Active thrusting and folding in the Qilian Shan, and decoupling between upper crust and mantle in northeastern Tibet

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Fieldwork south of the city of Gaotai (Gansu province, China) shows that active shortening of surface sediments in the foothills of the Yumu Shan, a large fore-mountain of the Qilian Shan, at the northeastern edge of Tibet, involves both overthrusting and flexural-slip folding. North of this mountain, we found and mapped a prominent north-facing thrust scarp that offsets a Holocene fan slope gently (3.4°) to the north. Part of this scarp appears to be related to the M = 7.5, 180 A.D. earthquake that may have led to the demise of the Han Dynasty city of Luo Tuo Chen, in the Hexi corridor. A set of 10, 100-150 m long profiles measured across this scarp, 3.2 m high on the average, can be made to fit the diffusion-degraded morphology of a surface break related to the 180 A.D. event using a value of about 3.3 m²/10⁷ yr for the mass diffusivity k of conglomerates in this part of Gansu province. Smaller mountain-facing scarps on a terrace-capped foothill result from a bedding-slip component with active folding of underlying, steeply north-dipping, Pleistocene sandstone, and conglomerate beds. Holocene uplift rates along the Yumu Shan, which is only one of the Qilian Shan ranges, are estimated to be between 0.4 and 0.9 mm/yr, which implies that much of the mountain formed in the Quaternary. The periclinal structure of the Pleistocene envelope under which the Paleozoic core of the Yumu Shan plunges towards the west suggests that the whole 5000 m high mountain is a basement ramp anticline. Mountains striking parallel to the Yumu Shan, with similar structure and comparable or greater sizes north and south of the Hexi corridor probably also correspond to recent, crustal ramp anticlines. This implies that the wide, mountainous upper crustal wedge making the northeastern edge of the Tibet-Qinghai plateau is detached from the underlying lower crust and upper mantle.

1. Introduction

South of the Ala Shan and Gobi deserts, the NW–SE trending Qilian mountain range marks the northeastern edge of the Tibet–Qinghai highlands (Fig. 1, insert). Its highest peaks tower to more than 5500 m above the Hexi or Gansu corridor, whose average elevation is about 1500 m (Fig. 1a). The Qilian Shan is a Caledonian orogenic belt that has been reactivated by the collision between India and Asia [e.g., 1,2]. The present-day elevation of the range, and its seismicity, have been interpreted to result from the widespread post-Eocene crustal shortening caused by this collision in the interior of Asia [e.g., 3]. From the analysis of Landsat images [4], the northern front of the range has been inferred to be limited by large, south-dipping, active thrusts. Several earthquakes with magnitudes equal to or greater than 6 have occurred along the Hexi corridor during history [5,6], (Figs. 1a and b). Among these, the earthquake of 180 A.D., near Gaotai, appears to have been the most destructive in the last 2000 years. This event, whose magnitude is estimated to have been 7.5, and whose macroseismic epicentral location is roughly at 39°24'N, 99°30'E (Figs. 1a and b), [5], is reported to have led to the abandonment of the ancient city of Luo Tuo Chen (city of camels), a flourishing caravan halt along the silk road under the Han Dynasty (206 BC, 220 A.D.). The walled city lies on the northern piedmont of Yumu Shan, a high fore-mountain north of the Qilian Shan. The mountains provided an essential source of water, which was fed across the desert to the city by an impressive system of aqueduct canals (Figs. 2a and b). Bank collapse or
Fig. 1a. Map of recent and active faults and folds in the northeastern corner of the Tibet-Qinghai highlands. Insert shows location of this area in eastern Asia. Box indicates location of area displayed in Fig. 1b. Faults and folds shown are only those whose existence has been verified by us in the field, or on Landsat and Spot images. Elevation contours are from O.N.C. chart G8 [3°]. Earthquake epicenters are from catalog of strong shocks of China [5], and from catalog of strong earthquakes in Shanxi, Gansu, and Qinghai from 177 B.C. to 1980 A.D. [6]. AB is approximate location of section on front side of diagram in Fig. 10b.
slope changes along the canals during the earthquake may have contributed to the demise of the city [7]. We describe here the evidence we found in the field for active faulting in this area, and the relationship it bears to the 180 A.D. earthquake. We also discuss the nature of several mountain facing fault scarps whose origin and link with active tectonic features were hitherto unclear and examine the mechanism by which their formation can be related to the regional deformation. We finally assess the rate and style of recent tectonic processes along this part of the Qilian Shan and, by inference, along the northeastern edge of Tibet.

