Ice-volume-forced erosion of the Chinese Loess Plateau global Quaternary stratotype site


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The International Commission on Stratigraphy (ICS) utilises benchmark chronostratigraphies to divide geologic time. The reliability of these records is fundamental to understand past global change. Here we use the most detailed luminescence dating age model yet published to show that the ICS chronology for the Quaternary terrestrial type section at Jingbian, desert marginal Chinese Loess Plateau, is inaccurate. There are large hiatuses and depositional changes expressed across a dynamic gully landform at the site, which demonstrates rapid environmental shifts at the East Asian desert margin. We propose a new independent age model and reconstruct monsoon climate and desert expansion/contraction for the last ~250 ka. Our record demonstrates the dominant influence of ice volume on desert expansion, dust dynamics and sediment preservation, and further shows that East Asian Summer Monsoon (EASM) variation closely matches that of ice volume, but lags insolation by ~5 ka. These observations show that the EASM at the monsoon margin does not respond directly to precessional forcing.
The margins of deserts are highly sensitive to climate change and human influences. Small changes in vegetation, climate or land use drive major changes in sand dune and dust activity, which in turn have major impacts on local populations, global dust emissions and climate forcing. By extension, sedimentary records from the margins of deserts are highly sensitive indicators of past environmental change in these crucial areas. The desert margin of the Chinese Loess Plateau (CLP; one of the world’s most important terrestrial climate archives) is especially significant in this regard. In addition to recording East Asian Monsoon climate and Asian aeolian dust dynamics in loess and palaeosol units (systems that alter global climate and now affect billions of people), the area also records expansion and contraction of a desert sand sea that has experienced significant Holocene and recent desertification. This is of particular relevance, given that the loess–palaeosol climate proxy record from the CLP desert marginal Jingbian site (Fig. 1) has been adopted as the terrestrial stratotype for the International Commission on Stratigraphy’s global benchmark Quaternary chronostratigraphic scheme. It is therefore of central importance in past climate research.

Nevertheless, our understanding of how processes in desert marginal environments impact the preserved sedimentary record is limited, and the longer-term driving forces behind sand activity remain debated due to the limited preservation potential of dune sediments. Sandy desert areas are known to be highly complex and dynamic environments, with the location of deposition and erosion shifting rapidly and across small distances in response to forcing by winds and precipitation. Jingbian lies in an area that has been covered by expanded sand dunes in the past. Such processes could therefore severely compromise the completeness of the stratigraphic record and undermine the integrity of correlation-based, non-independent chronostratigraphic models such as the one used in the ICS scheme. Furthermore, detailed optically stimulated luminescence (OSL) dating of more central CLP sites over the last glacial has shown that age models derived from correlation-based methods contain significant inaccuracies of up to 10 ka. More fundamentally, a recent proposal argues that the CLP is a highly dynamic environment which leads to substantial internal aeolian recycling of deposited material and a reduction in CLP area size. Such sediment recycling would undermine routine desert marginal CLP paleoenvironmental reconstruction as well as the basis of understanding of past monsoon, dust and desert dynamics in this region. By implication this hypothesis also calls into question the accuracy of the ICS Jingbian chronostratigraphy. It is thus crucial that the past loess and desert record at Jingbian is independently constrained.

Here we develop a fully independent age model for the Jingbian section over the last ~250 ka using a combination of quartz OSL and K-feldspar post-IR Infra-Red Stimulated Luminescence (IRSL) techniques applied at high sampling resolution. This model shows that Jingbian is characterised by numerous hiatuses of up to ~60 ka that are highly spatially variable across a heavily eroded gully section (Fig. 2). This radically changes the palaeoclimatic interpretation of the sedimentary sequence preserved at the site, supports a revised model for development of the CLP, provides new insights into East Asian Summer and Winter Monsoon (EASM/EAWM) dynamics, and requires a major revision of the ICS chronostratigraphic scheme for Jingbian.

Results
A luminescence-based chronostratigraphy for the past 250 ka.
Our new luminescence age model is based on 220 ages on

