault; F6-Liulengshan basin northern margin fault; F7-Yuguang basin southern margin fault; F9-Sunzhuangzi-Wulonggou fault; and F10-Xuanhua Basin southern margin fault.

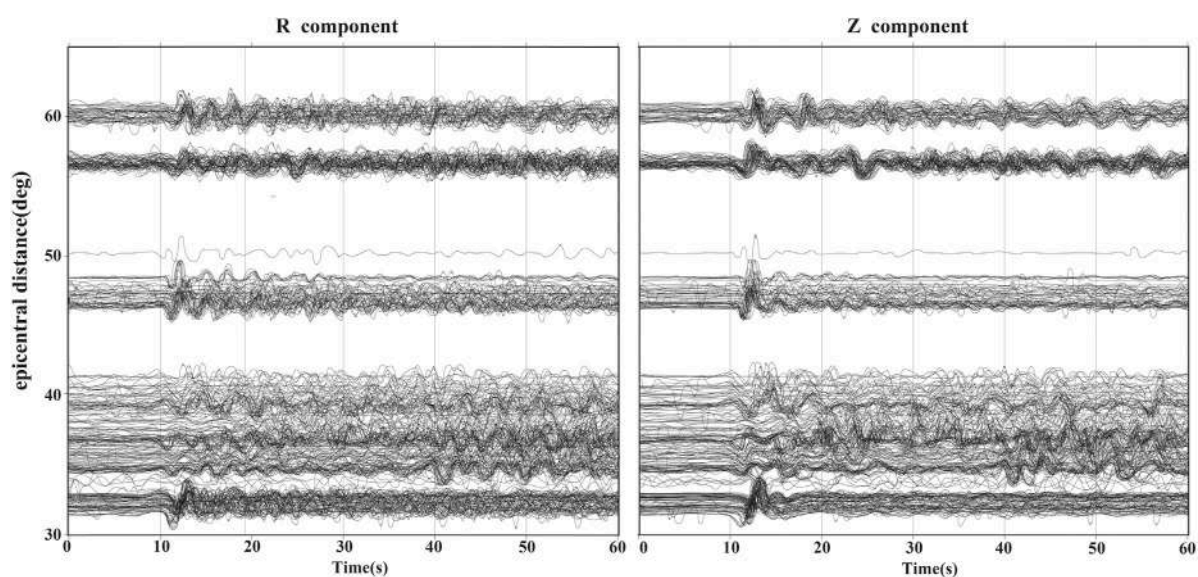


FIGURE 2

Observed teleseismic waveforms of R and Z components (after 0.02–2 Hz bandpass filtering).

tectonic processes. Several scholars have reported the strong concordance of surface topographic changes in the area and the E-W variations in the crustal and lithospheric structures (e.g., Chen, 2010; Cheng et al., 2013; Wei et al., 2015; Zheng et al., 2017). NCC and several other cratons in the world have restructured as a result of subduction (Zhu et al., 2021), and the pre-existing structures in the lithosphere primarily control such continental reconfiguration. However, deep imaging results of satisfactory scales in this local area are still lacking. In this context, the reconfiguration of the ancient tectonic amalgamation belt was investigated by procuring the higher precision imaging results of deep structures. For this purpose, a short-period array was deployed in the Huai'an-Zhangjiakou area to apply the receiver function method for exploring deep structures. The present tectonic activity explained the activation of the ancient tectonic structures based on the imaging results.

2 Method and data

2.1 Method

The receiver function is a reliable seismological method widely applied for the detection of discontinuities in the crust and upper mantle. This approach extracts the signatures of the structure under the station by the deconvolution of the R and Z directions of seismic wave records. The depth of the discontinuities, the crustal thickness, and the average velocity ratio of the crust can be estimated by analyzing the time and amplitude of the P-to-S wave of the crust or upper mantle discontinuities. This method is also used for the

development of broadband seismology (Langston, 1979; Yuan et al., 1997; Zhu and Kanamori, 2000) and understanding the crustal and upper mantle structure globally and regionally with a multitude of intriguing conclusions. However, the distance between stations in broadband seismic array observation is generally more than 10 km, with longer periods of observation, typically several years, due to the constraints of observation cost and environment, which also limits its implementation. Apart from this, the spatial resolution is also limited due to the minimum distance of such stations' separation. However, this method has been successfully applied recently in dense array observation data for a shorter time period. In petroleum exploration, a dense array of stations are established flexibly and quickly at several hundred meters apart by the operation of low-cost and simple nodal seismographs. After a short period of continuous observation (1–2 months), suitable teleseismic earthquakes can be identified for high-resolution crustal structure imaging. Several studies have been successfully carried out globally on the investigation of short-period dense array observation data. For example, a series of innovative research studies have been performed in Long Beach (e.g., Lin et al., 2013; Bowden et al., 2015; Bianco et al., 2019), a representative study area where a shallow magma chamber was imaged under the Wudalianchi volcanic field using highly dense arrays (Li et al., 2016). Similarly, Wang et al. (2019) imaged the fault damage zone of the San Jacinto Fault near Anza using a dense array of ambient noise tomography. The unprecedented details of the growing crustal structure along various boundaries of the Tibetan Plateau were unveiled by a series of short-period dense arrays (e.g., Liu X et al., 2017; Shen et al., 2020, 2022; Tian et al., 2021). Furthermore, the existence of Paleoproterozoic residual subduction in the craton

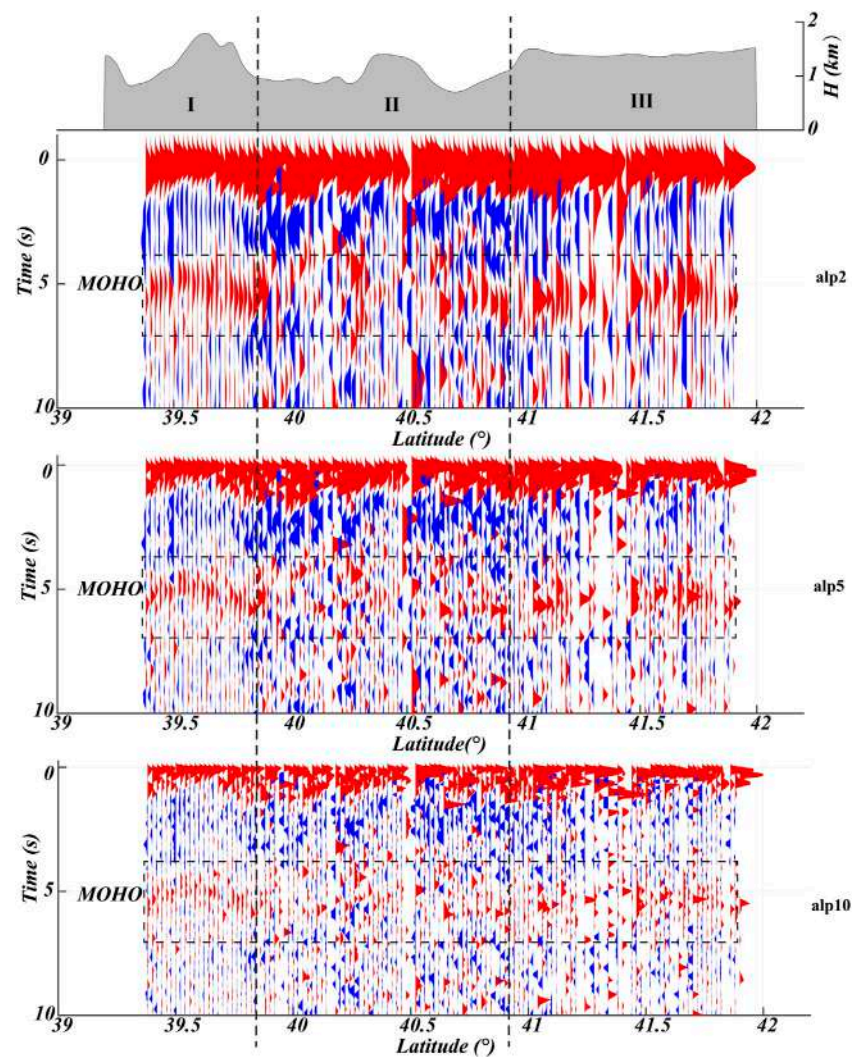


FIGURE 3

Stacking profile of receiver functions with different Gaussian filter factors (alp). Alp = 2 (1.18 s pulse width), 5 (0.75 s pulse width), and 10 (0.53 s pulse width) were used to compare the signal stability.

margin was revealed by deploying a high-density seismic array across the boundary between the Yinshan block and the Ordos block (the Khondalite belt) (Wan et al., 2020).

2.2 High dense seismic array and data source

From August 2020 to September 2020, we deployed approximately 230-km-long seismic profiles with 140 short-period seismometers, crossing the ancient tectonic boundary and rift system from south to north in our study area (Figure 1 b and c). EPS seismometers with a flat velocity response ranging from 0.2 to 200 Hz were kept nearly 1–3 km apart, and the

observations were performed continuously for nearly one month with a sampling rate of 100 samples/s. The teleseismic waveform records of $M_s > 5.5$ with distinct P-wave signal onsets were carefully identified within an epicentral distance range of 30° – 95° (Figure 1D). The earthquake catalog was obtained from the global United States Geological Survey (USGS) (http://neic.usgs.gov/neis/epic/epic_global.html), and the event waveforms with obvious P-wave onset signals are shown in Figure 2.

2.3 Data processing procedure

Each seismic event's theoretical p-wave arrival time was calculated based on the IASP91 model (Kennett and Engdahl,

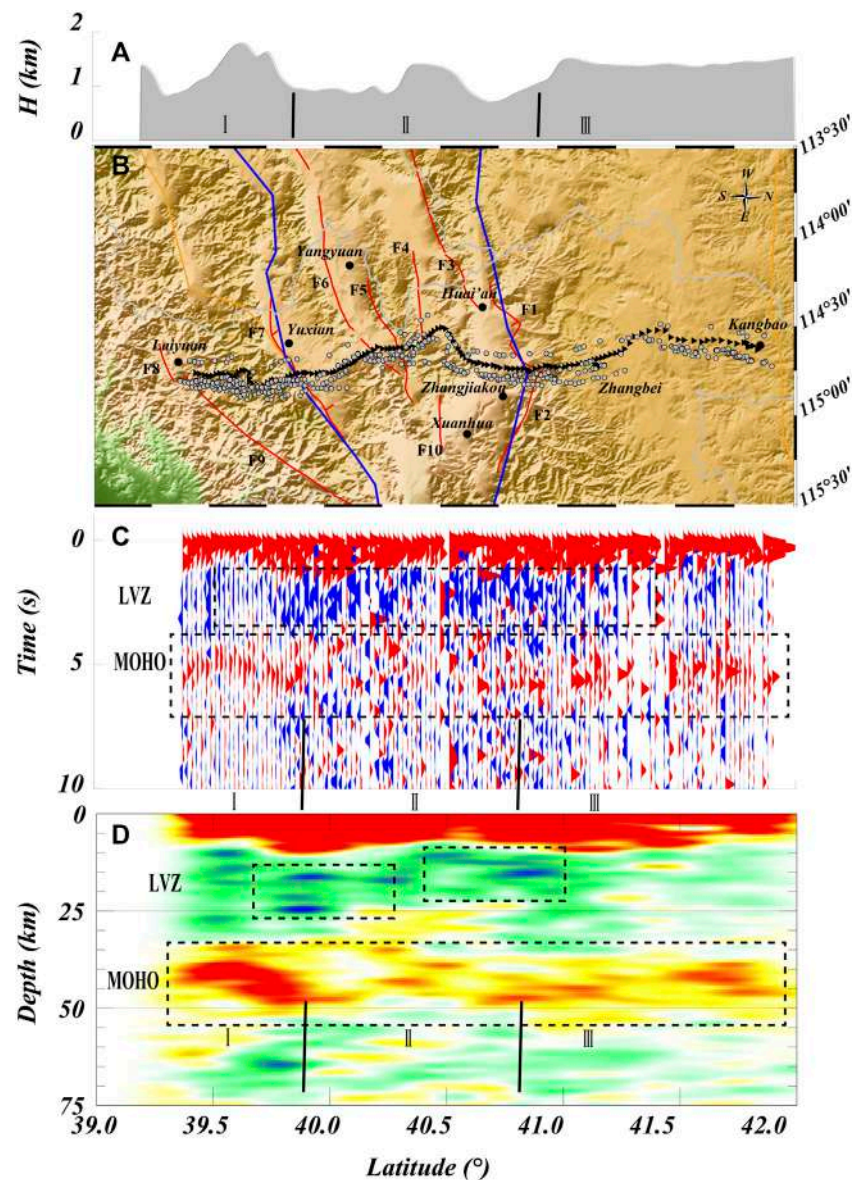


FIGURE 4

Topography, receiver function, and migration image. (A) Topographic elevation of the survey line, (B) topography and geomorphology of the study area and the distribution plan of the station, and (C) profile under the survey line superimposed by the receiving function. (D) CCP result. F1-Huai'an–Wanquan Basin northern margin fault; F2-Zhangjiakou fault; F3-Tianzhen–Yanggao basin northern margin fault; F4-Huai'an Basin southern margin fault; F5-Yangyuan basin northern margin fault; F6-Liulengshan basin northern margin fault; F7-Yuguang basin southern margin fault; F9-Sunzhuangzi–Wulonggou fault; F10-Xuanhua Basin southern margin fault.

1991). The seismic data between 20 s before and 80 s after the P-wave were collected after calibrating the P-wave arrival. Furthermore, the mean, the linear trend, and other filtering processes were removed to check the waveform quality. The original ENZ component data were rotated to the RTZ component per the circle path. The waveforms with relatively clear recorded events were deconvoluted with different Gaussian filters to obtain the receiver functions and subsequently selected

the clear receiver functions of the PmS seismic phase (Ligorria and Ammon, 1999). A Gaussian filter of 5 with an approximate pulse width of 0.75 s was used to maintain the detailed information in receiver function traces. Also, the receiver functions with a Gaussian filter of 2 (1.18 s pulse width), 5 (0.75 s pulse width), and 10 (0.53 s pulse width) were calculated to test the stability of signal characteristics (Figure 3). The crustal structure under the station was

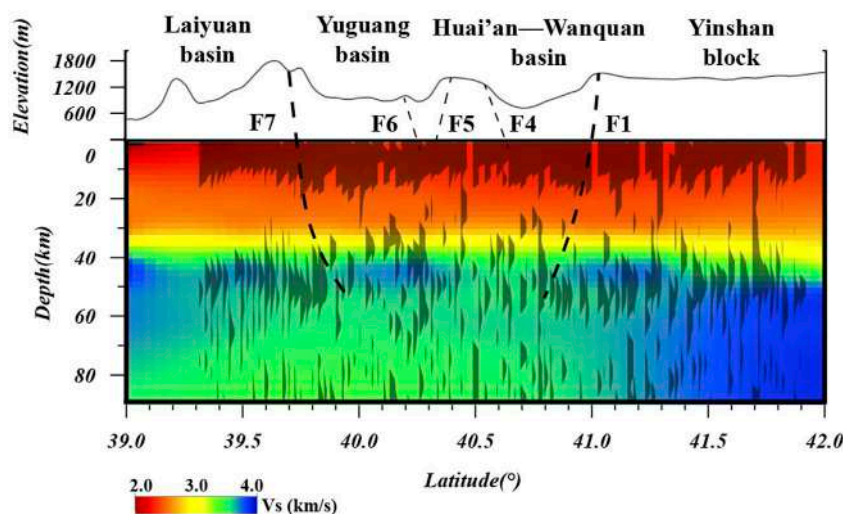


FIGURE 5
Tomography profile and receiver functions. The tomography results are combined with the results reported by Tao et al. (2018).

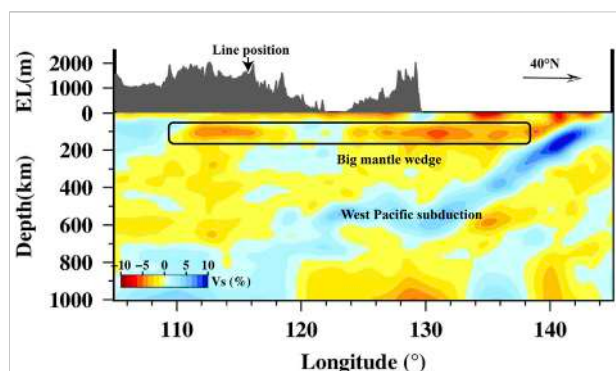


FIGURE 6
Tomographic profile of the Pacific subduction with latitude = 40°N (according to Tao et al., 2018)

obtained by stacking in accordance with the transition point of projection and subsequently selected 560 high-quality receiver functions. The influence of other interfaces at different epicentral distances was reduced by correcting the time of receiver functions with P-to-S moveout correction by setting a reference distance of 65° and a focal length of 20 km. Thus, the receiver functions may be configured to display the undulations of the Moho or the interfacial features of the crust. The profiles of stacking receiver functions after P-to-S moveout correction during different periods along the longitude of PmS pierce points, calculated based on the IASP91 model, are shown in Figure 3.

Furthermore, the receiver functions on the temporal domain were converted into the spatial domain along the

ray path and P-to-S conversion arrivals by the common conversion point (CCP) stacking method (Yuan et al., 1997). For migration purposes, the crust and upper mantle within 75 km depth were divided into 0.5 km segments in both the horizontal and vertical directions, and subsequently, the quality of migration images was improved by staking the receiver functions with the pierce points along the ray path using the IASP91 model to generate the final migration images ultimately. The migration images obtained along the profiles from receiver functions with the Gaussian filter of 5 are shown in Figure 4D.

3 Results

The reliable signals in the receiver functions are stable with variations in time and αp . A smaller αp factor provides the Moho basic and smoothed interface data, while a larger αp factor provides more crustal information (Wei et al., 2016). Even though the signal of the Moho is about 5 s (depth of approximately 40 km), it has a distinct segmented pattern from south to north. The profile is subdivided into three characteristic segments based on the shape of the Moho signal: (I) 39°N–39.8°N (south of F7), where the Moho is clear and continuous in the staking profile, gradually descending from south to north; (II) 39.8°N–40.7°N (between F7 and F1), where the morphology is complex with weak continuity, exhibiting no single layer; and (III) 40.7°N–42°N (north of F1), where the Moho is relatively stable and flat. In addition, some relatively stable negative signals shallower than 25 km depth observed in the range of

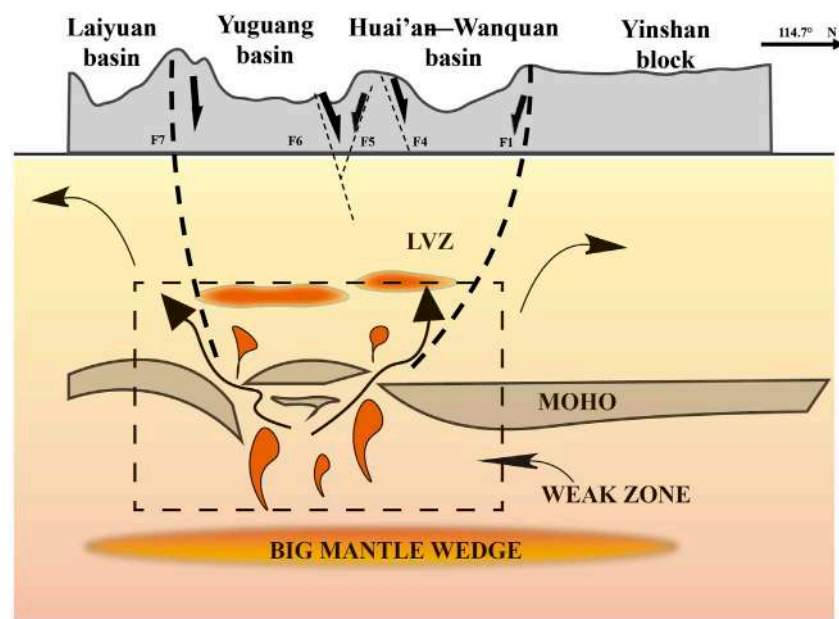


FIGURE 7

Sketched model of the interaction between the mantle and crust.

36.7°N–41.5°N may represent a low-velocity zone (LVZ) within the crust. A comparative analysis of topographic elevation, surface structure, receiver function results, and CCP results are shown in Figure 4. Based on the CCP results, the Moho exhibits a uniform pattern. However, the Moho in region I is curved downward, while that in region II is relatively complex and that in region III is relatively flat. The LVZ is mainly concentrated in region II and is discontinuous. The complex features of the Moho represent a series of surfaces closely related to structural features and deep structures and are visible below the densely constructed basin ridges.

4 Discussion

NCC is regarded as the most severely damaged craton on earth (Carlson et al., 2005; Zhu et al., 2011). The tectonic belts around cratons are relatively mechanically weak zones compared to the cratons and are primary sites of intense heating and strain localization (Hu et al., 2018; Liu et al., 2021). Based on the imaging results, the average depth of the Moho beneath the profile is 40 km. A comparative study was carried out on the basic crustal conditions under the adjacent regions in North China by the receiver function method and by using the broadband seismograph data (Zheng et al., 2007; Chen, 2009; Tang et al., 2010; Lou et al., 2017; Si et al., 2017; Wei et al., 2020; Zuo et al., 2020), which are comparable with crustal

thickness obtained by the measured receiver functions. The imaging results of the Moho obtained from region I and region III are comparable to the results from the western Khondalite belt (Wan et al., 2020). Moreover, the results of wide-angle reflection and refraction profile data of deep seismic sounding indicate the existence of a deep crust-scale fault zone extending to the Moho beneath the Zhangjiakou area, as well as the presence of LVZ within the crust (Zhu et al., 1999), which may suggest the presence of intense magmatic activity.

Numerical experiments demonstrate that inheritance is the primary controlling factor of lithospheric rifting and layer stiffness. The rift styles mostly depend on the inherited weaknesses embedded in the model (Chenin and Beaumont, 2013). The study area crosses through Yuguang and Wanquan basins from south to north and crosses a series of Holocene normal faults, making the area active. The normal faults are distributed in the east of the western parts, and their characteristic stress conditions are generally in a north–south extensional environment. On the other hand, the combined analysis of previous tomographic results using the receiver function shows that the Moho in region II is discontinuous (Figure 5). Also, the complex deep structure in region II corresponds rather well with the distribution of the faults and basins, implying that the surface topography variation is generally congruent with the deep dynamic processes. Faults 1 and 7 seem to represent two boundaries of the whole fault and basin system, governed by the migration

of deep material on both sides. Frequently, the axis of lithosphere extension and continental breakup is parallel to the orogenic belts and suture zones (van Wijk, 2005). This area is probably related to the weak zone formed by the collage of Yinshan and Ordos blocks, which is evidenced by the combined analysis of the area with the E-W trend of the surface structure.

F1-Huai'an-Wanquan Basin northern margin fault; F4-Huai'an Basin southern margin fault; F5-Yangyuan basin northern margin fault; F6-Liulengshan basin northern margin fault; F7-Yuguang basin southern margin fault.

Based on the tomography results, subduction of the western Pacific has extended westward into the central and WNCC toward the Datong volcano (Wei et al., 2012; Liu Z et al., 2017; Zhao, 2017; Lei et al., 2018), with a large mantle wedge in the middle and upper mantle. The depth profile of the S-wave perturbation velocity along the latitude of 40°N is illustrated in Figure 6 based on the topography results reported by Tao et al. (2018).

The Pacific plate is deeply subducted toward the west, forming a large mantle wedge, and its influence extends westward, up to a longitude of 110°E. The uneven mantle flow in the upper mantle caused by the subducting slab front poses strong action on the upper lithosphere, thereby influencing the tectonic evolution of the crustal and upper mantle structure beneath NC (Tian et al., 2019). The high calorific value in North China at a depth of ~100 km is also evidenced by the geothermal results reported by Wang and Cheng, (2012). The low-velocity anomalies in the crust and upper mantle below the Zhangjiakou area are shown by the local ambient noise (Jiang et al., 2021). The aforementioned findings show that certain Cenozoic tectonic deformation and fault development in the study area are strongly connected to the internal restructuring of the craton caused by the subduction of the western Pacific plate. The area is tectonically characterized by the western subduction of the Pacific plate and the distinct local restructuring evidenced by the formation of a large mantle wedge. The tomography images from ambient noise also reveal that the crust and upper mantle beneath the Shanxi fault depression belt, Zhangjiakou, and Datong volcanic areas are characterized by low-velocity anomalies associated with a large area of intense Cenozoic magmatic activity in Datong triggered by the upwelling of hot materials from the mantle (Zuo et al., 2020), and the slow movement of the magma surface causing negative signals.

LVZ-low-velocity zone; F1-Huai'an-Wanquan Basin northern margin fault; F4-Huai'an Basin southern margin fault; F5-Yangyuan basin northern margin fault; F6-Liulengshan basin northern margin fault; F7-Yuguang basin southern margin fault.

The results of the receiver function and tomographic imaging were combined to generate a conceptual diagram of the tectonic process of this area, as shown in Figure 7. Since region II in Figure 3 and Figure 4 was relatively weak in the ancient craton collage belt earlier, it is more sensitive to get impacted by the mantle wedge (Zuo et al., 2020). These weak zones are reactivated by the mantle plume resulting in the upwelling of the Moho as channels, which ultimately restructures the crust. The shallow crust's materials are pushed to move on both sides by the circulatory power of deep materials, resulting in a series of normal faults and grabens on the earth's surface. As a result, region I and region III may still retain the shape of the ancient Moho without substantial changes.

5 Conclusion

In the present study, a short-period dense array is used to generate the receiver function image of two ancient tectonic belt boundaries of NCC. The extensive continental restructuring of the area might have affected the structure and properties of the continental lithosphere and impacted both surface processes and mantle dynamics regionally. For example, the series of nearly parallel E-W basins and lithospheric scale normal faults in the study area demonstrate their relationship with the pre-existing weak zone apart from the signatures representing current tectonic processes. The present study proposed a model illustrating the damage of the ancient weak zone caused by the dynamic processes of western Pacific subduction. The mantle wedge activates the local thermal circulation in the lithosphere to reactivate the weak zone to enter the crust as a channel which forms an extensional environment. This is evidenced by the slowly cooling magma upwelling of LVZ at the bottom of the upper crust in Zhangbei-Huai'an and Yangyuan basins.

Data availability statement

The original contributions presented in the study are included in the article/Supplementary Material; further inquiries can be directed to the corresponding author.

Author contributions

Conception and design of the study: QZ and XS; acquisition of data: QZ and XS; analysis and/or interpretation of the data: QZ, XS, HH, SC, and JZ; drafting the manuscript: QZ and XS; revising the manuscript critically for important intellectual

content: QZ.

Funding

Some figures are plotted using the Generic Mapping Tools (Wessel et al., 2019). This study was financially supported by the National Key Research and Development Program of China (2017YFC1500100), the National Natural Science Foundation of China (Grant Nos. 41874052 and 41730212), the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0701), the Guangdong Province Introduced Innovative R&D Team (2016ZT06N331 and 2017ZT07Z066), and the Guangdong Collaborative Innovation Center for Earthquake Prevention and Mitigation (2018B020207011).

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 14 July 2022

ACCEPTED 04 August 2022

PUBLISHED 25 August 2022

CITATION

Wang S, Shen X, Chevalier M-L,
Replumaz A, Zheng Y, Li H, Pan J, Li K
and Xu X (2022), Illite K-Ar and (U-Th)/
He low-temperature
thermochronology reveal onset timing
of Yadong-Gulu rift in southern
Tibetan Plateau.
Front. Earth Sci. 10:993796.
doi: 10.3389/feart.2022.993796

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Illite K-Ar and (U-Th)/He low-temperature thermochronology reveal onset timing of Yadong-Gulu rift in southern Tibetan Plateau

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Determining the timing of E-W extension across the NS-trending rifts in southern Tibet is key to test the mechanical models of the latest evolution in the collision between India and Asia. We focus on the southern half of the largest of the seven main rifts, the Yadong-Gulu rift (YGR), which, despite being the focus of numerous studies thanks to its easy access, still lacks direct time constraints. Using illite K-Ar ages of fault gouge from the active Yadong normal fault of the YGR, we directly constrain its onset timing at 9 ± 1 Ma. (U-Th)/He dating of the footwall leucogranite reveals a rapid exhumation of the southern YGR since ~ 9 Ma, attesting to its onset activity. Such timing is similar to that estimated for the northern half of the YGR at 8 ± 1 Ma, suggesting that the entire YGR formed at approximately the same time. Our synthesis of published initiation ages of the other main rifts in southern Tibet shows that they mostly fall between ~ 23 and 8 Ma, suggesting a clear spatial and temporal pattern of old initiation ages to the west and young to the east. In this case, the formation of rifts in southern Tibet is unlikely caused by slab tearing of the underthrusting Indian plate or orogenic collapse. Our study supports that E-W extension in Tibetan Plateau is triggered by a combination of eastward propagation of the Karakorum-Jiali fault zone and divergent thrusting along the curved Himalayan arc.

KEYWORDS

Southern Tibetan Plateau, illite K-Ar, low-temperature thermochronology, Yadong-Gulu rift, onset timing

Introduction

The Tibetan Plateau, resulting from the continent-continent collision of the Indian and Asian plates, shows an emblematic paradox that, despite ongoing ~N-S shortening between the Tibetan Plateau and the Indian plate, the latest deformation is dominated by ~E-W extension in southern Tibet (e.g., Tapponnier et al., 1981; Armijo et al., 1986; Coleman and Hodges 1995; Chevalier et al., 2020). Indeed, seven main, parallel ~NS-trending rifts, resulting from E-W extension, attest to the latest and strongest tectonic activity in the southern half of the plateau (Figures 1A,B). Two main models have been proposed to explain the rifts formation. On one hand, one type of model focuses on lithospheric dynamics, in which either orogenic collapse (Dewey, 1988), delamination of thickened mantle lithosphere (England and Houseman, 1989), eastward flow of low-viscosity lower crust (Royden et al., 1997; Yin and Taylor, 2011), or tearing of subducting Indian lithospheric slab (Chen et al., 2015; Bian et al., 2020) is regarded as the main mechanism. On the other hand, another type of model emphasizes the boundary forces and regional stress fields. In such case, the rifts open either as pull-apart basins at the end of strike-slip faults accommodating the eastward extrusion of Tibet (Tapponnier et al., 1982; Armijo et al., 1986), northward propagation of underthrusting Indian slab (DeCelles et al., 2002), inward bending of orogen rims (Klotwijk et al., 1985), or basal shear from oblique underthrusting of the Indian plate (McCaffery and Nabelek, 1998).

Although numerous studies have documented the initiation age of the south Tibetan rifts to test these models, the ages are spread out, ranging from ~23 to 2 Ma (Figure 1C; Supplementary Table S1, and references therein), partly due to the different dating methods used, thus resulting in a hot debate on the dynamic mechanisms responsible for the rifts' formation. For example, Bian et al. (2022) suggested that the Yadong-Gulu rift (YGR) initiated at ~13–11 Ma based on apatite fission track and (U-Th)/He dating. They further proposed that the onset timing of the YGR is younging northward, due to outward expansion of the Himalayan arc. Using (U-Th)/He thermochronology, Sundell et al. (2013) determined initiation ages for the South Lunggar rifts to be >10 Ma with a rifting acceleration at 5–2 Ma, which they explained was caused by a combination of gravitational orogenic collapse and underthrusting Indian lithosphere. Therefore, the exact determination of the rifts' onset timing and their spatio-temporal variations are of great importance for the analysis of the E-W extension dynamics in southern Tibetan Plateau.

Previous studies usually used low-temperature (U-Th)/He or fission-track thermochronology of pluton or basin sediments in the rifts to indirectly determine their initiation timing (e.g., Mahéo et al., 2007; Styron et al., 2013; Sundell et al., 2013; McCallister et al., 2014; Shen et al., 2016; Bian et al., 2020, 2022; Shen et al., 2022a). In recent years, thanks to the advances in low-temperature dating techniques, fault gouge dating has also

been used to offer direct information on the timing of faulting events (e.g., Van der Pluijm et al., 2001; Duvall et al., 2011; Zheng et al., 2016). In this paper, we use illite K-Ar dating of fault gouge, as well as zircon U-Pb, and zircon and apatite (U-Th)/He thermochronology of leucogranite from the footwall of the active Yadong normal fault to determine the onset timing of the southern YGR. We further reconstruct the thermal history and activity of the YGR and discuss the spatio-temporal pattern and dynamics of E-W extension in southern Tibet.

Geological setting

The YGR consists of several Neogene-Quaternary fill basins and NNE-NE-trending bounding normal faults. It is the longest (~500 km) and most prominent rift in southern Tibet, extending from Yadong in the Himalayas to north of Gulu in the Lhasa terrane. It cuts across the Himalayan terrane, South Tibetan Detachment System (STDS), Yarlung Zangbo suture, and Lhasa terrane (Figure 1B). It is bounded to the north and south by the Karakorum-Jiali fault zone (KJFZ) and Main Frontal Thrust (MFT), respectively (Figure 1A; e.g., Armijo et al., 1986; Chevalier et al., 2020; Wang et al., 2020). The YGR is also the most active rift in southern Tibet, with 14 $M \geq 6.0$ earthquakes since 1264 A.D. (Wu et al., 2011), as well as significant hydrothermal activity. In view of the geometry of the bounding faults, Armijo et al. (1986) divided the YGR into three sections: northern, central, and southern section, with a total of seven grabens or half-grabens. In the south, the Yadong rift runs from Yadong town to the Yarlung Zangbo suture (YZS), and is characterized by W-dipping normal faults along three right-stepping en-echelon grabens (Pagri, Nierudui, and Relong from south to north). In the center, from the YZS to Yangbajing town, the Angang and Yangbajing grabens are bounded by nearly symmetrical normal faults dipping to the west and east. By contrast, while the northern section from Yangbajing to the Gulu bend is bounded by faults with both left-lateral and normal components along the SE side of the Nyainqentanglha mountain range, only a large normal fault exists north of the Gulu bend.

Our study area is located in the ~130 km-long, ~20 km-wide Pagri graben, southernmost Yadong rift (Figure 2A). Boundary normal faults in this graben dip to the W-NW at an angle of 50°–70° (Figures 3H, 4H; Wang et al., 2020), cutting and offsetting the ~EW-trending South Tibetan Detachment System (STDS) (Wu et al., 1998; Dong et al., 2021). Field evidence showed rapid normal faulting along the Pagri graben in the Quaternary, with, for example, >2.6 km of topographic relief (Armijo et al., 1986), and $<17 \pm 1$ m offset since the global Last Glacial Maximum (~20 ka) (Wang et al., 2020; Chevalier et al., 2022). Additional evidence such as earthquake-triggered landslide at 13–15 ka (Peng et al., 2018), as well as fresh surface rupture and hot springs (Wang et al., 2020) are also present. The

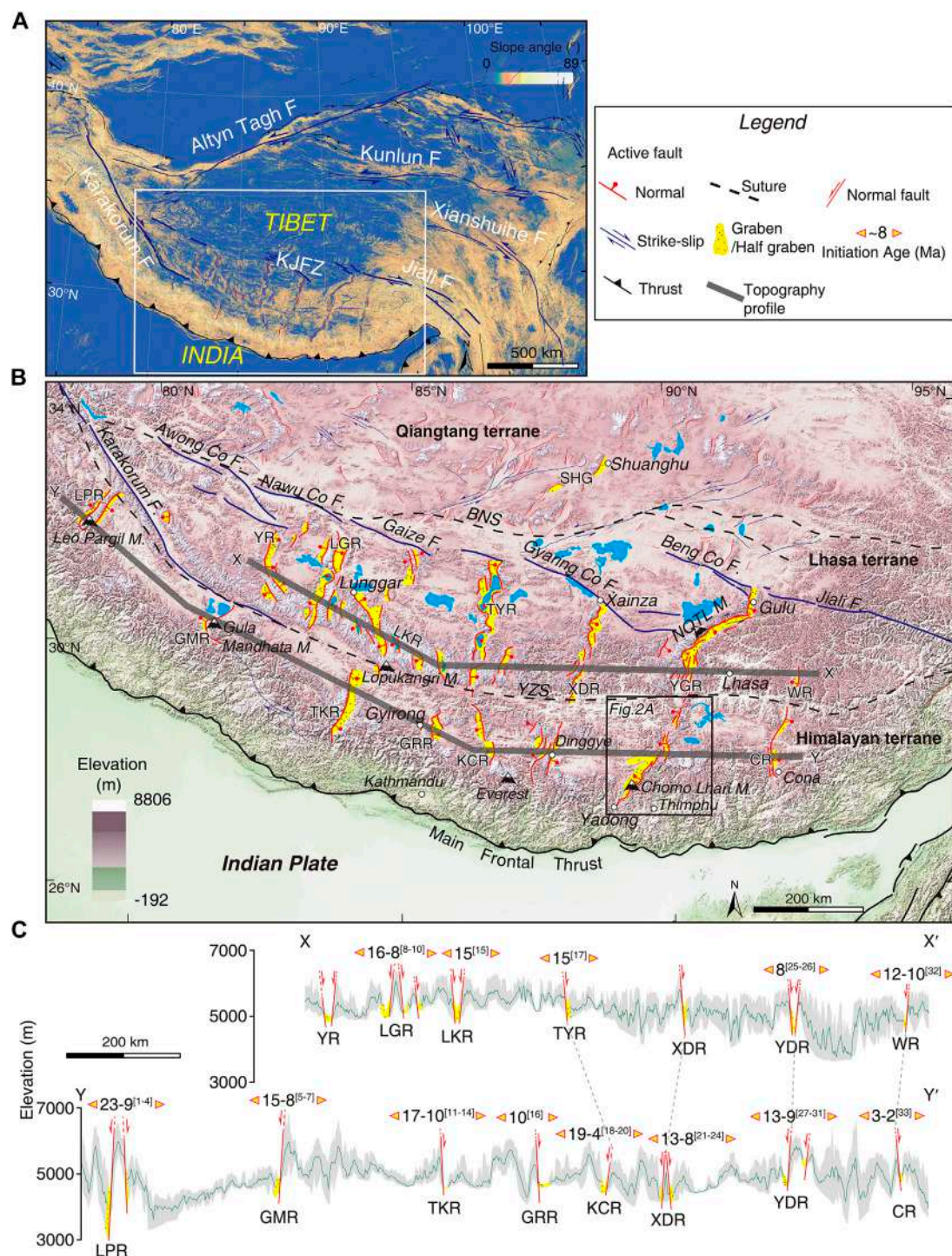


FIGURE 1

(A) Slope map and main active faults of Tibetan Plateau. (B) Active faults of Tibetan Plateau interior with main rifts, modified from [Tapponnier et al. \(2001\)](#) and [Chevalier et al. \(2020\)](#) (C) Topographic profiles across the south Tibetan rifts with initiation ages in Ma (references as in [Supplementary Table S1](#)). Main rifts and sutures are: CR, Cona rift; GMR, Gula Mandhata rift; GRR, Gyirong rift; KCR, Kung Co rift; LGR, Lunggar rift; LPR, Leo Pargil rift; SHG, Shuanghu graben; TKR, Thakkhola rift; TYR, Tangra Yumco rift; WR, Woka rift; XDR, Xainza-Dinggye rift; YGR, Yadong-Gulu rift; YR, Yari rift; NQTL, Nyainqentanglha; KJFZ, Karakoram-Jiali fault zone; BNS, Bangong-Nujiang suture; YZS, Yarlung Zangbo suture.

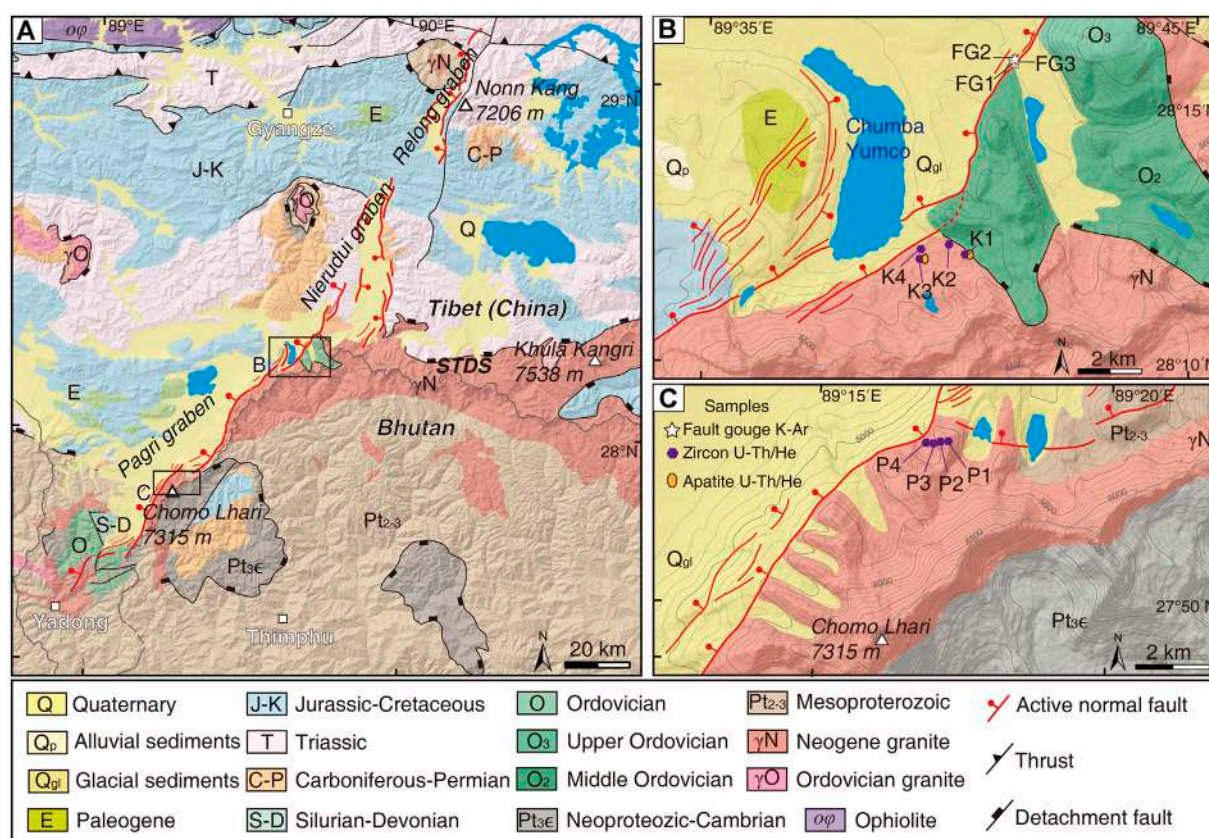


FIGURE 2

(A) Geological map of the southern YGR (location in Figure 1B) [revised from China University of Geosciences (Beijing) (CUGB), 2004]. (B) Geological map of Chumba Yumco area with sample locations (FG1–FG3, K1–K4) (location in A). (C) Geological map of Chomo Lhari area with sample locations (P1–P4) (location in A). STDS–South Tibetan Detachment System.

hanging wall of the active Yadong normal fault consists of Holocene and Pleistocene conglomerate, as well as Paleogene conglomerate and sandstone. The footwall is composed of Cenozoic leucogranite, Paleozoic limestone, and intensely deformed Proterozoic metamorphic rocks such as gneiss and schist (Figure 2). Mesozoic strata including sandstone, limestone, and slate are underlying the Quaternary deposits to the north and overlie on top of Proterozoic metamorphic rocks to the south (Figure 2).

Field observation and sampling strategy

Cenozoic leucogranite is uplifted in the footwall of the YGR along the rugged NE-trending Chomo Lhari Range (Figure 2A). About 6 km north of Chomo Lhari peak (7,315 m), a series of triangular facets, with ~400–600 m relief (Figure 3A), lie at the foot of the mountain range. Active normal faults dipping to the NW at a high angle of 60°–70° (Figures 3C,H, 4H) separate the

leucogranite in the footwall from Pleistocene conglomerate in the hanging wall. We collected four leucogranite samples (P1–P4, Figures 2C, Figures 3A–G locations in Table 1) along a vertical transect on one of the triangular facets at altitudes ranging from 5,350 to 5,700 m (Figures 3D,E) for zircon U–Pb (P1) and (U–Th)/He dating (P1–P4). Just SE of Chumba Yumco Lake, the active, NW-dipping normal faults (at a high angle of 65°–75°) (Figure 4H) offset the STDS (Figure 2A), leucogranite, and Pleistocene–Holocene conglomerate. We also collected four similar samples (K1–K4; Figures 2B, Figures 4A–G, locations in Table 1) in the footwall's leucogranite for zircon U–Pb (sample K3) and (U–Th)/He dating between altitudes of 4,560 and 4,770 m (Figures 4B,C).

Northeast of Chumba Yumco, ~6 km north of samples K1–K4, active normal faults strike NE, dip at ~50° to the NW, and offset Holocene travertine deposits resulting in fresh ruptures considered as paleoearthquake relics (Zuo et al., 2021; Figures 5A–C). The footwall is made of Ordovician limestone where at least four different fault branches offset the Holocene travertine and conglomerate deposits made of

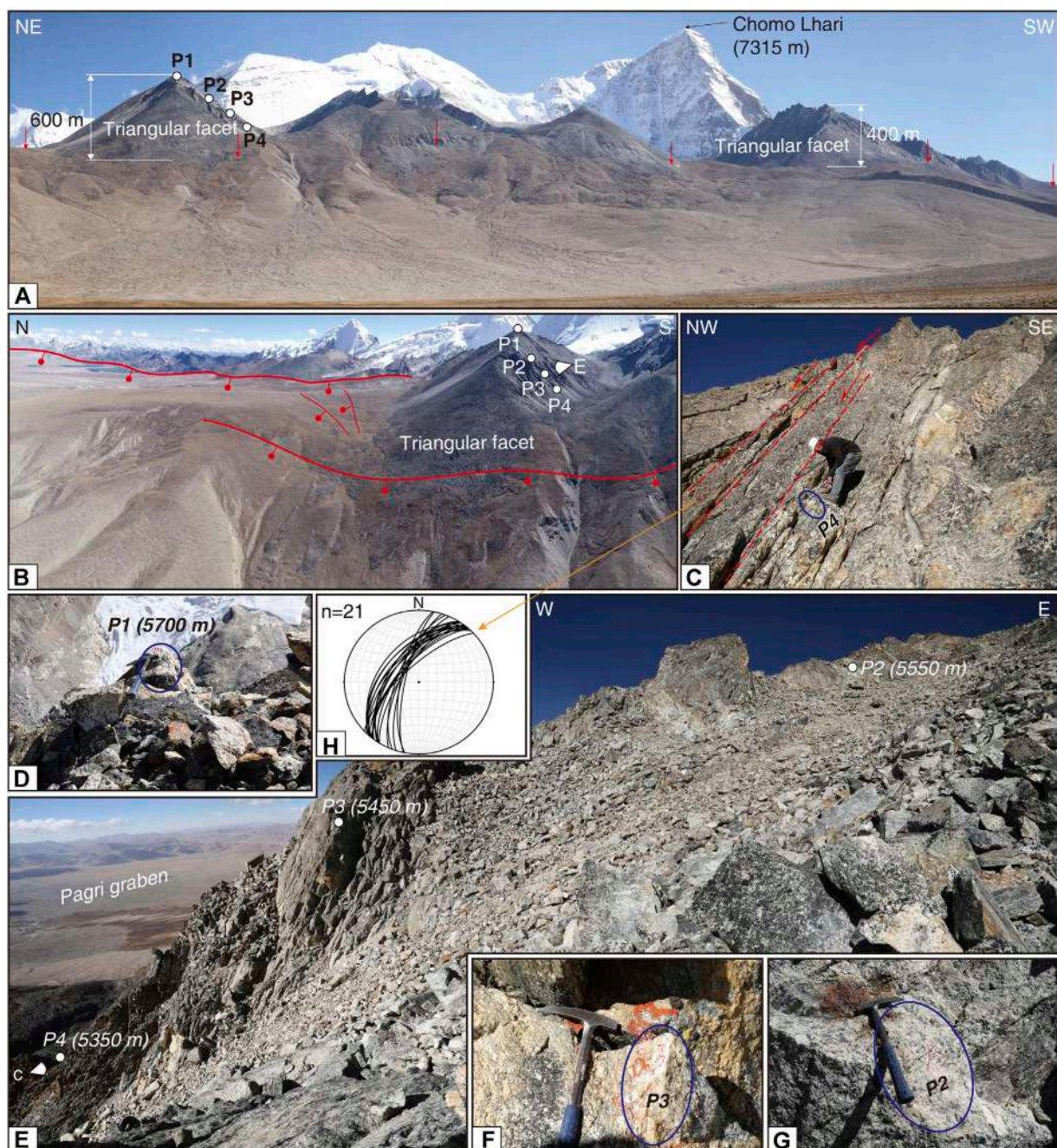


FIGURE 3

Field photos of active normal faults north of Chomo Lhari Mountain along southernmost YGR and location of samples P1–P4. (A) Triangular facets along YGR active normal faults and location of samples P1–P4. Red arrow refers to the active Yadong normal fault. Note Chomo Lhari peak (7,315 m) in the background. (B) Unmanned aerial vehicle (UAV) photo of samples P1–P4 and active normal faults. (C) Outcrop of sample P4 and brittle normal fault. (D–G) Outcrop and location with elevations of samples P1 (5,700 m), P2 (5,550 m), P3 (5,450 m), and P4 (5,350 m). (H) Foliations in the YGR normal fault north of Chomo Lhari Mountain, indicating the high-angle of active normal fault.

~50 m of cataclastic rocks and ~120 m of fault gouge/breccia zone (Figure 5A). Below the travertine, along the fault plane, the outcrop reveals two ~1 m-wide fresh black fault gouge layers (Figures 5B–F), which can be used to obtain direct constraints for

the onset timing of normal faulting (e.g., Duvall et al., 2011; Mottram et al., 2020). The black fault gouge is slightly foliated and half-consolidated with a few calcite veins. Within the strong foliated black fault gouge layers, we collected three clay-rich

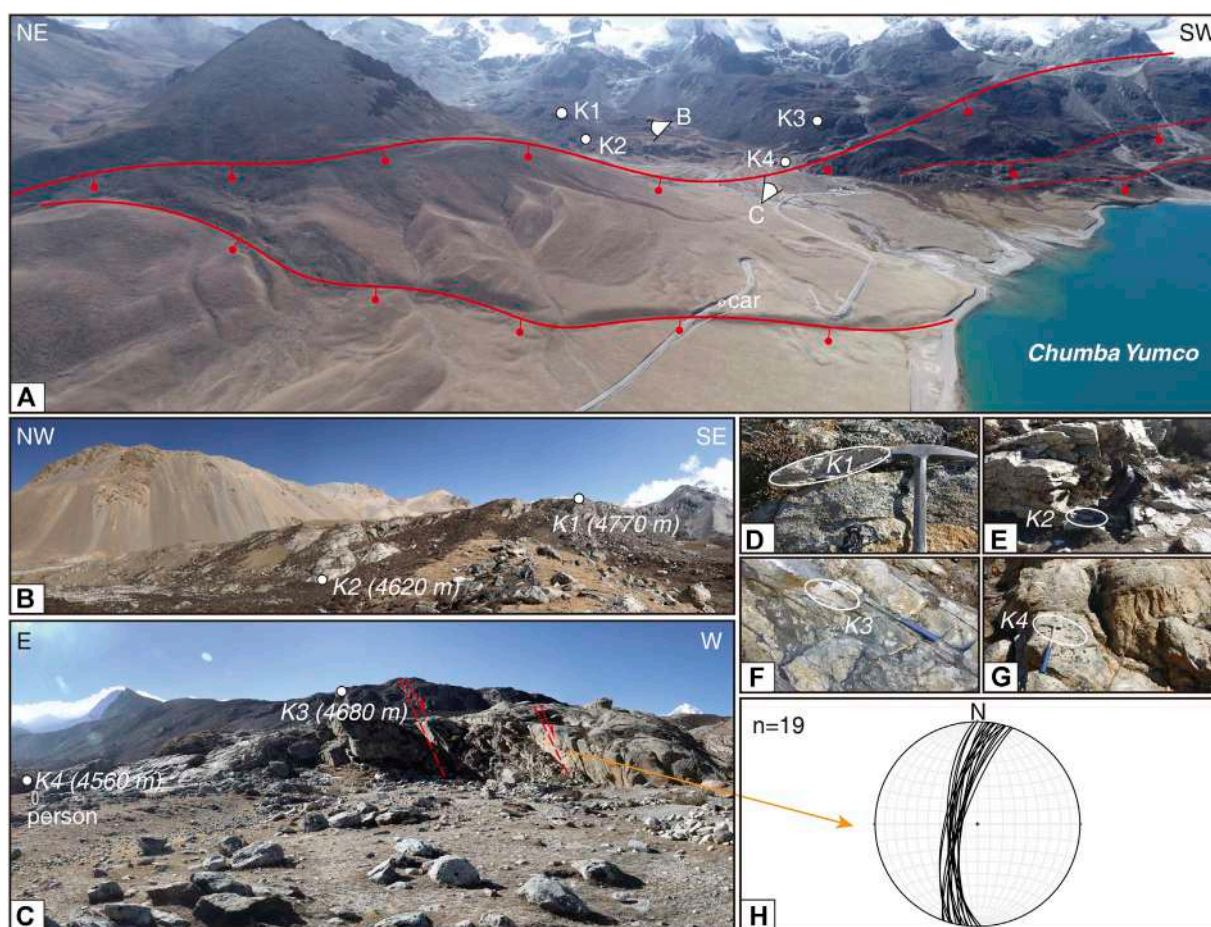


FIGURE 4

Field photos of active Yadong normal faults south of Chumba Yumco lake. (A) UAV photo of active faults south of the lake and location of samples K1–K4. (B,C) Location of samples K1 (4,770 m), K2 (4,620 m), K3 (4,680 m), and K4 (4,560 m). (D–G) Outcrops of samples K1–K4. (H) Foliations in the YGR normal fault south of Chumba Yumco lake, indicating the high-angle of active normal fault.

samples for K–Ar dating (FG1–FG3, Figures 5D,F), located next to a ~2 m-high waterfall, most likely resulting from the last large earthquake, which offset the active river (Figures 5A,B) (Wang et al., 2020).

Materials and methods

In order to reveal the exhumation history of the YGR, clay mineral illite K–Ar dating was used to date the fault gouge samples (FG1–FG3), i.e., to date the onset timing of the active, bounding normal faults of the southern YGR. Low-temperature zircon and apatite (U–Th)/He thermochronology, as well as high-temperature zircon U–Pb dating were also applied to the leucogranite samples (K1–K4 and P1–P4) to constrain the activity timing of the YGR.

Fault gouge authigenic illite K–Ar dating

$2M_1$ and $1M_d$ polytypes of illite usually coexist in fault gouge, which is generated in a shallow brittle fault zone in nature (Van der Pluijm et al., 2001; Duvall et al., 2011). The former $2M_1$ phase, which is a more stable phase that forms above 280°C (Velde, 1965), is usually derived from the surrounding wall rocks, whereas the latter $1M_d$ polytype is also referred to as the authigenic illite that usually grows below 200°C in fault gouge (Haines and Van der Pluijm, 2010; Duvall et al., 2011). Due to the ratio of fine size authigenic illite (<2 μm) and $2M_1$ polytype illite becoming higher in finer grain size, the authigenic illite age, representing the active timing of brittle faults, is normally determined by plotting K–Ar ages versus wall rocks-derived illite percentages in various sizes of isolating illite. The fitting line intercept with 0% and 100% of detrital ($2M_1$) illite end

TABLE 1 Single-grain zircon an apatite (U-Th)/He results from Leucogranite along southern section of YGR.

Sample	⁴ He	²³⁸ U	²³² Th	Raw age	Error	Rs	FT	Corrected	Error	Mean age	eU
No.	(mol)	(ppm)	(ppm)	(Ma)	(±1σ)	(μm)		age (Ma)	(±1σ)	(±1σ) (Ma)	(ppm)
<u>Zircon (U-Th)/He results</u>											
P1, 89.286366°E, 27.876033°N, 5,700 m											
Z1	6.01E-13	1,421.52	67.66	11.25	0.21	55.33	0.78	11.25	0.21	11.22 ± 1.42	1,437.4
Z2	1.14E-12	2,567.19	55.65	13.13	0.23	52.13	0.77	13.13	0.23		2,580.3
Z3	2.18E-13	1,112.59	96.36	9.77	0.18	45.84	0.74	9.77	0.18		1,135.2
Z4	6.04E-13	1,022.07	86.42	10.72	0.18	52.16	0.77	10.72	0.18		1,042.4
P2, 89.286629°E, 27.872261°N, 5,550 m											
Z1	3.17E-13	614.23	112.43	10.17	0.19	61.46	0.80	10.17	0.19	10.46 ± 1.04	640.7
Z2	7.90E-13	1,098.25	80.81	11.48	0.21	66.34	0.81	11.48	0.21		1,117.2
Z3	4.48E-13	676.56	427.91	9.13	0.22	65.21	0.80	9.13	0.22		777.1
Z4	3.28E-13	410.08	246.19	11.07	0.19	67.28	0.81	11.07	0.19		467.9
P3, 89.285261°E, 27.871767°N, 5,450 m											
Z1	3.12E-13	801.313	63.646	10.13	0.51	53.19	0.77	10.13	0.51	10.88 ± 0.65	816.3
Z2	9.77E-13	778.217	96.039	25.93	0.46	57.83	0.79	25.93	0.46		800.8
Z3	5.16E-13	830.741	121.41	11.31	0.20	59.01	0.79	11.31	0.20		859.3
Z4	9.37E-13	1,408.36	61.468	11.20	0.21	62.55	0.80	11.20	0.21		1,422.8
P4, 89.283822°E, 27.871983°N, 5,350 m											
Z1	1.45E-13	540.16	62.86	9.55	0.17	49.11	0.75	9.55	0.17	9.58 ± 0.05	554.9
Z2	1.17E-13	585.72	84.75	9.56	0.16	45.63	0.73	9.56	0.16		605.6
Z3	2.15E-13	786.54	56.77	9.54	0.17	50.24	0.76	9.54	0.17		799.9
Z4	1.50E-13	701.95	90.56	9.66	0.18	45.77	0.73	9.66	0.18		723.2
K1, 89.672762°E, 28.200880°N, 4,770 m											
Z1	1.49E-12	1,495.59	66.93	8.72	0.17	74.60	0.83	8.72	0.17	8.78 ± 1.54	1,511.3
Z2	6.73E-13	1,323.91	66.93	10.95	0.20	57.54	0.79	10.95	0.20		1,339.6
Z3	5.57E-13	1,535.88	73.87	8.04	0.14	54.48	0.78	8.04	0.14		1,553.2
Z4	4.25E-13	2,249.98	179.40	7.42	0.13	47.15	0.74	7.42	0.13		2,292.1
K2, 89.667226°E, 28.203987°N, 4,620 m											
Z1	1.61E-13	873.28	87.24	5.62	0.09	50.53	0.76	5.62	0.09	7.97 ± 1.61	893.8
Z2	4.28E-13	1767.31	242.05	8.80	0.16	46.82	0.74	8.80	0.16		1824.2
Z3	2.63E-13	965.07	137.37	8.32	0.16	51.63	0.76	8.32	0.16		997.3
Z4	3.94E-13	1906.24	315.52	9.16	0.18	46.44	0.74	9.16	0.18		1980.4
K3, 89.657206°E, 28.199202°N, 4,680 m											
Z1	1.23E-12	1743.07	107.02	10.07	0.21	66.09	0.81	10.07	0.21	8.61 ± 1.08	1,768.2
Z2	4.49E-13	1,072.65	92.66	8.61	0.15	57.26	0.79	8.61	0.15		1,094.4
Z3	1.45E-13	513.18	120.94	7.49	0.14	54.37	0.77	7.49	0.14		541.6
Z4	2.85E-13	734.91	175.30	8.29	0.16	53.34	0.77	8.29	0.16		776.1
K4, 89.657435°E, 28.202538°N, 4,560 m											
Z1	3.35E-13	1,033.40	322.28	8.09	0.14	52.18	0.76	8.09	0.14	7.8 ± 0.43	1,109.1
Z2	2.17E-13	922.67	229.08	7.74	0.14	51.58	0.76	7.74	0.14		976.5
Z3	1.75E-13	1,136.33	164.91	7.21	0.12	44.67	0.73	7.21	0.12		1,175.1
Z4	1.66E-13	1,091.44	297.83	8.14	0.14	44.38	0.72	8.14	0.14		1,161.4
<u>Apatite (U-Th)/He results</u>											
K1, 89.672762°E, 28.200880°N, 4,770 m											
A1	1.20E-11	1,515.86	56.06	5.1	0.11	47.36	0.62	8.22	0.18	8.71 ± 0.7	1,529.0

(Continued on following page)

TABLE 1 (Continued) Single-grain zircon and apatite (U-Th)/He results from Leucogranite along southern section of YGR.

Sample	⁴ He	²³⁸ U	²³² Th	Raw age	Error	Rs	FT	Corrected	Error	Mean age	eU
No.	(mol)	(ppm)	(ppm)	(Ma)	(±1σ)	(μm)		age (Ma)	(±1σ)	(±1σ) (Ma)	(ppm)
A2	2.61E-13	14.22	1.90	6.51	0.16	61.59	0.71	9.21	0.23		14.7
K3, 89.657206°E, 28.199202°N, 4,680 m											
A1	4.08E-13	53.62	5.01	5.13	0.12	46.61	0.61	8.36	0.20	8.53 ± 0.24	54.8
A2	7.60E-13	105.12	7.90	5.27	0.12	45.7	0.61	8.69	0.20		107.0

Rs, Radius of a sphere with the equivalent surface area-to-volume ratio as cylindrical crystals (Meesters and Dunai (2002)). Ft: α -ejection correction factor (Farley et al. (1996)). Mean age: Weighted means calculated using IsoplotR (Vermeesch (2018)). Clear older age outliers are highlighted in bold and are excluded from calculation of the weighted mean age. eU, Effective uranium content, [eU] = [U] + 0.235 × [Th] (Flowers et al. (2009)). Shaded data are outliers

members represents the fault-related authigenic illite age and wall rock-derived detrital illite age, respectively (Van der Pluijm et al., 2001; Yamasaki et al., 2013).

Illite isolation was carried out at the Institute of Geology, Chinese Academy of Geological Sciences, Beijing, following the detailed procedures of Zheng et al. (2014, 2016). Using gravitational floating in deionized water and ultra-centrifugation separation methods, fault gouge samples FG2 and FG3 were subdivided into four grain size fractions: 1–2, 0.5–1, 0.25–0.5, and 0.25 μm, while FG1 was subdivided into three size fractions (no 0.5–1 μm). These aliquots were characterized by X-ray diffraction (XRD) to determine the content of illites at the Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing. Quantification of the 2M₁ and 1M_d polytypes of illite was calculated by the BGMN software (Bergmann et al., 1998; Ufer et al., 2012). Then, each fraction was analyzed by K-Ar analytical method at the Analytical Laboratory of Beijing Research Institute of Uranium Geology, following the detailed methods of Zwingmann and Mancktelow (2004). K-Ar ages and errors <1% (2σ) were calculated based on the K and radiogenic ⁴⁰Ar concentrations with the usual decay constants (Steiger and Jäger, 1977).

Zircon U-Pb method

Zircon crystals were selected at the Langfang Regional Geological Survey, China, by combining magnetic and heavy liquid separation techniques from whole-rock samples. Under a binocular microscope, well-shaped zircons without inclusions, cracks and corrosion were handpicked. All zircons were photographed using MIRA 3 scanning electron microscope and cathodoluminescence (CL) (Supplementary Figure S1) to document their internal structures. Zircon U-Pb dating was obtained by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) with an Agilent 7500 quadrupole ICP-MS coupled with a resolution SE 193-nm excimer laser at the National Institute of Natural Hazards, Ministry of Emergency

Management of China (NINH-MEMC), following the detailed procedures described in Liu et al. (2008). Zircon ²⁹Si concentrations in NIST 610 were used for internal standardization and zircon 91500 was used for external standards. A 30 μm spot size was used in LA-ICP-MS analyses and the common lead correction used the method of Anderson (2002). Concordia ages and diagrams were determined using Isoplot R (Vermeesch, 2018). All calculated ages are quoted with a 1 sigma error.

Zircon and apatite (U-Th)/He dating

Zircon and apatite (U-Th)/He (ZHe and AHe) ages were measured at the NINH-MEMC, Beijing, following standard procedures (Farley and Stockli, 2002; Reiners et al., 2004; Shen et al., 2021; Shen et al. 2022b). After being separated by magnetic and standard heavy liquid mineral separation techniques, >70 μm diameter, euhedral ZHe and AHe grains without visible inclusions, were selected under a microscope for dating. Each grain dimension was measured from digital photographs for an α ejection correction, then individual grains were wrapped into a platinum capsule for thermal outgas. Subsequently, He extraction and measurement of ZHe and AHe were conducted by a PrismaPlus QME 220 quadrupole mass spectrometer. Determination of U and Th was carried out on a coupled Agilent 7500 quadrupole ICP-MS. The α ejection correction (Farley et al., 1996) was applied to each crystal to calculate and correct the (U-Th)/He ages. We analyzed four single-grain aliquots for ZHe dating and at least two grains for AHe dating for chosen samples.

Thermal history modeling for ZHe and AHe data

To model the thermal evolution of multiple samples, we used the program QTQt (Gallagher, 2012) to simulate the

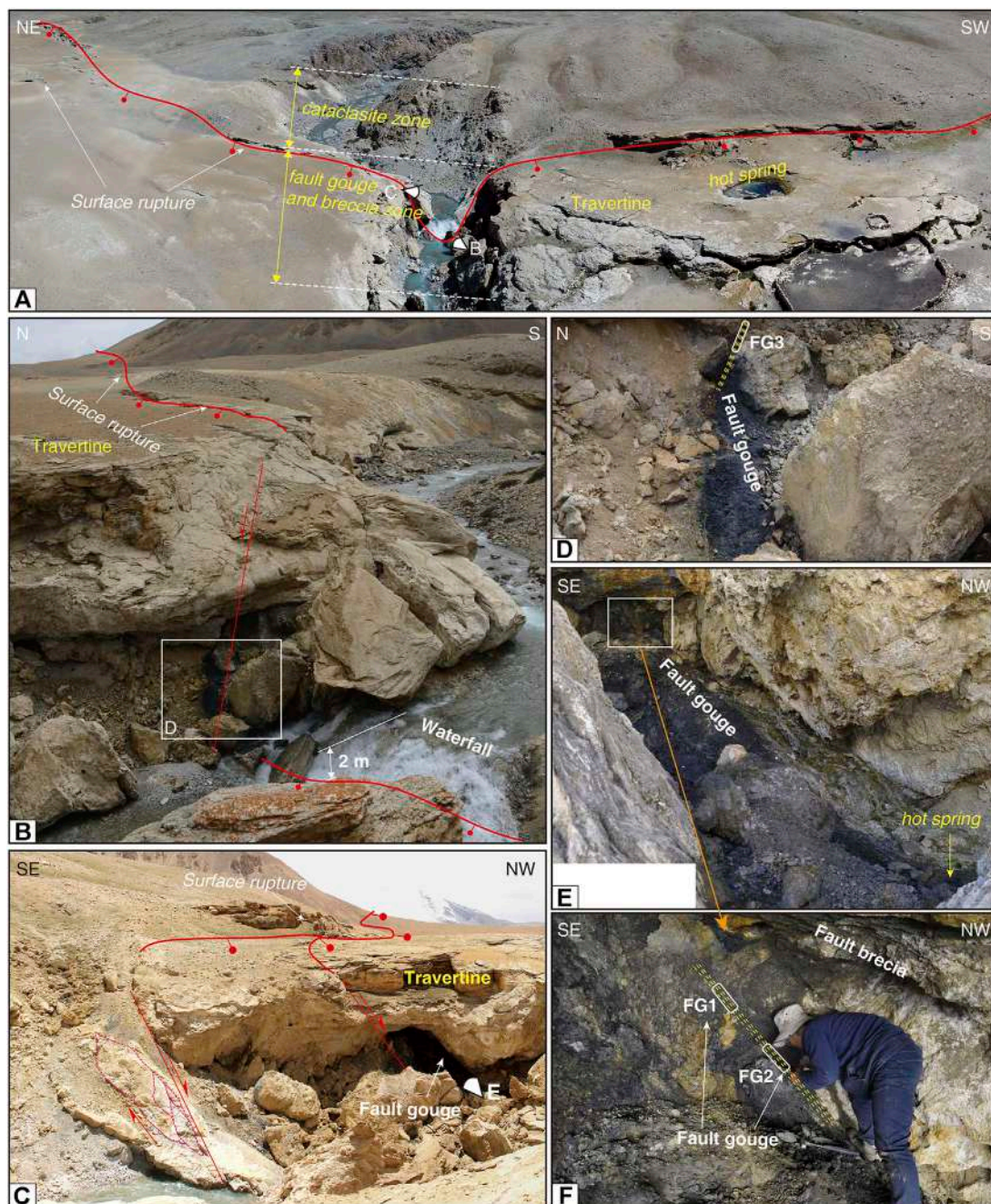


FIGURE 5

Field photos of active normal faults and location of fault gouge samples north of Chumba Yumco lake. (A) UAV photo of surface cracks in the travertine and fault rock distribution along active faults. (B,C) Exposure of fault gouge beneath the travertine as well as ~2 m-high waterfall due to normal throw. (D–F) Black fault gouge where three samples, FG1–FG3, were collected for authigenic illite K-Ar age analyses.

thermal history of footwall leucogranite. In the modeling, He diffusion parameters based on α radiation damage (Flower et al., 2009) and the annealing model of Ketcham et al. (2007) were applied for both ZHe and AHe systems. All samples follow the geological constraints: 1) mean temperature of $10 \pm 10^\circ\text{C}$ for the present-day earth surface; 2) initial time-

temperature at $\sim 27 \pm 3$ Ma (Zircon U-Pb age in our study) and $700 \pm 50^\circ\text{C}$ (Zircon U-Pb closure temperature, Copeland et al., 1990); 3) activity time-temperature at 9 ± 1 Ma and $200 \pm 50^\circ\text{C}$ of the fault gouge in our study; 4) geothermal gradient is set to $\sim 30^\circ\text{C}/\text{km}$ (Clark et al., 2005), and is allowed to vary over time.

TABLE 2 K–Ar ages for the different size fractions of illite isolated from the fault gouge samples.