2. Evidence for active faulting south of Gaotai

South of Gaotai, the Yumu Shan forms an isolated, NW-SE trending mountain ridge, about 60 km long, 15 km wide and over 3000 m high, approximately 30 km north of the front of the main Qilian Shan range (Figs. 1b, 2 and 3a). It marks the southern boundary of the wide alluvial plain of the Hei He, largest river to flow out of Qinghai north into the Gobi desert. This plain is part of the Hexi corridor, which parallels the Qilian Shan to the north (Fig. 1), and is filled by Quaternary deposits with thicknesses locally in excess of 2500 m (e.g., B on Fig. 1b). The deposits are continental clastics, whose lithology displays an alternation reflecting glacial advances and retreats. This lithological alternation provides a basis for the first-order regional chronostratigraphy (e.g., 8). \(Q_4\) (Holocene: \(\approx 10^3\) yr B.P., to present) is chiefly composed of fluvial conglomerates with dark grey pebbles, usually capped by a thin loess layer. These conglomerates generally lie conformably upon \(Q_3\) (Wurm: \(\approx 10^5\) to \(\approx 10^4\) yr B.P.), which is made of coarser, glacial conglomerates with sand intercalations, generally above a loess layer. Regionally, \(Q_3\) lies unconformably upon \(Q_2\) (\(\approx 7 \times 10^5\) to \(\approx 10^5\) yr B.P.), which is divided into \(Q_{2,1}\) (glacial conglomerate), and \(Q_{2,2}\) (conglomerates with intercalations of sand and loess). \(Q_2\) is generally conformable upon \(Q_1\) (\(\approx 2 \times 10^6\) to \(\approx 1 \times 10^5\) yr B.P.), which is divided into \(Q_{1,1}\) (glacial conglomerate with large boulders that
reach as much as 20 cm in diameter), and $Q_{1,2}$ (fluviolacustrine sandstones with clay intercalations).

North of the steep front of the Yumu Shan, the broad, gently sloping sedimentary apron — or piedmont bajada — visible on Figs. 2a and 3a results from the coalescence of large fans fed by the mountain. All the fans have active braiding channels (dark on Figs. 2a and 3a). The surfaces of the fans are recent, still mostly depositional. They are not deeply incised by the active channels, except near the mountain front where steeply dipping conglomerates, which form the substratum of the fans, are exposed in narrow gullies. Within the fans, the upper layers of conglomerates visible in outcrop are mostly composed of small, grey pebbles. Hence, both in view of its morphology and of the regional Quaternary stratigraphy, the north Yumu Shan piedmont bajada appears to be for the most part of late Quaternary age ($Q_4$).

N100–140°E striking faults slice the Palaeozoic and Mesozoic rocks that make the Yumu Shan core (Figs. 1b, 2b, and 3b). Along much of the mountain front, which is particularly sharp on the Landsat image of Fig. 2a and on the photograph of Fig. 3a, such faults place these older rocks against or on top of the late Quaternary conglomerates of the piedmont bajada (Figs. 1b, 2b, and 3b). There is evidence of recent tectonic activity along most segments of the mountain-front faults (Fig. 3a and sites 1, 2, 3, 4, Fig 2b). On them we found a particularly young, north-facing scarp about 15 km SE of the city of Lian Tun Chen, and 2–3 km northwest of Wutong Quan monastery (site 2 in Fig. 2b). The trace of the scarp, which is visible on the air photograph of Fig. 4a, is outlined on Fig. 4b. Three late Quaternary alluvial fan surfaces ($a_1$, $a_2$, $a_3$) may be distinguished on the photograph (Fig. 4). The relative ages of these surfaces may be assessed from their relative elevations and the degree to which they have been incised by regressive erosion channels. The north-facing scarp appears to offset only the intermediate fan surface $a_2$, not the lower, active channel beds $a_3$. Small south-facing scarps (not drawn to scale on Fig. 4b), which do not connect in map view with the large north-facing scarp above, cut the upper fan surface $a_1$ at Wutong Quan.

2.1. Nature and age of the main north-facing scarp

In the field, the large scarp outlined on Fig. 4b is a prominent topographic step (Fig. 5a) that can be traced continuously for nearly 2 km across the alluvial fan surface $a_3$. As inferred from the air photograph, it cannot be followed across presently active braided channels towards the east or west (surface $a_2$, Fig. 4). In map view between these active channels, it has a variable strike (between N80 and N140°E), and a festooned geometry with two north bulging lobes and one south pointing cusp (Fig. 4b). Hence, this scarp appears to mark the surface trace of a south dipping thrust uplifting the southern, proximal part of the fan. On the fan surface $a_2$, two kinds of stream channels cross the scarp (Fig. 5a). Large channels (R), coming from far upstream and dissecting the whole fan, are entrenched by more than the total scarp height. By contrast, smaller rills (r) taking shape and dying out only about 15 m respectively upstream and downstream from the scarp incise the fan surface by a maximum of only about 30 cm at the scarp. We infer these rills to have formed as a result of progressive headward spring retreat since the last important throw increment on the scarp occurred.