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**Fig. 1** Map of Chinese Loess Plateau showing surrounding deserts and rivers and location of the ICS stratotype site Jingbian. The Jingbian site location is marked with a filled red triangle and other well-known loess sites (filled red circles) and the Sanbao cave site (filled pink circle) are also indicated. Inset shows location of map in China and main prevailing wind directions of East Asian Winter Monsoon (EAWM), East Asian Summer Monsoon (EASM) and Indian Summer Monsoon (ISM). The data set is provided by Data Center for Resources and Environmental Sciences, Chinese Academy of Sciences (RESDC) (http://www.resdc.cn). The base map is a coloured DEM map derived from SRTM 90 m data and the inset map is based on http://www.arcgis.com/home/item.html?id=c-3265f30461440c2999add34bcae8e0a. A detailed aerial photograph of the Jingbian site with the studied loess profiles (A, B, C, D, E) marked is given in Supplementary Fig. 1.
samples taken with a vertical spacing of between 5 and 40 cm at 5 Jingbian sections dug at the ICS stratotype location (Fig. 2). It constitutes the largest and most detailed luminescence data set to date, and to our knowledge is the most comprehensive geochronological analysis yet undertaken at a single site. Details on site location, sampling, luminescence dating methodology, age depth modelling and proxy analyses are given in Methods. There are two striking features of the age–depth models for the Jingbian sections (Fig. 2). Large jumps in ages are found in many sections, indicative of large hiatuses in the record of up to 60 ka. Crucially, these substantial gaps are not observable in the visual or proxy stratigraphy and have not been demonstrated previously at Chinese loess sites, yet have major implications for the chron stratigraphic model and climate reconstruction. In addition, while the age ranges of some of the sections overlap, the nature of the preserved record at each section is inconsistent, indicating a highly spatially variable relationship between age, depth, sediment type and preservation.

As with many CLP-desert marginal sites, Jingbian is located in a relatively flat plateau landscape with the sections exposed in a deeply incised gully system (Figs. 1 and 2, Supplementary Fig. 1). Some sections show hiatuses where others concurrently show deposition, and yet other sections exhibit extremely high accumulation rate phases of short duration (Fig. 2). No single section preserves the full sequence covered at the site, as shown in our composite climate records (Fig. 3). As such, these gully sequences require consideration as dynamic landforms, where gully geomorphology and local morphological context must be taken into account, together with the stratigraphic sequence. One consequence of this is that while at many CLP sites the Holocene record has been partly disturbed by human activity21, a uniquely undisturbed 2 m Holocene sequence is preserved at Jingbian (see Fig. 2, section D), protected by unconformable deposition within the gully system and dated by 31 luminescence ages. Thus, the luminescence results reveal that the gully must pre-date the Holocene and that the gully landform itself is recorded in the stratigraphic record at the site. While this dynamism adds to the complexity of interpreting these stratigraphic sequences, our independent dating demonstrates that a detailed composite environmental history can be obtained through luminescence dating of multiple overlapping sections (Fig. 3). This is also reflected in the detailed record of the last interglacial in section E (Fig. 2).

**Ice-volume-forced processes in a desert marginal environment.**

When our climate proxy and stratigraphic records are plotted on our new age model against 65°N July insolation22, marine oxygen isotope stratigraphy LR04 stack23 and Lake Baikal biogenic silica13 (Fig. 3), some striking patterns become apparent. Notably, there is a near total lack of preserved record during the last two glacial phases (MIS 2–4 and 6), but with preserved material from the glacial stage MIS 8, as well as interglacials MIS 1, 5 and 7. The large hiatuses appear to terminate close to or following the rapid shift away from peak Northern Hemisphere ice volume at the end of the MIS 2–4 and 6 glacial stages (Terminations I and II). During less positive marine δ¹⁸O isotope stages when Northern Hemisphere ice volume was lower, loess sediments are generally preserved. During MIS 7 and the second half of MIS 8, there is relatively low amplitude variability in ice volume and full
**Fig. 3** Comparison of global/Northern Hemisphere proxy records with the Jingbian records on independent timescales for the last 300 ka. 

- **a** 65°N July insolation record.
- **b** Benthic LR04 δ^{18}O stack.
- **c** Lake Baikal biogenic silica record.
- **d** Low-field magnetic susceptibility (MS) and sand content (>63 μm) of the Jingbian loess-paleosol record.
- **f** Individual luminescence age determinations for each section are shown with 1 s.d. age uncertainties.

Grey shading behind the MS/sand fraction curves indicates 1 s.d. age model uncertainties. Marine oxygen isotope stages/boundaries shown are from Lisiecki and Raymo. 

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preservation of the loess record, including in the comparatively low ice volume glaciation of MIS 8. Two palaeosols associated with the two ice volume minima of MIS 7 are also preserved, separated by a loess unit representing the deep stadial during MIS 7, while MIS 5 and 1, which have no such deep stadials, are only represented as palaeosols at Jingbian.

