

MIS 3 climate transitions revealed by high-resolution loess records from the eastern margin of the Tibetan Plateau

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ABSTRACT

Humans evolved rapidly and significantly between 60 and 20 ka, placing unprecedented demands on detailed climate evolution during the marine isotope stage (MIS) 3 period. In this study, a high-resolution loess record from the eastern margin of the Tibetan Plateau revealed that the climate was optimal in 62–43.5 ka, relatively stable in 43.5–34 ka, and continuously cooled in 34–20 ka. Such a three-stage pattern can be well correlated with changes in the temperature and CO₂ content in Antarctica, climate evolution in Europe, and advances in the Laurentide Ice Sheet. The climate transition at 43.5 ka was closely related to the declining boreal summer insolation, CO₂ content, temperature, and sea level, whereas the climate transition at 34 ka corresponded to a decrease in the CO₂ concentration and development of polar ice sheets. These results are of great scientific significance because they provide insights into MIS 3 climate transitions and human migration.

1. Introduction

Marine isotope stage 3 (MIS 3) is the period between 60 and 25 ka during which humans evolved rapidly and significantly on Earth (e.g., Roberts et al., 1990; Lambeck and Chappell, 2001; Mellars and French, 2011; Zhang et al., 2018). Humans were present in Lake Mungo in Australia 50–46 ka ago (Bowler et al., 2003), synchronously with or soon after the initial occupation of northern (Roberts et al., 1990, 1994) and western Australia (Turney et al., 2001). The well-known Neanderthals transitioned to modern humans, with a ten-fold population increase in Western Europe at ~40 ka (Mellars and French, 2011). In Asia, the earliest human occupied the high-altitude Tibetan Plateau (TP) 40–30 ka ago (Zhang et al., 2018). These place unprecedented demands on detailed climate evolution during the MIS 3 period.

Comprehensive models of sea-level change enabled the prediction of coastline migration during the last glacial cycle (Lambeck and Chappell, 2001). During the MIS 3 period, the sea level was between 60 and 90 m

below the present (e.g., Chappell, 2002; Siddall et al., 2003), and it was ~20 m higher during the first half of MIS 3 than during the latter half (Siddall et al., 2008). This sounds reasonable to have been driven by higher summer insolation in the Northern Hemisphere during the first half of MIS 3 (e.g., Huybers, 2006; Siddall et al., 2008; Berger et al., 2024). However, it remains enigmatic whether the MIS 3 climate transition should be divided into two or three stages.

Most studies suggest that the MIS 3 climate evolution can be divided into two stages, but transition timings differ. First, various sea-level reconstructions showed that the transition from early to late MIS 3 occurred at 45 ka (Siddall et al., 2008). Secondly, isotopically derived sea-surface temperatures (SSTs) from offshore northern New Zealand record average changes of +2 °C and – 2 °C for the 59–43 and 43–24 ka time periods, respectively (Wright et al., 1995). Third, loess in China is a unique terrestrial archive that documented the MIS 3 climate transition at 48 ka at several sites such as Hongtong in the Shanxi Province (Jiang et al., 2023), Lijiayuan (Ding et al., 1999), and Caotan (Sun and Huang,

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2006) in the northwestern Loess Plateau. Finally, stalagmite $\delta^{18}\text{O}$ records in China provide information about millennial-scale climate events (e.g., Wang et al., 2001, 2008), which did not show a clear timing point of climate transition during the MIS 3 period. Therefore, high-resolution records with reliable dates are required to explore MIS 3 climate transitions and possible forcing mechanisms in detail.

In this study, a high-resolution (< 100 yr) loess record from the eastern TP was studied and compared with data for other loess sections in China to investigate the MIS 3 climate transition. This work is of great scientific significance because it provides insights into the evolution of global climate and modern humans (e.g., Higham et al., 2011; Zhang et al., 2018; Sprafke et al., 2020).

2. Geographical and geological settings

The Sichuan Basin, with an elevation ranging from 450 to 650 m is surrounded by the high mountains of the eastern TP (~4500 m) to the west, E-W-trending Qinling Mountains (~1600–2000 m) to the north, and Daliang–Wu Mountains (1500–2000 m) to the south and east (Fig. 1). And it consists of low hills and alluvial plains. The basement of the basin is composed of highly metamorphosed pre-Cambrian rocks from the Yangtze Block and less-deformed Mesozoic rocks. Cretaceous red sandstone is widely distributed in the basin without Tertiary sediments. The Sichuan Basin is covered by silty clay sediments called Chengdu Clay or red earth, and its thickness varies from 2 to 10 m, with a maximum thickness exceeding 20 m (Thorpe and Dye, 1936; Liu, 1983). Geological and geochemical evidence, such as spatial variations in thickness, grain size, distributive features, Sr isotopes, oxygen isotope compositions of quartz particles and mineralogy, demonstrated the aeolian origin of Chengdu Clay in the Sichuan Basin (Feng et al., 2014).

Extensive loess deposits were developed in the interior and south of the Tibetan Plateau, which were mainly formed after the last deglaciation (Sun et al., 2007; Huang et al., 2022; Liu et al., 2023). The TP loess did not accumulate during the last glacial stage despite the presence of dust in the environment, but began depositing in the Holocene when the relatively warm and moist climates increased the vegetation density (Huang et al., 2022). The study area, located in the Sichuan Basin around the Tibetan Plateau, is a negative orography that could capture

dust and accumulate loess during the last glacial, which has also been evidenced by previous studies (Han et al., 2010; Feng et al., 2014; Li et al., 2023).

The Sichuan Basin remains little affected by the East Asian winter monsoon and westerlies during glacial periods (Feng et al., 2014) but is dominated by the Indian Summer Monsoon (ISM, Han et al., 2010; Huang and Cui, 2015). During summer, the ISM penetrates and brings considerable rainfall into the Sichuan Basin due to solar heating. The Sichuan Basin is characterized by a subtropical monsoon climate. According to data from the Chengdu Meteorological Administration (Fig. 2), the mean annual temperature (MAT) is 16.1 °C, with a July average of 25.2 °C and January average of 5.6 °C. The mean annual precipitation (MAP) is 870.2 mm, and over 74 % of the precipitation falls between June and September, with a peak mean rainfall of 224.5 mm in July. Local soils are subtropical yellow soils (luvisols), and most

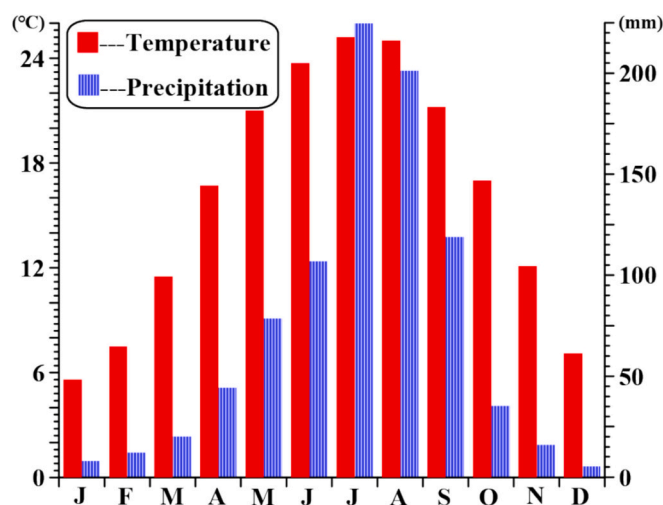


Fig. 2. Monthly changes of mean annual temperature and mean annual precipitation in the Sichuan Basin. Data from the Chengdu Meteorological Administration.