With a high precision theodolite-distancemeter (Wild T 2000-DI 3000), and from a single base, we levelled ten, 100–150 m long profiles perpendicular to the thrust scarp (Figs. 6a and c, location on Fig. 4b). For a survey at that scale with this type of instrument, uncertainties on profile point positions are mainly due to the roughness of the topographic surface [9]: here, on the smooth, loess powdered fan surface $a_3$ (Fig. 5a) they were quite small, less than a few centimeters on the average. In the following, we thus take error bars on profile point elevations to be 5 cm. The average slope of the fan surface on either side of the scarp is 3.4°, and the average vertical offset of that surface, 3.2 m (Figs. 5a and 6a). The measured profiles are different from one another (Fig. 6a). Because the profiles are long, the regional slopes are well constrained, at least on one side of the scarp, despite the distance (a few tens of meters) at which the farfield fan slope is reached. Irregularities in the profiles appear to be due to three-dimensional effects, such as erosion by small gullies whose courses are not perpendicular to the scarp, or
Fig. 2. (a) Detail of Landsat image 144-33 (07/31/86), showing morphology of Yumu Shan and surroundings. Railway line north of Yumu Shan is clearly visible. Northern front of Yumu Shan is limited by particularly sharp fault. Fan baiada north of this fault appears more youthful than any other on the image. (b) Map of Holocene and Quaternary faults near Yumu Shan (box in Fig. 1b). Elevation contours are from Gaotai topographic map (scale: 1/100,000). Numbers refer to sites discussed in text. Box indicates location of Fig. 4a,b. A'B', location of cross section in Fig. 3b.
Fig. 3. (a) Oblique south-looking view of Yumu Shan mountain ridge from the air. Ancient Luo Tuo Chen aqueduct canals are clearly visible south of railway, across active bajada fans. Prominent Quaternary thrust is also clear along Yumu Shan front, particularly at site 4. (b) Interpretative section A'B' across Yumu Shan (location on Figs. 1b and 2b). Only schematic structure of basement is shown. Plausible geometry of main Quaternary faults and of deformed Jurassic and Upper Neogene unconformities is outlined (see discussion in text).

deposition in small fans at the foot of the scarp (e.g., profiles P3 and P5 in Figs. 6a and b). Scarp heights deduced from the profiles are within about 50 cm of 3 m, with the exception of those on profiles P4 and P6. In addition to yielding the greatest scarp height (4.4 m), profile P6 is that which displays the gentlest regional slope (2.6 °C). Besides, that slope is poorly constrained on the down-side of the scarp. Hence, it is possible that the scarp height is somewhat overestimated on this profile. The small height (1.5 m) of the scarp on profile P4, on the other hand, is reasonably well constrained by the measurements, and the regional slope is nearly the same as that in adjacent profiles. The height of the scarp here is about half that in adjacent profiles P3 and P5 (Figs. 6a and b). Moreover, the uplifted surface south of the scarp in P4 stands uniformly lower, by about 1.7 m, than that in P3 and P5 (Fig. 6b). This suggests that the conglomerate surface offset along P4 is younger than that offset along P3, P5 or the other profiles, and consequently that along much of its length the scarp may be the result of two seismic events. The complexity of the scarp shapes su-

Fig. 3. Aerial view (a) and map (b) of Pliocene-Pleistocene deposits and young fault scarps in area near Wutong Quan monastery (W), sites 2 and 3, Fig. 2b. Three alluvial surfaces of different ages may be distinguished: \( a_1, a_2, a_3 \), for older, intermediate and recent surface respectively. \( N_2 \) is Neogene. W, Wutong Quan monastery. Note that oldest surface is perched on top of Neogene and that large fault scarp ("Luo Tuo Chen" scarp) cuts only intermediate surface. Numbers 1–10 indicate locations of profiles transverse to large scarp shown in Fig. 6a. Note also small mountain facing scarps on \( a_1 \) near Wutong Quan. \( a \) and \( b \) refer to specific features discussed in text.

port this inference: on most profiles, two distinct breaks in the scarp slope are apparent (Fig. 6a). Note however that, with the exception of those on P5 and P9, such slope breaks are not symmetric about the point of steepest slope. Even with allowance for three-dimensional effects, it is un-
likely that differences in erosion or deposition modes over a short distance along the same scarp could account for such asymmetries. They are better ascribed to the along-strike changes in geometry and shifts in position that typify thrust scarps [e.g., 10]: such scarps often form irregular echelon arrays of pressure ridges, and may not break the surface at the same place during consecutive events.

Using the ten profiles, we have performed a mathematical analysis of the scarp morphology based on a diffusion model of scarp degradation (Figs. 6a and d). Analyses of this type are useful because they provide morphological dating in regions where degradation rates are known [11,12] or, conversely, yield information on such rates when the age of a scarp is known. Here we use a method somewhat different from that of the reduced scarp slope [e.g., 13,14]: assuming a two-dimensional geometry, the morphological evolution of an active scarp formed in unconsolidated fanglomerates as a result of an earthquake can be modelled as the convolution of the initial scarp shape by a gaussian degradation function whose variance \( \sigma^2 = 2\tau \) increases with time. This approach is valid whatever the degradation process (rainsplash, slopeswash, solutution, geoturbation or any other kind of surface creep proportional to topographic gradient) for scarps that are relatively small (a few meters high) and have no remnant freeface. If the mass diffusivity \( \kappa \) is constant, the diffusion equation leads to the simple result that the variance of the degradation function is proportional to the age of the scarp: \( \tau = \kappa t = \sigma^2 / 2 \). \( \tau \) is expressed in square meters and may be called erosion-age. It is thus possible to synthesize theoretical profiles as the superposition of a planar rectangular topography with one or several scarps of different erosion-ages and positions and to search for the best fit between such profiles and those measured.

