Incision rate of the Yellow River in Northeastern Tibet constrained by ¹⁰Be and ²⁶Al cosmogenic isotope dating of fluvial terraces: implications for catchment evolution and plateau building

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Abstract: Unlike other large rivers flowing out of Tibet, the Yellow River escapes from the plateau towards the NE crossing no less than five NW–SE striking, actively growing ranges and intervening basins. Thick Plio-Quaternary deposits and fluvial terraces testify to a phase of aggradation and sediment infill up to the average surface elevation (3200-3250 m a.s.l.) of the Gonghe, Guide and Qinghai Lake basins. A set of seven main terraces across the Gonghe Basin suggests progressive down-cutting of the Yellow River carving the 500 m deep Longyang gorge at the basin exit. ¹⁰Be and ²⁶Al concentrations in quartz of surface and sub-surface samples of four terraces constrain the timing of incision by determining the burial age of the deposit and the ²⁶Al/¹⁰Be ratio allows us to constrain the onset of the ongoing phase of incision to 120-250 ka. These ages suggest long-term incision rates between 2-6 mm a⁻¹. Together with the present morphology of the Yellow River terraces across the Gonghe winthe the present imply rapid river catchment evolution and interaction between river dynamics, tectonic and climate in northeastern Tibet.

How and when the Tibetan Plateau reached its present altitude and shape are important to investigate as the topography influences atmospheric circulations, thus the climate, but also because it may be used to constrain models of Tibetan Plateau evolution (e.g. Tapponnier et al. 1982; England & Houseman 1986: Peltzer & Tapponnier 1988; Houseman & England 1993; Royden et al. 1997; Fluteau et al. 1999; Tapponnier et al. 2001; Liu-Zeng et al. 2008). In the stepped growth model of Tibet of Tapponnier et al. (2001), northeastern Tibet is the youngest part of the plateau that formed from Pliocene time to present (Meyer et al. 1998; Métivier et al. 1998). The region is characterized by narrow actively growing ranges (e.g. Tapponnier et al. 1990; Meyer et al. 1998; Van der Woerd *et al.* 2001) that separate flat rapidly filling closed sedimentary basins (Meyer *et al.* 1998; Métivier *et al.* 1998; Van der Woerd 1998). Such process may have been active in the central part of the Tibetan Plateau, leading to the formation of a high topography at *c.* 5000 m a.s.l. From this point of view, the northeastern Tibetan region is a key place to unravel processes of the formation of the Tibetan Plateau. Understanding the different roles of tectonics, climate and the evolving drainage is thus important to constrain the formation of the present topography of Tibet (e.g. Clark *et al.* 2004; Liu-Zeng *et al.* 2008).

The Yellow River, together with the Jinsha, Lancang (Mekong), and Nu (Salween) rivers, drains the Tibetan Plateau on its eastern rim (Fig. 1). While

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Fig. 1. Seismotectonic map of the upper reaches of the Yellow River across northeastern Tibet. Fault modified from Van der Woerd (1998), Meyer *et al.* (1998), Tapponnier *et al.* (2001). Seismicity from Iris catalogue, focal mechanisms from Harvard. Gonghe 1990 M 7 earthquake isoseismal from Chen *et al.* (1996). Topographical background is contours extracted from SRTM DEM with a 30 s precision.

its source is located on the high summer monsoon influenced central plateau, like the other large rivers, it does not escape the plateau to the SE but is captured to the north after a hairpin loop in the Zoige (Roergai) basin (Fig. 1; e.g. Harkins et al. 2007). As it flows back to the NW, north of the Anyemagen Shan, it enters regions of lesser precipitation that barely feed the river. This particularity makes the river sensitive to any north-south shift of the northern monsoon limit, which depends on climate variations (e.g. Benn & Owen 1998; Joussaume 1999). Both the fact that the river crosses a relatively dry northeastern Tibet allowing long-term preservation of the terraces and its high sensitivity to climate change probably in part explains the long sequence of fluvial terraces found all along the river from the Anyemagen to the Ordos, in northern China.

The terraces of the Yellow River, near Gonghe (Fig. 2) or near Lanzhou (e.g. Li 1991; Wang et al. 2010), have been commonly interpreted both to testify the first occurrence of the Yellow River in NE Tibet and as a record of regional plateau surface uplift (e.g. Li 1991; Li et al. 1997; Harkins et al. 2007). In this study, we will show that the terraces instead more likely indicate rapid incision of the river after a phase of damming and deposition of the Yellow River. In fact, probably better than anywhere else in Tibet, the terraces of the Yellow River across the Gonghe basin illustrate a process of plateau building that involves damming of the drainage by active range growth and rapid filling of closed sedimentary basins (e.g. Meyer et al. 1998; Métivier et al. 1998).

In this paper, we describe the terraces of the Yellow River in the Gonghe basin. We present the ¹⁰Be and ²⁶Al exposure ages obtained on four different terrace levels that allow the constraint of both the incision history of the Yellow River across the Gonghe basin and the uplift rate of the surrounding ranges. We then discuss the implications on the basin history, and on the local and regional tectonics.

Geological and geomorphological setting

The Yellow River, basins and ranges

From the Zoige basin, east of Anyemaqen and south of the Kunlun fault to the Gobi platform, over a length of 500 km (or river length of >1000 km), the elevation drop of the Yellow River reaches c. 2500 m as it crosses several sedimentary basins (Tongde, Gonghe, Guide, Xunhua, Linxia), all separated by actively growing ranges (Heka Shan, Waligong Shan, Kongma Shan, Laji Shan) (Figs 1–3). Along this stretch, the river crosses no less than 3 active strike–slip fault systems and

c. six active thrusts implying interactions of river dynamics and active tectonics. Across both the Kunlun and Haiyuan left-lateral strike-slip faults (Fig. 1), the Yellow River is deflected 80-90 km left-laterally (e.g. Gaudemer et al. 1989, 1995; Van der Woerd et al. 2002). Right-lateral movement along the Haiyen fault that connects with the more easterly-trending Rivue-Laii-Kongma Shan thrusts contribute to the range growth along the northeastern border of the Guide basin. The Tongde basin is bounded to the north by the south-vergent Heka Shan (Fig. 1; Harkins et al. 2007). The Gonghe basin is limited to the north and south by active thrusts. In the south, thrusts are north-vergent in the west and south-vergent in the east, and their activity is attested by the magnitude Mw 6.9 1990 seismic event (Chen et al. 1996). In the north, the Oinghainan Shan thrusts are south-vergent to the west and are attested by folded and uplifted alluvial fans and terraces (Van der Woerd 1998) and northvergent to the east (Waligong Shan). Intensely faulted and folded Neogene series in front of the Ryue-Laji Shan (Fang et al. 2005) together with detrital zircon ages and their provenance from the Laji Shan (Lease et al. 2007) imply major shortening phases during the Plio-Quaternary along the northern Guide basin. Additional but less important thrusting also occurs along the western Linxia basin (Fang et al. 2003; Zheng et al. 2003), across the Lanzhou thrusts (Wang et al. 2010) and across the Tianjing-Mibo Shan north of the Haiyuan fault (Gaudemer et al. 1995). Most of the elevation change of the Yellow River occurs between Zoige and Linxia and the steepest river gradients are found in the gorges across the Waligong and Kongma Shan (Fig. 3).

At the northeastern exit of the Gonghe basin, the Yellow River crosses the Waligong mountains carving the 500 m deep Longyang Gorge (Longyangxia in Chinese) (Figs 2 & 3). Since 1986, an artificial dam has been built at the entrance of the Longyang gorges forming the c. 100 m deep Longyangxia Reservoir (water level at 2568 m in October 2004). The NW-SE-trending Waligong Shan is composed of two massifs, the western and eastern Waligong, each one bounded to the east by active west- to SW-dipping thrusts (Fig. 2). The western Waligong thrust juxtaposes the Waligong granite on top of red sandstones of Late Miocene-Pliocene age (Qinghai Geological Bureau 1988; Pan 1994; Fig. 4e), similarly to the thrust of the eastern Waligong at the Guide basin western margin.