Based on this pattern and its relationship to the δ18O record, we propose an explanation of the mechanisms behind desert marginal sediment accumulation, preservation and erosion, and hence the controls on desert dynamics. The greatly enhanced maximal extent of Northern Hemisphere ice limits during peak MIS 2 and 6 is known to have strengthened the Siberian High and moved the polar front southwards, enhancing cold air outbreaks, and strengthening winds and aridity. The associated water stress would have reduced vegetation stabilisation of dunes while cold air outbreaks would have driven seasonally strong winds, promoting deflation and sediment movement; these erosive processes changed Jingbian from a depocentre into a dust source and account for the hiatuses at the site. On shorter timescales, the polar front, modulated by Atlantic Meridional Overturning, has been shown to drive strengthened EAWM circulation and dust deposition on the Loess Plateau, while the strength of the Siberian High is tied to ice volume and snow cover over multiple timescales, supporting this model. While the MIS 2–4 and 6 hiatuses cover most of these glacials (Fig. 3), accumulation may still have occurred locally, but the strong erosional events during peak glaciation would have removed previously deposited material. Sand was deposited at the end of both hiatuses, indicating both an expansion of the Mu Us desert and some dune stability. As no sand was preserved during the prior glacial episodes, a highly mobile sand sea is implied, close by or covering the site and providing a plentiful supply of saltating impactor grains to promote deflation of existing deposited material. Thus, the two major hiatuses at Jingbian during MIS 2–4 and 6 are interpreted as erosional unconformities resulting from enhanced dust mobility driving erosion of underlying strata. During glacial MIS 8 and the stadials within MIS 7 and 5, glaciation did not extend as far as during MIS 2 and 6 (Fig. 3) and so cold air outbreaks, winter monsoon intensity and aridity was not sufficient to drive such dune expansion and dust deflation.

Our revised age model and resulting sedimentary history has a fundamental impact on the interpretation of the global benchmark record at Jingbian. In traditional Loess Plateau chronostratigraphic models, loess/sand units and palaeosols are considered of glacial and interglacial age, respectively. Here we propose a different model for Jingbian. In our view, palaeosol units are indeed indicative of interglacial phases of enhanced EASM (high magnetic susceptibility; MS) and weaker EAWM (finer grain size). However, rather than representative of glacial phases, loess units in the upper part of Jingbian appear to be mainly associated with stadials within interglacials, during which relatively increased ice volume drives cold air outbreaks, aridity and enhanced EAWM circulation, with associated silt transport and dust trapping at Jingbian. Deep glacial phases are removed from the record due to erosion, and sand units occur over more restricted time intervals, both within interglacials and glacials, with both indicating enhanced dust activity and expansion of the Mu Us (Fig. 3). Although the proxy records show general antiphase behaviour of the EASM with the EAWM (Fig. 3), sand accumulation can also occur during enhanced summer monsoon conditions (e.g., MIS 5e). This suggests that sediment availability, EAWM/Siberian High driven winter aridity, and cold air outbreaks and enhanced wind strength drive dune mobility, desert expansion and sand deposition at desert marginal sites. This is in contrast to the idea that dust expansion and deposition is controlled by summer monsoon-driven moisture availability. Recent identification of relic dune sediment from the LGM preserved in isolated frost wedges in the Mu Us confirms intense aeolian activity at this time, but also argues for the domination of net erosion due to high winds and aridity. This explains the lack of dune record from the last glacial in the Mu Us and supports our deep glacial erosional unconformity model. Thus, desert sand dune activity in this part of China is controlled by the intensity of EAWM circulation in Asia, in turn driven by ice volume in the Northern Hemisphere through the Siberian High. Our findings suggest that during peak ice volume phases, this climatically driven erosion in the Mu Us also extended onto the edge of the CLP, driving development of multiple unconformities in one of the global benchmark Quaternary sediment records. This clearly limits the use of Jingbian as a benchmark site for the Quaternary stratigraphic column, and we suggest that a more central CLP site may be more appropriate for use in the Quaternary chronostratigraphic subdivision. Currently, our results demonstrate that the present ICS scheme for Jingbian is incorrect and should be revised.