In all the best fits of profiles at site 2, we have assumed the initial slope of the scarp to have been close to the angle of repose of unconsolidated fanglomerates (\( \approx 35^\circ \)). “One-event” best fits lead to scatter in erosion age values and variable confidence intervals for such values (Fig. 6d). In addition, except in the case of P4, there remains discrepancies between theoretical and measured profiles near the base or top of the scarp. Hence, as inferred from inspection of Figs. 6a and b, it seems inappropriate to ascribe the formation of the scarp to one seismic event only. The shapes of the measured profiles were therefore best fitted
Fig. 6. (a) Profiles transverse to Luo Tuo Chen thrust scarp at site 2 (locations on Fig. 4b), and at site 4 (box). Crosses are measured points. Thin parallel straight lines approximate fan surface slope far from scarp. Small numbers indicate values of far field fan surface slope (in degrees) and of fan surface vertical offsets (m). For profile P12, section of underlying geological structure and bedding attitudes (Sb), as observed in the field, are shown. (b) Three-dimensional block diagram showing detailed morphology of scarp between P3 and P5. Note that terrace surface upslope from scarp along P4 stands uniformly lower (by about 1.7 m) than that on either side (P3 and P5). (c) Best fits to scarp profiles assuming two surface offsetting events with nearly same vertical offset but different erosion-ages and positions. Vertical dashed lines indicate positions of initial surface offset for each event. Circled points in P3, which are interpreted to reflect irregularities due to three-dimensional effects (see discussion in text), have been discarded in the analysis. Parameters characterizing the best fitted diffusion degraded profiles are given in Table 1. (d) Vertical bars represent confidence intervals about minima defining best-fitting erosion-ages (small crosses) of youngest event. Size of confidence bars correspond to uncertainties on best fits minima that result from introducing 5 cm uncertainty on vertical position of all measured profile points. Note how poorly constrained erosion-ages are in one event analysis.
with theoretical profiles involving two "events" (Figs. 6c and d). These fits are satisfactory only if the two surface deforming "events" have roughly the same height but different positions and erosion-ages (Fig. 6c and Table 1). The fact that the positions of the surface traces of the two "events" do not coincide is essential to account for the variable asymmetries of the measured profiles. On most profiles the "two-events" fits single out a recent "event" with similar erosion-ages that average about 6 ± 3 m² and have confidence intervals narrower than those found with one-event fits (Fig. 6d and Table 1). The other, older "event" has variable, ill-defined erosion-ages between 70 and 370 m². The great scatter of the latter erosion-age values arises from the fact that the analysis of the residual topography corresponding to the older "event" becomes sensitive to rather small irregularities in the profiles. The clustering, about the value found for P4, of all the erosion ages corresponding to the recent "event" on the other profiles justifies the "two-event" analysis. Thus, we interpret the present-day morphological profiles measured at site 2 to represent a degraded 1.6 m high thrust earthquake scarp, with an erosion-age of about 6 ± 3 m², offsetting a similarly high but
TABLE 1

Confidence intervals and erosion ages for most recent scarp forming event (1) in site 2, and erosion ages of older event (2), deduced from Fig. 6a, are listed in first and second column, respectively; values of minimum standard deviation between measured and computed profiles are shown in third column; downslope shift of trace of second event relative to that of first is also given for each profile.

<table>
<thead>
<tr>
<th>Profile</th>
<th>Event 1 (m²)</th>
<th>Event 2 (m²)</th>
<th>Std. dev. (m)</th>
<th>Shift of event trace (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1 ≤ 6 ≤ 21</td>
<td>360</td>
<td>0.10</td>
<td>11.5</td>
</tr>
<tr>
<td>2</td>
<td>1 ≤ 3.6 ≤ 11</td>
<td>140</td>
<td>0.12</td>
<td>12</td>
</tr>
<tr>
<td>3</td>
<td>3 ≤ 6 ≤ 11</td>
<td>260</td>
<td>0.08</td>
<td>-12</td>
</tr>
<tr>
<td>4</td>
<td>3 ≤ 6 ≤ 18</td>
<td>-</td>
<td>0.07</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>0 ≤ 3 ≤ 10</td>
<td>110</td>
<td>0.12</td>
<td>0</td>
</tr>
<tr>
<td>6</td>
<td>1.5 ≤ 6 ≤ 21</td>
<td>370</td>
<td>0.10</td>
<td>11</td>
</tr>
<tr>
<td>7</td>
<td>1.5 ≤ 2.5 ≤ 15</td>
<td>70</td>
<td>0.06</td>
<td>-9.5</td>
</tr>
<tr>
<td>8</td>
<td>1.5 ≤ 2.5 ≤ 21</td>
<td>90</td>
<td>0.12</td>
<td>-6.5</td>
</tr>
<tr>
<td>9</td>
<td>0 ≤ 5 ≤ 22</td>
<td>80</td>
<td>0.12</td>
<td>0</td>
</tr>
<tr>
<td>10</td>
<td>5 ≤ 10.5 ≤ 24</td>
<td>150</td>
<td>0.06</td>
<td>12.5</td>
</tr>
</tbody>
</table>

more degraded step in the fan slope, corresponding to the much smoothed trace of an older, similar earthquake.