The most striking evidence for ongoing uplift is the impressive Longyang gorges carved by the Yellow River as it crosses the Waligong range (Fig. 4a, b). The slope of the Yellow River increases dramatically at the entrance of the Longyang gorges to $c. 0.5^{\circ}$ (Fig. 3). The height of the bedrock cliff



Fig. 2. (a) SPOT satellite image mosaic of eastern Gonghe basin. White rectangle is Figure 2b. (b) CORONA image showing the northern part of the basin aross Yellow River terrace flights, Waligong Shan and Longyang gorge.

reaches c. 300–500 m in the gorge (Fig. 4). Other evidence for active tectonic uplift are given by the abraded bedrock abandoned above the gorge. At present, the bedrock slopes towards the west as expected if it were back-tilted to the west due to uplift of the Waligong on top of a west-dipping thrust (Fig. 5).

The Yellow River terraces across the Gonghe Basin

The Gonghe Basin is a NW–SE-trending 250×50 km wide sedimentary basin (Figs 1 & 2). It is bounded to the north by the actively growing Qinghainan Shan and to the south by the



Fig. 2. Continued. (c) Digital elevation model (SRTM, 90 m, GeomapApp) of most parts of the Gonghe, Guide and Qinghai Lake basins. Yellow River terraces clearly show up in central part of figure as wide fan shape stairs. Inset is elevation data histogram of same area. The tight peak of elevation around 3200 m a.s.l. is the wide plateau surface corresponding to surface elevation of the Gonghe and Qinghai Lake basins, as well as part of the Guide and Tongde basins.

Ngola Shan (or Gonghenan Shan) and the Heka Shan. It is filled with up to 1200 m of Quaternary sediments (Qinghai Geological Bureau 1988; Métivier et al. 1998). Semi-arid climatic conditions prevail over most of the basin (annual precipitation average 250-350 mm; Domroes & Peng 1988) and sand dune fields occupy the eastern part of it (Figs 2 & 4c). The basin is almost entirely internally drained except in its eastern part where it is crossed by the Yellow River. In contrast with the Qinghai Lake basin located just north of the Qinghainan Shan (Fig. 1) and occupied by the largest freshwater lake of China (4500 sq. m and about 27 m deep; Qin 1992), only small lakes are visible across the Gonghe basin. Evidence for higher lake levels can be seen in abandoned lake shore lines or lake deposits (e.g. around Dalian lake or Chaka salt lake; Fig. 6; Van der Woerd 1998). Despite their proximity, both basins have slightly different present climatic regimes characterized by dryer conditions in Gonghe (Domroes & Peng 1988).

As it crosses the Gonghe basin and incises through the sediment pile (Fig. 4d), the Yellow

River has abandoned a set of terraces that form huge fan shaped stairs across the entire 50 km width of the basin (Fig. 2). The freshness of the terraced landscape (Fig. 4c) together with the fact that regressive erosion from the Yellow River gorge itself has not yet captured the central drainage of the Gonghe basin indicate that the Yellow River incision occurred recently as suggested by several authors (e.g. Li 1991; Métivier et al. 1998; Van der Woerd 1998; Zhang et al. 2004). The large, up to 10 km wide and 40-50 km long, terraces across the basin are mostly flat with slopes toward the north comparable to the present Yellow River slope $(0.15-0.3^{\circ})$ in the same area (Fig. 3a, c). From the present river bed to the top we numbered the main and largest terrace levels T0-T7 (Figs 2 & 3). We designate the highest and largest terrace T7 knowing that it also corresponds to the regional surface level of basins around 3200-3250 m a.s.l. (Figs 2 & 3). This surface can be easily recognized on satellite images or on digital elevation models (SRTM DEM) as it forms flat areas that merge smoothly with the piedmont bajadas of the surrounding ranges (Figs 2 & 7a). A similar surface



Fig. 2. *Continued.* (d) Geomorphological map of the Yellow River terraces in the eastern Gonghe basin. White filled circles are sampling site on terraces T7 and T5. (e) Topographic cross-section of terraces following black line of Figure 2d (projection N85°E).



Fig. 3. (a) Slope variations along the Yellow River. River slope reaches its maximum values $(>0.5^{\circ})$ across the Longyang and Kongma gorges. (b) Profile of Yellow River bed from the southern Anyemaqen to Xunhua Basin. Red lines are active faults at basin margins (strike-slip, normal or reverse). Vertical scale is exaggerated 100 times. Profiles were extracted from SRTM DEM. River profile across Longyangxia Reservoir from Chinese 1:50 000 topographic maps. Position and terrace ages upstream from Gonghe from Harkins *et al.* (2007). (c) Enlarged Yellow River profile between Tongde and Xunhua, with projection of the Gonghe terraces (relative heights from Fig. 2), upper T7 regional top fill, and samples locations. Height of terraces in Guide basin from Fang *et al.* (2005). Colours of terrace threads in different basins are arbitrary.

level correlates with the top of the sediment fill in the Guide and Tongde basin (Fothergill & Ma 1999; Fang *et al.* 2005; Harkins *et al.* 2007). Interestingly, this level is also the highest palaeo-lake stand of the Qinghai lake basin to the north (c. 3250 m; e.g. Lister et al. 1991; Fang et al. 2005; Rhode et al. 2007, 2010; Liu et al. 2010).

T7 is clearly recognized as a fluvial terrace deposit at the entrance of the Gonghe basin north of Heka Shan (Figs 2 & 7b). T7 is now covered by



Fig. 4. (a) Field photo of the Longyang gorges, view towards the east. (b) Steep gorge cliff on right bank of river incising flat abraded bedrock surface, view towards the south. (c) Terrace T7/T6 riser inside Gonghe basin (see Fig. 2), view towards the north. Riser forms a wind shelter where dunes accumulate (barkanes). (d) Yellow River gorge in the south of the Gonghe basin, view towards north. River can be seen cutting trough the sediment fill and northern bedrock sliver of Heka Shan (e.g. Métivier *et al.* 1998). Flat incised surface is level T7. (e) View to west at the steep thrust contact between the Mesozoic granodiorite of the western Waligong and the Tertiary red sandstones in the Yellow River gorge.



Fig. 5. (a) Panoramic view of Waligong gorge. (b) View to west, towards Longyangxia Reservoir. In front, rim of bedrock strath can be seen. Note westward sloping flat top surface above the gorge. (c) View to east of stepped slope of western Waligong near the gorge that corresponds to the progressive entrenching and narrowing of the Yellow River bed. (d) View to west of the terraces in southwestern flank of Waligong Shan.



Fig. 6. Google Earth view of the Gonghe and Qinghai basins with present lakes or lake remnants (salt flats, abandoned shore lines) highlighted.

a 2 m thick layer of aeolian loess overlying a 15 cm thick reddish palaeosol (B-horizon; Fig. 8a, b; Pan 1994). The abandonment of T7 marks the onset of incision of the Yellow River in the Gonghe basin. The other major terrace levels (T6–T1) are also composed of fluvial pebbles and gravels and are devoid or only covered by a thin loess cover across the Gonghe basin. Clear steep risers up to several tens of metres high separate the main terrace levels (Figs 2d & 4c). In several places, as can be clearly seen on satellite images (Fig. 2), these risers disappear under wind-blown sand dune fields because of favourable accumulation at the north–south-trending risers which are almost perpendicular to the main NW blowing winds.

For a long time the Yellow River terraces have been recognized as geomorphic markers of the active tectonics of northeastern Tibet during the Quaternary (e.g. Li 1991; Li et al. 1997; Lu et al. 2004; Sun 2005). From Lanzhou to Gonghe, their Early Pleistocene-Late Pleistocene ages (e.g. Li et al. 1997) have been interpreted to represent phases of regional plateau uplift. In contrast, Harkins et al. (2007) have recently proposed, from a geomorphological analysis of the upper Yellow River gorge and its tributaries between Tongde and Zoige basins that the Yellow River propagated upstream due to headward incision. From OSL and C-14 ages of low terraces 10-140 m above the present river bed in Tongde and upstream that range from 9-140 ka, (Fig. 3), they infer an onset of incision from the top of the basin fill in Tongde at about 400–500 ka (Craddock *et al.* 2010). In more details, their results imply the upward propagation of a wave of incision starting around 500 ka in Tongde with incision rates of *c*. 2.5 mm a⁻¹ that slowed down as incision progressed to less than 1 mm a⁻¹ at present (Harkins *et al.* 2007; Fig. 3).