In addition, our new chronostratigraphic model has a number of significant implications for understanding the CLP, desert sand and atmospheric dust dynamics, as well as monsoon climate. Jingbian lies just south of an escarpment marking the boundary between the Ordos Platform (including the Mu Us desert) and the northern margin of the CLP. Based on the presence of yardangs and wind-gaps cut into Quaternary strata north of this boundary, as well as on loess provenance data, it has recently been proposed that the escarpment has retreated south and east due to wind erosion during peak glacials in a process of ‘aeolian cannibalism’ of pre-deposited loess material. Our finding that large amounts of sand and dust are eroded during peak glacial conditions at Jingbian supports the reinterpretation of the CLP as a dynamic landform, with deposits undergoing reworking and recycling along the boundary with the Mu Us desert. This is the first direct, independent evidence to support ‘aeolian cannibalism’ of pre-existing loess alongside reworked Yellow River alluvial sediments as the source of Quaternary loess to the central CLP and may indicate that indeed the CLP is being reduced in size due to peak glacial wind erosion. That this reworking at Jingbian occurs during those glacial phases (the most recent) with greatest ice volume is also consistent with a long-term increase in aeolian dust CLP accumulation rates over the Quaternary. As glacial stage ice volumes increase and cold air surges penetrate further south, generating large, erosive NW to SE tracking dust storms over the Mu Us, Yellow River alluvial platform and northern CLP, material is reworked and incorporated into younger CLP deposits further south. As such, this apparent long-term accumulation rate increase may be more tied to increasing ice volume and loess recycling rather than to changing aridity or dust source alone. Kang et al. and Stevens et al. noted that independently dated central CLP sites show enhanced dust accumulation during the peak of the last glacial (23–19 ka). This general peak in last glacial dust activity coincides with the hiatuses at Jingbian, and with erosive activity in the Mu Us. We propose that enhanced ice volume may then also be the driver of enhanced Asian dustiness during short phases of the late Quaternary, and erosion of desert marginal loess will likely directly contribute to increased atmospheric dust loading downwind on the central CLP.

EASM, ice volume and lagged response to insolation forcing. EASM-driven MS peaks preserved at Jingbian show a remarkable match with reductions in ice volume (Fig. 3). MS also shows variability at the same frequency as precessional cycles in the Northern Hemisphere summer insolation record, but systemically lags behind July insolation at 65°N (Fig. 4). Multiple independent records and models support a role for precessional forcing in driving Asian summer monsoon intensity and
monsoon variation generally is regarded as a function of low latitude solar insolation\(^3\,7\). However, over geologic timescales the degree to which there is a direct, singular forcing response of the monsoon to precession, or one where multiple factors such as CO\(_2\) and sea level modulate a lagged EASM response, is unclear. Some authors advocate a direct response with no lag, based often on speleothem \(^{18}\)O records\(^3\,3\,4\), while others argue for a c. 8 ka lag compared to absolute annual maximum insolation, based mainly on marine records\(^3\,5\, 3\,8\). As previous studies only focus on the last glacial termination\(^3\,9\, 4\,0\), our results permit the first independently
dated analysis of multiple precessional cycle phase lags between EASM proxies and insolation forcing in the loess record, and provide an independent test of these conflicting hypotheses. A clear, consistent phase lag between 21st July 65°N isolation and the Jingbian EASM is seen across all transitions in our dataset (Fig. 4, Supplementary Table 2). The insolation lag calculation and the effect of different life-time averaged water content assumptions on this lag are outlined in Methods. While the size varies due to age model uncertainty, the average lag is 4.9 ka (s.e.m. = 0.7 ka, n = 9), which would increase to ~7 ka if the target reference curve for phase measurement is taken as the absolute maximum insolation curve, as suggested by Clemens et al.15. This is within uncertainties of the lag proposed from marine records such as the Arabian Sea summer monsoon stack11, and contrasts sharply with results from speleothem δ18O14. We argue that the observed lag is not related to delays in MS signal acquisition; both theoretical models and empirical evidence point to rapid oxidation/reduction response of iron oxides and formation of superparamagnetic minerals that enhance MS22–25. Although transmission of the forcing signal through the climate system may account for some of the lag, we also note that there is remarkable similarity between our independently dated MS record and global ice volume as represented in the LR04 stack23 (Fig. 3). The only exception is during MIS 7 where a peak in ice volume has no MS/EASM equivalent peak in our Jingbian record (Fig. 3). However this may be an artefact of preservation; the missing peak occurs at the point of increased sand content bracketed by a deeper glacial phase (Fig. 3). Given the larger absolute age uncertainty at this time point and the more scattered ages in the section this data set comes from (B, Fig. 2), it is quite possible an undetected erosional event has removed this peak.