Sample	Size	K cont.	$(^{40}\text{Ar}/^{38}\text{Ar})_{\text{m}}$	$(^{38}\text{Ar}/^{36}\text{Ar})_{\text{m}}$	Radiogenic	^{40}K	Age
No.	(μm)	(%)			^{40}Ar (mol/g)	(mol/g)	(Ma, 1σ)
FG1	1–2	5.87	7.94	310.03	6.87×10^{-10}	1.41×10^{-7}	82.2 ± 1.2
	0.25–0.5	5.65	3.70	412.58	4.35×10^{-10}	1.35×10^{-7}	54.4 ± 0.6
	< 0.25	2.24	3.36	298.57	2.20×10^{-10}	5.37×10^{-8}	69.1 ± 0.6
FG2	1–2	3.35	7.33	916.88	7.94×10^{-10}	8.03×10^{-8}	162.6 ± 1.7
	0.5–1	4.52	4.66	791.85	7.84×10^{-10}	1.08×10^{-7}	120.5 ± 1.3
	0.25–0.5	4.7	4.72	361.87	5.90×10^{-10}	1.13×10^{-7}	87.9 ± 0.9
	< 0.25	7.83	5.14	268.01	4.69×10^{-10}	1.88×10^{-7}	42.5 ± 0.6
FG3	1–2	6.15	6.51	404.05	7.15×10^{-10}	1.47×10^{-7}	81.6 ± 0.8
	0.5–1	5.7	4.30	363.84	5.74×10^{-10}	1.37×10^{-7}	70.9 ± 0.6
	0.25–0.5	6.24	3.78	553.42	4.99×10^{-10}	1.50×10^{-7}	56.5 ± 0.7
	< 0.25	5.34	3.12	281.03	2.61×10^{-10}	1.28×10^{-7}	34.8 ± 0.5

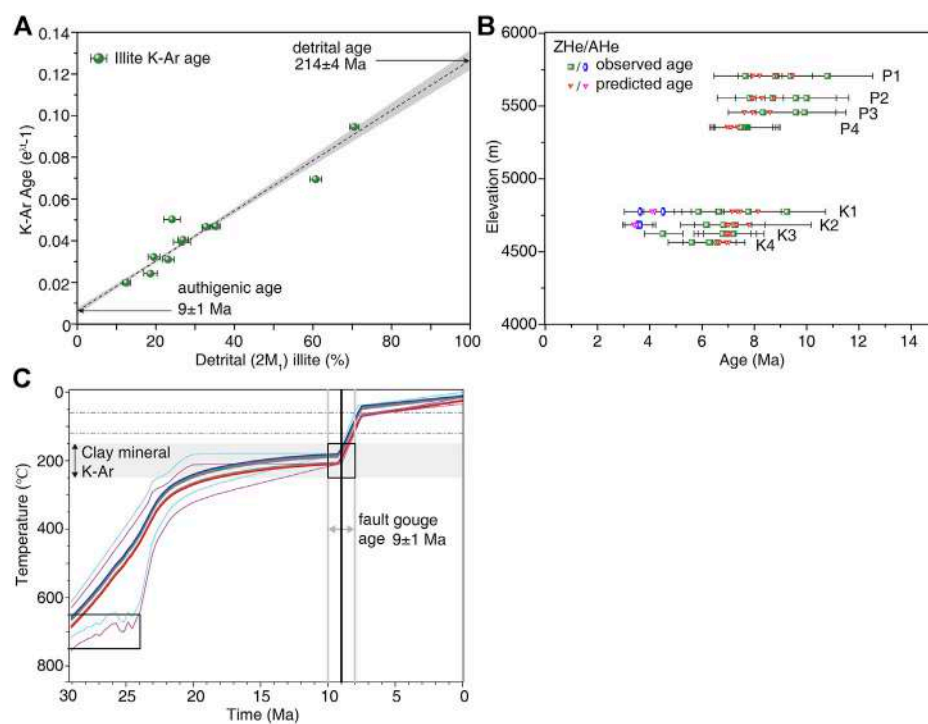


FIGURE 6

Zircon, apatite (U-Th)/He results and fault gouge authigenic illite ages. (A) K-Ar ages correlate with the percentage detrital illite component. Authigenic illite ages are late Miocene at 9 ± 1 Ma and detrital illite ages are late Triassic at 214 ± 4 Ma. (B) Age-elevation relationship for the northern Chomo Lhari Mountain and southern Chumba Yumco profiles, showing both observed and predicted single-grain ages modeled by QTQt. (C) Thermal modeling results of northern Chomo Lhari Mountain and southern Chumba Yumco leucogranite using QTQt (Gallagher (2012)) with fault gouge dating in the normal Yadong fault. The thick blue, grey, and red lines show the uppermost, intermediate, and lowermost samples, respectively, representing the 95% confidence intervals for the uppermost, intermediate, and lowermost thermal histories. The grey box shows the temperature range of K-Ar dating of Clay mineral in fault gouge (Grathoff and Moore (1996)).

Results

Fault gouge authigenic illite K-Ar ages

Eleven XRD results and K-Ar ages of different size fractions in three fault gouge samples were obtained. According to the XRD results, no calcite and K-feldspar was visible in the illite of various size fractions (1–2, 0.5–1, 0.25–0.5, and 0.25 μm) (Supplementary Figure S4), with mainly containing $2M_1$ and $1M_d$ polytype illites. Quantification of mineral compositions in the fractions is calculated by the BGMN software. The best-fitting mineral composition curves obtained from BGMN are consistent with the original plot of XRD results (Supplementary Figure S5). The detailed composition and content of illite aliquots are summarized in Supplementary Table S2. In principle, the percentage content of detrital illite ($2M_1$ polytype) decreases with the grain size (Supplementary Table S2).

The K-Ar ages for each illite fraction are listed in Table 2. Aliquot sizes decrease with decreasing $2M_1$ polytype illites, and ages also gradually become younger (Figure 6A). The fine, <0.25 μm fractions in the samples yield ages ranging between 34.8 ± 0.5 and 69.1 ± 0.6 Ma, and the coarse, 1–2 μm fractions yield apparent older ages varying from 81.6 ± 0.8 to 162.6 ± 1.7 Ma. We plot the detrital illite content of different fractions against their K-Ar ages ($e^{\lambda t}-1$) (Figure 6A) with linear fitting used by the York regression (York, 1968) to constrain the end-member ages. We obtain authigenic illite ages of 9 ± 1 Ma at the authigenic end-member, and of 214 ± 4 Ma at the detrital end-member (Figure 6A).

Zircon U–Pb isotopic ages

To determine the crystallization age of the footwall leucogranite, samples P1 and K3 were selected for zircon U–Pb dating. Measurement results are listed in Supplementary Table S3. Zircon grains are euhedral to subhedral in shape, exhibiting clear oscillatory zoning and rim-core domains pattern (Supplementary Figure S1). Both samples show similar characteristics of concordant ages. In sample P1, 119 spots were analyzed. On the concordia plot, the 13 crystal borders of neocrystallisation yielded a late Oligocene $^{206}\text{Pb}/^{238}\text{U}$ age of 28.80 ± 0.17 Ma. Other spots in crystal cores are concordant and reveal a Cambrian inherited $^{206}\text{Pb}/^{238}\text{U}$ age of 505.04 ± 0.76 Ma (Supplementary Figures S2A,B). Eighteen zircon grains in sample K1 are concordant and dated at 23.30 ± 0.23 Ma, i.e., late Oligocene–Early Miocene. Other analyzed spots in sample K1 have a Cambrian inheritance $^{206}\text{Pb}/^{238}\text{U}$ age of 507.79 ± 0.87 Ma (Supplementary Figures S2C,D). Few zircon grains implying a large inheritance from 200 to 500 Ma (Supplementary Figures S2, S3) probably correspond to the mixed value, due to the small grain size or narrow crystal rims.

Zircon and apatite (U–Th)/He ages and thermal modeling results

Four zircon grains were selected in each sample (P1–P4, K1–K4) for ZHe dating. Two apatite grains were collected in each of the samples K1 and K3 for AHe dating. All ZHe cooling ages from the eight samples range from 5.6 to 13.1 Ma, with most ages ranging between ca. 7 and 11 Ma, except one outlier replicate in sample P2 at 25.9 Ma (Table 1), suggesting a possible influence from U-rich neighbouring minerals or U-rich mineral inclusions in the analyzed grain (Shen et al., 2021). All AHe ages in the two samples fall between 8.2 and 9.2 Ma (Figure 6B; Table 1).

Modeling results suggest that the footwall leucogranite may have experienced two stages of rapid cooling since the late Oligocene (Figure 6C). The first stage started at ~ 28 Ma when the leucogranite formed, as constrained by the zircon U–Pb isotopic ages. During this stage, the leucogranite rapidly cooled at a rate of $\sim 50^\circ\text{C}/\text{Myr}$ between ~ 28 and 20 Ma, after which the leucogranite was in a relatively slow cooling period until ~ 9 Ma. Then, the second rapid cooling stage started at ~ 9 Ma and lasted until ~ 8 Ma with an exhumation rate of $\sim 110^\circ\text{C}/\text{Myr}$, derived from the closure temperature of ZHe ($180 \pm 20^\circ\text{C}$, Mahéo et al., 2007) to AHe ($70 \pm 15^\circ\text{C}$, Mahéo et al., 2007).

Discussion

Onset timing of the Yadong–Gulu at ~ 9 –8 Ma

Isotopic dating of authigenic illite in faults provides an important approach to directly determining the last primary episode of fault motion in the shallow crust (Zwingmann and Mancktelow, 2004; Van der Pluijm et al., 2006; Duvall et al., 2011; Yamasaki et al., 2013). Previous studies suggest that each well-defined fitting line in the percentage of detrital illite and the K-Ar age usually reveals one single fault event (Duvall et al., 2011; Mottram et al., 2020; Zhao et al., 2022). Linearly extrapolating to the 0% detrital illite of all different size fractions at 9 ± 1 Ma represents the integrated or mean age at which authigenic illite grew (Yamasaki et al., 2013). This average age, 9 ± 1 Ma, also reconciles the rapid exhumation stage determined by ZHe and AHe results (Figure 6C), which most likely indicates the onset timing of normal faulting of the southern YGR (Duvall et al., 2011). Clay minerals K-Ar in fault gouge samples are usually generated in thermal events at $200 \pm 50^\circ\text{C}$ (Grathoff and Moore, 1996). Authigenic illite, the $1M_d$ polytype, is thought to reflect the relatively lower temperature of thermal activity at ~ 200 – 150°C (Zwingmann and Mancktelow, 2004). Therefore, the formation depth of the YGR fault gouge can be calculated as 9–4 km (using a geothermal gradient of ~ 23 – $40^\circ\text{C}/\text{km}$, Francheteau et al., 1984), which is at a similar depth as that of the footwall leucogranite:

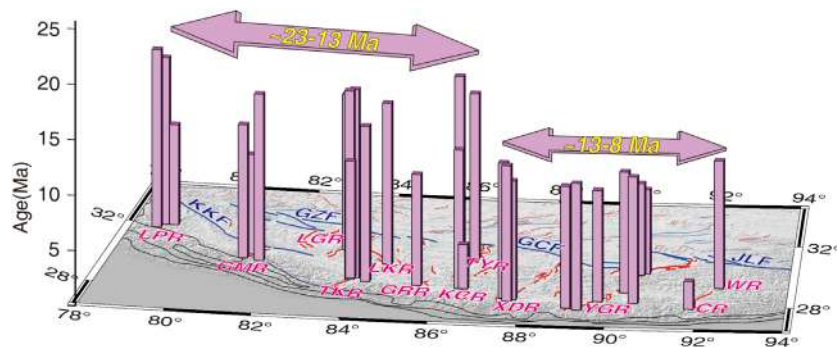


FIGURE 7

Distribution of onset timing for the main ~NS-trending rifts in southern Tibet. The majority of ages shows old initiation timing from ~23 to 13 Ma west of Xainza-Dinggye rift (XDR) and young onset timing from ~13 to 8 Ma east of XDR. CR, Cona rift; GMR, Gurla Mandhata rift; GRR, Gyirong rift; KCR, Kung Co rift; LGR, Lunggar rift; LPR, Leo Pargil rift; TKR, Thakkhola rift; WR, Woka rift; XDR, Xainza-Dinggye rift; YGR, Yadong-Gulu rift. GCF, Gyaring Co fault; GZF, Gaize fault; JLF, Jiali fault; KKF, Karakorum fault.

7.8–4.5 km at ~9 Ma. Contemporaneous rapid exhumation of the footwall leucogranite and the mean age of authigenic illite of fault with their consistent depth at ~9 Ma provides clear evidence that rapid exhumation directly results from the activity of the boundary normal fault of the southern YGR.

Near the Khula Kangri Mountain (Figure 2A), located ~90 km east of Chumba Yumco lake, Edwards and Harrison (1997) suggested that the detachment motion of the STDS continued to 12.5 ± 0.4 Ma, the date when monazite crystallized in the leucogranite. Therefore, these authors further inferred that the Yadong normal fault formed after the youngest crystallization age (~10 Ma), cutting across the STDS. By measuring U-Pb dating of granite cut by the normal fault bounding the eastern part of the southernmost YGR, Ratschbacher et al. (2011) obtained a maximum onset timing of 11.5 ± 0.4 Ma. Both results suggest an upper onset timing limit of the southernmost YGR. Thus, the authigenic illite with 9 ± 1 Ma K-Ar age probably derives from the hydrothermal effect created by the initiation timing of the YGR, following a rapid and strong activity of the boundary fault. Considering the initiation age of 8 ± 1 Ma in the northern YGR along the Nyainqentanglha mountain (Figure 1B; Harrison et al., 1995; Kapp et al., 2005), we propose that E-W extension along the different sections of the YGR occurred contemporaneously at ~9–8 Ma.

Recently, Bian et al. (2022) proposed that the southern Yadong-Gulu rift initiated at ~13–11 Ma in the south, using (U-Th)/He and fission track low-temperature thermochronology, and later northwards, at ~8 Ma (Harrison et al., 1995). They argued that such spatial and temporal pattern was influenced by the outward expansion of the Himalayan arc. However, since most of their samples are located quite far, ~20 km for some samples, from the main active Yadong normal fault of the southern YGR, and closer to the active thrusts along the Himalaya (such as the Main Boundary Thrust, MBT), their uplift may instead be influenced more by

the MBT or regional uplift of Himalayan orogenic belt. Our fault gouge and leucogranite samples were taken in or near the fault plane, therefore, we argue that our new illite K-Ar age in fault gouge and (U-Th)/He dating of footwall leucogranite provide direct evidence of the fault activity and represent the onset timing of the YGR. Thus, the 9 ± 1 Ma onset timing in the southern YGR (this study) and 8 ± 1 Ma (Harrison et al., 1995) in the northern YGR, suggest that E-W extension along the YGR occurred contemporaneously along its ~500 km length from south to north.

Cooling history of the Yadong-Gulu footwall leucogranite

Although the two elevation transects, P1–P4 and K1–K4, are located in different leucogranite bodies, the two transect samples yielded similar zircon U-Pb ages (Supplementary Figures S2, S3). Thus, samples P1–P4 and K1–K4 probably experienced a uniform cooling history in the Pagri graben. For samples P1 and K3, both zircon U-Pb dating show similar rim ages ranging between 28.8 ± 0.17 and 23.3 ± 0.23 Ma. These two late Oligocene ages suggest that the leucogranite crystallization formed due to the motion of the STDS, which primarily occurred at ~23 Ma and ended before ~13 Ma (e.g., Edwards and Harrison, 1997; Grujic et al., 2002; Leloup et al., 2010; Cooper et al., 2015). Combined with the ~22–14 Ma biotite ^{40}Ar – ^{39}Ar ages in the leucogranite north of Chomo Lhari mountain (Regional, 2004), we constrain the first rapid cooling of the footwall leucogranite at $\sim 50 \pm 20^\circ\text{C}/\text{Myr}$ from the $700 \pm 50^\circ\text{C}$ zircon U-Pb closure temperature (Copeland et al., 1990) to $300 \pm 30^\circ\text{C}$ for Ar in biotite (Mahéo et al., 2007).

The age-elevation section and the modeling results of ZHe and AHe imply that the second rapid cooling of the

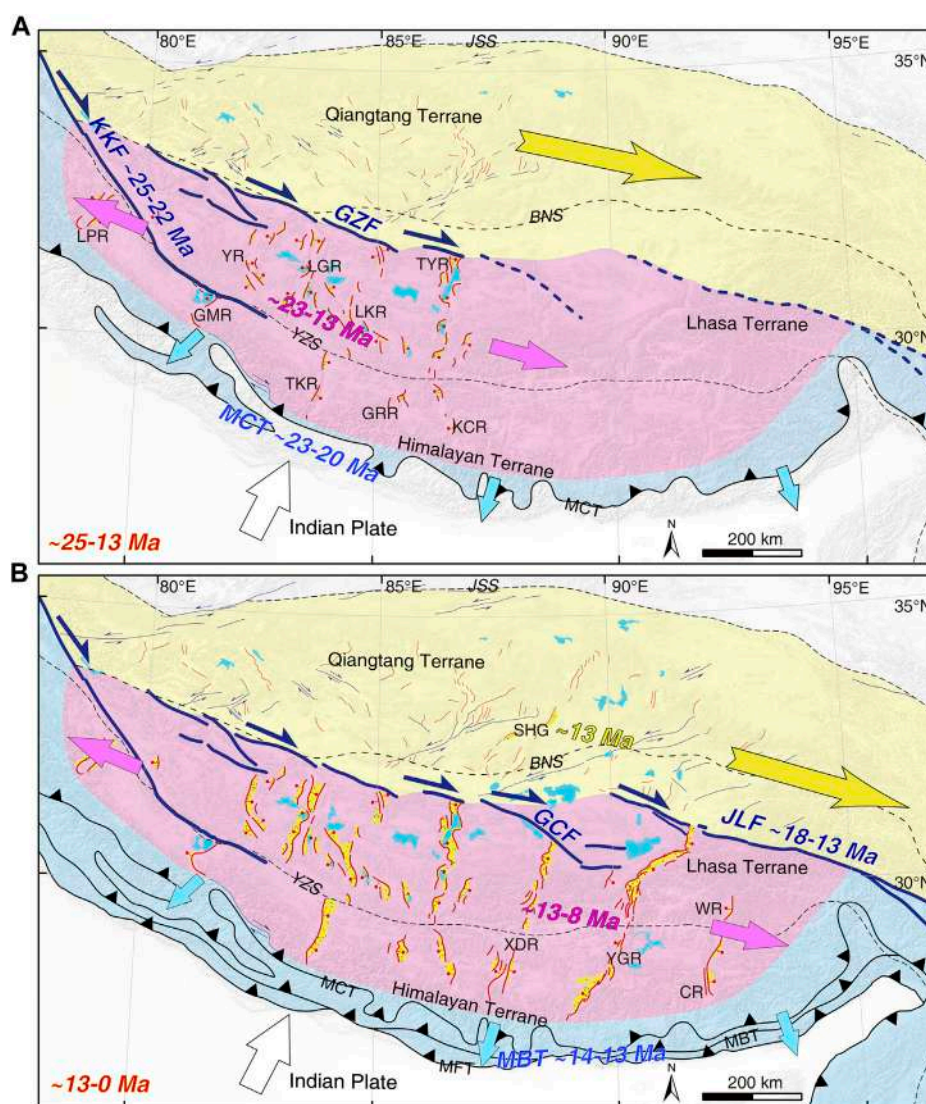


FIGURE 8

E-W extension kinematic model of southern Tibet. (A) Formation of rifts in the western half of southern Tibet are controlled by the Karakoram-Gaize fault (~25–22 Ma) and Main Central Tibet (~23–20 Ma). (B) Rifts in the eastern half of southern Tibet formed by eastward propagation of Gyaring Co–Jiali fault (~18–13 Ma) and Main Boundary fault (~14–13 Ma) along the curved Himalayan arc. Age references are listed in [Supplementary Tables S1, S4](#). Yellow area and arrow represent the eastward extrusion of Qiangtang terrane. Pink area and arrows represent the E-W extension of arcuate southern Tibet. Blue area and arrows represent the divergent orthogonal thrusting along the curved Himalayan arc. Main rifts and sutures are: CR, Cona rift; GMR, Gurla Mandhata rift; GRR, Gyirong rift; KCR, Kung Co rift; LGR, Lunggar rift; LPR, Leo Pargil rift; SHG, Shuanghu graben; TKR, Thakkhola rift; TYR, Tangra Yumco rift; WR, Woka rift; XDR, Xainza-Dinggye rift; YGR, Yadong-Gulu rift; YR, Yari rift; GCF, Gyaring Co fault; GZF, Gaize fault; JLF, Jiali fault; KKF, Karakoram fault; MBT, Main boundary fault; MCT, Main central fault; MFT, Main frontal thrust; NQTL, Nyainqentanglha; BNS, Bangong-Nujiang suture; JSS, Jinsha suture; YZS, Yarlung Zangbo suture.

leucogranite especially occurred around ~9–8 Ma (Figures 6B,C). ZHe mean ages of samples K1 and K3 are 8.8 ± 1.5 and 8.6 ± 1.1 Ma, respectively, and AHe ages are 8.7 ± 0.4 and 8.5 ± 0.2 Ma (Figure 6B; Table 1). In the modeling results (Figure 6C), the leucogranite had a rapid exhumation from ~9 Ma to ~8 Ma. Given the ~23–40°C/km geothermal gradient in southern Tibet (Francheteau et al., 1984) and

the closure temperature of ZHe ($180 \pm 20^\circ\text{C}$, Mahéo et al., 2007) and AHe ($70 \pm 15^\circ\text{C}$, Mahéo et al., 2007), we can calculate a minimum exhumation rate of $\sim 3.5 \pm 1$ mm/yr in this narrow timescale (~9–8 Ma) from depths of 7.8–4.5 to 3.0–1.7 km. After ~8 Ma, the leucogranite started its slow exhumation near the Earth's surface at a rate of 0.4–0.2 mm/yr.

Implications for E-W extension in the southern Tibetan Plateau

In recent decades, several detailed geochronology studies have been conducted for the E-W extension in southern Tibet, using zircon or monazite U-Th/Pb, mica $^{40}\text{Ar}/^{39}\text{Ar}$, and low-temperature thermochronology [fission track, (U-Th)/He] methods (Supplementary Table S1). At least seven main, ~NS-trending rifts resulting from E-W extension are located in southern Tibet (e.g., Armijo et al., 1986). West of the Xainza-Dinggye rift (XDR), the oldest, published onset timing for each of the main rifts, such as the Leo Pargil, Yari, Gurla Mandhata, Lunggar, Thakkhola, Gyirong, Tangra Yumco, and Kung Co rifts, were constrained at ~23–15 Ma by zircon U-Pb, apatite and zircon (U-Th)/He, as well as fission track dating (Figure 1C; e.g., Langille et al., 2012; Mitsuishi et al., 2012; Styron et al., 2013; McCallister et al., 2014; Shen et al., 2016; Wolff et al., 2019; Brubacher et al., 2021; Larson et al., 2020). By contrast, east of the XDR, the YGR, Cona, and Woka rifts show relatively young initiation timing at <13 Ma, and as recent as 2 Ma (Figure 1C; Supplementary Table S1; Harrison et al., 1995; Kapp et al., 2005; Cottle et al., 2009; Kali et al., 2010; Leloup et al., 2010; Bian et al., 2020, 2022; Shen et al., 2022a). Overall, initiation ages indicate an old to the west and young to the east of the XDR accommodating E-W extension in southern Tibet (Figure 7).

Such eastward younging of the south Tibetan rifts, and synchronous onset timing of the entire YGR have important implications for understanding and re-evaluating the proposed tectonic models of E-W extension in southern Tibet. Orogenic collapse model (Dewey, 1988) has been proposed as a mechanism for E-W extension, in which the collapse mainly occurs eastward, and extension should be predominant to the east (e.g., Copley, 2008). Delamination of thickened mantle lithosphere (England and Houseman, 1989), predicts that E-W extension occurred simultaneously in all rifts southern Tibet. Both models are inconsistent with the eastward younging pattern we suggest here. On the other hand, dynamics from the tearing of subducting Indian lithospheric slab (e.g., Replumaz et al., 2010; Chen et al., 2015), yields ages to decrease eastwards, but more related to a eastward wave of uplift, than a wave of extension. Thus, lithospheric dynamic models are inappropriate in dealing with the observed deformation.

Given the absence of mechanical coupling between the underthrusting Indian lithosphere beneath southern Tibet and upper crustal deformation (Wang et al., 2022), the main dynamics driving the initiation of E-W extension should instead be related to the boundary forces, in link with the Karakorum-Jiali fault zone (KJFZ) accommodating eastward extrusion of central/northern Tibet, as well as the thrusting along the Himalayan arc. In this model, the development of the south Tibetan rifts may be divided in two stages (Figures 8A,B, Supplementary Table S4, and references therein): the first stage at ~23–13 Ma, during which rifting in the arcuate western half of the southern Tibet bounded by

the Main Central Thrust (active since ~23–20 Ma) was triggered by the right-lateral strike slip of Karakorum-Gaize faults (active since ~25–22 Ma, Figure 8A; Supplementary Table S4); the second stage started at ~13 Ma, during which rifting in the eastern half of the southern Tibet bounded by the Main Boundary Thrust (active since ~14–13 Ma) was triggered by the right-lateral strike slip of Gyaring Co-Jiali faults (active since ~18–13 Ma, Figure 8B; Supplementary Table S4). Such E-W extension accommodates the ongoing orthogonal and divergent thrusting along the curved Himalayan arc (Armijo et al., 1986, 1989).

Conclusion

Our new authigenic illite K-Ar age and (U-Th)/He thermochronology confirm the onset timing of southern Yadong-Gulu rift at 9 ± 1 Ma. Combined with previous data from the northern YGR, our results suggest contemporary formation of the entire YGR at ~9–8 Ma. Overall, the onset timing of the seven ~NS-trending south Tibetan rifts appear to formed earlier to the west from ~23–13 Ma and later to the east from ~13 Ma. Therefore, we suggest that E-W extension in southern Tibet was driven by a combination of eastward propagation along the Karakorum-Jiali fault zone, together with divergent thrusting along the curved Himalayan arc.

Data availability statement

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author.

Author contributions

SW did the field work, interpreted the data and wrote the paper. XS interpreted the data and contributed to the modeling. M-LC and AR discussed, interpreted the data, and improved the manuscript including figures. YZ and JP did the field work. YZ, HL, JP, KL and XX discussed and improved the figures.

Funding

This research was financially supported by the second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0708), Scientific Research Fund of the National Institute of Natural Hazards, MEMC (ZDJ2019-19), the Ministry of Science and Technology of China (2019QZKK0901, 2021FY100101), the Key Special Project for Introduced Talents Team of the Southern Marine Science and Engineering Guangdong Laboratory (GML2019ZD0201), the China Geological Survey (DD20221630), and Institute of Geology, Chinese Academy of Geological Sciences (J1901-30, J2201).

Acknowledgments

We thank Mingkun Bai and Fucai Liu for their help in the field. We thank the Lab at NINH-MEMC for analyzing the samples and Haijia Lei for experiment. Constructive reviews by the reviewers as well as editorial work are gratefully appreciated.

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.993796/full#supplementary-material>

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 12 May 2022

ACCEPTED 04 August 2022

PUBLISHED 30 August 2022

CITATION

Liu X, Gao Z, Lei Q, Shao Y and Yao Y
(2022), Timing of Paleoeearthquakes and
Seismic Hazard of the Zhuozishan West
Piedmont Fault in the Northwestern
Ordos Block.
Front. Earth Sci. 10:942242.
doi: 10.3389/feart.2022.942242

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Timing of Paleoeearthquakes and Seismic Hazard of the Zhuozishan West Piedmont Fault in the Northwestern Ordos Block

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The Zhuozishan West Piedmont Fault (ZWPF) is an active normal fault in the northwestern corner of the Ordos Block. Studying the recurrence characteristics of paleoeearthquakes is useful for understanding the regional seismic hazard potential. In this study, paleoseismic trench and optically stimulated luminescence (OSL) dating methods are used to determine the temporal sequence of paleoeearthquakes. Seven earthquakes were constrained to be E7 (73.4 ± 12.9 ka to 75.3 ± 10.6 ka), E6 (63.7 ± 7.4 ka to 64.4 ± 8.2 ka), E5 (53.2 ± 7.0 ka to 58.7 ± 6.5 ka), E4 (42.6 ± 6.3 ka to 47.7 ± 5.8 ka), E3 ($31.0 - 3.4$ ka to 31.3 ± 3.4 ka), E2 (26.1 ± 2.7 ka to 28.1 ± 3.8 ka) and E1 (15.8 ± 1.9 ka to 20.2 ± 2.6 ka). The recurrence interval of strong earthquakes in the ZWPF is roughly constrained to be 8.1–14 ka, with magnitudes of Mw 7.0–7.4 that ruptured the entire fault. Based on the timing of the latest earthquake event, the elapsed time approached or exceeded the recurrence interval revealed from paleoeearthquakes. Hence, we suggest that the northwestern corner of the Ordos Block has a significantly high seismic hazard potential and that the occurrence of a strong earthquake should not be underestimated.

KEYWORDS

zhuozishan west piedmont fault, ordos block, normal fault, paleoseismic event, seismic hazard

Introduction

The boundary zone of Ordos Block are composed of normal faults and affected by regional large scale active faults, which are capable of generating strong earthquakes, which are one of the important strong seismic fault zones in China ([Research Group of Active Fault System around Ordos Massif, 1988](#)). Many earthquakes with $M \geq 8$, such as the Baotou earthquake in 849 ([Research Group of Active Fault System around Ordos Massif, 1988](#); [Jiang et al., 2000](#); [Nie et al., 2010](#); [Li et al., 2015](#)), the Hongdong earthquake

in 1303 (Meng et al., 1985; Jiang et al., 2004; Xu et al., 2018), the Huaxian earthquake in 1556 (Rao et al., 2015; Ma et al., 2016; Du et al., 2017; Feng et al., 2020), the Pingluo earthquake in 1739 (Zhang et al., 1986; Deng and Liao, 1996; Lin et al., 2013; Middleton et al., 2016) and the Haiyuan earthquake in 1920 (Deng et al., 1984; Zhang et al., 1988; Ren et al., 2016; Ou et al., 2020), have occurred within these basin systems and have caused severe casualties and economic losses (Figure 1). Research on active faults and seismic hazard assessment are important for reducing earthquake disasters.

Paleoearthquakes are related to prehistoric earthquake events preserved in geological records or events that were not recorded in history (Ran and Deng, 1999; Deng, 2002; Liu et al., 2021). The advantage of paleoseismic research is to compensate the

shortcomings and limitations of instruments and historical seismic records to a great extent so that we can understand the long-term activity and behavior of faults and estimate the hazard of future earthquakes over the period of several repeated earthquake cycles. In this study, we present field trench evidence for the occurrence of paleoearthquakes along the Zhuozishan West Piedmont Fault (ZWPF) in the northwestern corner of the Ordos Block (Figures 1, 2). In the study area, there is a lack of earthquakes with $M \geq 7$ recorded by historical and instrumental recordings. Therefore, we attempt to enrich data of paleoearthquakes by optically stimulated luminescence (OSL) dating in an investigation of the timing and recurrence intervals of large earthquakes during the Pleistocene along the ZWPF. Our results of paleoseismological analysis are important for assessing

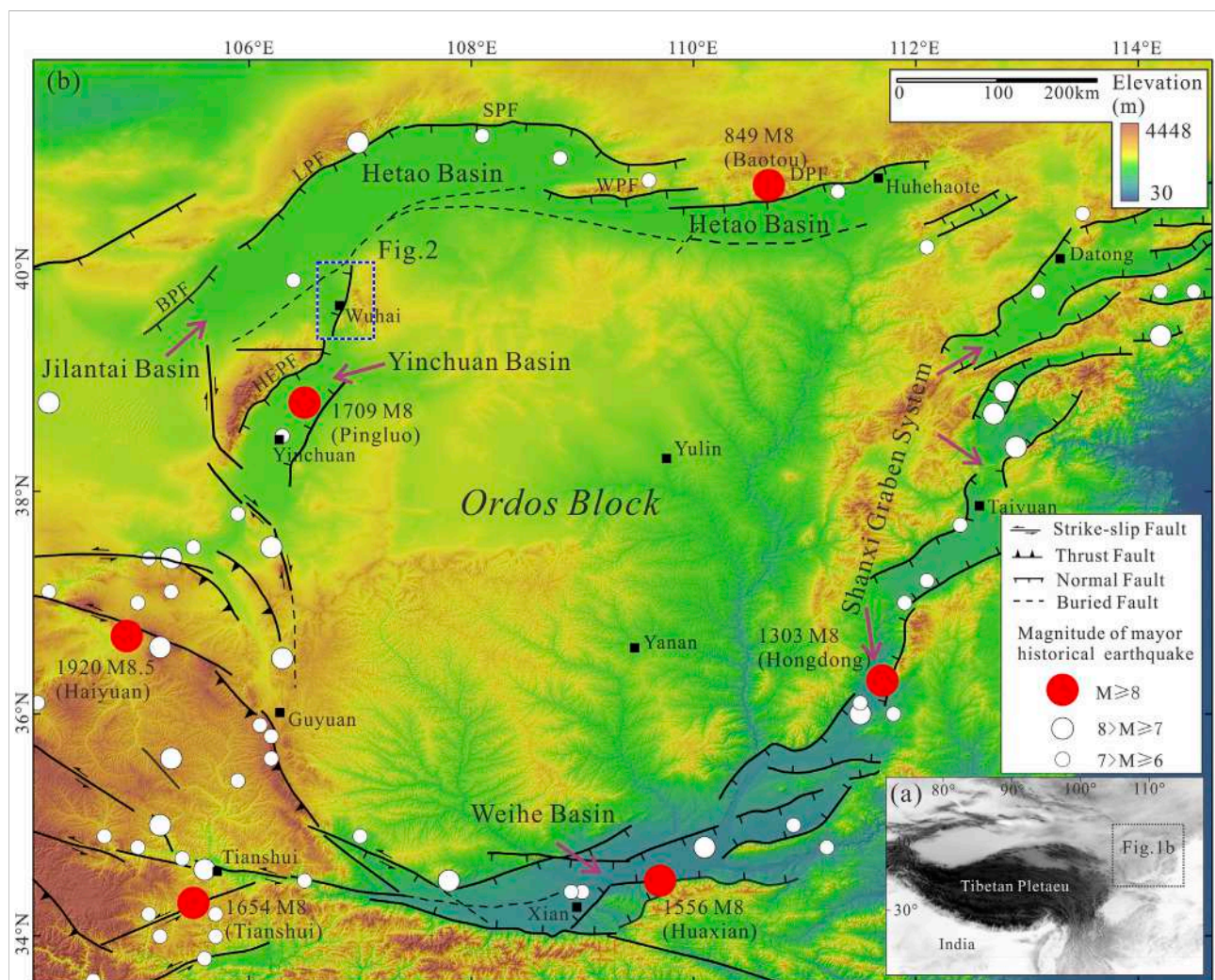


FIGURE 1

(A) Location of the Ordos Block. (B) Color-shaded relief map showing the distribution of large historical earthquakes and major active faults around the Ordos Block. The blue dashed rectangle shows the location of Figure 2. Abbreviations of names of active faults: DPF, Daqingshan Piedmont Fault; WPF, Wulashan Piedmont Fault; SPF, Seertengshan Piedmont Fault; LPF, Langshan Piedmont Fault; BPF, Bayanwulashan Piedmont Fault; HEPF, Helanshan East Piedmont Fault.

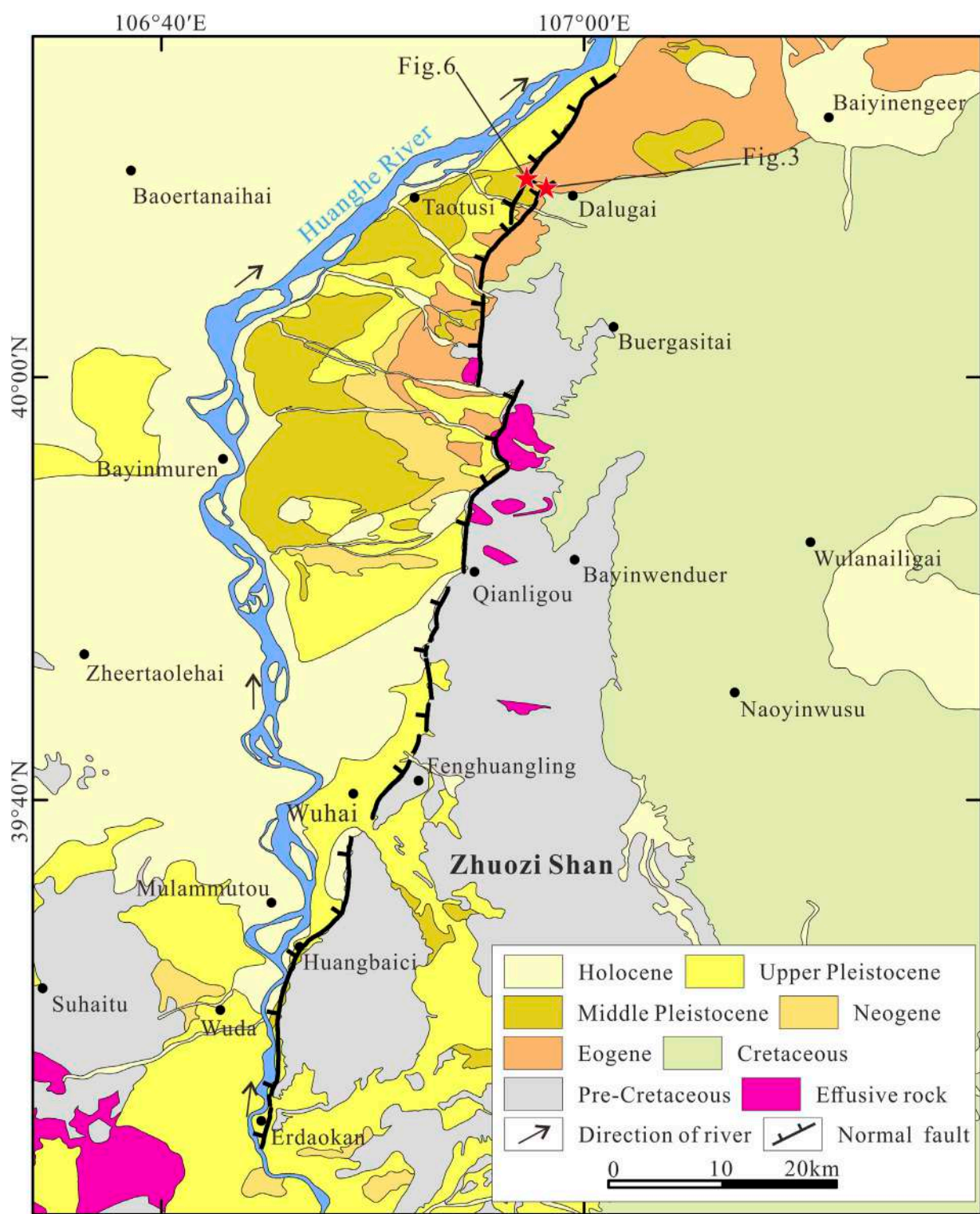


FIGURE 2
Geological map of the study area and its adjacent region, showing the distribution of bedrock lithology and the Zhuozishan West Piedmont Fault.

the seismic hazard in the northwestern corner of the Ordos Block and for a better general understanding of the seismogenic characteristics of normal faults in active basins.

Geological background

The Ordos Block, located in the western North China Craton, is a rigid block that has played an important role in tectonic activities in the Cenozoic in China (Research Group of Active Fault System around Ordos Massif, 1988; Zhang et al., 2002, 2003). There are no large-scale active faults developed within Ordos Block. It has good integrity and no records of earthquakes with a magnitude greater than M6 (Research Group of Active Fault System around Ordos Massif, 1988; Zhang et al., 2013), indicating the block is internally stable. However, there are series of large faulted basins controlled by active normal faults developed along the Ordos Block margin (Figure 1). A large number of studies have shown that regional tectonic activity and strong earthquakes mainly occur on active crustal block boundaries (Zhang et al., 2003; Zhang et al., 2004; Zhang et al., 2013; Zheng et al., 2020). According to historical data, more than 10 earthquakes with $M \geq 7$ have occurred around the Ordos Block, including five earthquakes with $M \geq 8$ (Research Group of Active Fault System around Ordos Massif, 1988; Jiang et al., 2000; Zhang et al., 2003). Therefore, the fault activities around the Ordos Block have always been the focus of active fault research (Research Group of Active Fault System around Ordos Massif, 1988; Jiang et al., 2000; Rao et al., 2015, 2016, 2018; Middleton et al., 2016; Dong et al., 2018; Xu et al., 2018; Zheng et al., 2019; Yu et al., 2021).

The study area is located in the northwestern corner of Ordos Block. A series of basins are developed in this area, including the Yinchuan Basin, the Jilantai Basin and the Hetao Basin (Research Group of Active Fault System around Ordos Massif, 1988) (Figure 1B). These basin boundaries are controlled by active normal faults that have caused destructive earthquakes. As the eastern boundary of the Yinchuan Basin, the Helanshan East Piedmont Fault (HEPF) is the seismogenic fault of the Pingluo M8 earthquake in 1739 (Zhang et al., 1986; Deng and Liao, 1996; Lin et al., 2013; Middleton et al., 2016). Previous studies suggested that the Daqingshan Piedmont Fault was the seismogenic fault of the Baotou M8 earthquake in 849 in the Hetao Basin (Research Group of Active Fault System around Ordos Massif, 1988; Jiang et al., 2000; Nie et al., 2010; Li et al., 2015). Another historical strong earthquake in the basin, an earthquake with a magnitude of nearly 8 in 7 BC, is considered to have occurred on the Langshan Piedmont Fault (LPF) (Li et al., 2015; Rao et al., 2016). There are also many paleoseismic records on the Sertengshan Piedmont Fault (SPF) and the Wulashan Piedmont Fault (WPF) (Yang et al., 2002; Yang et al., 2003; Chen et al., 2003; He et al., 2020). However, there are no records of

strong earthquakes and systematic studies of paleoearthquakes in the Jilantai Basin.

AS eastern boundary of the Jilantai Basin, the ~90-km-long NNE–SSW-trending Zhuozishan West Piedmont Fault (ZWPF) starts from the southeastern part of Dengkou County; extends southward by Dalugai, Qianligou, Fenghuangling, Wuhai, and Huangbaici; and terminates at Erdaokan from north to south, with dips ~50°–80° to the WNW (Figure 2). The fault traces are discontinuous and composed of several segments (Gao, 2020). If a strong earthquake occurs on the ZWPF, it will cause serious damages and losses to the Wuhai city. However, research on the ZWPF mainly focuses on the slip rate, the paleoseismic sequence and recurrence characteristics have not been studied (Zhang et al., 2014; Liang et al., 2019).

Materials and methods

Paleoseismic trenching has been the most effective method to study paleoearthquakes (Ran and Deng, 1999; McCalpin, 2009; Liu et al., 2021). In the exploration trench, the deformed strata and fault traces caused by strong earthquakes could be recorded. To identify normal fault paleoseismic events, colluvial wedges, filling wedges, and sudden changes in the displacement of different stratigraphic units are usually used as indicators (Deng et al., 1984; McCalpin, 2009; Ran et al., 2014). According to the recurrence model and evolution of paleoearthquakes along normal faults (McCalpin, 2009; Ran et al., 2014; Liu et al., 2021), after an earthquake, the fault scarp may collapse and result in the deposition of loose gravels, which finally become a colluvial wedge or filling wedges. Therefore, the fault planes together with the observed colluvial or filling wedges suggest an earthquake event. In the trenches we excavated, many colluvial or filling wedges are observed, which are the key indicators to identify paleoseismic events. Dating of the sediments below and overlying the wedges can be used to determine the lower and upper bound ages of paleoearthquakes (Ran and Deng, 1999; McCalpin, 2009; Ran et al., 2014; Liu et al., 2021).

Through remote sensing image interpretation and field investigation, we excavated three trenches near the Dalugai site along the northern segment of the ZWPF (Figure 2). Before excavation, we obtained high-resolution digital elevation model (DEM) data by using the structure from motion (SfM) technique for the study site (Bemis et al., 2014; Bi et al., 2017). Moreover, the SfM technique was also used to obtain photo mosaics of trench walls (Bemis et al., 2014; Gao et al., 2017).

To constrain the timing of paleoseismic events, OSL samples were collected from the sedimentary strata deposited before and after earthquakes, which can offer a time interval for the events (McCalpin, 2009; Ran et al., 2014). The dating was performed at the Laboratory of Neotectonic

TABLE 1 Result of OSL analysis for the layers in the trench.

	Depth (m)	U-238 (Bq/Kg)	Th-232 (Bq/Kg)	K-40 (Bq/Kg)	Dose rate (Gy/ka)	Equivalent dose (Gy)	OSL age (ka)
ZZS-OSL-11	5.0	25.4 ± 0.3	35.8 ± 4.3	607.1 ± 1.8	3.3 ± 0.1	229.8 ± 32.9	73.4 ± 12.9
ZZS-OSL-12	4.6	28.3 ± 1.4	21.7 ± 5.0	535.2 ± 0.4	2.7 ± 0.1	206.0 ± 19.8	75.3 ± 10.6
ZZS-OSL-13	4.1	49.0 ± 8.4	37.8 ± 3.1	554.6 ± 8.7	3.5 ± 0.1	222.4 ± 16.6	64.4 ± 8.2
ZZS-OSL-14	3.8	29.8 ± 1.2	31.1 ± 1.3	503.0 ± 1.4	3.1 ± 0.1	196.5 ± 10.4	63.7 ± 7.4
ZZS-OSL-18	0.6	45.5 ± 2.2	42.4 ± 3.1	503.8 ± 7.2	3.6 ± 0.1	56.4 ± 3.5	15.8 ± 1.9
ZZS-OSL-19	1.6	33.6 ± 2.5	37.0 ± 5.4	528.6 ± 9.2	3.3 ± 0.1	384.1 ± 23.8	117.0 ± 14.0
ZZS-OSL-20	0.9	38.7 ± 2.1	38.3 ± 2.2	456.8 ± 1.6	3.2 ± 0.1	290.4 ± 80.5	91.1 ± 26.9
ZZS-OSL-21	2.4	95.4 ± 1.0	88.6 ± 2.7	505.3 ± 10	4.5 ± 0.1	213.7 ± 13.6	47.7 ± 5.8
ZZS-OSL-22	2.0	59.1 ± 1.6	54.5 ± 1.0	536.2 ± 6.4	4.2 ± 0.1	178.7 ± 19.1	42.6 ± 6.3
ZZS-OSL-23	3.2	46.8 ± 1.3	46.9 ± 2.9	516.3 ± 3.2	3.6 ± 0.1	210.7 ± 8.4	58.7 ± 6.5
ZZS-OSL-24	2.8	75.6 ± 0.6	70.7 ± 1.4	638.1 ± 4.7	4.2 ± 0.1	224.1 ± 18.2	53.2 ± 7.0
ZZS-OSL-25	0.8	45.7 ± 1.0	41.7 ± 4.7	529.0 ± 5.8	3.6 ± 0.1	219.9 ± 20.6	61.1 ± 8.5
ZZS-OSL-26	0.4	39.1 ± 1.7	38.9 ± 1.5	532.9 ± 7.2	3.5 ± 0.1	211.9 ± 6.8	60.8 ± 6.5
ZZS-OSL-27	3.5	60.1 ± 2.3	41.7 ± 2.1	556.5 ± 11	3.8 ± 0.1	122.9 ± 10.1	32.3 ± 4.3
ZZS-OSL-28	3.2	50.9 ± 0.9	50.4 ± 1.1	544.9 ± 5.6	3.9 ± 0.1	122.2 ± 4.3	31.0 ± 3.4
ZZS-OSL-31	1.9	43.1 ± 1.2	32.7 ± 5.5	534.0 ± 0.6	3.2 ± 0.1	90.8 ± 8.1	28.1 ± 3.8
ZZS-OSL-32	1.6	50.1 ± 2.1	48.1 ± 1.9	506.3 ± 3.2	3.8 ± 0.1	81.3 ± 1.4	21.2 ± 2.2
ZZS-OSL-33	1.2	84.8 ± 0.9	76.3 ± 1.8	534.9 ± 5.8	4.3 ± 0.1	87.4 ± 6.5	20.2 ± 2.6
ZZS-OSL-36	2.6	39.1 ± 0.5	33.9 ± 3.4	539.0 ± 3.5	3.3 ± 0.1	314.6 ± 38.8	96.4 ± 15.5
ZZS-OSL-37	0.7	26.1 ± 5.0	26.8 ± 3.4	532.2 ± 0.2	3.0 ± 0.1	172.2 ± 8.0	57.2 ± 6.4
ZZS-OSL-38	5.0	39.3 ± 0.5	45.3 ± 3.2	537.1 ± 2.6	3.6 ± 0.1	170.4 ± 4.8	47.7 ± 5.1
ZZS-OSL-39	4.6	28.6 ± 3.8	45.6 ± 3.4	535.8 ± 3.5	3.5 ± 0.1	160.9 ± 6.6	45.7 ± 5.1
ZZS-OSL-40	2.4	40.6 ± 9.7	39.7 ± 8.3	523.7 ± 11	3.4 ± 0.1	142.0 ± 27.0	42.3 ± 9.1
ZZS-OSL-41	2.3	67.5 ± 1.9	51.5 ± 4.0	534.1 ± 0.6	4.2 ± 0.1	145.6 ± 3.0	35.0 ± 3.7
ZZS-OSL-42	2.2	61.5 ± 3.5	58.6 ± 2.1	606.5 ± 16	4.4 ± 0.1	130.7 ± 5.2	29.7 ± 3.3
ZZS-OSL-43	1.8	56.1 ± 0.6	53.5 ± 2.4	541.9 ± 4.4	4.1 ± 0.1	80.4 ± 7.0	19.5 ± 2.6
ZZS-OSL-44	1.6	55.8 ± 2.2	51.4 ± 2.7	525.2 ± 3.9	4.0 ± 0.1	138.2 ± 9.2	34.3 ± 4.2

(Continued on following page)

TABLE 1 (Continued) Result of OSL analysis for the layers in the trench.

	Depth (m)	U-238 (Bq/Kg)	Th-232 (Bq/Kg)	K-40 (Bq/Kg)	Dose rate (Gy/ka)	Equivalent dose (Gy)	OSL age (ka)
ZZS- OSL-45	1.2	67.0 ± 1.1	52.1 ± 2.1	510.4 ± 15	4.1 ± 0.1	104.3 ± 5.7	25.7 ± 3.0
ZZS- OSL-46	1.0	77.1 ± 0.3	64.6 ± 2.4	546.0 ± 2.6	4.7 ± 0.1	99.7 ± 6.0	21.0 ± 2.5
ZZS- OSL-51	1.7	29.7 ± 0.8	30.6 ± 2.5	524.4 ± 9.9	3.1 ± 0.1	98.1 ± 3.2	31.3 ± 3.4
ZZS- OSL-52	1.2	66.4 ± 15	50.7 ± 10.7	457.9 ± 0.5	3.9 ± 0.1	102.5 ± 2.3	26.1 ± 2.7
ZZS- 9	0.6	18.0 ± 2.0	30 ± 1.30	509 ± 9.51	3.1 ± 0.1	96.4 ± 1.6	31.5 ± 1.4
ZZS- 25	0.3	17.0 ± 2.6	28 ± 1.58	500 ± 6.36	3.32 ± 0.14	12.2 ± 0.3	3.7 ± 0.2
ZZS- 35	0.6	12.0 ± 0.9	34 ± 2.03	535 ± 9.51	3.08 ± 0.13	23.7 ± 0.5	7.7 ± 0.4

Geochronology of Institute of Disaster Prevention, China Earthquake Administration, by standard procedures for Chinese loess (Forman 1991; Wang, 2006). Fine-grain quartz fractions of 4–11 μm were obtained after a series of physical and chemical processes (Aitken, 1998). The equivalent doses were obtained based on the single-aliquot regenerative-dose (SAR) dating protocol (Wang, 2006). The determination of the environmental dose rate is also considered the standard rule (Aitken 1998; Wang, 2006). The dating information is listed in Table 1.

Paleoseismic observations and results

The northern segment of the ZWPF is composed of two left-step oblique faults (Figure 2). Trench 1 and trench 2 are located on the eastern side, and trench 3 is located on the western side (Figure 2).

Trench 1

Trench 1 and trench 2 are located on the flat alluvial fan far from the Zhuozi Shan. Based on the hillshade map interpretation and field investigation, in addition to the seasonal gullies, only one stage of alluvial fan development occurred around the trenches (Figure 3A). The fault traces cut the alluvial fan, which is clearly observed in the hillshade map and validated during field investigation (Figures 3A–C). DEM-based topographic profiles indicate that fault scarps at trench 1 and trench 2 sites are 2.0 ± 0.2 m and 2.6 ± 0.2 m, respectively (Figure 3D).

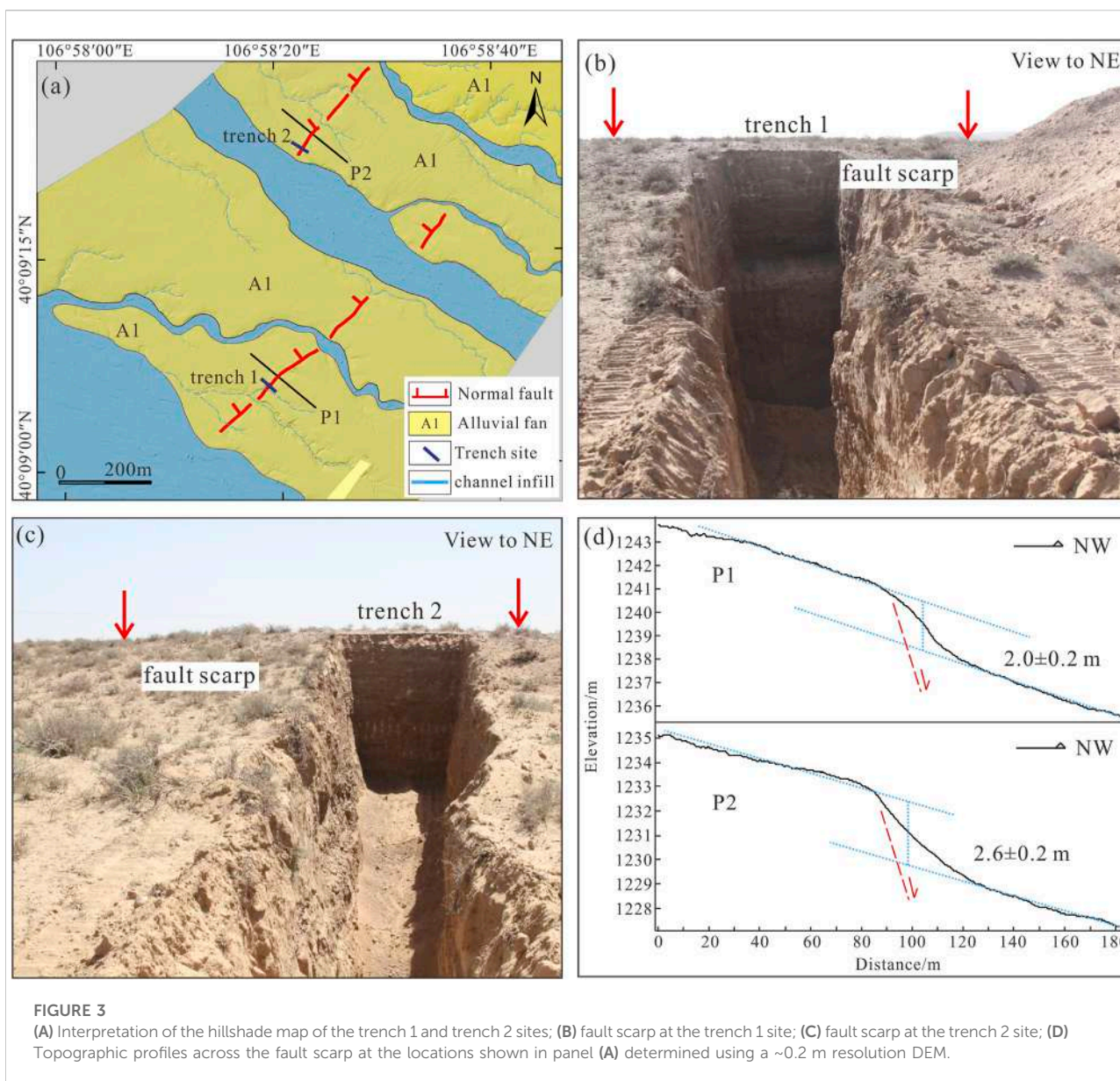
Faults can be easily identified according to the directional arrangement of gravel and the difference in strata on both sides of the faults plane in the mosaic photo of trench exposure (Figure 4). The three main faults (f1, f2, and f3) dip to the

west at dips of 63° – 67° , while McCalpin, 2009 other secondary faults dip to the east.

The strata exposed by trench 1 can be divided into at least 15 units (Figure 4). All units are displaced except for unit U15 which covers the main deformation zone. Unit U1 comprises the footwall in the trench exposure, while units U2–U14 constitute the hanging wall. U1 is mainly composed of a light yellow sand gravel layer with nearly horizontal bedding and contains multiple fine sand layers, in which only developed at the foot wall. Two OSL samples collected from U1 yielded ages of 96.4 ± 15.5 ka and 57.2 ± 6.4 ka (Figure 4; Table 1). These ages are much older than those of the hanging wall strata. The footwall strata appear as regular deposits except for U5. U5 is a unit of gravel layers with a thickness of approximately 2 m. Four cycles of the rhythmic deposits are discernible from the trench mosaic picture and log (Figure 4). Every sedimentary sequence contains a lower gray loose layer (U2, U6, U9, and U12) and upper light red and hard buried paleosol layer (U3, U7, U10, and U13). Buried paleosols reflect the long-standing paleosurface environment and are also important indicators to identify paleoseismic events (Chen et al., 2003; Ran et al., 2014). U4 and U8 are interpreted as colluvial wedges, while U11 and U14 are interpreted as filling wedges. U4 does not have the characteristics of typical colluvial wedge mixed accumulation and may also be the gravel accumulation in the early stage of U5.

Based on the identified colluvial wedges, filling wedges and buried paleosol, four paleoseismic events in trench 1 are identified, named E1-1, E1-2, E1-3, and E1-4 from youngest to oldest. By dating the sediments formed before and after the earthquake, the corresponding lower and upper ages of seismic events can be obtained.

The timing of E1-4 may be after the formation of U3 but before that of U5. The ages of the two samples collected from U2 and U3 are relatively consistent, which are 45.7 ± 5.1 ka and 47.7 ± 5.1 ka, respectively. (Figure 4; Table 1). We use the age of 47.7 ± 5.1 ka from U3 as the lower bound of E1-4. A lack of

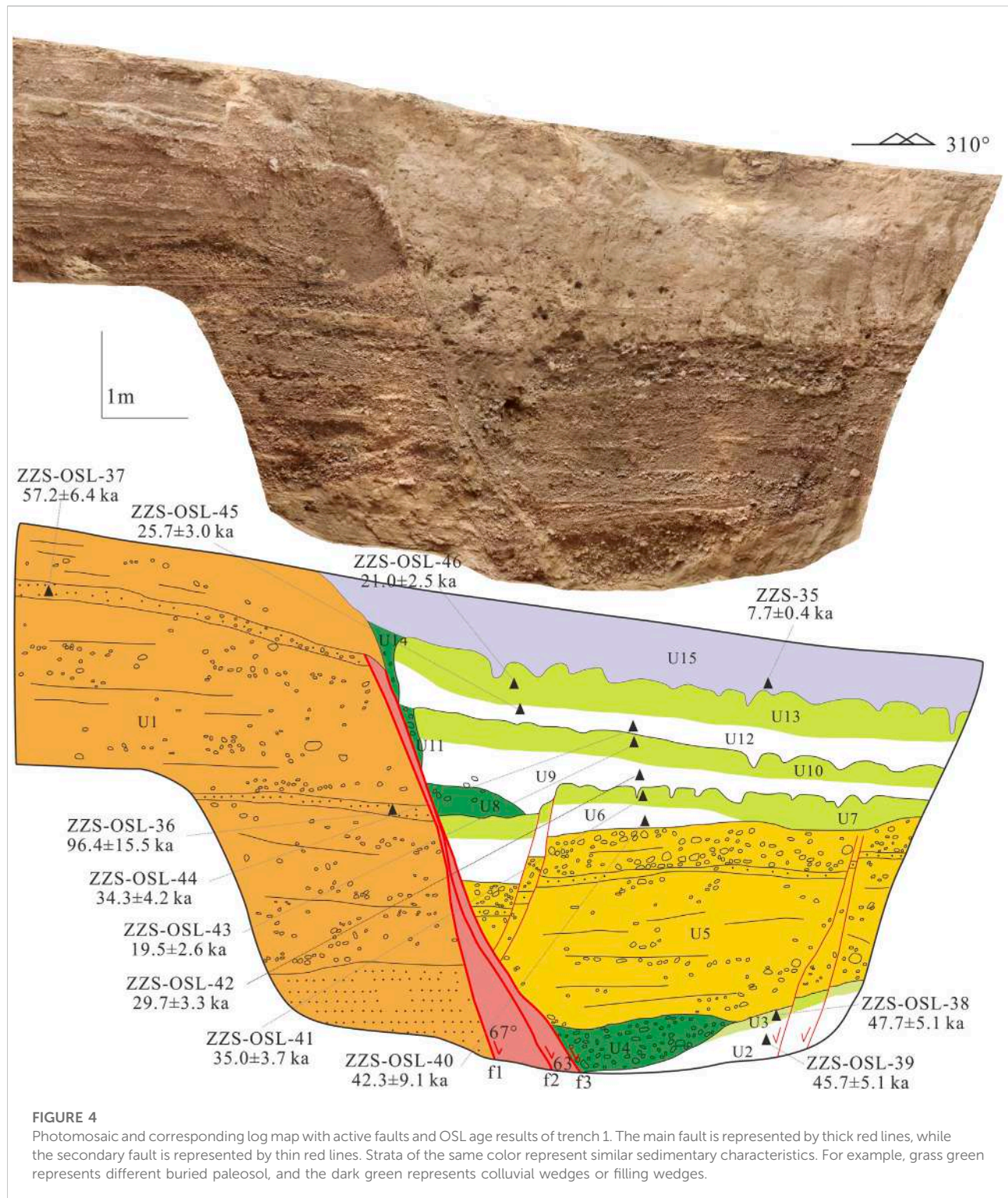


dateable material in U5 indicates that the timing of E1-4 cannot be accurately constrained. However, the sample collected from U6 can offer an upper bound age for E1-4. Thus, we roughly constrained that E1-4 occurred between 42.3 ± 9.1 ka and 47.7 ± 5.1 ka by dating U6 and U3 (Figure 4; Table 1). Two samples collected from U7 and U9 yielded ages of 35.0 ± 3.7 ka and 29.7 ± 3.3 ka, respectively (Figure 4; Table 1). These two ages provide the lower and upper bound ages for the occurrence of E1-3. The timing of E1-2 can be limited by samples from U12 and U10. However, the results of the ages are inverted and mismatched with the upper and lower units (Figure 4; Table 1). This may be related to incomplete bleached sediments or possible irregularities in the sampling process. Considering the ages and sedimentary relationship of the upper and lower units,

samples ZZS-OSL-45 and ZZS-OSL-42 are used to limit the timing of this event (Figure 4; Table 1). Thus, E1-2 is constrained to between 25.7 ± 3.0 ka and 29.7 ± 3.3 ka. E1-1 is the latest earthquake event that occurred on the fault. U14 is not faulted and is in sedimentary conformity with other strata. The samples collected from U15 and U13 constrain the timing of E1-1 to 7.7 ± 0.4 – 21.0 ± 2.2 ka (Figure 4; Table 1).

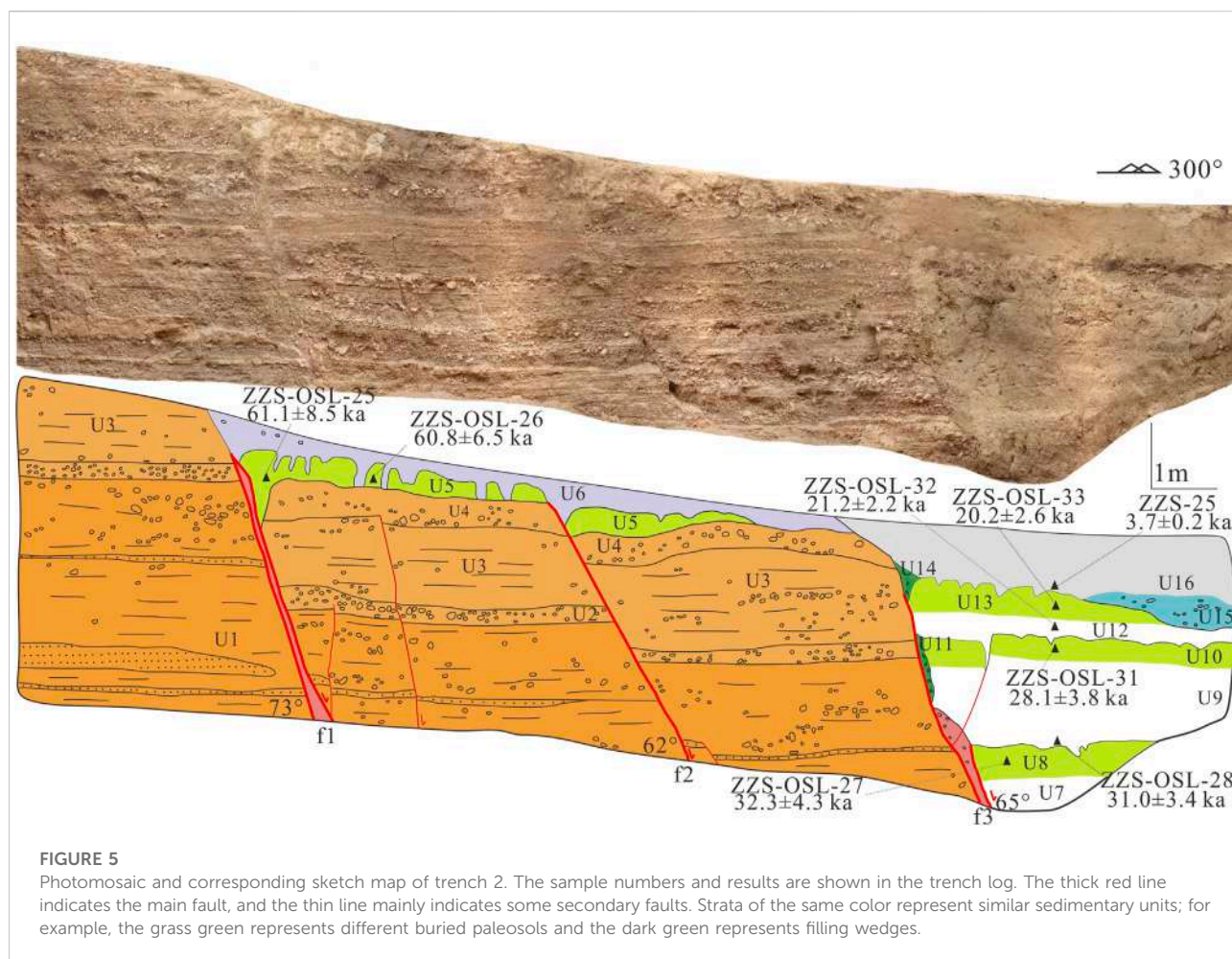
Trench 2

Three subparallel normal main faults, f1-f3, with dips of 62° – 73° to the west can be found in the trench based on stratigraphic correlation (Figure 5). Some secondary faults are



also distinguished (Figure 5). The strata can be classified into units U1–U16 from the field observations and mosaic picture of trench 2 (Figure 5). Units of U1–U4 are sand and gravel deposited units that formed in the early period. U5 is a buried paleosol that

experienced severe erosion. Then, it was covered by a loose U6 after an earthquake event. Similar to trench 1, trench 2 also contains three rhythmic sedimentary cycles; each cycle is composed of two units, U7 and U8, U9 and U10, and U12 and



U13, respectively (Figure 5). Among them, U7, U9, and U12 are paleosols, the surface has suffered a certain degree of erosion, and fissures have developed in them. U11 and U14 are interpreted as filling wedges, and U15 and U16 were deposited after the latest earthquake (Figure 5).

At least four earthquake events, named E2-1, E2-2, E2-3, and E2-4 from youngest to oldest, can be dated based on the colluvial wedges, filling wedges, and buried paleosol (Figure 5). In E2-4, f1 and f2 displaced units U1-U5 and were covered by U6. However, the fault throw on both sides of f1 is significantly greater than that of f2, which may indicate an earlier earthquake (Figure 5). Unfortunately, no other stratigraphic and sedimentary evidence remains.

The formation ages of U5 and U6 can offer the lower and upper bound ages for E2-4. However, the thickness of U6 is too thin to be suitable for the collection of OSL samples. Two samples collected from U5 yielded ages of 61.8 ± 8.5 ka and 60.8 ± 6.5 ka (Figure 5; Table 1). Therefore, we suggest that E2-4 occurred after 60.8 ± 6.5 ka. The two samples from U9 and U8 constrain E2-3 to between 31.0 ± 3.4 ka and 32.3 ± 4.3 ka (Figure 5; Table 1). Based on two samples collected from U12 and U10, E2-2 can be constrained to

between 21.2 ± 2.2 ka and 28.1 ± 3.8 ka. The timing of the latest E2-1 is limited by samples ZZS-25 and ZZS-OSL-33, i.e., to between 3.7 ± 0.2 ka and 20.2 ± 2.6 ka (Figure 5; Table 1).

Trench 3

Trench 3 is located west of the trench 1 and trench 2 (Figure 2). A reverse dipping fault scarp is developed at the trench 3 site (Figure 6A). Two periods of alluvial fans are identified according to the hillshade map and field investigation (Figure 6A). Fault scarps dipping to the east are well preserved on alluvial fans A1 and A2 (Figures 6B,C). The scarp heights on the two alluvial fans are obviously different, reflecting the long-term continuous activity of the fault. The vertical throws determined from the DEM are approximately 3.0 ± 0.1 m and 5.8 ± 0.2 m, respectively (Figure 6D).