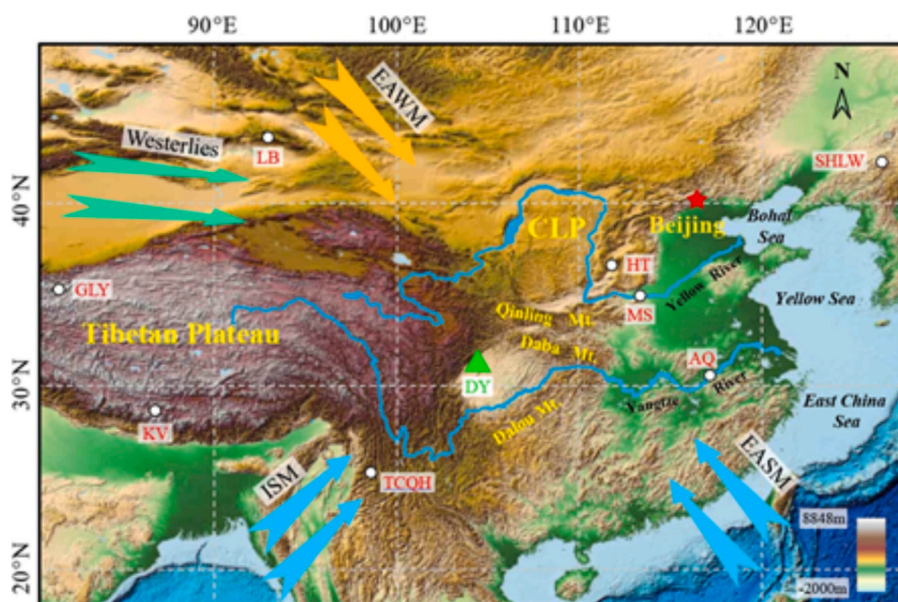


Fig. 1. Locations of the Deyang section (Sichuan Basin) and comparable sites in this study. Trajectories of the Indian Summer Monsoon (ISM), East Asian Summer Monsoon (EASM), East Asian Winter Monsoon (EAWM) and the westerlies are indicated by blue, orange and green arrows, respectively. SHLW, Sihailongwan Lake; LB, Lake Barkol; CLP, Chinese Loess Plateau; HT, Hongtong section; MS, Mangshan; DY, Deyang (this study); AQ, Anqing; TCQH, Tengchong Qinghai lake; KV, Kashmir Valley; GLY, Guliya ice core. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

of the original modern soils in the basin have been removed by erosion or changed by human cultivation.

The developed Quaternary clayey strata in the study area can be divided into three units from top to bottom (Zhang, 1988; Yang et al., 2010; Feng et al., 2014). The upper horizon is Chengdu Clay with brownish yellow in color (Munsell, 10YR 6/8), and is rich in calcareous concretions (maximum thickness: ~ 3.0 m). It forms a discontinuous blanket over the low hills and high river terraces in the western and northwestern Sichuan Basin (Thorp and Dye, 1936; Feng et al., 2014). The middle horizon is brown clay with a dominant dark brown (7.5YR 5/6) or brownish yellow (10YR 6/8) color (Feng et al., 2014). Its thickness ranges from 1.5 to 6.0 m, and is rich in ferromanganese nodules. The calcareous concretions that appear at the top of this horizon enclose the ferromanganese nodules. This syntagmatic relationship indicates that part of the carbonate is leached and transported downward from the overlying Chengdu Clay horizon and precipitates within the top brown clay horizon (Feng et al., 2014). The lower horizon consists of red (2.5YR 4/8) clays with a maximum thickness of ~ 6.0 m. Throughout the profile, the Chengdu Clay and the underlying clay horizons exhibit a clayey texture, a massive and blocky structure and a lack of visible bedding features. A reticulately-mottled structure is developed within the brown clay (middle horizon) and red clay (lower horizon). Recently, the Chengdu Clay is identified as Late Pleistocene based on fossils (Liu, 1983) and different dating methods (^{14}C and OSL, Chen and He, 1990; Han et al., 2010; Yang et al., 2010).

In this study, a well-exposed loess sequence is selected at Deyang ($31^\circ 14' \text{N}$, $104^\circ 25' \text{E}$, 513 ± 5 m a.s.l.), ~ 40 km away of Chengdu City, Sichuan Province (Fig. 1). The ~ 3.6 m thick section is located in a brickyard on the second terrace of the Mianyu River, and represents the uppermost part of Late Pleistocene loess deposits in the Sichuan Basin. In the field, the sequence exhibited homogeneous dust deposits with massive structures and vertical cracks (Fig. 3). No depositional discontinuities such as color mutation and gravel interlayer were observed. The sequence is divided into five units according to lithology, color and degree of weathering (Fig. 3).

Unit 1: 0–50 cm is cultivated soil with dark brownish color and

blocky structure, containing some plant roots.

Unit 2: 50–100 cm is loam soil with brownish gray color, massive structure, and vertical cracks.

Unit 3: 100–150 cm is homogeneous dust deposit with a grayish-yellowish color, massive structure, and vertical cracks.

Unit 4: 150–280 cm is weakly developed soil with brownish color and some Fe–Mn coatings.

Unit 5: 280–360 cm is moderately developed soil with dark brownish color and many Fe–Mn coatings.

In the laboratory, the grain size results with 0.5 cm spacing did not show significant mutation, like those occurring at Diexi and Lixian areas (Jiang et al., 2014, 2017). The severe discontinuous loess deposition was discovered in the northwest margin of the Loess Plateau mainly due to strong winds during the LGM period revealed (Stevens et al., 2006, 2007). Comparatively, the Sichuan Basin is a subtropical and relatively closed humid area without the influence of strong winds, and the nearby mountains continue to provide material sources, ensuring continuous deposition. This is also evidenced by the loess study near the study area (Han et al., 2010). Therefore, it is reasonable to consider the loess sequence as continuous deposition.