The terraces in Gonghe have not been directly dated so far. Thermoluminescence (TL) dating of loess covering seven terraces in Linxia give ages that range from 10-160 ka (Li et al. 1997) and due to the similar setting of terraces in Gonghe, the oldest of these ages has been attributed to the highest terrace in Gonghe (Li 1991; Li et al. 1997; Fothergill & Ma 1999; Zhang et al. 2004). However, from palaeomagnetostratigraphical studies in Guide and Linxia, it has been argued that the highest terrace in Guide postdates closely the shallowest dated sediments at 1.8 Ma (Fang et al. 2005), and that the Yellow River started incising in Gonghe at c. 1.1 Ma (Yan et al. 2004), in agreement with the inference of a post 500 ka incision of the Yellow River in Tongde and the upward propagation of a wave of incision (Harkins et al. 2007).

It is thus necessary to directly date the terraces, to determine their age of deposition and abandonment to constrain the incision history of the Yellow River and the rates of tectonic uplift of the narrow ranges cross-cut by the river.



Fig. 7. (a) View towards south of terrace T7 surface, (see location in Fig. 2). Surface T7 merges smoothly with range slopes at the margin of the basin. (b) View towards north of T7 on the right bank of the Yellow River, in southern Gonghe basin. In front, gravel quarry that allowed access to first 10 m of terrace deposit.



Fig. 8. (a) The top 10 m of terrace T7 in a gravel quarry on the right bank of the Yellow River, north of Heka Shan (see location in Fig. 2). The terrace conglomerate is covered by a 2 m thick loess deposit. Yellow vertical line (hanging tape) shows position of depth profile. Schematic section is shown on left with position of samples. Single pebbles, amalgamated samples and sand lenses were collected from 2-10 m depth. (b) Detail of palaeosol between loess cover and conglomerate. (c) Detail of sand lenses.

Terrace dating, results and interpretation

Cosmogenic nuclide dating method

Cosmogenic nuclides are produced *in situ* by the interactions between secondary cosmic ray particles and surface rocks. Cosmogenic nuclides accumulate in rocks as long as they stay at/near to the surface of the Earth. ¹⁰Be and ²⁶Al are mainly produced in quartz from O and Si by spallation and muonic capture, and the ubiquity of this mineral makes it a favoured target. The concentration of cosmogenic radioactive nuclides in a rock, *N*, is given by:

$$\frac{dN}{dt} = P - \lambda N,\tag{1}$$

where λ is the decay constant of the nuclide considered and *P* is the production rate, which depends on the geographic location and altitude. For sub-surface samples, the production rate decreases exponentially with depth following:

$$P(z) = P(0) \exp\left(-z\frac{\rho}{\Lambda}\right), \qquad (2)$$

where P(0) is the production rate at the surface, *z* the depth of the sample, ρ the density of the overlying material and Λ the attenuation length of the interacting particle. Analysing deep (shielded) samples may give access to the amount of inheritance. Indeed, samples situated below several times the penetration depth are shielded from cosmic radiations and their cosmogenic nuclide concentrations thus correspond to nuclides accumulated before being deposited, that is, the inherited component. When inheritance and erosion are assumed to be negligible, then the minimum exposure age of a sample is given by:

$$t_{\min} = \frac{1}{\lambda} \ln \left(\frac{1 - \lambda N}{P} \right). \tag{3}$$

Taking into account erosion, inheritance and production by muons, the number of ¹⁰Be atoms grows with time according to:

$$N(z, \Delta t) = N_0(z, \Delta t) \exp(-\lambda \Delta t)$$

$$+ \left[\frac{P_n}{\lambda + \frac{\rho}{\Lambda_n} \varepsilon} \exp\left(-\frac{\rho}{\Lambda_n} z\right) \right]$$

$$\times \left(1 - \exp\left(-\left(\lambda + \frac{\rho}{\Lambda_n} \varepsilon\right) \Delta t\right) \right)$$

$$+ \left[\frac{P_m}{\lambda + \frac{\rho}{\Lambda_m} \varepsilon} \exp\left(-\frac{\rho}{\Lambda_m} z\right) \right]$$

$$\times \left(1 - \exp\left(-\left(\lambda + \frac{\rho}{\Lambda_m} \varepsilon\right) \Delta t\right) \right), \quad (4)$$

where $N_0(z, \Delta t)$ is the inherited component, (n, m) subscripts for spallation by neutron and muons, respectively.

Samples were processed in the Cosmogenic Nuclides Laboratory of the University of Strasbourg following standard methods described in Kohl & Nishiizumi (1992). ¹⁰Be/⁹Be and ²⁶Al/²⁷Al ratios were determined by accelerator mass spectrometry (AMS) at the ASTER (CEREGE, Aix-en-Provence) facility except two samples (Qi4-29-1 and Qi4-29-19, Table 1), which were measured at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory (Table 1). Natural ⁹Be has been assumed to be not present in our samples after a set of samples has been checked by ICP-MS. The total ⁹Be concentration is thus determined by the amount of carrier added to the sample.

The ²⁶Ål concentrations were measured in a subset of samples to better constrain the exposure history of the terraces. As the stable isotope ²⁷Ål is naturally present in quartz minerals, sample content of natural ²⁷Ål was measured by ICP-MS before addition of ²⁷Ål carrier. ¹⁰Be and ²⁶Ål have different decay constants and production rates, but the same production profiles. They can thus be analysed together and compared. In case of a simple exposure history, without a period of burial, the ²⁶Ål/¹⁰Be nuclide ratio approximates 6.75 (Balco 2009).

Nuclide concentrations were determined from the AMS measurements and the amount of corresponding stable isotope (Table 1). Ages were modelled using the CRONUS online calculator (Balco et al. 2008) version 2.2 (April 2009). Sub-surface sample concentrations have been modelled by fitting the sample concentrations following equation (2) (see also Granger & Smith 2000) and using the constant surface production rate determined by the CRONUS calculator (see details below). We used the revised ¹⁰Be decay constant of $5.1 \pm 0.26 \times 10^{-7} a^{-1}$ of Nishiizumi *et al.* (2007) and the ²⁶Al decay constant of $9.83 \times 10^{-7} a^{-1}$, following Balco et al. (2008). Penetration depths used for depth profiles ages calculation are 160 g cm^{-2} for neutrons and 1500 g cm^{-2} for muons (Granger & Smith 2000; Gosse & Phillips 2001). We used densities of 2.65 for quartz, 2.7 for granite, 2.5 for a mix gravel/pebble/cobble deposit, 1.8 for soil, 1.4 for loess. No topographic shielding correction was needed given the extent and flatness of the terraces. All other data used in the calculator, as advised by Balco et al. (2008) and Frankel (2010), are given in Table 1.