There are 21 stratigraphic units identified from the trench log as presented in Figure 7. U1 is mainly a set of yellowish brown sand gravel layers that can be found in both the hanging wall and

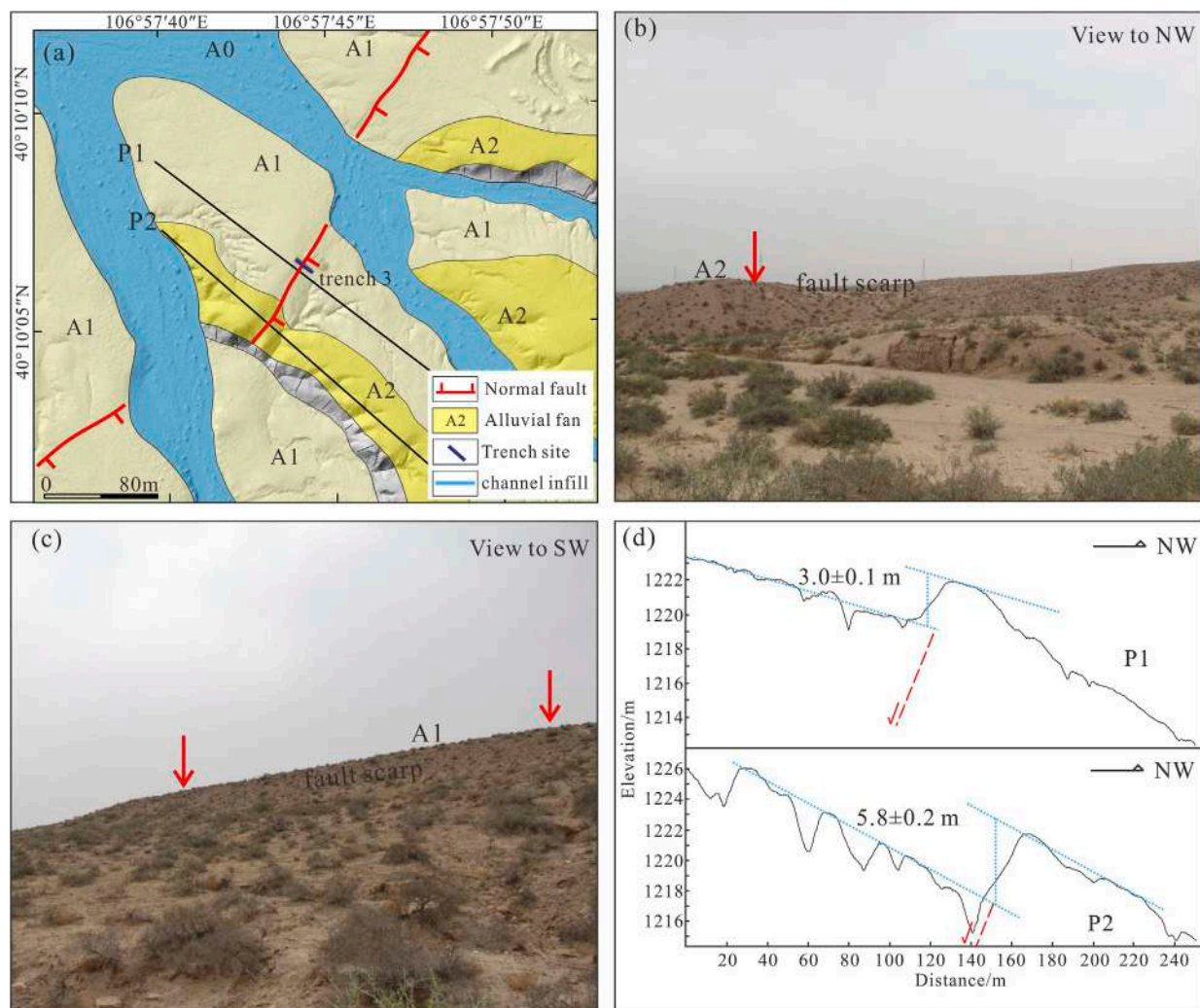


FIGURE 6

(A) Interpretation of the hillshade map at the trench 3 site; (B) fault scarp of the A2 alluvial fan; (C) fault scarp of the A1 alluvial fan; (D) topographic profiles across the fault scarp at the locations shown in panel (A) determined using a ~ 0.2 m resolution DEM.

the footwall. According to the results of two samples (117.0 ± 14.0 ka and 91.1 ± 26.9 ka) of the intercalated fine sand layer (Figure 7; Table 1), the formation age of U1 is ca. 100 ka. Similar to trench 1 and trench 2, six cycles of rhythmic deposits are distinguished in trench 3 (Figure 7). Each cycle is composed of a buried paleosol in the upper part (U3, U6, U10, U13, U16, and U19) and loose sediments in the lower part (U2, U5, U9, U12, U15, and U18). U4 and U8 are interpreted as colluvial wedges, while U1, U14, U17, and U20 are interpreted as filling wedges. U21 was deposited after the latest earthquake and interpreted as an unfaulted unit (Figure 7).

A series of normal faults with different dips are in the ca. 8-m-wide deformation zone identified (Figure 7). The main faults, f1 and f2 which have opposite dips, form a small graben structure, in which many secondary faults are developed

(Figure 7). According to the stratigraphic correlation of the hanging wall and footwall, the cumulative displacement of f1 and f2 is approximately 5.2 m. The f3 and f4, which are nearly vertical and dip to the east, are developed on the east side of the trench. According to U3, the cumulative displacement of the two faults is approximately 1 m. The total cumulative displacement of strata is significantly greater than the height of the fault scarp, which may be due to the short measurement profile, which cannot remove the influence of scarp-derived deposit. According to the sedimentary characteristics of the strata and paleoseismic identification marks (Figure 7), six earthquake events, named E3-1, E3-2, E3-3, E3-4, E3-5, and E3-6 from youngest to oldest, can be determined from the trench.

The formation ages of U3 and U5 can offer the lower and upper bound ages for E3-6. The dating results of two samples

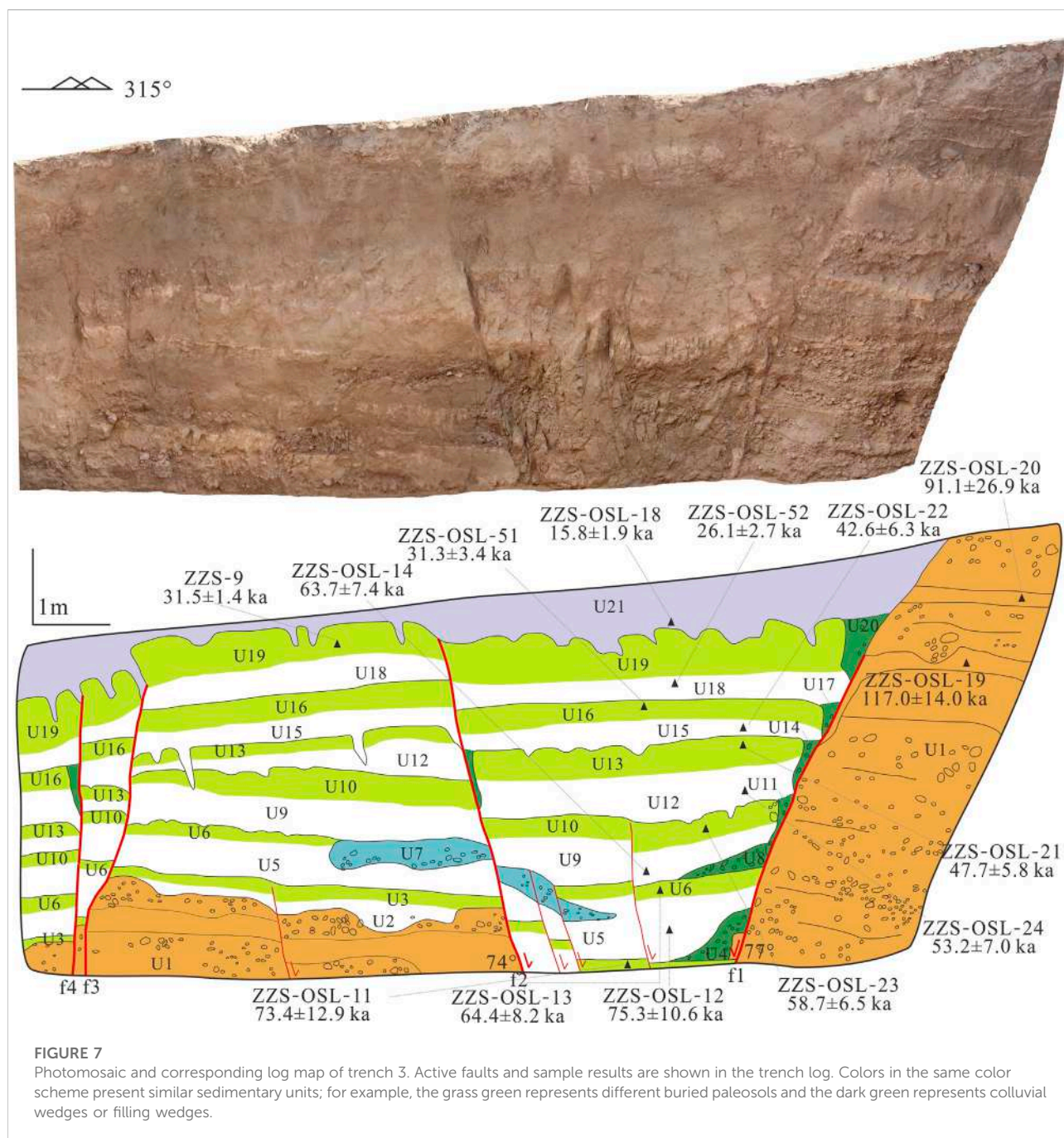
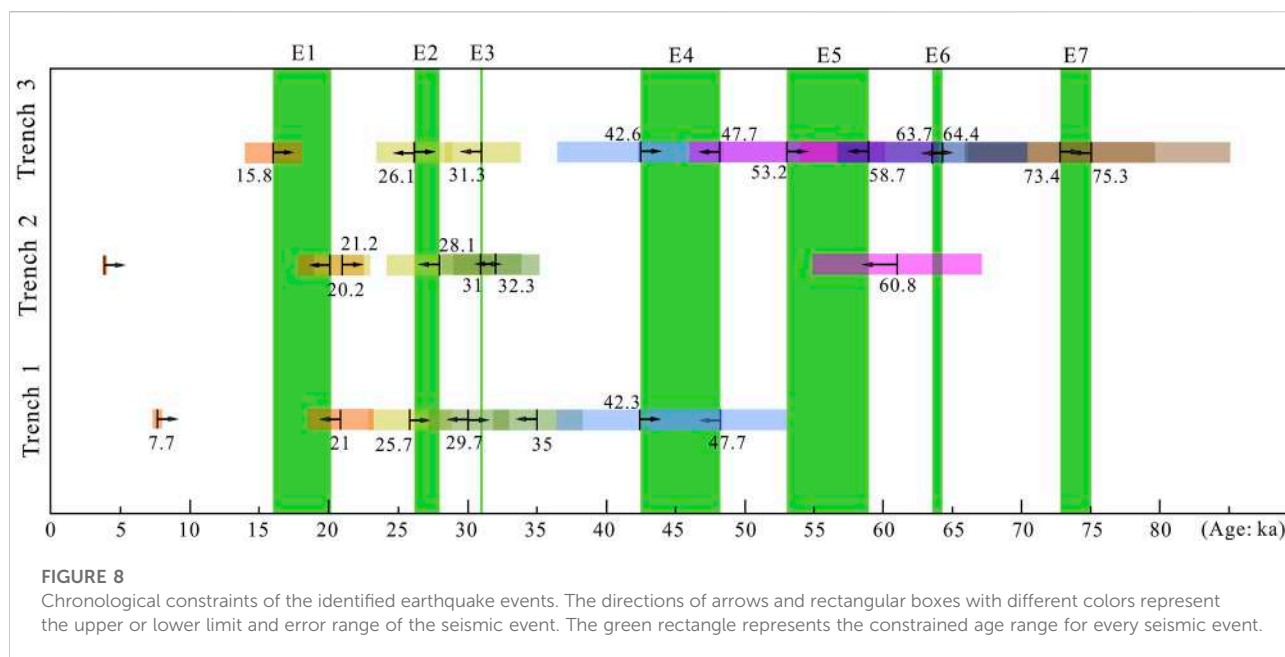


FIGURE 7

Photomosaic and corresponding log map of trench 3. Active faults and sample results are shown in the trench log. Colors in the same color scheme present similar sedimentary units; for example, the grass green represents different buried paleosols and the dark green represents colluvial wedges or filling wedges.

from U3 and U5 are 73.4 ± 12.9 ka and 75.3 ± 10.6 ka, respectively (Figure 7; Table 1). The results of the two samples are relatively consistent. Finally, the timing of E3-6 is determined to be between 73.4 ± 12.9 ka and 75.3 ± 10.6 ka. Two samples collected from U6 and U9 constrain the timing of the E3-5. The dating results yield ages of 64.4 ± 8.2 ka and 63.7 ± 7.4 ka for U6 and U9, respectively (Figure 7; Table 1). Therefore, the timing of E3-5 is between 64.4 ± 8.2 ka and 63.7 ± 7.4 ka.

The formation ages of U10 and U12 constrain the timing of E3-4 to between 53.2 ± 7.0 ka and 58.7 ± 6.5 ka (Figure 7; Table 1). Two filling wedges are developed in both E3-3 and E3-2 (Figure 7). The samples collected from U13, U15, U16, and U18 constrain the timing of E3-3 and E3-2 to 42.6 ± 6.3 – 47.7 ± 5.8 ka and 26.1 ± 2.7 – 31.3 ± 3.4 ka, respectively (Figure 7; Table 1). Filling wedge U20 is formed by sedimentation and is not faulted. Loose U21 formed after the earthquake and covered



U19, whose surface has been eroded. The timing of E3-1 should be constrained by the formation ages of U21 and U19. Sample ZZS-9 collected from U19 yields an age of 31.5 ± 1.4 ka, which is older than the underlying U18. By comprehensively comparing the age series of strata, we use the age of U18 as the lower bound of E3-1. Therefore, the timing of E3-1 is limited to between 15.8 ± 1.9 ka and 26.1 ± 2.7 ka (Figure 7; Table 1).

Discussion

Paleoseismic sequence of the Zhuozishan west piedmont fault

The three trenches have revealed many earthquake events, some of which have large age ranges, or some have only the upper or lower age bounds. Using the progressive constraining method (Mao and Zhang, 1995), these events can be constrained to a narrow range to better discuss the recurrence and interval mode of paleoearthquakes. This method assumes that several trenches are excavated on a fault segment, and the number of events revealed by each trench is not exactly the same. However, there are always some events with similar ages in different trenches. Although some events determine only the upper or lower bound ages and some determine the range of the event ages, they may jointly constrain a time range for the events. Seven earthquake events, named E1 to E7 from youngest to oldest, can be determined from the 3 trenches by using the progressive constraining method (Figure 8).

The event E7 is only recorded in trench 3 and has an age range of 73.4 ± 12.9 ka to 75.3 ± 10.6 ka (Figure 8). This event

occurred early with big dating the error range. Therefore, it can be used as a reference for discussing seismic events.

The event E6 is also only found in trench 3, with an age between 63.7 ± 7.4 ka and 64.4 ± 8.2 ka (Figure 8). E6 and E7 are not recorded in trenches 1 and 2, which may be because the formation age of the landform at the sites of trenches 1 and 2 is younger than that of E6.

The event E5 is revealed in trench 2 and trench 3. Samples collected from U5 in trench 2 give the lower age bound of E5. However, this event can be constrained to a narrower range of 53.2 ± 7.0 ka to 58.7 ± 6.5 ka based on the dating results from trench 3 (Figure 8).

The evidence of event E4 is observed in trench 1 and trench 3. We infer that this event is also recorded in trench 2, if the hanging wall of main fault f3 was excavated deeper. Buried paleosol U3 in trench 1 and U13 in trench 3 show similar ages of ca. 47.7 ka (Figures 4, 7; Table 1), indicating that this may be the paleosurface existing at the same time. Finally, the occurring time of E4 is determined to be between 42.6 ± 6.3 ka and 47.7 ± 5.8 ka (Figure 8). Zhang (2014) reported that a paleoseismic event occurred in ca. 43.8 ± 4.6 ka in the southern segment of the ZWPF. We infer that they are the same event.

Evidence of rupture associated with the event E3 can be found in all trenches, which indicates the occurrence of E3. To constrain the age of E3, the formation age of buried paleosol U7 in trench 1, U8 in trench 2 and U16 in trench 3 are determined. The results of three samples from these units are 35.0 ± 3.7 ka, 32.3 ± 4.3 ka and 31.3 ± 3.4 ka, respectively (Figures 4, 5, 7; Table 1), which show consistency. The age result at the bottom of their upper unit is 31.0 ± 3.4 ka (Figure 5; Table 1). Thus, 31.0 ± 3.4 ka to 31.3 ± 3.4 ka can be considered to be the

timing of E3 (Figure 8). Notably, the age we determined for the timing of E3-2 is between 26.1 ± 2.7 ka and 31.3 ± 3.4 ka. Given these uncertainties, E3-2 can be assigned to either E2 or E3. However, given that U17 (the colluvial wedge) seems to extend and flatten from the top of U16, the bottom of U18 (capped above both U16 and U17) is relatively horizontal. The earthquake event represented by the formation of U17 may have an age closer to the U16, which is 31.3 ± 3.4 ka. Therefore, E3-2 should be assigned to event E3.

Clues of surface rupturing due to event E2 are observed only in trench 1 and trench 2. The dating results of the two samples for the penultimate period of the buried paleosol are quite different. The result of U10 in trench 1 is 19.5 ± 2.6 ka, and that of U10 in trench 2 is 28.1 ± 3.8 ka (Figures 4, 5; Table 1). According to the analysis of the dating results of the upper and lower stratigraphic units, we suggest that 28.1 ± 3.8 ka can better represent the formation age of the paleosol in this period. The dating result at the bottom of its upper unit is 26.1 ± 2.7 ka (Figure 7; Table 1). The timing of E2 is determined to be between 26.1 ± 2.7 ka and 28.1 ± 3.8 ka (Figure 8).

The event E1 is the most recent earthquake on the ZWPF, and was disclosed by TC1, TC2, and TC3. It faulted the latest buried paleosol and was covered by a loose sedimentary layer on the surface. Two samples collected from U13 in trenches 1 and 2 yield similar ages of 21.0 ± 2.5 ka and 20.2 ± 2.6 ka (Figures 4, 5; Table 1). Thus, we use the age of 20.2 ± 2.6 ka as the lower bound of the event. The ages of the loose surface sedimentary layer vary greatly, ranging from 3.7 ± 0.2 ka to 15.8 ± 1.9 ka (Figures 4, 5, 7; Table 1). The bottom age of 15.8 ± 1.9 ka can be used as the upper bound of the event. Therefore, the timing of E1 can be constrained to 15.8 ± 1.9 ka to 20.2 ± 2.6 ka (Figure 8). In the southern segment of the ZWPF, the age of the latest seismic event revealed by the trench is 20.0 ± 2.5 ka to 23.2 ± 2.5 ka, which may be the same event (Zhang, 2014).

In summary, we determine that the timing of seven paleoseismic events observed in the trenches on the ZWPF are 73.4 ± 12.9 ka to 75.3 ± 10.6 ka (E7), 63.7 ± 7.4 ka to 64.4 ± 8.2 ka (E6), 53.2 ± 7.0 ka to 58.7 ± 6.5 ka (E5), 42.6 ± 6.3 ka to 47.7 ± 5.8 ka (E4), 31.0 ± 3.4 ka to 31.3 ± 3.4 ka (E3), 26.1 ± 2.7 ka to 28.1 ± 3.8 ka (E2) and 15.8 ± 1.9 ka to 20.2 ± 2.6 ka (E1) (Figure 8).

Implications for the seismic hazard

Paleoseismic trenches are mainly located in the northern segment of the ZWPF. However, the study of paleoearthquakes is lacking in the middle and southern segments of the fault due to the lack of continuous fine-grained sediments and the influence of human activities. Studies have shown that two seismic events have occurred in the southern segment of the fault at ca. 20 ka and 43 ka (Zhang, 2014), which are more consistent with the two events revealed by the trench in the northern segment. These

results suggest that earthquake events have ruptured along the entire length of the fault. By applying the empirical relationships between moment magnitude (M_w) and surface rupture length (Wells and Coppersmith 1994), we take the equations for a normal fault as $M_w = 4.86 + 1.32 \times \log(\text{SRL})$. With a length of ~ 90 km, we obtain an M_w value of 7.4.

The total vertical displacement of the six seismic events revealed by trench 3 is approximately 6 m. Then, the average vertical displacement of each earthquake is inferred to be at least 1 m. There is no corresponding stratum in the hanging walls and footwalls of the faults in trench 1 and trench 2, so it is difficult to determine the total vertical displacement. The colluvial wedge formed by a single earthquake is generally half of the vertical displacement (Wallace, 1977; McCalpin, 2009). According to the height of colluvial wedge U8 in trench 1, it is inferred that the displacement of a single event is also ~ 1 m. The northern segment of the ZWPF is divided into two branches (Figure 2). The vertical displacement on each fault is approximately 1 m. Then, the total displacement of a single earthquake is approximately 2 m. Considering the fault dips of 70° , it is inferred that the coseismic displacement of each earthquake is approximately 2.2 m. In the middle segment of the ZWPF near Wuhai city, the height of the fault scarps formed on the geomorphic surface of ca. 20 ka is approximately 2.5 m (Gao, 2020). The coseismic displacement is inferred to be approximately 2.8 m based on the dip of 65° . Based on the empirical relationships between moment magnitude (M_w) and average displacement (Wells and Coppersmith 1994), we take the equations for a normal fault as $M_w = 6.78 + 0.65 \times \log(\text{AD})$. With an average displacement of 2.5 m, we obtain an M_w value of 7.0. Therefore, we suggest a preferred magnitude in the range of M_w 7.0 to M_w 7.4 for the ZWPF.

Seven paleoseismic events have been revealed in the northern segment of ZWPF. Considering the intermediate value of each earthquake event, the ages of each earthquake are 74.4 ka (E7), 64.1 ka (E6), 56 ka (E5), 45.2 ka (E4), 31.2 ka (E3), 27.1 ka (E2), and 18 ka (E1). Thus, the recurrence intervals of each earthquake are 10.3, 8.1, 10.8, 14, 4.1 and 9.1 ka, respectively. The relevant evidence for E2 is recorded only in trench 1 and trench 2. No evidence is found in trench 3, although it has a good sedimentary sequence. Meanwhile, no events having a similar time were found in the middle and southern segments of the fault (Zhang, 2014; Gao, 2020). When talking about the recurrence characteristics, we consider only the seismic events that ruptured the entire fault length. Thus, E2 can be excluded from the events list when calculating the recurrence interval. Then, the recurrence intervals of the remaining major earthquake events are 10.3, 8.1, 10.8, 14 and 13.2 ka, with a range of 8.1–14 ka.

Considering both the paleoseismic characteristics and the rupturing behavior of the ZWPF, a strong earthquake (M_w 7.0–7.4) may be caused by the ZWPF, which will generate a significant seismic hazard to Wuhai and adjacent areas in the

northwestern corner of the Ordos Block. As shown above, we conclude that the recurrence interval of strong seismic events in the ZWPF is 8.1–14 ka based on the paleoseismic sequence obtained in this study (Figure 8). Moreover, the most recent surface-rupturing event occurred at 15.8 ± 1.9 ka to 20.2 ± 2.6 ka, which suggests that the elapsed time in the ZWPF has approached or exceeded the recurrence interval of paleoearthquakes. Hence, we infer the ZWPF has a high seismic potential.

Conclusion

Based on the results of the analysis of paleoseismic trenches and OSL dating, we have obtained the following conclusions.

1) At least seven surface-rupturing earthquakes have occurred on the ZWPF since the late Pleistocene. The times of each earthquake are constrained to 73.4 ± 12.9 ka to 75.3 ± 10.6 ka (E7), 63.7 ± 7.4 ka to 64.4 ± 8.2 ka (E6), 53.2 ± 7.0 ka to 58.7 ± 6.5 ka (E5), 42.6 ± 6.3 ka to 47.7 ± 5.8 ka (E4), 31.0 ± 3.4 ka to 31.3 ± 3.4 ka (E3), 26.1 ± 2.7 ka to 28.1 ± 3.8 ka (E2) and 15.8 ± 1.9 ka to 20.2 ± 2.6 ka (E1). Furthermore, E2 is inferred to have been an event of lower magnitude that partially ruptured the fault, while other events are considered to be strong earthquakes that ruptured the entire length of the ZWPF.

2) Based on the paleoseismic event sequence of the ZWPF, the earthquake recurrence interval is estimated to be 8.1–14 ka. Compared with the most recent earthquake (E1), the elapsed time has approached or exceeded the recurrence interval. We infer that the ZWPF has a high seismic potential and will pose a significant seismic hazard in the western corner of the Ordos Block.

To better understand the seismogenic behavior and seismic hazard of the ZWPF, we suggest that further paleoseismic studies are needed in the middle and southern segments of the fault.

Data availability statement

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author.

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Author contributions

XL carried out field work, drew the figures and wrote the manuscript. ZG, QL, YS, and YY performed the fieldwork, samples collection, and aerial photogrammetry. All authors contributed to the manuscript and approved the submitted version.

Funding

This work was supported by the National Key Research and Development Program of China (2017YFC1500100) and the Innovation Team Project of the Earthquake Administration of Gansu Province (2019TD0101).

Acknowledgments

We greatly thank the reviewers and editors for their detailed comments and suggestions, which helped improve the manuscript.

Conflict of interest

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 13 June 2022

ACCEPTED 08 August 2022

PUBLISHED 01 September 2022

CITATION

Ge J, Shi X, Chen H, Lin X, Ge W, Wei X,
Li F, Chen J, Li H, Cheng X and Zhang F
(2022), Two kinematic transformations
of the Pamir salient since the Mid-
Cenozoic: Constraints from multi-
timescale deformation analysis.
Front. Earth Sci. 10:967529.
doi: 10.3389/feart.2022.967529

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Two kinematic transformations of the Pamir salient since the Mid-Cenozoic: Constraints from multi-timescale deformation analysis

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The Pamir salient is a key part of the Himalayan–Tibetan Plateau orogenic system and has undergone intense tectonic deformation during the India–Asian collision. Delineating the Cenozoic kinematics and geodynamics of the Pamir salient requires a comprehensive understanding of the active arcuate structures along its frontal margin, from the perspective of the multi-spatiotemporal evolution of deformation patterns. Here, we reviewed the deformation rates of the major structures at different timescales, reanalyzed the published Global Positioning System velocities, and examined the present-day seismicity to constrain the kinematics of the Pamir salient since the Late Cenozoic. Integrated with the crustal evolution history during the Middle–Late Cenozoic and the deep structure, we proposed a new model to explain the multi-stage kinematics and associated geodynamics of the Pamir salient. During ~37–24 Ma, the initial Pamir salient moved northward via radial thrusting that rotated the basins on both sides, which was driven by the continuous compression of the Indian slab after the breakoff of its oceanic part. During ~23–12 Ma, the gravitational collapse of the Central and South Pamir crusts, which was induced by the breakoff of the continental part of the Indian slab, triggered the extension within the Pamir and foreland-ward movement of the upper crust. The upper crustal materials moved in varying directions due to the differential strength of the foreland areas, transforming the crustal kinematics from radial thrusting into a combination of radial thrusting and transfer faulting. Since the coupling of the Indian and Pamir slabs at ~12–11 Ma, the deformation propagation towards the forelands accelerated, after which the kinematics of the Pamir salient exhibited asymmetric radial thrusting that has been sustained until the present. The asymmetric radial thrusting was likely driven by the compressive stress effect of the lithospheric basal shear generated by the underthrusting of the cratonic

Indian lithosphere, which further led to the rollback of the Pamir slab and the consequent migratory extension in the South Pamir.

KEYWORDS

Pamir salient, multi-timescale, middle-late Cenozoic, kinematic models, geodynamic processes

1 Introduction

The Cenozoic India–Asia collision has driven the growth of the Pamir Plateau, at the western tip of the Himalayan–Tibetan orogen (Burtman and Molnar, 1993; Yin and Harrison, 2000). The Pamir Plateau, convexing northward and known as the Pamir salient, consists of a series of arcuate structures that extend for hundreds of kilometers and straddles neighboring China, Kyrgyzstan, Pakistan, and Tajikistan (Figure 1). These active

structures, including thrust, normal, and strike-slip faults, have played a significant role in the regional geodynamic evolution and pose major seismic and related hazards within and around the Pamir salient (Figure 1B).

Numerous studies have been conducted on these structures to explore how the Pamir salient evolved during the Cenozoic. Currently, two main end-member models exist to explain its formation. One argues that the arcuate shape developed as a result of the large-scale (~300 km) northward overthrusting of

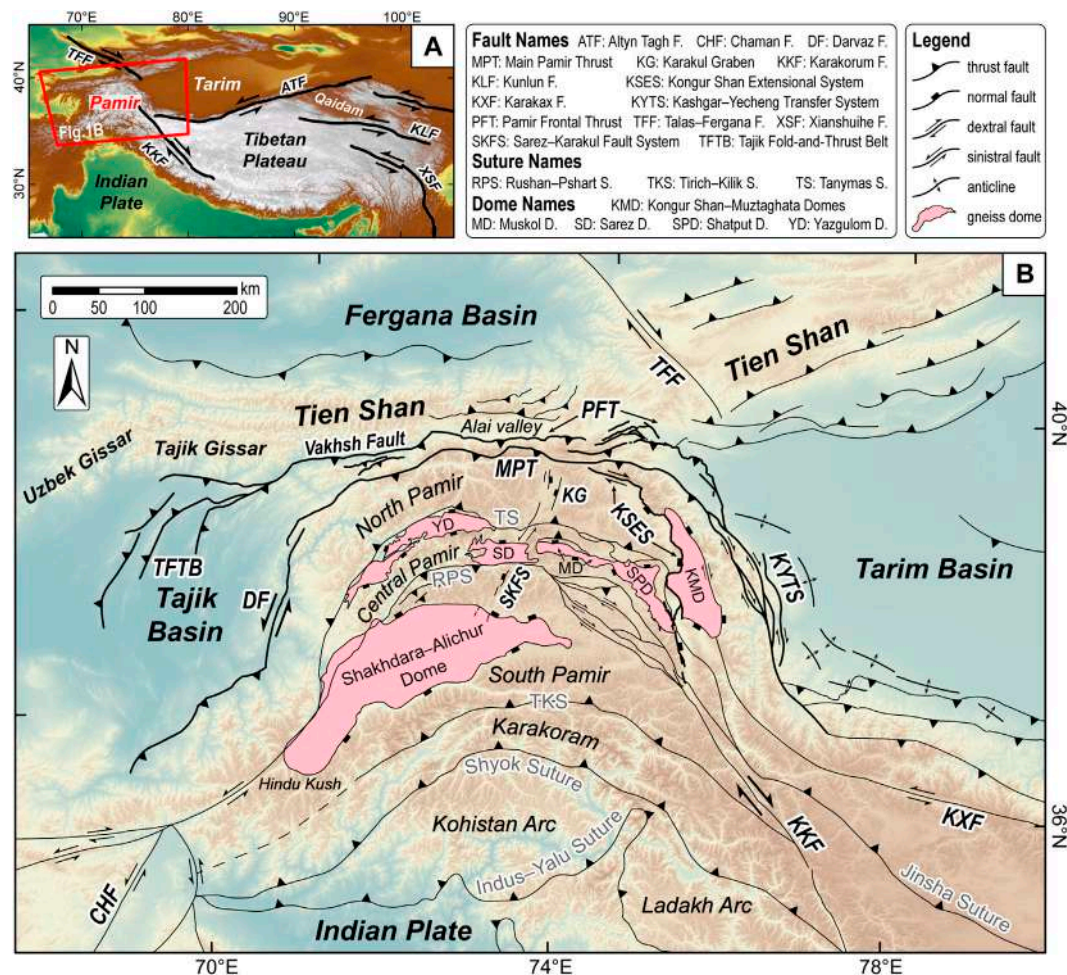


FIGURE 1

Tectonic setting of the study area and its surrounding regions. (A) The location of the Pamir salient within the tectonic context of the Himalayan–Tibetan orogen. (B) Tectonic map of the Pamir salient and its adjacent areas (modified from Wei et al. (2018) after Robinson et al. (2004), Robinson et al. (2007), Cowgill (2010), and Stübner et al. (2013a)). The bold black lines denote the arc-like active structures along the frontal Pamir salient. The pink-filled polygons define the Cenozoic gneiss domes.

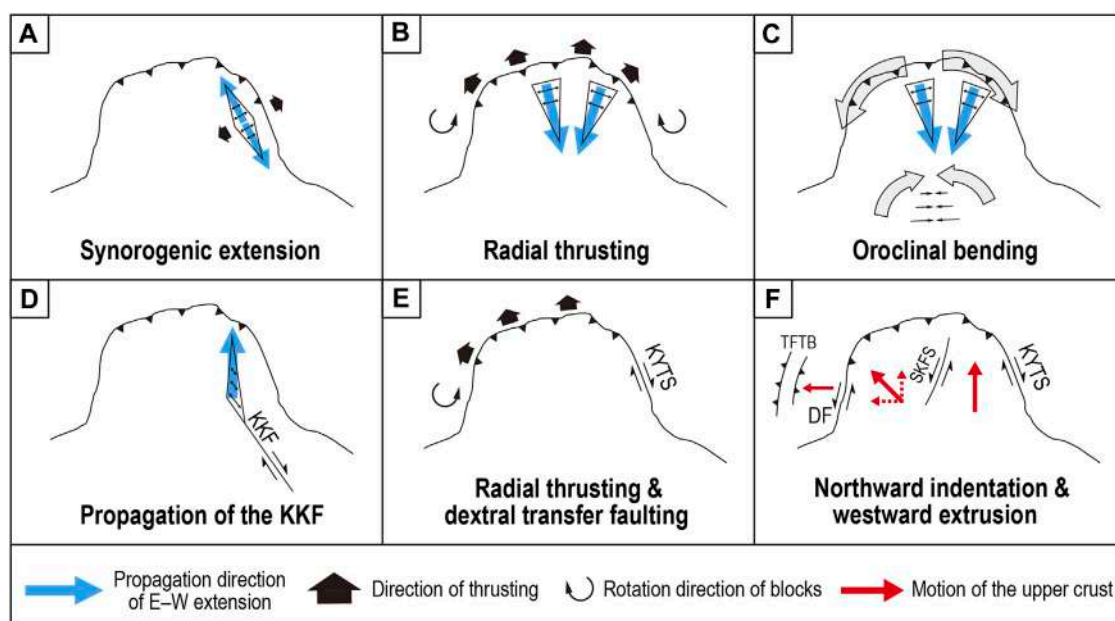


FIGURE 2

Previously proposed kinematic models of the Pamir salient (modified from Robinson et al. (2004), Cowgill (2010), Bosboom et al. (2014); Schurr et al. (2014)). See detailed description of these models in the text. Fault names same as Figure 1.

the Pamir upon the foreland areas (e.g., Burtman and Molnar, 1993; Cowgill, 2010; Sobel et al., 2013); and the other proposes that the arcuate shape was mostly inherited from pre-Cenozoic tectonics with <110 km northward movement during the Cenozoic (e.g., Chen et al., 2018; Li et al., 2020a).

Based on these two end-member models, several kinematic models have been proposed to describe the dynamic evolution of the Pamir salient (Figure 2). In early research, some of these models were established to explain the origin of E-W extension within the Pamir, with different predictions on the distribution and timing of extension. Brunel et al. (1994) proposed the synorogenic extension model, suggesting that the extension in the NE Pamir developed coevally with frontal thrusting to accommodate the topographic rise in the Kongur Shan–Muztaghata area. The model predicts the extension to propagate both northwesterly and southeasterly from the center (Figure 2A). Strecker et al. (1995) proposed the “radial thrusting” model, in which the Pamir salient moves northward by radial thrusting. Such a process may generate two southward-propagating extension zones in the plateau interior and drive anticlockwise and clockwise rotations on the west and east sides of the Pamir, respectively (Figure 2B). Yin et al. (2001) proposed that the oroclinal bending of the Pamir synchronously drove the extension within the Pamir and compression in the Nanga Parbat region (Figure 2C). In comparison with the radial thrusting model, this model predicts a similar extension mode, but its extension mechanic is different. A fourth kinematic model

proposed that the extension in the NE Pamir was caused by the northward propagation of the dextral-slip Karakoram Fault, hence the northward propagation of the extension (Figure 2D) (e.g., Ratschbacher et al., 1994; Strecker et al., 1995; Murphy et al., 2000). Recent research has proposed two hybrid models to reconcile the diverse kinematics of the Pamir. One is the radial thrusting and dextral transfer faulting model (Figure 2E). This model argues that the combined faulting may explain the tectonic transformation after the Miocene in the eastern flank of the Pamir, where the large-scale dextral slip initiated while the rotation of the Tarim Basin ceased (Cowgill, 2010; Bosboom et al., 2014). The other hybrid model is the northward indentation and westward extrusion model (Figure 2F). It proposes that the eastern Pamir moves northward *en bloc*, while the western Pamir shows bulk northward movement accompanying gravitational collapse and westward extrusion of Pamir’s western flank. This model has been used to explain observations from the Global Positioning System (GPS) velocity field and modern seismicity (Schurr et al., 2014; Metzger et al., 2020).

Despite significant progress, the above models for depicting the kinematic evolution of the Pamir salient are typically used to explain the formation of a certain structure or the kinematics of the Pamir salient during a particular period. Moreover, the relationship between most of these models and underlying geodynamic processes has rarely been discussed. All these hinder our comprehensive understanding of the complex

tectonic evolution of the Pamir salient during the Cenozoic, highlighting the need for a multi-timescale, interdisciplinary analysis of the regional tectonics.

Here, we targeted the frontal Pamir, attempting to examine its multi-timescale (million-year, millennial, and decadal) deformation and infer the regional tectonic evolution. First, we reviewed and evaluated the multi-timescale deformation rates of the main structures along the frontal Pamir salient. We then collated and reanalyzed the published GPS data to constrain the crustal kinematics of the Pamir salient and its adjacent regions and present deformation rates across the major structures. We also analyzed the modern shallow seismicity to observe the deformation characteristics. Finally, by integrating the multi-timescale data and the burial and exhumation history of the Cenozoic gneiss domes and the deep structure under the Pamir, we examined the spatiotemporal pattern of fault kinematics and associated geodynamics of the Pamir salient since the Middle Cenozoic.

2 Background

The Pamir salient consists of three terranes (North, Central, and South Pamir) that evolved along or accreted to the southern margin of Asia during the Late Paleozoic–Mesozoic (Burtman and Molnar, 1993; Schwab et al., 2004; Robinson et al., 2012). During the Cenozoic evolution of the Pamir, significant crustal deformation and complex deep geodynamic processes occurred within the continent, (e.g., Burtman and Molnar, 1993; Chapman et al., 2017). These processes established the present tectonic framework and subcrustal structure of the Pamir salient. In this section, we summarize the main structures, burial and exhumation history of the Cenozoic gneiss domes, and deep structure of the Pamir, which provide a general context for analyzing the crustal kinematics and related geodynamic mechanisms.

2.1 Main structures in the Pamir

The northern front of the Pamir is defined by thrust faulting, namely the south-dipping Pamir Thrust System (PTS), which has absorbed the major N–S shortening across the system (Coutand et al., 2002; Zubovich et al., 2010). The PTS is divided from south to north into the Main Pamir Thrust (MPT) and Pamir Frontal Thrust (PFT). The MPT has been thrusting the North Pamir over the Trans-Alai Range since its initiation at ~25–20 Ma (Sobel and Dumitru, 1997; Coutand et al., 2002; Bershaw et al., 2012). In places, this fault appears to have been inactive during the Quaternary (Sobel et al., 2013; Thompson et al., 2015), and most deformation across the northern Pamir front is now being accommodated by the PFT, which initiated at ~6–5 Ma and is the latest deformation front of the northward propagation

of the Pamir (Thompson et al., 2015; Chen et al., 2019; Chen et al., 2021). The PFT lies north of the MPT and thrusts the Trans-Alai Range over the Alai Valley and consists of segmented thrust faults and dextral transfer zones (Arrowsmith and Strecker, 1999; Li et al., 2012; Patyniak et al., 2021). The western section of the PFT (hereafter referred to as the western PFT) stretches along the southern margin of the Alai Valley between 72°E and 73.8°E, while the eastern section (hereafter referred to as the eastern PFT) is located in the northeastern corner of the Pamir, with strikes varying from E–W to NW–SE (Figure 1B).

Along both flanks of the Pamir, two large strike-slip fault systems accommodate the northward motion of the Pamir with respect to its adjacent blocks. In the west, the sinistral-slip Darvaz Fault (DF) marks the boundary between the Pamir and Tajik Basin. The DF has displaced the Panj River sinistrally by ~100 km, where the river cuts across the DF, compared to the reconstruction of the river at ~10 Ma (Brookfield, 2008). To the west, the Tajik Fold-and-Thrust Belt (TFTB) has absorbed the major shortening in the Tajik Basin, which has been generated by the foreland-ward expansion of the Pamir (e.g., Chapman et al., 2017; Gągala et al., 2020). In the east, the dextral-slip Kashgar–Yecheng Transfer System (KYTS) separates the Tarim Basin from the eastern Pamir. The KYTS comprises four nearly parallel faults: the Kusilaf, Kumtag, Yarkand, and Aertashi Faults (Cowgill, 2010; Sobel et al., 2011). The dextral slip of the KYTS commenced at ~25–18 Ma, based on sediment provenance, magnetostratigraphic, and thermochronological analyses (Sobel and Dumitru, 1997; Blayney et al., 2016; Blayney et al., 2019).

The interior of the Pamir is characterized by the extension of normal fault systems and domes during the Cenozoic. The normal fault systems dominated by E–W extension are mainly represented by the Karakul Graben (KG) in the north, and the Kongur Shan Extensional System (KSES) in the northeast (Figure 1B).

The NNE–SSW-trending KG is bounded by high-angle (~60°) normal faults on its western and eastern sides with weak tectonic activity (Strecker et al., 1995). The total extension across the KG may not exceed ~3 km, and the estimated extension rate is ~0.5–1 mm/yr, implying that the extension may have started between 9 and 3 Ma (Amidon and Hynek, 2010). The normal faults in the KG, together with sinistral-slip and oblique normal faults in the south, form the Sarez–Karakul Fault System (SKFS, Figure 1B) (Schurr et al., 2014).

The ~250 km-long KSES consists of, along its strike from NWW to SSE, the Muji dextral-slip fault, the Kongur Shan normal fault, the Taheman normal fault, and the Tashkurgan normal fault (Arnaud et al., 1993; Robinson et al., 2004). Previous thermochronology research on the Kongur Shan and Tashkurgan Faults suggested that the KSES started to develop in the Late Miocene. In the north, the Kongur Shan Fault along

the Kongur Shan Dome initiated at ~7 Ma (Robinson et al., 2010) or ~6–5.5 Ma (Cao et al., 2013a); in the middle, the fault along the Muztaghata Dome began to be active at ~6–5 Ma (Robinson et al., 2007; Cao et al., 2013a; Cao et al., 2013b; Thiede et al., 2013); and in the south, the Tashkurgan Fault initiated coevally with the Kongur Shan Fault, whose initiation age is ~10–5 Ma, most likely ~6–5 Ma (Chen and Chen, 2020).

2.2 Burial and exhumation history of the Cenozoic gneiss domes

The Cenozoic gneiss domes are widely exposed throughout the Pamir, occupying ~1/3 of its surface area (Figure 1B) (Stübner et al., 2013a). These gneiss domes are bounded by normal-sense shear zones and have been exhumed from depths of 30–40 km (e.g., Robinson et al., 2004; Schmidt et al., 2011), recording the deep crustal processes to be discussed in this paper. Therefore, we briefly summarize the burial and exhumation history of these domes below.

The burial and exhumation processes of the gneiss domes may be related to regional crustal thickening and thinning (e.g., Stearns et al., 2015). The gneiss domes in the Pamir include the 1) Yazgolom, Sarez, and Muskol-Shatput Domes in the Central Pamir, 2) Shakhhdara–Alichur Domes in the South Pamir, and 3) Kongur Shan–Muztaghata Domes in the eastern Pamir (Figure 1B). The extension direction of the Central and South Pamir domes is nearly N–S, whereas the exhumation of the Kongur Shan Dome has been accommodated by the E–W extension of the KSES. The Muztaghata Dome, considered the eastern continuation of the Muskol-Shatput doming (Robinson et al., 2007; Cao et al., 2013b; Thiede et al., 2013), records both the earlier N–S extension of the Central Pamir domes and the later E–W extension along the KSES (e.g., Robinson et al., 2007; Cao et al., 2013b).

The Central and South Pamir domes experienced coeval prograde metamorphism in the middle–lower crust during ~37–18 Ma (e.g., Schmidt et al., 2011; Stearns et al., 2013; Smit et al., 2014; Stearns et al., 2015). The Yazgolom Dome was buried to start prograde metamorphism at ~35 Ma, reaching the peak metamorphism at ~22 Ma and lasting until ~18 Ma (Stearns et al., 2015; Hacker et al., 2017). The other Central Pamir domes began to experience prograde burial from depths of 15–20 km at 35–27 Ma to 25–30 km by 22–19 Ma (Hacker et al., 2017). In the South Pamir, the prograde metamorphism of the Shakhhdara Dome started at ~37 Ma and reached its peak at ~19 Ma when the maximum burial depth was over 50 km (Smit et al., 2014; Hacker et al., 2017).

The initial exhumation of the Central and South Pamir domes was also almost synchronous, and the synorogenic extension of these domes started at ~23–20 Ma (Lukens et al., 2012; Stübner et al., 2013b; Stearns et al., 2015; Rutte et al., 2017a; Worthington et al., 2020). However, the duration of exhumation

differs between the Central and South Pamir domes: the doming in the South Pamir was longer lasting. The Central Pamir domes experienced rapid extensional exhumation at an average rate of ~3 mm/yr before 12 Ma, which was followed by a phase of slow cooling (Rutte et al., 2017a). The total amount of ~N–S extension in the Central Pamir ranges from 17 to 75 km (Rutte et al., 2017b). In detail, the bulk exhumation of the Shakhhdara Dome lasted from ~18 to 15 Ma to ~2 Ma at rates of 1–3 mm/yr, resulting in ~90 km of ~N–S extension (Stübner et al., 2013a; Stübner et al., 2013b). In comparison, the neighboring Alichur Dome lasted until ~4 Ma at rates of ~1.1 mm/yr, and the total ~N–S extension is much smaller, at 27–41 km (Worthington et al., 2020).

The formation of the Kongur Shan–Muztaghata Domes occurred later than those in the Central and South Pamir. The Muztaghata Dome experienced prograde metamorphism during ~30–12 Ma due to crustal thickening in the northeastern Pamir (Robinson et al., 2007; Cai et al., 2017; Li et al., 2020b). The dome then underwent two stages of exhumation. The first-stage rapid exhumation occurred during ~12–8 Ma, reflecting ~N–S extension driven by the south-dipping Shen-ti normal fault (Robinson et al., 2007; Cao et al., 2013b). The second-stage exhumation slowed and has continued to the present, relating to the E–W extension along the KSES since ~6–5 Ma and focused glacial and fluvial erosion since ~4–3 Ma (Robinson et al., 2007; Sobel et al., 2011; Cao et al., 2013a; Cao et al., 2013b; Thiede et al., 2013). The Kongur Shan Dome was buried at a crustal depth of ~30 km at ~9 Ma, when it experienced peak metamorphism (Robinson et al., 2004). In contrast to the Muztaghata Dome, the Kongur Shan Dome is characterized by rapid exhumation at a rate of >3 mm/yr due to the onset of the KSES at ~7 or ~6–5 Ma (Robinson et al., 2010; Cao et al., 2013a; Thiede et al., 2013). This dome experienced another phase of enhanced exhumation during ~3–1 Ma, caused by glaciation and/or fluvial incision and erosion in response to the continuous extension of the KSES (Cao et al., 2013a).

2.3 Deep structure under the Pamir

Beneath the Pamir–Hindu Kush region, a ~500-km-long belt of intermediate-depth earthquakes extends from the Hindu Kush northeastward into the Central Pamir and then bends eastward into the eastern Pamir (Kufner et al., 2018; Bloch et al., 2021). Recent studies on the intermediate-depth earthquakes and teleseismic tomography have distinctly depicted the lithospheric configuration of the Pamir (e.g., Sippl et al., 2013b; Kufner et al., 2016; Schneider et al., 2019; Liang et al., 2020; Xu Q. et al., 2021). The intermediate-depth earthquakes characterize two separate continental Wadati–Benioff zones: the Hindu Kush seismic zone and the Pamir seismic zone (e.g., Burtman and Molnar, 1993; Fan et al., 1994; Kufner et al., 2018).

Beneath the Hindu Kush, the seismicity-imaged slab has generally been interpreted as the northward-subducting Indian

plate, with its dip steepening downward into sub-vertical or even overturned directions (Kufner et al., 2016; Kufner et al., 2017; Liang et al., 2020; Peng et al., 2020). Tomographic images show a low-velocity zone at ~60–160 km depth and a high-velocity zone at ~160–600 km depth, interpreted as the subducted crust overlying the lithospheric slab (Kufner et al., 2017; Kufner et al., 2021). The elongated shape of the deep-reaching lithospheric slab, together with high tension strains within it, suggests ongoing slab break-off and thinning of the subducting marginal Indian lithosphere (Lister et al., 2008; Kufner et al., 2017; Kufner et al., 2021).

Beneath the Pamir, receiver function images, local earthquake tomography, and observation of guided waves have revealed an arcuate Pamir slab (lower crust and upper mantle) with various dips (Sippl et al., 2013a; Sippl et al., 2013b; Schneider et al., 2013; Li et al., 2018; Mechie et al., 2019; Bloch et al., 2021; Xu Q. et al., 2021). The Pamir slab dips to the east under the western South Pamir and then bends south-dipping under the Central Pamir. To the east of a seismicity gap between 74.4°E and 74.9°E, the south-dipping slab abruptly overturns to the north. A slab tear may have occurred at the boundary, showing flipping of the slab dips (Bloch et al., 2021). Below the southern portion of the eastern Pamir, the observed double Moho structure, P-wave velocity structure, and seismicity distribution indicate that the Pamir slab has underthrust eastward beneath the Tarim lithospheric mantle (Bloch et al., 2021; Xu Q. et al., 2021). South of the arcuate Pamir slab, the presence of a high-velocity anomaly was interpreted as the underthrusting cratonic Indian lithosphere (Sippl et al., 2013a; Li et al., 2018; Liang et al., 2020; Bloch et al., 2021). The northward-advancing cratonic Indian lithosphere, as an indenter, was inferred to have forced the delamination and rollback of the Pamir/Asian lithosphere, forming the arcuate Pamir slab (Kufner et al., 2016; Bloch et al., 2021). The slab tear or overturns underneath the eastern Pamir may be related to the along-strike variability of the indenter tip and/or lateral differences in the strength and density of the indented lithospheres (Bloch et al., 2021; Kelly and Beaumont, 2021). The collision of the cratonic Indian lithosphere and the Pamir slab could have started at ~12–11 Ma, based on the India–Asia convergence rate of ~34 mm/yr (Molnar and Stock, 2009) and ~380–400 km delamination of the Pamir slab (Kufner et al., 2016).

3 Multi-timescale tectonic deformation along the frontal Pamir salient

The arcuate tectonic belt in the frontal Pamir salient consists of several main structures with different tectonic styles, including the DF, TFTB, PTS, KSES, and KYTS. We summarized the million-year, millennial, and decadal deformation rates of these structures (Table 1) based on previous research.

3.1 Darvaz fault

The timing, total displacement, and slip rates of the entire DF remain largely unquantified; in particular, there are few constraints on the million-year-averaged slip rate. The millennial sinistral-slip rates of the DF were mainly constrained based on Late Quaternary fault-displaced geomorphic markers. In the southern part of the DF (at ~37.8°N), the offset geomorphic features inside a pull-apart basin (i.e., 20 m of Late Holocene landforms, 120–150 m of Early Holocene terraces and alluvial fans, and 300 m of Late Pleistocene landforms) suggest a sinistral slip rate of 12–14 mm/yr (Trifonov, 1978). Farther north, at ~38.5–38.7°N, many Late Holocene dry gullies have been displaced left-laterally from a few meters to 100 m (Burtman and Molnar, 1993). Among these offsets, a 21 m displacement of a human-made wall with an inferred age of ~1,500–2,100 years suggests a sinistral slip rate of 10–14 mm/yr (Kuchai and Trifonov, 1977; Burtman and Molnar, 1993). In the same region, Nikonov (1975) reported a 3 km offset of conglomerates deposited between 2.5 and 0.7 Ma, combined with offsets of Holocene (~160 m) and Late Pleistocene (~800 m) valleys, the slip rate along this part of the DF is estimated to be 4–16 mm/yr (Nikonov, 1977; Trifonov, 1983). Farther northeast, around 39°N and 71°20'E, two moraines, considered to have formed since the Last Glacial Maximum (LGM), have been displaced sinistrally by 50 and 170 m, respectively, implying 3–4 to 8 mm/yr sinistral-transpressive slip rates (Trifonov, 1983; Burtman and Molnar, 1993). In summary, most slip rates mentioned above were derived from offset geomorphic markers lacking solid-constrained ages. Burtman and Molnar (1993) inferred that the sinistral slip rate on the DF is 10–15 mm/yr based on three relatively reliable offset features.

The previous sparse GPS data showed the present-day sinistral slip rate of ~10 mm/yr along the nearly N–S trending segment of the DF (Mohadjer et al., 2010; Ischuk et al., 2013; Zhou et al., 2016). More recently, denser GPS data have yielded rates of 8 ± 1 mm/yr and ~15 mm/yr for the ~N–S trending and NEE-trending segments, respectively (Metzger et al., 2020).

3.2 Tajik fold-and-thrust belt

The Tajik Fold-and-Thrust Belt constitutes a series of westward-convex thrust faults and folds that are generally parallel to the western margin of the Pamir salient (Chapman et al., 2017). Due to the difference in the availability of subsurface data of the Tajik Basin in previous research, the resulting diverse structural models of the TFTB imply variable magnitudes of ~E–W shortening, ranging from ~70 to ~240 km (Bourgeois et al., 1997; Burtman, 2000; Chapman et al., 2017; Gagala et al., 2020). A recent study (Gagala et al., 2020) defined the TFTB as a thin-skinned belt and showed an increase in the ~E–W

TABLE 1 Multi-timescale deformation rates of main structures along the frontal Pamir salient.

Structure	Segment	Deformation rate		Timescale	References
Darvaz Fault	Northern segment (38.4–39°N)	<i>Sinistral</i>			
		4–8		Millennial	Burtman and Molnar (1993) Trifonov (1983)
		10–14		Millennial	Burtman and Molnar (1993) Kuchai and Trifonov (1977)
		4–16		Millennial	Nikonov (1977) Trifonov (1983)
	Southern segment (36.9–38.4°N)	15		Decadal	Metzger et al. (2020)
		12–14		Millennial	Trifonov (1978)
		10 ± 1		Decadal	Ischuk et al. (2013)
		8 ± 1		Decadal	Metzger et al. (2020)
		10 ± 2		Decadal	This study
		<11.4 ± 2		Decadal	Mohadjer et al. (2010)
		10		Decadal	Zhou et al. (2016)
Tajik Fold-and-Thrust Belt	<i>Shortening</i>				
	6–9 (17 Ma)		Million-year	Gagala et al. (2020) Chapman et al. (2017)	
	8–12 (12 Ma)		Million-year	Gagala et al. (2020) Abdulhameed et al. (2020)	
	8–16		Decadal	Metzger et al. (2021)	
	6.2 ± 1 ^a		Decadal	Mohadjer et al. (2010)	
	5–7 ^a		Decadal	Zhou et al. (2016)	
	>8 ^a		Decadal	Metzger et al. (2020)	
	10 ± 2		Decadal	This study	
Pamir Thrust System	Across the MPT and PFT	<i>Shortening</i>			
		<11.8 ± 2		Decadal	Mohadjer et al. (2010)
		10–15		Decadal	Zubovich et al. (2010)
		15 ± 1		Decadal	Ischuk et al. (2013)
		11.8		Decadal	Jay et al. (2017)
		14 ± 2		Decadal	Jay et al. (2018)
	Western PFT (Alai Valley)	9 ± 2	5.6 ± 1	Decadal	This study
		0.66–0.78 (25 Ma)		Million-year	Coutand et al. (2002)
		2–4		Millennial	Burtman and Molnar (1993) Nikonov et al. (1983)
		>5.2	2.5–5.7	Millennial	Arrowsmith and Strecker (1999)
		4.1 ± 1.5		Millennial	Patyniak et al. (2021)
		2.2 ± 0.8	5.6 ± 0.8	Decadal	Zubovich et al. (2016)
	Eastern PFT (NE Pamir)	1.1–2.2 (25–20 Ma)		Million-year	Li et al. (2019a)
		4–6 (6–5 Ma)		Million-year	Li et al. (2019a)
		6.4 (0.35 Ma)		Million-year	Li et al. (2012)
		6.2 (0.34 Ma)		Million-year	Chen et al. (2011)
		5.9		Millennial	Li et al. (2012)
		7.4		Millennial	Thompson et al. (2018)
		5.6		Millennial	Thompson et al. (2017)
			4 ± 1	Millennial	Fu et al. (2010)
		6.6 ± 1.1		Decadal	Li et al. (2012)
			<i>Dextral</i>		

(Continued on following page)

TABLE 1 (Continued) Multi-timescale deformation rates of main structures along the frontal Pamir salient.

Structure	Segment	Deformation rate	Timescale	References
		6 ± 2.8	Decadal	Bufe et al. (2017)
<i>E–W extension</i>				
Kongur Shan Extensional System	Muji Fault ^b	>3–4	Millennial	Chevalier et al. (2015)
		4.5–11	Millennial	Chevalier et al. (2011)
		6–9	Millennial	Li et al. (2019b)
		11.3 ± 2.4	Millennial	Deng et al. (2020)
		8.1 ± 0.9	Decadal	Chen et al. (2011)
	Kongur Shan Fault	7	Decadal	Li et al. (2019b)
		5	Million-year	Robinson et al. (2010)
		3–4	Million-year	Robinson et al. (2007)
		3.8 ± 1.2	Millennial	Jianhong Xu et al. (2021)
		1.3–1.9	Millennial	Pan et al. (2013)
		1.6 ± 0.3	Millennial	Li (2013)
		2.1 ± 0.2	Millennial	Li (2013)
		1.1 ± 0.1	Millennial	Li (2013)
		0.8–1.8	Millennial	Chevalier et al. (2015)
		0.9 ± 0.1	Millennial	Li (2013)
	Taheman Faut	8 ± 1	Decadal	This study
		5.1 ± 0.8	Decadal	Chen et al. (2011)
	Tashkurgan Fault	2.6 ± 0.1	Millennial	Li (2013)
		1.6–3.3	Millennial	Li (2013)
		0.3–0.6	Million-year	Chen and Chen (2020)
		0.9–1.3	Millennial	Li (2013)
		0.7 ± 0.1	Millennial	Li (2013)
		<0.1	Millennial	Li (2013)
		1 ± 0.5	Decadal	This study
<i>Dextral</i>				
Kashgar-Yecheng Transfer System		11–15 (280 km, 25–18 Ma)	Million-year	Cowgill (2010)
		4–6 (110 km, 25–18 Ma)	Million-year	Chen et al. (2018)
		1.7–5.3 (5–3 Ma)	Million-year	Sobel et al. (2011)
		<2 ± 2	Decadal	Ischuk et al. (2013)
		<2	Decadal	Zhou et al. (2016)

^aUnderestimated rates due to sparse GPS data or including partial Tajik Fold-and-Thrust Belt.

^bThe E–W extension rates of the Muji fault are actually its dextral-slip rates because its strike is nearly east to west.

shortening, from ~93 km in the south to ~148 km in the north, based on analyses of rich boreholes and 2-D seismic data.

Regarding to the onset and rates of shortening across the TFTB, Chapman et al. (2017) reported the first low-temperature thermochronology data of the TFTB, indicating that its shortening started before 17 Ma. By utilizing new low-temperature thermochronology data, Abdulhameed et al. (2020) confirmed that the major shortening commenced at ~12 Ma, which is also reflected in Chapman et al.'s (2017) thermokinematic modeling results. Therefore, with the newest estimate of the amount of total shortening, the average ~E–W

shortening rates of the TFTB since ~17 Ma and ~12 Ma are ~6–9 mm/yr and ~8–12 mm/yr, respectively.

In early studies, GPS velocity profiles across the western Pamir and Tajik Basin yielded ~E–W shortening rates of ~6 mm/yr within the basin (Mohadjer et al., 2010; Zhou et al., 2016). However, recent, denser GPS data has indicated a higher shortening rate of >8 mm/yr (Metzger et al., 2020), consistent with those of ~8–16 mm/yr estimated from the InSAR rate map (Metzger et al., 2021). The denser GPS data also revealed a 2 ± 2 mm/yr extension of the DF, which cannot be detected in early research owing to the limited GPS sites. Therefore, these early

studies involved the extension of the DF when calculating the shortening rates within the basin, resulting in smaller rates.

3.3 Pamir thrust system

Deformation along the PTS (including both the MPT and PFT) is dominated by thrust faulting (with variable verging from N15°W in the west to N30°E in the east) and local dextral slip (Figure 1B) (Arrowsmith and Strecker, 1999; Sobel et al., 2013). Studies on the deformation rates have focused on the recent deformation front, the PFT, while the deformation rates of the MPT have rarely been constrained.

On the million-year timescale, structural restorations based on regional cross-sections and stratigraphy indicate two main deformation periods of the PTS (Coutand et al., 2002; Cheng et al., 2016; Li et al., 2019a). The total shortening across the Alai Valley in the northernmost Pamir was estimated to be 33–37 km, including two shortening events starting in the Late Oligocene–Early Miocene and Middle Miocene, respectively (Coutand et al., 2002). The estimated average shortening rate is 0.66–0.78 mm/yr over the past 25 Myr, which was largely underestimated, possibly due to the incomplete shortening restoration across parts of the PTS (Gągała et al., 2020). In the NE Pamir, the total shortening across the PTS decreases eastward, ranging from 43 to 29 km (Li et al., 2019a), which has also been accomplished by two key stages of shortening (Thompson et al., 2015). The first stage commenced at 25–20 Ma with average rates of 1.1–2.2 mm/yr, and the second stage started at 6–5 Ma with elevated average rates of 4–6 mm/yr (Thompson et al., 2015; Li et al., 2019a). The latter shortening rates are consistent with the rate since 0.35 Ma (6.2–6.4 mm/yr), which was constrained based on field mapping and magnetostratigraphy (Chen et al., 2011; Li et al., 2012).

On the millennial timescale, Arrowsmith and Strecker (1999) constrained the dip-slip rate of one central strand of the western PFT to >6 mm/yr (equivalent to >5.2 mm/yr shortening rate) by reconstructing the fault offset and associated chronology of a terrace at the Syrinadjar River. A recent detailed paleoseismic study on the same fault strand obtained a slightly lower rate of 4.1 ± 1.5 mm/yr over the past ~5 kyrs (Patyniak et al., 2021). This slip rate is compatible with the other determination of a minimum of 2–4 mm/yr based on the overthrusting of Stone Age artifacts farther west (Nikonov et al., 1983; Burtman and Molnar, 1993). In the vicinity of the artifacts, Arrowsmith and Strecker (1999) also estimated the dextral-slip rate across the Altyndara River to be 2.5–5.7 mm/yr based on the offset channels that were assumed to have formed 10 kyrs ago. Along the eastern PFT, a terrace surface developed in the Mayikake Basin and in the fault's footwall was displaced by ~38 and ~15 m along two fault strands, respectively. Li et al. (2012) suggested that the terrace was abandoned at ~18.4 ka based on optically stimulated

luminescence (OSL) dating and quantified PFT's shortening rate to be ~5.9 mm/yr. Thompson et al. (2018) dated the same terrace to be ~14.2 ka using a ^{10}Be depth profile, yielding a slightly higher rate of ~7.4 mm/yr. To the southwest, the PFT deformed another three flights of terraces in the hanging wall, forming fold scarps. Thompson et al. (2017) calculated PFT's total slip across these terraces based on the fault-bend fold scarp model (Li et al., 2015 and references therein). A fault slip of ~540 m across the highest terrace, with its OSL age (~97.9 ka), yielding a shortening rate of ~5.6 mm/yr. Farther southeast, the eastmost segment of the eastern PFT shows 4 ± 1 mm/yr of dextral slip, as inferred from co-seismic channel offsets and average earthquake recurrence intervals (Fu et al., 2010).

On the decadal timescale, GPS measurements have shown that present-day ~N–S shortening rates across the PTS range from 10 to 15 mm/yr (Zubovich et al., 2010; Ischuk et al., 2013; Zhou et al., 2016). The rates are consistent with the modeling rates (11.8–14.4 mm/yr) from a kinematic model based on recent GPS velocities and Quaternary fault slip rates (Jay et al., 2017; Jay et al., 2018). The velocities from four GPS stations across one strand of the western PFT indicated a shortening rate of 2.2 ± 0.8 mm/yr and dextral-slip rate of 5.6 ± 0.8 mm/yr near the Altyndara River (Zubovich et al., 2016). The latter rate is consistent with the estimated millennial rate. In the eastern PFT, Li et al. (2012) obtained a minimum shortening rate of 6.6 ± 1.1 mm/yr by analyzing previous GPS velocities (Yang et al., 2008; Zubovich et al., 2010), consistent with the InSAR observed rate of 6 ± 2.8 mm/yr (Bufe et al., 2017).

3.4 Kongur Shan extensional system

In comparison with the fault systems mentioned above, previous studies have made more substantial progress to quantify the kinematics of the KSES in the NE Pamir. Here, we summarize the published deformation rates of the KSES's four major fault segments: the Muji, Kongur Shan, Taheman, and Tashkurgan Faults.

3.4.1 Muji fault

The WNW-trending Muji Fault belongs to the northernmost portion of the KSES, connecting with the Kongur Shan Fault in the east and nearly intersecting the MPT in the west. The fault can be divided into two segments based on its along-strike variations in slip sense: the eastern segment with dominant dextral slip and a significant normal-faulting component and the western segment characterized by pure dextral slip or dextral slip with a reverse-faulting component (Li et al., 2019b).

Along the Muji Fault, various landforms displaced by dextral slip provide opportunities to determine the millennial slip rates of the fault's different segments. In the west segment at 39.272°N, 73.989°E, the dextral-slip rate was estimated to be > 3–4 mm/yr

based on three offset channels that were assumed to be preserved during the maximum glacial advances at 8–10, 16, and 50–80 ka, respectively (Chevalier et al., 2015). In the eastern segment near the Akesayi village, Chevalier et al. (2011) constrained the dextral-slip rate to ~ 4.5 – 11 mm/yr based on fault displacements across fluvial terrace risers and associated cosmogenic ^{10}Be ages. Li et al. (2019b) dated the same terraces again using ^{10}Be depth profiles and added a terrace's surface age, refining the rate to ~ 6 – 9 mm/yr. Approximately 20 km farther east, the dextral slip of the Muji Fault increases to 11.3 ± 2.4 mm/yr, based on the offsets and ages of moraines along the fault (Deng et al., 2020).

The present-day lateral slip rate of the Muji Fault is ~ 7 – 8 mm/yr near Akesayi, as estimated from the published GPS velocity data (Yang et al., 2008; Zubovich et al., 2010; Zhou et al., 2016) by Chen et al. (2011) and Li et al. (2019b). The accuracy of the decadal-slip rate could be improved in the future when denser GPS sites are available in this region.

3.4.2 Kongur Shan fault

The Kongur Shan Fault generally trends NNW, dipping moderately (35° – 45°) to the west and southwest (Robinson et al., 2004). It is characterized by dominant normal slip and minor dextral slip (Robinson et al., 2004; Chevalier et al., 2015).

On the million-year timescale, the thermo-kinematic modeling results showed that the fault's constant rates of dip-slip and E–W extension along the Kongur Shan Dome have been 6.5 and 5 mm/yr, respectively, since ~ 7 Ma (Robinson et al., 2010). To the south, Robinson et al. (2007) reported the E–W extension along the fault west of the Muztaghata Dome, ~ 20 km since 6–5 Ma, yielding an average extension rate of 3–4 mm/yr.

On the millennial timescale, fault offsets reconstructed from an alluvial fan (38.912°N , 74.785°E) north of the Kongur Shan and associated ^{10}Be surface exposure ages, suggest a Holocene vertical slip rate of 1.8–2.6 mm/yr (Pan et al., 2013). This slip rate is consistent with that obtained from an adjacent fluvial terrace (2.2 ± 0.7 mm/yr, Xu J. et al., 2021). The latter vertical rate corresponds to an E–W extension rate of 3.8 ± 1.2 mm/yr, for a fault dip of 30° (Robinson et al., 2004). In the southern piedmont of the Muztaghata Dome, a landslide body formed at ~ 7 ka (Yuan et al., 2013) has been vertically displaced by ~ 15 m, yielding a vertical slip rate of 2.2 ± 0.4 mm/yr; the E–W extension rate was determined to be 1.6 ± 0.3 mm/yr based on the fault strike (330°) and dip (50°) (Li, 2013). The adjacent debris flow fan also shows a dextral-slip rate of 0.9 ± 0.2 mm/yr for this fault segment (Chevalier et al., 2015). About 3 km farther southeast, the fault displaced an ~ 62 ka-old moraine by ~ 98 m, providing the vertical-slip and E–W extension rates of 1.8 ± 0.2 and 2.1 ± 0.2 mm/yr, respectively (Li, 2013). South of the Kongur Shan Fault's junction with the Taheman Fault (37.982°N , 75.255°E), the fault's vertical slip rate is ~ 1.7 – 2.3 mm/yr (Li, 2013; Chevalier et al., 2015), based on alluvial fan scarp offsets (~ 35 m) and chronology (~ 15 – 20 ka). The corresponding E–W

extension rate is ~ 1.1 – 1.5 mm/yr with fault dip of 40° and strike of 304° . Farther south, the activity of the Kongur Shan Fault has been significantly reduced; its vertical-slip and E–W extension rates reach 0.7 ± 0.1 and 0.9 ± 0.1 mm/yr, respectively (Li, 2013).

On the decadal timescale, the velocity analysis from sparse GPS stations indicated that the E–W extension rate of the Kongur Shan Fault generally decreases from north to south (Zhou et al., 2016; Li et al., 2019b). Specifically, the E–W extension rate west of the Kongur Shan is 5.1 ± 0.8 mm/yr (Chen et al., 2011), consistent with the million-year-averaged rate.

3.4.3 Taheman fault

The ~ 20 -km-long Taheman Fault has been interpreted as a transfer fault, linking the Kongur Shan and Tashkurgan Faults (Robinson et al., 2007). This transfer fault can be divided into two segments based on their orientations: the ~ 13 -km-long northern segment striking $\text{N}25^\circ\text{E}$ and the ~ 7 -km-long southern segment striking $\text{N}5^\circ\text{E}$ (Robinson et al., 2007).

To date, only millennial slip rates of the Taheman Fault have been published, constrained from two offset sites in alluvial fans on the northern segment (Li, 2013). The first alluvial fan (37.998°N , 75.235°E) is cut by multiple step-like fault scarps containing coseismic scarps of the 1895 Tashkurgan M7.0 earthquake (Li et al., 2011). The total offsets (~ 32 – 37 m) and the fan's exposure age (~ 15 ka) together indicate that the fault's vertical-slip and E–W extension rates are 2.2–2.7 and 2.2–3.1 mm/yr, respectively. The second alluvial fan (37.981°N , 75.219°E) exhibits a vertical offset of ~ 22.6 m over the past ~ 7 – 15 kyr, suggesting the vertical-slip and E–W extension rates of 1.5–3.1 and 1.6–3.3 mm/yr, respectively.

3.4.4 Tashkurgan fault

As the southernmost segment of the KSES, the Tashkurgan Fault strikes NNW and dips to the east. The low-temperature thermochronological analysis indicated that the fault's footwall has been exhumed at an average rate of ~ 0.6 – 0.9 mm/yr since ~ 6 – 5 Ma (Chen and Chen, 2020). For the fault dipping 60° (Robinson et al., 2007) and striking 162° , the million-year-averaged E–W extension rate of this fault is ~ 0.3 – 0.6 mm/yr.

The millennial fault slip rates were constrained by offset moraines at three sites from north to south (Li, 2013). At the mouth of the Kuzigun valley (37.652°N , 75.233°E), the vertical offsets (~ 19.6 – 33.1 m) and the ^{10}Be exposure ages (~ 19 – 24 ka) of the moraines yield the vertical-slip and E–W extension rates of ~ 1 – 1.4 and ~ 0.9 – 1.3 mm/yr, respectively. Farther south, at the mouth of the Jialongqiete valley (37.550°N , 75.263°E), the moraines that formed ~ 19 kyr ago have been offset by ~ 11.5 m, yielding the vertical-slip and E–W extension rates of ~ 0.6 and ~ 0.7 mm/yr, respectively. In the southernmost portion of the fault (37.256°N , 75.382°E), an isolated moraine complex with a ^{10}Be exposure age of ~ 65 ka has been vertically offset by ~ 8 m. These data suggest that the vertical-slip and E–W

extension rates are 0.1 and <0.1 mm/yr, respectively. In general, the E–W extension rates of the Tashkurgan Fault decrease from north to south, consistent with the trend from the GPS velocity data (Zhou et al., 2016; Li et al., 2019b).