3. Methods

Nine samples were collected for optically stimulated luminescence (OSL) dating from the Deyang loess section at intervals of 20–35 cm (Fig. 3 and Table 1). According to the OSL dating results, 178 and 458 samples were collected for major element and grain-size analyses with time resolutions of 237 and 92 years, respectively.

For OSL dating, the samples were extracted under subdued red light and pretreated with 30 % HCl and 30 % H_2O_2 to remove carbonates and organic materials, respectively. Fine silt (4–11 μm) was then separated using sedimentation procedures based on Stokes' law. Polymineral fine grains were immersed in hydrofluorosilicic acid, H_2SiF_6 (30 %), for 5–7 days in an ultrasonic bath to obtain the fine-grained quartz component. The purity of the isolated quartz was verified using infrared (IR) stimulation. No significant IRSL signal was detected for the isolated quartz (Wang et al., 2006). All signals were measured with an automated OSL reader, equipped with a combined blue (470 ± 5 nm) and infrared (880 ± 80 nm) LED OSL unit, two $^{90}\text{Sr}/^{90}\text{Y}$ beta sources (0.19 and 0.157 Gy/s) and an Am-241 alpha source ($0.062 \mu\text{m}^{-2} \text{s}^{-1}$) for irradiation. The 4–11 mm quartz samples were measured using a sensitivity-corrected multiple aliquot regenerative-dose (SMAR) protocol (Wang et al., 2005; Lu et al., 2007). The U, Th, and K concentrations were determined using neutron activation analysis (NAA). The water content was calculated from the sample weights before and after oven drying at 105°C for 24 h (Table 1). The contribution of cosmic rays to the dose rate was calculated from the burial depth and altitude of the samples (Prescott and Hutton, 1994).

Subsamples for major element analysis were finely ground in an agate mortar before acid dissolution. To remove organic matter and carbonates and determine the chemical composition of the silicate minerals, a part of each sample (10–15 g) was ground and immersed in 40 mL of 20 % H_2O_2 at room temperature for more than 5 h. Subsequently, 40 mL of 20 % acetic acid (HAc) solution was added, and the suspension was occasionally shaken for 2 h to dissolve the carbonates. To remove all carbonates, 40 mL of a 20 % HAc solution was added for more than 12 h. After digestion, acid-soluble and acid-insoluble (AI) fractions were separated via centrifugation at 2500 rpm for 5 min. AI phases were washed two times with deionized water, separated by centrifuging, and dried at 50°C . The AI samples were then ground to less than 250 mesh ($< 61 \mu\text{m}$) and weighed. Approximately 4 g of each sample was used to determine the major element composition. Major element abundances were determined using a PANalytical B.V. Axios Advanced X-ray Fluorescence Spectrometer at the Institute of Geophysical and Geochemical Exploration, Chinese Academy of Geological Sciences. Approximately 0.6 g of dry-ground samples were

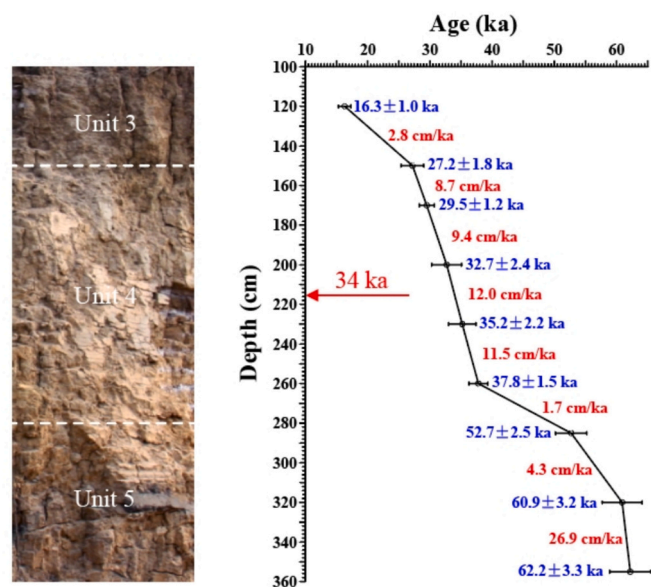


Fig. 3. Photograph of the Deyang section in the northwestern Sichuan Basin, Southwest China, and distribution of sampling points for OSL age control. Note that the bottom part of the section appears much redder than the middle and upper parts, and the deposition rate maintains high values during 37.8–27.2 ka. The red arrow indicates the stratigraphic position corresponding to 34 ka. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 1
OSL dating results for samples collected from the Deyang section.

Sample	Depth (cm)	Water content (%)	U (ppm)	Th (ppm)	K (%)	Equivalent Dose (Gy)	Dose rate (Gy/ka)	OSL Age (ka)
1-355	355	17.0 ± 2	3.03 ± 0.11	14.5 ± 0.42	1.67 ± 0.05	247.15 ± 8.8	3.97 ± 0.16	62.2 ± 3.3
2-320	320	17.6 ± 2	3.16 ± 0.12	14.9 ± 0.43	1.56 ± 0.05	239.26 ± 8.3	3.93 ± 0.16	60.9 ± 3.2
3-285	285	15.5 ± 2	3.64 ± 0.13	14.5 ± 0.42	1.49 ± 0.04	212.42 ± 5.4	4.03 ± 0.16	52.7 ± 2.5
4-260	260	14.7 ± 2	3.28 ± 0.11	14.4 ± 0.42	1.66 ± 0.05	154.11 ± 1.1	4.09 ± 0.16	37.8 ± 1.5
5-230	230	14.6 ± 2	3.41 ± 0.12	14.8 ± 0.43	1.75 ± 0.05	150.01 ± 7.4	4.26 ± 0.17	35.2 ± 2.2
6-200	200	12.7 ± 2	3.60 ± 0.13	14.3 ± 0.41	1.75 ± 0.05	141.53 ± 8.5	4.33 ± 0.17	32.7 ± 2.4
7-170	170	13.3 ± 2	3.61 ± 0.13	14.8 ± 0.43	1.84 ± 0.05	131.86 ± 1.8	4.47 ± 0.18	29.5 ± 1.2
8-150	150	16.1 ± 2	3.22 ± 0.12	15.6 ± 0.44	1.69 ± 0.05	114.64 ± 5.9	4.21 ± 0.17	27.2 ± 1.8
9-120	120	16.0 ± 2	2.70 ± 0.10	14.7 ± 0.43	1.72 ± 0.05	64.98 ± 3.2	3.99 ± 0.16	16.3 ± 1.0

OSL: optically stimulated luminescence. Sensitivity-corrected multiple aliquot regenerative-dose (SMAR) protocol: Six aliquots were used to establish the standard growth curve and another six aliquots were averaged to calculate the equivalent dose.

mixed with 6 g of $\text{Li}_2\text{B}_4\text{O}_7\text{-Li}_2\text{CO}_3$ fusion reagent in Platinum crucibles, heated to 1100 °C in a furnace, and cooled down to form a glass disc for the major element analysis. Analytical uncertainties were less than 3 % for the major elements (SiO_2 , Al_2O_3 , CaO , $\text{Fe}_2\text{O}_{3(\text{Total})}$, K_2O , TiO_2 , Na_2O , P_2O_5 and MnO). The loss on ignition was determined by weighing after 1 h of calcination at 950 °C.