While low latitude insolation directly drives monsoon variability at the precessional band37, the lagged MS record shows there cannot be a direct response at Jingbian; there must be other factors that heavily modulate the monsoon response in this region. This seems plausible given that the summer monsoon only penetrates as far north as Jingbian due to factors such as land–sea configuration37. As such, changes in this configuration due to ice volume would be expected to alter summer monsoon patterns at the site. The match between the MS record and the LR04 stack implies a response of the monsoon at Jingbian to insolation forcing that is similar to the response of the Northern Hemisphere ice sheets, potentially controlled by combined eccentricity, obliquity and precession, or alternatively that Northern Hemisphere ice volume dominates the forcing of the EASM28, 30. We suggest that variation in the EASM at Jingbian over the last 250 ka can be explained by combined insolation, ice volume, and CO2 forcing, supported by results from δ13C of loess organic matter, recent climate model simulations and many marine records35, 38, 40. Coupled, ocean–atmosphere–sea ice–land surface climate modelling of the last glacial monsoon suggests that atmospheric CO2 driven high latitude temperature changes drive latitudinal shifts in zonal circulation and the Intertropical Convergence Zone (ITCZ), in turn affecting monsoon precipitation40. These shifts would also have affected meridional temperature gradients, snow and ice cover on high ground, ice sheet dynamics, and hence global sea level (land–sea configuration), which in turn will also directly modulate summer monsoon circulation36, 46–49. Additional temperature forcing is driven by insolation at high latitude40. In monsoon marginal areas like Jingbian, such factors are likely to be the dominant control on summer monsoon dynamics, even if direct precessional forcing dominates monsoon intensity in core monsoon areas37. Variations in sea-level and CO2 forcing will alter the spatial extent and coverage of the summer monsoon, which will cause significant changes to precipitation levels at monsoon marginal sites, consistent with our record at Jingbian. As such, the previously widely accepted hypothesis of dominant direct low-latitude precessional forcing of EASM precipitation patterns seems increasingly implausible at the monsoon margin. Variation in monsoon proxies in various archives is consistent with this geographic effect with regard to monsoon forcing35, with high latitude forcing exerting a dominant control on monsoon precipitation patterns in monsoon marginal areas. Our data apparently conflict with some speleothem δ18O records of summer monsoon rainfall. However, reinterpretations of speleothem δ18O data suggest that either this proxy is not solely influenced by summer monsoon intensity or that δ18O is a function of integrated rainfall amounts between monsoon source and the cave site. If the latter is true it would imply this integrated rainfall was a function of low latitude precessional forcing. However, this is still consistent with our model as we would expect that integrated summer monsoon rainfall prior to precipitation at cave sites close to the southern part of the CLP would be dominated by low latitude precessional forcing, as this rainfall occurs dominantly in the core monsoon region. However, the extent of the summer monsoon in raining climate of the monsoon margins like on the north CLP, would still be dominantly controlled by the spatial extent and coverage of the summer monsoon, itself modulated by ice volume–sea level–CO2 forcing.

Methods

Study site. The study site is located in Jingbian County and comprises five sections (A, B, C, D, E; where >1 m of material was removed to freshly expose the sediment) (see Supplementary Fig. 1 where locations of individual sections are also given). The elevations of the individual sections were measured to within a few cm using differential GPS and our coordinates measured at section A were 37°29’52.8”N, 108°54’14.4”E. It should be noted that these are different to the coordinates for the Jingbian site by Ding et al.10. However, as we outline below, there appears to be an error in the site coordinates quoted in Ding et al.10 and we here demonstrate that in fact we are working on the same site; the ICS stratotype section. Firstly, coordinates for the stratotype site position subsequently given to us by E. Derbyshire are 37°29’58.7”N and 108°54’2.7”E (E. Derbyshire, personal communication 2015), with an elevation of ~1700 m above sea level (a.s.l.). Note that these coordinates refer to the position of a pylons 100 m to the immediate left of the proposed from marine records such as the Arabian Sea summer monsoon stack35. This is within uncertainties of the lag given by Ding et al.10 in which Derbyshire is a co-author. The coordinates provided by Derbyshire are also ~330 m from our differential GPS measured location of section A (see above), consistent with the position of the section on the east side of the gully ~300 m from the pylons (Supplementary Fig. 1). Furthermore, Ding et al.52 first presented the Jingbian section, which was subsequently analysed in Ding et al.10. Here they noted that the section was located near the settlement of Guojialiang. Indeed, the nearest settlement to both our sampling site and the revised coordinates provided by Derbyshire is Guojialiang. However, the coordinates given in Ding et al.10 provide a location ~40 km from Guojialiang, inside the Mu Us desert sand field, with this location also consistent with the site descriptions given in Ding et al.10, 52 and lacking any obvious gully exposure. During our fieldwork in a, local farmer confirmed that a group of Chinese scientists had worked previously at our sections D and E, and we could distinguish prior sampling (presumably for grain size and/or MS) at many sections within the gully. We therefore conclude that the coordinates given in Ding et al.52 are erroneous. Given the revised coordinates from Derbyshire and the match of our sections with the site descriptions and nearby settlements in Ding et al.10, 52 we are very confident that we were working at the same site as is described in Ding et al.10 and therefore the ICS stratotype site.