3.5 Kashgar–Yecheng transfer system

The total dextral-slip offset of the KYTS remains debated under two end-member models (see descriptions in Section 1). Cowgill (2010) estimated the cumulative dextral-slip offset to be ~280 km based on the assumption that the North Pamir and the western Kunlun were situated at a linear belt before the Cenozoic indentation of the Pamir. The million-year-averaged slip rate was thus estimated to be ~11–15 mm/yr since its initiation at 25–18 Ma (Cowgill, 2010). However, the magnitude of the dextral-slip offset is much lower (<110 km) in the inherited-arc model (Chen et al., 2018), yielding a lower million-year-averaged rate of ~4–6 mm/yr. Recent thermochronological data and geomorphic observations suggest that the KYTS has been inactive, except for the Aertashi strand, with an average rate of 1.7–5.3 mm/yr since at least 5–3 Ma (Sobel et al., 2011). This rate range brackets the present-day rate of relative motion between the eastern Pamir and Tarim block (~2 mm/yr), as revealed by GPS velocity analyses (Ischuk et al., 2013; Zhou et al., 2016).

4 Present-day deformation of the Pamir salient

To reveal the bulk pattern of the present-day deformation field of the Pamir salient, we collected and reanalyzed all previously published GPS velocity data and instrumental seismicity recorded from 1976 to 2021 (Global Centroid Moment Tensor Catalog; <https://www.globalcmt.org/>) within and around the Pamir.

4.1 GPS velocity field

4.1.1 GPS velocity data acquisition and analysis

The published GPS data in the Pamir and its surrounding regions (34–42°N, 66–79.4°E) (Zubovich et al., 2010; Li, 2012; Ischuk et al., 2013; Jouanne et al., 2014; Zhou et al., 2016; Metzger et al., 2020; Wang and Shen, 2020) were acquired using the updated dataset for the same sites. Most of the GPS velocities are based on the campaign mode, whereas the measurements of the continuous stations span at least 3 years. Because some of these velocity datasets are referenced to different reference frames, we combined them in the International Terrestrial Reference Frame 2008 (ITRF 2008, Altamimi et al., 2012) through an application of the rigid body rotation to each set using the VELROT program of GAMIT/GLOBK software (Herring et al., 2010). Then, we

transformed the GPS velocity vectors from ITRF2008 to the Eurasia-fixed reference frame using the rotation Euler pole (54.23°, –98.85°, 0.257/Myr). Finally, the GPS velocities of 362 stations formed the integrated GPS velocity field relative to Eurasia (Figure 3; data are provided in the Supplementary Material).

To constrain the deformation rates of the main structures in the Pamir, we selected 91 GPS stations that constitute four swath profiles: Swath 1 across the western Pamir and the Tajik Basin; Swath 2, crossing from the northern margin of the southwesternmost Tien Shan to the southern Pamir; Swath 3 across the northern KSES; and Swath 4 from the western Pamir to the southwestern Tarim Basin (Figure 3). The horizontal GPS velocity vectors within the four swaths were separately projected onto three directions in parallel with the swaths or perpendicular to the strikes of the main structures (Figure 3). In detail, the GPS velocity vectors in Swath 1 were projected on profile A–A' striking 95°, those in Swath 2 were projected on profile B–B' striking 175°, and those in Swaths 3 and 4 were projected on profile C–C' striking 82°.

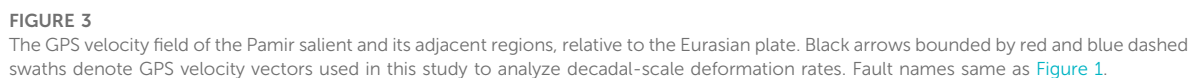
4.1.2 Results

Two key kinematic patterns of the Pamir salient can be observed in the GPS velocity field. First, the GPS velocity vectors show bulk northward movement accompanied by an anticlockwise rotation of the Pamir salient (Figure 3), as has been noted in previous studies (e.g., Zhou et al., 2016; Pan et al., 2018). Second, the GPS velocity vectors radiate asymmetrically from the interior of the Pamir to its foreland. The vector directions change anticlockwise from NNW to NW or WNW in the west, while they rotate clockwise to NNE in the east with a much smaller magnitude of rotation.

The three GPS profiles reveal the present deformation of the Pamir salient. Along profile A–A', the velocities parallel to the profile can be divided into two groups within the Tajik Basin: the -15.5 ± 1.3 mm/yr in the east of the TFTB and the -5.4 ± 0.5 mm/yr in the west (Figure 4A). The difference between the two groups suggests 10 ± 2 mm/yr of shortening of the TFTB. The velocities perpendicular to the profile decrease from 13.6 ± 1.5 mm/yr in the western Pamir to 3.9 ± 0.5 mm/yr in the Tajik Basin (Figure 4B), indicating the sinistral slip of 10 ± 2 mm/yr between the Pamir and the Tajik Basin, mainly accommodated by the DF.

Along profile B–B', the velocities parallel to the profile increase significantly across the PTS from north to south (Figure 5A), indicating a crustal shortening of 9 ± 2 mm/yr. Within the South Pamir, the velocities exhibit a decreasing trend from north to south, suggesting a 1.8 ± 2 mm/yr N–S extension rate. The velocities perpendicular to the profile on both sides of the PTS show an obvious difference (Figure 5B), representing a dextral slip rate of 5.6 ± 1 mm/yr between the Pamir and Alai valley, which is mainly accommodated by the PTS.

Along profile C–C', the velocities parallel to the profile from Swath 3 indicate that the E–W extension rate of the northern



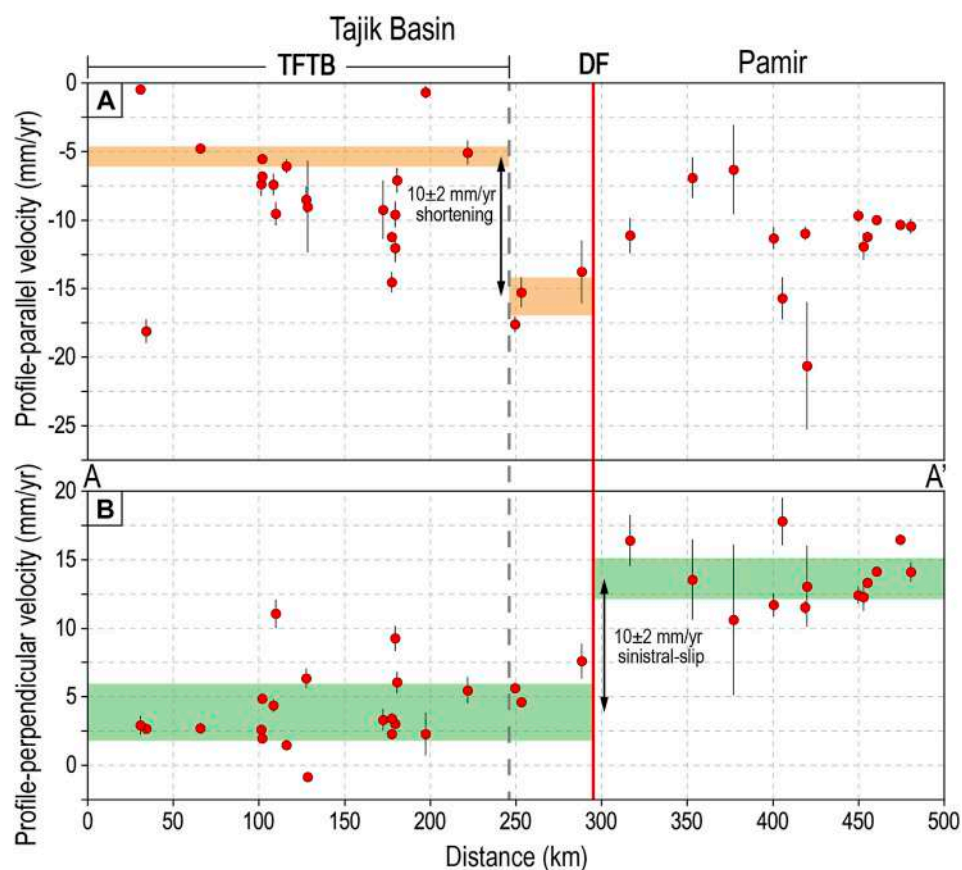


FIGURE 4

GPS velocity profile A–A' (see location in Figure 3). (A) Velocities parallel to the profile. (B) Velocities perpendicular to the profile. Colored stripes show velocity ranges. The dark grey dashed line and bold red line mark the tectonic domain boundary and fault location, respectively. Fault names same as Figure 1.

northernmost to the southernmost part of the Pamir. The earthquake clustering is conspicuous in the South Pamir. These nearly N–S striking earthquakes related to normal faulting occurred in the middle and upper crust according to their depths, but there appears to be no significant evidence of normal faulting or extensional activity on the surface. In contrast, the main extensional structures in the northern and northeastern Pamir, such as KG and KSES, generated fewer normal fault earthquakes.

5 Discussion

5.1 Spatiotemporal variations of deformation rates along the frontal Pamir salient

The above summary of multi-timescale deformation rates of the main structures along the frontal Pamir salient allows for further analysis of the spatiotemporal variations in fault

kinematics. To do this, we projected the rates onto an arcuate profile along the margin of the Pamir salient (Figure 8).

The strike-slip rates of each segment have remained stable since the Late Quaternary, and show an anticlockwise decrease along the arcuate profile (Figure 8B). The millennial and decadal left-slip rates of the main DF are generally consistent, within the range of 10–15 mm/yr (Burtman and Molnar, 1993). The millennial slip rate decreases to 4–8 mm/yr at the east end of the DF (Trifonov, 1983), indicating that the deformation along the DF zone may be partly accommodated by the thrusting at the location with changes in fault strikes (see Section 4.2). The millennial dextral-slip rate along the northern Pamir is also consistent with decadal rates of 4–5 mm/yr (Arrowsmith and Strecker, 1999; Zubovich et al., 2016). Along the eastern margin, the dextral-slip rate of the KYTS has remained at <2–4 mm/yr since 5–3 Ma (Fu et al., 2010; Sobel et al., 2011; Ischuk et al., 2013; Zhou et al., 2016). This rate is lower than the fault's average dextral-slip rate since its initiation at 25–20 Ma, at 4–6 or 11–15 mm/yr, depending on different end-member models

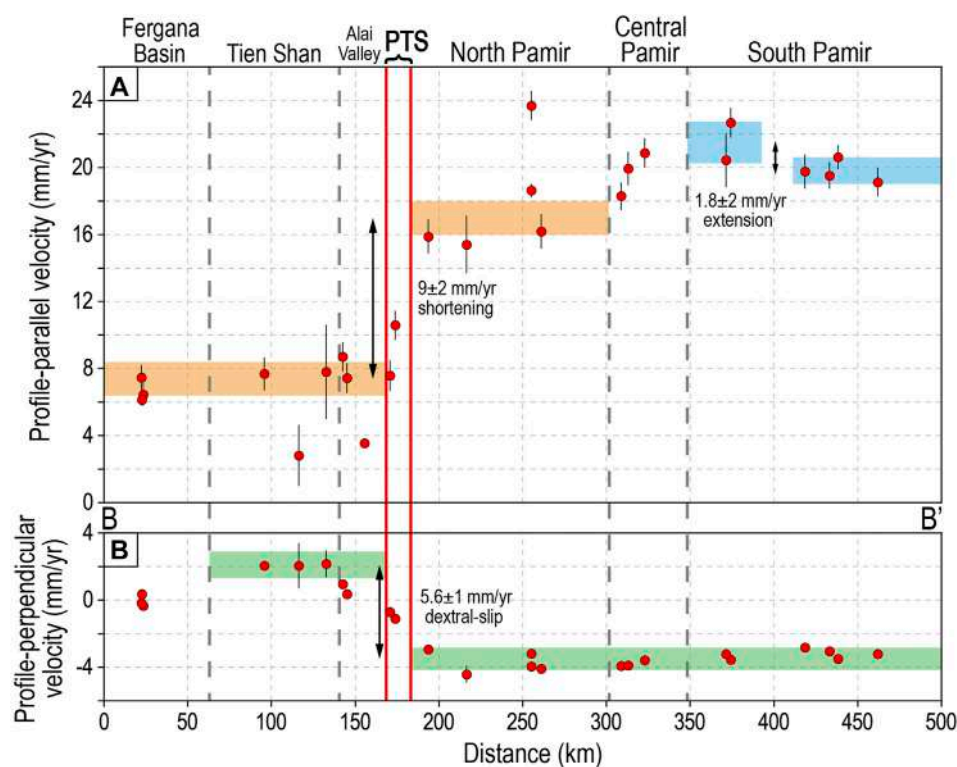


FIGURE 5

GPS velocity profile B–B' (see location in Figure 3). (A) Velocities parallel to the profile. (B) Velocities perpendicular to the profile. Colored stripes show velocity ranges. The dark grey dashed lines and bold red lines mark the tectonic domain boundaries and fault locations, respectively. PTS: Pamir Thrust System. Other fault names same as Figure 1.

(see Section 3.5), suggesting a decrease in the dextral-slip rate. Such observation is consistent with thermochronologic and geomorphic evidence that also suggests the deceleration of the KYTS between the Late Miocene and Pliocene (Sobel et al., 2011).

The shortening rates along the western and northern Pamir were relatively lower during the early deformation stage of each structure, then experienced a significant increase, and have remained roughly constant until the present (Figure 8C). Spatially, the million-year-averaged and decadal shortening rates within the Tajik Basin both show a northward increasing trend. Temporally, the shortening accelerated during the Middle Miocene: the average rates since the onset of shortening at ~17 Ma range from 6 to 9 mm/yr (Chapman et al., 2017; Gałala et al., 2020), while the rates after ~12 Ma are larger, within the range of 8–16 mm/yr (Abdulhameed et al., 2020; Gałala et al., 2020; Metzger et al., 2021). Consistency can be seen between the million-year-averaged rates since ~12 Ma and decadal rates when excluding the underestimated rates (transparent blue diamonds in Figure 8C; Section 3.2). Despite lacking millennial rates of the TFTB, considering the constant slip rate since the Late Quaternary of the DF, which has been

experiencing similar kinematics in the western Pamir (Schurr et al., 2014; Metzger et al., 2020), we suggest that the shortening rate of the TFTB has remained stable at 8–16 mm/yr since ~12 Ma. The average N–S shortening rate on the northern margin has been only ~0.7 mm/yr during the past 25 Myr (Coutand et al., 2002), much smaller than current rates of 10–15 mm/yr (e.g., Zubovich et al., 2010). The millennial rates of a branch of the western PFT are 3–5 mm/yr (Arrowsmith and Strecker, 1999; Patyniak et al., 2021), consistent with the decadal rate of 3.8 ± 0.8 mm/yr for the same segment (Zubovich et al., 2016). Based on the relatively uniform slip rate of this branch fault, the millennial rate on the northern margin can be reasonably extrapolated as also 10–15 mm/yr based on the same context of thrust faulting along the northern margin. The jump in the rate between the million-year and millennial timescales may have begun after the Middle Miocene (Coutand et al., 2002). Along the northeastern margin, the average shortening rate since 25 Ma is 1.7–2.2 mm/yr, followed by an increase to 4–6 mm/yr since 6–5 Ma (Li et al., 2019a); the latter is compatible with the eastern PFT's rate of ~5–7 mm/yr since the Quaternary (e.g., Li et al., 2012; Thompson et al., 2017; Thompson et al., 2018).

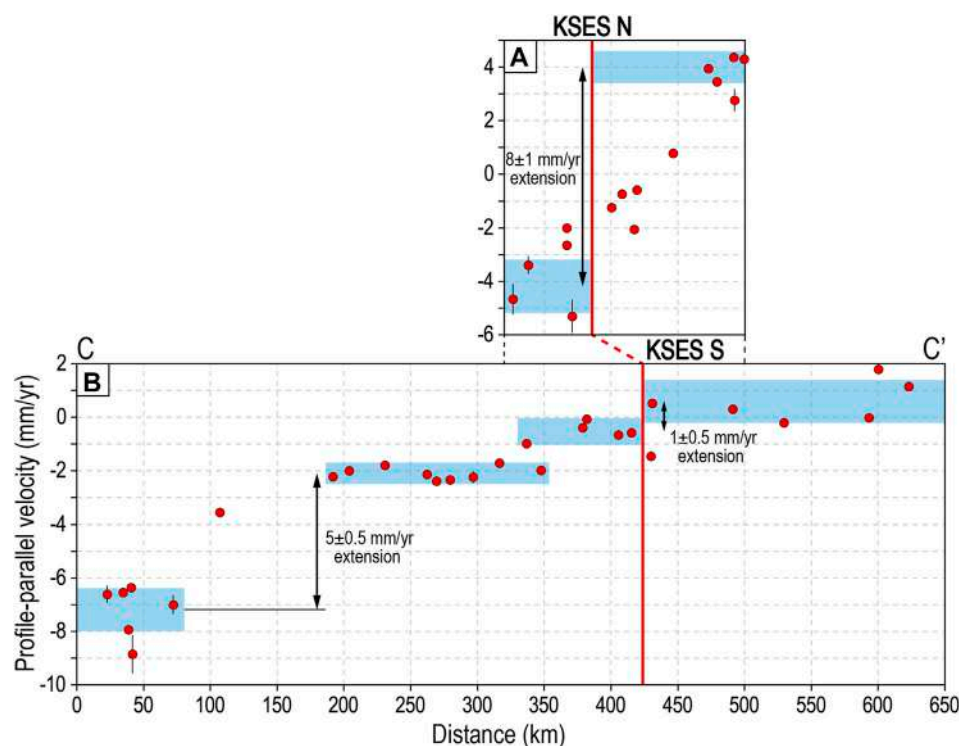


FIGURE 6

GPS velocities parallel to the profile C–C' (see location in Figure 3). (A,B) show velocities in swaths 3 and 4, respectively. Colored stripes show velocity ranges. The red bold line denotes the fault location of the Kongur Shan Extensional System.

The E–W extension rates of the KSES at three timescales all show an overall decrease from north to south, with similar trends for the change in rates (Figure 8C). Constraints on million-year-averaged extension rates only exist at several localities near the Kongur Shan Dome (~5 mm/yr, Robinson et al., 2010), Muztaghata Dome (~3.5 mm/yr, Robinson et al., 2007), and northern Tashkurgan Fault (~0.5 mm/yr, Chen and Chen, 2020). In contrast, more abundant millennial rates, despite the gap in data of the middle segment of the KSES, show an overall rapid decrease in extension rates from ~7 to 10 mm/yr on the Muji Fault (Chevalier et al., 2011; Li et al., 2019b) to ~3.8 mm/yr on the northern Kongur Shan Fault (Xu J. et al., 2021), and then a gradual, southward decrease to ~0.1 mm/yr on the southernmost Tashkurgan Fault (Li, 2013). The decadal rates are limited due to the sparse GPS stations installed in the region but are consistent with those on the other two timescales. The decadal rate of the central Muji Fault is 7–8 mm/yr (Chen et al., 2011), consistent with the millennial rate of 6–9 mm/yr (Li et al., 2019b), and the rate near the Kongur Shan Dome is the same as the million-year rate, at ~5 mm/yr. In the southern KSES, the decadal rate of ~1 mm/y is compatible with the millennial and million-year rates.

In summary, the limited deformation rate data indicates two deformation stages since the Late Oligocene–Early Miocene. The kinematics of most structures along the frontal Pamir salient have remained relatively stable since the Middle–Late Miocene.

5.2 Middle–late Cenozoic tectonic evolution of the Pamir salient and geodynamic processes

The multistage deformation pattern (see Sections 4 and 5.1) denotes various kinematic models of the Pamir salient since the Late Oligocene–Early Miocene, which should have been driven by coeval geodynamic processes. Combined with the burial and exhumation history of the Cenozoic gneiss domes and deep structure under the Pamir, we depict the tectonic evolution of the Pamir salient as three main stages during the Middle–Late Cenozoic, with two tectonic transformations.

5.2.1 Stage 1: ~37–24 Ma

Following the breakoff of the Neo-Tethyan oceanic lithosphere at ~45 Ma, the subduction of the buoyant Indian continental lithosphere forced the compression within the Pamir, and the

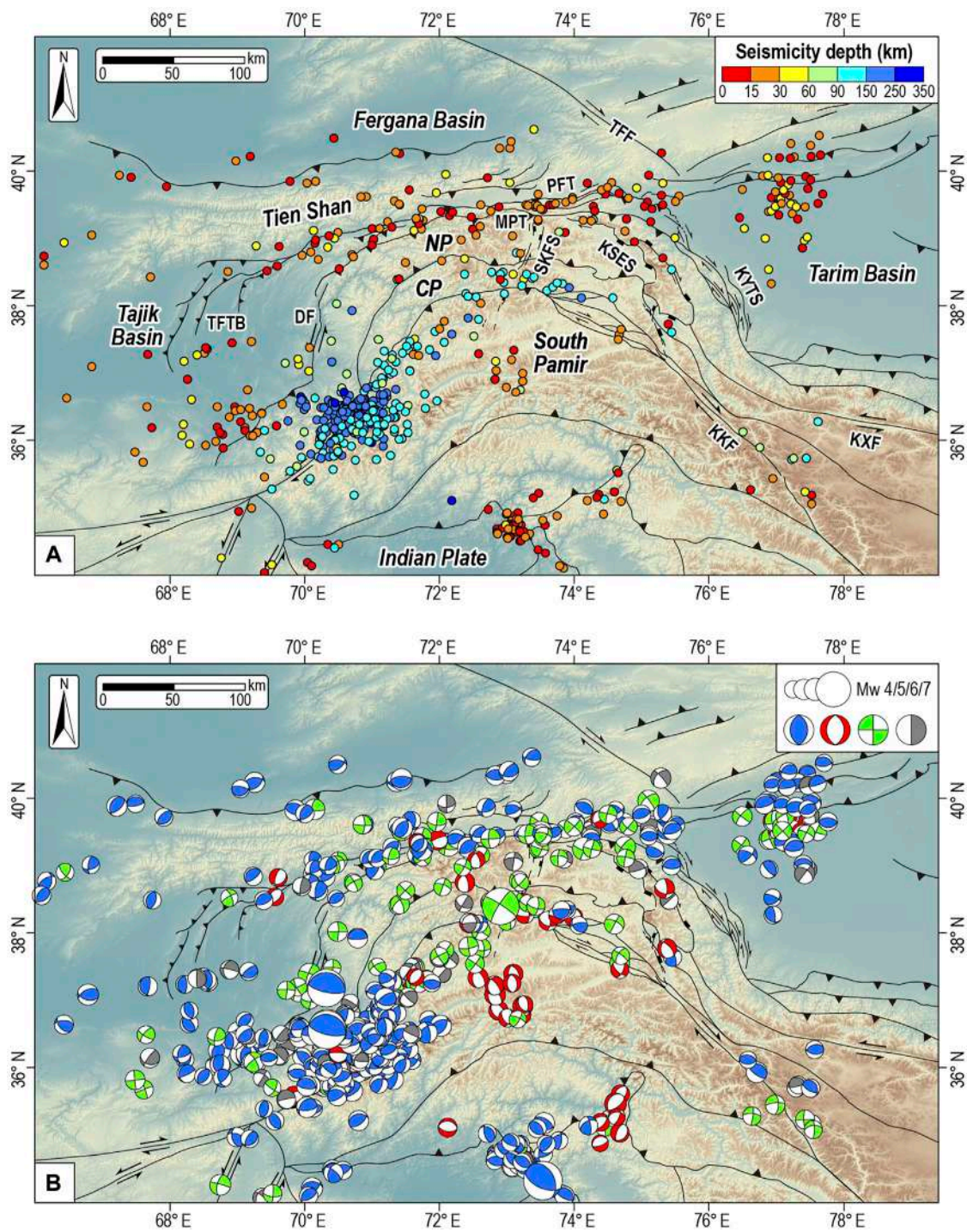


FIGURE 7

Seismicity of the Pamir salient and its adjacent regions. (A) Earthquake focal depth. (B) Focal mechanisms. Blue, red, and green beachballs represent the thrust, normal-fault, and strike-slip earthquakes, respectively. The earthquake data are from Global Centroid Moment Tensor Catalog (GCMT). Fault names same as Figure 1. NP, North Pamir; CP, Central Pamir.

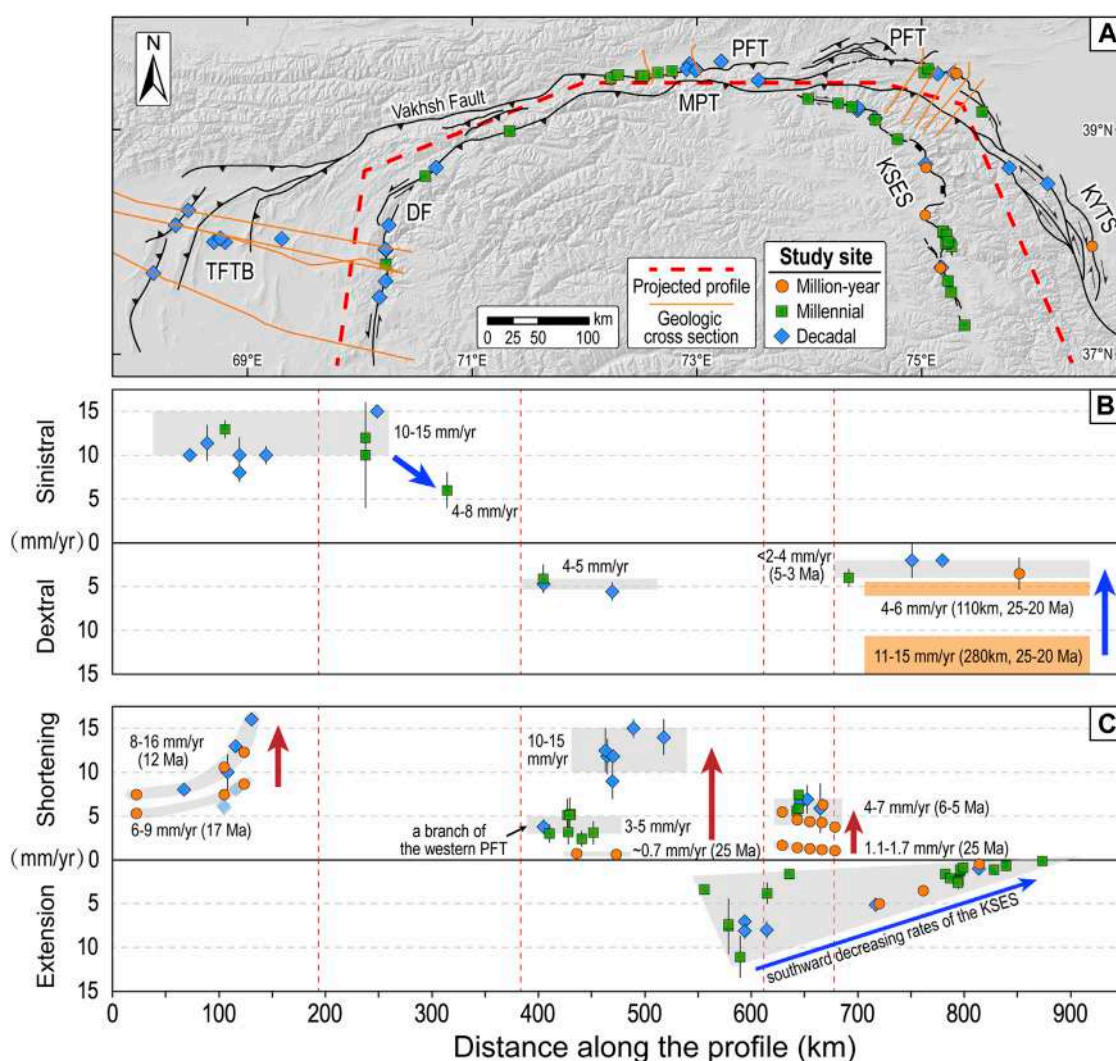


FIGURE 8

Projections of deformation rates at three timescales along the arcuate front of the Pamir salient (modified from Ge et al. (2022)). (A) Study locations of geologic cross-sections (thin, orange lines) and deformation rates at million-year, millennial and decadal timescales. (B) Deformation rates parallel to the profile denote strike-slip rates. (C) Deformation rates perpendicular to the profile denote shortening and extensional rates. The direction of extension rates is east to west. In Figures (B,C), the red dotted lines represent the turning points of the projected profile; the grey shadows show approximate ranges or trends of rates; the brownish-red and blue arrows show increasing and decreasing trends of rates across different timescales, respectively. The transparent blue diamonds in Figure C are the underestimated decadal rates of the TFTB. Fault names see Figure 1.

sinking oceanic slab induced the asthenospheric upwelling beneath the Karakoram–Pamir region (Figure 9A) (Negredo et al., 2007; Mahéo et al., 2009). The consequent intense and sustained crustal thickening in the Central and South Pamir, together with heating by the asthenospheric heat flow, caused the prolonged metamorphism in the middle–lower crust during ~37–20 Ma (Section 2.2). Meanwhile, the subduction of the Indian plate drove the northward expansion of the Pamir salient. Paleomagnetic studies have observed the anticlockwise rotation of the Tajik Basin and the clockwise rotation along the southwestern Tarim Basin that occurred

during the Late Eocene to Late Oligocene (Bosboom et al., 2014 and references therein). The rotations on both sides of the nascent Pamir salient indicate that the deformation propagated from the Pamir towards the forelands by radial thrusting (Figure 9A) (Bosboom et al., 2014).

5.2.2 Stage 2: ~23–12 Ma

When the subducted Indian continental lithosphere broke off during ~25–20 Ma (Replumaz et al., 2010; DeCelles et al., 2011), the remainder was released from the slab pull force, and

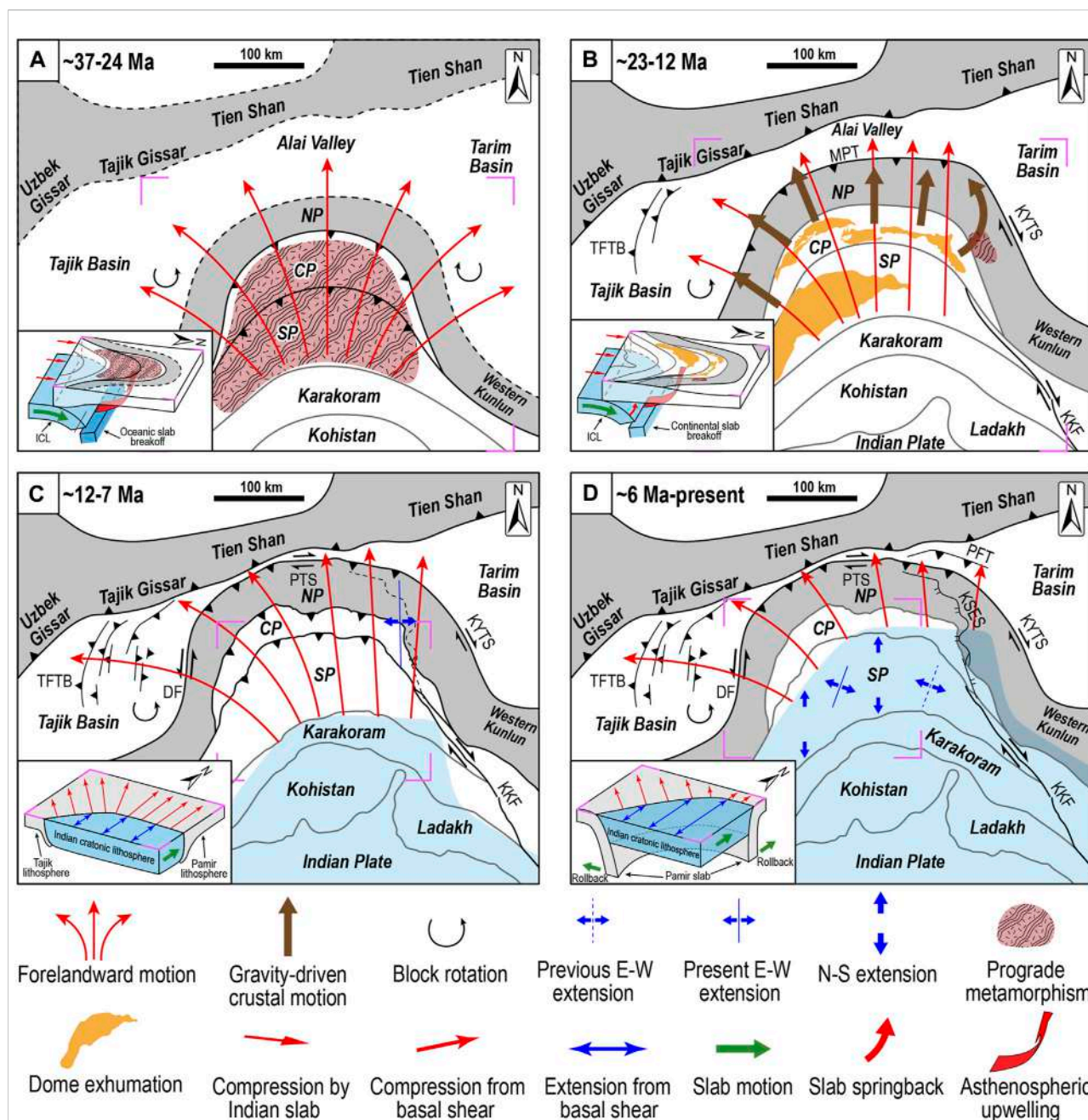


FIGURE 9

Tectonic evolution and associated geodynamic processes of the Pamir salient since the Middle–Late Cenozoic. Insets show schematic diagrams of geodynamic processes within pink boxes in corresponding plane graphs, based on Stearns et al. (2015), Kufner et al. (2016), and Rutte et al. (2017b). Insets in Figures (C,D) illustrate interactions between the Indian mantle lithosphere (blue-shaded polygons in plane graphs) and the Pamir slab. (A) ~37–24 Ma: after the breakoff of the Neo-Tethyan oceanic lithosphere, the subduction of the buoyant Indian continental lithosphere (ICL) pushed the Pamir salient northward via radial thrusting, causing block rotations in the Tarim and Tajik basins; the crust in the Central Pamir (CP) and South Pamir (SP) experienced intense thickening and prograde metamorphism. (B) ~23–12 Ma: the breakoff of the ICL caused asthenospheric upwelling and spring-back of the remainder, inducing the gravitational collapse in the CP and SP; the movement of the gravity-driven upper crust caused the migration of the deformation front to the North Pamir (NP) and kinematic transformation in the eastern Pamir into the large-scale dextral slip. (C) ~12–7 Ma: the stress effects of basal shear imposed by the underthrusting of the cratonic ICL have governed the kinematics of the Pamir salient since ~12 Ma; due to impedance of the rigid Tarim block in the east and the irregular shape of the Indian promontory, the compression induced by basal shear facilitated the foreland-ward expansion of the Pamir by asymmetric radial thrusting, generating ~E–W extension in the northeastern Pamir that formed the KSES. (D) During ~6 Ma–present: as the underthrusting continued, the earlier distributed deformation concentrated in the northern margin and propagated northward, forming the PFT; the extension resulting from basal shear and rollback of the Pamir slab also began to affect the SP, showing westward migration of the E–W extension. Fault names same as Figure 1.

asthenospheric materials upwelled along the detachment zone (Figure 9B). These would not only cause a rapid topographic rise that built up local gravitational potential energy (GPE) in the South and Central Pamir, but also heat and weaken the crust. Given the existing thickened crust during the Oligocene in the two regions, the excess GPE generated a large gradient between the Pamir and its foreland areas. In addition, the lithosphere of the South and Central Pamir had been rheologically weakened before by a long history of subduction, arc formation, accretion, and tectonism (e.g., Schwab et al., 2004), and the asthenospheric heating further weakened their crust. These factors favor the gravitational collapse in the Central and South Pamir (Rey et al., 2010). There are two fundamental modes of gravitational collapse called “fixed-boundary collapse” and “free-boundary collapse”; the “fixed-boundary collapse” mode can be classified into three types based on various behaviors of crustal deformation accommodating gravitational collapse, including “upper crustal deformation only,” “lower crustal flow only,” and “both upper crustal deformation and lower crustal flow” (Rey et al., 2001).

Two lines of evidence point to the “both upper crustal deformation and lower crustal flow” mode of gravitational collapse in the South and Central Pamir. First, the extensional exhumation of the Central and South Pamir domes was synchronous with the onset of the shortening in the forelands. The exhumation of the domes started at ~23–20 Ma (Section 2.2), while the MPT in the north and TFTB in the west initiated at ~25–20 Ma and >17 Ma, respectively (Sections 2.1, 3.2). Second, many observations of ductile deformation in the detachment faults bounding the domes indicate abundant partial melting of the middle–lower crust (Stübner et al., 2013a; Rutte et al., 2017b; Worthington et al., 2020). In this mode of gravitational collapse, materials from the upper crust and the lower crustal flow are transferred towards the forelands driven by the GPE, contributing to the lateral growth of the orogen (Rey et al., 2001). The upper crustal materials moving towards the forelands would cause the shortening of the forelands and the extension of the orogen, which affects the horizontal deformation of the upper crust much more than the lower crustal flow (Rey et al., 2001 and references therein). Therefore, here we focus on the effects of the movement of upper crustal materials during gravitational collapse on the kinematics of the Pamir salient.

The different fault kinematics along the frontal Pamir during this period imply that the gravity-driven upper crust tended to move towards the northern and western forelands. Within the Tajik Basin in the west, the TFTB initiated before 17 Ma and absorbed ~6–9 mm/yr of shortening (Section 3.2). The initiation of the MPT at ~25–20 Ma (Section 2.1) and the accelerated exhumation in the central North Pamir during ~25–16 Ma (Amidon and Hynek, 2010) suggest the shortening in the northern foreland. During the initiation of the MPT, the shortening rate on the northern margin was low (<1–2 mm/yr, Section 5.1), consistent with the limited exhumation of the

North Pamir (Sobel et al., 2013). We ascribe this low rate of shortening to the fact that the major shortening was still accommodated by crust thickening around the Central–South Pamir, as reflected by syn-extensional peak metamorphism of the Central and South Pamir domes during ~22–19 Ma, and long-lasting prograde metamorphism of the Muztaghata Dome until ~12 Ma (Section 2.2). However, for the Tarim Basin in the east, most (~94–100%) of the Cenozoic shortening deformation is concentrated along its margin (Laborde et al., 2019). Moreover, the KYTS, as the boundary fault between the Pamir and Tarim Basin, was transformed from thrusting to major dextral slip at ~25–18 Ma (Cowgill, 2010; Bosboom et al., 2014). This means that few materials from the upper crust were transferred into the eastern foreland. From the extension in the interior of the Pamir, the scheme is consistent with the dominant N–S extension accompanying a dextral shear component of shear zones bounding the Central and South Pamir domes (Stübner et al., 2013a; Rutte et al., 2017b), and also matches the asymmetric distribution and arcuate shape of these domes (Figure 1B).

The strength of the surrounding upper crust controls the movement of the gravity-driven upper crust (Rey et al., 2001). The crustal materials will be stopped or forced in another direction if the GPE fails to overcome the compressive strength of the surrounding upper crust. This can explain why the collapsed upper crust in the Pamir tended to move northward and westward. The cratonic Tarim Basin is a rigid block with high compressive strength, which would hinder the orogenic expansion. In contrast, the upper crust of the northern foreland and Tajik Basin is relatively weak, thus favoring the foreland-ward movement of crustal materials. In particular, the Tajik Basin has a weak, upper Jurassic evaporite detachment layer (e.g., Gağala et al., 2020).

The asymmetric foreland-ward movement of the upper crust during the gravitational collapse could have transformed the kinematics of the Pamir salient after ~23 Ma. The northward and westward movement of the upper crust sustained and may have strengthened the radial thrusting in the western and northern Pamir (Figure 9B), causing the continued rotation within the Tajik Basin (e.g., Thomas et al., 1994). The gravity-driven upper crust that was forced to move northward along the eastern boundary (Figure 9B), together with the overall northward movement of the Pamir salient, may have caused the decoupling of the eastern Pamir from the Tarim Basin. This decoupling, therefore, led to the tectonic transformation from radial thrusting to dextral-slip faulting along the eastern margin of the Pamir. Therefore, the rotation along the eastern flank of the Pamir mostly ceased at ~24 Ma (Bosboom et al., 2014), but a major dextral-slip fault system (KYTS) developed at 25–18 Ma (see Section 3.5).

5.2.3 Stage 3: ~12 Ma–present

After the breakoff of the Indian continental lithosphere during ~25–20 Ma, the western passive margin of India (Marginal India) was separated from Cratonic India and has been subducting under the Hindu Kush, while the cratonic

Indian lithosphere (Cratonic India) switched from subduction to underthrusting (Kufner et al., 2016; Kufner et al., 2018). At ~12–11 Ma, the cratonic Indian lithosphere collided with the Pamir slab (Kufner et al., 2016), i.e., the Pamir and Tajik lithospheres (Figure 9C).

Geodynamic modeling shows that the underthrusting Indian lithosphere would impose basal shear on the overlying plate, resulting in two stress regimes: 1) extension above the underthrusting Indian lithosphere due to the counteraction between basal shear and horizontal compression, and 2) additional compression from basal shear in front of the underthrusting edge (e.g., Liu and Yang, 2003; Copley et al., 2011). For the Pamir, the two regimes were interpreted to have sustained the exhumation of domes (i.e., gravitational collapse) until ~4–2 Ma in the underthrust South Pamir and terminated the dome exhumation at ~12 Ma in the non-underthrust Central Pamir (Worthington et al., 2020). This deformation pattern is also well illustrated by the GPS velocity field (Figure 3), which shows ~ N–S extension in the South Pamir and distinct shortening north of the Central Pamir (Figure 5). Considering the current position of the underthrusting Indian lithosphere beneath the Central Pamir (see Section 2.3), we suggest that the effects of basal shear on the Pamir are subsistent. Due to the irregular curved shape of the underthrusting Indian promontory (Bloch et al., 2021), the northward compression resulting from basal shear was decomposed into asymmetric radial compressive stresses perpendicular to the underthrusting edge (Figures 9C,D). The radial compressive stresses replaced the diminished gravitational collapse as the main driving force, facilitating the foreland-ward expansion of the Pamir salient after ~12 Ma.

The present divergent interaction of the Indian lithosphere with the foreland lithospheres beneath the Pamir (see Section 2.3) indicates an asymmetric kinematic model of the expansion. On the one hand, under the western and central Pamir, the Pamir slab exhibits westward and northward delamination and rollback forced by the Indian lithosphere's underthrusting. This suggests that driven by radial compressive stresses, the Pamir salient has been spreading towards the forelands in these two directions. Therefore, the shortening rates of the TFTB and northern margin significantly increased from ~12 Ma and the Middle Miocene, respectively (Section 5.1). On the other hand, under the eastern margin of the Pamir, the Pamir slab has been overturned and is sandwiched between the Indian and Tarim mantle lithospheres. Moreover, the Moho offset was observed under the PFT in the northeastern Pamir, suggesting pure shear shortening of the Pamir crust (Xu Q. et al., 2021). These two observations imply the “hard” collision between the Pamir and the rigid Tarim lithospheres. The compressive stresses insufficient to deform the Tarim may have been transformed in two ways. One is the thickening of the crust in the northeastern Pamir. The Muztaghata Dome thus experienced the first rapid exhumation during ~12–8 Ma due to the gravitational imbalance of the new

thickened crust (Cai et al., 2017). Then, the compressive stress and the gravity-driven crustal materials from the N–S extension of the Muztaghata Dome thickened the crust around the Kongur Shan region, leading to the peak metamorphism of the Kongur Shan Dome at ~9 Ma (Section 2.2). The other is that the compressive stresses may have been transferred to the central Tien Shan through the Tarim block. This not only contributed to the intense mountain-building events throughout the central Tien Shan during the Middle–Late Miocene (e.g., Glorie et al., 2011; Jia et al., 2015; Qiao et al., 2016), but also drove the basinward deformation propagation during the Middle Miocene to Pliocene, transferring the deformation front in the NE Pamir from the MPT to the PFT at ~6–5 Ma (e.g., Thompson et al., 2015). An episode of strong thrusting-related exhumation during ~10–6 Ma was also reported across the KYTS (Cao et al., 2013a). Therefore, in this context, the Tarim Basin might have coupled with the Pamir and began to move northward after ~10 Ma, causing the strike-slip deceleration of the KYTS (Sobel et al., 2011).

In summary, the kinematic model of the Pamir salient since the ~12 Ma shifted from the radial thrusting and dextral transfer faulting model during ~23–12 Ma to another asymmetric model due to the change in the driving mechanism. The asymmetric model is discussed below.

5.3 Asymmetric radial thrusting of the Pamir salient and its internal migrating extension since ~12 Ma

Present-day crustal deformation of the Pamir constrained by the GPS and seismic strain rate, and earthquake focal stress suggests an asymmetric radial thrusting model of the Pamir salient (Pan et al., 2020). Moreover, the multi-timescale fault kinematics of the Pamir salient support that asymmetric radial thrusting has been the dominant kinematic model of the Pamir salient from ~12 Ma until the present. This model is represented by the anticlockwise rotation of symmetric radial thrusting (Ge et al., 2022).

Firstly, the fault kinematics along the frontal Pamir salient may have remained stable since the Middle–Late Miocene. In Section 5.1, existing studies indicate that the TFTB since ~12 Ma, eastern PFT, and KSES since their initiation at ~6–5 Ma have maintained relatively constant deformation rates. Deformation in the northern margin (i.e., the PTS) also resumed in the Middle Miocene (Coutand et al., 2002), but the initial deformation rate might be much smaller than those on the millennial and decadal timescales. This is because the out-of-sequence thrusting and dextral wrenching were also reactivated within the Central and southeastern Pamir during ~12–10 Ma (Rutte et al., 2017a; Rutte et al., 2017b). This suggests that the deformation was not concentrated in the

northern margin during the early foreland-ward expansion of the Pamir. As the ongoing northward advance of the Indian plate, the deformation has been concentrated in the northern margin with consistent rates on both the millennial and decadal timescales (Section 5.1). For the N–S shortening, if the rate is calculated using the total shortening throughout the Pamir (similar to the TFTB using the shortening throughout the Tajik Basin), which is equal to the amount of the delaminated Pamir crust (~150 km, Sippl et al., 2013a), the average rate would be ~12–13 mm/yr since ~12 Ma, consistent with the millennial and decadal rates of 10–15 mm/yr (Section 5.1). For the dextral slip component, despite the lack of an equivalent to determine the total magnitude of slip, the result may be similar to that of shortening because these two deformations have been governed by the same geodynamic process. Therefore, the deformation in the Pamir north of the underthrusting edge since ~12 Ma can be considered as steady. As for the strike-slip fault, the DF in the western flank has had a steady slip rate since the Late Quaternary, but the million-year rate is devoid. Based on the ~100 km left-lateral offset of the Panj River since ~10 Ma (see Section 2.1), we speculate that the DF became active before ~10 Ma, and the average slip rate is calculated to be ~10 mm/yr, consistent with the rate of 10–15 mm/yr since the Late Quaternary (Section 5.1). The dextral slip of the KYTS in the eastern flank decelerated to ~2–4 mm/yr at least by ~5–3 Ma (Section 5.1). As stated in Section 5.2.3, the deceleration may have started after ~10 Ma and was caused by the coupling of the Pamir and Tarim forced by the radial compressive stress since ~12 Ma. Therefore, the slip of the KYTS has been steady since the Middle–Late Miocene.

Secondly, along the Pamir front, the shortening rates decrease bilaterally, from the northwestern margin rather than the northern margin. The shortening rates of the TFTB in the west decrease southward from 16 mm/yr to 8 mm/yr, while the rates of the PTS in the north decrease eastward from 10 to 15 mm/yr in the northern margin to 4–7 mm/yr in the northeastern margin (Figure 8C). In the northwestern margin between the TFTB and PTS, the sinistral slip of the DF was partly transformed into thrusting (see Section 5.1). Although the shortening rate of the DF is unknown, the shortening rate of the Vakhsh Fault (Figure 1B) is 13–19 mm/yr (Metzger et al., 2020), implying that the summit of the rate is in the northwestern margin. In summary, the asymmetric distribution of the shortening rates indicates the anticlockwise rotation of radial thrusting.

Thirdly, the sinistral slip along the western flank and the dextral slip along the northern and eastern flank denote that the Pamir salient that moves northward *en bloc* has been accompanied by a counterclockwise rotation, as also reflected in the GPS data (Figure 3, and see Section 4.1). The GPS velocity vectors display a radial asymmetry that rotates anticlockwise. The rotation pole is located west of the Pamir (Zhou et al.,

2016), thus the strike-slip motion along the Pamir front should become weaker as it is away from the pole, which is consistent with the anticlockwise decreasing trend of the strike-slip rates (see Section 5.1).

Finally, the KSES, as the large and active E–W extensional fault system within the Pamir, exhibits a decrease from north to south in its E–W extension rate (see Section 5.1), matching the prediction of the radial thrusting (see Section 1). However, the fact that no counterpart of the KSES exists in the western Pamir suggests that the radial thrusting is asymmetric. The anticlockwise radial thrusting would form larger E–W extension stress diminishing from north to south in the northeastern Pamir (Figure 9C). The extension stress may have induced the gravitational collapse of the thickened crust in the northeastern Pamir, forming the KSES at ~6–5 Ma (Figure 9D) (Cao et al., 2013a; Cheng et al., 2016; Chen and Chen, 2020).

The above four lines of evidence from fault and crustal kinematics suggest that anticlockwise radial thrusting has been driving the foreland-ward expansion of the Pamir salient since ~12 Ma. In addition to the foreland-ward expansion mode, observations from the GPS velocity analysis and modern seismicity may imply the migration of the extension within the Pamir.

The GPS velocity profile C–C' shows that the E–W extension rate of 5 ± 1 mm/yr in the central South Pamir is larger than that of the southern KSES (Figure 6). Modern seismicity also indicates that N–S striking normal-fault earthquakes cluster in the same area of the South Pamir and spread roughly in the N–S direction, whereas the main E–W extensional structures in the northern and northeastern Pamir, such as KG and KSES, show low seismicity (Figure 7, and see Section 4.2). These both indicate the stronger E–W extension in the central South Pamir, but it is unreasonable under anticlockwise radial thrusting where the E–W extension in the east is stronger than that in the west.

Two events could have caused the stronger E–W extension in the central South Pamir: 1) the westward gravitational collapse in the South Pamir (e.g., Schurr et al., 2014; Rutte et al., 2017a; Rutte et al., 2017b); and 2) the westward rollback of the Pamir slab (Section 2.3). The average exhumation rate of the South Pamir domes was ~1–3 mm/yr, consistent with ~2–4 mm/yr of E–W extension based on the geometry of faults bounding the domes (Stübner et al., 2013a; Stübner et al., 2013b; Worthington et al., 2020). The figure is smaller than the present rate of 5 ± 1 mm/yr and the dome exhumation ceased at ~2 Ma (Stübner et al., 2013b; Worthington et al., 2020). Therefore, the westward gravitational collapse in the South Pamir might have partly contributed to the E–W extension in the South Pamir before its cessation, but could not enhance the present E–W extension. The slab's rollback would result in the extension parallel to the rollback direction within the area above the slab. The arcuate Pamir slab is now beneath the western South Pamir and the Central Pamir (see Section 2.3), thus the rollback of the slab can affect the central South Pamir, causing the stronger E–W extension. The ongoing westward rollback of the Pamir slab since ~12 Ma may have produced the migratory tensile stress from east to

west in the South Pamir, migrating the E–W extension in the same direction (Figure 9D). Moreover, except for the extension resulting from the basal shear of the underthrusting Indian lithosphere, the northward rollback of the Pamir slab is likely to have contributed to the current N–S extension in the South Pamir (Figure 5).

6 Conclusion

We used previously published multi-timescale deformation rates of the main structures, re-analyses of the GPS velocity and seismicity data, and existing deep structure and tectonic background, to interpret the kinematics and geodynamics of the Pamir salient since the Middle Cenozoic. From our analysis of the deformation rates along the frontal Pamir salient at million-year, millennial, and decadal timescales, we observed two deformation stages since the Late Oligocene–Early Miocene, which correspond to two transformations of the geodynamic processes, respectively. The kinematics since the second stage starting in the Middle Miocene was enhanced and has remained steady, which is consistent with the present-day deformation pattern revealed by the GPS velocity and seismicity. Based on these observations and inferences, we established multistage kinematic models for the foreland-ward expansion of the Pamir salient since the Middle Cenozoic and explore the underlying geodynamic processes and their potential causal mechanisms. The Middle–Late Cenozoic tectonic evolution can be divided into three stages.

- 1) ~37–24 Ma: The subduction of the buoyant Indian plate continued northward indenting the Pamir after the breakoff of the oceanic lithosphere at ~45 Ma, causing intense crustal shortening and thickening, and northward movement of the Pamir salient. The Pamir salient moved northward via radial thrusting to propagate deformation into the forelands.
- 2) ~23–12 Ma: The breakoff of the Indian continental lithosphere during ~25–20 Ma induced the gravitational collapse in the Central and South Pamir. The collapsed upper crust driving the orogenic expansion tended to move westward and northward, due to the different strengths of the Pamir's forelands. The kinematic model of the Pamir salient thus was transformed into the radial thrusting and dextral transfer faulting model.
- 3) ~12 Ma–present: After the collision between the underthrusting Indian lithosphere and the Pamir slab at ~12–11 Ma, the basal shear and Pamir slab's rollback forced by the underthrusting Indian lithosphere began to govern the deformation of the Pamir salient. The kinematic model of the Pamir salient was transformed into the anticlockwise radial thrusting model, due to the asymmetric radial compression induced by the basal shear and the irregularly shaped Indian promontory, as well as the resistant Tarim block in the east. The extension resulting from the lithospheric basal shear and northward rollback of the Pamir slab have caused the N–S extension within the South Pamir.

Meanwhile, the westward rollback of the Pamir slab may dominate a westward migration of the E–W extension within the Pamir.

Author contributions

XS and JG conceived the study. JG, XS, and HC wrote the manuscript. WG helped reanalyze the GPS data. All authors discussed, commented and edited the manuscript.

Funding

This work was supported by projects funded by National Natural Science Foundation of China (41720104003, 51988101, 41941016, 41972227), the Second Tibetan Plateau Scientific Expedition and Research Grant (2019QZKK0708) and the Fundamental Research Funds for the Central Universities (2021XZZX005), the Qianjiang Talent Funds of Zhejiang Province (QJD190202) and the 100 Talents Program of Zhejiang University.

Acknowledgments

We thank Dr. Xin Wang for discussing the focal mechanisms. We are grateful to the reviewers for their constructive comments and suggestions, and the editors and the Frontiers editorial team for handling this manuscript.

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Supplementary material

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 10 June 2022

ACCEPTED 26 July 2022

PUBLISHED 02 September 2022

CITATION

Peng Y, Sun X, Zhan Y, Zhao L, Luo Q,
Liu X, Ran Z and Wang L (2022), 3D deep
electrical structure and seismogenic
environment in the western section of
the Zhangjiakou-Bohai fault zone.
Front. Earth Sci. 10:966192.
doi: 10.3389/feart.2022.966192

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3D deep electrical structure and seismogenic environment in the western section of the Zhangjiakou-Bohai fault zone

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The Zhangjiakou-Bohai fault zone (ZBFZ) is an important NW-trending active tectonic zone in North China. The western section of ZBFZ is characterized by frequent moderate and strong earthquakes. This is a typical tectonic area for studying seismic and volcanic activities. The three-dimensional (3D) deep electrical structure of the region was revealed by using 3D electromagnetic inversion of 143 magnetotelluric stations. The results suggest that the deep electrical structure in the western part of the ZBFZ is laterally heterogeneous. To the north of the Shangyi-Chicheng Fault (F4), the upper crust has a high resistivity structure (R), corresponding to the exposed Hanuoba basalts and metamorphic rocks. To the south of F4, high and low resistivity zones alternate, which is attributed to the complex faults and basin-range structures. In the middle crust, the low-resistivity structures (C1 and C2) gradually appear in the northwest and southeast of the study area, showing the characteristics of a high-resistivity structure sandwiched between two low-resistivity structures. In the lower crust, the two low-resistivity structures (C1 and C2) gradually join each other and the whole region becomes characterized by low resistivity. The Shangyi-Chicheng Fault (F4), the northern margin fault of the Huai'an Basin (F7), and the northern margin fault of the Huaizhuo Basin (F11) are definite electrical difference zones, which are connected to a low-resistivity layer in the crust. The results indicate that the focal depth of the 1998 M6.2 Zhangbei earthquake was between 12 km and 15 km and its seismogenic fault was the Dahezhen-Hailiutu Fault (F3), which is concealed beneath the Hanuoba basalt area. It is speculated that owing to the neotectonic activities, the magma originated in the asthenosphere mantle rose and erupted on the surface near Datong, forming the Datong volcanic group. Then the magma continued to migrate and overflow along with the fractures in the NE direction, forming the Hannuoba basalts in the Zhangbei area. The existence of mantle thermal material in this area led to the migration of accompanying fluids to the middle and upper crust, resulting in fault sliding and frequent occurrence of moderate and strong earthquakes in this region.

KEYWORDS

Zhangjiakou-Bohai fault zone, 3D magnetotelluric imaging, electrical structure, 1998 Zhangbei earthquake, seismogenic environment

Introduction

The Zhangjiakou-Bohai fault zone (ZBFZ) is a vast structural belt with a width of tens of kilometers. It is one of the important NW-trending active tectonic zones in North China. It comprises a series of discontinuous, NW-trending active tectonic belts in the northern part of North China Craton and the northeastern part of Shanxi Graben (Fu et al., 2000; Figure 1A). The ZBFZ is tectonically situated in the central and eastern part of the North China Craton, which experienced destruction during the Phanerozoic (Zhu et al., 2012; Figure 1A). It extends from the west of Zhangjiakou, through the north of Beijing, Sanhe, the south of Tangshan, the south of Tianjin, to Penglai, and finally intersects with the NNE-trending Tan-Lu Fault Zone (TLFZ) in Bohai Bay. From west to southeast, the ZBFZ can be divided into the Zhangjiakou (western), Beijing (central-western), Tangshan (central-eastern) and Penglai (eastern) sections (Gao, 2001; Suo et al., 2013).

The ZBFZ is an area of concentrated strong earthquakes and plays a significant role in controlling the distribution of moderate and strong earthquakes in North China. Since the beginning of recorded history, there have been more than 20 moderate and strong earthquakes above M6 in the middle-western and middle-eastern

sections of the ZBFZ, such as the 1679 M8 Sanhe-Pinggu earthquake, 1969 M7.4 Bohai earthquake and 1976 M7.8 Tangshan earthquake (Liu et al., 1982; Xiang et al., 1988, Zhang et al., 1995; Wang et al., 2017; Figure 1A). The western section of the ZBFZ has also been seismically active since records are available, with frequent moderate and strong earthquakes above M5, such as the 1720 M6.7 Shacheng earthquake (SCEQ), 1989 M5.6–5.9 Datong-Yanggao earthquake swarm (DT-YGEQ) and 1998 M6.2 Zhangbei earthquake (ZBEQ 1998) (Xu et al., 1998; Feng et al., 2016). In recent years, moderate and small earthquakes have frequently occurred in the ZBFZ. In particular, in 2021, a significant number of moderate and small earthquakes with a magnitude of about 3 were observed near the Zhangbei seismic area, indicating that the western section of the ZBFZ has become more seismically active (Figure 1B).

The western section of the ZBFZ is also characterized by intensive volcanic activities. The Datong volcanic group (DVG) located in this area is the only volcanic group on the Loess Plateau. The basaltic magma originated from melting of the upper mantle (Deng et al., 1987; Li, 1988; Figure 1B). The volcanic activities in Datong were intermittent and lasted for a long time. The latest volcanic activity occurred in the early Late Pleistocene (Zhai et al., 2011). A study of mantle xenoliths

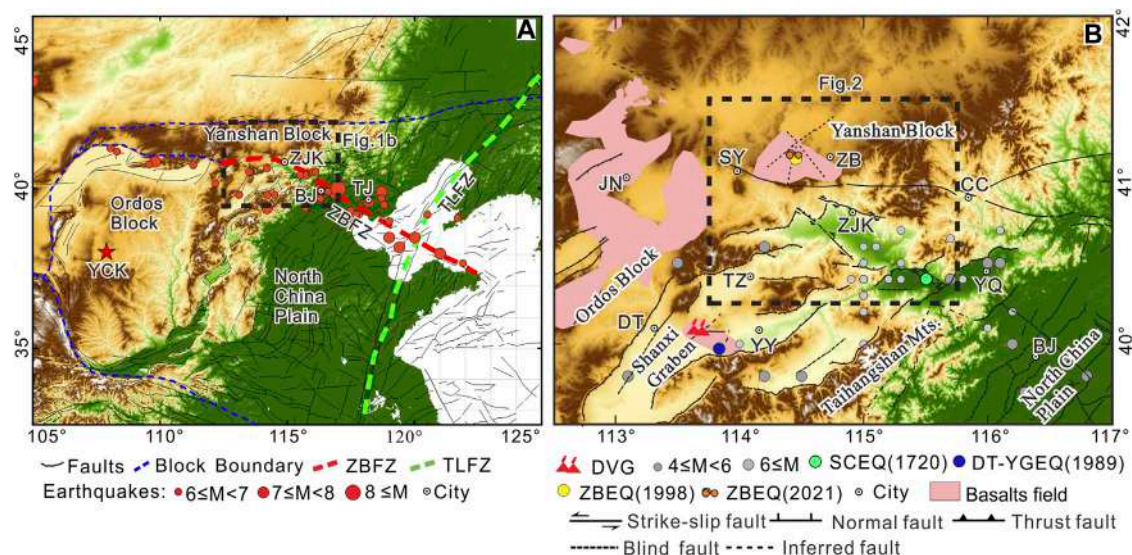


FIGURE 1

(A) Location of the study region. (B) Topographic map of the Western segment of Zhangbo fault zone. Faults are modified from Zhou (2011) and Luo et al. (2021); Moderately and strong earthquakes (BC 231-AD 2018) are from Liu et al. (2017) and Hu et al. (2021); Abbreviations are: ZBFZ, Zhangbo fault zone; TLFZ, Tanlu fault zone; DVG, Datong Volcanoes group; JN, Jinjing; SY, Shangyi; ZB, Zhangbei; CC, Chicheng; ZJK, Zhangjiakou; TZ, Tianzhen; YQ, Yanqing; DT, Datong; YY, Yangyuan; BJ, Beijing; YCK, Remote reference station; LPF, Langshan Piedmont fault; LHF, Linhe fault; DKF, Dengkou fault; QS, Qingshan; HJHQ, Hangjinhouqi; LH, Linhe; DK, Dengkou; WH, Wuhai; WLTHQ, Wulatehouqi; WY, Wuyuan; WLTHQ, Wulateqianqi; WLTHQ, Wulatezhongqi; HJQ, Hangjingqi.

indicates that the DVG resulted from destruction of the lithosphere and upwelling of mantle materials (Cen et al., 2015). Zhang (2021) observed that magmatic rocks are rarely found in Tianzhen and Yanggao in the northeast of the Datong volcanic area. It is speculated that magma in this area has not erupted, and there are still magma chambers or high-temperature anomalies at depth. Cenozoic fissures or overflow basalt (Hannoba basalt, HNBB) are distributed along the borders of Shangyi, Zhangbei and Zhangjiakou (Liu et al., 1992; Figure 1B). They are a result of an overflow of mantle-derived magma along the approximately EW-trending Shangyi-Chicheng Fault and the NW-trending ZBFZ. The numerous volcanic activities in the western section of the ZBFZ also strongly suggest that the crust in this area was active in the Cenozoic (Hao, 2010).

Carrying out 3D deep structure exploration of the western segment of ZBFZ and its adjacent area will help to comprehensively understand the seismogenic mechanisms and environment in the seismically active area and is practically very important for the medium and long-term seismic risk analysis in this area. The magnetotelluric method can be used for detection of even small changes in the resistivity of a rock mass. It can be successfully applied to reveal the seismogenic environment and the deep extension of fault zones. It is one of the effective means of studying the structure of moderate and strong earthquake and seismogenic environments (Tan et al., 2004; Unsworth et al., 2004; Becken et al., 2011; Zhao et al., 2012; Zhang et al., 2016; Cai et al., 2017; Yu et al., 2017; Lin et al., 2017; Zhan et al., 2017, 2021; Ye et al., 2018a; Wang et al., 2018; Yang et al., 2019; Li et al., 2021). In the past decade, moderate and strong earthquakes have occurred on the eastern edge of the Qinghai-Tibet Plateau, such as in Wenchuan, Lushan, and Jiuzhaigou. Magnetotelluric exploration has been carried out in these areas and the surrounding Longmenshan, Altyn Tagh, and Qilian-Haiyuan structural belts. It revealed the different deep resistivity structures along different sections of the active faults as well as the deep tectonic environment of a series of moderate and strong earthquakes. Literature suggests that moderate and strong earthquakes and aftershocks are closely associated with the resistivity structure of the underground medium (Zhao et al., 2012; Zhan et al., 2013; Wang et al., 2014; Xiao et al., 2017; Cai et al., 2017; Sun et al., 2019, 2020). The 3D magnetotelluric imaging before the 2021 M7.4 Maduo, 2021 M6.4 Yangbi and 2022 M6.9 Menyuan earthquakes (Ye et al., 2021; Zhan et al., 2021; Zhao et al., 2022) also showed that the seismogenic positions of these earthquakes were near the resistivity change zone in the deep medium. In addition, by using the magnetotelluric 3D electromagnetic imaging technology, Ye et al., 2018b, Ye et al., 2020 identified the 3D resistivity characteristics of the crust beneath the Tengchong volcanic belt and the distribution of magma and revealed multiple seismogenic structures in the area.

Moderate and strong earthquakes and volcanoes abound in the western section of the ZBFZ and its adjacent areas. The 3D

magnetotelluric imaging technology enables determining the 3D structure of the resistivity in the deep medium in the western section of the ZBFZ, which can provide a reference for studying the material state and deep seismogenic environment in this area. Using a 3D electromagnetic inversion of 143 magnetotelluric stations, this study reveals the 3D deep electrical structure of the region, clarifies the seismogenic structure of the M6.2 Zhangbei earthquake, interprets the Zhangbei earthquake area and the deep breeding environment of the recent frequent small and medium-sized earthquakes, and provides deep data for seismic risk assessment in this area.

Regional geological structure and magnetotelluric profile

The study area is located at the intersection of the EW-trending Yanshan Orogen, the northeastern section of the NE-trending Shanxi fault depression belt and the NW-trending ZBFZ (Figure 1B). The geological structure and the distribution of magnetotelluric survey points in the western section of the ZBFZ are shown in Figure 2. Bounded by the approximately EW-trending Shangyi-Chicheng Fault (F4), the Yanshan Orogen is located in the north, while the complex basin-range structural area is located in the south, including the Zhangjiakou Fault (F5), Huai'an Basin (HAB), Yanggao-Tianzhen Basin (YTB), Xuanhua Basin (XHB), and Huaizhuo Basin (HZB) (Xu et al., 2002; Zhou, 2011; Peng et al., 2017; Figure 2).

The 150-km long Shangyi-Chicheng Fault (F4) is the largest fault in this region. The fault plane is steep and dips to north or south in different places, exhibiting features of high-angle compression and shear (Zhou et al., 2019). In the northern part of F4, three hidden faults are located near the Zhangbei earthquake area, namely the Daman-Qianheishatu (F1), Miaodongying-Dayingtian (F2), and Dahezhen-Hailiutu (F3) Fault (Xu et al., 1998; Cai et al., 2003). In the southern part of F4, three groups of NW-, NE- and approximately EW-trending faults are found. The NW-trending faults, including the Zhangjiakou (F5) and Xialin-Shuiquan (F6) Fault, are situated in the middle of the study area. The NE-trending faults are distributed in the southwest of the study area and include mainly the northern margin fault of Huai'an Basin (F7) and the northern margin fault of Yanggao-Tianzhen Basin (F8), which are characterized by large scale and long extensions. The approximately EW-trending faults are developed in the south of the study area. Most of them are concealed, discontinuous, and arranged as an echelon. These EW-trending faults are represented by the southern margin fault of Huai'an Basin (F9), the southern margin fault of Xuanhua Basin (F10), and the northern margin fault of Huaizhuo Basin (F11) (Ran et al., 2001; Zhou, 2011; Qi, 2017; Luo et al., 2021).

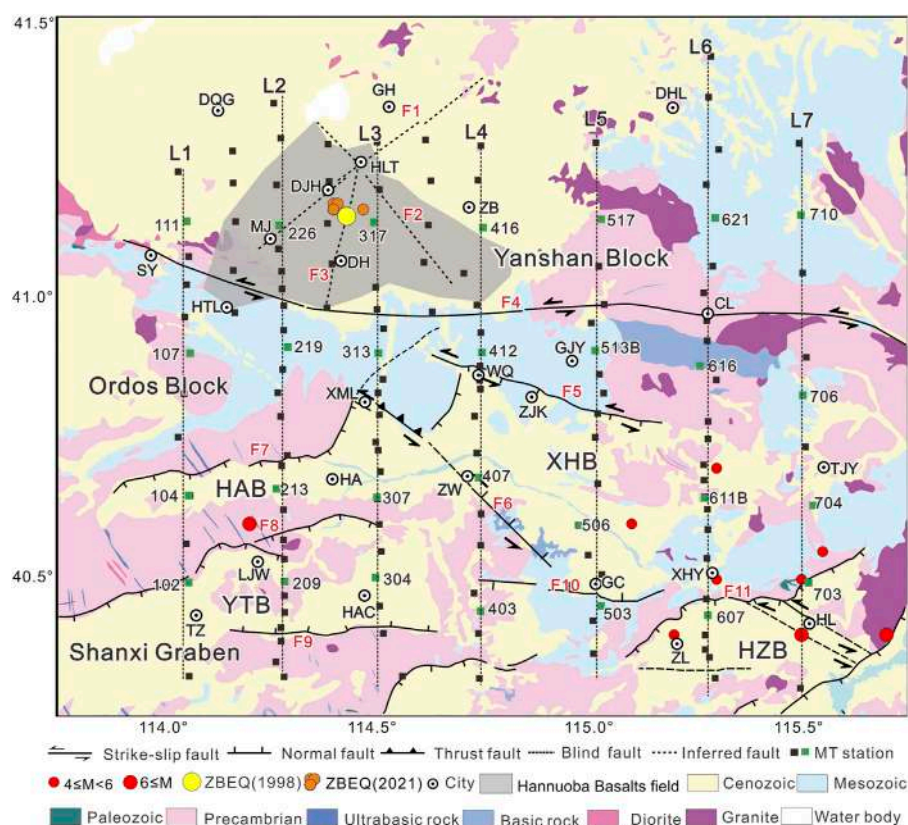


FIGURE 2

Geological setting [modified from Peng et al. (2017), Zhou (2011); Luo et al. (2021)] and MT stations in studied region. Black Squares denote MT stations; Green square denote typical MT stations in Figure 2; Abbreviations are: HABA, Huaian basin; YTB, Yanggao-Tianzhen basin; XHB, Xuanhua basin; HZB, Huaizhuo basin; F1, Daman-Qianheishatu fault; F2, Miaodongying-Dayingtan fault; F3, Dahezhen-Hailiutu fault; F4, Shangyi-Chicheng fault; F5, Zhangjiakou fault; F6, Ximalin-shuiquan fault; F7, Northern margin fault of the Huaian basin; F8, Northern margin fault of the Yanggao-Tianzhen basin; F9, Southern margin fault of the Huaian basin; F10, Southern margin fault of the Xuanhua basin; F11, Northern margin fault of the Huaizhuo basin; DQG, Daqinggou; GH, Gonghui; DHL, Dahulun; HLT, Hailiutu; DJH, Danjinghe; MJ, Manjing; DH, Dahe; SY, Shangyi; CL, Chongli; HTL, Hongtuliang; GJY, Gaojiaying; WQ, Wanquan; XML, Ximalin; ZJK, Zhangjiakou; HA, Huaian; ZW, Zuowei; TJY, Tianjiayao; LJW, Lujiawan; HAC, Huaiancheng; GC, Guocun; XHY, Xiahuyuan; TZ, Tianzhen; ZL, Zhuolu; HL, Huilai.