The thrust scarp at site 2 is the longest, highest and best preserved paleo-earthquake scarp in the region west of Gaotai. It is also the only scarp we found to cut a fan or terrace level immediately above — hence only just older than — the active, braided stream channels presently draining the northern slope of the Yumu Shan. It follows that this scarp should be related to particularly large and rather recent seismic events in that region. In this part of the Hexi corridor, the historical catalog of strong shocks of China [5], which spans about 3000 years, records only three earthquakes with $M \geq 6$ (Fig. 1). The epicentral location, about 30 km west of Wutong Quan, of the 1609 A.D., $M = 6.7$ earthquake suggests that this event is likely to have activated the Qilian Shan frontal thrust (Fig. 1b) rather than a more distant thrust north of Yumu Shan. The epicenter of the 756 A.D., $M = 7.0$ earthquake, on the other hand, has a location (about 10 km southwest of site 2) compatible with occurrence on a thrust dipping south under the western extremity of the Yumu Shan. If its magnitude is not grossly underestimated however, we infer this event to have been too small to create a nearly 2 m high surface scarp; thrust earthquakes of this size are rarely observed to break the surface. The third and oldest earthquake, which damaged in 180 A.D. not only the city of Luo Tuo Chen but probably also the archipelago canals stretching several tens of kilometers southeast of that city along the Yumu Shan piedmont [7], appears to have had the largest mesoseismic area of all three events and has been assigned the largest magnitude ($M = 7.5$). The mesoseismic area encompasses the zone where the scarp at site 2 was found. If as commonly observed for events with magnitudes greater than 7, the dislocation on the fault that slipped during the 180 A.D. "Luo Tuo Chen" earthquake reached the surface, the most recent part of the scarp at site 2 could thus well represent one remnant segment of the surface break of this earthquake. Following this latter inference, which we find the most plausible, we will refer to this scarp as the "Luo Tuo Chen" scarp (Fig. 4b), and assume that the 180 A.D. earthquake ruptured part of the mountain-front thrust responsible for the Quaternary uplift of the Yumu Shan above the Hei He plain. Taking an age of about 1.8 ka for the "Luo Tuo Chen" scarp, which we inferred to have an erosion-age of about $6 + 3$ m² from the diffusion profiles fits of Fig. 6, and assuming that the mass diffusivity $\kappa$ has been constant for the last 2000 years, yields for $\kappa$ a value of $3.3 + 1.7$ m²/10³ yr. Note that in making such an estimate we also implicitly assume that the scarp never had a fresh face and that its initial shape was that of a "fold-scarp" (Fig. 5b), with a maximum initial slope on the order of the angle of repose of unconsolidated conglomerates ($\approx 35^\circ$) soon after the earthquake. This assumption is justified by the fact that such fold-scars are common along thrust surface breaks and typically have maximum initial slopes of $40^\circ$ or less (e.g., Fig. 5c), [10]. The value found for $\kappa$ is in fair agreement with those obtained for scarp in similar unconsolidated conglomerates under comparable semi-arid climates [e.g., 12,15]. It is larger, however, than the value ($\kappa = 1.1$ m²/10³ yr) used by Zhang et al. [16] for the diffusion degraded profile offset by the relatively young scarp of the 1739 A.D., $M = 8$, Yinchen earthquake in Ningxia province, about 600 km to the east. This difference may be interpreted to be a consequence of the high elevation of the Hexi corridor and of greater rain and snowfall along the Qilian Shan front, which marks the north edge of the Tibet–Qinghai highlands, than in the more desertic lowlands west of the Ordos.
Fig. 7. (a) View of mountain facing scarp a (Fig. 4b) at site 3. Figure 2b. Regional slope of upper terrace a, is towards North (right-hand side of photograph). Flat plain in right-background is He Xi corridor. (b) Upper extremity of small north dipping reverse fault in trench b (Figure 4b) at site 3, Fig. 2b. North is towards left-hand side of photograph. Ballpen gives scale. Note reverse drag of pebbles in Q4 conglomerate layer.