At the State Key Laboratory of Earthquake Dynamics, Institute of Geology, China Earthquake Administration, the grain size of 458 samples from the Deyang section was analyzed with a time resolution of 92 years using a computer-operated Malvern Mastersizer 3000 laser grain-size analyzer. Approximately 0.2 g of sediment was pretreated with 20 mL of 30 % H_2O_2 to remove organic matter. Subsequently, the sample was treated with 10 mL of 10 % HCl to remove carbonates. The sample residue was dispersed in 10 mL of 0.05 M (NaPO_3)₆ on an ultrasonic vibrator for 10 min before grain-size measurement. The grain-size analyzer automatically outputs the percentages of related size fractions of a sample with a relative error of less than 1 %.

Redness measurements were conducted on all grain-size samples. For redness (a^*) measurements, the sample (~1.5 g) was first dried at 40 °C

for 24 h and then crushed into dust without changing the grain size. Finally, the reflectance was measured using a handheld SPAD-503 spectrophotometer.

4. Results

4.1. OSL dating

Both the growth and decay curves show that the OSL signal was strong and significantly decayed within the first two second of stimulation, suggesting the dominance of fast components (Fig. 4). For growth curve construction, the OSL signal was not saturated at the level of the natural signal and the recuperation was close to zero (Fig. 4), implying reliable dating results.

Noticeably, three samples at the base of the Deyang section exhibited D_e values >200 Gy, which may indicate saturation and age underestimation, as expressed in eastern China (Lai et al., 2010). However, several pieces of evidence support the reliability of the dating results. First, D_e values of sample 1-355, 2-320, 3-285 are 247.15 ± 8.8 Gy,

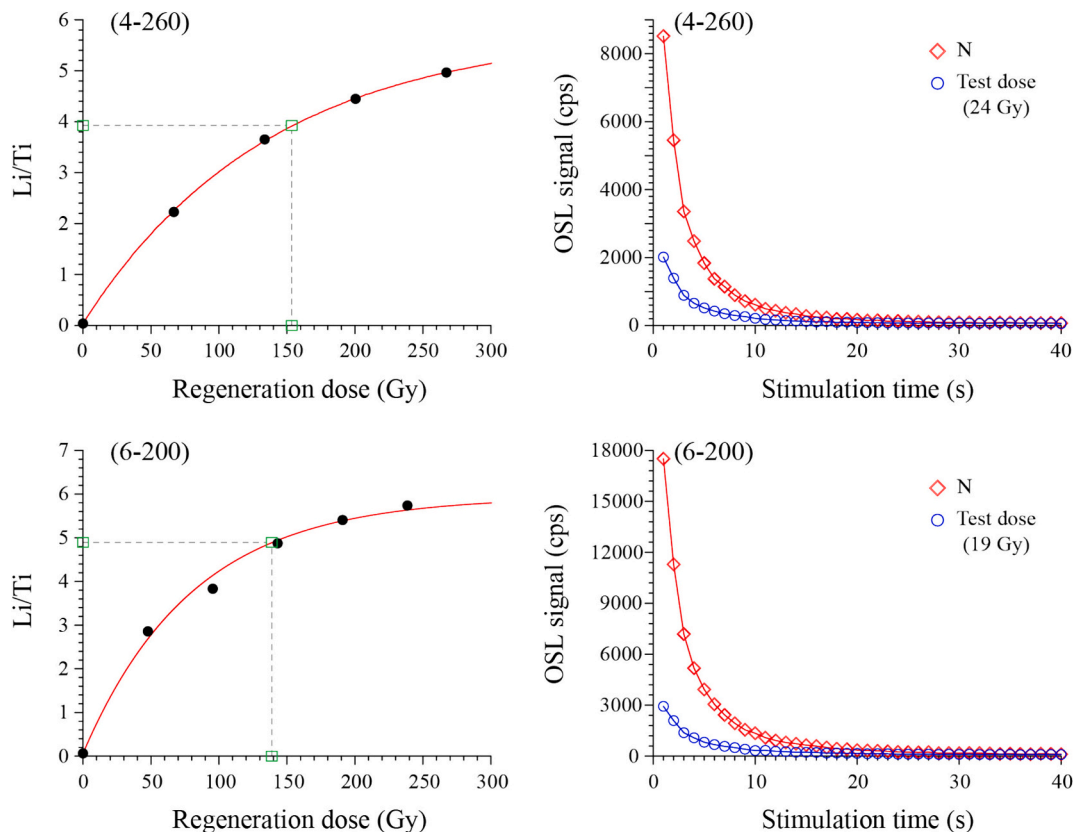


Fig. 4. Growth and luminescence decay curves for two representative samples from the Deyang section of the northwestern Sichuan Basin.

239.26 ± 8.3 Gy, 212.42 ± 5.4 Gy, respectively, which are less than $2D_0$ (344.12 Gy, 328.96 Gy, 286.38 Gy). These indicate that they are not saturated and are relatively reliable (Wintle and Murray, 2006; Lai, 2010). And Chinese loess dating allowed reliable equivalent dose (D_e) determination up to at least 400 Gy (Lai, 2010). Second, the dating results of the Deyang section exhibit a good linear relationship with the stratigraphic depth (Fig. 3). Third, the map of linear relationship shows a rapid dust deposition (8.7–12.0 cm/ka, mean 10.4 cm/ka) interval (37.8–27.2 ka), which correlates in timing with the continuously high dust concentration (0.63–20 μm) revealed by the Guliya ice core on the northern TP during 37.5–27.5 ka (Thompson et al., 1997). Finally, the peak values of the superfine fraction, clay, and chemical index of alteration (CIA; Fig. 6a, b, d) in this study at ~ 56.5 ka correlate with high $\delta^{18}\text{O}$ values in the Guliya ice core (Fig. 6f, Thompson et al., 1997) and insolation (Fig. 6k–l, Berger and Loutre, 1991; Berger et al., 2024). Therefore, nine OSL ages established a reliable timescale for the Deyang section in the Sichuan Basin.

We acknowledge the ongoing debate regarding age uncertainties of the Guliya ice core (e.g. Cheng et al., 2012; Hou et al., 2019; Tian et al., 2019a; Zhong et al., 2019; Hou, 2022; Wang et al., 2022), but this discussion about the age of the Guliya ice core is beyond our attention. The precise comparison of multi-proxies between the Deyang section and the Guliya ice core well supports the dating reliability of the Guliya ice core (Thompson et al., 1997).