The strata, lithology, and magmatic activities in the western section of the ZBFZ are obviously controlled by fault structures (Qi, 2017). Precambrian, Mesozoic, and Cenozoic strata are exposed to the north of F4. The Cenozoic Hannuoba basalts are distributed in the Shangyi and Zhangbei regions (Liu et al., 1992), Precambrian strata are mainly found to the south of F4, Paleozoic strata occur sporadically around Shangyi. Mesozoic strata are mainly distributed between Shangyi and Zhangjiakou, and Cenozoic strata in the Huaian (HAB), Yanggao-Tianzhen (YTB), Xuanhua (XHB), and Huaizhuo (HZB) Basins as well as their margins. Ultramafic, mafic, diorite, and granite are also sporadically exposed in the study area. The Precambrian strata are dominated by metamorphic rocks, the Paleozoic strata are mainly composed of carbonate rocks and the Mesozoic strata consist mostly of volcanic sedimentary rocks in continental basins. The Cenozoic strata are exposed in the mountains and comprised plagioclase

granulite and gneiss. The Cenozoic strata in the basin are mainly loess, sandy clay and glutenite (Wang et al., 1996; Peng et al., 2017; Figure 2).

In view of the complex fault system and distribution characteristics of the main faults in the study area, an magnetotelluric detection array was adopted, which had Zhangjiakou as its center and extended from it, forming seven survey lines along the SN direction (numbered L1, L2, L3, L4, L5, L6 and L7 from west to east) as seen in Figure 2. The spacing of adjacent lines was about 20 km. Survey line L6 was the longest, with a length of 125 km, and survey line L1 was the shortest, with a length of 103 km. The measurement points were evenly distributed and relatively dense along the survey lines, with most of them less than 5 km apart. In order to reveal the deep seismogenic structure of the Zhangbei earthquake area, 15 additional measurement points were arranged, bringing the total number of magnetotelluric measurement points to 143.

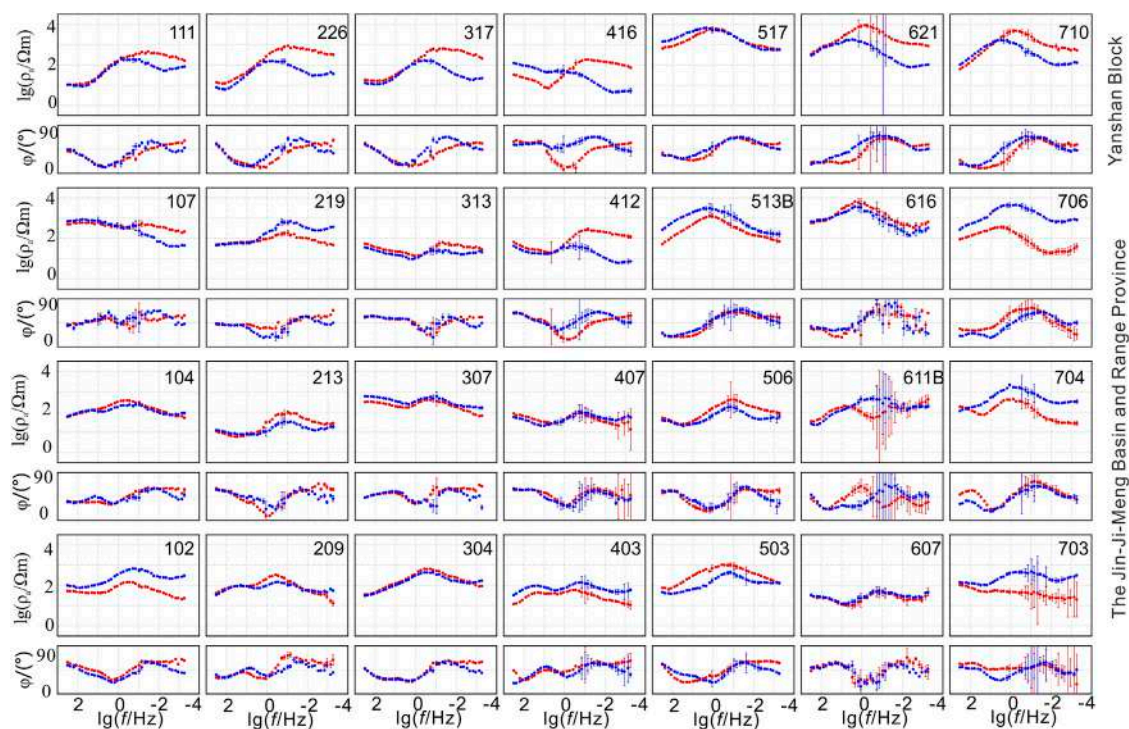


FIGURE 3

Apparent resistivity and impedance phase curve of typical MT stations (Red dots denote XY mode, blue dots denote YX mode).

Field observations and data analysis

Data collection, processing and apparent resistivity and phase curve characteristics

The magnetotelluric field survey was carried out between July and September 2020. The MTU-5A magnetotelluric observation system manufactured by Phoenix Geophysics Ltd. of Canada was used for data acquisition. The electric field components in the SN and EW directions and the magnetic field components in the SN, EW, and vertical directions were observed at each measurement point. Transport and electrical power infrastructure, such as electrified railways and wind power stations, are densely distributed over the entire study area. During field data collection, remote-reference was used to process the data for enhanced quality (Gamble et al., 1979). A remote reference station (YCK) is located near the Maowusu Desert in Otog Front Banner, Inner Mongolia, about 700 km away from the study area (Figure 1A). The data recording period at each measurement point was more than 40 h, and at some it reached 50 h. The quality of data was significantly improved after the remote reference processing. The apparent resistivity and phase curves of 12 measurement points before and after remote reference processing are shown in Supplementary Figure S1 as examples.

The apparent resistivity and impedance phase curves of 28 typical measurement points from north to south along the seven survey lines are shown in Figure 3 (the locations of measurement points are shown as green squares in Figure 2). As can be seen in Figure 3, the apparent resistivity curves at all measurement points in the Yanshan Orogen to the north of F4 are essentially the same, i.e., they first have low values, then increase, to fall again with increasing period. The apparent resistivity at the intermediate frequencies is close to tens of thousands of Ωm , indicating the presence of high resistivity anomalies in the middle and upper crust. The decrease in low-frequency apparent resistivity implies the existence of deep low-resistivity structures. Moreover, the apparent resistivity is low in the west and high in the east. To the south of F4, the shapes of apparent resistivity curves in medium-high frequency band at each measurement point are distinct, corresponding to the complex Jin-Ji-Meng Basin and Range Province. The apparent resistivity values at lower frequencies are generally low, indicating the presence of low-resistivity structures in the middle and lower crust in this region.

Regional dimension analysis

The phase tensor decomposition technique (Caldwell et al., 2004; Bibby et al., 2005) is usually used to qualitatively analyze the

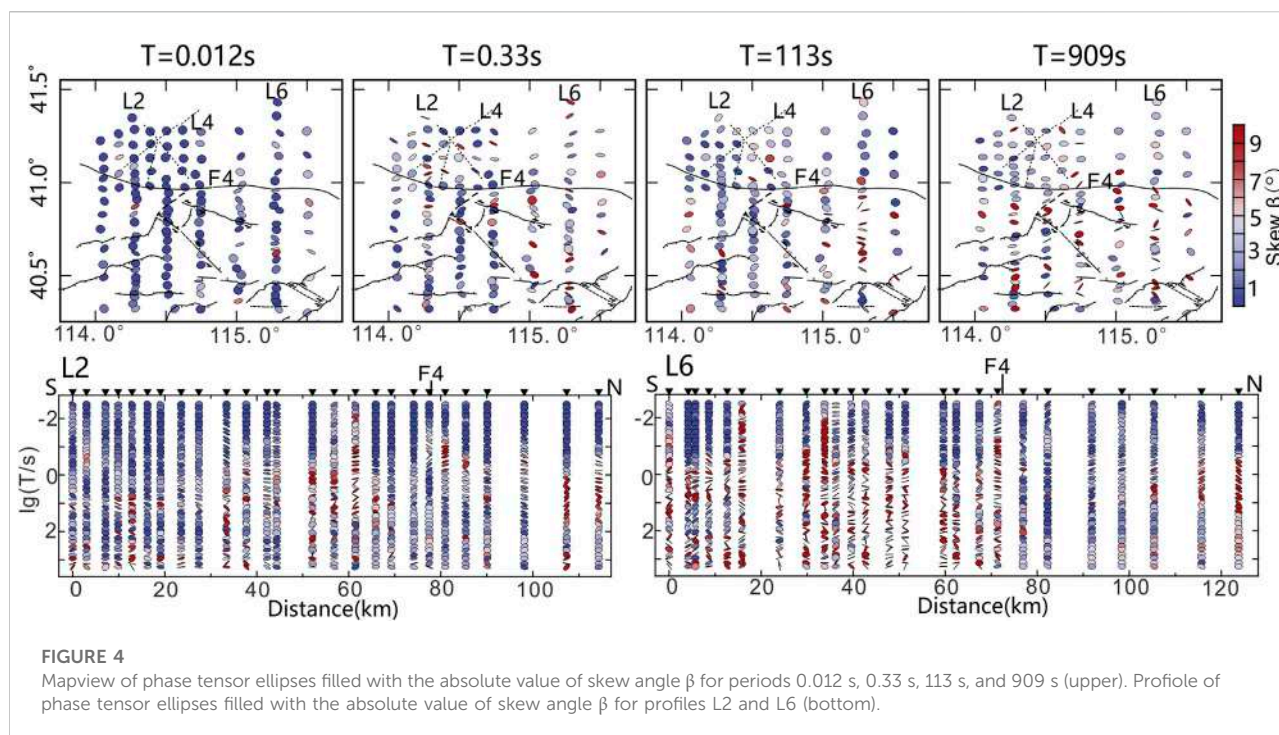


FIGURE 4

Mapview of phase tensor ellipses filled with the absolute value of skew angle β for periods 0.012 s, 0.33 s, 113 s, and 909 s (upper). Profile of phase tensor ellipses filled with the absolute value of skew angle β for profiles L2 and L6 (bottom).

dimensional features of underground media. Considering that the actual observation data contain uncertainty, it is generally believed that when two-dimensional deviation degree of the phase tensor, $|\beta|$, is larger than 5, the 3D property of an underground structure is strong (Booker, 2014; Cai et al., 2017). Figure 4 shows the values of $|\beta|$ of all measurement points in four periods, as well as the periodic changes along survey lines L2 and L6, where it can be seen that in the moderate and high frequency bands, the values of $|\beta|$ are below 5. With increasing period, especially to the south of F4, the values of $|\beta|$ markedly increase. A large number of measurement points exhibit $|\beta|$ values exceeding 5, indicating that the shallow structure of the survey area is relatively simple. The middle and deep structures, especially those to the south of F4, are relatively complex. The deep electrical structure is 3D, which is related to the distribution of complex fault systems in the south of the study area. In order to reveal the deep electrical structure of such complex areas, it is necessary to apply the 3D electromagnetic inversion technology.

Regional electrical difference analysis

In the qualitative analysis of regional electrical differences, the phase tensor invariant, φ_2 (Heise et al., 2008), is adopted, which only contains the resistivity gradient and variation information. Although it cannot be used to determine the specific resistivity value of an underground structure, it can reflect its trend. A large value of φ_2 indicates that the resistivity decreases with depth, otherwise it indicates that it

increases with depth. Figure 5 shows the distribution map of phase invariants at all measurement points over four periods as well as the periodic changes along survey lines L2 and L6. The φ_2 values increase with the period in the moderate and high frequency bands to the west of survey line L4 and to the south of F4. The φ_2 values are clearly smaller than those in the to the east of survey line L4. This shows that the decreasing trend in the west and south of the study area is more obvious. With the period increasing in the low-frequency band below tens of seconds, the φ_2 values become much larger and more uniform, indicating that the deep strata of the study area generally has a decreasing trend.

3D inversion

The 3D inversion was carried out using the ModEM software (Egbert and Kelbert, 2012; Kelbert et al., 2014). The full impedance data at 35 frequency points from all 143 measurement points between 320 Hz–0.000137 Hz (periods 0.0031 s–7,300 s) were utilized for inversion, including nine frequency points above 1 s and 28 frequency points below 1 s. Among the 143 measurement points, 132 points provided valid data with a period of more than 1,000 s.

The horizontal grid generation for the 3D inversion model covers the core of each measurement point and its boundary. The horizontal grid covering the core of each measurement point was evenly divided. A 48 (SN direction) $\times$ 48 (EW direction) grid

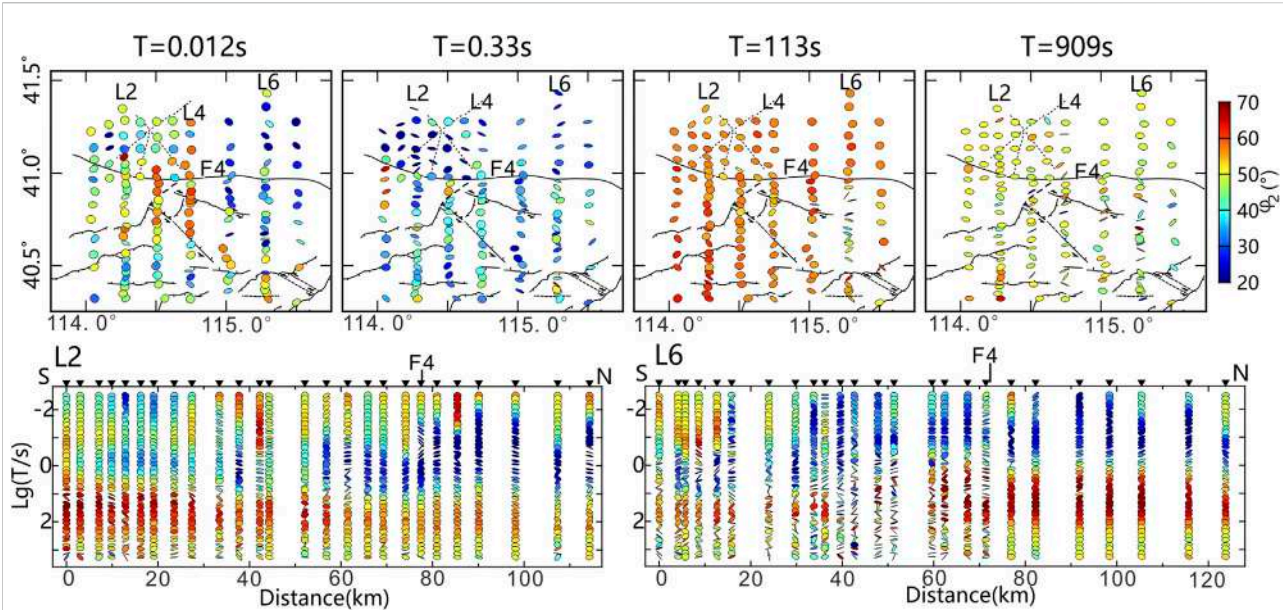


FIGURE 5

Mapview of phase tensor ellipses filled with the phase tensor invariant for periods 0.012 s, 0.33 s, 113 s, and 909 s (upper). Profile of phase tensor ellipses filled with the phase tensor invariant for profiles L2 and L6 (bottom).

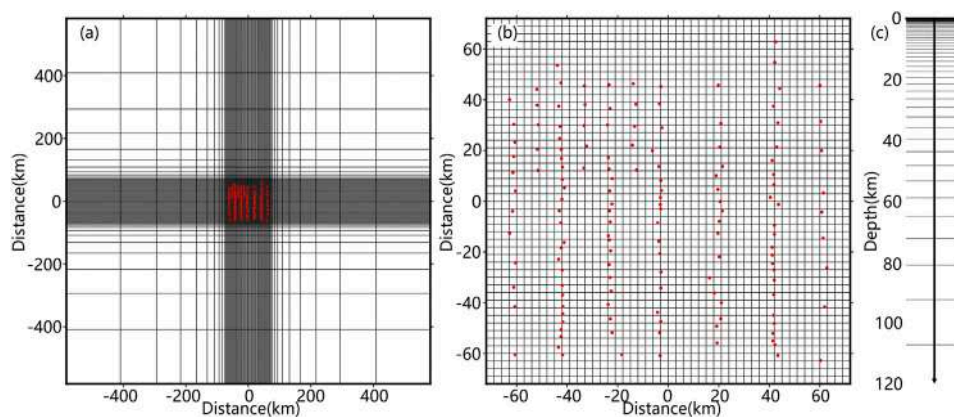


FIGURE 6

Grid used for the 3D inversion (A) Horizontal grid; (B) Horizontal grid of central part; (C) Vertical grid; The red dots denote MT stations.

(3.0 km $\times$ 3.0 km in the centre of the study area) was adopted. There were 10 extended grids for each horizontal boundary, and, therefore, 68 $\times$ 68 grid lines in total. The vertical grids were denser closer to the surface but became gradually coarser with the increasing depth. The thickness of the uppermost layer was 10 m and increased at different rates with depth. These rates were 1.2, 1.1, 1.2, 1.3, and 1.4 within 2 km, 2–60 km, 60–80 km, 80–150 km, and 150–800 km, respectively. The vertical grid had a total of 69 layers (Figure 6).

During the 3D inversion, the error floor of the diagonal elements and the off-diagonal elements were set to 10% and 5%, respectively. The errors of a few frequency points obviously disturbed by noise were enlarged to reduce their weight in inversion. The smoothing factors used in the three directions for the 3D inversion equalled 0.3. A 100 Ω m uniform half-space was utilized as the initial model with an automatically updated regularization factor of an initial value of 5,000. When the inversion no longer converged, the regularization factor was

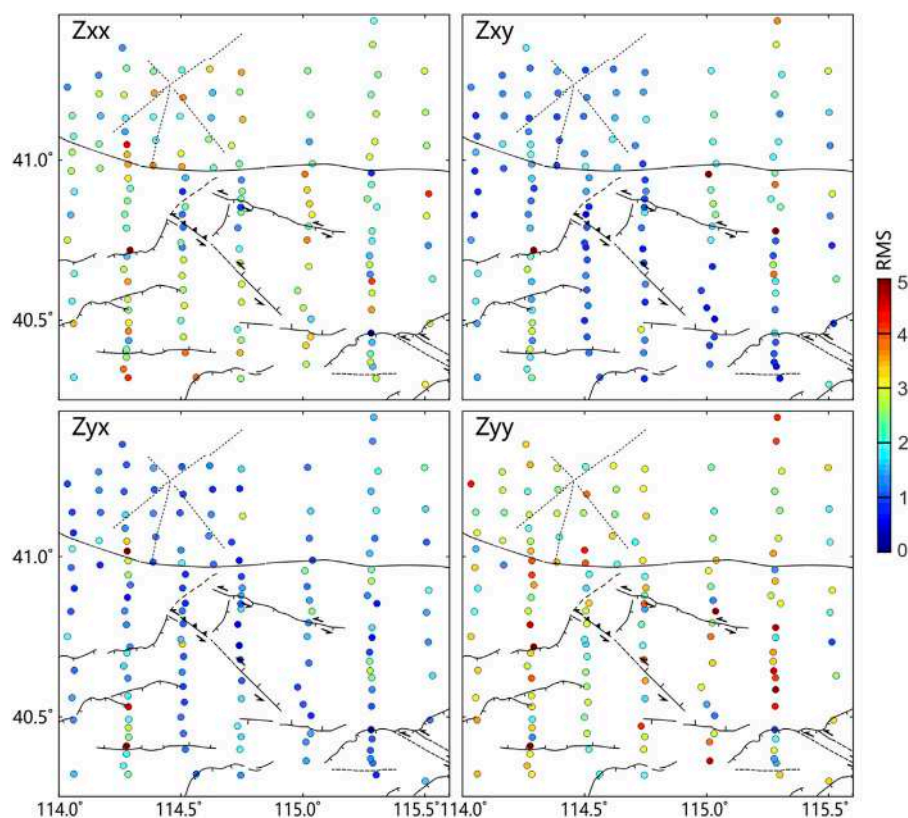


FIGURE 7
The four component RMS misfit at each station.

reset to one tenth of the original value and inversion continued. After 79 iterations, the inversion was completed, and the root mean square (RMS) error of the model was 2.22. The 3D inversion results at the measurement points agreed well with the measured main diagonal elements and sub diagonal elements (Figure 7). The lines fitted to the 3D inversion model response at all the measurement points are shown in Supplementary Figure S2. Regarding the robustness of the major model features, sensitivity tests of C1 and C2 have been carried out, as shown in Supplementary Figures S3, S4.

Results and discussions

The 3D electromagnetic inversion revealed the 3D deep electrical structure of the western section of the ZBFZ. The deep electrical structures of seven sections along the SN direction are shown in Figure 8. The planar sections at nine depths are displayed in Figure 9, while two vertical sections across the Zhangbei seismic area are presented in Figure 10. The positions of these faults from surface geological data were compared with the positions of magnetotelluric measurement

points. The deep extension of each fault is illustrated in Figure 8. Three concealed faults in the northern part of F4 and the seismic structure in Zhangbei area are detailed in Figure 10. From the electrical structure characteristics shown in these figures, the main faults in the study area are situated in obvious electrical interface or electrical gradient zones and the electrical structures of the survey area are strongly horizontally heterogeneous.

Deep extension features of faults

The approximately EW-trending F4 is one of the largest faults in the study area. The fault dips to north or south, exhibiting high-angle compression and shear characteristics (Zhou et al., 2019). Along survey lines L1 to L7, F4 is consistent with the electrical difference zone (Figure 8). The deep extension of the electrical difference from west to east is segmented. In the western survey lines L1 to L4, there is an obvious electrical difference zone. The part to the north of F4 is characterized by high resistivity structure, while the part to the south of it has markedly low resistivity. From west to east, the downward extension angle of the electrical difference zone

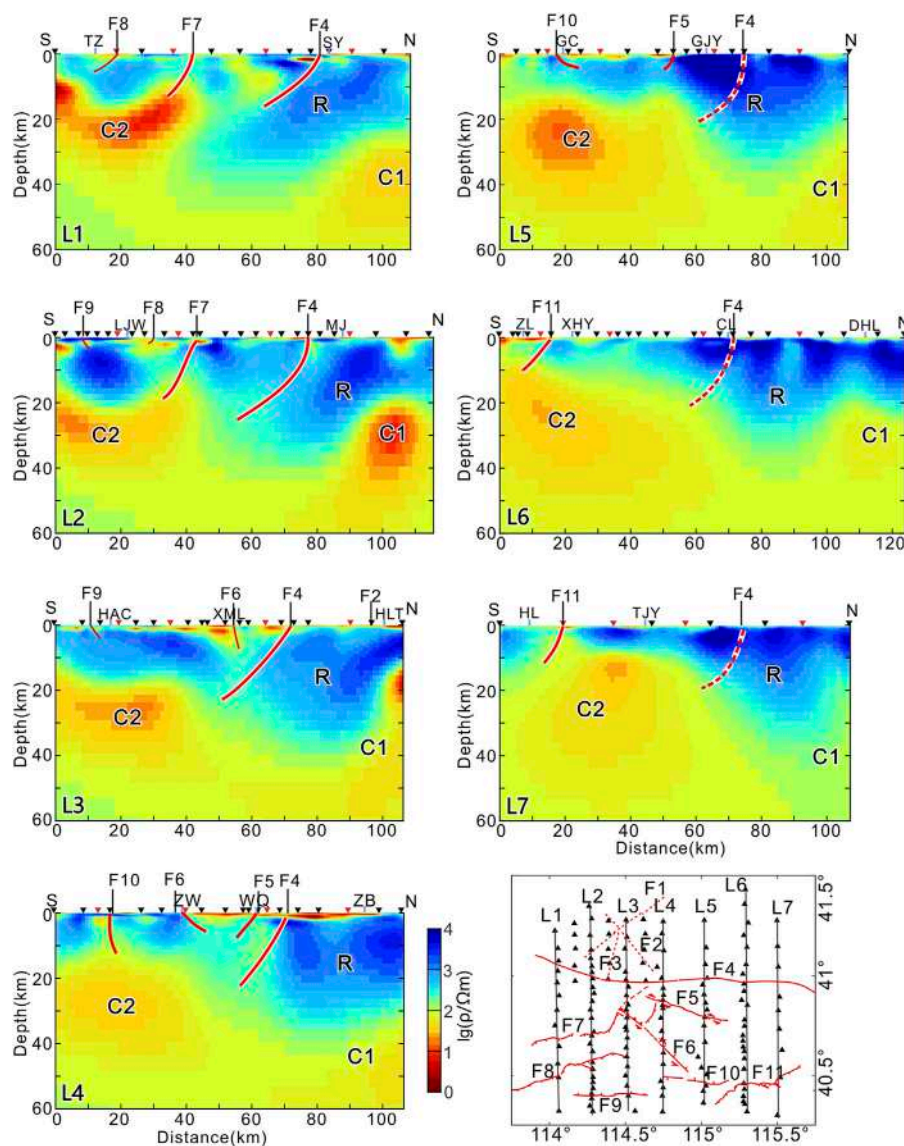


FIGURE 8

Cross sections of seven profiles (L1, L2, L3, L4, L5, L6, and L7) extracted from the 3D inversion model.

gradually becomes steeper from dipping south. The extension depth this zone can reach is below 15 km and it is connected with the deep low-resistivity structure C2. To the east of survey line L4, F4 is developed inside the high-resistivity structure R, with little electrical difference. This difference along the EW direction may be related to the fact that the western segment of F4 is highly active and relatively new, while the eastern segment of F4 is relatively stable. We speculate that the fault may be concealed in the high-resistivity body in the eastern section, which is also consistent with the conclusions of the geological survey (Cai et al., 2002; Zhou et al., 2019).

The Zhangjiakou (F5) and Ximalin-Shuiquan (F6) Faults are two NW-trending faults in the middle of the study area and are the southern and northern boundaries of the Xuanhua basin (XHB). Along survey lines L4 and L5, F5 is an electrical difference zone with a slight southward dip. The electrical difference tends to decrease from west to east, which may be related to the intense activity of the western section and the relatively weak activity of the eastern section (Zhou and You, 2010). Along survey lines L3 and L4, F6 is an electrical boundary zone with a slight dip to the north. These two faults extend to a depth of about 3 km and are small in scale.

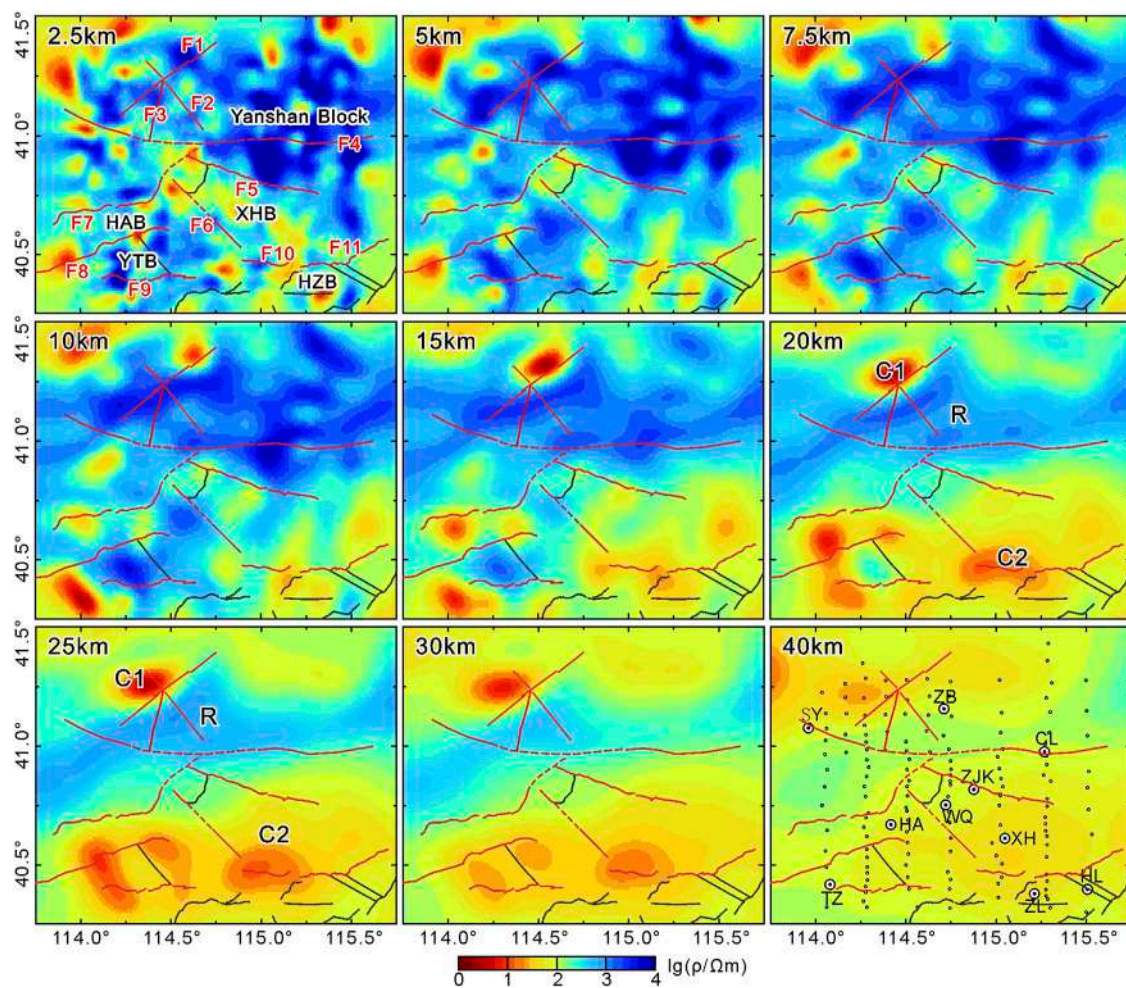


FIGURE 9
Map view of the 3D inversion model.

F7 is mainly observed in survey lines L1 and L2 and has a southward sloping electrical boundary below it. It extends below 15 km and merges with the moderate-low resistivity structure (C2). F8 is the northern boundary of Yanggao-Tianzhen Basin (YTB) and is present as an electrical differential gradient zone in survey lines L1 and L2, with a depth of 2–3 km. In the southwest of the study area, the fault in the southern margin of Huai'an Basin (F9) shows as an upright and slightly northward dipping electrical difference zone in survey line L2, with an extended depth of about 10 km. The fault in the southern margin of Xuanhua Basin (F10) manifests as a gently northward dipping electrical difference zone in survey lines L4 and L5, extending for about 5 km. The fault in the northern margin of Huaizhuo Basin (F11) is evident as a southward dipping electrical difference zone in survey lines L6 and L7, with an extension depth of about 10 km. In the deep part, it may merge with the low resistivity

layer (C2). The distribution of faults indicated by the deep electrical structure image reveal a clear electrical anomaly difference zone with a deep extension under the Shangyi-Chicheng Fault (F4), the northern margin fault of Huai'an Basin (F7) and the northern margin fault of Huaizhuo Basin (F11). It is connected with the low-resistivity layer in the middle and lower crust, indicating that these three faults may be large faults.

Lateral characteristics of deep electrical structure

The horizontal slices of the resistivity structure at nine depths are shown in Figure 9. As can be seen from the figure, the electrical structures in the shallow and deep parts of the study area are clearly distinct. From the shallow surface to the

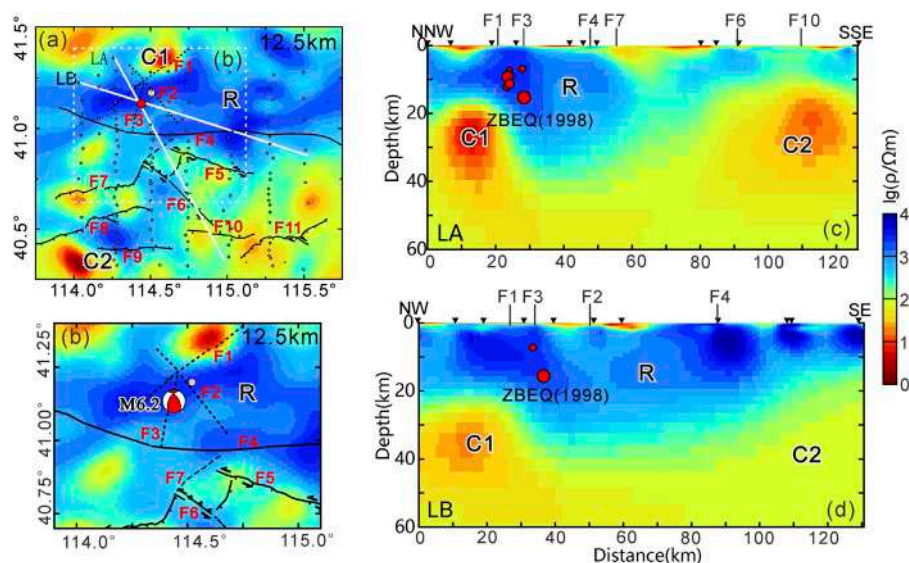


FIGURE 10

Electrical structure beneath the Zhangbei earthquake. (A) Map view of the 3D inversion model in the Western section of ZBFZ at depth of 12.5 km; (B) Map view of the 3D inversion model near the Zhangbei earthquake at depth of 12.5 km; (C) and (D) Cross sections of 3D inversion model along profiles LA and LB.

depth of 7.5 km, the electrical structures to the north and south of F4 are significantly different. To the north of F4, the deep electrical structure of the surrounding area centered on Zhangbei is generally characterized by high resistivity (R) reaching up to thousands of Ωm and corresponding to the Cenozoic Hannuoba basalts and metamorphic rocks exposed in Zhangbei and its adjacent areas (Liu et al., 1992). To the south of F4, zones of high and low resistivities are liberally distributed, consistent with the shallow faults and complex basin-range structural areas comprising the Huai'an (HAB), Yanggao-Tianzhen (YTB), Xuanhua (XHB), and Huaizhuo (HZB) Basins (Xu et al., 2002; Zhou, 2011; Peng et al., 2017). At depths of 7.5–15 km, in the northwest and southeast regions low resistivity structures C1 and C2 exist, respectively. The deep electrical structure shows a zone of high resistivity sandwiched between two zones of low resistivity. At depths of 20–30 km, with the increase in depth, C1 extends to the southwest and C2 extends to the northwest. The high-resistivity structure R becomes an elliptic band spreading from southwest to northeast. Below 40 km, the resistivity of the whole region is generally low (Figure 8).

The sections shown in Figure 8 enable further elaboration on the spatial distribution characteristics of the low resistivity structures C1, C2, and the high-resistivity structure R. Across the seven survey lines, the high-resistivity structure R is narrow in the west and wide in the east, and deep in the west and shallow in the east. Along the western survey line L1,

R is the narrowest, with a depth of about 40 km. Moving across survey lines L2, L3 and, L4, its width gradually increases, while its depth gradually decreases. Along survey lines L5 and L6, it is the widest, but its depth is reduced to about 30 km. Along the easternmost survey line L7, the range of R is reduced. The low-resistivity structure C1 in the northwest part of the survey area is generally deep in the south and shallow in the north. The low-resistivity structure C2 developed in the southeast region is shallow in the south and deep in the north, with a depth of about 30 km in the northern part. Along survey line L1, C2 changes from deep in the south to shallow in the north. The depth of the southern section of C2 exceeds 20 km, while in the northern section it is shallower, about 10 km. It joins F7 in the upper crust. Along survey line L7, C2 is about several kilometers below the northern margin fault of Huaizhuo Basin (F11). On the whole, the deep electrical structure in the study area varies greatly in the horizontal direction. The areas with the greatest variation are consistent with the locations of major faults. At a depth of 20 km, the shallow part of the basin-range structural area to the south of F4 is characterized by a low-resistivity structure, while the overlying area covered by Hannuoba basalts and metamorphic rocks has a high-resistivity structure. At depths 20–40 km, the high-resistivity structure R is sandwiched between the low-resistivity bodies C1 and C2. Further down, the medium is characterized by uniform low resistivity.

Deep seismogenic environment of M6.2 Zhangbei earthquake

In 1998, an earthquake with a magnitude of 6.2 occurred near Zhangbei at the northwest end of the ZBFZ. Since 2021, medium and small earthquakes have frequently been reoccurring in the Zhangbei seismic area, several of them with magnitudes of about 3. The occurrences of these earthquakes suggests that the region may have become seismically active. Figure 10 shows the electrical structure planar section (a and b) at the depth of 12.5 km and two deep electrical sections (LA and LB) across the Zhangbei seismic area. The fine 3D deep electrical structure characteristics of Zhangbei can reveal the seismogenic structure and the possible location of the source of the Zhangbei earthquake and help to understand the deep seismogenic environment of the region.

Existing seismological research literature proposes two estimates of the focal depth of the M6.2 Zhangbei earthquake; one is 8–10 km (Zheng et al., 1998; Gao et al., 2002), and the other is 12–15 km (Yang et al., 2002; Yang and Chen, 2004). There are also controversies with regard to the seismogenic faults (Xu et al., 1998; Cai et al., 2001, 2002; Gao, 2001; Zhou et al., 2019). The deep electrical structure diagrams (LA and LB) across two sections of the Zhangbei seismic area shown in Figure 10 demonstrate that there is a high-low resistivity electrical interface in the crust at the depth of about 15 km in the focal area, which is consistent with the estimation that the focal depth of the Zhangbei earthquake was 12–15 km (Yang et al., 2002; Yang and Chen, 2004). The depths of multiple M3 earthquakes that occurred near Zhangbei in 2021 are projected on the sections LA and LB in Figure 10, where it can be seen that they are located near the Dahezhen-Hailiutu Fault (F3). Ekstrom et al. (2012) suggested that the 1998 M6.2 Zhangbei earthquake resulted from thrust and dextral tectonic activities. Chen et al. (2020) argued that the main compressive stress in Zhangbei was in the NEE-SWW direction, which is almost consistent with the trend of F3. In addition, the east side of F3 is blocked by relatively high resistivity structures and should be characterized by thrust movement, which is in accordance with the focal mechanism solution of the Zhangbei earthquake (Ma et al., 1998). Therefore, the seismogenic fault of the 1988 M6.2 Zhangbei earthquake was the Dahezhen-Hailiutu Fault (F3). The fine structure of seismic region revealed by the magnetotelluric method can be used to constrain the focal depth as well as the seismogenic fault of the earthquake.

Ye et al. (2021) concluded that the M6.4 Yunnan Yangbi earthquake occurred near the high and low resistivity transition zone, on the high resistivity side. He described this phenomenon as “strong earthquake gestation and occurrence in the rigid and flexible transition zone”. This is consistent with the analyses of the 2021 M7.4 Qinghai Maduo earthquake (Zhan et al., 2021). The high-resistivity structure in the upper crust of Zhangbei may represent the Hanuoba basalts, which are very rigid. The faults developed in the Hanuoba basalts can gradually accumulate enough stress to induce moderate to strong earthquakes. The

low-resistivity structure beneath the Hanuoba basalts may comprise mantle magmatic materials (Zhao, 2009; Zhang et al., 2016; Wu et al., 2021). The continuous intrusion and upwelling of these materials may have weakened the stability of the Dahezhen-Hailiutu Fault (F3) in the Hanuoba basalt region, leading to heterogeneous stress accumulation and deformation as well as weakening of the fault, finally resulting in the occurrence of the Zhangbei earthquake. This analysis suggests that the focal depth of Zhangbei earthquake was between 12 and 15 km and its seismogenic fault was F3. The deep seismogenic environment of the earthquake was associated with the magmatic activities in the middle and lower crust and mantle.

3D crust-mantle structure model and seismic risk in western part of Zhangjiakou-Bohai fault zone

Numerous studies have suggested that the occurrence of moderate and strong earthquakes is induced by the sliding of faults due to instability. In the process of fault sliding, layer decoupling is often required to provide conditions for fault dislocation (Unsworth et al., 1997; Becken et al., 2011). According to the magnetotelluric detection results of the 2008 M8.0 Wenchuan earthquake, 2013 M7.0 Lushan earthquake, 2016 M6.9 Menyuan earthquake, 2021 M7.4 Maduo earthquake and 2021 M6.4 Yunnan Yangbi earthquake (Zhao et al., 2012; Zhan et al., 2013, 2021; Zhao et al., 2015, 2019; Ye et al., 2021), the focal areas were near the transition zone of high and low resistivity, on the side of high resistivity. Similar results have been reported for velocity structures, i.e., strong earthquakes occurred in the transition zone between high and low seismic waves velocities, close to the side of high velocity (Zhang et al., 2005; Yang et al., 2011). The 3D deep structure of the western section of the ZBFZ has the clear electrical anomaly zone beneath the Shangyi-Chicheng Fault (F4), the northern margin fault of Huai'an Basin (F7), and the northern margin fault of Huaizhuo Basin (F11). It extends deep and connects with the low resistivity layer in the middle and lower crust. These observations suggest that the three faults are transcrust faults that have the electrical structural characteristics of faults that host moderate and strong earthquakes.

Since the Cenozoic, the collision between the Indian and Eurasian Plates has resulted in the eastward expansion of the asthenosphere (Li et al., 2010). The weak materials in the middle and lower crust of the western section of the ZBFZ were also squeezed eastward. Inhomogeneous bodies exist in the crust and the deep parts of the western segment of Shangyi-Chicheng Fault (F4) and the north margin fault of the Huai'an Basin (F7) are connected with the low resistivity structure (C2) in the crust, which becomes gradually shallower to the northeast along F7. These deep faults provided channels for upwelling of mantle-derived basaltic magma (Xu et al., 1998). In particular, the deep materials near F7 have a stronger effect on the shallow crust. This

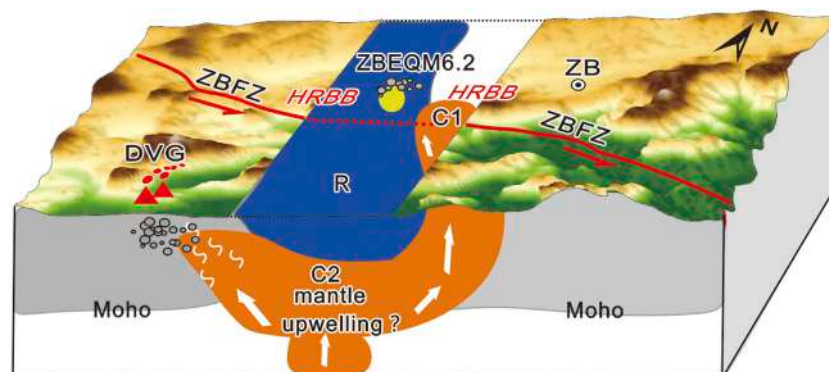


FIGURE 11

Schematic model of the seismogenic environments and formation mechanism for volcanism in the ZBFZ.

deep tectonic and geodynamic environment indicate that this region may be the main area of moderate and strong earthquakes in the future and deserves special attention.

The resistivity structure characteristics of the medium revealed by magnetotelluric imaging combined with previous geological surveys, geophysical exploration and deformation field analysis (Xu et al., 2002; Jia et al., 2009; Zhao, 2009; Li et al., 2010; Guo et al., 2016; Zhang et al., 2016; Shen et al., 2019; Chen et al., 2020; Wang and Shen, 2020; Wu et al., 2021) enabled formulating a model of the deep seismogenic environment and volcanic formation mechanism in the western ZBFZ (Figure 11). In the northwest part of the Zhangbei seismic area in the western ZBFZ, there is an obvious low-resistivity structure (C1), which extends from the depth upward to the depth of 15 km and then gradually disappears (Figure 9). Structure C1 is connected with the low-resistivity structure (C2) under the DVG (Zhou, 2021) in the southwest (Zhang et al., 2016), indicating that C1 may extend to the Datong region in the southwest. Combined with the results obtained by Zhao, 2009, it is speculated that the low-resistivity structure in the middle and lower crust beneath the Zhangbei-Datong region may have originated from partial melting of upper mantle materials. Moreover, there is an obvious electrical boundary zone between C1 and the southeastern high-resistivity structure (R) in the middle and lower crust, which may be a manifestation of the Dahezhen-Hailiutu Fault (F3) beneath the Hanuoba basalt area.

Conclusion

The ZBFZ is an important NW-trending active tectonic zone in North China. It is one of the well-known moderate and strong seismic tectonic zones of the mainland China. Zhangjiakou and other regions in the west section of the ZBFZ are not only characterized by frequent moderate and strong earthquakes, but also contain the DVG and widely exposed basalts. This area provides a good opportunity for studying earthquake and volcanic activity. In

order to reveal the deep electrical structure and seismogenic environment of the western section of the ZBFZ leading to the 1998 M6.2 Zhangbei earthquake, magnetotelluric array exploration was carried out in an area extending for about 113 km from north to south and 125 km from east to west with Zhangjiakou as the center. The 3D deep electrical structure of the area was obtained by 3D electromagnetic inversion imaging. The results show that:

- 1) The deep electrical structure in the western section of the ZBFZ is characterized by significant lateral heterogeneity. To the north of the Shangyi-Chicheng Fault (F4), the upper crust has a high resistivity structure (R), corresponding to the exposed Hanuoba basalts and metamorphic rocks. To the south of F4, zones of high and low resistivities alternate, which is attributed to the complex faults and basin-range structures. In the middle crust, low-resistivity structures (C1 and C2) gradually appear in the northwest and southeast of the study area, showing the characteristics of a high-resistivity structure sandwiched between two low-resistivity structures. In the lower crust, two low-resistivity structures (C1 and C2) gradually join each other and the whole region becomes characterized by low resistivity. The Shangyi-Chicheng Fault (F4), the northern margin fault of Huai'an Basin (F7) and the northern margin fault of Huaizhuo Basin (F11) are definite electrical difference zones, which are connected with the low-resistivity structure in the crust.
- 2) The electrical structure of the upper crust in the 1998 M6.2 Zhangbei earthquake area is dominated by high resistivity. In the distribution area of some fault zones, high and low resistivities alternate. The middle and lower crust mainly consists of large-scale and laterally discontinuous low-resistivity layers. The results of magnetotelluric exploration indicate that the focal depth of the Zhangbei earthquake was between 12 and 15 km and its seismogenic fault was the Dahezhen-Hailiutu Fault (F3), concealed beneath the Hanuoba basalt area. The northern margin fault of the Huai'an Basin (F7) is a transcrustal fault with a deep

extension, and it connects with the low resistivity structure in the middle and lower crust. The Zhangbei seismic area and the distribution region of F7 are potential seismic areas requiring further attention.

- 3) Based on the comprehensive analysis of geological surveys, geophysical exploration and deformation field, it is speculated that owing to the neotectonic activities, the magma originated from asthenosphere mantle rose and erupted on the surface near Datong, forming the DVG. The magma continued to migrate and overflow along the fractures in the NE direction, forming the Hannuoba basalts in the Zhangbei area. The existence of mantle thermal material in this area led to the formation of the Hannuoba basalt area and the accompanied fluids migrated to the middle and upper crust, resulting in fault sliding and frequent occurrence of moderate and strong earthquakes in this region. The continuous upwelling of mantle thermal materials weakened the stability of the fault in the Hannuoba basalts and gave rise to a heterogeneous stress accumulation and deformation. Thus, the fault was weakened and finally the Zhangbei earthquake occurred in 1998.

Data availability statement

The raw data supporting the conclusions of this article will be made available by the authors, without undue reservation.

Author contributions

YP, XS, LZ, and YZ completed the field data acquisition, the data analyses, and wrote the article. YZ provided funding for the study. XL and LW assisted in data acquisition. All authors have read and agreed to the published version of the manuscript.

Funding

This paper was supported by the National Key Research and Development Program of China (2017YFC1500103), the National Nonprofit Fundamental Research Grant of Institute of Geology, China Earthquake Administration (Grant No.

IGCEA2215), the Science and Technology Program of Hebei Province (19975412D), and the National Field Scientific Observation and Research Station Project of Continental Rift Dynamics in Taiyuan, Shanxi Province (NORSTY 20-07).

Acknowledgments

The field data acquisition was supported by Zhangjiakou Emergency Management Bureau and Zhangjiakou Central Station, Earthquake Administration of Hebei Province. We appreciate the access to the computing resources of the Computer Network Information Center at the Institute of Geology, China Earthquake Administration. Some figures were prepared using GMT (Wessel et al., 2013).

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2022.966192/full#supplementary-material>

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 15 June 2022

ACCEPTED 08 August 2022

PUBLISHED 06 September 2022

CITATION

Chen G, Li Z and Huang X (2022), Late
Quaternary faulting and
paleoearthquakes along the northern
section of the Bayanhaote fault and their
implications for
regional seismotectonics.
Front. Earth Sci. 10:970192.
doi: 10.3389/feart.2022.970192

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Late Quaternary faulting and paleoearthquakes along the northern section of the Bayanhaote fault and their implications for regional seismotectonics

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Analysis of the late Quaternary activity and paleoseismicity of the Bayanhaote fault is critical because it is part of the frontier of the propagating Tibetan Plateau and the boundary between the Alxa and Ordos blocks, and the development of a regional seismotectonic model surrounding the Helan Mountains is crucial. We studied offset landforms and paleoearthquakes along the northern section of the Bayanhaote fault and found that it deformed dextrally, with a slight reverse slip to the east. A channel and a deluvial edge on the alluvial terrace, aged between 56.28 ± 4.04 ka and 82.2 ± 5.78 ka, are right-laterally offset and a sag pond formed on the east side of the fault scarp. We calculated a dextral slip rate of 1.0–2.4 mm/a. Three surface-rupturing paleoearthquakes were discovered in the reversely offset strata in a trench south of Sumutu village. We infer that these three earthquakes might be a portion of surface-rupturing earthquakes by comparing them with documented paleoearthquake data along the southern half of the Bayanhaote fault. We established a regional seismotectonic model around the Helan Mountains using our new and published geological and geophysical data. The seismic risk along the dextral Bayanhaote fault to the west of the Helan Mountains is also substantial. The dextral Bayanhaote fault west of the Helan Mountains and the normal fault system in the east constitute the active boundary belt between the Alxa and Ordos blocks.

KEYWORDS

Bayanhaote fault, active fault, late Quaternary, regional seismotectonics, Ordos block, Alxa block

Introduction

The vast Tibetan Plateau was formed by the collision of India and Eurasia, and it is still expanding northward under this tectonic regime, which profoundly shaped the intracontinental tectonic framework and controlled seismicity in western China

(Tapponnier et al., 2001; Yuan et al., 2013). The Alxa block is a tectonic and seismically active area where the Tibetan Plateau interacts with the Ordos and Xingmeng blocks during the outward expansion of the northeastern margin of the Tibetan Plateau (Research Group on Active Fault System around Ordos Massif, 1988; Deng et al., 1999; Huang et al., 2012; Yang and Dong,

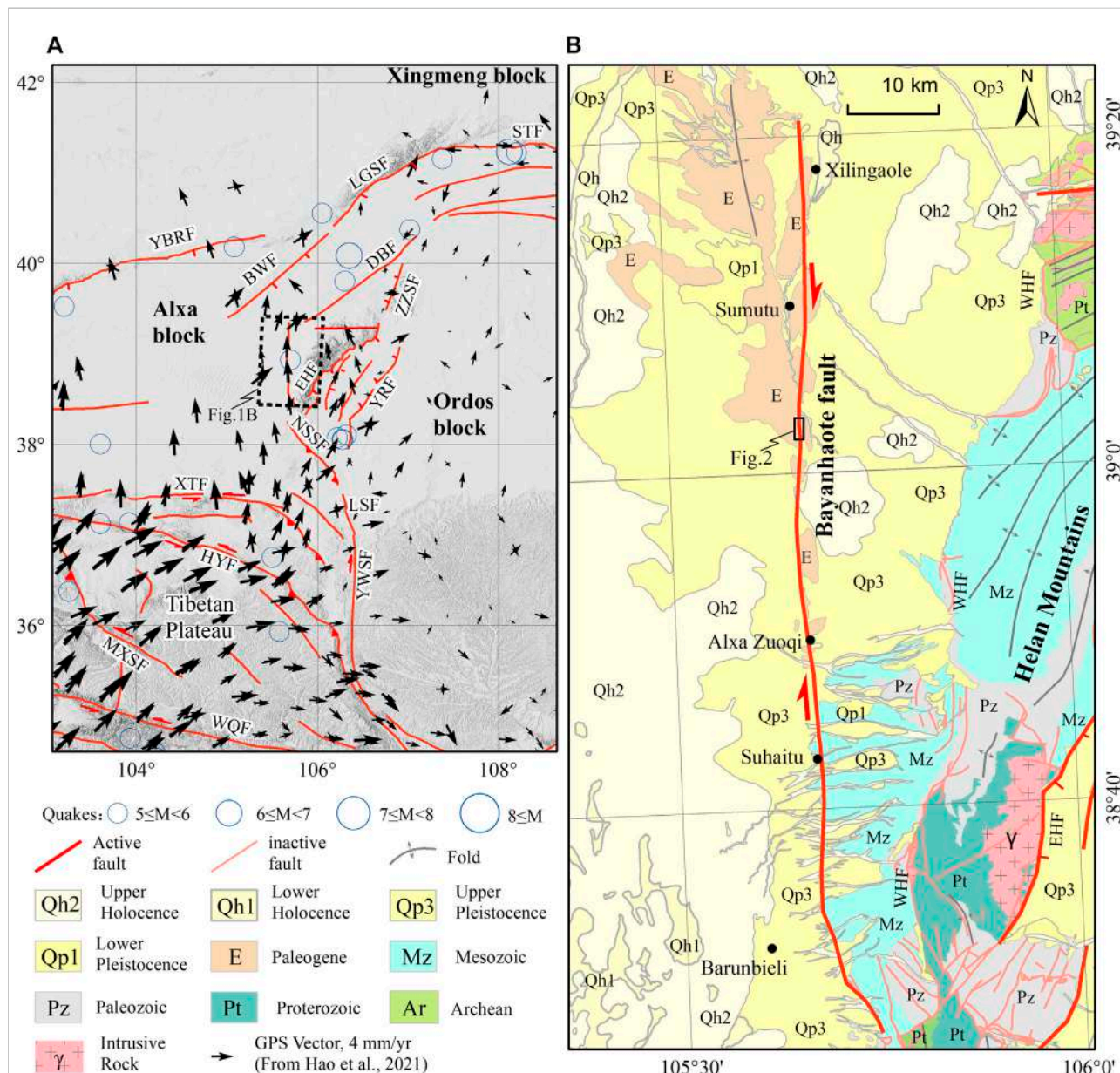


FIGURE 1

Regional geological setting of the Bayanhaote fault. (A) Major active faults, earthquakes, and GPS velocity fields in the western margin of Ordos (fault modified from Xu et al., 2016; earthquake catalog from China Earthquake Network Center; GPS velocity with respect to the Ordos block from Hao et al. (2021)). (B) Structure of the Bayanhaote fault with geological background (geological data from 1: 250,000 scale public geological map via geocloud. <http://geocloud.cgs.gov.cn>). BWF: Bayan Wula fault; EHF: East Helanshan fault; HYF: Haiyuan fault; LGSF: Langshan fault; LSF: Luoshan fault; MXSF: Maxianshan fault; NSSF: Niushoushan fault; STF: Seerteng fault; WHF: West Helanshan fault; WQF: West Qinlin fault; XTF: Xiangshan-Tianjingshan fault; YBRF: Yabrai fault; YWSF: Yunwushan fault; ZSF: Zhuozishan fault.

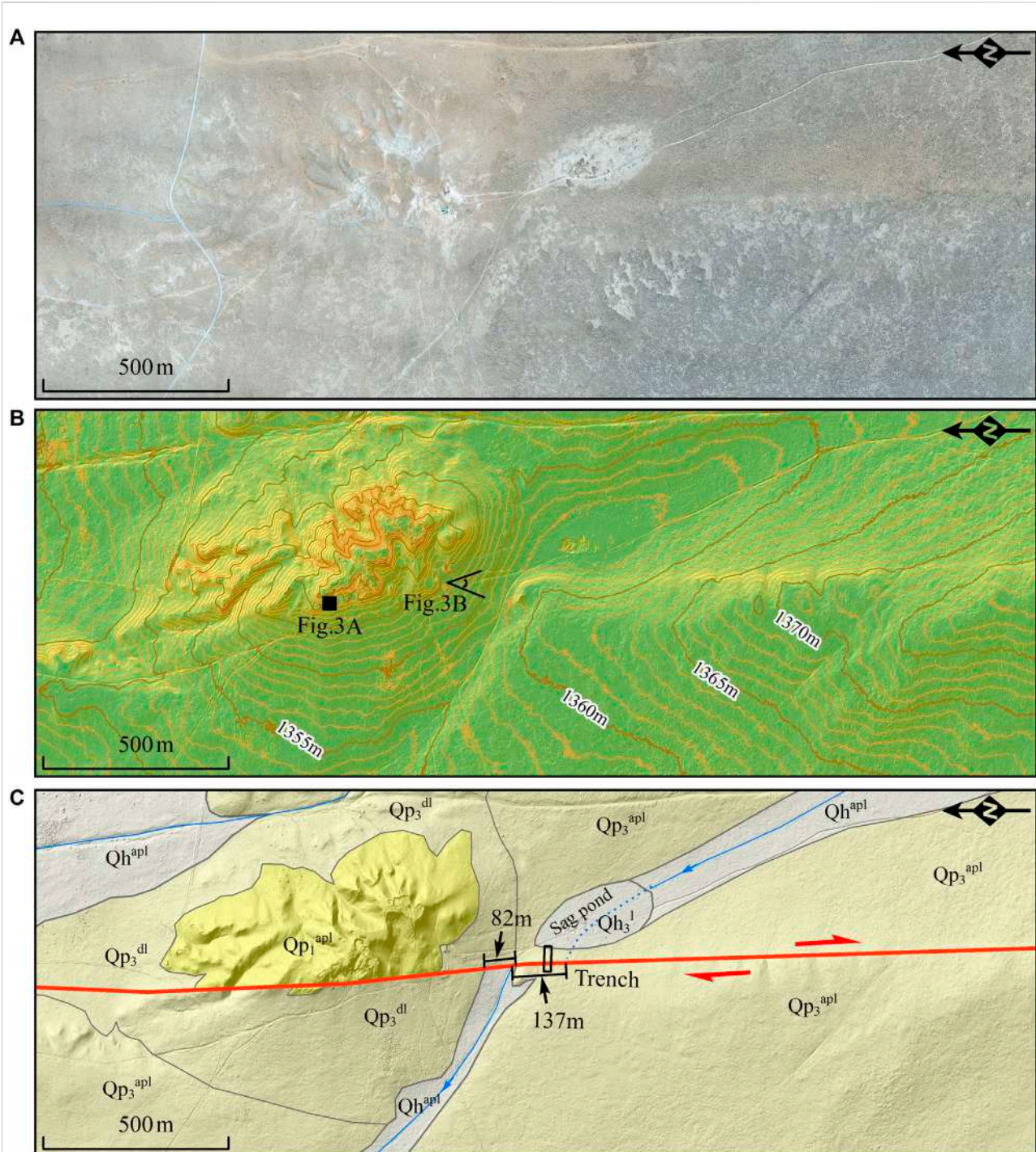


FIGURE 2

Site south of Sumutu. (A) DOM in 5 cm resolution. (B) Shaded relief from DEM in 10 cm resolution. with contours. (C) Offset Geological map and offset landforms. Qh₃^l: active sag pond sediment; Qh^{apl}: Holocene alluvial; Qp₃^{apl}: Late Pleistocene alluvial; Qp₃^{dl}: Late Pleistocene deluvial; Qp₁^{apl}: Early Pleistocene alluvial; Rectangle denotes the trench site.

2018; Zhang et al., 2021). The analysis of faulting and seismicity around Alxa is critical for understanding the deformation behavior and dynamics of intracontinental structures.

Since the Mesozoic Era, the eastern Alxa block has interacted with the Ordos block, resulting in the formation of several graben basins, such as the Bayanhaote, Jilantai, and Yinchuan basins and



FIGURE 3

Field photos from the site south of Sumutu. (A) Fracture zone. (B) Sag pond and fault scarp. See Figure 2B for localities.

the Helan Mountains, which are bordered by these basins (Liu et al., 2010; Yang and Dong, 2018; Ma and He, 2019). A series of active faults developed between the basins and mountains and generated a series of strong historical earthquakes (Figure 1A). Certain faults have been investigated around the Yinchuan Graben, such as the East Helanshan, Yinchuan–Pingluo, and Yellow River faults (Middleton, et al., 2016a; Middleton, et al., 2016b; Lei, 2016; Wei et al., 2021); yet several faults remain unstudied (Zhang et al., 2021). Clarifying the late Quaternary activity of these faults and the occurrence of paleoearthquakes is critical for establishing a regional seismotectonic model. It provides the foundational data for analyzing intracontinental structures in this region, identifying seismogenic sources, and mitigating disasters caused by strong earthquakes.

The Bayanhaote fault is a major active fault in this region that has received little attention, possibly because the fault activated within the Bayanhaote Basin after its formation (Liu and Liu, 2002); however, the fault may be relatively weak in activity (Deng et al., 1999; Lei et al., 2022). In recent years, some scholars have conducted research and mapping work, particularly on the Bayanhaote fault, and they have published new data. However, evidence of late Quaternary activity of the fault is mainly concentrated in the southern section to the south of Alxa Zuoqi (Bayanhaote Town) (Jing et al., 2019; Bi et al., 2020; Lei et al., 2022). There is only evidence of synchronous gully twisting in the northern section, and reliable late Quaternary activity and paleoseismic data remain lacking. The current geology shows that the bedrock along the section to the north of Bayanhaote Town is Paleogene sandstone, whereas the exposed bedrock along the south section is Mesozoic strata (Figure 1B). This variation

may have resulted from the differential activity between the southern and northern sections of the Bayanhaote fault.

We carried out UAV aerial photogrammetry along the northern section of the Bayanhaote fault and obtained a high-precision and high-resolution digital elevation model (DEM) and orthophoto image (DOM) along the fault. The offset late Quaternary landform was analyzed at a site south of Sumutu village, and a trench was excavated, revealing the latest activity and paleoearthquake history of the northern section of the Bayanhaote fault.

Regional tectonics and seismicity

In some studies, the Bayanhaote fault within the Bayanhaote Basin is sometimes referred to as the West Helanshan fault, also known as the fault along the eastern boundary of the basin. In the research of the pre-Cenozoic structure, the Bayanhaote fault is considered a hidden fault in the basin, whereas the West Helanshan fault is a high-angle normal boundary fault between the Bayanhaote Basin and the Helan Mountains (Liu and Liu, 2002; Yang and Dong, 2018; Zhang et al., 2021). Some scholars refer to the active fault crossing Bayanhaote Town (Alxa Zuoqi), the target fault of this study, as the Bayanhaote fault (Huang et al., 2012; Jing et al., 2019; Zhang et al., 2021), while others refer to it as the West Helanshan fault (Bi et al., 2020; Lei et al., 2022). The West Helanshan fault between the Helan Mountains and the Bayanhaote Basin is currently considered dormant owing to lack of data on its activity (Huang et al., 2012). To distinguish the fault within the basin from that along the piedmont, the Bayanhaote fault discussed in this study refers to

TABLE 1 Dating results from Optically Stimulated Luminescence (OSL) samples.

Sample ID	Depth (m)	U (ppm)	Th (ppm)	K (%)	Dose rate (Gy/ka)	Equivalent dose (Gy)	OSL AgeF (ka)
SMT-OSL-9	3.0	0.996 ± 0.02	3.76 ± 0.02	1.72 ± 0.01	2.27 ± 0.10	186.43 ± 10.55	82.2 ± 5.78
SMT-OSL-10	2.0	2.63 ± 0.04	9.77 ± 0.03	2.04 ± 0.01	3.36 ± 0.14	215.73 ± 11.14	64.19 ± 4.22
SMT-OSL-11	2.0	1.69 ± 0.03	5.75 ± 0.03	1.90 ± 0.01	2.75 ± 0.11	242.09 ± 13.56	87.94 ± 6.11
SMT-OSL-12	1.5	2.21 ± 0.02	7.78 ± 0.10	1.97 ± 0.01	3.09 ± 0.13	173.65 ± 10.28	56.28 ± 4.04
SMT-OSL-13	3.0	2.06 ± 0.02	8.62 ± 0.03	1.98 ± 0.01	3.07 ± 0.13	170.05 ± 6.22	55.33 ± 3.04
SMT-OSL-14	2.7	2.66 ± 0.02	8.24 ± 0.10	1.90 ± 0.01	3.11 ± 0.13	153.0 ± 4.63	49.16 ± 2.5
SMT-OSL-15	2.0	5.07 ± 0.02	12.1 ± 0.02	2.18 ± 0.03	4.19 ± 0.17	137.38 ± 7.45	32.79 ± 2.22
SMT-OSL-16	1.6	2.52 ± 0.03	6.54 ± 0.09	1.91 ± 0.02	3.01 ± 0.12	115.96 ± 5.45	38.51 ± 2.4
SMT-OSL-17	1.2	1.89 ± 0.02	8.02 ± 0.10	1.97 ± 0.01	3.04 ± 0.12	41.85 ± 2.87	13.76 ± 1.1
SMT-OSL-18	0.7	1.70 ± 0.02	8.22 ± 0.04	1.96 ± 0.02	3.02 ± 0.12	23.76 ± 0.8	7.86 ± 0.43

Samples were analyzed using the coarse-grained (90–150 μm) quartz single aliquot regenerative-dose (SAR) method (Murray and Wintle, 2000). The OSL dating method followed the process described by Chen et al. (2022). We applied a Central Age Model (CAM) to the samples.



the active fault that developed in the Bayanhaote Basin, which passes through Alxa Zuoqi, as shown in Figure 1B.

The Bayanhaote Basin and other basins around the western margin of Ordos experienced at least four tectonic events after their formation, and substantial tectonic activity was recorded during the Yanshan and Himalayan movements. The latest major tectonic event occurred during the Miocene, with the formation of a series of NW-trending open folds and NE-trending strike-slip faults, which were related to the rapid uplift and expansion of the Tibetan Plateau (Liu et al., 2010; Yang and Dong, 2018; Ma

and He, 2019). According to the focal mechanism from seismic observations, the maximum compressional principal axis of the present stress field in the northern part of the western margin of Ordos is NE-trending (Sheng et al., 2015; Wang et al., 2015; Guo et al., 2017). According to the GPS velocity field, the Tibetan Plateau is obliquely compressed in the NE direction in the northern part of the western margin of Ordos, whereas the Alxa block moves northward relative to Ordos (Figure 1A).

The Bayanhaote fault extends from Xilingaole in the north to Barunbieli in a north–south strike; the strike then becomes eastward, with a total length of >100 km (Figure 1B). It was divided into three sections (Jing et al., 2019; Lei et al., 2022). South of Alxa Zuoqi, the Bayanhaote fault is exposed on alluvial fans to the west of the Helan Mountains. The Bayanhaote fault is associated with the folds north of Alxa Zuoqi. Cenozoic rocks are folded and intermittently distributed along the fault. The western plate of the fault is relatively uplifted, and a secondary sedimentary basin was formed on the east side of the West Helanshan fault (Figure 1B).

A study on the Quaternary activity of the Bayanhaote fault shows that the fault is mainly dextral strike-slip displaced Paleogene rocks for >800 m and gullies for 12–14 m in the northern section (Lei et al., 2022). The Holocene strike-slip rate in the southern section was 1.08 ± 0.1 mm/a, and the average slip rate was 1.38 ± 0.2 mm/a since the late Pleistocene Era (Jing et al., 2019). The vertical slip rate since the late Pleistocene Era was 0.64 mm/a, determined from an offset alluvial fan. Lei et al. (2022) reported a dextral slip rate of 0.2–0.4 mm/a for the southern section.

There were few instrumental records of strong earthquakes and no surface-rupturing historical earthquakes along the Bayanhaote fault. The maximum historical events occurred on 16 June 1991, near the city of Alxa Zuoqi. Well-constrained results on paleoearthquakes are still expected owing to the limited dating data. Lei et al. (2022) excavated three trenches in Xiaosuhaitu and reported three paleoearthquake events between 5.9 ± 0.4 ka and 6.2 ± 0.7 ka, between 13.0 ± 0.2 ka and 24.0 ± 2.0 ka, and >30.54–30.6 ka. Jing et al. (2019) reported that three paleoearthquake events occurred according to their trenches at ~3, 6.0, and 8–11 ka, with a recurrence interval of 3,000–4,000 years. Characteristic displacements of strong earthquakes were proposed from the statistics of displacements based on the DEM, which are a right-lateral slip of ~3 m and vertical displacement of ~1 m (Bi et al., 2020).

Offset landforms at the site south of Sumutu

Erosional landforms developed along the northern section of the Bayanhaote fault, which were accompanied by the uplift of Cenozoic strata. Linear fault scarps were observed along the fault traces. There are no widely distributed deposits or young

landforms that record the activities of the fault. Only a few Holocene rivers and their landforms passed through the fault, which might explain the lack of published data in this section. We discovered a sag pond south of Sumutu that recorded the latest deformation of the fault.

Weathered Paleogene red sandstone and its overlying early Pleistocene gravel layer were uplifted near the fault south of Sumutu (Figure 2), and secondary fractures and fault-damaged zones were developed (Figure 3A). The right-lateral offsetting fault prevents a gully from flowing through the fault from southeast to the west (Figure 2 and Figure 3B). The western plate of the fault is relatively uplifted, forming a linear fault scarp with a height of 3–4 m, and the beheaded gully was abandoned. The blocked gully on the eastern plate of the fault formed a sag pond that continuously stored deposits. At present, the northern boundary of the sag pond is almost aligned with the outlet, and the linear western boundary is approximately 274 m long along the fault scarp. Considering the sediment filling of the sag pond above the gully on the east side of the fault, we estimated the original gully centerline as the center point of the western boundary of the sag pond. Using the gully as a mark, we obtained a right-lateral offset of 137 m on the abandoned terrace at this site. The deluvial, eroded from the uplifted Paleogene sandstone and early Pleistocene gravel layers, covered on the late Pleistocene terraces. Taking the inflection line between the diluvial slope and the terrace as a mark, the offset is 82 m (Figure 2C).