The largest analytical uncertainty comes from the production rate estimation (Gosse & Phillips 2001). Balco *et al.* (2008) estimated this uncertainty to approximate 10%. It is however larger for depth samples because uncertainties on penetration

Sample name	Sample description	Latitude (°N)	Longitude (°E)	Elevation (m a.s.l.)	Depth (cm)	Sample thickness	¹⁰ Be concentration	err	²⁶ Al concentration	err	[²⁶ Al]/ [¹⁰ Be] ratio	err
	I.					(cm)	$(at g^{-1})$		$(at g^{-1})$			
Terrace T5												
Qi4-30	Quartz cobble	36.17584	100.52143	2920	0	3	2.98E + 06	8.44E + 04	na	na	na	na
Qi4-30-2 [‡]	C C						3.12E + 06	8.87E + 04	na	na	na	na
Qi4-33	Quartz cobble	36.17567	100.52181	2920	0	4	2.98E + 06	8.57E + 04	na	na	na	na
Qi4-34	Quartz cobble	36.17520	100.52123	2920	0	8	3.26E + 06	9.37E + 04	na	na	na	na
Qi4-36	Quartz cobble	36.17532	100.521	2918	0	5.5	3.27E + 06	9.27E + 04	na	na	na	na
Qi4-36-2 [‡]							3.34E + 06	9.77E + 04	na	na	na	na
Qi4-38	Quartz cobble	36.17555	100.52113	2920	0	6	3.05E + 06	9.43E + 04	na	na	na	na
Qi4-39	Quartz cobble	36.17566	100.52179	2919	0	8	3.74E + 06	1.02E + 05	na	na	na	na
Qi4-40	Quartz cobble	36.17595	100.52154	2920	0	5	2.84E + 06	8.52E + 04	na	na	na	na
Terrace T7												
Oi4-29-1*	Ouartz cobble	35.70043	100.29636	3196	200	5	9.55E + 05	2.56E + 04	4.64E + 06	5.33E + 05	4.86	0.57
Qi4-29-27a [†]	8 amalgamated pebbles	35.70043	100.29636	3194	200-220	2	1.53E + 06	4.39E + 04	na	na	na	na
Qi4-29-27b [†]	13 amalgamated	35.70043	100.29636	3194	200-220	2	1.32E + 06	4.04E + 04	na	na	na	na
$Qi4\text{-}29\text{-}27c^\dagger$	32 amalgamated	35.70043	100.29636	3194	200-220	1.5	1.47E + 06	6.43E + 04	8.88E + 06	4.31E + 05	6.03	0.39
Qi4-29-27d [†]	55 amalgamated pebbles	35.70043	100.29636	3194	200-220	1	1.54E + 06	5.53E + 04	9.11E + 06	3.94E + 05	5.91	0.33
	1				Weighted mean: 1.4615E +			$6 \pm 0.1031E + 06$				
Qi4-29-19*	Quartz cobble	35.70043	100.29636	3194	230	4	1.07E + 06	2.85E + 04	6.33E + 06	2.13E + 05	5.89	0.25
Qi4-29-9	Quartz cobble	35.70043	100.29636	3194	240	4.5	2.13E + 06	7.31E + 04	8.77E + 06	3.80E + 05	4.12	0.23
Qi4-29-13	Quartz cobble	35.70043	100.29636	3194	260	5.5	7.66E + 05	2.83E + 04	na	na	na	na
Qi4-29-11	Quartz cobble	35.70043	100.29636	3194	320	6	3.97E + 05	1.61E + 04	2.43E + 06	1.11E + 05	6.12	0.37
Qi4-29-11-2 [‡]							4.00E + 05	1.59E + 04	na	na	na	na

Table 1. ¹⁰Be and ²⁶Al analytical results of surface and sub-surface samples of the Yellow River terraces in the Gonghe basin

Qi4-29-14	Sand lens	35.70043	100.29636	3194	370	2	7.85E + 05	2.74E + 04	3.79E + 06	1.51E + 05	4.83	0.25
Qi4-29-15	Sand lens	35.70043	100.29636	3194	390	2	5.05E + 05	1.75E + 04	2.46E + 06	9.65E + 04	4.87	0.26
Qi4-29-16	Quartz cobble	35.70043	100.29636	3194	400	6	6.18E + 05	2.45E + 04	3.66E + 06	1.62E + 05	5.93	0.35
Qi4-29-21	Quartz cobble	35.70043	100.29636	3194	615	5	2.35E + 05	9.48E + 03	1.34E + 06	6.88E + 04	5.69	0.37
Qi4-29-22	Quartz cobble	35.70043	100.29636	3194	675	9	1.90E + 06	6.54E + 04	9.67E + 06	4.12E + 05	5.09	0.28
Qi4-29-23	Sand lens	35.70043	100.29636	3194	750	2	4.45E + 05	1.87E + 04	2.25E + 06	2.03E + 05	5.06	0.50
Qi4-29-24	Quartz cobble	35.70043	100.29636	3194	480	6	5.38E + 05	2.04E + 04	3.26E + 06	1.35E + 05	6.06	0.34
Terrace WRB												
Qi4-47	Granitic bedrock	36.12209	100.93520	2680	0	4	1.28E + 06	3.72E + 04	na	na	na	na
Qi4-47-2 [‡]							1.40E + 06	5.74E + 04	na	na	na	na
Qi4-48	Granitic bedrock	36.12221	100.93571	2680	0	3.5	1.18E + 06	3.44E + 04	na	na	na	na
Qi4-49	Granitic bedrock	36.12184	100.93532	2683	0	3.5	1.90E + 06	7.73E + 04	na	na	na	na
Qi4-50	Cobble	36.12196	100.93521	2680	0	8	1.91E + 06	5.46E + 04	na	na	na	na
Qi4-51	Cobble	36.12203	100.93511	2680	0	8	1.89E + 06	5.43E + 04	na	na	na	na
Qi4-52	Cobble	36.12212	100.93505	2682	0	8	1.85E + 06	6.26E + 04	na	na	na	na
Qi4-53	Cobble	36.12212	100.93505	2681	0	4	1.90E + 06	6.11E + 04	na	na	na	na
Qi4-54	Cobble	36.12214	100.93503	2681	0	4	1.89E + 06	6.28E + 04	1.17E + 07	5.83E + 05	6.16	0.37
Terrace WLB												
Qi4-59	Quartz vein in	36.13491	100.95922	2925	0	8	5.61E + 05	1.91E + 04	na	na	na	na
	bedrock											
Qi4-60	Granitic bedrock	36.13455	100.95926	2934	0	3.5	1.82E + 06	6.16E + 04	na	na	na	na
Qi4-61	Cobble	36.13449	100.96021	2938	0	4	2.64E + 06	9.86E + 04	na	na	na	na
Qi4-61-2 [‡]							2.58E + 06	8.39E + 04	na	na	na	na
Qi4-62	Granitic bedrock	36.13405	100.96105	2940	0	3	2.49E + 06	7.56E + 04	1.30E + 07	5.30E + 05	5.23	0.27
Yellow River sand												
Qi4-43A	River sand	36.12972	100.9967	2367	0		3.09E + 05	1.24E + 04	1.63E + 06	7.63E + 04	5.28	0.32

na, not analysed. *AMS measurements at CAMS at Lawrence Livermore National Laboratory. [†]See details of sample characteristics in Figure 9. ^{*}Duplicate of the same sample after quartz leaching.



Fig. 9. Mass distribution plot (probability density function) of amalgamated pebbles of samples Qi4-29-27a, b, c and d collected about 2.1 \pm 0.1 m deep below T7 surface. Number of pebbles, average mass and approximate size are indicated. Concentrations of ¹⁰Be measured in amalgams are shown (see Table 1) together with weighted mean.

depths (6%), sample depth (0.5%) and density of overlying material (2–3%) must be added. Final uncertainties on production rates are then comprised between 10 and 12%. Uncertainties from the chemical processing are low (1–2%). Uncertainties from the AMS measurement are comprised between 1.8 and 3.5%. Total analytical uncertainties on ages, not considering modelling errors, thus range from 11-13% (Table 1).

We sampled four different terrace levels for cosmogenic dating, T7 and T5 in the basin, and two strath terrace levels in the gorge, WRB (Waligong right bank) and WLB (Waligong left bank) (Fig. 2, Table 1).