Luminescence dating. Samples for luminescence dating were collected by hammering stainless steel tubes (diameter 2.5 or 5 cm; length up to 25 cm) with a vertical spacing of 5–40 cm into freshly cleaned sediment profiles. The tubes were opened under subdued orange light at the Nordic Laboratory for Luminescence Dating (Aarhus University, DTU Risø campus, Denmark). The outer ~5 cm of each tube end was removed and reserved for dose rate analysis (see below). The inner material was wet-sieved to extract the 63–90 and 90–180 µm grain size fractions. These fractions were treated with HCl and H2O2 to remove carbonates and organic material, respectively. The fractions were etched for 20 min in 10% HF to remove feldspar. The purity of the quartz OSL signal was confirmed by the absence of a significant IRSL signal using the OSL IR depletion ratio. Both quartz and K-
feldspar rich fractions were mounted as multi-grain aliquots containing hundreds of grains on stainless steel cups.

All luminescence measurements were carried out using Riso TL/OSL DA20 luminescence readers equipped with calibrated 90Sr/90Y beta sources delivering between ~0.10 and ~0.20 Gy s⁻¹ to multi-grain aliquots in stainless steel cups. Quartz grains were stimulated using blue LEDs (470 nm; ~80 mW cm⁻²) and the OSL signal was detected through 7.5 mm of U-340 glass filter. Feldspar grains were stimulated using IR LEDs (870 nm; ~140 mW cm⁻²) with the OSL signal detected through a blue filter pack (combination of 2 mm BG-39 and 4 mm CN-7-59 glass filters). Single aliquot regenerative-dose (SAR) protocols were used to determine the quartz OSL and K-feldspar post-IR OSL equivalent protocols (Supplementary Table 1).

For the quartz measurements, a preheat of 260 °C (duration: 10 s) and cut heat to 220 °C was used; each SAR cycle ended with a high temperature (280 °C) blue light stimulation for 40 s. Natural, regenerative and test dose signals were measured at 125 °C for 40 s. The initial 0.00–0.32 s of the signal minus an earlier background (0.32–0.64 s) was used for dose calculation. Feldspar aliquots were preheated at 320 °C for 60 s for natural, regenerative and test dose signals. They were then stimulated twice with infra-red light for 200 s. The first IR stimulation temperature was 200 °C (IR signal) and the subsequent IR stimulation temperature was 290 °C (post-IR OSL signal, pIRIR200,290). The IR clean-out at the end of each SAR cycle was carried out at 325 °C for 200 s. The first 2 s of the post-IR OSL signal minus a background estimated from the last 50 s was used for dose calculation.

It is well-known that quartz OSL from Chinese loess is dominated by the fast component and generally behaves well in a SAR protocol. However, age underestimation is observed when doses >150 Gy are measured in loess using quartz SAR OSL. Therefore, we restricted the use of the quartz OSL signal to samples from the upper 480 cm in section D. Below this limit the quartz SAR OSL Dₜ values are ≤240 Gy and these results were not used for age modelling. Figure 5a shows the results of a preheat plateau test on sample D38141. It can be seen that over a wide temperature interval quartz Dₜ is independent of preheat temperature, recycling ratio is close to unity and recuperation is low. A dose recovery test using the SAR protocol outlined in Supplementary Table 1a was carried out on 10 samples from section D (D38102, 104, 106, 107, 108, 109, 110, 111, 112, 113) with given doses ranging between 10 and 50 Gy. Prior to giving the laboratory dose, the natural quartz OSL signal was reset by two blue light stimulations (100 s each) separated by a 10,000 s pause to allow any photo-transferred charge in the 110 °C TL trap to decay. The results of the dose recovery test are shown as a histogram and a measured to given dose plot in Fig. 5c, d, respectively. It can be seen from these results that our SAR protocol (preheat 260 °C/10 s, cut-heat 220 °C) is able to satisfactorily recover aliquots with relatively poor recycling ratios (e.g., deviating >10% from unity) and that the quartz Dₜ values are insensitive to the levels of feldspar contamination present in these extracts. The quartz Dₜ values for section D are tabulated in Supplementary Data 1.

Since the discovery of more stable post-IR OSL signals compared to conventional IRSL signals measured at ambient temperature, several SAR protocols have been developed to use IRSL to date beyond the range of OSL dating. Here we present more laboratory tests of the pIRIR200,290 signal from the coarse-grained feldspar extracts.