In view of the terrace deposits on the western plate of the fault shown by the trench, the sediments were primarily alluvial deposits. The deposit on the top of the terrace belonged to the same horizon as the reddish-brown silty clay covered in the gravel layer exposed by the trench. The optically stimulated luminescence (OSL) sample SMT-OSL-12 collected in this silty layer was dated at 56.28 ± 4.04 ka (Table 1). The OSL sample SMT-OSL-9 from the alluvial sand layer 3.2 m below the surface was dated at 82.2 ± 5.78 ka. Therefore, we estimated the right-lateral slip rate to be 1.0–2.4 mm/a using the dextral offset of 137 m and 82 m, and the terrace age of between 56.28 ± 4.04 ka and 82.2 ± 5.78 ka.

Paleoearthquake trenching at the site south of Sumutu

We excavated a trench 23 m long and 3–4 m deep across the fault scarp on the west side of the sag pond at the site south of Sumutu. The north wall of the trench (Figure 4) showed that the deformation zone was approximately 18 m. West of fault F1, the uplift plate maintained its original sedimentary strata. The descending plate east of fault F3 continuously accumulated deposits with weak deformation. A series of secondary faults produce an asymmetric semi-flower structure in the deformation zone between faults F1 and F3. The deformation zone can be

divided into two domains based on deformation intensity and properties: the strike-slip zone between faults F1 and F2 is characterized by co-seismic strike-slip fractures and corresponding filling deposits, and the reverse zone between faults F2 and F3 is characterized by west-dipping reverse faults and colluvium in front of the scarp.

The strike-slip zone between faults F1 and F2 is approximately 9 m wide. The filling wedge adjacent to fault F1 is 5 m wide and filled with silt, which could be the result of a wide surface rupture caused by a strong earthquake. There are many vertical faults on both sides. At least two events were distinguished from the cutting relationship between the faults and strata. The filling wedge along fault F2 broke many times, which shows that the deformation in this strike-slip zone is very complex. In this strike-slip domain, the terrace began to suffer erosion after terrace sedimentation and abandonment. Without new deposits, the deformation history of the strike-slip could not be effectively recorded or revealed from the trench profile.

In the reverse zone between faults F2 and F3, the main deformation is concentrated in thrust fault F3, and the strata between faults F2 and F3 thrust eastward and tilt. The footwall of fault F3 receives accumulation from the hanging wall *via* short-distance transportation. The accumulated strata recorded the active history of fault F3.

The stratum units in the reverse zone include the following: yellow-brown silt (U1); yellow-brown silt with gravel (U2), in which the OSL sample SMT-OSL-18 is dated to 7.86 ± 0.43 ka; dark brown silt with gravels (U3), colluvial wedge accumulation; yellowish brown silt (U4), in which the OSL sample SMT-OSL-17 was dated to 13.76 ± 1.1 ka; dark brown sandy clay with gravels (U5), deposit in a filling wedge; yellow-brown silty clay (U6), in which the OSL sample SMT-OSL-15 was dated to 32.79 ± 2.22 ka; yellow-brown fine-silty sands with clay (U7), in which the OSL sample SMT-OSL-14 was dated to 49.16 ± 2.5 ka, and there was no abrupt variation between this layer and U6, which may be deposited under low hydrodynamic conditions in a long period of time and controlled by the filling process of sag pond and the climate change; dark brown gravel-bearing sandy clay (U8), deposit in a filling wedge; dark brown sandy clay with gravels (U9), deposit in a filling wedge; gray-brown coarse sands with gravels (U10); reddish-brown fine silt with gravels (U11), in which the OSL sample SMT-OSL-13 was dated to 55.33 ± 3.04 ka; reddish-brown silty clay (U12), in which the OSL sample SMT-OSL-12 was dated to 56.28 ± 4.04 ka; reddish brown gravel with silt (U13); dark brown sandy clay with gravels (U14), deposit in a filling wedge; yellowish-brown coarse sands with gravels (U15), in which the OSL sample SMT-OSL-11 was dated to 87.94 ± 6.11 ka; yellow-brown gravel with coarse sands (U16); yellow-brown gravel (U17); yellow-brown gravel with coarse sands (U18).

According to the change in sediment grain size, the strata revealed in the eastern part of the trench were divided into three sedimentary series from coarse to fine: U15–U12 is the lowest

sedimentary series from gravel to clay; U11–U7–U6 is a sedimentary series in the middle from silt with gravel to clay; and U4 and its upper layers comprise the uppermost sedimentary series, which is dominated by silt. U3 was a sandwiched layer of silt and gravel. Compared with the stable accumulation on the west side of fault F1, the sedimentary series east of fault F2 was transported under short-distance water flow from the terrace to the west. These sedimentary series were not correlated with regional climate change but were mainly controlled by local slope changes. The slope change corresponds to the fault scarp formed by local co-seismic dip slip. Therefore, the variations among the three sedimentary series may correspond to strong earthquakes, that is, the stratigraphic interfaces U12/U11 and U6/U4 represent two earthquakes with corresponding times between 56.28 ± 4.04 ka and 55.33 ± 3.04 ka and between 32.79 ± 2.22 ka and 13.76 ± 1.1 ka, respectively.

According to the layer thickness and occurrence, U11, U7, and U4 were thicker in the eastern plate of fault F3 than in the western plate. U3 is a colluvium wedge that is deposited rapidly in front of a fault scarp. After a ~ 0.29 -m offset of the top bedding of U4, a rapid deposit was formed on the foot of the fault scarp, which pinched out on the footwall (eastern plate) away from the fault. The dip-slip displacement of U12/U11 and U6/U4 along fault F3 is 1.08 and 0.64 m, respectively, which is consistent with the abovementioned sedimentary series analysis. The three different displacements of the three layers show that these layers recorded different times of co-seismic deformation.

With the constraint of OSL data for layers of event horizons, the three surface-rupturing strong earthquakes occurred between 56.28 ± 4.04 ka and 55.33 ± 3.04 ka, between 32.79 ± 2.22 ka and 13.76 ± 1.1 ka, and between 13.76 ± 1.1 ka and 7.86 ± 0.43 ka. According to the cumulative dip-slip, we calculated the dip slips of these three events to be 0.44, 0.35, and 0.29 m.

Discussion

Activity of the Bayanhaote fault and recurrence of strong earthquakes

According to the offset landforms at the site south of Sumutu and the deformation characteristics revealed from the trench, the northern section of the Bayanhaote fault was active in a right-lateral strike-slip with a certain reverse slip. The western plate of the fault is relatively uplifted, forming a fault scarp. The fault zone shows an asymmetric positive flower structure, as revealed by the trench at the site south of Sumutu. The main deformation zone is composed of vertical faults with a strong strike-slip. It also formed a thrust branch and fault scarp on the east side. The fault zone exposed from trenches at Suhaitu south of Alxa Zuoqi exhibits a negative flower structure (Lei, 2016; Jing et al., 2019; Lei et al., 2022). The southern section of the Bayanhaote fault deformed in dextral slip with normal slip. The fault scarp

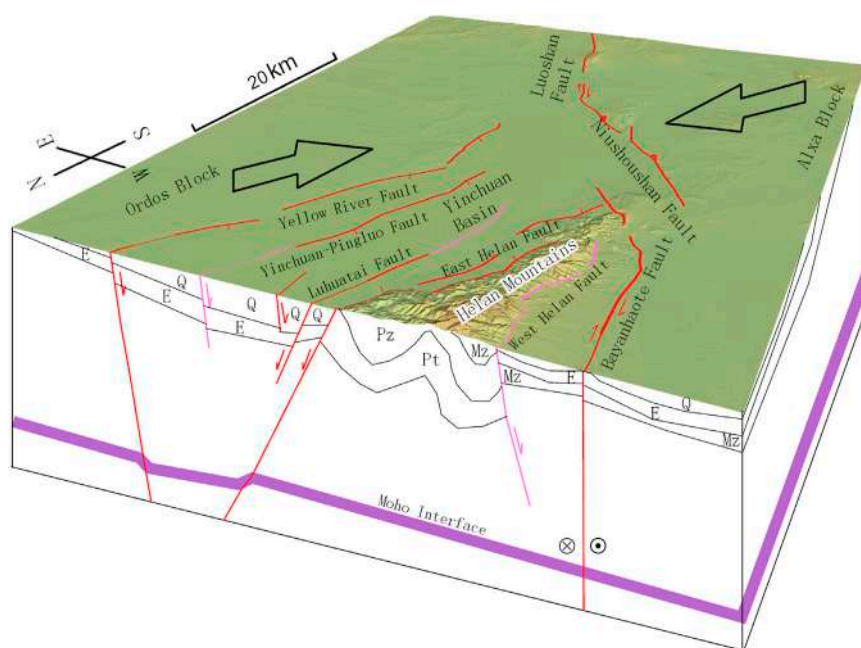


FIGURE 5
Seismotectonic model around the Helan Mountains.

shows relative uplift of the eastern plate along the southern section of the Bayanhaote fault. Although the entire Bayanhaote fault is dominated by strike-slip movement, it behaves differently in normal-slip in the southern section from reverse-slip in the north section.

In this study, we estimated the dextral strike-slip rate to be 1.0–2.4 mm/a in the northern section, which is much larger than that in the southern section, as reported by [Lei et al. \(2022\)](#). [Lei et al. \(2022\)](#) estimated the right-lateral strike-slip rate to be 0.2–0.4 mm/a in the southern section of the fault using the displaced terrace risers and the abandonment age of terraces. [Jing et al. \(2019\)](#) estimated the strike-slip rate to be 1.38 ± 0.2 mm/a using a 16.7-m offset of the lowest terrace with an abandonment age of ~12 ka, which is quite similar to the lower value of our result. We believe that this discrepancy may partially result from time constraints on the offset marks of landforms, especially for terrace risers ([Cowgill, 2007](#)). [Lei et al. \(2022\)](#) and [Jing et al. \(2019\)](#) used the abandonment ages of upper terraces. Considering the differences in fault properties, slip rates, and distribution of Neogene rocks between the southern and northern sections, we propose that the Bayanhaote fault deformed in a way similar to “the tearing” of the Alxa block with respect to the Helan Mountains, which implies the western plate of the northern section acts as the front of the active plate with compressing and uplifting, while the western plate of the

southern section acts as the tail of the active plate with stretching and subsidence.

As revealed by the dating data and structures from the trench at the site south of Sumutu, the Bayanhaote fault has been strongly deformed for 60,000 years in the northern section, and at least three strong surface-rupturing earthquakes have occurred. The most recent surface-rupturing earthquake occurred no earlier than the Holocene. We compared our paleoearthquake results in the northern section with those in the southern section reported by [Jing et al. \(2019\)](#) and [Lei et al. \(2022\)](#). The event between 13.76 ± 1.1 ka and 7.86 ± 0.43 ka is close to the events of 10.15–11.24 ka from [Lei \(2016\)](#) and 8.8–11.66 ka from [Jing et al. \(2019\)](#). The event between 32.79 ± 2.22 ka and 13.76 ± 1.1 ka is consistent with the event between 13.0 ± 0.2 ka and 24.0 ± 2.0 ka from [Lei et al. \(2022\)](#). These suggest two events ruptured the entire Bayanhaote fault. In fact, the three paleoearthquakes were revealed from the dip-slip deformation in the reverse zone in this study, and there are insufficient records to effectively identify paleoearthquakes related to the strike-slip zone in the trench. We infer that the three paleoearthquakes revealed in this study were only a part of these events. Although [Jing et al. \(2019\)](#) published more paleoearthquakes in the southern section, the current paleoearthquake data are insufficient to establish a complete recurrence model for the entire fault.

Seismotectonic model around the Helan Mountains

The tectonic deformation around the Helan Mountains was the product of the interaction between the Alxa and Ordos blocks. Based on GPS and stress field observations, the two blocks were under oblique compression with dextral shear along the north–south-trending boundary. The boundary zone consists of the active Bayanhaote Basin, Yinchuan Basin, and Helan Mountains. The Helan Mountains are triangular, and the hypotenuse of this triangle in the southeast is the Yinchuan Basin, which is controlled by a succession of normal faults running from NE to SW. The Bayanhaote Basin and Jilantai Basin lie to the west and north, respectively. The boundary zone between the Alxa and Ordos blocks was simplified into a structural model, as shown in Figure 5.

A study on the sedimentation of the Yinchuan Basin showed that the NE–SW trending Yinchuan Basin is controlled by a normal fault system under depression and receiving sediments (Wang et al., 2007; Guo et al., 2017). Cenozoic sediments are thick in the northwest part (Cai and Zhou, 2018; Yang and Dong, 2018; Yang and Yang, 2018). With the deep seismic reflection profiles across the Helan Mountains and Yinchuan Basin, the lithospheric structure of this region was revealed, and a negative flower structure was proposed as the geometric combination of normal faults (Fang et al., 2009; Liu et al., 2017). The published data show that the total vertical slip rate of the SE-dipping east Helanshan and Luhutai faults in the Yinchuan Basin is greater than that of the NW-dipping Yinchuan–Pingluo fault and Yellow River faults (Lei, 2016), and the Helan Mountains are relatively uplifted (Liu et al., 2010). We believe that the SE-dipping normal fault system on the west side of the Yinchuan Basin should not be cut and limited by the Yellow River fault in-depth and should extend downward to cut the interfaces between the upper and lower crust and the Moho interface. This model provides a better explanation for the depression at the Moho interface (Fang et al., 2009; Chen et al., 2020). The tectonic status of the Bayanhaote Basin changed from the Mesozoic to Cenozoic. The main active tectonics were transferred from the basin-bounding normal faults to the inner-basin strike-slip fault (Yang and Dong, 2018). The Bayanhaote fault developed as a new boundary fault. As a strike-slip fault, the fault could not extend eastward and connect to the Yinchuan Basin fault system. We believe that it directly cuts down the entire lithosphere and becomes a new block boundary between Alxa and Ordos. This updated tectonic model (Figure 5) integrating current data on the deep structure and fault kinematics provides a new interpretation of the shallow-to-deep structure in the boundary zone between the Alxa and Ordos blocks.

In this regional tectonic model centered around the Helan Mountains, the Luoshan fault is a relatively simple boundary structure between the Alxa and Ordos blocks in the Luoshan area. The right-lateral slip rate of the Luoshan fault (~4.3 mm/a) is considered to represent the relative movement between the Alxa and Ordos blocks (Middleton, et al., 2016b). The GPS stations west of the Bayanhaote fault advance northward at a rate of 2.0–2.5 mm/a relative to the Ordos block (Hao et al., 2021), which approximates the relative movement of the two blocks. These GPS values are similar to the geological data from the Luoshan fault. Lei (2016) calculated a 1.05-mm/a N–S strike-slip caused by the extension of the NE-trending normal faults in the Yinchuan Basin. We may derive a dextral strike-slip rate of about 1.0–1.5 mm/a on the Bayanhaote fault by deducting the slip of the Yinchuan Basin from the overall motion between the Alxa and Ordos blocks, which is close to the dextral strike-slip rate of 1.0–2.4 mm/a obtained from a field investigation in this study. Based on the slip rate, length of the Bayanhaote fault, and normal fault system in the Yinchuan Basin, we infer that the Bayanhaote fault may have the seismogenic capability and seismic risk similar to that of normal fault systems in the Yinchuan Basin.

Conclusion

Data from offset landforms and trench profiles at the site south of Sumutu show that the northern section of the Bayanhaote fault was dominated by right-lateral strike-slip. It also had some reverse slip, which resulted in the relative uplift of the western plate and the formation of a semi-flower structure in the shallow sediments. Based on the ~137 m dextral strike-slip of landforms and the OSL dating data from the trench, we estimated the dextral strike-slip rate of the fault to be 1.0–2.4 mm/a. This estimation of the slip rate was compared with that published by prior studies in the southern section and was consistent with the relative movement between the Alxa and Ordos blocks.

The analysis of strata and faulting in the trench found that three strong earthquakes with significant reverse slips were recorded in the reverse zone within the 18-m-wide fault zone. These events occurred between 56.28 ± 4.04 ka and 55.33 ± 3.04 ka, between 32.79 ± 2.22 ka and 13.76 ± 1.1 ka, and between 13.76 ± 1.1 ka and 7.86 ± 0.43 ka, with a reverse slip of 0.44, 0.35, and 0.29 m, respectively. The deformation distribution in the trench suggested that there were more than three paleoearthquakes. More study is needed to develop a comprehensive paleoearthquake recurrence model for the entire Bayanhaote fault.

We developed a regional tectonic model around the Helan Mountains, including the Bayanhaote fault developed in the Bayanhaote Basin west of the Helan Mountains, by integrating the sedimentation and activity of the normal fault system in the

Yinchuan Basin east of the Helan Mountains. This implies that the Bayanhaote fault was capable of producing strong earthquakes with similar magnitude as the normal fault system in the Yinchuan Basin. The Bayanhaote fault, along with the normal fault system in the Yinchuan Basin on the east side of the Helan Mountains, constitutes the boundary structural belt between the Alxa and Ordos blocks and absorbs the majority of the dextral shear deformation of the two blocks along the N–S boundary.

Data availability statement

The original contributions presented in the study are included in the article/Supplementary Material; further inquiries can be directed to the corresponding author.

Author contributions

GC: conceptualization, funding acquisition, investigation, formal analysis, and writing. ZL: investigation and formal analysis. XH: formal analysis, review, and editing.

Funding

This study was supported by the National Key Research and Development Program of China (Grant No. 2018YFC1504201).

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Acknowledgments

OSL samples were dated in the OSL Laboratory of the National Institute of Natural Hazards, Emergency Management Department. We would like to thank Zhao Junxiang for her help with sample testing. We thank Erdente Gus from the Civil Affairs Bureau of Alxa Zuoqi and Chen Hongbo from Yinchuan for coordinating land use for trench excavation. We are grateful to Shiming Zhang and Feng Li for their helpful discussions and suggestions. We also appreciate helpful comments from the reviewers and the editor Daoyang Yuan.

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OPEN ACCESS

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SPECIALTY SECTION
This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 11 June 2022
ACCEPTED 19 August 2022
PUBLISHED 09 September 2022

CITATION
Zuo H, Qin Y, Liang M, Sun K, Huang F,
Liao C, Zhou W, Wu W, Zhang H and
Yang Y (2022), Geological and
geomorphological evidence of
holocene activity along the maisu fault
in the northern Sichuan–Yunnan block,
Tibetan Plateau.
Front. Earth Sci. 10:966558.
doi: 10.3389/feart.2022.966558

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Geological and geomorphological evidence of holocene activity along the maisu fault in the northern Sichuan–Yunnan block, Tibetan Plateau

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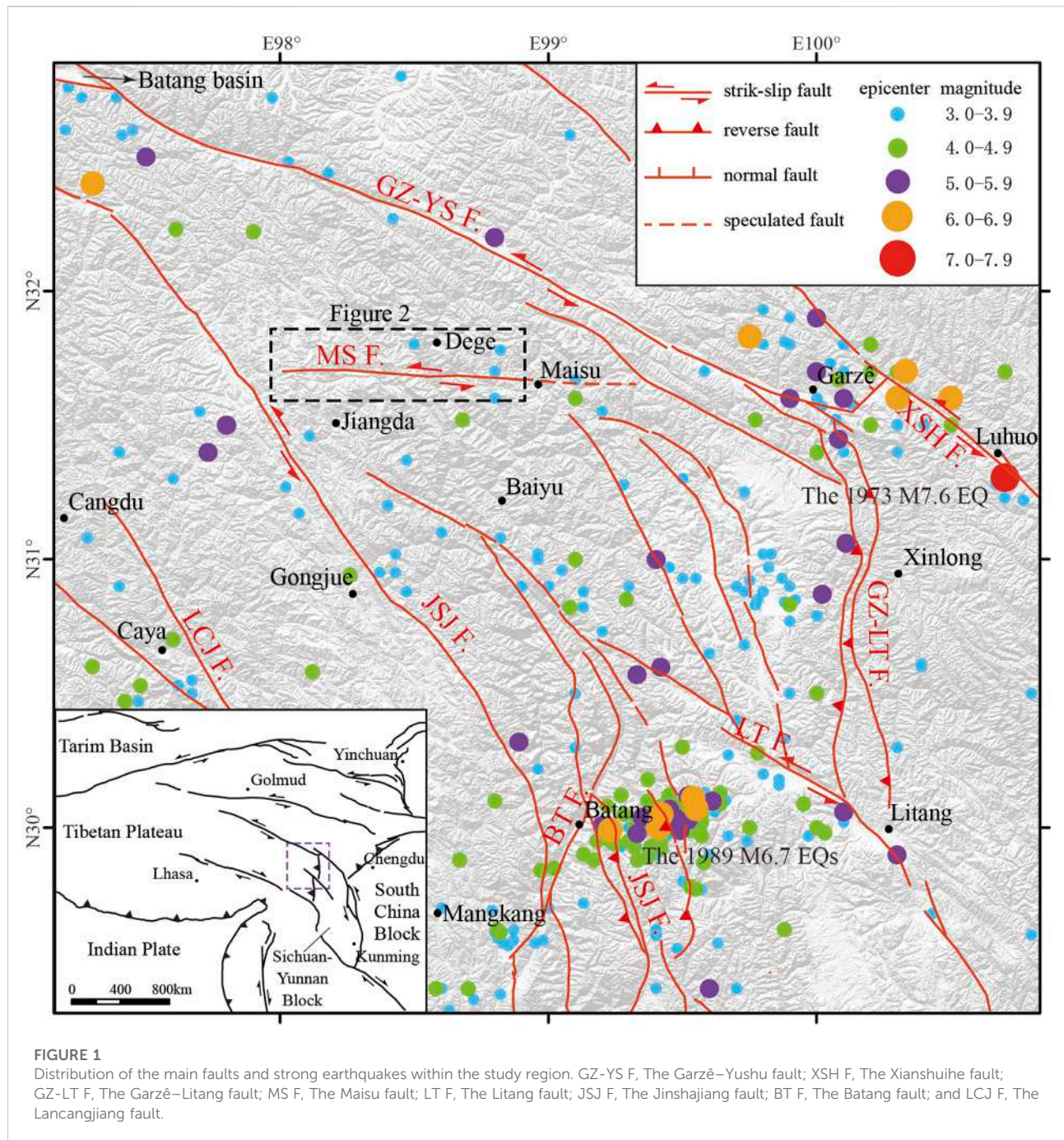
The Sichuan–Yunnan block is located in the eastern of Tibetan Plateau and exhibits strong tectonic and earthquake activity. The Maisu fault is an E–W-trending fault within this block. Via interpretations of remote-sensing imagery and field surveys, we identified a earthquake surface rupture zone that has developed along the Maisu fault; we then estimated its Holocene activity. The surface rupture extends westward from the town of Puma, Sichuan Province, to the village of Worilong, Xizang Province, and has a length of approximately 45 km. According to a fault outcrop and carbon-14 dating of a profile near the village of Yongqu, Xizang Province, the most recent earthquake along this rupture may have occurred after 1850 ± 30 BP. The Maisu fault extends eastward and may intersect the Garzê–Yushu fault. Accordingly, as a secondary fault, the Maisu fault likely accommodates the partitioned horizontal slip deformation of the Garzê–Yushu fault.

KEYWORDS

Maisu fault, Garzê–Yushu fault, Holocene activity, surface rupture, Sichuan–Yunnan block

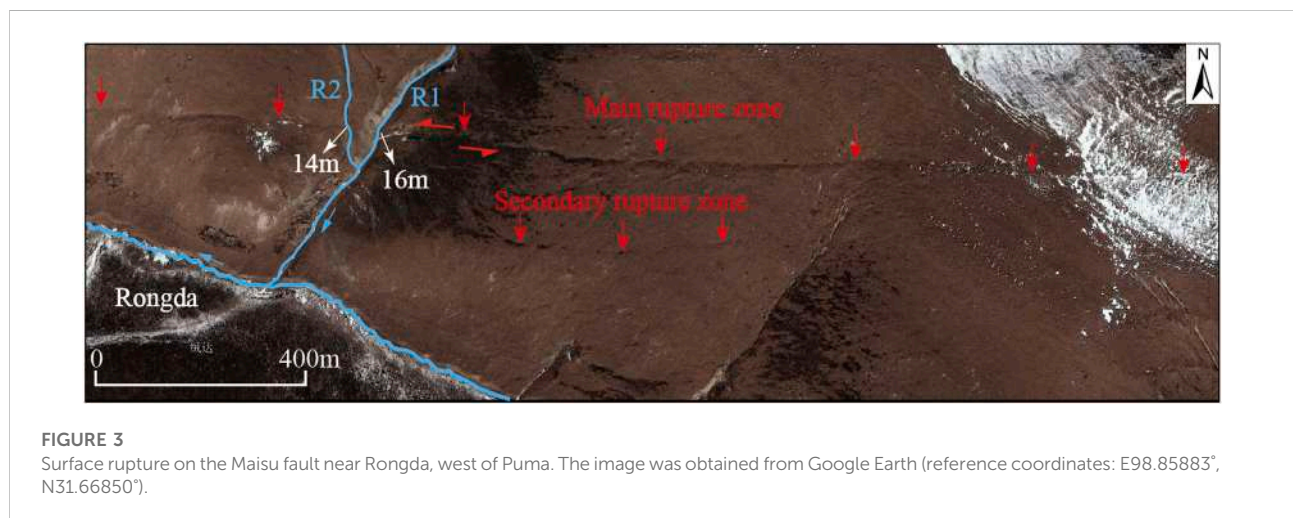
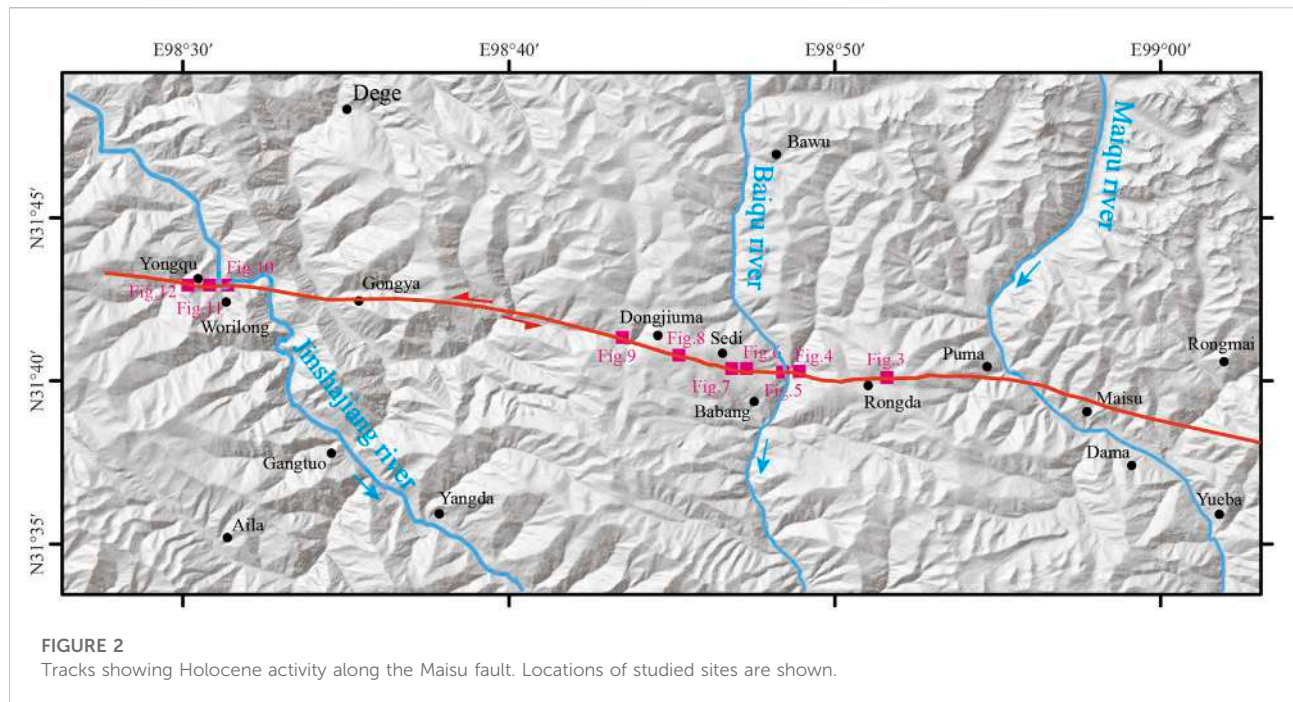
1 Introduction

The Sichuan–Yunnan block in the eastern Tibetan Plateau is located in a region of China that has strong tectonic and earthquake activity (Deng et al., 2014; Yuan et al., 2020). In the eastern margin of the plateau, several strong and major earthquakes (i.e., the 2008 Mw7.9 Wenchuan, 2010 Mw6.9 Yushu, 2013 Mw6.6 Lushan, and 2017 Mw6.5 Jiuzhaigou earthquakes) have occurred over the past 2 decades (Xu et al.,



2008; Wen et al., 2009; Lin et al., 2011; Zhou et al., 2014; Yi et al., 2017). The occurrence of these strong earthquakes is closely related to active faults along the eastern margin of the Tibetan Plateau, and the Garzê-Yushu fault is one such left-lateral strike-slip fault associated with major earthquake activity. The Garzê-Yushu fault and the Xianshuihe fault form a left-stepping pattern and together constitute the northern boundary of the Sichuan-Yunnan block (Zhou et al., 1996; Wen et al., 2003; Xu et al., 2003; Zhao et al., 2021). From the

southeast to the northwest, the Garzê-Yushu fault can be divided into five segments: the Garzê, Manigange, Dengke, Yushu, and Dangjiang segments. Major earthquakes have occurred along these fault segments, e.g., the 1866 M7.5 earthquake on the Garzê segment, the 1896 M7.3 earthquake on the Dengke segment, the Yushu M7.1 earthquake on the Yushu segment, and the 1738 M > 7 earthquake on the Dangjiang segment (Wen et al., 2003; Lü, 2017).

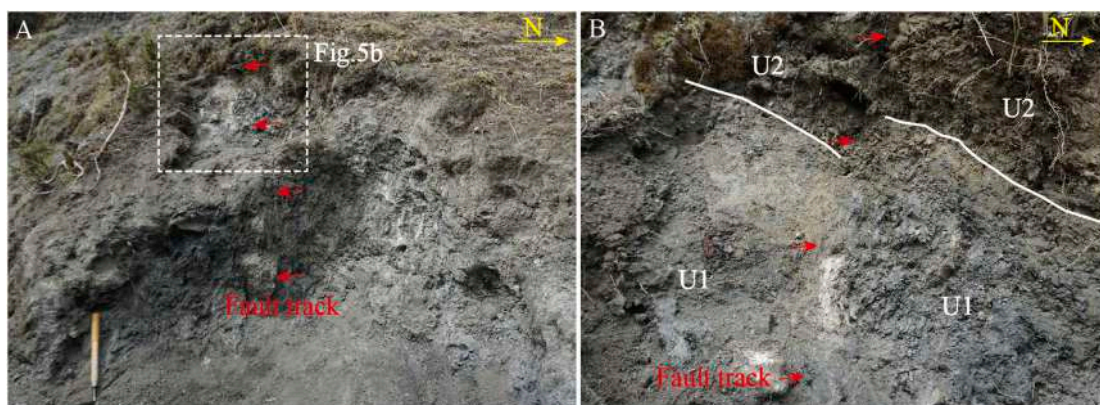


There are also several E–W-trending faults on the southern side of the Garzê–Yushu fault, such as the southern boundary fault of the Batang Basin. This fault, which intersects the Garzê–Yushu fault at an angle of 30°, not only exhibits Holocene activity but also accommodates the horizontal slip rate of the Garzê–Yushu fault (Huang et al., 2015; Lü, 2017). South of Dege County, there is another E–W-trending fault called the Maisu fault, which cuts the Triassic and Tertiary strata forming a linear negative topography. The Maisu fault is a left-lateral strike-slip fault that was thought to be a late-Pleistocene-

active fault (Xu et al., 2016). Our field study, however, indicates that the Maisu fault is a Holocene-active fault. This finding has important implications for our understanding of the tectonic deformation and regional strain partitioning of the Sichuan–Yunnan block. Furthermore, it is useful to study the characteristics of the regional strong earthquake activity in this block. In this article, on the basis of high-resolution satellite imagery and detailed fieldwork, we map the surficial track of the Maisu fault and present geological and geomorphological evidence of its Holocene activity.

**FIGURE 4**

Seismic surface rupture zone near the Baiqu River northeast of Babang. (A) Seismic scarp and waterfall developed along the surface rupture zone (E98.80997°, N31.67147°), and (B) seismic scarp with a height of 0.5 m (E98.80790°, N31.67189°).

**FIGURE 5**

Fault outcrop on a hillside along the Baiqu River near Babang (E98.80776°, N31.67190°). (A) Photograph showing that the fault has dislocated all of the stratigraphic units in the outcrop profile, and (B) close-up photograph of unit U2, which was dislocated by the fault.

2 Tectonic setting

The study area is located in the northwestern margin of the Sichuan–Yunnan block. It lies between the Jinshajiang suture and the Garzê–Litang suture but is closer to the former. The Jinshajiang suture experienced the Indosinian subduction–collision orogeny and the late Yanshanian–Himalayan strike-slip transition (Hou et al., 1995; Wu et al., 2014; Zeng and Xu, 2019; Xia and Zhu, 2020). Within the study region, there are NW–SE-trending, NNE–SSW-trending, E–W-trending, and S–N-trending fault systems controlled by the Jinshajiang, Garzê–Litang, and Garzê–Yushu fault zones. The S–N-trending Jinshajiang fault zone was produced by the closure of the Tethys Ocean. This complex fault zone has a width of approximately

50 km from east to west and comprises five main faults (Xu et al., 1992; Zhou et al., 2005). The southern segment of the Jinshajiang fault zone exhibits Holocene fault activity (Zhou et al., 2005; Xia and Zhu, 2020). Many M6.0–6.9 earthquakes have been documented as occurring along this segment (Zhou et al., 2005; Wu et al., 2019). Its northern segment also exhibits late Quaternary activity (Wu et al., 2019); however, no historical records of strong earthquakes exist. To the south, the NE–SW-trending Batang fault intersects with the Jinshajiang fault zone at an acute angle, showing obvious Holocene dextral strike-slip fault activity; in addition, the 1870 M7.3 earthquake occurred on this fault (Zhou et al., 2005). The NW–SE-trending Litang fault is located on the northeastern side of the Jinshajiang fault, showing left-lateral strike-slip fault activity during the Holocene (Xu et al., 2005;

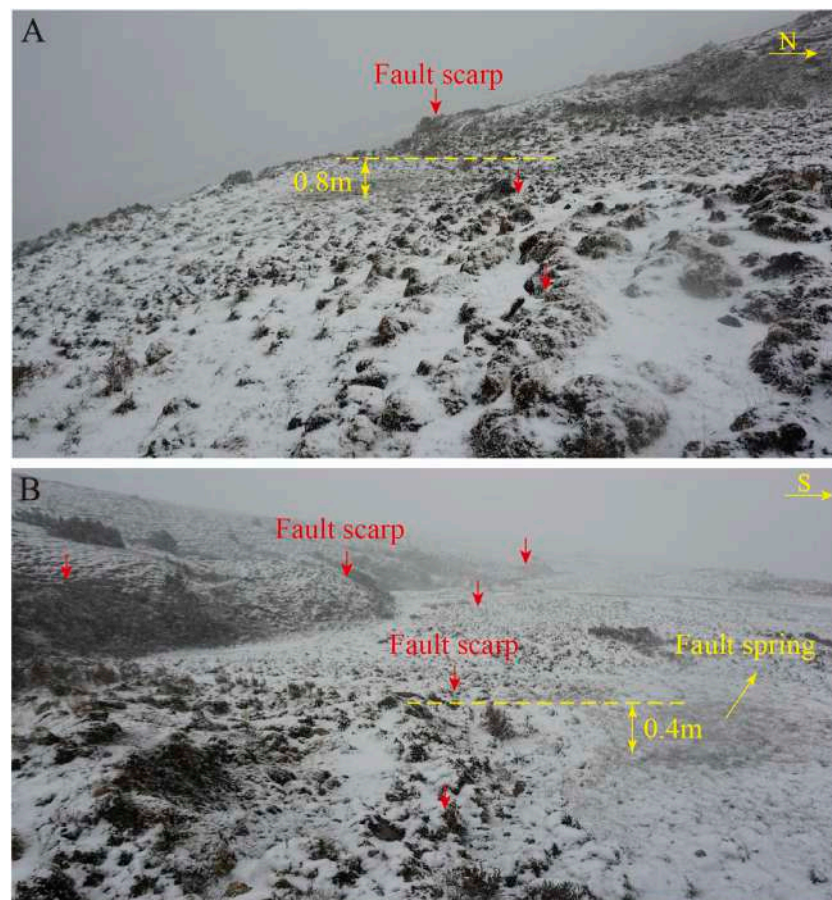


FIGURE 6

Track of the seismic surface rupture near the village of Sedi, close to Babang. **(A)** Seismic scarp developed on the alluvial fan (E98.78315°, N31.67171°), and **(B)** fault and seismic scarp developed on the alluvial fan, with a fault spring developed under the scarp (E98.78173°, N31.67198°).

Zhou et al., 2005; Chevalier et al., 2016). This fault controlled the formation and evolution of Tertiary–Quaternary basins such as the Maoyaba, Litang, and Jiawa basins. The Litang fault also produced the 1948 M7.3 and 1890 M > 7 earthquakes (Xu et al., 2005; Zhou et al., 2005; Chevalier et al., 2016). The Garzê–Yushu fault constitutes the northern boundary of the Sichuan–Yunnan block. It is a Holocene-active sinistral strike-slip fault, and its horizontal slip rate is estimated to have been 5–12 mm/a in the late Quaternary (Zhou et al., 1996, 1997; Wen et al., 2003; Peng et al., 2006; Chevalier et al., 2017). Earthquakes with magnitudes of M7.3 (1986), M7.7 (1854), M7.3 (1896), M7.1 (2010), and M > 7 (1738) have occurred along the Garzê–Yushu fault (Wen et al., 2003; Lin et al., 2011; Li et al., 2016; Lü, 2017). The Maisu fault is located to the south of the Garzê–Yushu fault, and its eastward extension may intersect the Garzê–Yushu fault. These active faults constitute the tectonic framework of the study area and control the regional tectonic deformation.

The current fault activity within the study area is unevenly distributed. Specifically, the distribution of destructive earthquakes is controlled by the active faults. Recent major

earthquakes have concentrated primarily along the Xianshuihe fault and in the Jinshajiang fault zone (Figure 1). For example, the Luhuo M7.6 earthquake in 1973 occurred on the Xianshuihe fault and the M6.7 earthquake in 1989 occurred in the northern section of the Jinshajiang fault zone. However, current fault activity in other areas, dominated by small earthquakes with magnitudes of $\leq M5$, is more scattered.

3 Materials and methods

In this study, we used high-resolution satellite images (Google Earth) and 30-m-resolution Shuttle Radar Topography Mission data to estimate the track of late Quaternary activity along the Maisu fault. On the basis of interpretations of these data and tectonic geomorphological survey data, we determined the fault track of the Maisu fault using geomorphological markers such as fault scarps, deflected stream channels and terraces, and linear fault valleys.

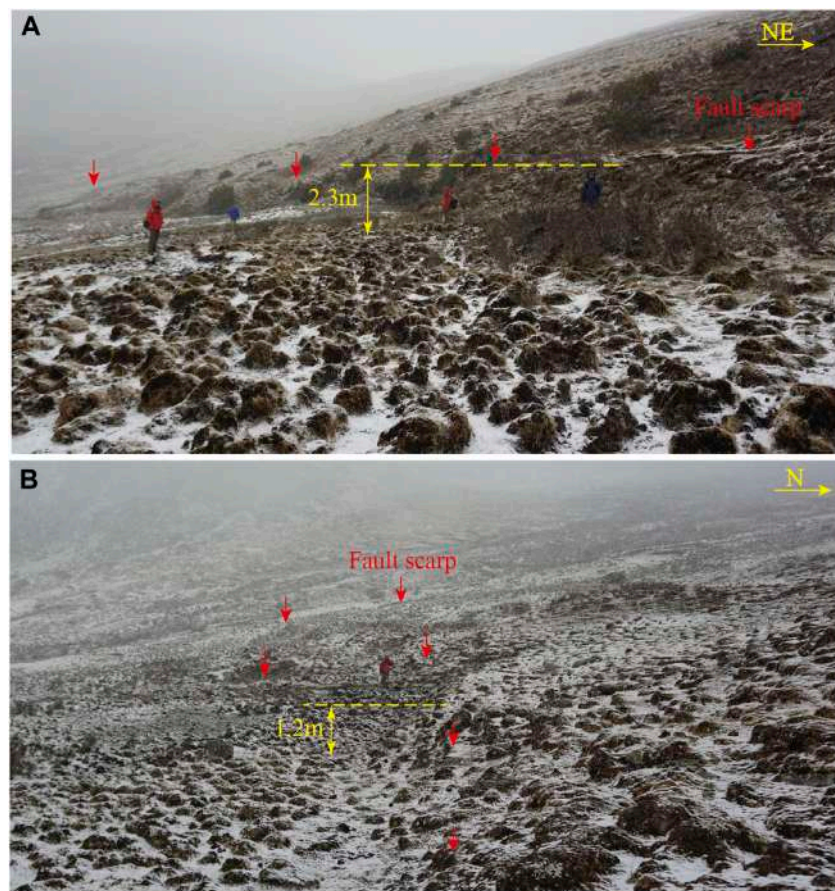


FIGURE 7

Track of the seismic surface rupture near Sedi, close to Babang. **(A)** Linear fault scarp and seismic scarp ($E98.78100^{\circ}$, $N31.67214^{\circ}$), and **(B)** partial segment rupture zone where the zone diverges into multiple branches ($E98.77832^{\circ}$, $N31.67261^{\circ}$).

In our field survey, unmanned aerial vehicle technology was used to obtain high-resolution geomorphological images. Using the ESRI ArcMap software, we generated digital images of the tectonic deformation. Via an analysis of these images and the field survey data, we examined the evolution of the deformed landforms and their cumulative displacement as a result of fault motion. Furthermore, a geological outcrop and the carbon-14 dating method allowed us to identify Holocene activity along the Maisu fault. The carbon-14 samples were analyzed at a laboratory operated by Beta Analytic Inc. United States. On the basis of the geomorphological markers and the radiocarbon dating of the strata displaced by the fault in the geological outcrop, we estimated the Holocene activity of the Maisu fault. Finally, we used formulas suggested by Wells and Coppersmith (1994) to calculate the magnitude of the latest strong earthquake that occurred along the Maisu fault.

4 Results

An analysis of the remote-sensing and field survey data revealed Holocene activity along the Maisu fault, as well as well-preserved surface ruptures. The Jinshajiang River and its main tributaries, such as the Baiqu and Maiqu rivers, pass through the Maisu fault and are obviously left-laterally displaced by the fault (Figure 2). Furthermore, the surface rupture zone on the fault extends from the town of Puma to north of the towns of Babang and Gongya in Dege County, Sichuan Province, and then finally disappears near the village of Worilong, near the town of Gangtuo, Xizang Province. The fault has a length of approximately 45 km (Figure 2). From east to west, some representative sites are described as following.

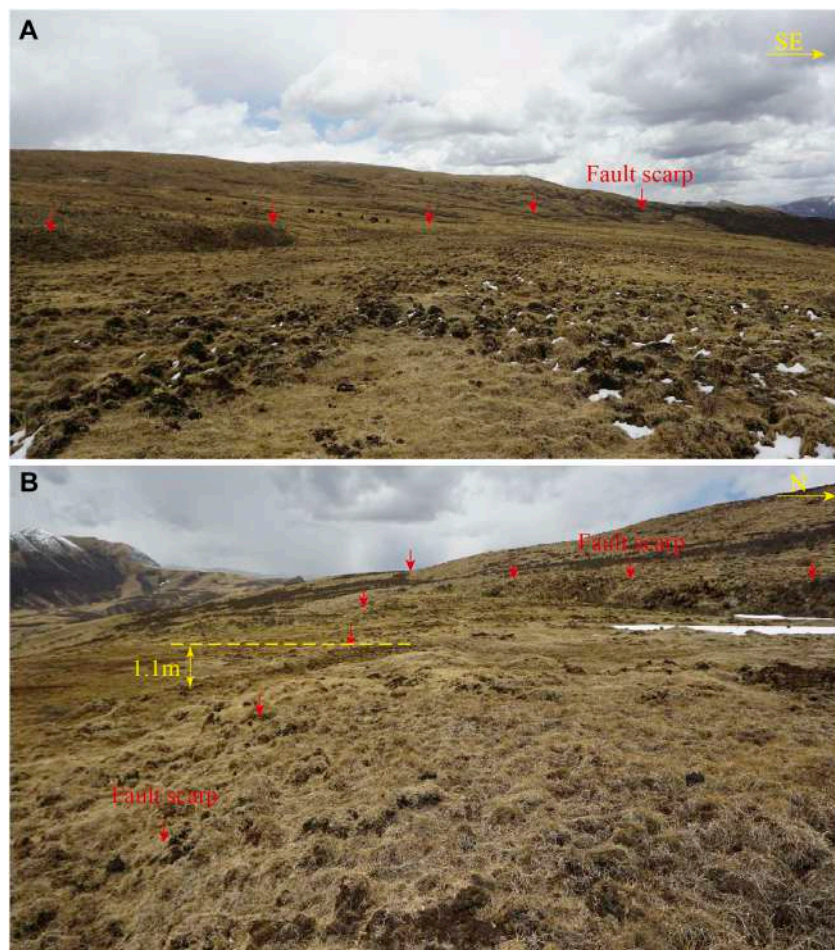


FIGURE 8

Seismic surface rupture track near Dongjiujima, close to Babang. **(A)** Fault scarp developed on the piedmont alluvial fan (E98.75844°, N31.67840°), and **(B)** seismic scarp developed in front of the older fault scarp (E98.75308°, N31.679,021°).

4.1 The rongda site

In Rongda, west of Puma, many geomorphological markers of recent Maisu fault activity are well preserved; for example, there exists a well-defined linear track of the recent fault activity, fault scarps, and gullies that are left-laterally displaced by 16 m (Gully R1) and 14 m (Gully R2) (Figure 3). This segment is close to the eastern end of the rupture zone, where it spreads out into multiple secondary rupture tracks (Figure 3).

4.2 The babang site

Approximately 3.2 km northeast of Babang, the surface rupture has formed a shallow linear trough on the hillside on the eastern bank of the Baiqu River; in addition, it has formed a fault scarp with a height of 0.6 m on terrace T1 and a waterfall on the Baiqu River

(Figure 4A). There is a linear fault scarp with a height of approximately 0.5 m on another hillside along the Baiqu River (Figure 4B). Moreover, a natural fault outcrop is exposed, revealing that the fault displaced the top alluvial gravel and sand layer (Figure 5A). As shown in Figure 5B, unit U1 is composed of dark gray silt and gravel, representing a sag pond deposit facies. Unit U2 is composed of brown sand and gravel, representing a diluvial deposit facies. The fault orientation is approximately 280°, and the fault is steeply dipping. The fault displaced unit U2 upward to the ground surface and formed the surface rupture, likely reflecting the most recent major earthquake (Figure 5B).

4.3 The sedi site

At the Sedi site (approximately 3.0 km northwest of Babang), the earthquake surface rupture is also obvious. The rupture zone

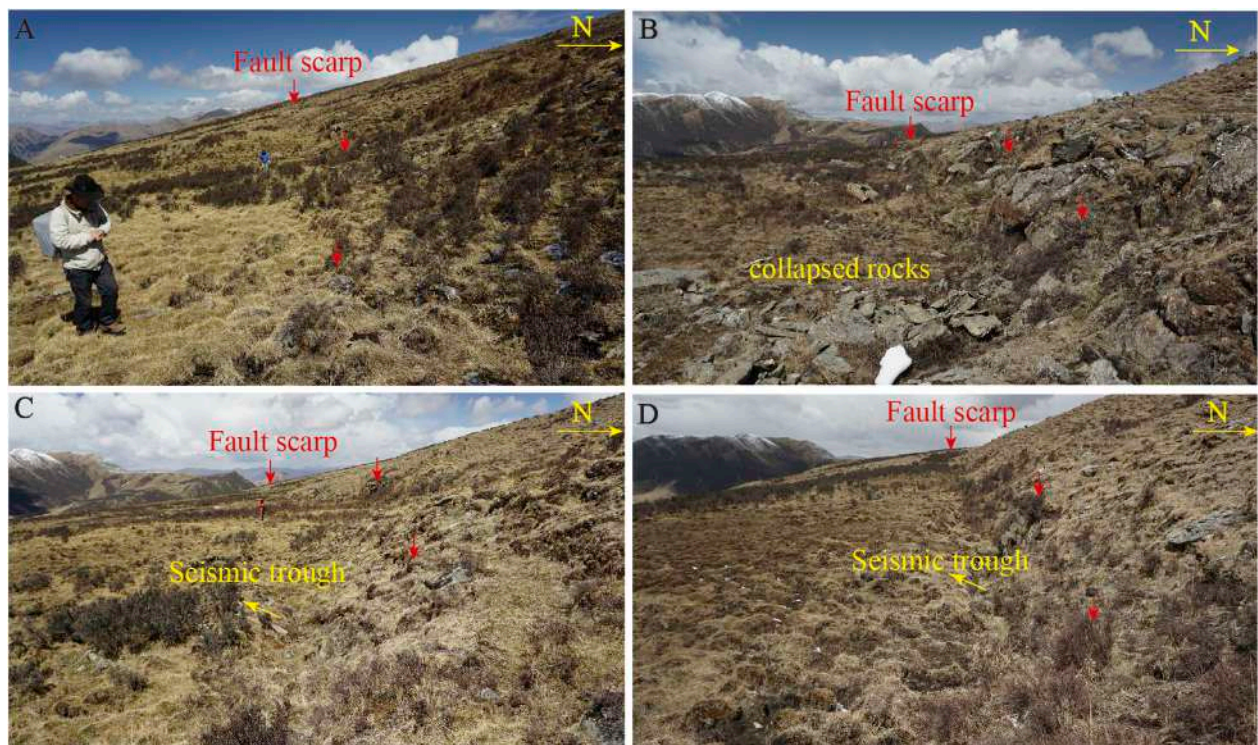


FIGURE 9

Seismic surface rupture track near Dongjiurna, close to Babang. (A) Seismic scarp (E98.74541°, N31.68088°), (B) seismic scarp and collapsed rocks under the scarp as a result of an earthquake (E98.74636°, N31.68085°), and (C) and (D) seismic troughs developed under the scarp as a result of an earthquake (coordinates E98.74887°, N31.6807° and E98.75156°, N31.67963°, respectively).

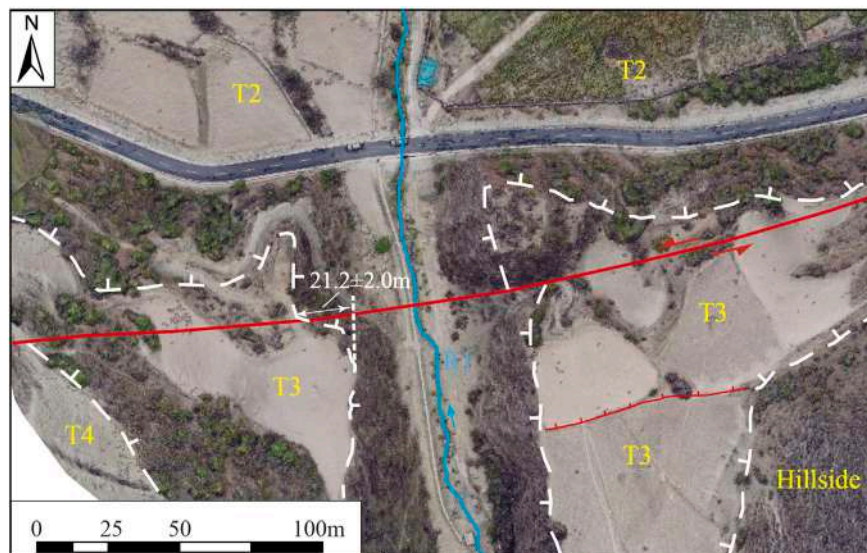


FIGURE 10

Terrace (T3) of the Jinshajiang River left-laterally displaced by the Maisu fault, near Worilong, close to Gangtuo, Xizang Province (E98.51997°, N31.71514°). The image was acquired using unmanned aerial vehicle photogrammetry. The white dashed lines represent the terrace riser. R1 is a tributary of the Jinshajiang river.

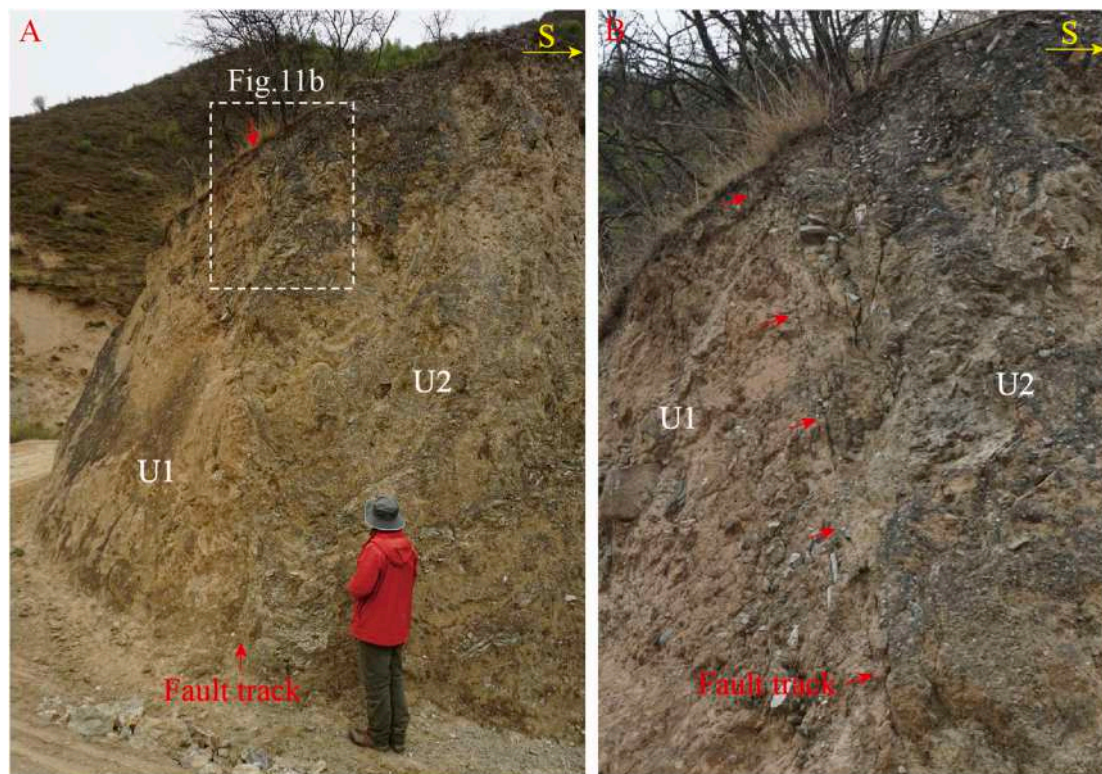


FIGURE 11

Fault outcrop of the Maisu fault near Yongqu, close to Gangtuo (E98.51560°, N31.71525°). (A) Photograph showing the fault outcrop, unit U1 is composed of alluvial sand–gravel, and unit U2 is composed of tectonic breccia and clastic rock from the fault zone. (B) Close-up photograph of the area shown by the dashed rectangle in panel A.

developed along the older fault scarp or formed a new fault scarp on the alluvial fan in the front of the older scarp. As shown in Figure 6A, the surface rupture zone developed on the latest alluvial fan with a scarp height of approximately 0.8 m, while the older fault scarp was heavily degraded (Figure 6A). Figure 6B shows that the surface rupture developed nearly parallel to the older fault scarp, forming a fault scarp with a height of 0.3–0.6 m (Figure 6B), and a fault spring was exposed on the front of the new scarp. The surface rupture also developed on top of the older fault scarp (Figure 7A) or the partial segment rupture zone diverged into multiple branches (Figure 7B).

4.4 The dongjiuma site

Just east of the village of Dongjiuma, the Maisu fault cuts through the piedmont alluvial fan, where it forms a linear fault scarp with a height of 1.5–3.0 m. The surface rupture developed along the bottom edge of the older scarp (Figure 8A) or cut through the newest alluvial fan in front of the older scarp to form a new fault scarp (Figure 8B). Extending to Dongjiuma, the fault cuts through the hillside, where a fault scarp has developed, as

shown in Figure 9A. The field survey revealed collapsed debris in front of some segments of the older scarps (Figure 9B) or a seismic trough under the fault scarp (Figures 9C,D).

4.5 The worilong site

The Maisu fault extends westward, cuts across the Jinshajiang River just south of Gongya, and then enters the Xizang Province. The surface rupture zone extends to Gangtuo and then gradually disappears. At the Worilong site (northwest of Gangtuo), the Maisu fault cuts through terrace T3 of the Jinshajiang River; the terrace has been left-laterally displaced by 21.2 ± 2.0 m (Figure 10). Furthermore, multiple fault scarps are developed on the terrace. We identified a fault outcrop exposed on the southern side of a road near the village of Yongqu, close to Gangtuo. The fault profile revealed two main units (Figure 11A). Unit U1 consists of Quaternary strata and is composed of sand and gravel, representing a diluvial deposit facies (Figure 11B). Unit U2 is composed of tectonic breccia and clastic rock from the fault zone. The fault has displaced these two strata up to the ground surface (Figure 11B).

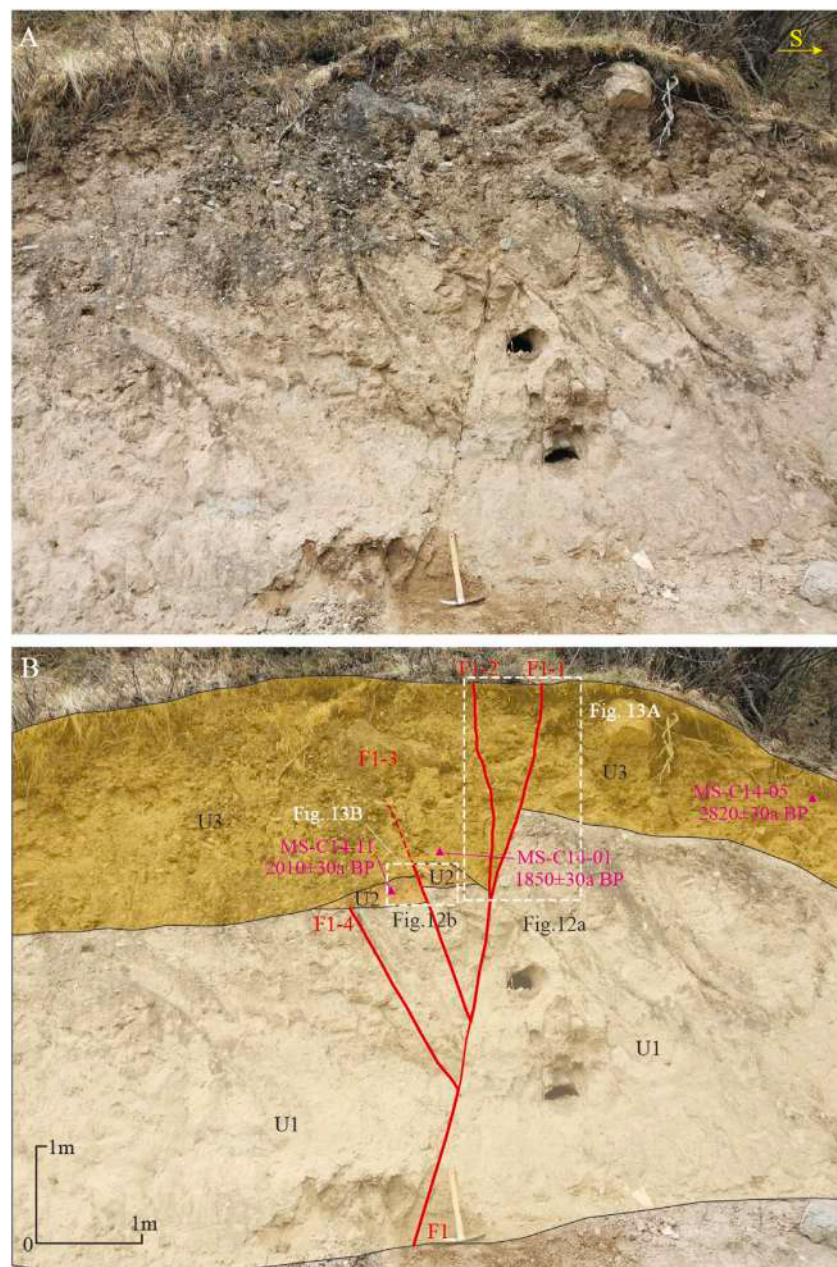


FIGURE 12

Fault outcrop of the Maisu fault near Yongqu, close to Gangtuo (E98.51321°, N31.71495°). According to the dating results of samples collected from the strata dislocated by the fault, the Maisu fault is a Holocene-active fault. **(A)** Photograph showing the fault outcrop. **(B)** Interpretation of the fault outcrop and the locations of the dating samples.

4.6 The yongqu site

We identified another fault exposure at the Yongqu site. Cleaning the outcrop revealed a very obvious upward-branching fault that cuts across the Quaternary stratum. Three stratigraphic units were revealed, labeled U1–U3 from bottom to top (Figures 12, 13). Unit U1 is composed of yellow silt–fine sand and contains some gravels. The

lower part of Unit U2 is composed of small gravel with poor roundness. Its upper part is a fine sand layer that pinches out toward the north. Unit U3 is a sand–gravel layer. The gravels in unit U3 are sub-rounded, and their diameters range widely from 5 cm to 20 cm. On the basis of the characteristics of the sedimentary strata and the deformation caused by the fault, we identified two earthquake events. In event I, unit U1 was displaced by the faults, forming a

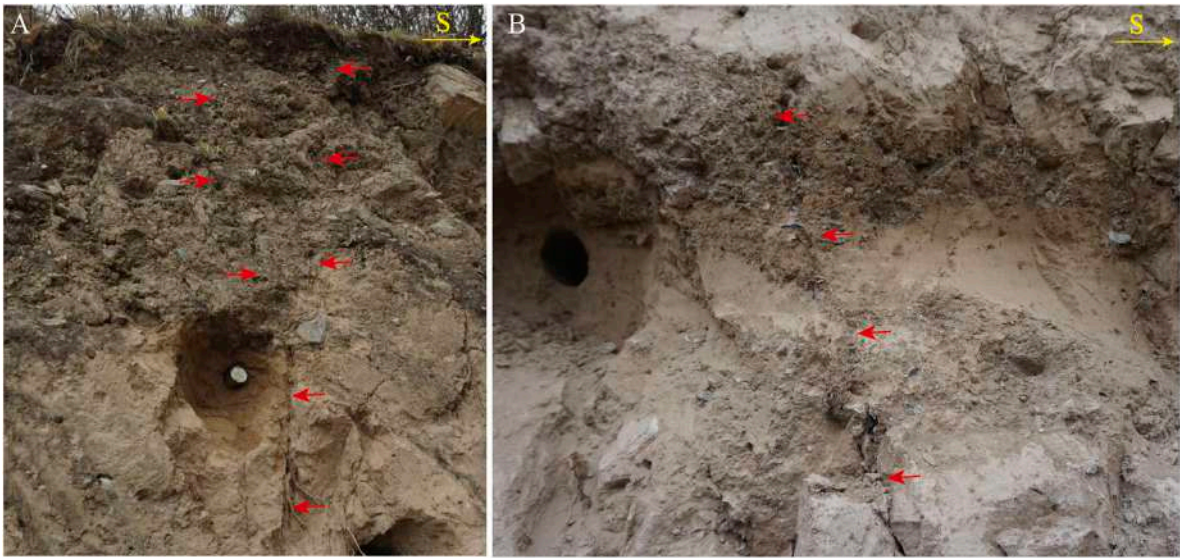


FIGURE 13
Photographs showing partial details of the Maisu fault. (A) and (B) close-up photographs of the areas defined by the large and small white dashed rectangles, respectively, shown in Figure 12B.

TABLE 1 Radiocarbon ages of the samples collected along the Maisu fault.

Sample	Lab.no	Description	13C/12C (o/oo)	Measured radiocarbon age (a BP)	Conventional radiocarbon age (a BP)
MS-C14-01	607305	Charcoal	−21.7	1800±30	1850±30
MS-C14-05	612636	Charcoal	−23.7	2800±30	2820±30
MS-C14-11	607307	Charcoal	−24.3	2000±30	2010±30

colluvial wedge (U2) and a fault scarp. Unit U3 is thicker on the hanging wall of the fault F_{1-1} than on the footwall. In event II, the fault displaced all the strata up to the surface. According to the test results of samples MS-C14-01, MS-C14-05, and MS-C14-11, collected from units U2 and U3, event I occurred prior to 2010 ± 30 BP. The test result for MS-C14-05 (2820 ± 30 BP) is older than the actual deposition age of unit U3. The sediments in sample MS-C14-05 may have originated from the erosion of older strata. Therefore, we suggest that event II may have occurred after 1850 ± 30 BP (Table 1).

The raw radiocarbon ages were calibrated by OxCal 4.3.1 (Ramsey and Lee, 2013).

5 Discussion

The Maisu fault is an E–W-trending left-lateral strike-slip fault located in the Sichuan–Yunnan block on the eastern

margin of the Tibetan Plateau. The Jinshajiang River and its main tributaries are obviously left-laterally displaced by the fault. At the Worilong site, terrace T3 of the Jinshajiang River has been left-laterally displaced by 21.2 ± 2.0 m. This indicates that the Maisu fault is primarily characterized by left-lateral strike-slip movement. An interpretation of the remote-sensing and field survey data further indicates a earthquake surface rupture zone with a length of 45 km on the fault, which extends from the town of Puma to north of the towns of Babang and Gongya in Dege County, Sichuan Province, finally disappearing near the village of Worilong (near the town of Gangtuo, Xizang Province). The magnitude of the latest strong earthquake calculated using the formulas suggested by Wells and Coppersmith (1994) is approximately Mw 7.

$$\text{Log(SRL)} = 0.74 * M - 3.55 \quad (1)$$

Here, SRL indicates the surface rupture length [km] and M indicates the moment magnitude. According to geological and chronological evidence, the latest earthquake occurred on this fault after 1850 ± 30 BP.

Previously, the Maisu fault was thought to be a late-Pleistocene-active fault (Xu et al., 2016). We estimated the Holocene activity of the Maisu fault based on the geomorphological markers and the radiocarbon dating of the displaced strata. In particular, the latest earthquake on the fault likely occurred after 1850 ± 30 BP with a magnitude of Mw 7. The discovery of Holocene activity along the Maisu fault changes our understanding of the regional kinematics and tectonic deformation. South of the Garzê–Yushu fault, which is the northern boundary fault of the Sichuan–Yunnan block, there are several E–W-trending strike-slip faults with Holocene activity, e.g., the Maisu and Batang faults (the southern boundary fault of the Batang Basin near Yushu County, Qinghai Province), which intersect the Garzê–Yushu fault. Both Huang et al. (2015) and Lü et al. (2017) have suggested that the Batang fault accommodates the partitioned horizontal slip of the Garzê–Yushu fault. The Maisu fault is similar to the Batang fault. It extends eastward and intersects the Garzê–Yushu fault. The intersection of the two faults could lead to partitioning of the horizontal strike-slip, although we have not obtained the horizontal slip rate of the Maisu fault. In addition, there is still no consensus concerning the horizontal slip rate of the southeastern segment of the Garzê–Yushu fault (Zhou et al., 1996, 1997; Wen et al., 2003; Peng et al., 2006; Chevalier et al., 2017). The Maisu fault may also play a role in absorbing the partitioned horizontal slip of the Garzê–Yushu fault.

6 Conclusion

The Maisu fault is one of the numerous E–W-trending strike-slip faults to the south of the Garzê–Yushu fault, which is the northern boundary fault of the Sichuan–Yunnan block. The Maisu fault shows a Holocene activity, and our study identified a well-preserved 45-km-long earthquake surface rupture zone along the fault. The magnitude of the latest major earthquake was estimated to be Mw 7. According to the geological and chronological evidence, this latest major earthquake on the Maisu fault may have occurred after 1850 ± 30 BP.

Data availability statement

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author.

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Author contributions

HZ: Field survey, Material and data collection, Writing (original draft); YQ: Field survey, Material and data collection; ML: Conceptualization, Field survey, Material and data collection, Writing (original draft), Writing (review and editing); KS: Field survey, Material and data collection; FH: Field survey, Material and data collection; CL: Field survey, Material and data collection; WZ: Material and data collection; WW: Material and data collection; HZ: Field survey, Conceptualization; YY: Material and data collection.

Funding

The earthquake catalog used for this study was provided by the Sichuan Earthquake Agency, Chengdu, China. This work was financially supported by the National Key Research and Development Program of China (2018YFC1504501-02, 2021YFC3000601), the Lhasa National Geophysical Observation and Research Station (NORSLS21-04), the National Survey on Seismic Disaster Risk Work (102152210240000009001), the State Key Laboratory of Earthquake Dynamics, Institute of Geology, CEA (LED 2020B02), and the Project of Science for Earthquake Resilience (XH21027).

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

The reviewer HY declared a shared affiliation with the authors HZ, ML, CL, WZ, WW, KS, FH, HZ to the handling editor at the time of review.

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OPEN ACCESS

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SPECIALTY SECTION

This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 09 July 2022

ACCEPTED 11 August 2022

PUBLISHED 12 September 2022

CITATION

Han M, Chen L, Li Y, Gao S and Feng J
(2022), Paleoeearthquakes of the
Yangda-Yaxu fault across the Nujiang
suture and Lancang river suture zone,
southeastern Tibetan Plateau.
Front. Earth Sci. 10:990187.
doi: 10.3389/feart.2022.990187

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Paleoeearthquakes of the Yangda-Yaxu fault across the Nujiang suture and Lancang river suture zone, southeastern Tibetan Plateau

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The WNW-trending Yangda-Yaxu fault (YYF) is located in the interior of the Qiangtang block (QTB). The YYF cuts through the Nujiang suture and Lancang river suture zone and divides Nujiang fault (NF) and Lancangjiang fault (LCJF) into two sections with significantly different activity levels, suggesting that the YYF may function as a specific structure in this region. In addition, a recent work argues that the YYF plays an important role in strain partitioning in southeastern Tibet and poses a high surface-faulting risk to the Sichuan-Tibet railway. However, no $M \geq 5.0$ earthquakes have been recorded, and no palaeoseismic research has been conducted along the fault, leading to limited knowledge regarding its rupture behavior, which is essential for understanding regional tectonic deformation and assessing the regional seismic potential. In this study, we constrained the timings and recurrence intervals of late Quaternary paleoseismic events along the YYF for the first time. Through trench excavations and exposure cleaning combined with radiocarbon dating, five faulting events were identified, namely, E1 through E5 from youngest to oldest (831–1,220, 3,307–6,703, 9,361–10,286, 12,729–14,651, and before 14,651 yr BP). The recurrence interval of major earthquakes along the YYF follows a quasi-periodic pattern with an interval of ~4,000 yr. Combining the clear linear geomorphic features along the fault and the paleoeearthquake results in this paper, we believe that YYF is a newly-generated active fault, and has a significant control effect on the late Quaternary evolution of the NF and the LCJF. Further analysis revealed that the YYF also plays an important role in accommodating crustal deformation.