Origin and age of the south facing scarps at Wutong Quan, and possible extension of the Luo Tuo Chen scarp along the Yumu Shan front

The evidence summarized above implies that the 3200 m high Yumu Shan continues to rise relative to its northern piedmont. The upper terrace surface a, just north of Wutong Quan monastery, however, is offset down to the south by several small parallel scarps that face the rising mountain (Fig. 2b, site 3, and Fig. 4b). The scarps strike between N95 and N105°E (Figs. 4b and 8, inset). One such scarp is shown on the photograph...
of Fig. 7a (site a, Fig. 4b). The late Quaternary fanglomerate layer forming the upper terrace dips gently north, about 6°, towards the Hei He plain, visible in the background. Trenches dug across some of the scarps show that the fanglomerates are generally back-tilted towards the mountain at the scarps [2,17]. One of the trenches (Fig. 7b, site b in Fig. 4b), reveals the upper extremity of a ≈40°N dipping shear zone in which reverse drag of the conglomeratic pebbles is clear. This change in pebble orientation, together with the attitude of the shear zone, implies that the scarps result from slip on multiple, parallel, north-dipping thrust-planes.

Observations made in the narrow stream canyons that incise the perched alluvial terrace a1 provide insight into the mechanism by which these scarps have formed. The late Quaternary fanglomerates of the terrace lie unconformably on a steeply north-dipping, conformable sequence comprising older, coarser Quaternary conglomerates at the top and pink Pliocene sandstones and mudstones at the base (Fig. 8). Bedding dips increase regularly upstream from about 70°N in the Quaternary conglomerates 500 m north of the monastery to almost 90° in the Pliocene sandstones under the monastery. Measured strikes of the steep beds in the Plioquaternary sequence coincide with the strikes of the mountain facing scarps on the terrace above (Fig. 8, inset). Furthermore, we found that bedding planes exposed in the stream canyons bore nearly dip-slip slickensides (pitch 86°W), compatible with thrust movement on these planes (Fig. 8, inset).

Folding of sediments at shallow levels in the crust often leads to bedding-plane slip between adjacent sedimentary layers, which tend to conserve their lengths and thicknesses. During the formation of such flexural-slip folds, slip of the outer over the inner layers is reverse updip in the plane orthogonal to the fold axis. Particularly well expressed slickensides on the inherently weaker stratification planes commonly indicate the slip direction [e.g., 18]. Thus, the Wutong Quan mountain-facing scarps are most simply interpreted as resulting from active bedding slip concurrent with incremental folding of the Plioquaternary sequence capped by the late Quaternary terrace. Bedding plane faults are also termed flexural-slip faults. Being a result of folding, such faults remain confined to the sedimentary cover and thus rarely reach depths sufficient for storing the elastic strain energy required to generate a great earthquake [19]. The Wutong Quan scarps thus probably reflect slip on secondary faults (in the sense of Lensen, [20]) on which movement is a consequence, not a cause, of the large earthquakes taking place along the northern piedmont of the
Yumu Shan. While common in ancient foldbelts, active flexural-slip folding has only recently been recognized to be an important process in zones of Quaternary shortening (e.g., [10,19,21,22]). To our knowledge, the example we document here is the first case of active flexural-slip folding reported in the mountain belts of China.

If the $M = 7.5$, 180 A.D. earthquake dislocation broke the surface along the north-facing Luo Tuo Chen scarp, then this $\approx 2$ km long scarp could represent only a short remnant of a surface break that might have extended many more kilometers towards the east or west. West of the narrow active channel $a_1$ (Fig. 4), the Luo Tuo Chen scarp appears to continue and merge with the surface trace of a south-dipping thrust fault displacing Neogene clastic onto the Quaternary bajada fans. We infer the 180 A.D. event to have ruptured this fault, along which a clear slope break is visible in most places, over a length of at least 15 km, as far west as Ke Shui Otu canyon (Figs. 1b and 2, site 1) whose section reveals a south-dipping thrust plane displacing superficial Quaternary layers in the manner shown in Fig. 3b (Fig. 9).

Whether or not the 180 A.D. break extended towards the east is less clear. Although the trace visible across fan $a_2$ projects eastwards about 1 km north of Wutong Quan monastery (Fig. 4b), no clear surface disruption of the coeval fan there is apparent on the air photo of Fig. 4a. Moreover, we found no evidence of a scarp in the field. We explored the possibility that the Luo Tuo Chen scarp might have veered south across the large active fan $a_2$, to join a fault separating the Neogene from the Quaternary at the base of Wutong Quan hill, a situation symmetrical to that seen in the west. But we found no clear trace of such a fault in the canyons on either side of the monastery. The evidence we have at hand thus suggests that the 180 A.D. dislocation did not reach the ground surface near Wutong Quan. Consequently, we infer that slip at depth on the eastward extension of the Luo Tuo Chen thrust could have been absorbed in this area by incremental folding of the surface sediments. If this inference were correct, the mountain-facing Wutong Quan scarps might have formed as a result of the 180 A.D. earthquake.