Cosmogenic nuclide measurements

Terrace T7 samples

Stratigraphy. On the southern side of the Gonghe basin, north of Heka Shan and on the right bank of the Yellow River, clean exposure of the top 10 m of the T7 level can be seen in a gravel quarry used for road constructions (Figs 2d, 7 & 8). The terrace conglomerate is a thick fill deposit, composed of gravels, pebbles and cobbles with a few 5-50 cm thick and several metres long sand lenses (Fig. 8c). It is covered by a 15 cm thick reddish palaeosol (Fig. 8b), itself covered by *c*. 2 m of aeolian loess (in the following sections, layer and sample depths are given relative to present ground surface, i.e., the top of the loess deposit). The conglomerate pile is marked by an angular

unconformity at c. 3.5-4 m depth. In an eastwest section, the c. 5° west-dipping pebble layers and sand lenses are intersected by overlying horizontal layers. In the north-south section, the angular unconformity disappears, but a clear layer transition can be seen in the section approximately at the same depth (Fig. 8). The two conglomerate units will be called C1 and C2 from bottom to top in the following sections. The top 20 cm of the C2 conglomerate, just below the palaeosol, is marked by a visible increase in the amount of whitish quartz-vein centimetric pebbles (Fig. 9; Table 1). Our field observations and inspection of the quarry cliffs have not allowed us to distinguish any other sedimentation interruption or other unconformities. Note that the latter would be difficult to detect since there is no obvious change in the nature and origin of the material deposited. Therefore, in the following discussion, we will consider the section analysed as a sequence of two depositional events (assuming that other sedimentation interruptions, if present, lasted only for short periods compared to the whole life time of the terrace), followed by the formation of a palaeosol, itself followed by the deposition of loess.

Sampling. From depths of 2-7.5 m we collected 16 quartz-rich samples for cosmogenic exposure dating along two nearby vertical profiles (Fig. 8, Table 1). At a depth of 2 m, a cobble was found above the palaeosol but below the loess cover (sample Qi4-29-1). Below the palaeosol, in unit C2, we collected a large number of centimetre-sized quartz-vein pebbles at a depth of 2.10 + 0.10 m. These pebbles were sorted by size and analysed as 4 separated amalgamated samples (samples Qi4-29-27a, b, c and d; Table 1; Fig. 9). Four additional samples are individual quartz-rich cobbles (quartz vein or granitoid) from 2.3 to 3.2 m. In unit C1, three sand lenses were sampled at 3.7, 3.9 and 7.5 m (samples Qi4-29-14, 15 and 23, respectively; Table 1). The 4 remaining samples are individual quartz-rich cobbles at 4.00, 4.80, 6.15 and 6.75 m (samples Qi4-29-16, 24, 21 and 22; Table 1). Among those 16 samples, 13 were analysed for ²⁶Al (Table 1).

¹⁰Be nuclide concentrations. ¹⁰Be concentrations range from 235 000 to 1 500 000 at g^{-1} (qtz) for most of the samples, except 2 that have highest concentrations of about 2 millions at g^{-1} (qtz) (Fig. 10a, Table 1). Lowest concentrations are those of cobble samples Qi4-29-21 and Qi4-29-11 with, 235 000 and 405 000 at g^{-1} (qtz), respectively. The sand lens samples have larger concentrations of 445 000, 505 000 and 785 000 at g^{-1} (qtz). The large concentrations of 1 500 000 at g^{-1} (qtz) is reached for samples right below the



Fig. 10. (a) Plot of ¹⁰Be concentrations of sub-surface samples of terrace T7 (see Table 1). Schematic stratigraphy of the section is shown. (b) 26 Al/ 10 Be ratio as a function of depth for a subset of samples of Figure 10a. Models 1, 2 and 3 are fitted to both the 10 Be and the 26 Al/ 10 Be ratio (see text and Table 2). (c) Models showing the impact of various loess accumulation scenarios, first number is duration of loess deposit, second number is duration of exposure at the end of loess deposit. Conglomerates C1 and C2 where exposed each during 15 ka (Table 2).

soil-conglomerate interface, samples Qi4-27a, b, c and d (Table 1). The whole sequence of 10 Be concentrations measured in the different samples do not show a monotonous exponential decrease with depth but imply either sequential deposition (e.g. Schaller *et al.* 2002; Matmon *et al.* 2009) or a variable inheritance between samples (e.g. Anderson *et al.* 1996; Repke *et al.* 1997; Mériaux *et al.* 2004, 2005; Le Dortz *et al.* 2009). In particular the different types of samples collected, individual cobbles, sand lenses or amalgamated pebbles may have various pre-exposure components (e.g. Oskin *et al.* 2008). However, within each unit, C1 and C2, the sample concentrations show an exponential decrease with depth linked to their respective exposure histories (Fig. 10).

 ^{26}Al nuclide concentrations. The ^{26}Al concentrations were measured in 13 samples (Table 1). The concentrations range from 2–9 millions at g⁻¹ (qtz) and imply $^{26}Al/^{10}Be$ ratios that vary between 4.12 to 6.06 (Figs 10 & 11; Table 1). Low $^{26}Al/^{10}Be$ ratio values usually result from differential decay of ^{10}Be and ^{26}Al caused by shielding from cosmic rays after a period of nuclide accumulation (Bierman 1994). None of the samples have typical simple exposure surface $^{26}Al/^{10}Be$ ratio of 6.75 implying either shielding of the section or



Fig. 11. ²⁶Al/¹⁰Be ratio v. ¹⁰Be for a subset of samples from the depth profile of terrace T7 (Table 1).

a significant component of inherited nuclides. Disregarding a few anomalous samples, such as, Qi4-29-1 that is probably a reworked cobble found above the palaeosol, Qi4-29-9 and Qi4-29-22 that have high ¹⁰Be concentrations (large inheritance) and the sand samples, the ${}^{26}Al/{}^{10}Be$ ratio decreases slightly from 6 to 5.7 between 2 m to 6.15 m depth (Fig. 10b). The sand lenses have lower ²⁶Al/¹⁰Be ratios of about 4.8 to 5.06, which may indicate incorporation of material with inherited low ²⁶Al/¹⁰Be ratios. We note that this might also be the case of present river sands, which have a ratio of about 5.28 (sample Qi4-43A, Table 1). We will jointly model (see below) the 10 Be concentration profile and the 26 Al/ 10 Be ratio to constrain the age of deposition and abandonment of the T7 level.

Terrace T5 samples

Geomorphological setting. Terrace T5 is one of the largest terraces SW of Gonghe situated about 150 m below T7 (Figs 2, 3 & 12). It is flat in its northern part where the surface is characterized by a loose gravel and pebble pavement (Fig. 12). At places, sand dunes (barkanes) travel southeastwards across the terrace with the probable consequence of incorporation of modern sand below the pavement and abandonment of sandy patches at the terrace surface. We selected the wide paved surface because such surfaces are usually characterized by low erosion rates and are thus well suited for surface exposure dating (e.g. Van der Woerd *et al.* 1998, 2006; Mériaux *et al.* 2004; Matmon *et al.* 2009). Sampling and measurements. We sampled 7 well embedded cobbles on terrace T5 (Fig. 12). The 7 samples have close ¹⁰Be concentrations on the order of 3×10^6 at g⁻¹ quartz (Table 1). This high nuclide concentration, which is twice the concentration of samples Qi4-29-27 from 2 m depth in T7, suggest that: (1) T7 must have been shielded quite rapidly after deposition; and (2) the T5 terrace surface has never been shielded; and (3) inheritance and/or erosion, if present, have similar effects on the samples of T5. We will use the weighted mean of the 7 sample concentrations (3.21 \pm 0.30 \times 10⁶ at g⁻¹ (qtz); Table 1) as the ¹⁰Be nuclide concentration of terrace T5 surface.

Waligong terrace samples

Geomorphology. It is across the Waligong ranges that the Yellow River has cut its most impressive gorge over a length of 30 km and a total drop of water level of 200 m (Figs 2 & 3). Across the western Waligong, the Yellow River has carved a deep gorge with walls reaching about 500 m (Fig. 3a, b). The flat top of the gorge mimics the overall shape of the western Waligong (Figs 4 & 5), and slopes towards the west into Longyangxia reservoir (Fig. 2). In fact, in more details, the top of the gorge is formed by the progressive down cutting of the river resulting in abandonment of a sequence of strath terraces carved in the bedrock, which are covered by a thin deposit of fluvial gravel and cobbles (Figs 5 & 13). These terraces now slope towards the west indicating that the western Waligong is actively growing and has progressively back tilted the terraces. While the main



Fig. 12. (a) View to east of terrace T5 surface, showing a loose pavement of gravels and pebbles. Note absence of loess cover (see location in Fig. 2). At places, terrace T5 is covered by sand dunes and grass patches. Surface cobbles were collected for ¹⁰Be cosmogenic dating in very flat areas with no sign of soil development or local incision. (**b**-**h**) close-up of cobbles sampled for surface exposure dating.