4.2. Grain size, redness, and geochemical records

The alluvial plain and alluvial-proluvial fan in the northwestern Sichuan Basin regions provide the source of the loess deposits in the basin, which are transported by an ancient katabatic wind over a short distance (Yang et al., 2010; Feng et al., 2014). The variations of grain size in loess sequence primarily reflect changes in sedimentation dynamics. Further, dust particles are affected by in-situ alteration after sedimentation, such as loessification (Sprafke and Obrecht, 2016) and pedogenesis (Sun et al., 2011; Sprafke et al., 2020). Pedogenesis in loess is linked to changes in the abundance of the superfine component (<1 μm , Sun et al., 2011), and mainly form in situ (Sprafke et al., 2020). Pedogenic hematite formation linearly correlates with redness (a^* ; Schwertmann et al., 1982) and depends on the pedogenic temperature (Gao et al., 2018). Therefore, grain size and redness measurements are often utilized to reveal rapid climate change during dust sedimentation (Porter and An, 1995; Liu and Ding, 1998; Ding et al., 1999; Gao et al., 2018; Sprafke et al., 2020).

Understanding the complex pedosedimentary evolution is prerequisite for using proxies to reconstruct the paleoenvironment (Sprafke et al., 2020). The top two units (0–100 cm) of Deyang section are the eluviated and illuviation horizons, which are seriously affected by human cultivation and microbial disturbance. And they were abandoned for palaeoenvironmental reconstruction. The following three units (100–360 cm) show homogeneous sediments with massive structure and color gradients associated with pedogenesis, containing a wealth of environmental information. The grayish-yellowish unit 3 may be the result of loessification under cold-dry climatic conditions during the last glacial maximum (Sprafke and Obrecht, 2016). From unit 5 to unit 4, the brownish color became lighter as the soil developed weaker (Fig. 3). The sedimentary proxies (superfine composition and redness) also show corresponding changes, which are mainly related to the in-situ pedogenesis after dust deposition (Sun et al., 2011; Sprafke et al., 2020). The whole profile does not show the characteristics of reworking and erosion, and pedogenesis occurred immediately after dust deposition without traces of downward migration. Correspondingly, the grain size and redness records show long-term trend changes without any signs of reworking or erosion. In this way, we combine a reliable chronological framework with continuous multi-proxies for palaeoenvironmental reconstruction.

The grain size record of the Deyang section was characterized by the

dominance of silt particles (mean 81.0 %). In contrast, clay (<2 μm) and sand (>64 μm) have low average contents of 16.4 % and 2.5 %, respectively (Fig. 5b–e). For the silt fraction, grains of 2–20 μm are most abundant with an average of 54.7 % (Fig. 5c), corresponding well to the feature of dominance of 0.63–20 μm fraction dust concentration in the Guliya ice core on the northern TP (Thompson et al., 1997). According to the varying grain size, redness, and major elements in Deyang loess records, three climatic evolution stages can be classified that correlate well with the pedogenesis features observed in the field (Fig. 3).

4.3. Early stage of MIS 3 (62.5–43.5 ka)

This stage witnessed the highest values of the superfine proportion (Fig. 5a), clay (Fig. 5b), and redness (Fig. 5f) of the entire sequence, reflecting the strongest pedogenesis observed in the field (Fig. 3) and the highest temperature of the MIS 3 period. The CIA is often used as an indicator of the degree of sediment weathering and is calculated in molecular proportions as follows: $\text{CIA} = [\text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{CaO}^* + \text{Na}_2\text{O} + \text{K}_2\text{O})] \times 100$ (Nesbitt and Young, 1982), where the CaO^* is the amount of CaO in the silicate minerals. During chemical weathering, the Ca, Na, and K in feldspar (main silicate) leach more easily than Al. With the enhancement of weathering, the contents of Ca, Na and K in the weathering products decrease, whereas the contents of Al remain relatively unchanged. Therefore, an increase in the CIA value indicates an enhancement in chemical weathering and climate amelioration (Meyer-Heintze et al., 2018). Similarly, Na_2O and Al_2O_3 can be used as indicators of chemical weathering because the mother rock undergoes chemical weathering in which unstable Na^+ is leached and the stable cation Al^{3+} remains unchanged and relatively enriched (Ding et al., 2001; Jiang et al., 2022). The content of Na_2O shows the lowest values (Fig. 5h), whereas Al_2O_3 (Fig. 5g) and CIA (Fig. 5i) present the highest values for the entire section, which agrees with the highest clay content during this period (Fig. 5b), suggesting the strongest weathering. Such an optimal MIS 3 climate in the early stage correlates well with the contemporary Hongtong loess record in the Shanxi Province (Jiang et al., 2023), reflecting the strongest summer monsoon during this stage. Moreover, it strongly corresponds with the high boreal summer insolation during early MIS 3 (Fig. 6k–l).

Approximately 62.5 ka ago, the superfine fraction, clay, Al_2O_3 content, and CIA values started to increase gradually; they reached their maxima at 56.5 ka ago (Fig. 5a, b, g, i), whereas the Na_2O content gradually decreased and reached its minimum at 56.5 ka ago (Fig. 5h). This is not only consistent with the gradual increase in temperature from 62 ka and to its maximum at 56.5 ka revealed by the Guliya $\delta^{18}\text{O}$ record (Fig. 6f, Thompson et al., 1997), but also agrees well with the occurrence of MIS 4 to MIS 3 transition in sea level between 60 and 57 ka (Siddall et al., 2008).

From 56.5 to 43.5 ka, the superfine fraction, clay, Al_2O_3 , and CIA gradually declined (Fig. 5a, b, g, i). They rapidly decreased from 48 to 43.5 ka in the Sichuan Basin. These can be well correlated with the Hongtong redness record in the Shanxi Province (Jiang et al., 2023), seemingly driven by the rapid decrease in boreal summer insolation (Fig. 6k and l).