Yet farther towards the east, in the vicinity of site 4 (Figs. 2b and 3a), a large cumulative scarp runs along the base of the Yumu Shan foothills. South of this scarp, an uplifted late Quaternary conglomerate veneer, analogous to the upper terrace $a_1$ at Wutong Quan, caps a steeply dipping clastic sequence that includes pink, grey or red sandstone, siltstone and conglomerate beds which are thrust over late Quaternary gravels (Fig. 6a, box). This sequence resembles that found in the canyons under Wutong Quan monastery, but may be older than Pliocene. It is cut by small thrusts which do not extend in the overlying conglomerate. The beds, which are slightly overturned with steep southern dips at the scarp, swing to more moderate northern dips towards the south. This suggests that they belong to the faulted, recumbent limb of a north-vergent anticline. Two long profiles leveled across the scarp (P12 and P13, Fig. 6a, box) reveal that it offsets the surface of the late Quaternary conglomerates, which slopes 6 to 7° towards the north, by 10 and 20 m. But there is little trace of a steeper gradient attesting to rather recent reactivation, with a couple meters of slip, of the thrust at the base of the large scarp. Nor is there any clear offset, due to flexural slip in the underlying beds, of the uplifted Quaternary surface south of this scarp. Lastly, along the strike of the large scarp, a lower, more gently sloping alluvial terrace nearby is offset by a smaller, 5 m high scarp (P11, Fig. 6a, box). An offset of this size appears to be too large to represent the surface break of just one $M = 7.5$ event, such as the 180 A.D. Luo Tuo Chen earthquake. Were it the result of two events, each with a throw of 50% greater than 1.6 m, then such events might be those identified at site 2 and the lower terrace at site 4 might be coeval with fan $a_2$ at site 2. This would imply that the 180 A.D. surface break extended as far as site 4. The evidence we gathered is insufficient to prove this inference, although all the scarps at site 4 appear to be of Holocene age.

3. Uplift rates along the north edge of the Qilian Shan

The existence of isolated patches of late-Quaternary fans ($a_1$) perched on top of folded Pliocene sediments (Fig. 4) implies that recent uplift of the northern foothills of the Yumu Shan relative to the Hei He base level as a result
of NE directed overthrusting has been relatively fast. Bounds on the uplift rate may be derived from our observations. As discussed earlier in this paper, most of the north Yumu Shan piedmont bajada fans appear to be of Holocene age. Moreover, it is probable that several terrace levels should have formed at the foot of such high mountains since the more intense downwasting concurrent with the Holocene global warming started. The fan surface a2 at site 2 in particular, which is next to the presently active drainage channels, must be younger than early Holocene (Fig. 4). Hence, assuming the onset of the Holocene warming to have been between 12 and 8 ka.

Fig 9. (a) Thrust fault emplacing Pliocene sandstones and mudstones on late Quaternary gravels at Ke Shui Gou (Site 1, Figs. 2b and 10a). Note that Pliocene beds in hanging wall are parallel to thrust plane. For correspondence with section in (b), this is a mirror image of river-cut, right bank of Ke Shui Gou canyon (north to the right). (b) Section across folded Pliocene and Quaternary formations exposed along Ke Shui Gou canyon (Site 1, Figs. 2b and 10a). Dashed lines are inferred. Small box indicates location of photograph of Fig. 9a.
ago [e.g. 9], the total offset (3.2 m) of this surface along the Luo Tuo Chen scarp could not be older than about 8 ka. This yields a lower bound of about 0.4 mm/yr for the uplift rate at site 2. More likely if a, whose age is closest to that of the presently active drainage surface a, were only a few thousand years old, the uplift rate would be faster. A maximum rate of 0.9 mm/yr would correspond to the impending occurrence of an earthquake similar to the 180 A.D. event in the next few years, with a recurrence time of about 1800 years. Note that such a maximum rate of uplift falls short of that obtained by assuming that the average 19 m vertical offset of the terrace at site 4 (Fig. 6a) is of early Holocene age, i.e. about 10 ka old (1.9 mm/yr). Such a fast value, however, is within the range of those proposed by Zhang et al. [7], who inferred 1.5 to 2.5 mm/yr from the 3 to 5 m entrenchment of present stream channels into the banks of the 2000 years old Luo Tuo Chen aqueduct canals, assuming it to be due to tectonic uplift only. The discrepancy between maximum uplift rates at sites 2 and 4 may simply be related to the position of these sites along the Yumu Shan front (Figs. 2b, 3a and 10a): faster rates are to be expected at site 4, which faces the high, central segment of the mountain ridge than at site 2, which is close to the northwestern extremity of this ridge. It might also be related to the fact that surface folding, not faulting, absorbs more of the shortening and corresponding uplift at certain sites (e.g. 2) than at others (e.g., 4).

Active crustal shortening across the northeastern edge of Tibet appears to be distributed over a width of several hundred kilometers (Fig. 1a), [4]. The Yumu Shan, for instance, which is located 30 km north of the northermost range of the Qilian Shan, is only one of many comparable mountains between the Tibet–Qinghai highlands and the Ala Shan–Gobi platform. Hence, uplift of the Yumu Shan at a rate on the order of 1 mm/yr, within the bounds discussed above, probably absorbs only a fraction of the convergence between Tibet and this platform. Nevertheless, at such a rate and with allowance for several hundred meters of erosion of its top, the mountain could have acquired its present elevation above the Hei He plain in about 2 Ma. This supports the qualitative inference, made earlier in this paper on the basis of morphological evidence alone, that the Yumu Shan is a particularly young mountain: much of it probably formed during the Quaternary.