Fig. 13. (a) View of Waligong left bank terrace (WLB) just above Longyang gorges (see Fig. 2 for location). The terrace is a smooth strath carved in granitic bedrock of Waligong Shan covered by a thin layer (>1 m) of gravels and pebbles. Both granitic bedrock strath and overlaying pebbles were targeted for ¹⁰Be exposure dating. (b) Schematic section with sample locations on WLB. (c) View of Waligong right bank (WRB) terrace with sample locations. Similarly to WLB, the strath terrace is smooth and covered by a thin layer of pebbles and cobbles. (d) Schematic section of WRB with sample locations. Samples Qi4-47 to -49 are bedrock, samples Qi4-50 to 54 are cobbles.

terrace levels can be clearly depicted in the landscape (Fig. 5c, d), their shape narrows near the gorge and their riser join with a low angle the gorge rims so that they become unclear near the gorge and we were unable to number and map them precisely.

Sampling and measurements. We sampled two sites on each side of the gorge (Figs 2 & 13), Waligong right bank (WRB) and Waligong left bank (WLB), located at 2680 and 2940 m a.s.l., respectively. Given the difficulties of terraces mapping near the gorge rim, these two sites are not located within the regional Gonghe basin terrace stratigraphy. Elevation only is not sufficient as we suspect uplift to have occurred since their abandonment (see below). At both sites we sampled the bedrock as well as the overlying cobbles, 4 samples at WLB and 8 samples at WRB (Fig. 13, Table 1).

On terrace WLB, samples have ¹⁰Be concentrations that range from $0.5\hat{6}$ to 2.6×10^6 at g⁻¹ (qtz) (Table 1). The lowest concentration $(0.561 \times 10^6 \text{ at})$ g^{-1} (qtz)) is found in a piece of a highly fractured quartz vein sampled near the gorge cliff edge (sample Qi4-59, Fig. 13a). Qi4-60 is a bedrock sample near the cliff edge and has a concentration of 1.82×10^6 at g⁻¹ (qtz). Both remaining samples have higher concentrations, Qi4-62 is a protruding bedrock high (Fig. 13a) and Qi4-61 is a cobble from the top of the terrace conglomerate cover (Fig. 13a). We think that the average concentration of these two latter samples of 2.57 + 0.11 at g⁻ (qtz) (Table 1) corresponds to the nuclide concentration accumulated since terrace abandonment and that the lower concentrations in the other samples are due to removal of the conglomerate cover and erosion of the fragile quartz vein outcrop.

On WRB terrace, samples have ¹⁰Be concentrations that range from 1.2 to 1.9×10^6 at g⁻¹ (qtz) (Table 1). The 35% difference in concentration among the WRB samples may be explained by the relative positions of the samples in the section sampled (Fig. 13c, d). It appears that samples with the lowest concentration, Oi4-48 and Oi4-47 (Fig. 11c) are bedrock samples close to the gorge cliff, which are no more covered by the terrace conglomerate. Sample Qi4-49 belongs to a bedrock high (Fig. 13c) standing at the same level as the pebbly conglomerate top and show a similar ¹⁰Be concentration as the 5 surface cobbles Qi4-50 to 54 (Table 1). The distribution of concentrations of the WRB samples is thus best explained by removal of the conglomerate cover near the gorge cliff. The ¹⁰Be concentration average of $1.89 \pm$ 0.02 e 6 at g^{-1} (qtz) of the 6 samples with highest concentrations correspond to the nuclides accumulated since terrace abandonment.

Age modelling

Model age of T7. We modelled jointly the ¹⁰Be concentration and the ${}^{26}Al/{}^{10}Be$ ratio in the depth profiles (e.g. Anderson et al. 1996; Granger & Smith 2000) considering a deposition scenario in three steps corresponding to the stratigraphy observed in the field (Figs 8 & 10). Modelling results are presented in Figure 10 and take into account progressive increase in muongenic production at depth following analytical solutions described in Granger & Smith (2000) and calibrated to our site (Table 2). Disregarding samples with large inheritance and the sand samples, model 1 is a best fit adjusting all the ²⁶Al/¹⁰Be ratios, and models 2 and 3 are limiting models toward short and long exposure times, respectively. In general, both the ¹⁰Be concentration and the values of the ${}^{26}\text{Al}/{}^{10}\text{Be}$ ratio imply shielding by the loess during more than 150 ka (model 1). Involving fast erosion rates of the loess (0.01 cm a^{-1} , model 3) very long time of loess shielding may be implied (400 ka) with the necessity to have an original loess thickness reaching several tens of metres.

Because the 2 m-thick loess deposit on top of the terrace is a shielding factor for cosmic rays (factor of decrease in production rate amounts to c. 6) (e.g. Hetzel et al. 2004; Matmon et al. 2009) we have to consider its mode and timing of deposition. The presence of a palaeosol below the loess (Fig. 8) implies that the terrace surface remained free of loess during a certain time, that is, quasi without any shielding. Without independent chronological constraints on the loess age, several loess deposition modes can be modelled. In the first mode, the 2 m thick loess is deposited instantaneously (or rapidly compared to the age of the terrace), in the second mode it was deposited continuously during a certain time (Fig. 10c). Some constraints can be added from additional observations in Gonghe and elsewhere. As we mentioned above, loess is absent on the lower Yellow River terraces across the Gonghe basin, particularly on T5, so that the age of T5 may be the time when loess deposition stopped in the area (see below). In addition, 1 to 2 m high loess sections described in northern Tibet generally show progressively younger ages towards the top (e.g. Peltzer et al. 1988; Van der Woerd et al. 2002; Hetzel et al. 2004; Rhode et al. 2007) over a time period of a few thousands of years. Another possible factor of shielding variation is involved if the loess cover is reworked by wind ablation implying a period of loess deposition followed by a period of loess erosion (model 3).

Assuming that sub-surface samples cannot lose any 10 Be, except by radioactive decay, the lowest concentrated and deepest samples in a depth-profile (below *c*. 1.5 m, Anderson *et al.* 1996) determine

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Model	Inheritance (at g ⁻¹ quartz)	Exposure of C1 (ka)	Erosion rate (cm a^{-1})	Exposure of C2 (ka)	Erosion rate (cm a ⁻¹)	Duration of loess deposit (ka)	Exposure loess (ka)	Erosion rate (cm a^{-1})	Total duration of exposure (ka)
1	235 000	15	0	45	0.005	1	150	0.0001	210
2	445 000	15	0	20	0	1	80	0	115
3	235 000	15	0	50	0		400	0.01	465
Models of Figure 10c	235 000 235 000 235 000 235 000 235 000	15 15 15 15	0 0 0 0	15 15 15 15	0 0 0 0	50 25 10 1	10 70 120 150	0 0 0 0	90 125 160 181

 Table 2. Depth profile model parameters (see Fig. 10)

the minimum inherited component of the samples. We have thus considered the inheritance given by sample Qi4-29-21 (i.e. 235 000 at g^{-1} , Table 1) in most models, and show only one alternative model (model 2) with an inheritance of 445 000 at g^{-1} fitting the profile to the two sand samples with the lowest concentration (Qi4-29-23 and 15; Fig. 10, Table 2). Note that the present Yellow River is carrying sand that has a ¹⁰Be concentration of about 300 000 at g^{-1} quartz (Table 1; see also Harkins *et al.* 2007) that may be considered as the average inherited ¹⁰Be component of the alluvial material transported by the river, assuming it is constant over time and not variable with material size.

While below the palaeosol, the fits to the data mostly depends on the choice of the inheritance value, above the palaeosol, the fits mostly depends on the onset of loess shielding (Fig. 10a, Table 2). These models also show that, in all cases, deposition of conglomerate C2 occurred shortly after deposition of conglomerate C1 (10–15 ka). Similarly, deposition of the loess cover occurred shortly (0-25 ka) after deposition of conglomerate C2.