4.4. Middle stage of MIS 3 (43.5–34.0 ka)

During this stage, the Deyang multi-proxies of the superfine fraction, clay, a^* , Al_2O_3 , Na_2O , and CIA slightly changed, with small amplitudes (Fig. 5), indicating an overall stable climate and environment in the study area. However, the data points became much denser 37.5 ka ago (Fig. 5a–f), which is an instantaneous response to the increases in both the deposition rate in the study area (Fig. 3) and Guliya dust concentration (0.63–20 μm) on the northern TP (Fig. 6g, Thompson et al., 1997). This is also consistent with the environmental decline of the Bering Strait region since 37 ka (Shapiro et al., 2004) and the rapid expansion of the Labrador Dome at 38 ka (Dalton et al., 2022). At ~ 38

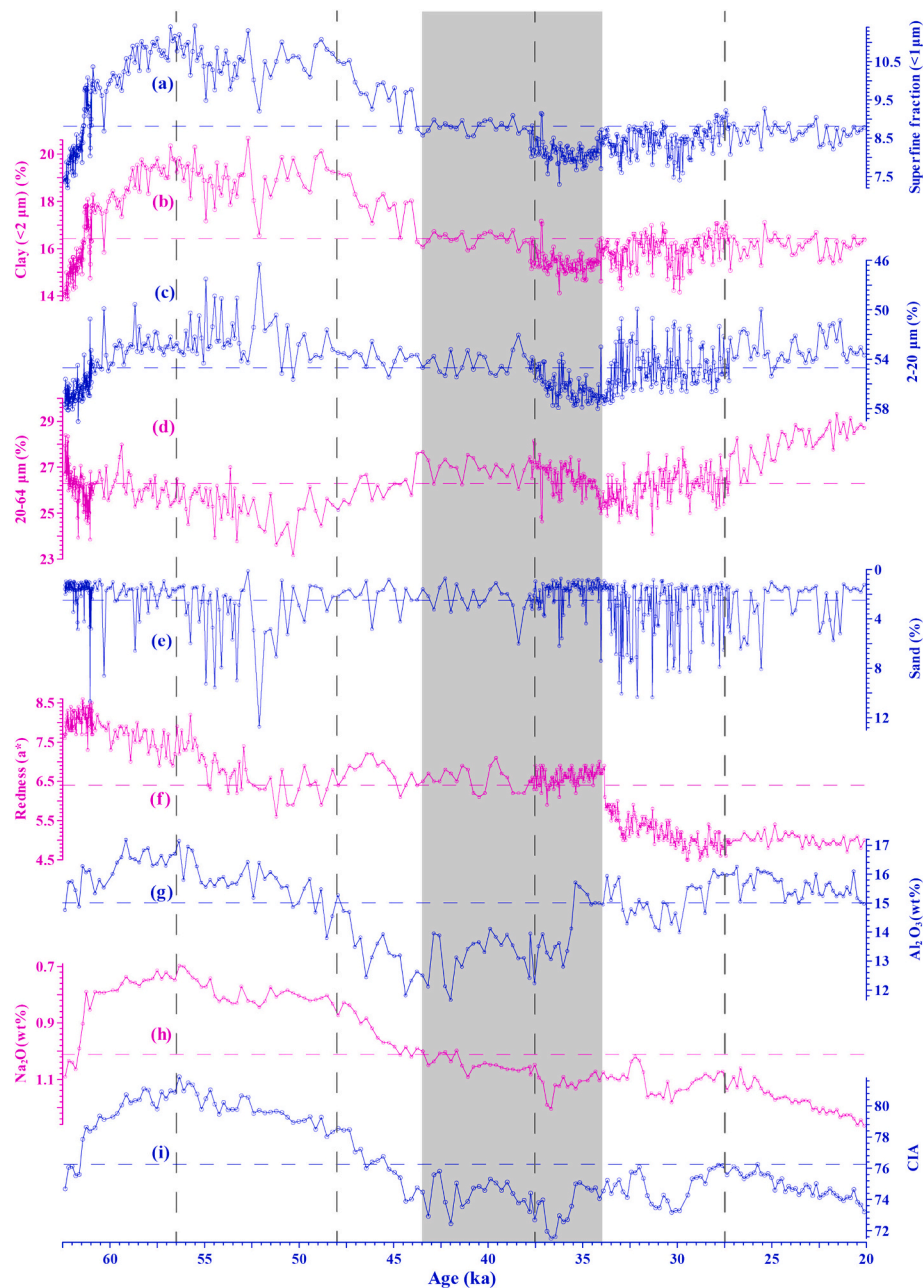


Fig. 5. Five grain-size fractions (superfine fraction, clay, 2–20 μm , 20–64 μm , and sand) of the Deyang section and their correlation with redness and geochemical records. Horizontal dashed lines indicate average values. Vertical dashed lines indicate important shifts discussed in the text. The gray belt indicates a relatively stable middle MIS 3 climatic period.

ka, both temperature (Jouzel et al., 2007) and CO_2 concentration (Bereiter et al., 2012) recorded by the Antarctic ice core decreased significantly (Fig. 6h–i). These suggest that significant climate decline at 37.5 ka could have been driven by some common factors. Obviously, both 65°N and 31°N insolation increased continuously at 37 ka though with relatively small amplitudes (Fig. 6k–l). These suggest that the growth of polar ice sheets and the decline in atmospheric CO_2 could have played an important role in global cooling and widespread drought and, consequently, a significant increase in atmospheric dust deposition (Petit et al., 1981; Lambert et al., 2008). These might have been responsible for the significant increases in both the Guliya dust concentration (Thompson et al., 1997) and deposition rate in the Sichuan Basin (Fig. 3). These deductions are supported by the continuous increase in the 2–20 μm fraction (Fig. 5c) instead of a continuous decrease in the superfine fraction and clay (Fig. 5a–b) and the generally low CIA

values below the average line (Fig. 5i) in the Deyang section from 37.5 to 34 ka.

4.5. Late stage of MIS 3 (34.0–20.0 ka)

From 34 to 20 ka, the coarse silt fraction (20–64 μm) in the Deyang section increased continuously (Fig. 5d), the fine silt fraction (2–20 μm , Fig. 5c) and redness (Fig. 5f) decreased continuously, and the CIA values remained below the average line (Fig. 5i).

This stage can be separated into an early stage (34–27.5 ka) and a late stage (27.5–20 ka). During the early stage, the redness decreased sharply (Fig. 5f), whereas the fine silt, coarse silt, and sand fraction strongly fluctuated (Fig. 5c–e), reflecting an unstable climate, probably caused by glacial cooling. This is supported by the high dust concentration documented by the Guliya ice core on the northern TP (Fig. 6g;