4. Origin of the Yumu Shan and similar mountains in Gansu and northern Qinghai provinces: implication for the deep structure of northeastern Tibet

As the elevation of the Yumu Shan gradually decreases towards the northwest, the Paleozoic–lower Mesozoic core of the mountain progressively plunges under the Pliocene and Quaternary (Figs. 1b, 2a, 3a and 10a). The structure of this Pliocenary envelope sheds light on the structure and origin of that mountain, and by analogy on those of other mountains of similar trend, age, and size north and south of the Hexi corridor. A good section across the Pliocenean west of the Yumu Shan is exposed at Ke Shui Gou (Site 1, Figs. 2b and 10a). There, the Yumu Shan frontal thrust emplaces pink Pliocene mudstones and sandstones, which include rare intercalations of fluvial conglomerates with small, well rounded pebbles, on top of late Quaternary fanglomerates. The active frontal thrust dips steeply, about 70° towards the south, parallel to the slightly overturned Pliocene beds (Fig. 9a and b). Near this frontal thrust, on a smaller thrust guided by a bedding plane, we measured slickensides with a pitch of 80°W (Fig. 9b). Upstream along the canyon, the attitude of these beds progressively swings from south-dipping to vertical, then north-dipping, horizontal and finally south-dipping again, albeit more gently, thus forming a simple anticline overturned towards the north. Where nearly vertical, the Pliocene beds are offset by a couple of small south-dipping thrust planes, and minor conjugate normal faults are observed at the apex of the anticline, where the beds are nearly horizontal (Fig. 9b). In the southern limb of the anticline, the Pliocene beds are capped in slight angular unconformity by coarse Quaternary conglomerates with gentler dips. The shape of this anticline (Fig. 9b) is that of a fault propagation fold [23–25], which implies that it results from ramping of the active Yumu Shan thrust to the surface from a decollement at depth. Near the surface, this upward propagating thrust appears to have splayed and to have been guided at places by the steeply south-dipping Pliocene beds (Figs. 9a and b).
Irrespective of complexities related to pre-Cenozoic tectonic strain in the deeper rocks of the Yumu Shan core, the simple anticlinal shape (Fig. 9b) of the fluvial, hence initially flat, Pliocene deposits forming the periclinal envelope of this core (Fig. 10a) suggests that the mountain as a whole, which is 15 km at its widest, over 60 km long and 3200 m high, is a basement ramp anticline of Quaternary age (Fig. 3b). By analogy with the mechanics of sedimentary cover overthrusting in foreland foldbelts [e.g., 26, 27], we infer the shallow-dipping decollement into which the Yumu Shan ramp thrust probably roots at depth (Fig. 3b) to extend southwestwards under the Qilian Shan (Figs. 1a and 10). In addition, because most of the elongated, NW–SE trending mountain
ridges on either side of the Hexi corridor appear to closely resemble the Yunnan Shan in their overall shape, structure and age, we infer them — including the great parallel ranges of the Tanghe Nan Shan, Gule Nan Shan, Ta Xue Shan, and Qilian Shan itself (Figs. 1 and 10a), [9] — to correspond also to recent, crustal ramp anticlines (Figs. 1a and 10).

The inference that these mountain ranges are crustal anticlines, together with the great width (between 300 and 400 km) of the zone across which they have risen (Fig. 1a), and the gradual northeastwards decrease in elevation across the northeastern border of the Tibet-Qinghai plateau may be taken to imply that much of the upper-crust of this region forms a northeastwards tapering wedge decoupled from the underlying lowermost-crust and mantle lithosphere along a large lower to mid-crustal decollement dipping at a shallow angle to the southwest (Fig. 10b) (see also [28]). Such a deformation process would thus be akin to that through which accretionary wedges grow at subduction zones [29], but would occur on a larger scale, involving much of the continental crust over a much broader surface. That such large scale decoupling between upper-crust and mantle-lithosphere might typify the recent tectonics of northeastern Tibet has far-reaching consequences. It implies that broadly distributed thrust faulting and earthquakes in the upper continental crust need not reflect similarly diffuse strain in the underlying mantle lithosphere which may instead founder elsewhere into the asthenosphere, a mechanism appropriately termed continental subduction [e.g., 20, 21]. In northern Tibet, the most plausible place for such continental subduction to occur appears to be under the Kun Lun range, where a discontinuous string of acidic Quaternary volcanoes extends for about 1000 km, from 81° to 93°E [4, 32], and where Neogene anatectic leucogranites analogous to those observed along the Himalayas — under which the Indian lithosphere now subducts [33] — have recently been discovered [34]. Were the inference of mantle subduction under the Kun Lun shown to be correct, using for instance the powerful visualization techniques of seismic tomography, then large scale crust–mantle decoupling might prove to be a better way to account for the diffuse nature of crustal shortening in continental mountains and play a more important role in thickening continental interiors than bulk shortening of the lithosphere [e.g., 35, 36].

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