KEYWORDS

Yangda-Yaxu fault, Nujiang suture zone, paleoeearthquake, active fault, newly-generated fault, Tibetan Plateau (TP)

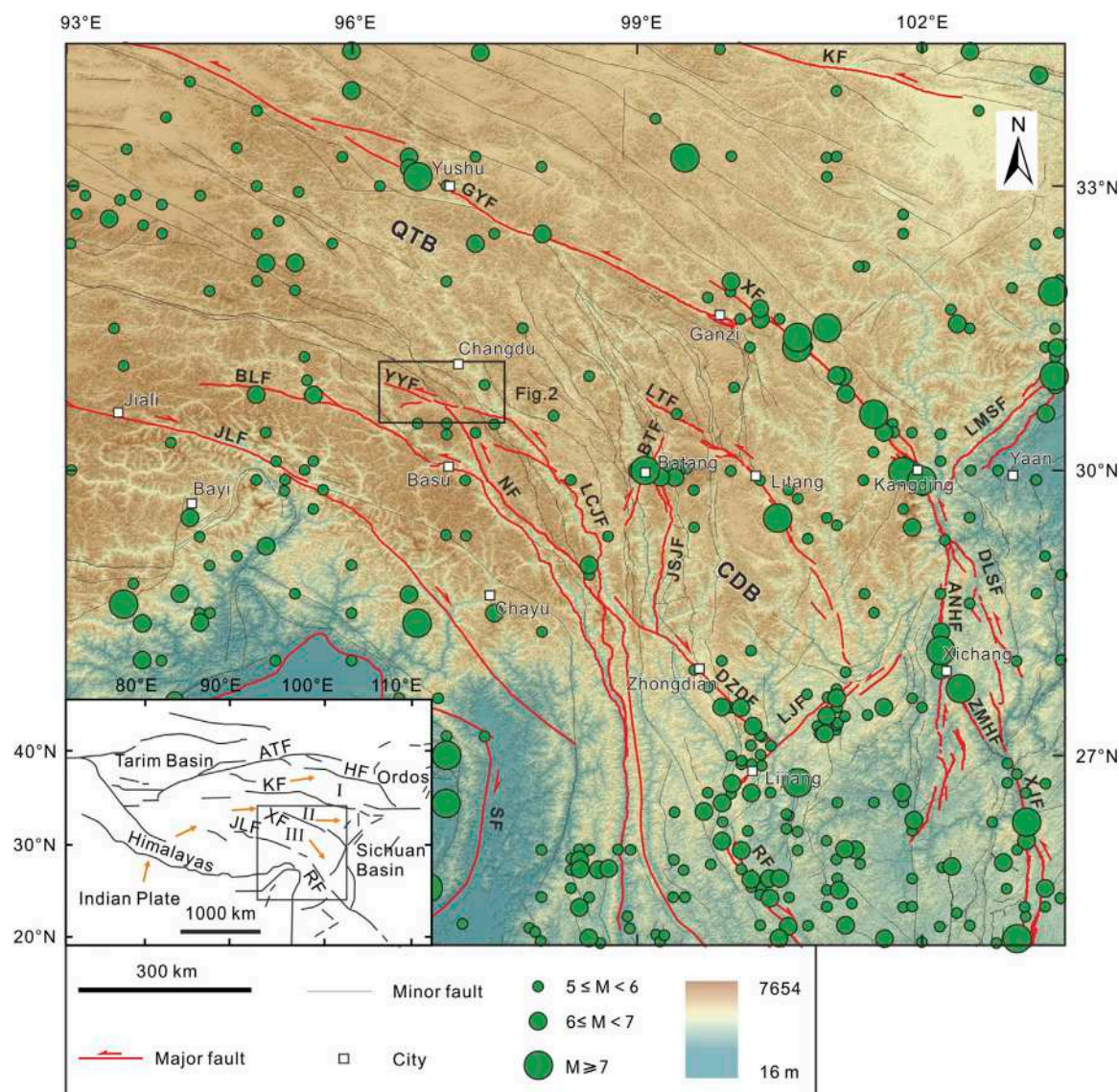


FIGURE 1

Regional tectonics and seismicity in southeastern Tibet. The red lines are major active fault traces. The green circles of different sizes indicate epicenters of earlier documented (historical and instrumental) earthquakes. The small white rectangles denote cities. The black rectangle indicates the study area of the YYF. QTB, Qiangtang block; CDB, Chuandian block. The hillshade map is generated from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) Global Digital Elevation Model (ASTGTM) (90-m resolution). Index map showing major boundary faults and blocks on the Tibetan Plateau. The orange arrows indicate the block motion direction according to GPS data (Gan et al., 2007). Blocks: I, Qaidam-Qilian block; II, Bayan Har block; III, Qiangtang-Chuandian block. Abbreviations of the active faults: ATF, Altyn Tagh fault; HF, Haiyuan fault; XF, Xianshuihe fault; JLF, Jiali fault; RF, Red River fault; BLF, Bianba-Luolong fault; YYF, Yangda-Yayu fault; NF, Nujiang fault; LTF, Litang fault; LCJF, Lancangjiang fault; JSJF, Jinshajiang fault; BTF, Batang fault; LMSF, Longmenshan fault; DLSF, Daliangshan fault; ANHF, Anninghe fault; ZMHF, Zemuhe fault; XJF, Xiaojiang fault; LJF, Lijiang-Xiaojinhe fault; DZDF, Deqin-Zhongdian-Daju fault; SF, Sagaing fault.

1 Introduction

Collision and continued convergence of the Indian and Eurasian plates have led to rapid Cenozoic uplift and intense crustal deformation (Yin and Harrison, 2000; Tapponnier et al., 2001). To accommodate uplift and crustal deformation on the

southeastern Tibetan Plateau, a large number of faults have been generated or reactivated, resulting in a very seismically active region (Figure 1; Yin and Harrison, 2000; Deng et al., 2002; Xu et al., 2003; Zhang et al., 2003; Taylor and Yin, 2009). In this region, historical documents and instruments have recorded more than 40 $M \geq 6.7$ earthquakes over the past

century (Li et al., 2019), most of which occurred along the boundary faults of tectonic blocks, such as the NW-SE-trending Xianshuihe fault (XF) and the NS-trending Xiaojiang fault (XJF) (Wang et al., 2013; Li et al., 2019; Guo et al., 2021). Because of the frequent occurrence of large earthquakes, these faults have stimulated great interest to geoscientists and have been considered important in accommodating crustal deformation of the Tibetan Plateau (Allen et al., 1991; Tapponnier et al., 2001; Wang et al., 2020; Guo et al., 2021). In contrast, few studies have considered the faults within active blocks; accordingly, the seismic risk and role in strain partitioning of these faults have generally been neglected. Over the past hundred years, a number of strong earthquakes, such as the 1933 M 7.5 Diexi earthquake (Ren et al., 2018), 1947 M 7 $\frac{3}{4}$ Dari earthquake (Liang et al., 2020a), 1948 M s 7 $\frac{1}{4}$ Litang earthquake (Xu et al., 2005a), and 2021 M w 7.3 Maduo earthquake (Ren et al., 2022a), have successively occurred along intrablock faults, indicating that the secondary faults within active blocks also exhibit the potential to produce devastating earthquakes and can play a role in strain partitioning (Zhan et al., 2021; Ren et al., 2022b; Yuan et al., 2022). As a consequence, investigation of the rupture behavior of these faults is vital not only to assess the regional seismic potential but also to better understand the deformation mechanism of tectonic blocks (Ren et al., 2013; Sun et al., 2015; Sun et al., 2017).

The Yangda-Yaxu fault (YYF), generally trending WNW, terminating at the Lancangjiang fault (LCJF) in the east, is one of the faults within the Qiangtang block (QTB). The YYF exhibits distinct linear features and faulted landforms revealed *via* image interpretation and field investigation studies. In addition, the YYF cuts through the Nujiang suture and Lancang river suture zone and divides Nujiang fault (NF) and Lancangjiang fault (LCJF) into two sections with significantly different activity levels, suggesting that the YYF may function as a specific structure in this region (Han, 2022a). However, for a long time, the fault geometry and Quaternary activity have remained poorly understood because of a lack of field data due to poor transportation conditions and high elevation, in addition to a lack of strong earthquake records. Recently, Ren et al. (2022b) first studied the geometric distribution and activity of the YYF *via* remote sensing interpretation, field investigation and chronological analysis. Further analysis revealed that the YYF has played an essential role in strain partitioning in eastern Tibet, and its presence poses a seismic potential on the Sichuan-Tibet railway, which traverses eastern Tibet and crosses the YYF (Ren et al., 2022b). However, until now, no paleoseismological investigation has been conducted, and little is known regarding the paleoearthquake history of the YYF, resulting in significant uncertainty in assessing the regional seismic potential and understanding its role in accommodating crustal deformation in eastern Tibet. Therefore, it is very crucial to investigate the faulting behavior and paleoearthquake history of the YYF.

In this paper, to constrain the rupture history of the YYF, we conducted field investigations and trench excavations based on the interpretation of high-resolution satellite images combined with radiocarbon dating. Then, we compared the latest Quaternary faulting behavior and kinematic properties of the YYF to those of neighboring faults. Finally, we described the tectonic implications of intrablock faults in seismic hazard assessment and regional strain partitioning.

2 Regional seismotectonic setting

The Qiangtang block (QTB) and Chuandian block (CDB) are being extruded south-southeastward, rotating clockwise around the Eastern Himalayan syntaxis, due to the ongoing collision of the Indian and Eurasian plates (Tapponnier et al., 2001; Xu et al., 2003; Zhang et al., 2004; Gan et al., 2007; Wang and Shen, 2020). Two groups of active faults with different strikes of nearly NEN and NW-WNW are located in the eastern QTB and northwest CDB (Figure 1).

The nearly NEN-striking fault mainly comprises the Batang fault (BTF). Six earthquakes of $M \geq 6.0$ have occurred along the BTF in recorded history (Department of Earthquake Disaster Prevention and State Seismological Bureau., 1995). Among these recorded earthquakes, the 1870 M 7 $\frac{1}{4}$ Batang earthquake, which was the largest earthquake, produced a right-lateral surface rupture zone (Xu et al., 2005b; Zhou et al., 2005), with a maximum right-lateral offset of 1.5 m, and a maximum vertical offset of 0.6 ± 0.2 m (Xu et al., 2005b). The late Quaternary strike-slip rate of the BTF has been estimated *via* geological methods as 1.3–2.7 mm/yr (Zhou et al., 2005). Recent trench work revealed that several surface-rupturing events have occurred during the late Quaternary, among which the latest event corresponds to the 1870 M 7 $\frac{1}{4}$ Batang earthquake (Gao, 2021). The activity of the BTF was weak during the early Holocene, increasing significantly since the late Holocene, and the recurrence interval was shortened to approximately 830 yr, as revealed *via* further analysis.

The nearly NW-WNW-striking faults mostly include the XF, Jiali fault (JLF), Deqin-Zhongdian-Daju fault (DZDF), Litang fault (LTF), Bianba-Luolong fault (BLF) and YYF (Figure 1). The left-lateral strike-slip XF, located along the northern boundary of the CDB, is seismically highly active, where more than ten large earthquakes ($M_s \geq 6.9$) occurred in recorded history (Allen et al., 1991; Wen et al., 2008). Based on multiple measurements, the strike-slip rate of the fault has been estimated to be ~ 2.5 – 5.6 mm/yr (Li et al., 2017; Liang, 2019; Liang et al., 2020b; Gao, 2021). Previous trench excavations have revealed that several paleoearthquakes have occurred since the Holocene, specifically, five events in the Qianning segment over the past $\sim 9,000$ yr (Li et al., 2017), six events in the Luhuo segment over the past $\sim 3,000$ yr (Liang et al., 2020b). The seismic sequences of the Luhuo and Qianning segments are consistent with clustering

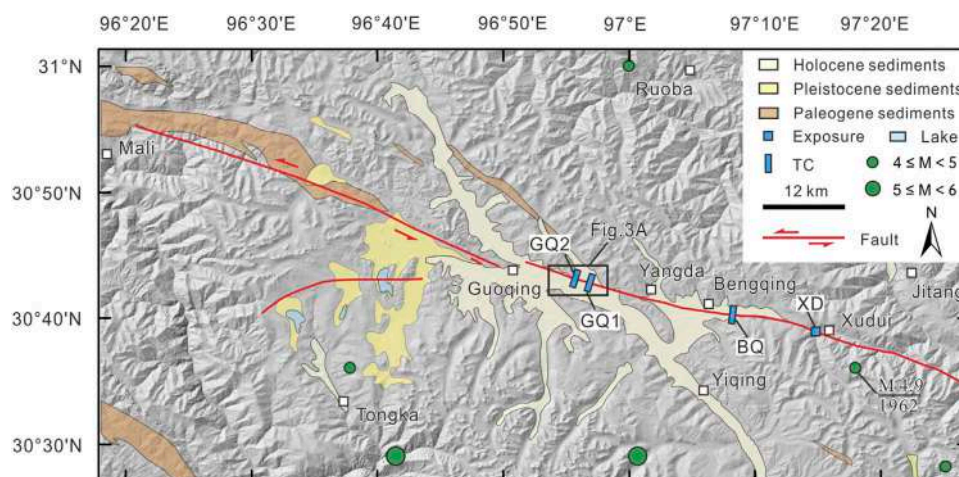


FIGURE 2

Geologic features and fault distribution of the YYF. Please refer to location in Figure 1. Geological data are adapted from the 1:250000 geologic map. The hillshade map is generated from the ASTGTM (30-m resolution). The black box indicates the extent of Figure 3A.

and abnormal accelerating stress release behavior patterns (Li et al., 2017; Liang et al., 2020b). The recent research data based on faulted landforms, paleoseismic trenches and GPS data reveal that the JLF is generally a Holocene active fault with dextral strike-slip and thrust component (Wang et al., 2020; Zhang et al., 2021). The fault has experienced at least two strong earthquake events since the Holocene (Wang et al., 2020), the latest of which occurred between 2,680–2,160 yr BP. The dextral strike-slip rate is about 2.4–3.8 mm/yr (Zhang et al., 2021). DZDF is characterized by dextral strike-slip with normal. The horizontal slip rate of the fault since Holocene is 1.7–2.0 mm/yr, and the vertical slip rate is 0.6–0.7 mm/yr (Chang et al., 2014).

The LTF, which is mostly parallel to the XF, is also quite active, with three historical earthquakes of $M > 7$ since 1700 (Zhou et al., 2015; Chevalier et al., 2016; Wang et al., 2021). Among these earthquakes, the 1890 $M_w \sim 7.3$ earthquake produced a surface rupture zone with a length of ~ 71 km (Wang et al., 2021), and the latest large earthquake occurred in 1948 ($M_s 7\frac{1}{4}$) with an ~ 40 -km surface rupture (Xu et al., 2005a). The LTF is a dominantly left-lateral strike-slip fault with a late Quaternary strike-slip rate of ~ 2 –4 mm/yr, as evidenced via geological and dating investigations (Xu et al., 2005a; Zhou et al., 2005, 2007; Gao, 2021; Wang et al., 2021). Recent trench excavations have revealed that four paleoseismic events have occurred since the Holocene, and the recurrence intervals were $\sim 1,415$ yr, $\sim 1,104$ yr, and ~ 775 yr (Wang et al., 2021). Historical records indicate that two large earthquakes, namely, the 1,642–1,654 $M \geq 7$ Luolong earthquake and the 1,791–1,804 $M 6\frac{3}{4}$ Bianba earthquake (Science and Technology Committee of the Tibetan Autonomous Region, 1982; Department of Earthquake Disaster Prevention and State Seismological Bureau, 1995), have occurred along the

BLF. A recent study based on the field survey and ^{14}C dating methods has suggested that the BLF is a Holocene active fault with the latest faulting event possibly corresponding to the 1,642–1,654 $M \geq 7$ Luolong earthquake, and the main movement pattern of the fault is sinistral strike-slip movement (Han et al., 2022b). The YYF is mainly a left-lateral strike-slip fault with a horizontal rate of ~ 1.7 mm/yr during the late Quaternary, as evidenced from displaced landforms integrated with corresponding dating results (Ren et al., 2022b). The clear surface ruptures, notable surface cracks and sand liquefaction phenomenon suggest that the last earthquake along the YYF did not occur long ago (Ren et al., 2022b). However, the YYF experienced very low seismic activity. Historic records indicate no $M \geq 5.0$ earthquakes along the YYF. The largest earthquake was the 1971 $M 4.9$ earthquake (Figure 2). Furthermore, small–moderate earthquakes with a magnitude less than 5.0 rarely occur along the fault, although the possibility that this phenomenon is also related to the sparse seismic stations in the adjacent areas cannot be ruled out. To date, no paleoearthquake sequence along the YYF since the late Quaternary has been reported.

3 Materials and methods

3.1 Field investigation and trench analysis

The surface traces of the YYF (Figure 2) were mapped in detail based on high-resolution Google Earth image interpretation and field investigation of faulted landforms, including fault scarps, fault troughs, sag ponds, alluvial fans, gullies, streams, and surface ruptures. To reveal fault

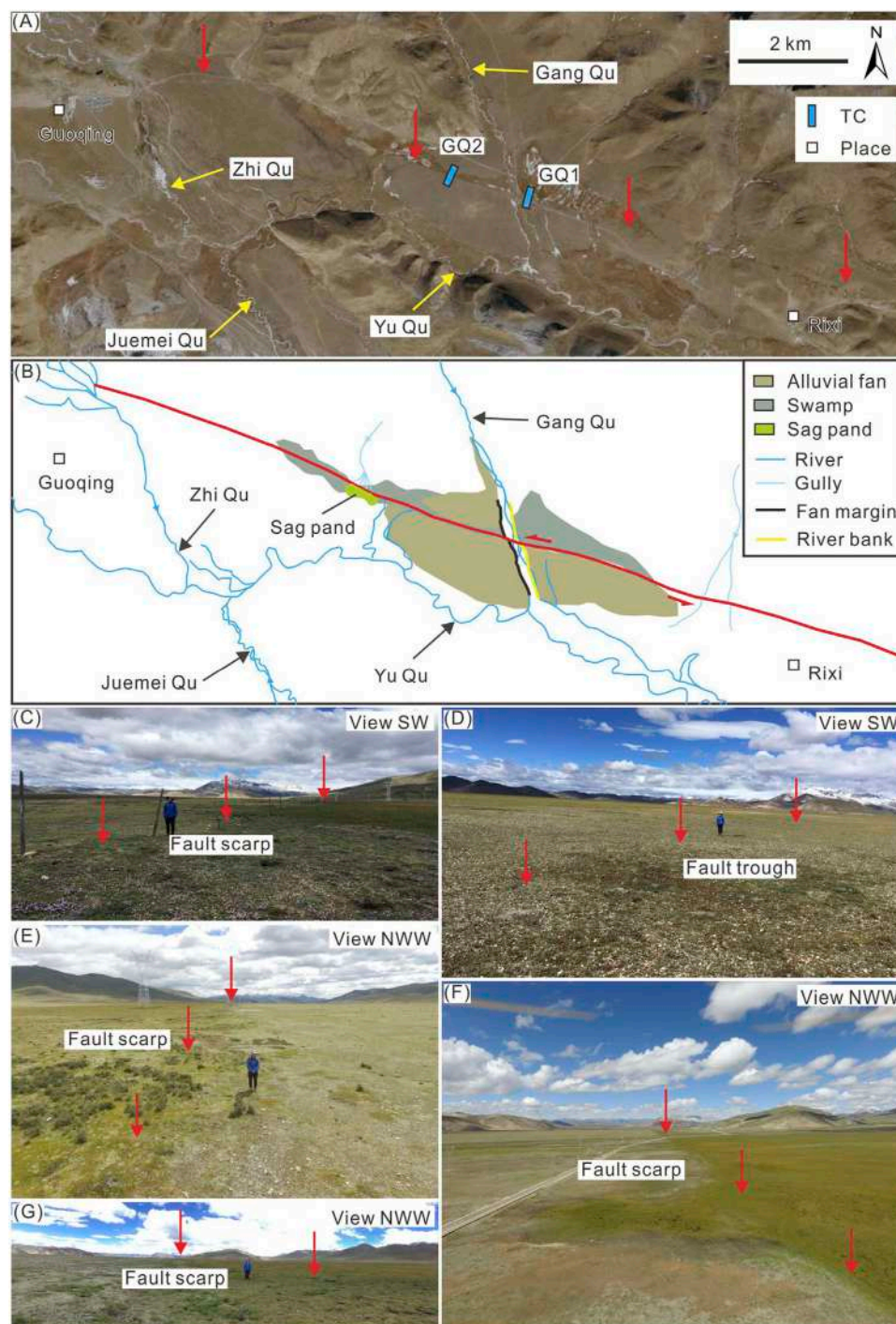


FIGURE 3

Fault traces and faulted landforms near the Guoqing township and the locations of trenches GQ1 and GQ2. (A) Satellite image and (B) interpreted map of the offset landforms near the Guoqing township. The location is shown in Figure 2. The red arrows and lines point to the surface traces of the fault. (C–G) Field photographs of the fault traces and representative faulted landforms.

structures and paleoseismic evidence, three trenches were excavated nearly perpendicular to the fault traces (Figure 3). Trench excavation sites were selected based on

locations where sediments were relatively continuous and stable, such as sag ponds or small depressions across distinct fault scarps. Unmanned aerial vehicle-based

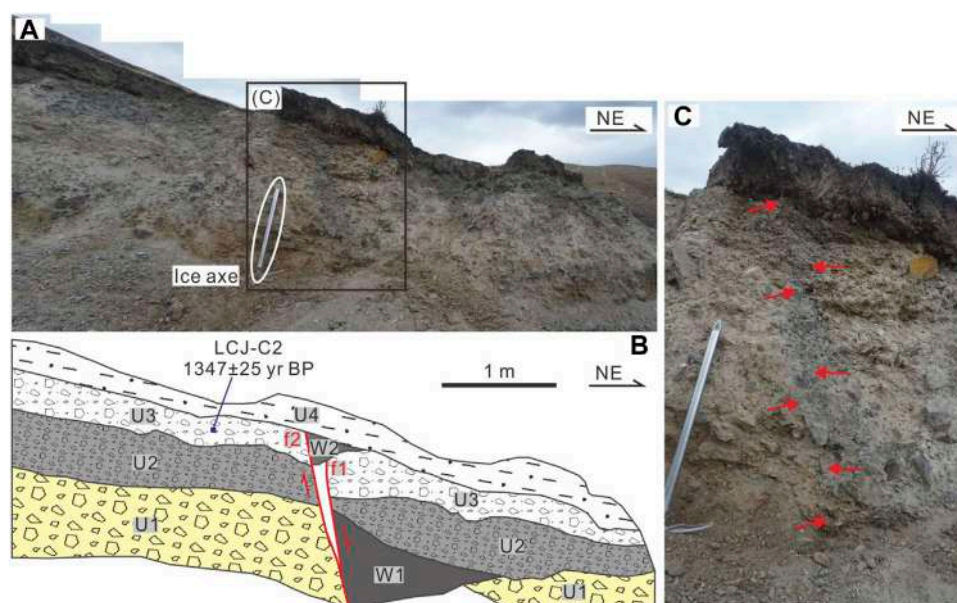


FIGURE 4

(A) Photograph and (B) interpreted sketch of the XD fault exposure. The location is shown in Figure 14C. (C) Enlarged photograph of the local details of the fault exposure. The U and red lines indicate the stratum units and the fault plane, respectively.

photogrammetry was employed to finely measure offset landforms at the trench sites and produce high-resolution digital elevation models (Bemis et al., 2014; Gao et al., 2017). The walls of each trench were systematically cleaned to expose clear evidence of paleoseismic events. After trench walls cleaning, 1 m by 1 m grids were constructed to capture photos. Field logging was conducted involving square gridded paper, and the field logs were then digitalized on a computer. In addition, along the YYF, a fault exposure, which was suitably exposed near Xudui village due to human activity, was cleaned and analyzed (Figures 3, 4).

3.2 Radiocarbon dating

Numerous radiocarbon samples were collected from the faulted and unfaulted strata in the three trenches and one outcrop. The ^{14}C samples were processed and analyzed by Beta Analytic Inc. Detailed accelerator mass spectrometry (AMS) ^{14}C sample preparation was provided by Beta Analytic Inc.; please refer to their website (<https://www.radiocarbon.cn/beta-AMS-lab/>). The ages of the samples were corrected as two-sigma calendar ages (95.4% confidence interval) with the OxCal 4.4 procedure (Ramsey, 2009; Reimer et al., 2013). The timings of the determined paleoseismic events based on the considered trenches and outcrop were also limited using the progressive confining method considering the sample age.

4 Paleoseismic investigation

4.1 Guoqing site

4.1.1 Site location and offset landforms

In the eastern part of the Guoqing township, the YYF cuts through piedmont alluvial fans, forming a well-extended linear scarp and a gentle trough landform on the surface. Moreover, a small sag pond is locally developed (Figures 3A,B). The trend of the linear scarp ranges from ~ 280 to 290° , and the height of the scarp ranges from ~ 0.6 to 1.7 m (Figures 5D–G, 6B). There are obvious differences in turf growth conditions between both sides of the scarp, and linear swamps are developed in low-lying areas near the scarp (Figure 3). In addition, due to fault motion, the sinistral offset occurs at the fan margin of the alluvial fan, Gang Qu (Qu means river in Chinese) river banks and several small gullies (Figure 3B), with a displacement ranging from ~ 20 to 40 m (Ren et al., 2022b). To reveal the complete sequence of paleoseismic events as much as possible, this paper excavated two trenches across the local low-lying area of the linear scarp (Figures 5A–B, 6A,C), denoted as trenches GQ1 and GQ2, from east to west, which were approximately 2 km away from each other (Figures 2, 3A).

4.1.2 Stratigraphy

Trench GQ1 is approximately 18 m long, 3 m wide, and 3 m deep (Figure 5C). Trench GQ2 is approximately 19 m long, 3 m wide, and 3 m deep (Figure 6D). Both the western and eastern walls of GQ1 and GQ2 exhibit a similar stratigraphy and

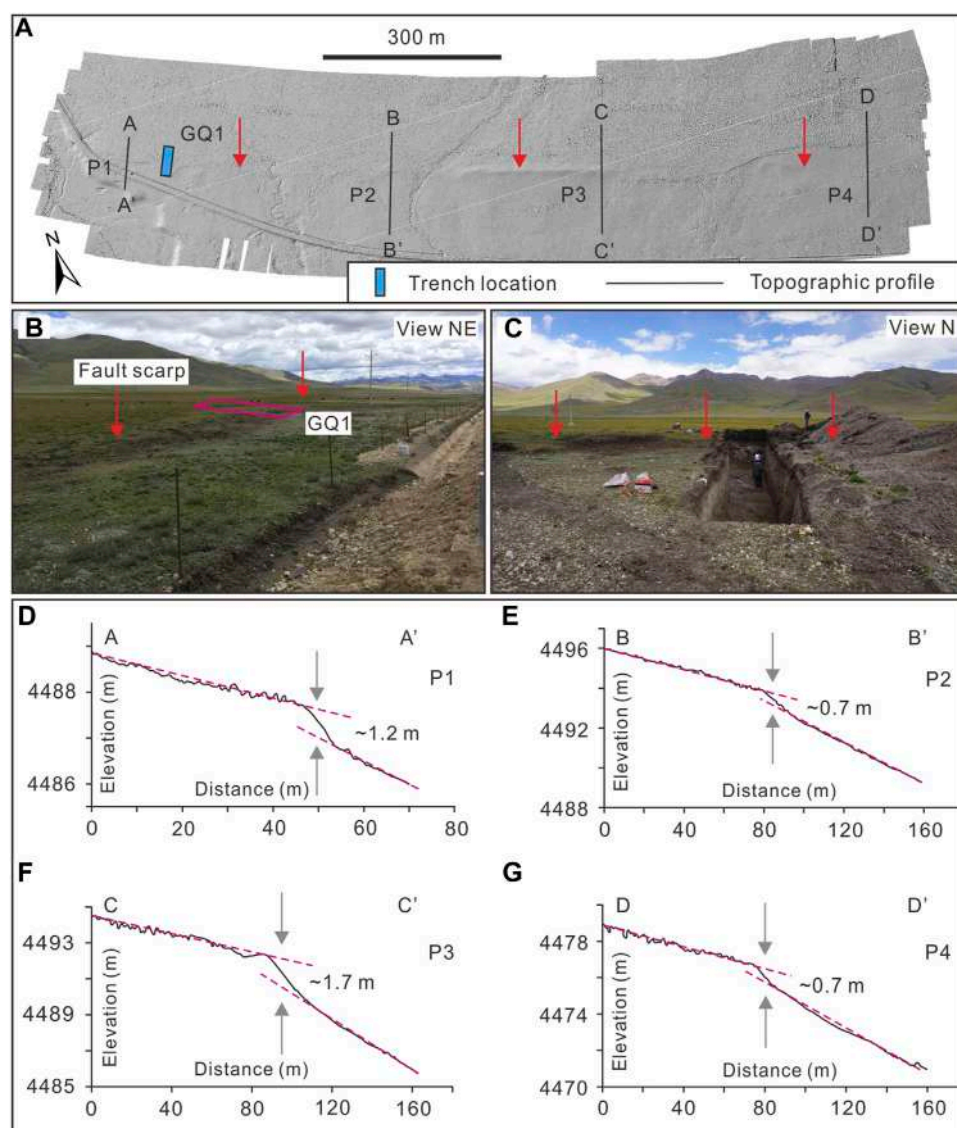


FIGURE 5

Faulted landforms around the trench GQ1 excavation site. (A) The hillshade map is generated from an unmanned aerial vehicle-based digital elevation model. The location is shown in Figure 3A. The red arrows point to the surface trace of the fault. (B,C) Field photos of the fault scarp and location of trench GQ1. (D–G) P1–P4 topographic profiles across the fault scarp.

deformation features. The strata mainly comprise alluvial gravel layers at the bottom and clay or sandy clay layers at the top, typical of fluvial facies. Five units of layers, including nine subunits, named in ascending order from bottom to top, could be recognized in the trenches sections (Figures 7, 8, 9). The description of these strata are provided in Tables 1, 2.

4.1.3 Evidence for events in trench GQ1

Based on the gravel alignment and deformation and lithological differences among the various strata, eight faults,

namely, f1–f8, and seven faults, namely, f1–f7, from the northeast to the southwest, could be identified in the eastern trench wall (Figure 7) and western trench wall (Figure 8), respectively. Specifically, on the eastern wall, these fault planes steeply dipped and formed a typical flower structure, which is common in strike-slip fault zones. On the western wall, the fault planes, as a whole, were subparallel. Based on analysis of the deformed strata and corresponding cut-and-cover structures and paleoseismic markers (Liu-Zeng et al., 2007; McCalpin, 2009; Ran et al., 2012; Wang et al., 2013; Wang et al., 2020; Guo et al.,

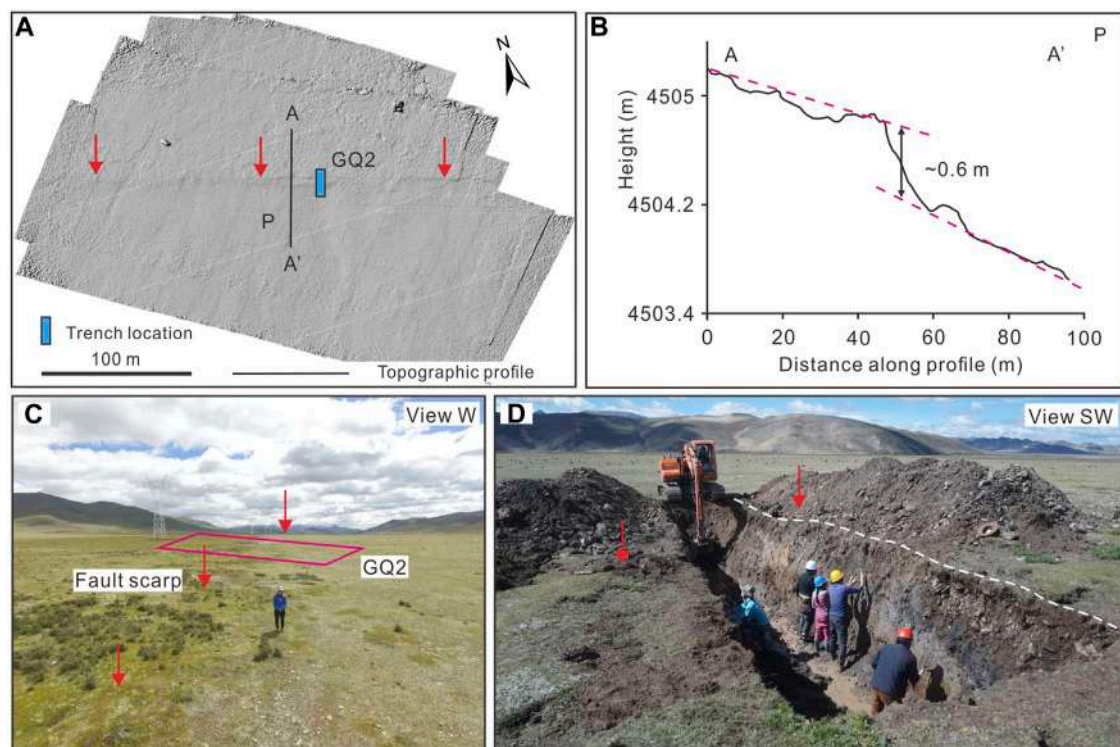


FIGURE 6

Faulted landforms around the trench GQ2 excavation site. (A) The hillshade map is generated from an unmanned aerial vehicle-based digital elevation model. The location is shown in Figure 3A. The red arrows point to the surface trace of the fault. (B) P topographic profile across the fault scarp. (C,D) Field photos of the fault scarp and location of trench GQ2.

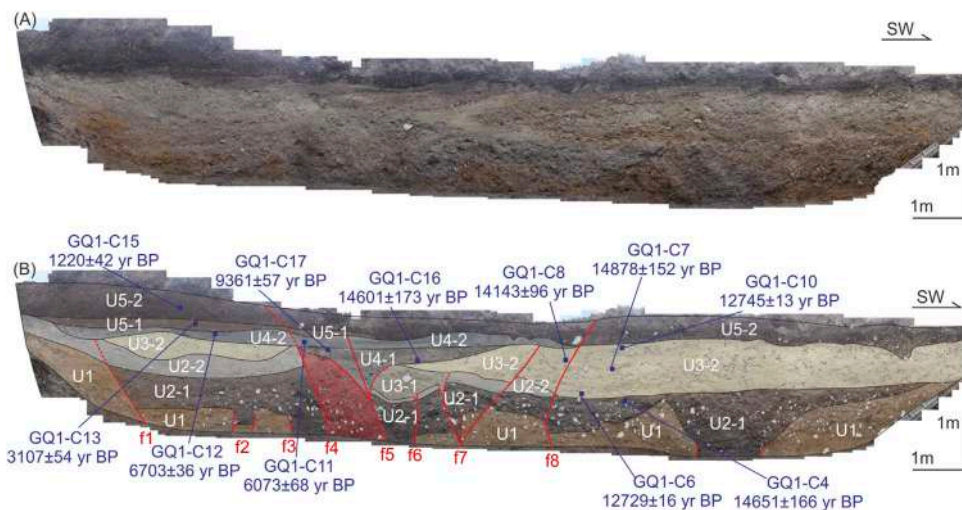


FIGURE 7

Photomosaic (A) and interpretation map (B) of the eastern wall of trench GQ1.

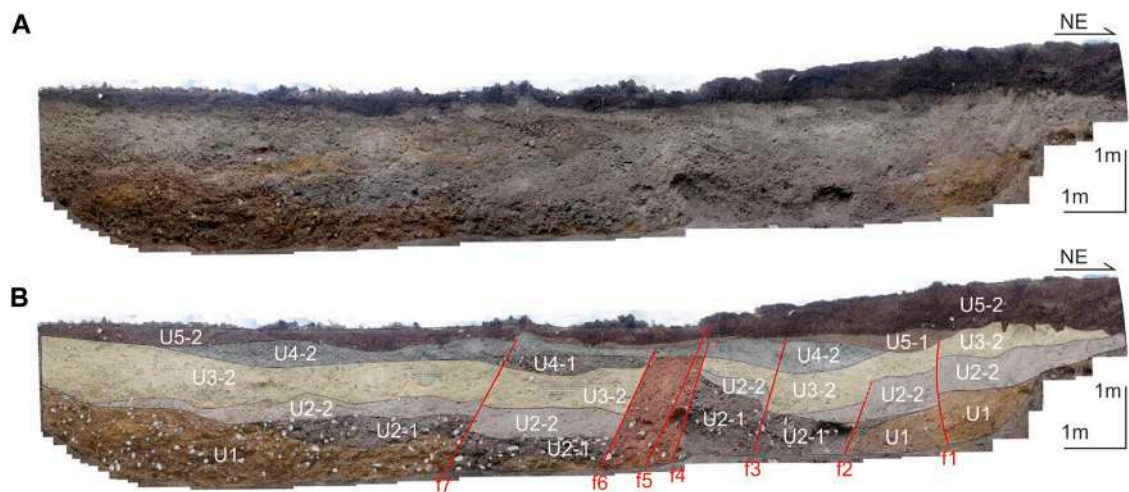


FIGURE 8
Photomosaic (A) and interpretation map (B) of the western wall of trench GQ1.

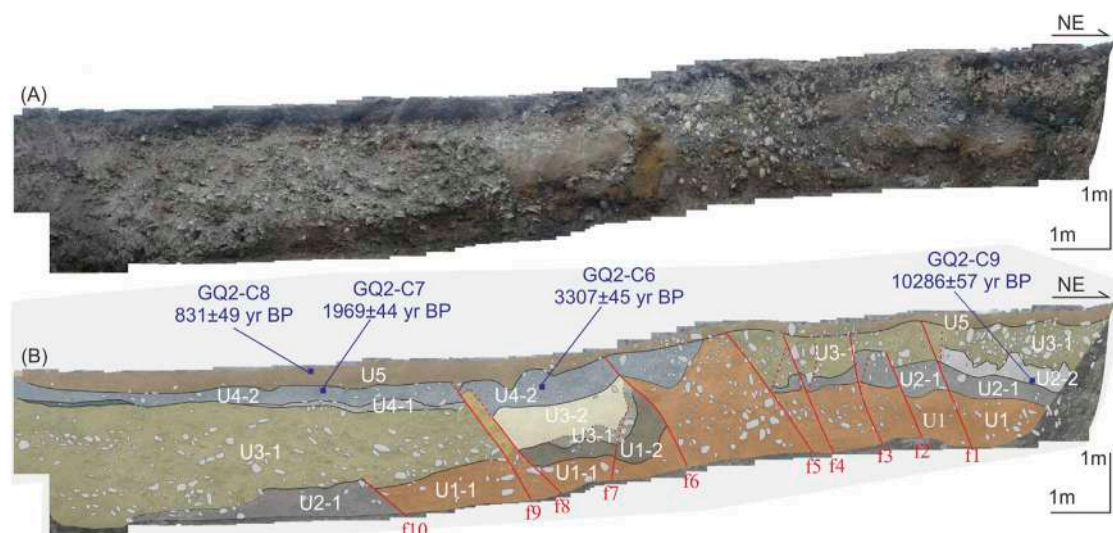


FIGURE 9
Photomosaic (A) and interpretation map (B) of the western wall of trench GQ2.

2021), five paleoearthquakes were identified, denoted as G1E1, G1E2, G1E3, G1E4, and G1E5, from youngest to oldest.

4.1.3.1 Event G1E5

This event was mainly revealed on the eastern wall (Figure 7). Faults f3, f2, middle branch of f7, and northeast branch of f8 faulted layer U1, which were overlain by layer U2-1, resulting in a faulted contact between layers U2-1 and U1. In conclusion, the event EGO5 occurred after U1 deposition and before U2-1 deposition.

4.1.3.2 Event G1E4

Western wall fault f2 faulted layer U2-2, overlain by layer U3-2, resulting in layer U3-2 occurring in unconformable contact with U2-2 (Figure 8). Eastern wall faults f6 and northeast branch of f7 faulted layer U2-2 and were overlain by U3-1 (Figure 7). Faulting and traction of f6 and f7-1 led to strong tilting deformation of U2-2 than that of the strata on both sides. Thus, the above evidence indicates that event EGO4 occurred between the deposition of layers U2-2 and U3-1.

TABLE 1 Unit description of trench GQ1.

Unit	Description
U1	Yellow-brown gravel layer containing a small amount of clay, with a certain sedimentary bedding. The gravel particles are angular
U2-1	Light gray-green gravel layer, mixed deposits. The gravel particles are angular
U2-2	Sandy clay layer with a high sand content, the color is mainly gray
U3-1	Gravel interbedded with clay layers. Versicolor mixed deposits, mainly gray-green. The gravel particles are well sorted. Gravel particles with a size of 3–5 cm are dominant, and the largest particles are up to 10 cm in size. This layer may be associated with multiple material sources
U3-2	Gray-green fine sandy clay layer containing a small amount of fine gravel particles. The gravel particles are angular
U4-1	Gravel interbedded with clay, the color is mainly brownish red
U4-2	Silty clay layer with local bedding, the color is mainly gray-black
U5-1	Thin layer of small gravel particles. This stratum is developed locally near the fault, with the characteristics of frontal fault scarp deposition
U5-2	Gray-black modern root soil layer. There is a small amount of small-sized gravel particles, and a few large gravel particles up to 13 cm in size

TABLE 2 Unit description of trench GQ2.

Unit	Description
U1-1	Gravel layer, containing a small amount of sand. The color is mainly gray-green to brown-green, with a slight light-yellow hue. The gravel particles are poorly rounded and exhibit obvious directional arrangement along the fault planes
U1-2	Gray-black to gray-green clay interbedded with a gravel layer. This stratum exhibits notable disturbance. The gravel particles are poorly sorted and poorly rounded
U2-1	Brown-green sand-gravel layer. The gravel particles are angular or subangular
U2-2	Sand layer with occasional gravel, and the bottom contains black strips with strong deformation.
U3-1	Gravel layer, the color is mainly gray-white. The gravel particles are poorly sorted and poorly rounded
U3-2	Thin layer of sand interbedded with clay, developed locally, containing a small amount of gravel particles
U4-1	Gray-green gravel layer, developed locally
U4-2	Black silty clay layer, containing a small amount of small gravel particles. The thickness is variable with a limited distribution.
U5	Gray-black topsoil layer rich in plant roots, containing a small amount of small gravel particles. Some large gravel particles are only visible near the fault at the northern end of the eastern wall.

4.1.3.3 Event G1E3

This event was mainly revealed on the eastern wall (Figure 7). Southwest branch of f7 faulted layer U3-2 and was overlain by U4-2. In addition, southwest branch of f5 faulted layer U3-1 and was overlain by U4-1. Therefore, event EGO3 occurred after the deposition of U3-2 and before the deposition of U4-1.

4.1.3.4 Event G1E2

On the eastern wall, the middle branch of fault f5 faulted layer U4-2, overlain by U5-1 (Figure 7). On the western wall, faults f3, f4, and f6 faulted layer U4-2 and was overlain by U5-1. Fault f7 faulted U4-2 and was overlain by U5-2 (Figure 8). Therefore, event EGO2 occurred between the deposition of U4-2 and U5-1.

4.1.3.5 Event G1E1

Western wall fault f5 and eastern wall faults f8 and f4 faulted layer U5-2 and extended to the ground surface (Figures 7, 8). A

well-preserved fault scarp at this location was interpreted to have been formed by this event. Based on the above, the EGO1 occurred after U5-2 deposition, which may correspond to a historical earthquake.

4.1.4 Evidence for events in trench GQ2

Similarly, ten fault branches were identified, denoted as f1–f10 from northeast to southwest (Figure 9). Based on analysis of the paleoseismic markers, three paleoseismic events were identified, namely, G2E1, G2E2, and G2E3, from youngest to oldest.

4.1.4.1 Event G2E3

Faults f2 and f10 faulted layer U2-1, overlain by layer U3-1. Along the above fault planes, there occurred multiple aligned small gravel particles. Therefore, these observations show that event EGT3 occurred before deposition of U3-1 and after deposition of U2-1.

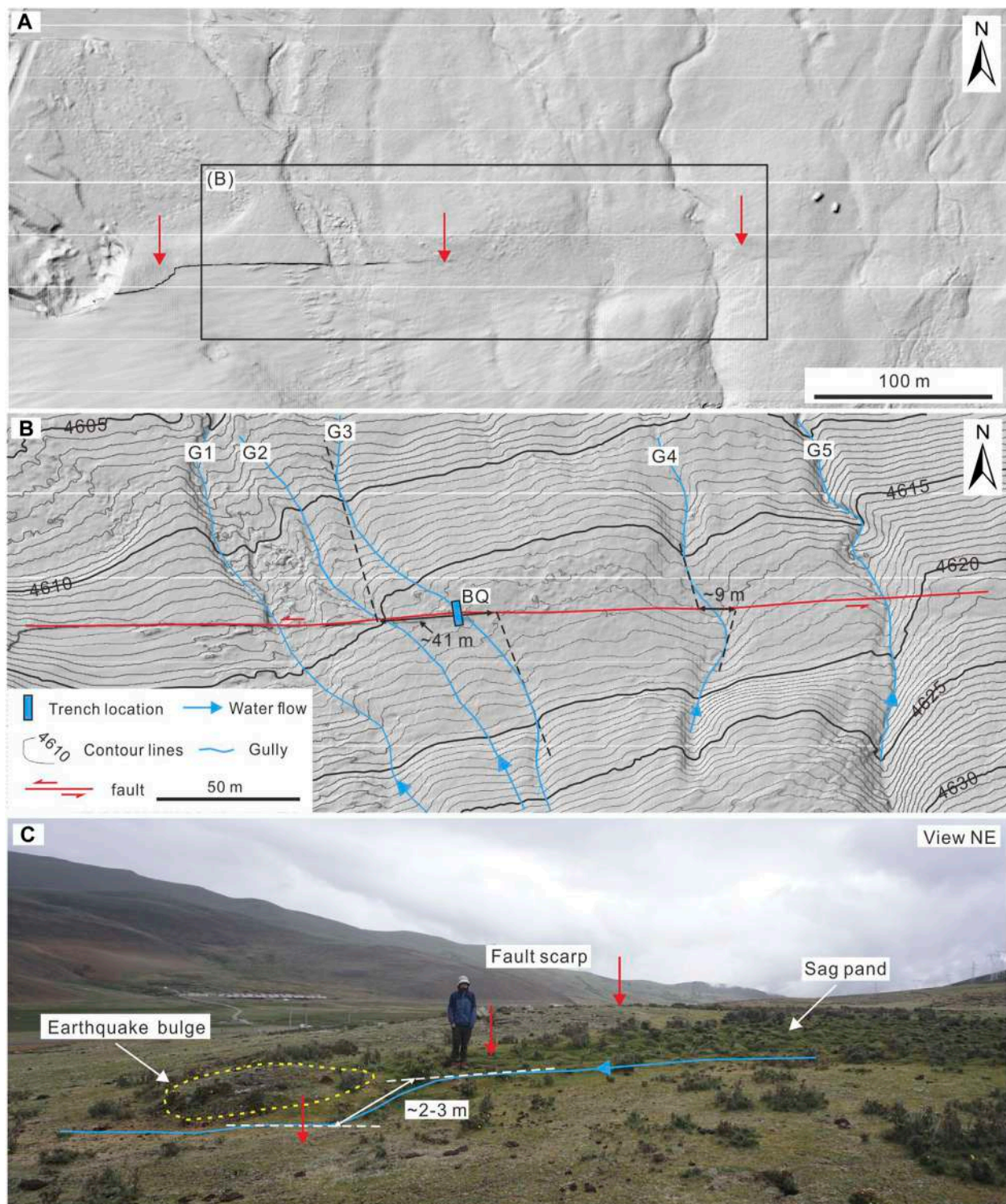


FIGURE 10

Faulted landforms around the trench BQ excavation site. **(A)** The hillshade map is generated from an unmanned aerial vehicle-based digital elevation model. The location is shown in Figure 2. The red arrows point to the surface trace of the fault. **(B)** Interpretation map of the offset landforms in (A). The contour interval is 0.5 m. **(C)** Field photo of the faulted landforms around trench BQ.

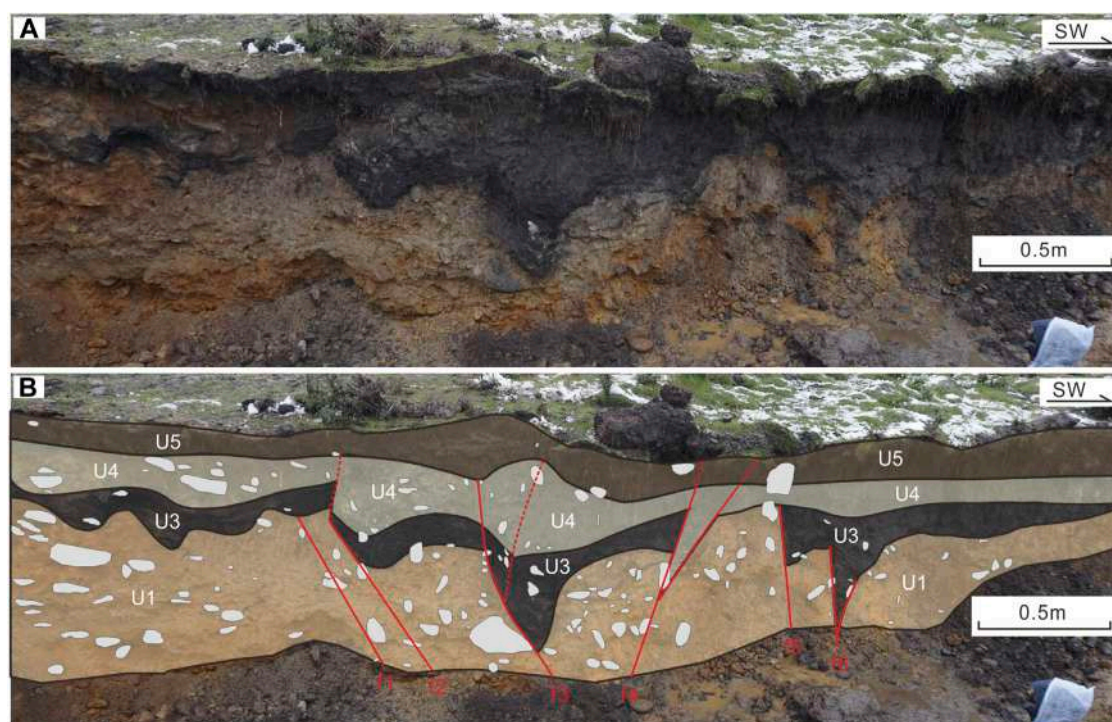


FIGURE 11
Photomosaic (A) and interpretation map (B) of the eastern wall of trench BQ.

4.1.4.2 Event G2E2

Fault f8 faulted layer U3-2 and was overlain by U4-2. The evidence suggests that event EGT2 occurred before deposition of U4-1 and after deposition of U3-2.

4.1.4.3 Event G2E1

Faults f6 and f9 faulted layer U4-2 and was overlain by U5. No obvious deformation was found in the overlying layer U5. The above evidence indicates that event EGT1 occurred between the deposition of layers U4-2 and U5.

4.2 Bengqing site

4.2.1 Site location and offset landforms

In the southeastern area of Bengqing village, ~21 km away from the trench GQ1 excavation location, the YYF cuts through the piedmont alluvial-proluvial fan (Figure 10A), forming striking faulted landforms, including fault scarps, earthquake bulges, and sag ponds (Figure 10C). Five large gullies (G1–G5) across the fault were synchronously left-laterally offset. The left-lateral displacements of gullies G3 and G4 were ~41 and ~9 m, respectively (Figure 10B). In addition, adjacent to the southern side of the fault scarp and the earthquake bulge, a small gully was dislocated by ~2–3 m, which can represent the horizontal

coseismic displacement of the latest earthquake event (Figure 10C).

4.2.2 Stratigraphy

Trench BQ is approximately 13.5 m long, 3 m wide, and 3 m deep. Because of the shallow depth of the ground water table and quick infiltration, in addition to weak properties of the sediments, the northern part of the eastern trench wall collapsed soon after the excavation, and only the southern part was retained. The strata on both walls are similar and hence correlated (Figures 11–13). The strata mainly comprise gravel particles at the bottom and pebbly clays or gravel interbedded clays at the top. Five sets of layers, named in ascending order from bottom to top, could be recognized in the trench sections, among which layer U2 was missing on the eastern wall (Figure 11). A description of the strata is provided in Table 3.

4.2.3 Evidence for events

According to the aligned gravel clasts within the gravel layer and the deformation and lithological differences among the strata, six faults, namely, f1–f6, and three faults, namely, f1–f3, from the northeast to the southwest, could be identified in the eastern trench wall (Figure 11) and western trench wall (Figure 12), respectively. Based on analysis of the deformed strata and paleoseismic markers, four paleoseismic events were identified, namely, BE1, BE2, and BE3, from youngest to the oldest.

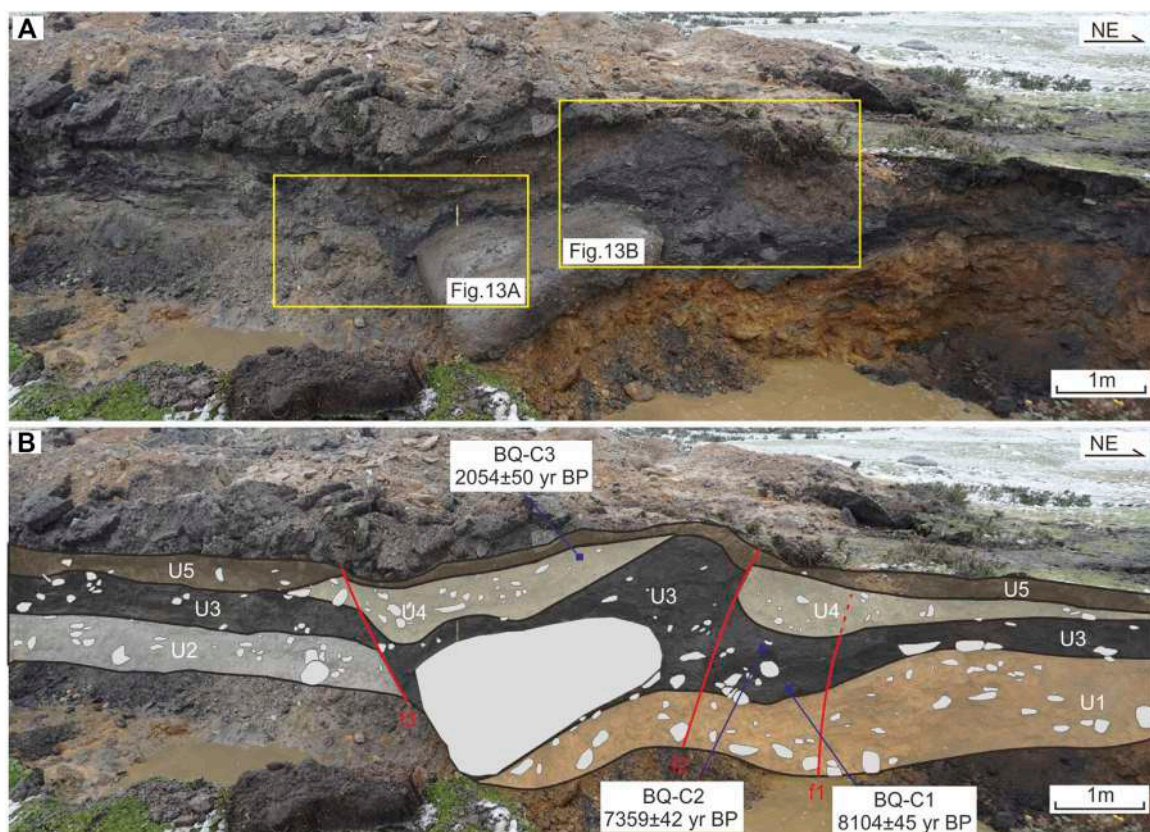


FIGURE 12
Photomosaic (A) and interpretation map (B) of the western wall of trench BQ.

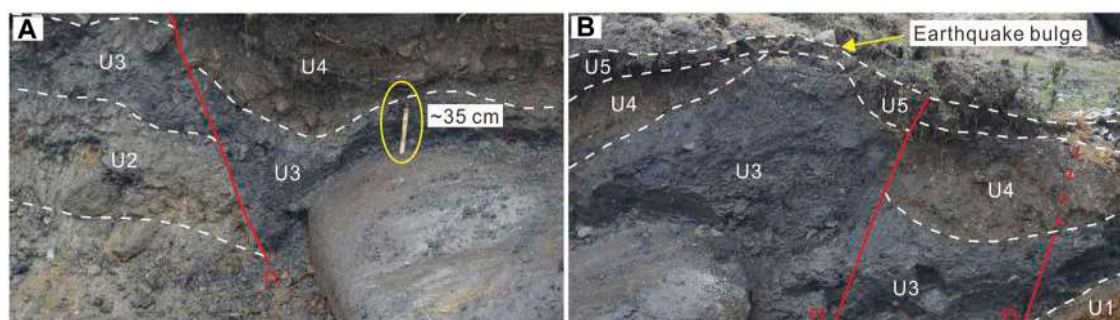


FIGURE 13
Local enlarged photographs of the western wall of trench BQ showing the faulted strata along the faults; the locations are shown in [Figure 12A](#).

4.2.3.1 Event BE3

This event was revealed only on the eastern wall. Faults f1 and f6 faulted layer U1 and was overlain by U3. Along faults f3 and f6, there occurred multiple obvious tensional cracks in the upper part of U1, which were penetrated by U3. In addition,

along fault f3, layer U1 exhibited a larger displacement amount and greater top surface disturbance than those exhibited by layer U3 ([Figure 11](#)). Therefore, the above evidence suggests that event BE3 occurred after U1 deposition and before U3 deposition.

TABLE 3 Unit description of trench BQ.

Unit	Description
U1	Yellow-brown to yellow-green alluvial-proluvial facies sand-gravel layer mixed with silty sandy clay. The gravel particles are angular and poorly sorted. The gravel particles are oriented along the fault plane near the fault and aligned nearly horizontal far from the fault
U2	Light yellow-brown to yellow-gray gravel-bearing silty clay layer. The gravel particles are angular or subangular and poorly sorted. Only developed locally in the southern part of the western wall
U3	Black clay layer interbedded with gravel. This stratum is strongly disturbed near the fault and relatively stable away from the fault zone. The gravel particles are poorly rounded and aligned near the fault. There are obvious differences in the gravel content and gravel size between the two walls. The eastern wall exhibits a lower content and smaller gravel size, while the western wall exhibits a higher content and larger gravel diameter, and a very large gravel particle occur in the fault zone of the western wall with a gravel diameter of more than 2 m
U4	Gray-black to gray-yellow gravel interbedded with a clay layer. This layer is thick near the fault and gradually thins away from the fault, forming a wedge-shaped deposition with a limited distribution. Gravel is dominant within the fault zone, while clay is dominant far from the fault. The gravel layer comprises mixed deposits and poorly sorted and poorly rounded. This stratum is likely a set of colluvial wedge deposit following an earthquake. The deposition of U4 is likely a colluvial deposit generated by a scarp after an earthquake.
U5	Surface turf layer rich in grass roots, mainly comprising black clay with a small amount of gravel particles

4.2.3.2 Event BE2

On the eastern wall, along fault f2, layer U3 exhibited a larger displacement amount and greater top surface disturbance than those exhibited by layer U4 (Figure 11). On the western wall, layer U4 was locally developed and exhibited a wedge shape, thickening near the fault, and gradually thinning to near-horizontal away from the fault. In layer U4, the gravel particles were mixed, and no horizontal bedding was developed (Figures 12, 13). We could infer that U4 may be a colluvial facies gravel layer formed by event BE2. Therefore, the above evidence indicates that event BE2 occurred after U3 deposition and before U4 deposition.

4.2.3.3 Event BE1

Faults f2 and f3 on the western wall faulted turf layer U5 and ruptured upward to the ground surface (Figure 12). Additionally, fault f4 on the eastern wall appeared to offset layer U5 (Figure 11). The depositional and geomorphological features indicate that event BE1 occurred after U5 deposition and might correspond to a historical earthquake. The well-preserved earthquake bulge was interpreted to have been caused by event BE1 (Figure 13).

4.3 Xudui site

4.3.1 Site location and offset landforms

Near Xudui village, the YYF cuts through the hillside. Satellite images and field investigation revealed that the fault exhibits a linear fault scarp, fault trough and bedrock scarp (Figure 14). Five gullies (G1–G5) were left-laterally offset when crossing the fault (Figure 14B). Close to the west bank of the G4 gully, we cleaned a fault outcrop

(Figures 4, 14C), which was widely exposed due to human activity, on the alluvial platform. Here, the exposure was recorded as XD.

4.3.2 Stratigraphy

Exposure XD is approximately 5.5 m long, and 1.5–2 m deep (Figure 4). The strata primarily comprised a sandy gravel layer, typical of the alluvial pluvial phase. Based on the sedimentary characteristics of the strata, four sets of layers, named in ascending order from bottom to top, could be recognized in the cleaned exposure. A description of the strata is provided in Table 4.

4.3.3 Evidence for events and radiocarbon dating of events

Two faults (f1 and f2) with steep dips and normal fault component were identified in the cleaned exposure. Fault f1 merged into fault f2 at the bottom of the exposure, forming a fault zone with a width of ~10 cm. Faulting led to clear gravel orientation and stratigraphic offset, and colluvial wedges were developed near the fault (Figure 4). Based on the cut-and-fill relationship and sedimentary structures, two paleoseismic events were recognized, namely, XE1 and XE2, from youngest to oldest.

4.3.3.1 Event XE2

The older, penultimate event occurred along f1, dislocated sandy gravel layer U1, and resulted in one typical colluvial wedge (W1). Colluvial wedge deposition generally occurred soon after the rupturing event, indicating that the older event occurred between the deposition of layer U1 and wedge W1, shortly before the deposition of W1.

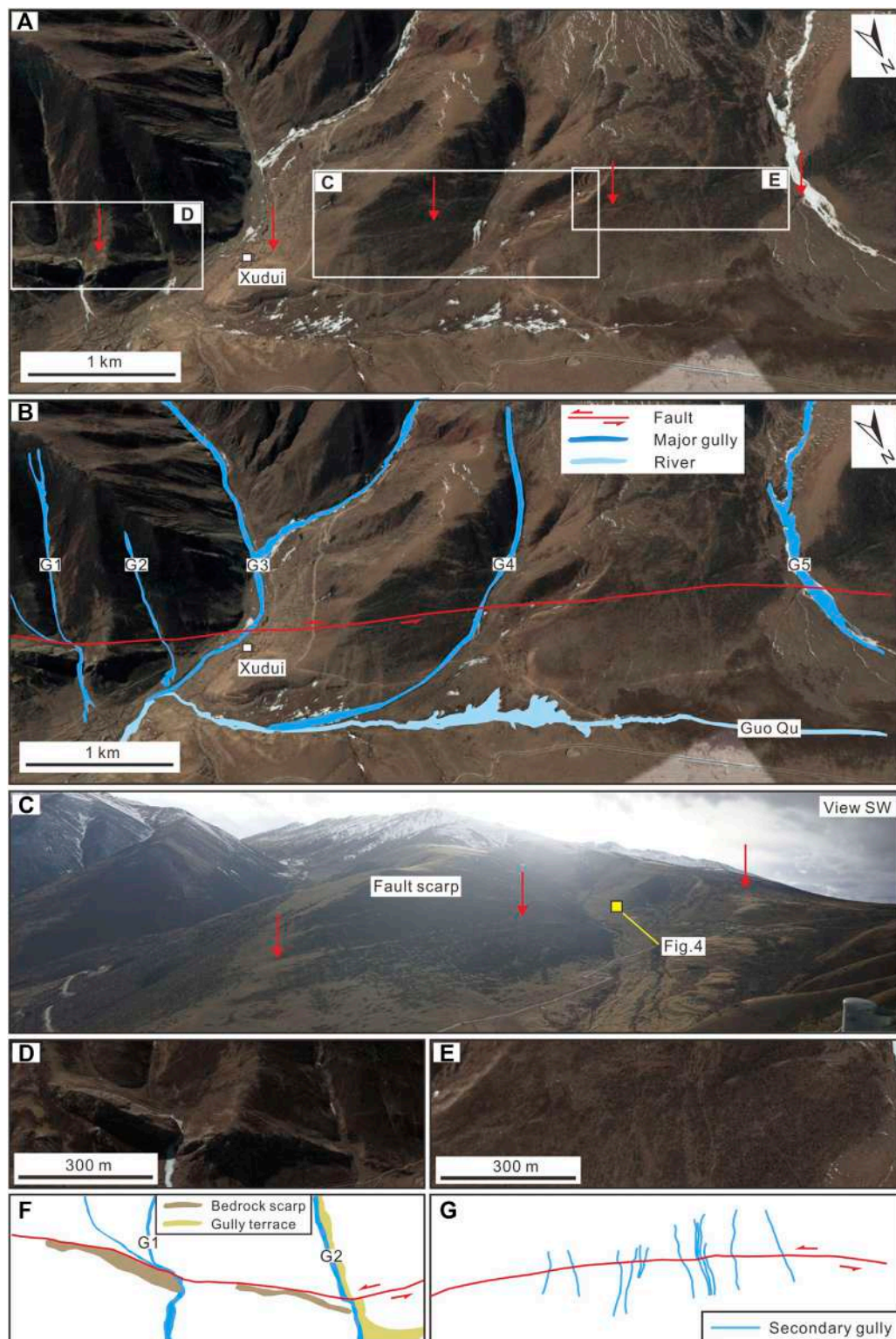


FIGURE 14

Offset landforms around Xudui village and location of the XD fault outcrop. (A) Satellite image and (B) interpreted map of the tectonic landforms near Xudui village. The location is shown in Figure 2. G1–G5 indicate the major gullies. (C–E) Local enlarged photograph of (A). (F) Interpretation map of the offset landforms in (D). (G) Interpretation map of the offset landforms in (E).

TABLE 4 Unit description of exposure XD.

Unit	Description
U1	Steel-gray sand-gravel layer. The gravel particles are angular
U2	Earthy-yellow sand-gravel layer. The gravel particles are poorly rounded. This layer exhibits a smaller gravel size than that of U1
U3	Steel-gray sand-gravel layer. The gravel particles are poorly rounded
U4	Black soil and turf layer

TABLE 5 Radiocarbon samples and dating results for the three trenches and one outcrop.

Sample	Lab code	Radiocarbon age (yr BP)	Calibration age (cal BP)	Analyzed material	Unit sampled
GQ1-C4	506534	12,480 ± 40	14,651 ± 166	Organic sediment	U2-1, GQ1
GQ1-C6	506536	10,750 ± 40	12,729 ± 16	Organic sediment	U3-2, GQ1
GQ1-C7	506537	12,550 ± 40	14,878 ± 152	Organic sediment	U3-2, GQ1
GQ1-C8	506538	12,220 ± 40	14,143 ± 96	Organic sediment	U3-2, GQ1
GQ1-C10	506539	10,800 ± 30	12,745 ± 13	Organic sediment	U3-2, GQ1
GQ1-C11	506540	5,280 ± 30	6,073 ± 68	Organic sediment	U4-2, GQ1
GQ1-C12	506541	5,880 ± 30	6,703 ± 36	Organic sediment	U4-2, GQ1
GQ1-C13	506542	2,950 ± 30	3,107 ± 54	Organic sediment	U5-1, GQ1
GQ1-C15	506543	1,270 ± 30	1,220 ± 42	Organic sediment	U5-2, GQ1
GQ1-C16	506544	12,440 ± 40	14,601 ± 173	Organic sediment	U4-1, GQ1
GQ1-C17	506545	8,340 ± 30	9,361 ± 57	Organic sediment	U4-1, GQ1
GQ2-C6	506548	3,100 ± 30	3,307 ± 45	Organic sediment	U4-2, GQ2
GQ2-C7	506549	2,030 ± 30	1,969 ± 44	Organic sediment	U4-2, GQ2
GQ2-C8	506550	920 ± 30	831 ± 49	Organic sediment	U5, GQ2
GQ2-C9	506551	9,120 ± 40	10,286 ± 57	Organic sediment	U2-2, GQ2
BQ-C1	532867	7,310 ± 30	8,104 ± 45	Organic sediment	U3, BQ
BQ-C2	532868	6,430 ± 30	7,359 ± 42	Organic sediment	U3, BQ
BQ-C3	532869	2,090 ± 30	2,054 ± 50	Organic sediment	U4, BQ
LCJ-C2	513060	1,470 ± 30	1,347 ± 25	Organic sediment	U3, XD

4.3.3.2 Event XE1

The youngest, most recent event occurred along faults f1 and f2 and faulted layers U1, U2 and U3. One colluvial wedge deposit W2 was rapidly formed along the f2 main fault plane and capped by layer U4. These characteristics suggest that the younger event occurred between the deposition of layers U3 and U4. To constrain the timing of this event, one soil sample (LCJ-C2) containing organic material was collected from the central sector of layer U3 for ^{14}C dating (Figure 4B). Based on the radiocarbon age, we suggest that the most recent event occurred after $1,347 \pm 25$ yr BP.

4.4 Radiocarbon dating of events

Numerous organic mud samples were collected in the above three trenches (Figures 7, 9, 12B) and one outcrop section (Figure 4B). Nineteen samples were sent to Beta Analytic Inc. for radiocarbon dating *via* the AMS method. Radiocarbon dates

are summarized in Table 5. Most radiocarbon dates of the trench samples generally occurred in correct stratigraphic order, such that younger samples overlaid older samples. The ages of three samples were questionable (Figure 7): the age of sample GQ1-C7 ($14,878 \pm 152$ yr BP) in U3-2 was older than that of GQ1-C4 ($14,651 \pm 166$ yr BP) in underlying layer U2-1. Similarly, the age of GQ1-C16 ($14,601 \pm 173$ yr BP) in U4-1 was older than that of GQ1-C6 ($12,729 \pm 16$ yr BP), GQ1-C10 ($12,745 \pm 13$ yr BP) and GQ1-C8 ($14,143 \pm 96$ yr BP) in underlying layer U3-2, revealing obvious stratigraphic inversion. The age of sample GQ1-C8 ($14,143 \pm 96$ yr BP) in layer U3-2 was much older than those of GQ1-C6 ($12,729 \pm 16$ yr BP) and GQ1-C10 ($12,745 \pm 13$ yr BP) in the same layer.