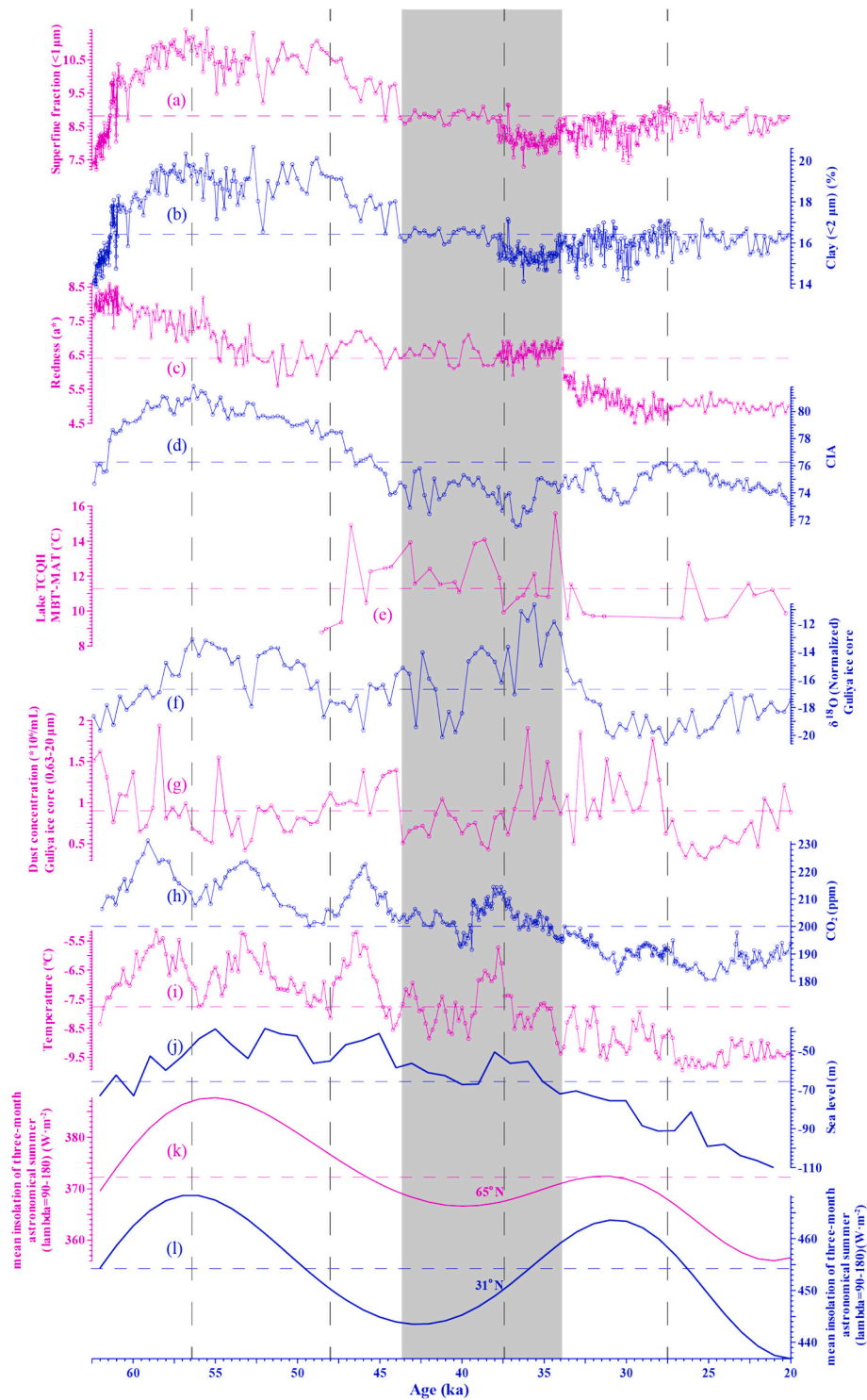


Fig. 6. Correlation analysis of superfine fraction, clay, redness, and CIA of the Deyang section (a, b, c, and d), the biomarker-derived temperature record of Lake TCQH in the Yunnan Province (e; [Tian et al., 2019b](#)), the Guliya ice core $\delta^{18}\text{O}$ -derived temperature and dust concentration records (f and g; [Thompson et al., 1997](#)), CO_2 and temperature changes documented by the Antarctic ice core (h and i; [Jouzel et al., 2007](#); [Bereiter et al., 2012](#)) and sea level estimates of a stack of 57 benthic $\delta^{18}\text{O}$ records (j; [Lisiecki and Raymo, 2005](#)), and mean insolation of astronomical summer at 65°N and 31°N (k and l; [Berger and Loutre, 1991](#); [Berger et al., 2024](#)).

[Thompson et al., 1997](#)) and strong temperature fluctuations recorded in the Antarctic ice core ([Jouzel et al., 2007](#)) during this period.

During the late stage (27.5–20 ka), the superfine fraction, clay, fine silt fraction, sand fraction, and redness in the Deyang section tended to stabilize ([Fig. 5a–c, e, f](#)), although the coarse silt fraction continued to increase significantly ([Fig. 5d](#)) and the CIA value continued to decrease ([Fig. 5i](#)).

5. Discussion

In this study, the high-resolution well-dated loess record of Deyang section for the first time revealed that the MIS 3 period can be clearly divided into three stages ([Fig. 5](#)): optimal climate in the early stage, relatively stable climate in the middle stage, and continuous cooling in the late stage. Such a three-stage pattern strongly correlates with

changes in the temperature and CO₂ content documented by the Antarctic ice core (Jouzel et al., 2007; Bereiter et al., 2012) and sea level estimates of a stack of 57 benthic $\delta^{18}\text{O}$ records (Lisiecki and Raymo, 2005). The temperature, CO₂ content, and sea level during MIS 3 were above their averages during the early stage, close to their averages during the middle stage, and continuously decreased below their averages during the late stage (Fig. 6h–j). Accordingly, we focused on two important climate transitions at 43.5 and 34 ka, which are of great scientific significance because they provide insights into the MIS 3 global climate transition and human migration.

5.1. Early-middle MIS 3 transition (43.5 ka)

Both the geophysical (superfine fraction, clay, redness) and geochemical (CIA, Na₂O, Al₂O₃) proxies in Southwest China reveal that the early to middle MIS 3 climate transition occurred at 43.5 ka (Fig. 5), which was characterized by a significant onset of fluvial sediments in Anqing (Anhui Province) in the lower Yangtze River Valley (Wu et al., 2024). The dust concentration recorded in the Guliya ice core from the northern TP decreased abruptly at 43.5 ka, distinctly separating the early and middle stages (Fig. 6g; Thompson et al., 1997).

Among the stalagmite $\delta^{18}\text{O}$ records in South China (e.g., Wang et al., 2001; Cai et al., 2006, 2015) and the organic carbon records of borehole 111 KL in the Indian Ocean (Schulz et al., 1998), Heinrich (H) and Dansgaard-Oeschger (D-O) events strongly correlate. The 43.5 ka transition is located at the cooling time interval between decadal-scale warming D-O events 11 and 12 and is submerged in numerous cooling events (Wang et al., 2001). However, the event frequency before the 43.5 ka transition (early MIS 3) was significantly lower than that after the transition (middle and late MIS 3).

Pollen records from France (Guiot et al., 1989) and Italy (Rossignol-Strick and Planchais, 1989) show that the climate was optimal in the early MIS 3 period, which ended at 43 ka. The recent synthesis of newly available ages shows that the most significant advance of the Laurentide Ice Sheet reached its terminus near 42° N during 45–40 ka with a strong peak at 42 ka (Kerr et al., 2021).