As shown in Figure 10c (Table 2), when a longlasting loess deposition process is considered, shielding starts to be efficient when this process lasts less than 20 ka. This implies that shielding by the loess must have started a few thousands of years after the terrace has been abandoned and that the loess must have been deposited in a short time. Note that this is fully in agreement with the observation, that the terraces below T7 are not loess covered, so that they must have been abandoned after the loess deposition process ceased in the region and on T7 in particular.

Sample Qi4-29-1, sampled above the palaeosol and below the loess may be used as an additional constraint. It is clearly a reworked cobble as it is not emplaced in a fluvial deposit, but it must have been reworked at the time of the formation of the palaeosol and before being capped by the loess. As can be seen in Figure 10a, models that are adjusted to this sample imply a minimum exposure of C2 of 10-15 ka. The same duration would be needed to account for soil formation above C2 and some surface reworking. The ¹⁰Be concentration of this sample implies exposition under the loess during 150 ka.

To summarize, T7, or more precisely the top C2 conglomerate may have been deposited c. 200 ka, and definitively incised by the Yellow River c. 150 ka, some time before the 2 m loess cap accumulated on top of C2.

Model age of T5. Without independent constraints on erosion and inheritance for terrace T5, these parameters have to be estimated. Inheritance may be taken similar as for terrace T7 (235 000–445 000 at g^{-1} quartz) or similar to the

present Yellow River sand ¹⁰Be concentration of $300\ 000\ at\ g^{-1}$ quartz (Table 1). Regarding с. erosion, studies have been carried out in the Gonghe basin following intensification of agriculture and its consequence on desertification. Using the ¹³⁷Cs technique to assess erosion in drylands and sand shifting areas, erosion rates of about 1-2 mm a^{-1} have been estimated over the last c. 40 years (Yan et al. 2002; Zhang et al. 2003). These values of erosion are rather high (Lal et al. 2003) and may only reflect the consequence of human activities. Lower values of $0.003-0.01 \text{ mm a}^{-1}$ were estimated for bedrock outcrops throughout Tibet (Lal et al. 2003). Again these values should be taken as maximum erosion rates for a flat alluvial terrace.

The age of T5 calculated from the weighted mean concentration of the surface samples $3.21\pm0.30\times10^6$ at g⁻¹ (qtz), is 106 ± 6 ka using a constant production rate model (Lal 1991; Stone 2000). Both the inheritance of 300 000 at g⁻¹quartz and a maximum erosion rate of 0.002 mm a⁻¹ (Fig. 14) imply subtractive and additive corrections of 15-20%, respectively. Without evidence that erosion rates may be larger, we thus conclude that the minimum age of T5 is 106 ± 15 ka.

Age model of Waligong. To calculate the age of the terraces we need to take into account the change in elevation due to their tectonic uplift (e.g. Hetzel *et al.* 2004; Ruszkiczay-Rüdiger *et al.* 2005*a, b*) (Fig. 15). Waligong Shan is composed of two subparallel ranges at the junction between the eastern termination of the Qinghainan and the Laji Shan closing the northeastern corner of the Gonghe basin (Fig. 2). The western Waligong is a 15 km wide and 50 km long arcuate range trending roughly NE–SE (Figs 1 & 2). Its asymmetric topographical shape, with a steep NE and eastern flank under the summital crest that reaches 3500 m a.s.l. above a west dipping thrust and a western flank



Fig. 14. Minimum exposure age range of T5 and influence of erosion. Erosion of about 0.001 mm a^{-1} cannot be excluded, but erosion of 0.003 mm a^{-1} may be a maximum (see text for details).



Fig. 15. (**a**–**d**) Schematic reconstruction of progressive incision of Yellow River across uplifting Waligong range. Phased incision leads to the formation of large stepped terraces across the basin, while strath terraces formed across Waligong bedrock are progressively uplifted and tilted. (**e**) Present-day section at same scale.

that slopes c. 3.5° westward, (Fig. 4e) indicate that it is an east-vergent crustal anticline. The eastern Waligong is less elevated but larger (20–30 km), trends NW–SE and makes up the western limit of the Guide basin. Similarly to the western Waligong, it has a steep eastern rim marked by a thrust at its base (Pan 1994) and grows as an east-vergent anticline (Fig. 2). As can be seen in the field (Fig. 13) the strath terraces are carved into the Waligong Mesozoic granodiorite (Geological map of Qinghai 1988).

A simple model (Fig. 15) accounting for the tilt of the Waligong Shan and uplift of the terraces

implies possible changes in elevation of 80 m for WRB terrace and of about 300 m for WLB terrace (Fig. 15). For WRB, the in-situ production rate changes from 25.73 at $g^{-1} a^{-1}$ at 2600 m to 27.04 at g^{-1} a⁻¹ at 2680 m. This change in production rate of c. 10% remains small with regard to other uncertainties and may be neglected. For WLB the production rate change is more significant, varving from 26.6 at $g^{-1} a^{-1}$ at 2640 m to 31.93 at $g^{-1} a^{-1}$ at 2940 m, that is, a change of 20%. The real mean production rate of this terrace lies certainly between these end-values and we will take the mean between these bounds with a large error of 15%. The terrace ages are thus 74 \pm 10 ka for WRB and 90 \pm 15 ka for WLB. The maximum changes in elevation mentioned above imply maximum uplift rates for Waligong Shan between $1-3 \text{ mm a}^{-1}$

However, this model does not take into account the low ${}^{26}\text{Al}/{}^{10}\text{Be}$ ratio (5.23 + 0.27) obtained for sample Qi4-62 (WRG), which is a bedrock sample near the gorge cliff (Fig. 13). This ratio implies that the surface bedrock needs to be exposed a first time (first cosmonuclide accumulation), then shielded during several hundred thousand years (ratio decreases; Fig. 11) and then re-exposed until present. The cobble cover, likely deposited by the river, covered the surface before being progressively eroded to the present state (Fig. 13). The beryllium concentrations of the cobbles would thus reflect the exposure duration since the river left, but bedrock concentrations are not totally related to this last phase of aggradation and erosion. The surface would thus be exposed since c. 80 ka. Initiation of uplift likely began shortly afterwards at a minimum rate of 0.2 mm a^{-1} . Despite the fact that the ¹⁰Be concentrations of the cobbles are higher on WLB than on WRB because of their higher elevation and production rate, simple calculations show that this scenario would lead to ¹⁰Be concentrations close to 2.0×10^6 at both sites, with a difference of $c. 2.0 \times 10^5$ atoms instead of the c. 6.0×10^5 measured (Table 1).

Discussion

Onset age of incision in the Gonghe Basin

Whether the beginning of river incision corresponds to the first appearance of the Yellow River in the basin is an open question. Terrace sequences along the Yellow River differ markedly from Gonghe to Lanzhou. As explained above, terraces in the different basins seem completely unrelated to each other (ages, loess thickness, elevations). We cannot simply rely on these terraces to compare our chronology in the Gonghe Basin, but the timing of river incision should be coherent between the different basins (Harkins *et al.* 2007).

Several authors have proposed that incision in the Gonghe Basin began at c. 150 ka (e.g. Li 1991; Pan 1994). Evidences for a major change along the Yellow River 150 ka ago have been summarized by Zhang et al. (2004). Periods of high regressive erosion from the Longyang gorges and the Sanmen gorges (in the middle reaches of the Yellow River north of the Ordos plateau, inset, Fig. 1) are reported around that time (Zhang et al. 2000; Wang et al. 2002). This period of erosion in the Longyang gorges could have followed the folding of the fluvial sediments of the Gonghe formation (Li 1991). An increased sedimentation in the marine delta of the river could be related to this period of high erosion and to the climate warming of MIS-5 (Imbrie et al. 1984).