Figure 7a shows a first IR stimulation plateau and a multi-elevated temperature (MET) Dₜ plateau (using the protocol described by Li and Li) for the deepest sample in section B. The first IR stimulation plateau results suggest that an apparently stable pIRIR signal is reached when the first IR stimulation temperature is 2170 °C. This is consistent with the observations of Li and Li who showed that for Chinese loess samples with Dₜ values >~400 Gy, the pIRIR200,290 Dₜ values are greater than pIRIR200,290 Dₜ values. Unfortunately, we did not observe a plateau region in the MET-pIRIR data from this sample and this protocol was not considered further. Based on the first IR stimulation plateau, we chose the pIRIR200,290 signal as the preferred dating signal for this study. Three other Chinese loess sections have also been successfully dated using the pIRIR200,290 signal from polynminerol coarse silt grains and from sand-sized K-rich felspars.

Based on extensive laboratory testing, Yi et al. concluded that in pIRIR dating, it is advisable to check for the dependence of the results on test dose size. Figure 7b presents the dependence of the dose recovery test on test dose size for sample D38135 (sample also used in Boyer et al.2). The dose recovery test was carried out by adding beta doses on top of the natural dose in the sample. From these data we deduce that small test doses (~20% of the dose to be measured) should not be used when large (>300 Gy) doses are measured, in agreement with the observations of Yi et al.68, Colarossi et al.69 have shown that their sample at least part of this effect could be attributed to charge carry-over from Lₜ to Tₜ. Figure 7c presents another dose recovery test on bleached (24 h in Hönle SOL2 lamp) 90–125 μm feldspar-rich grains from sample D38146 (test dose was ~40% of dose of interest). The residual dose in this sample after bleaching was 9.9 ± 0.2 Gy (n = 3) and this value was subtracted from the measured doses. It can be seen that for doses up to at least 800 Gy, our chosen SAR protocol is able to satisfactorily recover a dose given in the laboratory. Based on these results, the test dose size for all Dₜ measurements was kept between ~30% and ~70% of the measured dose.

Post-IR OSL signals bleach at a much slower rate than the quartz OSL signal and there appears to be a residual very-hard-to-bleach (or un-bleachable) component present in the pIRIR200,290 signal which needs to be taken into account. Based on a long-term (>80 days) bleaching experiment, Yi et al.68 concluded that a constant (or very difficult to bleach) residual pIRIR200,290 signal amounting to ~6 Gy is reached after bleaching for ~300 h in a Hönle SOL2 solar simulator with a lamp-sample distance of 80 cm. Even though a residual dose of...
5–10 Gy in our samples only makes up between 3% and 5% of the lowest pIRIR200,290 \(D_e\) in our pIRIR200,290 age data set (sample D38155), we have estimated this component by comparing the feldspar ages of the upper 450 cm in section D \((n = 50)\) with young quartz OSL ages. This is because it has been shown that fast component dominated quartz OSL signals can record very small doses\(^72\) and for loess the residual quartz OSL dose at deposition has been shown to be negligible\(^70\).

Fig. 7d, can be seen that there is overall good agreement between feldspar pIRIR200,290 and quartz OSL ages but that there is a small pIRIR200,290 offset of 1.43 ± 0.47 ka. Translating this age offset into dose using the average feldspar dose rate.
of 3.5 Gy ka$^{-1}$ gives a residual dose of 5 ± 2 Gy. This dose was subtracted from all the pIRIR$_{200,290}$ $D_i$ values prior to calculation of the age (Supplementary Data 1).

Material from the outer end of the tubes was used for dose rate analysis. Samples were first ignited at 450 °C for 24 h, homogenised using a ring-grinder and finally cast in wax in a cup or disc geometry. After storage for >21 days to allow $^{222}\text{Rn}$ to build up to equilibrium with its parent $^{226}\text{Ra}$, they were counted for at least 24 h on one of the six gamma spectrometers from the Nordic Laboratory for Luminescence Dating (Aarhus University). The calibration of the spectrometers is described in Murray et al.72. The resulting $^{238}\text{U}$, $^{226}\text{Ra}$, $^{232}\text{Th}$ and $^{40}\text{K}$ concentrations are given in Supplementary Data 1. Note that for some analyses, the data for $^{226}\text{Ra}$ is not available due to limited sensitivity of some detectors; in this case, the $^{226}\text{Ra}$ value was used for the entire U series. Radiocarbon concentrations were converted into dry beta and gamma dose rates using the conversion factors of Guérin et al.74. During calculation of the infinite matrix dry dose rate, we assumed a $^{238}\text{U}$ retention factor of 0.80 ± 0.10 on the $^{238}\text{U}$ chain; at two standard deviations, this covers a range from no $\text{Rn}$ loss to 40% $\text{Rn}$ loss. Total dose rates were calculated using life-time average water contents of 10 ± 5 and 15 ± 5% (weight water/dry sediment weight) for loess and soil units, respectively (this deviation, this covers a range from no Rn loss to 40% Rn loss. Total dose rates were calculated using life-time average water contents of 10 ± 5 and 15 ± 5% (weight water/dry sediment weight) for loess and soil units, respectively (this assumption is discussed in more detail below). A small cosmic ray contribution to the dose rate was added based on Prescott and Hutton75.