In Southeast Australia, the abundance of plants with the C₄ carbon fixation pathway was generally high (60–70 %) from 58 to 44 ka. By 43 kyr ago, the abundance of C₄ had dropped to 30 %, and biomass burning had increased (Lopes dos Santos et al., 2013). Similarly, in northern New Zealand, both isotopically derived SSTs and pollen records reveal average changes of +2 and – 2 °C for the 59–43 and 43–24 ka time intervals, respectively (Wright et al., 1995).

The mean summer insolation was relatively low at both 65° N and 31° N and reached a minimum at 31° N at 43.5 ka (Fig. 6k–l), corresponding well with the 43.5 ka transition observed in our loess records. Furthermore, the CO₂, temperature, and sea level decreased significantly at ~44 ka (Fig. 6h–j). Therefore, we propose that the climate transition at 43.5 ka was related to the combined effect of declining boreal summer insolation, CO₂ concentration, and global temperature.

Our recent loess record from the Hongdong section in the Shanxi Province indicates a unique cooling period of ~48 ka (Jiang et al., 2023), which is readily observed in the Lijiayuan (Liu and Ding, 1998; Ding et al., 1999) and Caotan (Sun and Huang, 2006) loess records at the same time and latitude (36° N). As the latitude moved southward, the cooling in the Sichuan Basin was possibly delayed to 43.5 ka. This is consistent with the timing and route of human migration from northern Asia to southern Tibet (Zhang et al., 2022). Northern Asia first experienced cooling at ~48 ka, and humans began migrating south to the Tibetan Plateau for survival (Zhang et al., 2022), where the climate was optimal until 43.5 ka. Furthermore, compared with the vague 40–30 ka (Zhang et al., 2018), the specific climate transition points (43.5 and 34 ka) proposed in this study provide a detailed climate evolution background for human migration during MIS 3. This will facilitate the detailed examination of human migration and evolution in different regions in the future.

5.2. Middle-to-late MIS 3 transition (34 ka)

At 34 ka, the Deyang loess record not only indicates a sharp and subsequently gradual decline in the redness value (Fig. 5f) as well as a continuous increase in the coarse silt fraction (20–64 μm ; Fig. 5d) and continuous decrease in the fine silt fraction (2–20 μm) since then (Fig. 5c). Both the biomarker-derived temperature record from Lake Tengchong Qinghai in the Yunnan Province (Fig. 6e; Tian et al., 2019a) and the Guliya $\delta^{18}\text{O}$ -derived temperature record (Fig. 6f; Thompson et al., 1997) significantly decreased at 34 ka. A decrease exceeding 5 °C was estimated from the fitting coefficient between temperature and oxygen isotopes (Schotterer et al., 1997), which may be associated with the amplifying effect of the TP with increasing height (Schotterer et al., 1997; Thompson et al., 2018). In subtropical regions across China, more peatlands were initiated from 41 to 34 ka and abruptly decreased to 34 ka (Zhao et al., 2014). Continuous fluvial sedimentation abruptly ended at 33 ka in Anqing (Anhui Province) in the lower Yangtze River Valley (Wu et al., 2024). In the Xinjiang Uygur Autonomous Region (Northwest China), an evaporation salt record from Lake Barkol shows that the total carbonate content significantly decreased at 34 ka and the lake water has deepened since then (Gu et al., 1998), which might be probably related to the moisture increase caused by global cooling and consequent strengthening of westerly circulation (Chen et al., 2008; Guo et al., 2022). The investigation of a Kashmir loess section in the western Himalayas showed that the CaO concentration increased from 34 to 15 ka, whereas CIA decreased, suggesting the retention of CaCO₃ in the soil due to dry climatic conditions (Shah et al., 2021).

The Laurentide Ice Sheet was the largest global ice mass to grow and decay during the last glacial cycle and showed gradual growth toward the last glacial maximum (LGM) starting at ~35 ka (Dalton et al., 2022). The ice advance known as the Lehigh Advance, occurred between ~34 and ~30 ka and is associated with the Keewatin Dome (Kerr et al., 2021). Previous reports of the local extinction of brown bears (Barnes et al., 2002) and hemionid horses (Guthrie, 2003) in Alaska from ~35–32 ka support the possibility of a large-scale environmental change affecting the populations of large mammals.

In southern Italy, *Quercus* and *Artemisia* pollen abundances became minimal after 33 ka, and Poaceae was the prevailing herbaceous pollen taxon (Rossignol-Strick and Planchais, 1989). Loess-paleosol sequences from NE Austria revealed that a marked transition from oxidized to reduced paleosols occurred at 34–35 ka, corresponding to the Middle-to-Upper Pleniglacial transition as part of a general cooling trend (Sprafke et al., 2020). The transition at 34 ka in our loess records is of broader significance, and it is particularly consistent with the growth of global ice sheets to their maxima that occurred between 33.0 and 26.5 ka and significant decreases in atmospheric CO₂ concentrations and the sea level (Fig. 6h and j). As summer insolation at both 65° N and 31° N showed an increasing trend at 34 ka (Fig. 6k–l), we propose that the 34 ka climate transition in our records is related to a decrease in atmospheric CO₂ contents and global cooling.

6. Conclusion

Multiple proxies of a high-resolution loess sequence from the eastern margin of the Tibetan Plateau revealed that the climate was optimal in 62–43.5 ka, relatively stable in 43.5–34 ka, and continuously cooled in 34–20 ka. Such a three-stage pattern can be well correlated with changes in the temperature and CO₂ content in Antarctica, climate evolution in Europe, and advances in the Laurentide Ice Sheet. The climate transition at 43.5 ka is closely related to the combined effect of the declining boreal summer insolation, CO₂ concentration, temperature, and sea level, whereas the climate transition at 34 ka corresponds to the combined decrease in atmospheric CO₂ contents and the significant development of polar ice sheets. These results are of great scientific significance because they provide insights into MIS 3 climate transitions and human migration.

CRediT authorship contribution statement

Xiaotong Wei: Writing – review & editing, Writing – original draft, Visualization, Methodology, Investigation, Formal analysis, Data curation. **Hanchao Jiang:** Writing – review & editing, Writing – original draft, Resources, Project administration, Methodology, Investigation, Funding acquisition, Conceptualization. **Youliang Bai:** Validation, Methodology, Investigation. **Wei Shi:** Validation, Methodology, Investigation. **Hongyan Xu:** Validation, Methodology, Investigation. **Xiaolin Ma:** Methodology, Investigation. **Ning Zhong:** Methodology, Investigation. **Shenglin Li:** Methodology. **Qiuzhen Yin:** Writing – review & editing, Validation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Data availability

All new data used in this study are available in Zenodo (<https://doi.org/10.5281/zenodo.14722508>).

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