However, it has been proposed that incision began c. 500 (\pm 200) ka ago in the Tongde Basin (Harkins et al. 2007; Craddock et al. 2010), following the last 1.8 Ma lacustrine deposit near the top of the section in Guide (Li et al. 1997). A direct implication is that incision in the Gonghe Basin began at the same time or before, but not after, which means that T7 must be much older than results from our investigations suggest. This can only be achieved if a significant amount of erosion or a long period of shielding occurred and may imply that all terraces are significantly older. A 140 ka (OSL) terrace located 140 m above the Yellow River is located at an elevation of c. 2830 m in the Tongde basin (Harkins et al. 2007; Craddock et al. 2010). For comparison, an equivalent terrace in the Gonghe Basin would be near to T1, which is expected to be much younger than 140 ka.

Incision rates of the Yellow River in the Gonghe Basin

Terrace ages are plotted against their elevation on Figure 16. The minimum and maximum long-term incision rates are c. 2 and $c. 6 \text{ mm a}^{-1}$, respectively, with a mean rate of c. 4 mm a⁻¹. This mean rate fits well with the ages of T5 and T7. However, for all data it is needed to account for changes in the incision rate. The incision rate between the formation of T7 and T5 depends strongly on the age of T7, if erosion on T5 remains small. A period of high incision is needed between T5/WLB and WRB. The incision rate may have reached 9 mm a^{-1} (10 mm a^{-1} maximum, if no elevation changes are taken into account for the Waligong terraces). Our data suggest that incision was first slow and then accelerated between c. 100-50 ka, before slowing down again. Such fast incision would have favoured the steep Waligong gorges formation.

A mean incision rate of 4 mm a^{-1} is higher than those mentioned in other places along the Yellow River in the literature, 0.7 mm a⁻¹ since 70 ka near Lanzhou (Wang *et al.* 2010), between 0.75–1 mm a⁻¹ since 166 ka near Linxia (Li *et al.* 1997) for the downstream part, between



Fig. 16. Incision rate of the Yellow River deduced from terrace ages and elevations. Coloured boxes are allowed age and elevation ranges. Yellow River elevation taken at Waligong gorges entrance (= WRB site). Several scenarios and associated incision rates are proposed. For Waligong terraces, rates are mean rates calculated from the middle of boxes. Maximum and mean total incision rates 6.0 and 4.0 mm a^{-1} , in orange.

 $0.9-1.0 \text{ mm a}^{-1}$ for the last 140 ka in the Tongde Basin (Harkins *et al.* 2007). But it is in agreement with the 6 mm a⁻¹since 93 ka in the Guide Basin proposed by Pan (1994).

Climatic correlations and formation of the set of terraces

Climatic fluctuations play a key role in shaping fluvial terraces. Correlation of terrace formation periods with climatic events is however not straightforward, because uncertainties inherent to the cosmogenic nuclides dating method often prevent from determining a precise correlation. Transitional periods between glacial and interglacial periods seem nevertheless to play a major role in forming fluvial terraces. These periods would be favourable for incision, while aggradation would occur during stable climatic phases, possibly both glacial and interglacial (e.g. Pan *et al.* 2003; Vandenberghe 2003).

Loess in northeastern China was mainly deposited during glacial periods, while palaeosols developed during interglacial phases, alternating loess and palaeosol layers being correlated to climatic fluctuations (e.g. Ding *et al.* 2002). The fact that T7 gravels are directly overlain by a palaeosol and this palaeosol by a loess cover suggests that this terrace was abandoned during a deglaciation period and covered by loess during an *a posteriori* glacial event (Gao *et al.* 2008; Pan *et al.* 2009). No such climatic indicators are available for the other terraces, although it is reasonable to assume that their formation followed a similar process, that is, a formation as a consequence of dry/cold to wet/warm climate transition.

The terraces in the Gonghe Basin could have been formed during glacial periods and abandoned at the transition with warmer periods, when the river high incision power enabled it to break dams formed by the Waligong Shan.

Evidence for a Gonghe-Guide-Qinghai palaeo-lake

Small lakes can be found in several places in the present Gonghe basin (Fig. 6); Chaka Lake, in the NW, has a present elevation of 3060 m; Dalian Lake, at an elevation of 2850 m (Yan *et al.* 2002); Duolonggou lake, north of Gonghe city, at an elevation of 3194 m. Lacustrine sediments can be found at several places in the Gonghe basin and at Dalian Lake for instance. These small lakes and sediments may be remnants of a larger lake occupying the bottom of the Gonghe basin (Fig. 17). Such a lake may have covered adjacent intra-mountainous basins. It has been reported that sedimentary sequences in the Guide Basin were deposited

under lacustrine conditions (Li et al. 1997; Pares et al. 2003; Fang et al. 2005). Moreover, it is striking that the present level of the Qinghai Lake is exactly at the same altitude as Duolonggou lake and as T7 sampling site (3200 m). Qinghai lake exists since c. 0.5 Ma (Yuan et al. 1990) and its present maximum depth approximates 20 m (Colman et al. 2007). Reconstructions of Oinghai lake water level palaeo-elevations lead to various chronologies (Colman et al. 2007; Madsen et al. 2008; Liu et al. 2010; Rhode et al. 2010). The lake reached a maximum elevation of 3260 m (palaeoshorelines older than 30 ka) and its level was particularly high during the penultimate glacial maximum (c. 150 ka). Other lakes in the region seemingly experienced the same high-stand event, the nearest in distance being Gahai Lake in the Qaidam Basin (Fan et al. 2010).

The Chinese literature discusses the existence of an eastern Qinghai palaeo-lake covering the Qinghai, Gonghe, Xinghai and Xining basin (e.g. Pan 1994). It is based on a possible stratigraphic continuity between sediments in the Guide and Oinghai Basin. Such a lake would have broken into several parts due to the recent tectonic activity. Tectonic activity along the Qinghainan Shan fault likely began at the end of the Miocene or at the beginning of the Pliocene (Métivier et al. 1998; Meyer et al. 1998; Van der Woerd 1998; Fang et al. 2005; Lease et al. 2007) separating the Qinghai and Gonghe Basin. A connection between the two basins was possible at the southeastern extremity of Qinghai Lake when the pass may have been at a lower elevation (present elevation: 3350 m). The Qinghai Basin and the western Gonghe Basin are now internally drained. Madsen et al. (2008) proposed that Qinghai Lake began to form during the Middle or Late Pleistocene when the Rivue Shan uplift separated the Qinghai Basin from the Yellow River drainage system. Uplifting mountain ranges crossing the river course would have acted as boundaries for the eastern Oinghai palaeo-lake, near the Longyang gorges (Gonghe Basin exit), Shanba gorges (Guide Basin exit) or Jishi Gorges (Xunhua Basin exit) (Fig. 17). The change from a lacustrine to a fluviatile environment would in this case correspond to the breaching of the topographic barriers and emplacement of the modern course of the Yellow River across the basins (Pares et al. 2003; Craddock et al. 2010).

However, no large lacustrine deposit can be found at present to attest more clearly the existence of such a lake, in particular at our sampling site of T7. Processes that lead to the formation of the wide palaeo-regional base-level around 3200 m and of lakes present at this elevation stills need to be investigated, as well as the possible past connections between the different basins.



Fig. 17. Topographical contours at 3150, 3200 and 3250 m a.s.l. showing possible extensions of a regional palaeolake or closed basin in the area of Qinghai and Gonghe basins before their separation due to lake capture by Yellow River and growth of the Qinghainan Shan. Lake surface elevation may have reached 3200–3250 m a.s.l. This mapping, drawn from the present topography, does not account for large tectonic and morphological changes.

Conclusion

The Gonghe Basin is one of the widest basins of northeastern Tibet. It contains noteworthy wide terraces that testify to an important and recent water level decrease. Our results indicate that terraces were formed at the transition between glacial and interglacial periods. The upper one probably dates from the penultimate glacial maximum (MIS-6; Imbrie et al. 1984; Thompson et al. 1997). Obtained ages show that incision occurs at a mean rate of c. 4 mm a^{-1} . Higher incision rates during some periods may be related to Waligong Shan uplift, although the tectonic signal remains difficult to separate from the river incision. Our results highlight the consequences of mountain growth and the interplay between tectonic and climatic processes in building the Tibetan Plateau. We provide chronological constraints supporting that this part of the plateau evolved rapidly in recent time from an internally drained but interconnected set of closed basins to an externally river drained plateau margin.

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