In addition, we observed that the ages of the above three samples, which were regarded as outliers, were all close to the age of sample collected from older strata. Specifically, the ages of samples GQ1-C7 ($14,878 \pm 152$ yr BP), GQ1-C8 ($14,143 \pm 96$ yr BP) and GQ1-C16 ($14,601 \pm 173$ yr BP) were similar to that of GQ1-C4 ($14,651 \pm 165$ yr BP). Therefore, we interpreted them as

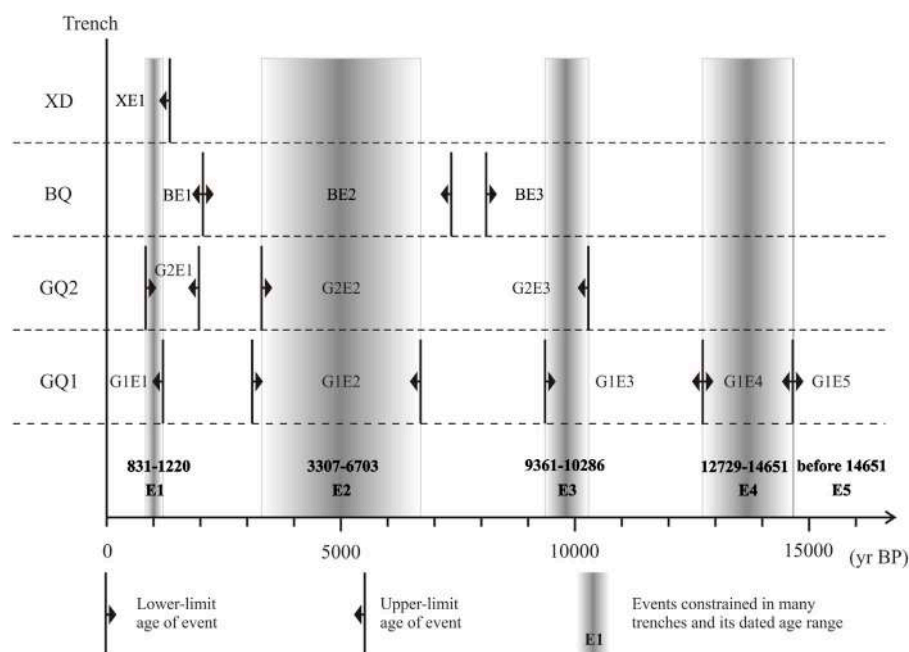


FIGURE 15

Constraints of the paleoseismic events and ages from the three trenches (GQ1, GQ2, and BQ) and one exposure (XD).

redeposited materials, and their ages represented early stratigraphic ages. The ages of these three samples were not used when constraining the ages of paleoearthquakes. Based on the ages of the other sixteen samples, we used the progressive confining method to constrain the timings of paleoseismic events recorded in the three trenches and one exposure. The constraint results are shown in Figure 15.

5 Discussion

5.1 Paleoseismic sequence and seismic recurrence behavior of the Yangda-Yaxu fault

Through trench excavations at the Guoqing and Bengqing sites, combined with one fault exposure at the Xudui site, five faulting events, named E5 through E1 from oldest to youngest, were identified at before 14,651, 12,729–14,651, 9,361–10,286, 3,307–6,703, and 831–1,220 yr BP, respectively (Figure 15). Unfortunately, due to the lack of recorded historical data in this region, no surface-faulting event in the Historical Strong Earthquake Catalog of China could be associated with the latest event (Science and Technology Committee of the Tibetan Autonomous Region, 1982; Department of Earthquake Disaster Prevention and State Seismological Bureau., 1995).

Based on the abovementioned event-constrained ages, the intervals between events E5, E4, E3, E2, and E1 were

approximately >961, 3,866, 4,819, and 3,979 yr. Trenches GQ1 and GQ2 were excavated across the local low-lying area of the fault scarps, and trench BQ was excavated across sag pond. The excavation locations of the three trenches were all in a relative depositional environment rather than a denudation environment. Combined with the stratigraphic depositional sequence revealed by these trenches and one fault outcrop, we propose that the sedimentary sequence in this paper is likely relatively continuous, and the paleoseismic sequence may be relatively complete. If the events are complete, the recurrence interval of surface-faulting earthquakes along the YYF follows a quasi-periodic pattern with an interval of ~4,000 yr.

Despite the lack of strong earthquake records, the trench work in this paper revealed that at least five surface-rupturing events have occurred since the late Pleistocene. According to the horizontal coseismic displacement (~2–3 m) and the recurrence interval of the latest two earthquake events (~3,979 yr) obtained in this paper, we obtain a horizontal slip rate of ~0.5–0.8 mm/yr since the late Holocene.

5.2 Tectonic implications and the regional seismic risk

As mentioned above, the YYF cuts through the NF and the LCJF, which are ancient regional deep faults with a long and complex history of development and evolution, and divides them into two sections with significantly different activity levels.

Specifically, on the south side of YYF, both the NF and the LCJF are Holocene active faults. On the north side of YYF, the NF is active in the late Pleistocene, and the LCJF is active in the early and middle Pleistocene (Han, 2022a). Therefore, the YYF is an important fault structure in the region, which has a significant control effect on the late Quaternary evolution of the NF and the LCJF. Combining the clear linear geomorphic features along the fault and the paleoearthquake results in this paper, we believe that YYF is a newly-generated active fault with a relatively high maturity.

In recent years, the increasingly abundant observation data also shows that the interior of the QTB and CDB are not rigid and fixed, and many internal faults or local segments, for example, LTF, BTF, and BLF, also exhibit obvious Holocene activity characteristics and relatively strong seismogenic capacity, as evidenced from geological, paleoseismic, seismic data investigations (Gao, 2021; Wang et al., 2021; Han, 2022a; Han et al., 2022b; Ren et al., 2022b). For example, the BTF undergoes a dextral rate of $\sim 1.3\text{--}2.7$ mm/yr during late Quaternary (Zhou et al., 2005), and the LTF is a predominantly left-lateral strike-slip fault with a late Quaternary strike-slip rate ranging from $\sim 2\text{--}4$ mm/yr (Xu et al., 2005b; Zhou et al., 2005; Zhou et al., 2007). Our results reflect that the YYF is mainly a left-lateral strike-slip fault with a horizontal rate of $\sim 0.5\text{--}0.8$ mm/yr during the late Holocene, comparable to the BTF, LTF and other major faults within the QTB and CDB. Studies have shown that these faults have occurred multiple large earthquakes since the late Quaternary (Department of Earthquake Disaster Prevention and State Seismological Bureau, 1995; Zhou et al., 2005; Gao, 2021). The occurrence of strong earthquakes helped to release the accumulated strain energy attributed to Indian-Eurasian plate collision and accommodate the tectonic deformation associated with eastward extrusion of the upper crust of the Tibetan Plateau. Therefore, the active faults within the QTB and CDB, including YYF, play important role in accommodating crustal deformation, and should be given more attention when assessing the regional seismic risk and examining regional deformation in the future.

6 Conclusion

1) Our geomorphic observations and paleoseismic trenches demonstrated that the YYF is an active fault with a left-lateral strike-slip rate ranging from $\sim 0.5\text{--}0.8$ mm/yr since the late Holocene. At least five surface-faulting events with constrained ages ranging from before 14,651, 12,729–14,651, 9,361–10,286, 3,307–6,703, and 831–1,220 yr BP have occurred since the late Pleistocene. The recurrence interval of major earthquakes along the YYF follows a quasi-periodic pattern with an interval of $\sim 4,000$ yr.

2) The YYF is a newly-generated active fault. The fault has a significant control effect on the late Quaternary evolution of the NF and the LCJF, and also plays an important role in accommodating crustal deformation.

Data availability statement

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author.

Author contributions

The study was designed by LC, the field work was done by MH, YL, SG, JF, and LC, the writing and preparation of the original draft were conducted by MH, the review and editing of the final manuscript were performed by LC and YL. All authors contributed to the article and approved the submitted version.

Funding

This research was supported financially by the Second Tibetan Plateau Scientific Expedition and Research Program (STEP) (2019QZKK0901); National Natural Science Foundation of China (41872228); and Basic Scientific Work of the Institute of Geology, China Earthquake Administration (IGCEA1418).

Acknowledgments

The authors greatly thank Associate Professor Hu Wang at the Faculty of Geosciences and Environmental Engineering, Southwest Jiaotong University, for his constructive and detailed comments to improve the manuscript. The authors also thank Peng Guo and Guodong Bao for their advice on our preliminary results.

Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

The reviewer JR declared a shared affiliation with the authors MH, YL, SG, and JF to the handling editor at the time of review.

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OPEN ACCESS

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SPECIALTY SECTION
This article was submitted to Structural
Geology and Tectonics,
a section of the journal
Frontiers in Earth Science

RECEIVED 27 May 2022
ACCEPTED 07 September 2022
PUBLISHED 16 January 2023

CITATION
Sun W, He H, Shi F, Wei Z, Sun H and Su P
(2023), Late-Quaternary
paleoearthquakes along the
Liulengshan Fault on the northern
Shanxi Rift system.
Front. Earth Sci. 10:954335.
doi: 10.3389/feart.2022.954335

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Late-Quaternary paleoearthquakes along the Liulengshan Fault on the northern Shanxi Rift system

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The Liulengshan Fault (LLSF), which lies on the northeastern edge of the Ordos Plateau, is a controlling boundary fault in the northern part of the Shanxi Rift system (SRS). The displaced landforms show that the fault has undergone strong and frequent late-Quaternary seismic activities. In 1989 and 1991, two moderate–strong earthquake swarms ($M_s=6.1$ and $M_s=5.8$) successively occurred in the LLSF, and GPS velocity shows that the areas are extending at around 1–2 mm/a. However, there is no surface-rupturing earthquake reported on the LLSF in historical records. Thus, the study of paleoseismic history and rupture behavior of paleoearthquakes in late-Quaternary on the LLSF is of fundamental importance for understanding the future seismic risk of this fault. To solve these problems, we conducted paleoseismological trench excavations at two sites on the LLSF to establish its paleoearthquake history. On the basis of the field geological survey and interpretation of high-precision topographic data, we carried out large-scale fault mapping and excavated two trenches in Xujiabao and Luofengwa across the LLSF. Then, four events in the Xujiabao trench and three events in the Luofengwa trench are identified. Finally, combined with radiocarbon dating (C_{14}), optically stimulated luminescence (OSL) and OxCal modeling, we constrained the ages of these events. Together with the previous results of paleoseismology in Yin et al. (1997), we consider that different segments of the LLSF may rupture together at the same time. Therefore, a total of six paleoearthquake events since late-Quaternary have been finally confirmed at 44,151–30881a, 40,163–28045a, 28,233–19215a, 16,742–12915a, 12,788–8252a, and 8203–2300a BP. According to the empirical relationships between moment magnitude and rupture length, the best estimated magnitude is inferred to be in the range between M_w 6.9 and M_w 7.7. Considering the strong late-Quaternary activity and a long earthquake elapsed time, we propose that the LLSF might have a high seismic hazard potential in the near future.

KEYWORDS

paleoseismology, trenching, geological survey, Liulengshan Fault (LLSF), Shanxi Rift system (SRS)

Introduction

The Shanxi Rift system (SRS) of northern China is wellknown for its intense neo-tectonics and strong earthquakes (Molnar and Tapponnier, 1977; Northrup et al., 1995; Ren et al., 2002). The main boundary faults of the SRS are active, and in the past thousand years, a number of destructive earthquakes ($M \geq 7$) occurred along these faults (Lee et al., 1978; Liu et al., 2004; Deng et al., 2007; Rao et al., 2018; Xu et al., 2018) (Figure 1). Previous studies in the northern Shanxi Grabens, at the northeastern corner of the Ordos Plateau in northern China, have found extension rates of anywhere between 0 and 6 mm

a⁻¹ at an azimuth between 95° and 180° (Zhang et al., 1998; Shen et al., 2000; Wang et al., 2011; Wang et al., 2012; Zhang et al., 2018). On the basis of Quaternary, seismic, and geodetic rates of strain, Middleton et al. (2017) conclude that the northern Shanxi Grabens are extending at around 1–2 mm/a at an azimuth of $\approx 151^\circ$.

Contrastingly, there is no historical record of large-magnitude earthquakes ($M \geq 7$) in the northern Shanxi Rift system (Datong Graben) (Figure 1) (Wesnousky et al., 1984; China Earthquake Administration, 1999a; Liu et al., 2007). Determining whether large-magnitude earthquakes can occur in the northern Shanxi Rift system is very important for

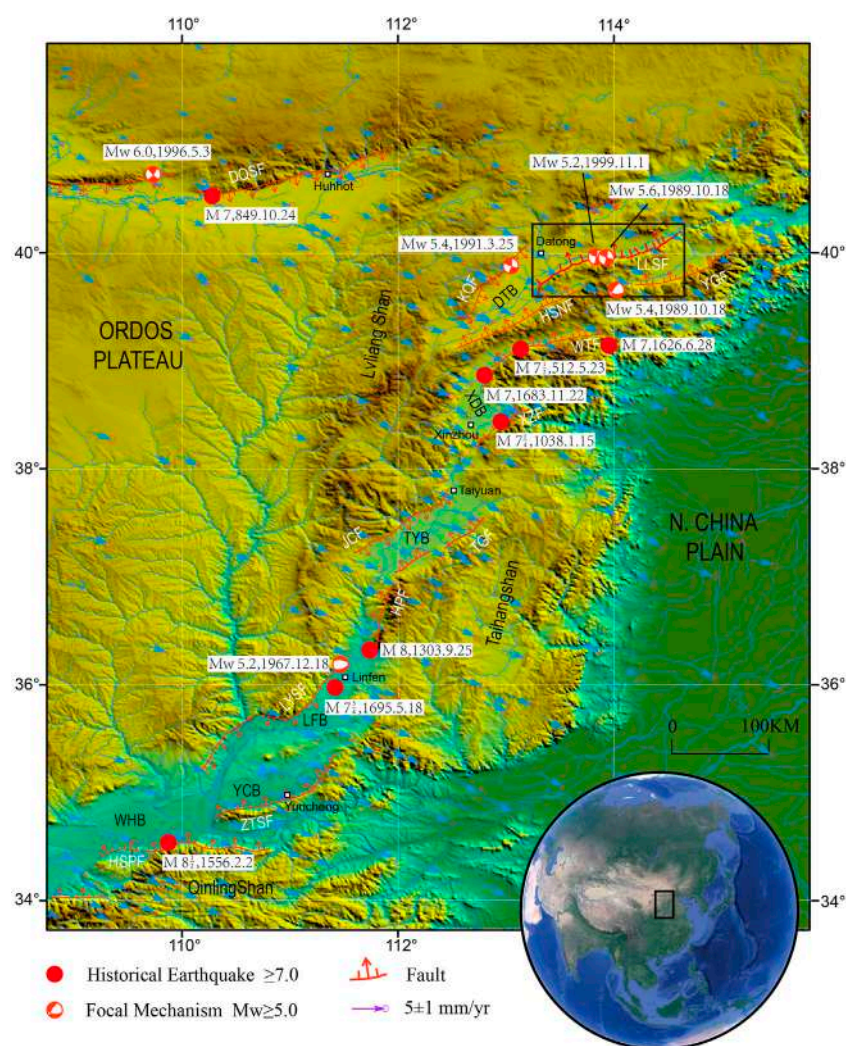


FIGURE 1

Shuttle Radar Topography Mission (SRTM) topography of the Shanxi Rift system in North China. The red lines and red circles represent major active faults and major historical earthquakes with magnitudes greater than $M7$, respectively (Working Group on Historical Earthquake Compilation of Shanxi Province Earthquake Administration, 1991). Horizontal GPS (global position system) velocities relative to the stable Eurasian plate (Wang and Shen, 2020), focal mechanisms of instrumental earthquakes with $Mw > 5$ (CMT catalog, Ekström et al., 2012). Inset globe shows the location of Figure 1 with Asia. Major faulted basins and faults: WHB, Weihe Basin; YCB, Yuncheng Basin; LFB, Linfen Basin; TYB, Taiyuan Basin; XDB, Xinding Basin; DTB, Datong Basin; HSPF, Huashan Piedmont Fault; ZTSF, Zhongtiaoshan Fault; LYSF, Luoyunshan Fault; HPF, Huoshan Piedmont Fault; JCF, Jiaocheng Fault; TGF, Taigu Fault; XZF, Xizhou Shan Fault; WTF: Wutaishan Fault; HSNF: Hengshan Northern Fault; YGF, Yangyuan Graben Fault; KQF, Kouquan Fault; DQSF, Daqingshan Fault; and LLSF, Liulengshan Fault.

TABLE 1 Unit descriptions from the exposed walls of the Xujiabao trench.

Unit no.	Description
U1	Sequence of earthy pebbly fine sand-silt layer with several brown gravel beds, with clear stratification. This unit can be further divided into many subunits, but there is no obvious layer opposite to it in the hanging side. The gravel diameter in gravel beds is generally 3–5 cm, and the psephicity is normal
U2	The unit consists of two subunits, u2-1 and u2-2, with a gravel layer at the top and a silt layer containing fine gravel at the bottom
U3	u3-1, u3-2, and u3-3 of the main fault's hanging side are composed of two silty sand layers sandwiched by a sand–gravel layer about 60 cm thick, which is covered by colluvial wedge and loess
W1/W2	Triangular red brown colluvial wedge (W1 and W2) near the fault, with sand and gravel mixed together
U4	Loess silty layer with an average thickness of about 3 m in the upper wall of the main fault covers the colluvial wedge W1
U5	Set of loess layers with relatively developed joints. The loess at its top contains more small gravel
U6	Light gray–brown loess with a few gravel and well jointed

TABLE 2 Summary of luminescence dating results from the Xujiabao profile on the LLSF.

Sample name	Depth (m)	Water (%)	Ra ^a (Bq/kg)	Th ^a (Bg/Kg)	K ^a (Bg/Kg)	U ^a (Bg/Kg)	Da ^b (Gy/ka)	De ^c (Gy)	Age (ka)
XJB201704	2.50	1.51	31.7 ± 1.1	45.8 ± 1.52	555.2 ± 26.9	32.6 ± 7.9	3.2 ± 0.2	67.8 ± 2.6	21.2 ± 1.9
XJB201705	3.20	2.00	29.4 ± 1.0	41.6 ± 1.4	517.0 ± 25.0	48.3 ± 10.1	3.0 ± 0.2	78.8 ± 4.4	25.9 ± 2.6
XJB201710*	7.0	1	21.2 ± 0.4	31 ± 0.4	430.3 ± 8.0	38.5 ± 6.0	2.9 ± 0.1	119.4 ± 5.9	41.1 ± 2.4
XJB201713*	3.5	1	25.1 ± 0.5	34.3 ± 0.5	474.1 ± 8.9	39.8 ± 6.7	3.3 ± 0.1	95.7 ± 2.5	29.2 ± 1.2
XJB201722*	0.5	1	21.2 ± 0.5	39.7 ± 0.6	477.9 ± 10.1	22.6 ± 5.1	3.4 ± 0.1	11.3 ± 0.4	3.4 ± 0.2

^aUranium, thorium, and potassium concentrations were measured using an inductively coupled plasma mass spectrometer (ICP-MS).

^bDa, annual dose.

^cDe, equivalent dose.

Note: XJB201704 and XJB201705 were prepared and analyzed by Key Laboratory of Crustal Dynamics, Institute of Crustal Dynamics, and CEA (Beijing, China), whereas XJB201710, XJB201713, and XJB201722 (marked by *) were analyzed by the State Key Laboratory of Earthquake Dynamics, Institute of Geology, and China Earthquake Administration (IGCEA) (Beijing, China).

understanding both the geodynamic mechanisms and strong earthquake hazard assessment of the SRS (Xu et al., 1992b, 1993; Xu and Ma, 1992).

The northern SRS lies at the northeastern corner of the Ordos Plateau, a piece of rigid continental lithosphere (Zhao et al., 2017). The seismicity in the northern SRS is of significant concern because it is adjacent to North China, one of the most densely populated regions on Earth. The cities of Beijing, Datong, Shijiazhuang, and Taiyuan, with a total population of more than 35 million people, all lie in or next to the northern Shanxi Rift system (Xu et al., 2002; Middleton et al., 2017; Rao et al., 2018).

The Liulengshan Fault is one of the major faults of the northern SRS, in which two moderate–strong earthquake swarms ($M_s=6.1$ and $M_s=5.8$) successively occurred in 1989 and 1991 (Wang and Wang, 1992; Deng et al., 1994; Xu et al., 1996). Strangely, previous studies suggest that the Liulengshan Fault is dominated by normal faulting in the late-Quaternary and seems to have been slipping at a rate of 0.5 mm/a (Liu et al., 1991; Xu et al., 1996; Sun, 2018; Luo et al., 2022), but the focal mechanisms of the two moderate–strong earthquake are both strike–slip. The discrepancies between the focal mechanisms of the

moderate–strong earthquakes and the fault's sense from Quaternary geology are important for our understanding of how both crustal deformation and seismic hazard may vary through time. In addition, the previous work laid more emphasis on its geometrical structure and the fault slip rate. Because of human activity and intense erosion, the traces of the Liulengshan Fault are indistinguishable, and only Yin et al. (1997) carried out the paleoseismic research work, and the research on the paleoseismic activity history of the fault is weak at present.

To solve those problems, we mapped the fault traces using high-resolution satellite images and aerial photographs and conducted paleoseismic trench excavations at two sites along the Liulengshan Fault. Combined with the radiocarbon and optically stimulated luminescence (OSL) dating techniques, we worked out the paleoearthquake history in the late-Quaternary and analyzed the future seismic potential of this fault.

Geological background

The SRS varies in width from 20–80 km and is ~1,200 km long (Figure 1; State Seismological Bureau (SSB), 1988). The SRS

TABLE 3 Unit descriptions from the exposed walls of the Luofengwa trench.

Unit no.	Description
U1	The unit locates in the footwall side of the fault and has a total height of about 5 m. It can be divided into eight subunits based on the proportion, size, roundness, and other factors of the unit: u1-1 (the bottom), the gravel is 3–10 cm in diameter with medium roundness; u1-2–u1-5 (the middle), the diameter of the gravels which varies greatly, ranging from 5 to 40 cm, and the gravel with large diameter has better roundness, while the other gravel has medium roundness; u1-6–u1-8 (the top), the gravel content decreases and its roundness is medium, and the sandy clay content increases
U2	The unit locates at the bottom of the hanging side and can be divided into two subunits: u2-1 is the gravel layer, the gravel diameter is small, generally less than 2 cm, the roundness of which is good; the u2-2 is the red sand layer
U3	Red sandy clay layer about 1.5 m thick, of which u3-2 is the reddish brown ancient soil layer
U4	Reddish brown silty clay layer
U5	Reddish brown layer of silty soil with a small amount of fine gravel
W1	Triangular grayish brown wedges, with sand and gravel mixed together
W2	Reddish brown gravel silt, wedge shaped
W3	Gravel mixed packing wedge, gravel in all shapes and sizes, and poor roundness

is bound by north and northeast striking normal faults that form an S-shaped series of en echelon basins. These basins form a narrow, north–northeast-trending right-lateral transtensional zone in the middle sections and two wide, east–northeast-trending extensional segments marking its northern and southern terminations (Li et al., 1998; Yu, 2004; Luo et al., 2021). The initial extension in the SRS commenced at the end of the Miocene (Ye et al., 1987; Zhang et al., 1998, 2003), and the main phases of rifting occurred throughout the Shanxi Rift system during the Pliocene and Quaternary (State Seismological Bureau, 1988; Xu and Ma, 1992). It is likely that the location and geometry of the grabens are, in part, determined by preexisting, northeast striking, and fold-and-thrust belt

structures of the Late Triassic and Middle Jurassic-Cretaceous orogenies (Xu and Ma, 1992, 1993). Manifestations of these older orogenies are still evident today as a thickened crust (40–44 km) and a corresponding gravity anomaly over the region (Xu and Ma, 1992). Tapponnier and Molnar (1977) note that the width and length of individual grabens decrease from south to north, while the mean elevations of the graben floors increase (Li et al., 2015), suggesting a northward propagation of the onset of rifting through time.

The Liulengshan Fault runs along the northwestern side of the Liulengshan, an S-shaped mountain range in the center of the northern SRS, forming part of the eastern edge of the Datong Graben and the southern edge of the subsidiary Yangyuan Graben

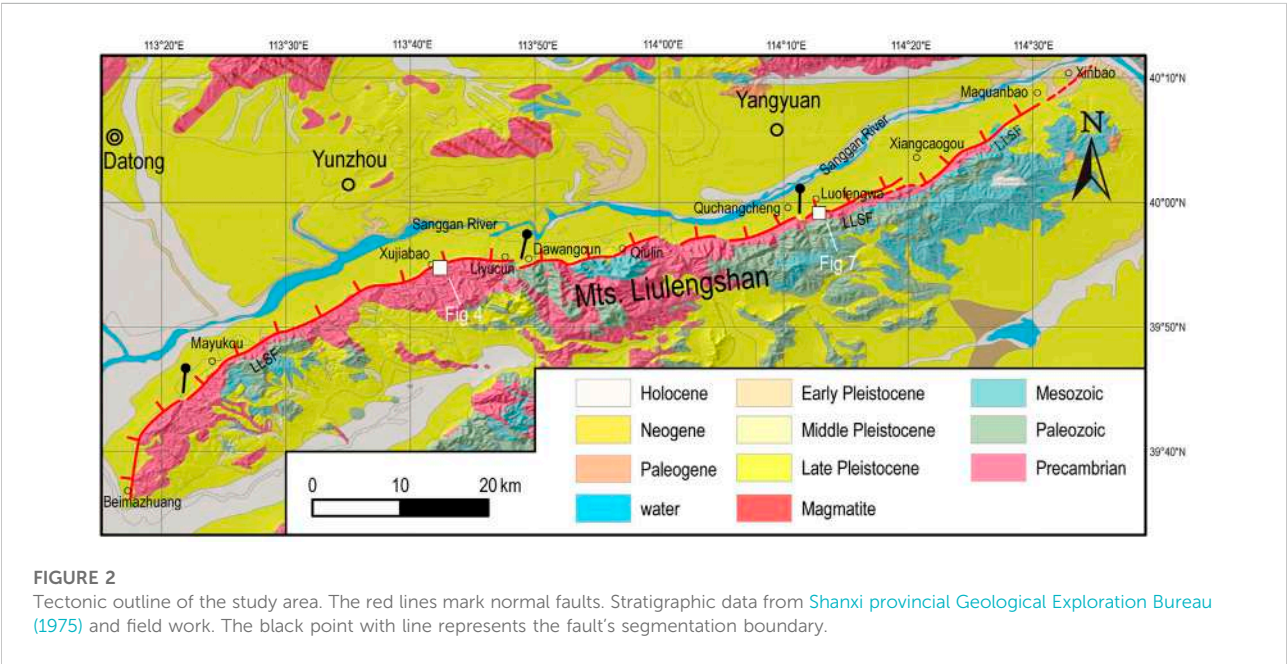


FIGURE 2
Tectonic outline of the study area. The red lines mark normal faults. Stratigraphic data from Shanxi provincial Geological Exploration Bureau (1975) and field work. The black point with line represents the fault's segmentation boundary.

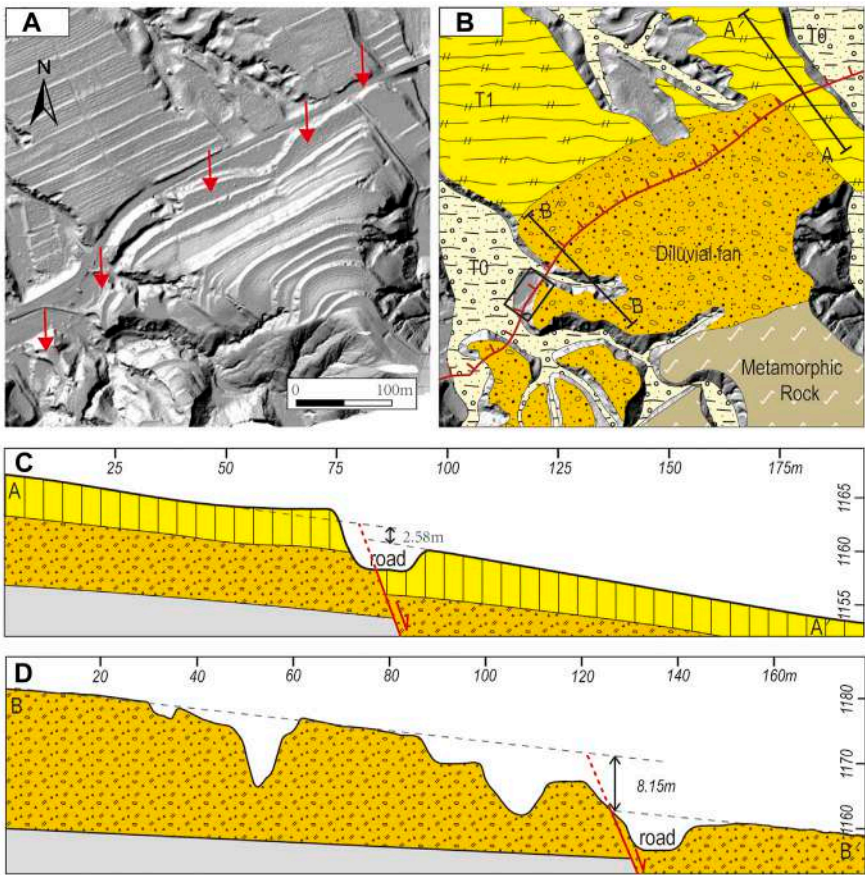


FIGURE 3
Tectonic landforms and simplified stratigraphy profile near the Xujiabao trenching site. (A) s-UAV (small unmanned aerial vehicle) DEM imagery of the Xujiabao alluvial fan, where red arrows represent the trace of the Liulengshan Fault; (B) interpreted map. The red line indicates the trace of the Liulengshan Fault; T0–T1, alluvial phases; the black box in Figure 3B indicates the position of the trench; (C) and (D) as the topographic profiles for A–A' and B–B' in (B), respectively. The color version of the figure is available only in the electronic edition.

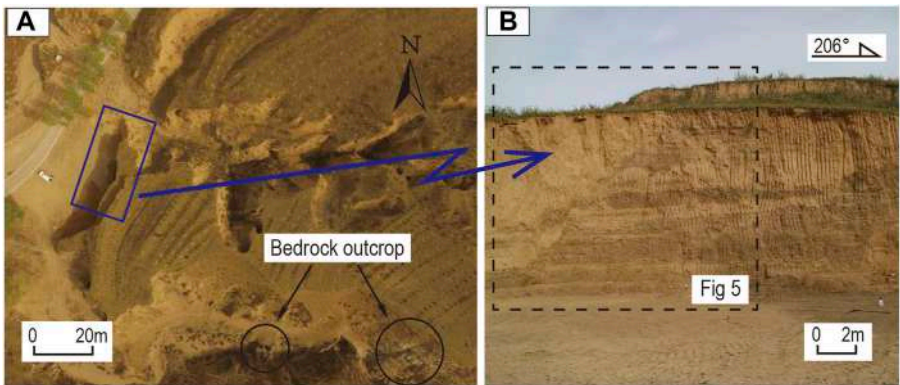


FIGURE 4
(A) Top view shows Site 1 near Xujiabao village. See Figure 1 for the location. (B) Field photograph of the fault zone.

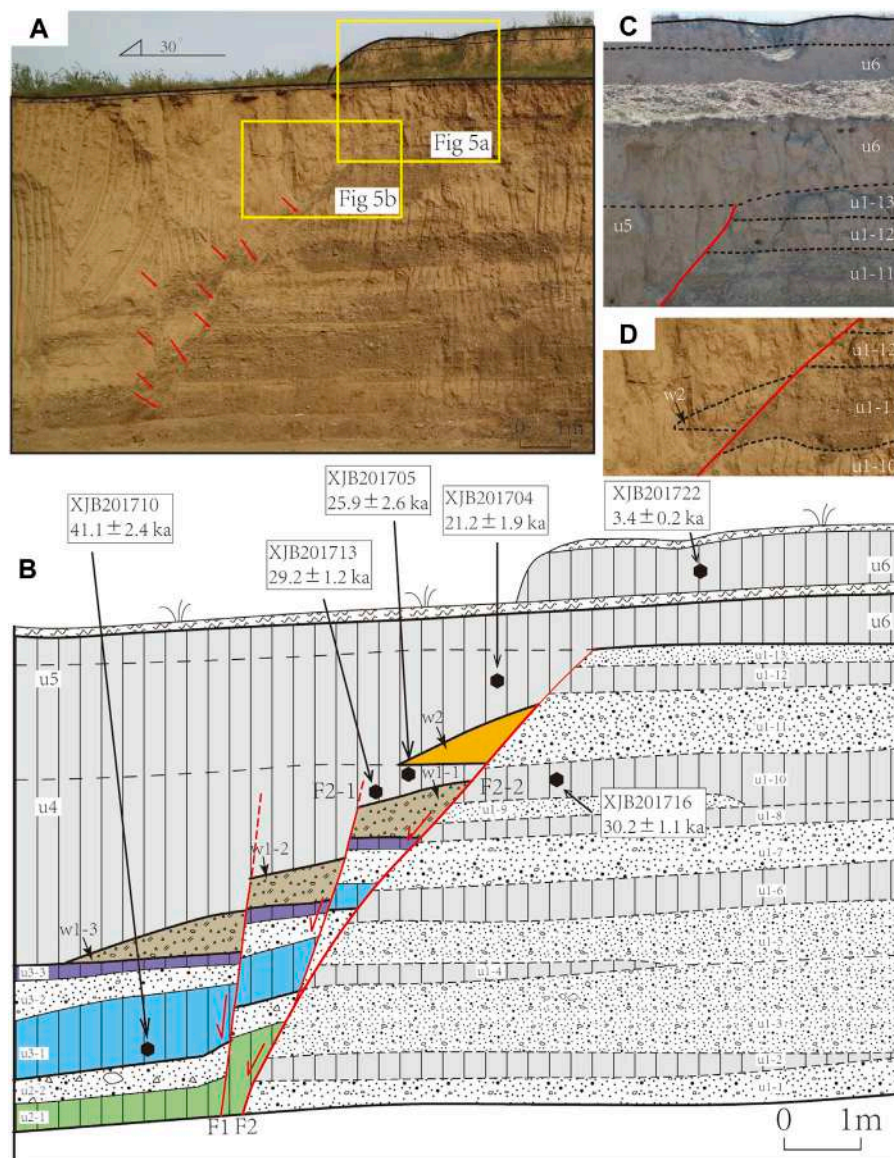


FIGURE 5

Photograph mosaic (A) and the interpreted map (B) of the Xujiabao profile. U denotes the stratum unit, and black lines show the stratigraphic contacts between units or subunits. Red lines show the fault planes. Small black filled hexagons show the locations of OSL samples, labeled with sample numbers and their ages (Table 2). (C, D), Local enlarged photographs, showing the details of the trench exposures; see locations in (A).

(Figure 2). The LLSF is a complex fault zone composed of a series of secondary faults. It starts from Xinbao of Yangyuan County in the east; passes Maquanbao, Luofengwa, Dawangcun, and Xujiabao in the west; and gradually transforms as the northeast-trending piedmont fault near Mayukou. It generally strikes east–northeast and dips north–northwest, for a total length of about 130 km. Based on previous geological and geomorphological investigations and studies of the geometric structure and segmentation of the fault, this study considers that the LLSF can be divided into four sections: Xinbao–Quchangcheng (XQ, 39 km), Quchangcheng–Dawangcun

(QD, 31 km), Dawangcun–Mayukou (DM, 34 km), and Mayukou–Beizhuang (MB, 27 km) (Deng et al., 1994; Duan and Fang, 1995; Cheng et al., 1996). There are inclined sequence discontinuities with a width greater than or equal to 1.5 km, which are generally covered by upper Pleistocene Malan loess, as fault boundaries between each section. The Xinbao–Quchangcheng section was active in the early Pleistocene, and the other three sections were active in the Late Pleistocene to Holocene (Xu et al., 1996).

The Liulengshan Fault has been active since the Late Pleistocene, with various displaced landforms, including linear

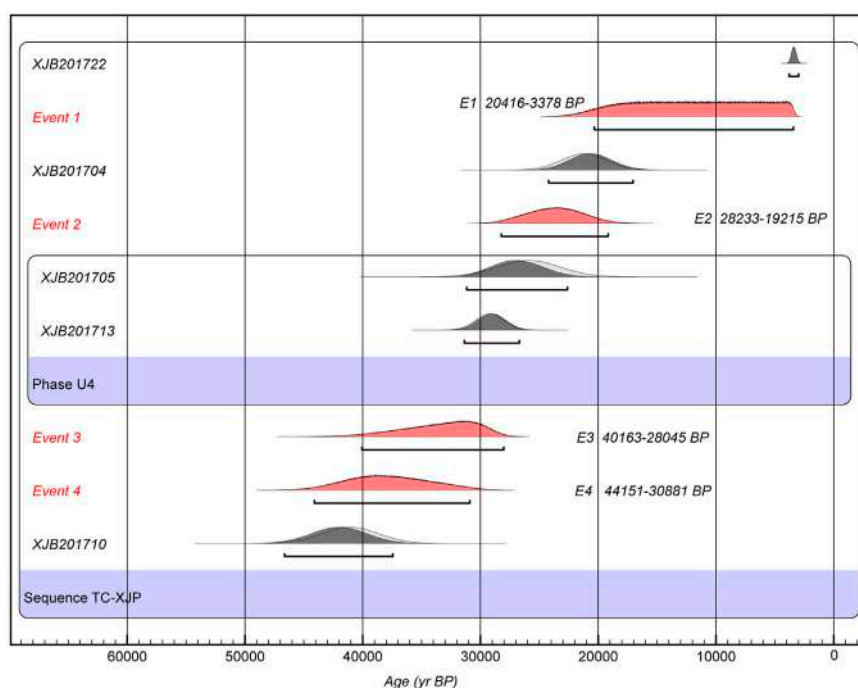


FIGURE 6

Results of OxCal analysis of luminescence dates from the Xujiabao trenching site. Unfilled areas are prior probability distributions, and solid curves are posterior distributions after OxCal analysis. Phases are summed probability for units with multiple luminescence dates.

fault scarps, bedrock fault surfaces, and offset streams. Based on detailed topographic measurements of typical displaced landforms, the vertical slip rates range from 0.2–0.6 mm/a, with an average value of 0.3 mm/a (Sun, 2018). Two earthquake swarms, with maximum magnitudes of M_s 6.1 and M_s 5.8, occurred on the Liulengshan Fault in October 1989 and March 1991, respectively (Xu et al., 2016). The focal depths of these earthquakes are thought to be in the range of 10–20 km (Xu et al., 1993) and there are no reports of surface ruptures associated with these events.

Study methods

Small unmanned aerial vehicle measurement

The study area is located in the north of Shanxi Province of China, a semi-arid area with sparse vegetation, most of which are low shrubs. We obtained the DEM (digital elevation model) and DOM (digital orthophoto map) data of the target area by using the s-UAV measurement (small unmanned aerial vehicle measurement). These data can truly reflect the topographic and geomorphological features and can provide assistance for identification of active faults

and their strike, as well as for the precise determination of parameters like displacement of the faults (James and Robson, 2012; Fonstad et al., 2013).

In this study, the DJI Phantom 3 unmanned aerial vehicle was used for image acquisition of the target area. The operating height of the aerial vehicle is 120 m, and the repetition rate of adjacent photos is up to 80%. The light intensity is relatively uniform throughout the whole process, without obvious shadow change. A total of 1,788 images in the RAW format have been collected. Agisoft PhotoScan, a semi-automatic software, was used for image processing. First, the images collected were stitched so as to reconstruct the point cloud data of the target area; second, the mesh grid model and orthophoto were generated. Finally, ground control points were added to generate the orthophoto with geographic coordinate information and the DEM (digital elevation model) (Verhoeven, 2011; Johnson et al., 2014; Wei et al., 2015).

Normal faults tend to form steep fault scarps and composite fault scarps, and these geomorphological phenomena can be clearly seen from the derived images (Figures 3,7). However, in some areas, fault scarps have been destroyed by human activities, which has a certain impact on extracting profile features from the images. In this article, two topographic profiles (Figures 3C,D) have been extracted from the area

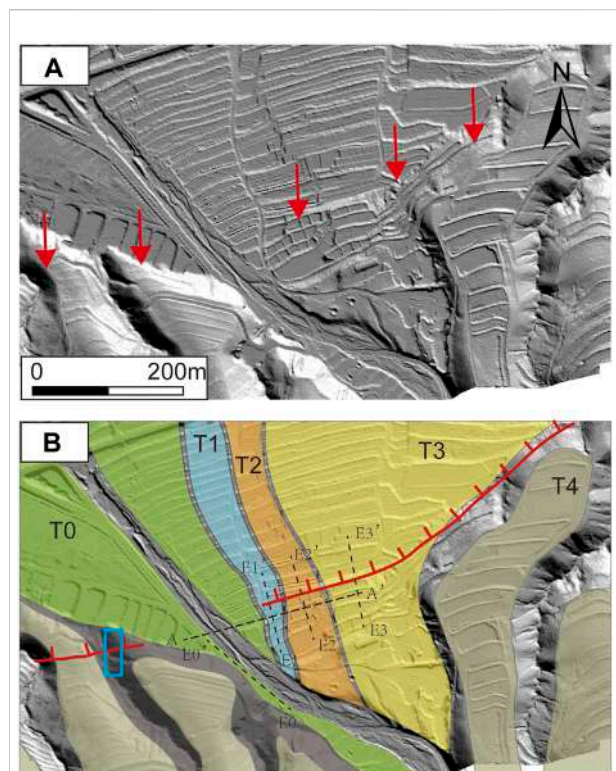


FIGURE 7

Tectonic landforms near the Luofengwa trenching site. (A) s-UAV DEM imagery of the Luofengwa alluvial fan, red arrows represent the trace of the Liulengshan Fault; (B) interpreted map. The red line indicates the trace of the LLSF; T0–T4, alluvial phases; A–A', E1–E1', E2–E2', and E3–E3' are topographic profiles; the blue box in Figure 7B indicates the position of the trench. The color version of the figure is available only in the electronic edition.

near Xujiabao. The vertical displacements of the faults measured in the profiles are 2.58 and 8.15 m, respectively. Thus, the position of the faults has been determined from the geomorphology.

Trench excavation

Trenching is the most direct and efficient approach in paleoseismology to reveal evidence of faulting recorded in surface sediments. This method enables determining the paleoearthquake history of a fault (including the timing of earthquake occurrences, co-seismic displacement, and elapsed time) and understanding its rupture behavior, which is of great significance in assessing its seismic risk (Wallace, 1981; Sun et al., 2015; Liang et al., 2021). The key problem in the study of the paleoearthquake is the selection of the excavation site, which determines the success of its paleoseismic research. The selection of the location is based on the following two factors: whether we

can determine the period of earthquake events and whether enough samples are available for dating the events (Deng et al., 2004; Ran et al., 2012).

In front of the fault triangle or fault scarp, the channel becomes wider, and the waterpower weakens when it flows out of the mountains. In addition, the materials carried by the river would be accumulated and develop the alluvial–diluvial fan in the piedmont. The loess and alluvial–diluvial sediments with clear stratification since the Late Pleistocene epoch can not only record multiple seismic rupture events but also preserve enough sediment for dating, thus making it a relatively ideal environment for the study of paleoearthquakes.

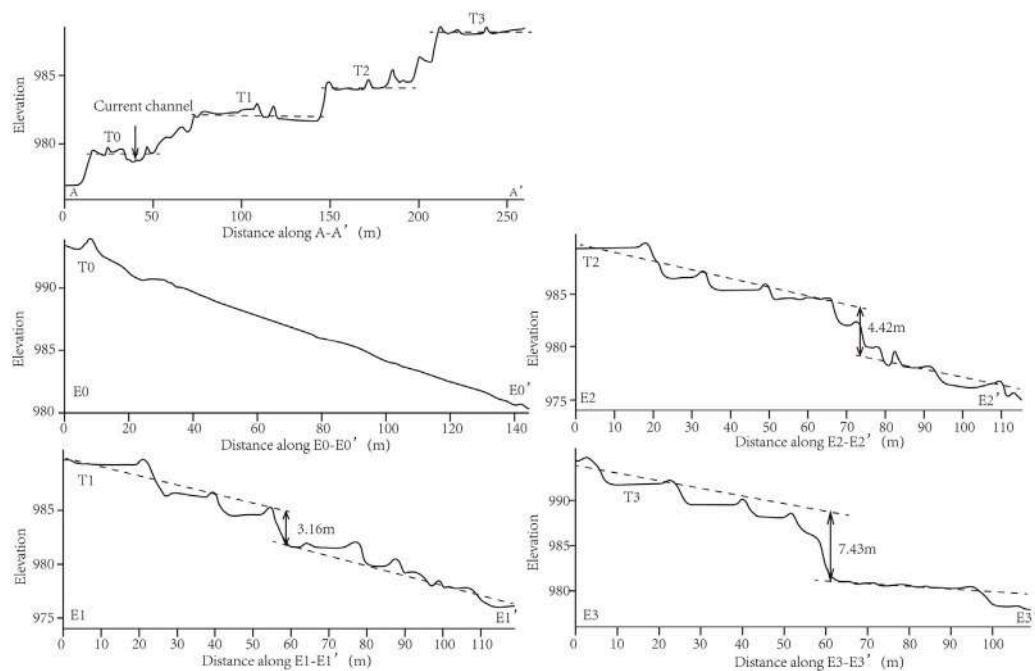
To confirm the location, geometry, and fault activity of the Liulengshan Fault, two trenches were chosen in Xujiabao and Luofengwa (for locations, see Figure 2). A single trench was excavated and described at each site, and strata were sampled for OSL dating. The details of the geomorphology, stratigraphy, and deformation of each site are described below.

Paleoseismic trenching

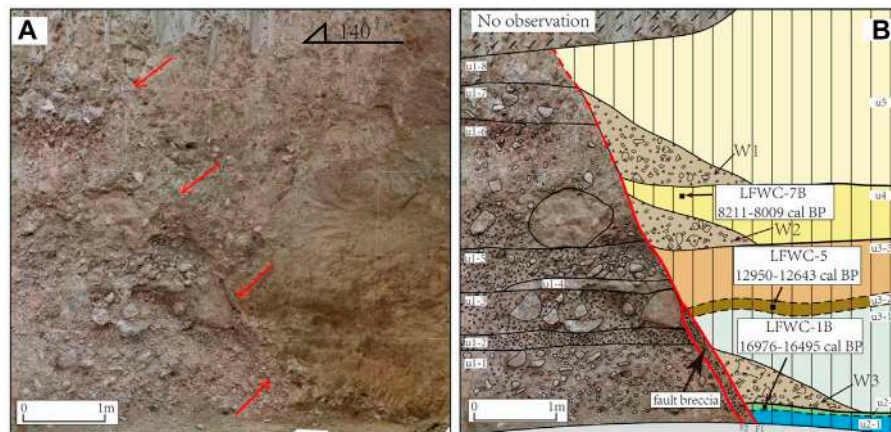
Xujiabao trench (39°54'59.65"N, 113°41'36.41" E)

The Xujiabao trench is located in the central section of the LLSF, 25 km to the southwest of the Yangyuan County (Figure 2). At Xujiabao, the fault forms a prominent fault scarp that ranges in height from 2.5 to 8.5 m (Figure 3). The sediments to the north side of the LLSF are predominately fluvial deposits and alluvium covered by loess, and the sediments to the south side of the LLSF are predominately bedrock and pediments covered by thin strath terraces of alluvium and loess. This area is heavily modified by anthropogenic terracing of agricultural fields. We used optically stimulated luminescence (OSL) (Rapp and Aitken, 1998; Rhodes, 2011) to determine the age of the displaced buried sediments in the trenches. The samples were dated by the single-aliquot regeneration (SAR) method at the State Key Laboratory of Earthquake Dynamics (SKLED), Institute of Geology, China Earthquake Administration.

The trench at Xujiabao was excavated in the T1 terrace. The profile is about 8 m high and 12 m wide. It should be noted that the steep slope with a height of about half a meter left on the northeast side of the top of the profile due to removal of soil for road construction is not a fault slope but an artificial one (Figure 4). Six distinct lithologic units were identified in the trench walls and are described in detail in Table 1. The footwall is composed primarily of gravel on the southwestern side of the fault, overlain by U6 (Figure 5). The presence of gravel at the footwall has allowed accumulation of at least two distinct sediment wedges in the hanging wall. We have recognized three fault planes in the trench: faults F1 and F2-1 cut the lower units and terminates at the base of U5, and

**FIGURE 8**

Topographic profiles showing river terraces and scarp heights. See Figure 7B for locations.

**FIGURE 9**

Photograph mosaic (A) and the interpreted map (B) of the Luofengwa profile. U denotes stratum unit, and black lines show the stratigraphic contacts between units or subunits. Red lines show the fault planes. Small black filled squares show the locations of radiocarbon samples, labeled with sample numbers and their ages (Table 4).

fault F2-2 cuts the higher units and terminates at the base of U6.

On the trench wall, we recognized three fault strands (Figure 5), F1, F2-1, and F2-2. F2-2 and F2-1 sole into a single fault at depth, F2, that dips to the northeast and forms

the main tectonic contact between the hanging wall, with mainly loess sediments, and the footwall, with mainly sand and gravel sediments. The fault is present from the top to the bottom of the profile and controls the boundary of the colluvial wedge as the secondary order-faults of F2-1 and F1 have different activity

TABLE 4 Radiocarbon ages and calibrated dates for the samples from the Luofengwa trench.

Sample	Lab no ^a	Radiocarbon age (a BP) ^b	Calibration years (cal BP, 2σ uncertainties) ^c	Analyzed material ^d	Unit sampled
LFWC-1B	Beta-480767	13,830 ± 50	16,976–16495	Organic sediment	u2-2
LFWC-5	Beta-480770	10,850 ± 80	12,950–12643	Organic sediment	u3-2
LFWC-7B	Beta-480771	7,320 ± 50	8,211–8,009	Organic sediment	u4

^aSamples were prepared and analyzed by the Beta Analytic Inc., United States.

^bUsing the accelerator mass spectrometry (AMS) method, the “conventional radiocarbon age” was calculated using the Libby half-life (5,568 years), which is corrected for total isotopic fraction and was used for calendar calibration where applicable. The age is rounded to the nearest 10 years and is reported as radiocarbon years before present (BP), “present” = AD 1950.

^cCalibration was calculated by the IntCal 13 database (Reimer et al., 2013), with two standard deviation uncertainties and a confidence level of 95%.

^dTypes of samples collected for ¹⁴C age dating.

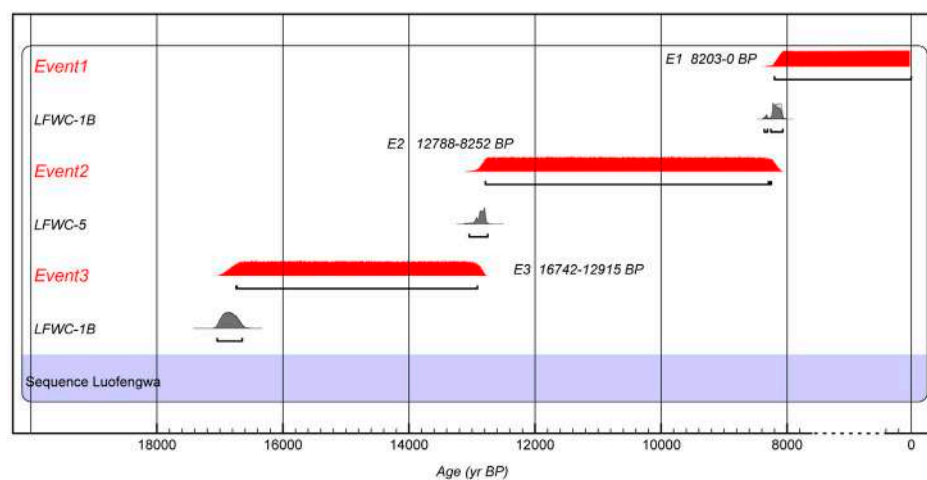


FIGURE 10

Results of OxCal analysis of radiocarbon dates from the Luofengwa trenching site. Discordant samples are not included. Unfilled areas are prior probability distributions, and solid curves are posterior distributions after OxCal analysis. Phases are summed probability for units with multiple radiocarbon dates.

stages and break different layers. The three faults are similar in geometry, dipping northeast. F1 and F2-1 are contemporaneous active faults, which fault u2 and u3 strata and colluvial wedge w1, with a vertical dislocation of about 1.2 m.

At least four paleoearthquake events can be identified in the profile, and we numbered E4–E1 for these events from old to new (Figure 6).

E4 event: after the deposition of u3-3, the main fault F2 (F2-2) was active, and all the strata below u3-3 were faulted. This event occurred between 41.1 ± 2.4 and 29.2 ± 1.2 ka, after which colluvial wedge W1 was accumulated and U4 was deposited.

E3 event: after the deposition of U4, the secondary faults F2-1 and F1 were simultaneously active and U4 and the colluvial wedge w1 accumulated in the E4 event were offset by this event and divided into w1-1, w1-2, and w1-3. Because the U4 stratum is mainly composed of a uniformly deposited loess silt layer, the colluvial wedge produced by this event is

difficult to recognize. This event also occurred between 41.1 ± 2.4 and 29.2 ± 1.2 ka.

E2 event: after the E3 event, the U4 stratum continues to deposit. The main fault F2 (F2-2) was active, and the colluvial wedge w2 was accumulated at the left of F2-2 after this event. Then, the U5 stratum was deposited. This event occurred at about or slightly before 21.2 ± 1.9 ka.

E1 event: because the colluvial wedge w2 and the bottom of U5 were offset, there was at least one event after E2, and the upper break point was not clear.

Luofengwa trench (39°59′02.17″N, 114°14′03.77″E)

Luofengwa village is located in the southern margin of the Yangyuan Basin, adjacent to the LLSF. The fault cuts along the

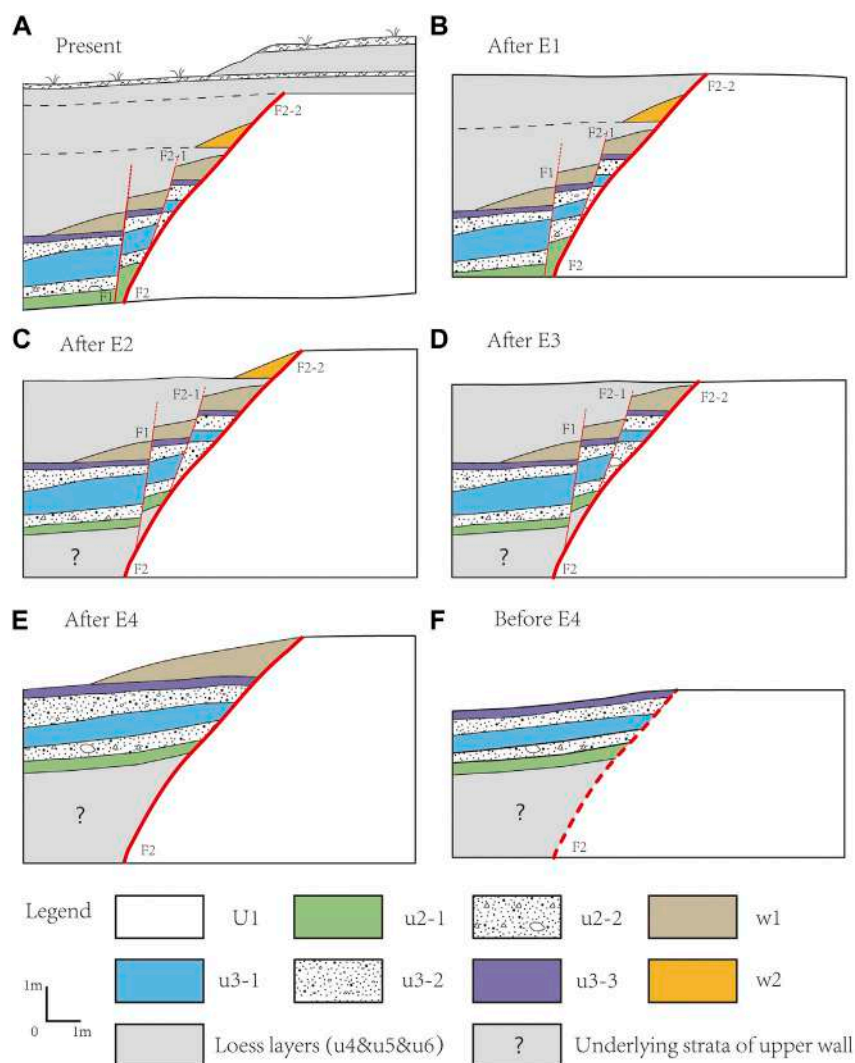


FIGURE 11

Possible restoration of the stratigraphy of the Xujiabao profile, showing the deformation associated with events E1 to E4. **(A)** Present-day situation; **(B)** colluvial wedge w2 and the bottom of U5 was offset, so there was at least one event after E2, and the upper break point was not clear; **(C)** E2 occurred after the deposition of U4 strata. The main fault F2 (F2-2) was active, and the colluvial wedge w2 was accumulated; **(D)** E3 occurred after the deposition of U4, the secondary fault F2-1 and F1 were simultaneously active, the U4 and the colluvial wedge w1 accumulated in the E4 event were offset by this event and divided into w1-1, w1-2, and w1-3; **(E)** E4 occurred after the deposition of u3-3, and the colluvial wedge W1 formed; and **(F)** possible original situation before E4.

back edge of a large alluvial fan south of Luofengwa village, cutting the Late Pleistocene to Holocene alluvial deposits within the basin, which strike 40–55° and dip ~75°NW (Deng et al., 1994; Xu et al., 1996).

About 1 km south of Luofengwa village, alluvial–diluvial fans are accumulated at the exit of the mountain and five terrace surfaces, namely, T0, T1, T2, T3, and T4 from low to high, were formed under the action of river erosion. The geomorphology of this section is well-preserved. Since the Late Pleistocene, the repetitive dislocation of faults has cut the geomorphic surface of different periods, and the height and vertical displacement of

fault scarps increase regularly with the periods of the geomorphic surface of faults. High-resolution measurement of the terrace profile shows (Figure 8) that terrace T0 maintains the original terrain slope, indicating that the fault has not been active since the terrace was formed. The latest faulting has dislocated T1 with a dislocation of 3.16 m. The dislocations of terrace T2 and T3 are 4.42 and 7.45 m, respectively.

The highest platform (T4) consists of loess capping sand and gravel. Near the fault, stream incision exposed an outcrop that reveals the lithology: the top is Malan loess, 3–5 m thick; the middle is gray–gray white gravel, silty sand, and gray–green clay

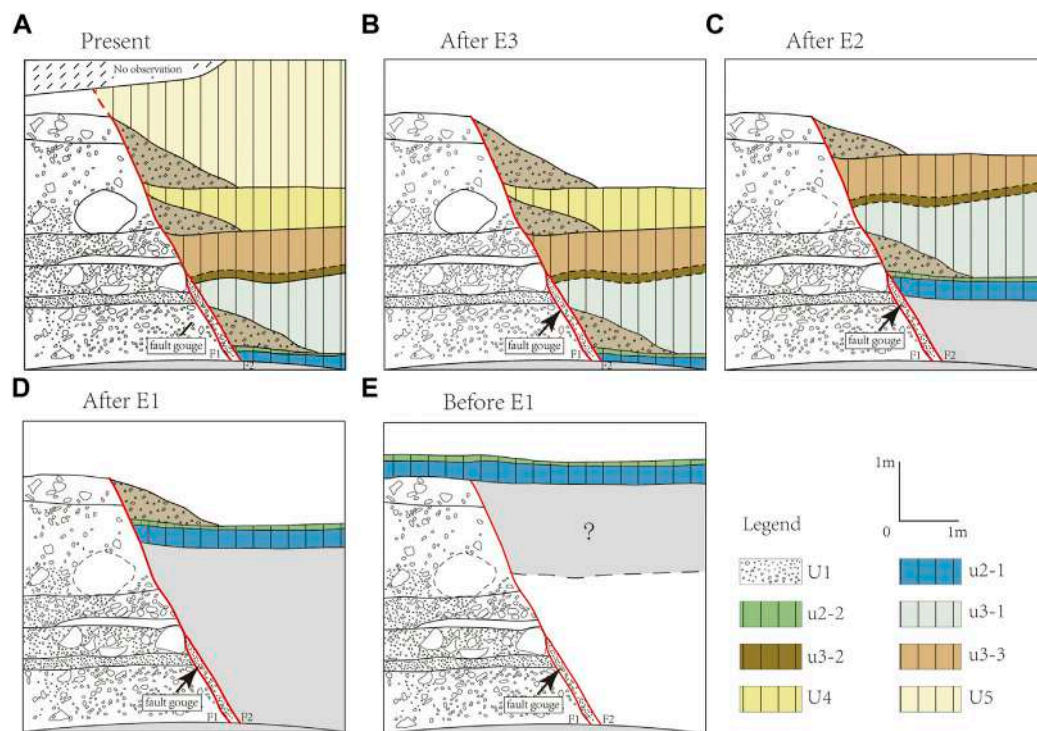


FIGURE 12

Possible restoration of the stratigraphy of the Luofengwa profile, showing the deformation associated with events E1 to E3. (A) Present-day situation; (B) E1 occurred after the deposition of u4, and colluvial wedge W1 formed; (C) E2 occurred after the deposition of u3-3, and colluvial wedge W2 formed; (D) E3 occurred after the deposition of u2-2, and colluvial wedge W3 formed; and (E) possible original situation before E3.

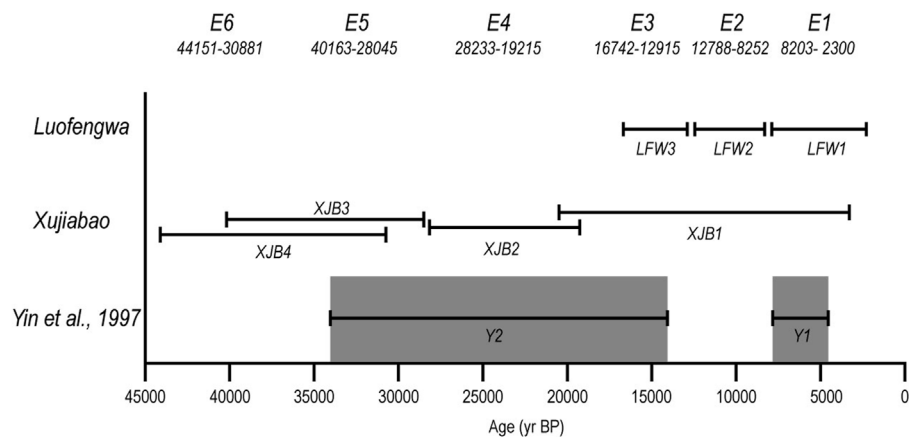


FIGURE 13

Late Pleistocene paleoearthquakes on the LLSF.

layer of the fluvio-lacustrine facies, which is about 12 m thick; and the bottom is metamorphic Proterozoic bedrock.

The excavated profile across the fault is about 5.5 m long and 5 m high. Figure 9 shows the overview photography and

interpreted map of the stratigraphy exposed on the profile. The sediments on the two sides of the fault are different. The footwall mainly consists of alluvial gravels, and the hanging wall is mainly composed of sand and clay layers. The stratigraphy is

mainly divided into five units, wherein U1/U2/U3 are divided into several subunits, and they are numbered in ascending order from the bottom to the top (Figure 9). W1/W2 and W3 are identified as the colluvial wedges.

The description of each unit is summarized in Table 3.

On the trench wall, fault F1 cuts the sediments from the top to the bottom of the trench (Figure 9). Fault F2 is only present at the bottom of the profile. As the master fault, F1 dips to the northeast and forms a sharp boundary between the stratified Unit 1 and the overlying Units 2–5 that are exposed on the downthrown side.

Among various indicators for recognizing paleoearthquakes in unconsolidated sediments in a normal fault setting, the best and clearest evidence for fault rupturing of the ground surface includes the morphology of colluvial sediments (Burbank and Anderson, 2012). Based on the relation of the fault and colluvial wedges, at least three paleoearthquake events can be identified in the profile, and we numbered E1–E3 for these events from old to new (Figures 2–9).

In the trench, three samples for AMS- ^{14}C dating have been collected and dated. Analyses were conducted at Beta Analytic Inc., United States, while detailed methods for sample preparation can be found at <http://www.radiocarbon.com>. Most of the result ages were calibrated using the INTCAL 04 database (Reimer, 2004). Calibrated radiocarbon ages are given in Table 4.

The E3 event occurred after the deposition of u2-2 and before u3-1, interpreting W3 as a colluvial wedge produced by the rupture. ^{14}C dating of the organic sediment from u2-2 and u3-2 gave the age of 16,976–16,495 and 12,950–12,643 years BP, respectively (Table 4). Therefore, the E3 event is constrained between 16,742–12,915 years BP, and we suggest that the age of the event is much closer to 12,915 years BP, as this was close to the ground surface prior to the deposition of the W3 wedge (Figures 9B,10; Table 4).

The E2 event occurred before U4 and after u3-3, again using the colluvial wedge of W2 as a key marker. U4 yields a ^{14}C age of 8,211–8,009 years BP (LFWC-7B; Table 4). Combining the age of u3-2, we constrained the E2 event to the period between 12,788 and 8,252 years BP, and we suggest that the age of the event is much closer to 8,252 years BP (Figures 9B,10; Table 4).

The E1 event is the youngest event in the trench and occurred before the deposition of u5 and after that of u4, using the colluvial wedge W1 that infilled a scarp produced by rupture along the fault. Accordingly, the E3 event that caused rupturing on this fault postdates the formation of u4, but a lack of datable material in the u5 means that the timing of this earthquake event cannot be accurately constrained. We are certain that the earthquake event occurred close to or slightly after ~8,203 years BP (Figures 9B,10; Table 4).

Discussion

Seismogenic capability of the Liulengshan Fault

Our paleoseismic analysis of the LLSF provides new evidence and more precise age constraints for the Late Pleistocene surface-faulting events. We determined four paleoseismic events in the Xujiabao trench, and five OSL samples with ages in correct stratigraphic order are used to constrain the events with very broad time ranges. On the Luofengwa trench wall, another three surface-rupturing events are correlated to colluvial wedges, and their timings are constrained by only three radiocarbon samples as organic materials were only found in the intervening sedimentary layers. Based on the lines of evidence, two restoration models of strata on trench walls are introduced to illustrate the seismic deformation and subsequent deposition in each trench (Figures 11, 12).

Yin et al. (1997) identified two events through trench excavations near our Luofengwa trench and determined their ages to be 14.2 ± 1.0 – 34.0 ± 2.2 ka and 4.6 ± 0.5 – 7.8 ± 0.6 ka based on IRSL dating (infrared stimulated luminescence). Among the events near Luofengwa, the age of the most recent paleoseismic event in the trench of Yin et al. (1997), Y1, overlaps with that of events LFW1 and XJB1, while the age of their older event overlaps with that of three paleoearthquakes XJB2, XJB3, and XJB4 in our trench. Yin et al. (1997) suggested that their latest event occurred around 6.7 ka. Therefore, we propose this event is the same paleoearthquake as our event E1. According to the overlaps of the ages of the events determined in the three trenches (Figure 13), we further determined six paleoearthquakes ruptured the LLSF with ages of 44,151–30,881, 40,163–28,045, 28,233–19,215, 16,742–12,915, 12,788–8,252, and 8,203–2,300 years BP.

The age of the latest paleoseismic event in the Xujiabao trench, event XJB1, is very wide (20,416–3,378 years BP) and coincides with the ages of all three paleoseismic events from Luofengwa, implying that they ruptured simultaneously during a single multi-segment rupture. Therefore, there are four possibilities for the faulting behavior of the LLSF: it either 1) ruptured each fault segment independently (independent 34 and 39 km ruptures), 2) ruptured between both trench sites (>50 km), 3) ruptured both two segments (34 + 39 km = 73 km), or 4) ruptured these two segments plus the additional two segments (130 km). Based on the empirical relationship between surface rupture length and moment magnitude for normal faults (Wells and Coppersmith, 1994; Blaser et al., 2010), these four rupture scenarios correspond to magnitudes M_w 6.9, 7.1, 7.3, and 7.7, respectively. Therefore, we suggest that the LLSF can generate an earthquake with a magnitude of 6.9–7.7.

Implications for seismic hazards

Due to the discontinuous stratigraphic deposition and poorly constrained ages of the determined paleoearthquakes, we cannot exactly tell the elapsed time of the most recent strong earthquakes on the LLSF. Datong, Shanxi Province, is the closest city to the fault and is about 30 km away. Datong was an important frontier county in ancient China and was built in the Warring States Period (about 300 BC), so it has a relatively complete historical record. Thus, information regarding damages caused by historic earthquakes was recorded very well since 300 BC in the near region. However, no strong earthquake was reported to happen on or near the LLSF in the Historical Strong Earthquake Catalog of China (China Earthquake Administration, 1999b). Therefore, we can estimate an elapsed time of >2,300 years since the latest strong earthquake. On the other hand, Deng et al. (1994) proposed a vertical slip rate of 0.4–0.8 mm/a through measuring and dating of several fault scarps, which were produced by multiple earthquake ruptures. According to the vertical slip rate and estimated elapsed time since the latest strong earthquake, the LLSF has accumulated an unreleased slip of 1.0–1.7 m, which is greater than the slip per event we observed in our trenches. That is to say, the LLSF has accumulated a seismic energy equivalent to moment magnitude of 6.8–7.0 according to the empirical relations between magnitude and slip for normal faults (Wells and Coppersmith, 1994). Therefore, we suggest there is a high seismic risk along the LLSF and in the nearby region.

Conclusion

Through two paleoseismological trenches at Xujiabao and Luofengwa, we revealed a ~42,000-year paleoearthquake history of the LLSF and confirmed that the fault has been active since the Late Quaternary and could produce surface-rupturing earthquakes. Based on the trench interpretations and the radiocarbon (C14) and optically stimulated luminescence (OSL) dating, we determined at least six surface-rupturing paleoearthquakes on the LLSF. The ages of these events, labeled E6–E1 from oldest to youngest, are limited to the following time ranges: 44,151–30881a, 40,163–28045a, 28,233–19215a, 16,742–12915a, 12,788–8252a, and 8203–2300a BP, respectively.

Compared with the previous results of Yin et al. (1997), we consider that different segments of the LLSF may rupture together at the same time. According to the empirical relationships between moment magnitude and rupture length, the best estimated magnitude is inferred to be in the range between Mw 6.9 and Mw 7.7. Considering the strong Late-Quaternary activity and a long earthquake elapsed time, we propose that the LLSF might have a high seismic hazard potential in the near future.

Data availability statement

The original contributions presented in the study are included in the article/Supplementary Material; further inquiries can be directed to the corresponding authors.

Author contributions

WS: conceptualization, methodology, and writing—original draft. HH: methodology, writing—review and editing, and funding acquisition. FS: conceptualization, field investigation, and writing—review and editing. HS: software and formal analysis. ZW: field investigation, and writing—review and editing. PS: field investigation and validation.

Funding

This work is funded by the State Key Laboratory of Earthquake Dynamics of China (LED2021A02) and Fundamental Scientific Research Project of the Institute of Geology, China Earthquake Administration (IGCEA2205).

Acknowledgments

We are especially grateful to Xibin Tan and Peng Guo as well as other colleagues for helpful suggestions. The authors sincerely appreciate the people who helped with the analysis of paleo-seismic trench results and processing of dating data. Thanks to Dr. Junjie Zou for field investigation. Thanks to Dr. Wei Gao and Dr. Peng Guo for their support in OxCal analysis of luminescence dates. Comments from three reviewers significantly improved the manuscript.

Conflict of interest

The author WS was employed by China Railway Engineering Design and Consulting Group Co., Ltd.

The remaining authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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