For K-feldspar grains, we have added an internal beta dose rate based on a K concentration of 400 ± 100 ppm was also included. A lower internal alpha dose rate was also included. A lower internal alpha dose rate.

The effect on the insolation lag of different lifetime-averaged water content assumptions in luminescence dating is important in this study. Our choice of water content and its uncertainty is first discussed with respect to literature values and the relevance to individual samples is then considered using the section containing the Holocene soil (section D). We then investigate the dependence on different water content assumptions of the apparent lag between our luminescence dated MS records and the insolation record. Firstly, although there is some variability in the published water content values for Chinese loess (see discussion in Stevens et al.81), the values used in this study, of 15 ± 5% w.c. for soil and 10 ± 5% w.c. for loess layers, are similar to previous water content assessments for loess deposits from sites in the CLP.82 In addition, Chen et al.83 used a value of 10 ± 5% for a single sample collected in the S8 palaeosol at Jingbian.

We next consider the water content required to reduce the EASM lag to 0 ka for section D. For the two Holocene samples (D38132, w.c. 15% and D38136, w.c. 10%), this would require increasing the water content by 1.5% and ~3%, respectively. These water contents are 3 standard deviations from the values used and are close to saturation for sandy loess deposits. However, it is likely that the upper loess-palaeosol units at Jingbian have been well-drained since deposition; the gully is at least 10 ka old since the Holocene soil is inset into the gully system and the current water table is now around 300 m below the sampling level in a 280-m-deep gully system.10 The river into which the gully flows has incised into Pliocene red clay below the Quaternary loess. The age of this feature is unknown but is likely to be at least multiple glacial-interglacial cycles. It is thus expected that the upper loess-palaeosol units have remained at least several tens of metres above the water table for the majority and probably all of their burial lifetime. Thus, we consider it unlikely that the lifetime-average water content of this site could have approached the levels that would be required to reduce the lag to 0 ka. It is also worth noting that the water content values required for a zero lag would exceed almost all published values for even southern CLP sites, where precipitation levels are double those at Jingbian. If on the other hand, our estimates are too high, the dose rates would be higher, the luminescence ages lower and the lag with the insolation larger. Thus, in all likely water content scenarios, there is a significant lag between the EASM recorded in loess and insolation.

If we now make the additional assumption that the underlying mechanisms causing the insolation lag have not varied systematically with time (which ought to be safe given the lack of an obvious systematic change in ice volume and CO2, back in time at insolation inflection points), we would in turn expect the insolation lag to have remained constant within some bounds over the past ~230 ka. This is precisely what is observed in our data (Supplementary Table 2 and black symbols in Fig. 6). However, increasing the water content by only 5% (i.e., from 15% to 20% for soil and from 10% to 15% for loess) causes an increase of 4.7% and 4.1% in the quartz and feldspar ages, respectively. Recalculating the insolation lag using ages based on these higher water contents introduces a negative trend in the insolation lag versus insolation inflection point graph in Fig. 9 (red symbols). Indeed, using these higher water content estimates supports the physically unrealistic scenario that prior to ~130 ka the loess record of monsoon variability formed before the change occurred in the driving force (change in insolation). A similar
Fig. 9 Insolation lag as a function of insolation inflection point time for different water content assumptions. Dashed line is drawn at the average of the insolation lag data given in Supplementary Table 2 (black symbols) and shown in Fig. 4. Solid line indicates no lag

but positive trend in the size of the lag is observed when the water content is decreased by 5% (green symbols in Fig. 9). In summary, changing the water content by ±5% introduces trends in the lag with time and increases the standard deviation of the lags from the original ~2 ka (Supplementary Table 2) to ~3 ka. We conclude that our current water content assumption remains the most likely. It does not produce any systematic trend in the insolation lag with time and, if anything, the uncertainty on the water content has been overestimated.

Data availability
The data that support the findings of this research can be found in Supplementary Data 1 or upon request from the corresponding author.

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