

Article

Temperature Variations and Possible Forcing Mechanisms over the Past 300 Years Recorded at Lake Chaonaqiu in the Western Loess Plateau

Keke Yu ^{1,2}, Le Wang ¹, Lipeng Liu ¹, Enguo Sheng ³, Xingxing Liu ^{2,4} and Jianghu Lan ^{2,4,5,*} 

¹ College of Geography and Environment, Baoji University of Arts and Sciences, Baoji 721013, China; yukezan2005@163.com (K.Y.); lengfeng1995@126.com (L.W.); bwl1lp@163.com (L.L.)

² State Key Laboratory of Loess and Quaternary Geology, Institute of Earth Environment, Chinese Academy of Sciences, Xi'an 710061, China; liuxx@ieecas.cn

³ Department of Resource and Environment, Zunyi Normal College, Zunyi 563002, China; shengeg1985@126.com

⁴ Center for Excellence in Quaternary Science and Global Change, Chinese Academy of Sciences, Xi'an 710061, China

⁵ China-Pakistan Joint Research Center on Earth Sciences, CAS-HEC, Islamabad 45320, Pakistan

* Correspondence: lanj@ieecas.cn or lanj115@163.com; Tel./Fax: +86-29-6233-6295

Abstract: Understanding the synchronicity of and discrepancy among temperature variations on the western Loess Plateau (WLP), China, is critical for establishing the drivers of regional temperature variability. Here we present an authigenic carbonate-content timeseries spanning the last 300 years from sediments collected from Lake Chaonaqiu in the Liupan Mountains, WLP, as a decadal-scale record of temperature. Our results reveal six periods of relatively low temperature, during the intervals AD 1743–1750, 1770–1780, 1792–1803, 1834–1898, 1930–1946, and 1970–1995, and three periods of relatively high temperature during 1813–1822, 1910–1928, and since 2000. These findings are consistent with tree-ring datasets from the WLP and correlate well with extreme cold and warm events documented in historical literature. Our temperature reconstruction is also potentially representative of large-scale climate patterns over northern China and more broadly over the Northern Hemisphere. The Pacific Decadal Oscillation (PDO) might be the dominant factor affecting temperature variations over the WLP on decadal timescales.

Keywords: authigenic carbonate; temperature variations; Lake Chaonaqiu; western loess plateau



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1. Introduction

Anthropogenic global warming is a critical issue of broad scientific and socio-economic concern [1]. Since the mid-20th century, severe heatwaves of increasing duration and intensity have impacted many regions of the globe [2–5]. For example, the 2010 heatwave in western Russia caused a widespread decline in ecosystem productivity, while concurrently increasing respiration [6]. It is therefore important to investigate temperature variations in different regions in order to gain further understanding of 20th century warming in the context of the previous several hundred years or the previous millennium. Numerous studies focusing on the last millennium has greatly improved our understanding of climate change and the relative roles of natural and anthropogenic forcings [7–17]. It is well known that temperature varies on different timescales and the corresponding forcings are also variable [18]. This would reasonably result in variable temperature patterns in different regions on different timescales. Because most large-scale temperature curves are generated from data averaged over broad geographic areas, differences in regional temperature variations could possibly be masked. This may limit our understanding of regional temperature variations and weaken the reliability of regional climatic predictions. Therefore, it is crucial to master the details in temperature variations for different regions and shed light on the underlying dynamics.

Located at the juncture of the East Asia monsoon (EAM) and northwestern arid zone, the ecological fragility and environmental sensitivity of the western Loess Plateau (WLP) make this an ideal region for studying global climatic changes. Mastering the similarities and differences in temperature variations along the WLP transect is thus vital to understand the mechanisms of temperature and precipitation variations. Despite the dearth of long-term meteorological data, scientists have made great efforts to reconstruct the paleoclimates by tree-rings throughout the WLP [19–22]. Yet, although common features are evident among these previous reconstructions, there are notable discrepancies in both the timing and magnitude of reconstructed events that remain unresolved. For example, Liu et al. [19] has reconstructed temperatures over the past 100 years at Huangling based on $\delta^{13}\text{C}$ in tree rings. Using the tree ring width data at Kongtong Mt., Song et al. [22] has further reconstructed temperature variations over the past 283 years. However, the possible forcing mechanisms have not been comprehensively discussed [19–21]. To help address these inconsistencies, we extracted temperature proxy indices from sediments collected from Lake Chaonaqiu in the Liupan Mountains, WLP, and compared these data with existing records of decadal temperature variability along a transect across the WLP. We focused specifically on the phase relationship of decadal climatic variations over the past three centuries and explored possible forcing mechanisms for these changes.

2. Materials and Methods

2.1. Background and Sampling

Lake Chaonaqiu (2430 m elevation; also known as Lake Tianchi) is a small alpine barrier freshwater lake in the Liupan Mountains, WLP, located ~30 km northeast of Zhuanglang County (Figure 1A). With an area of 0.02 km² and maximum water depth of 9 m, the lake is fed primarily by rainfall and drains seasonally via a topographic low on its western shore. The underlying bedrock throughout the lake basin is red sandstone. The salinity and pH of lake water are 0.17 g/L and 7.83, respectively [23], and total phosphorus (TP) and nitrogen (TN) concentrations are 40.2 µg/L and 1096.8 µg/L [24], respectively. Mean annual precipitation at Lake Chaonaqiu is 615 mm, with a mean annual air temperature of 3.4 °C [25].

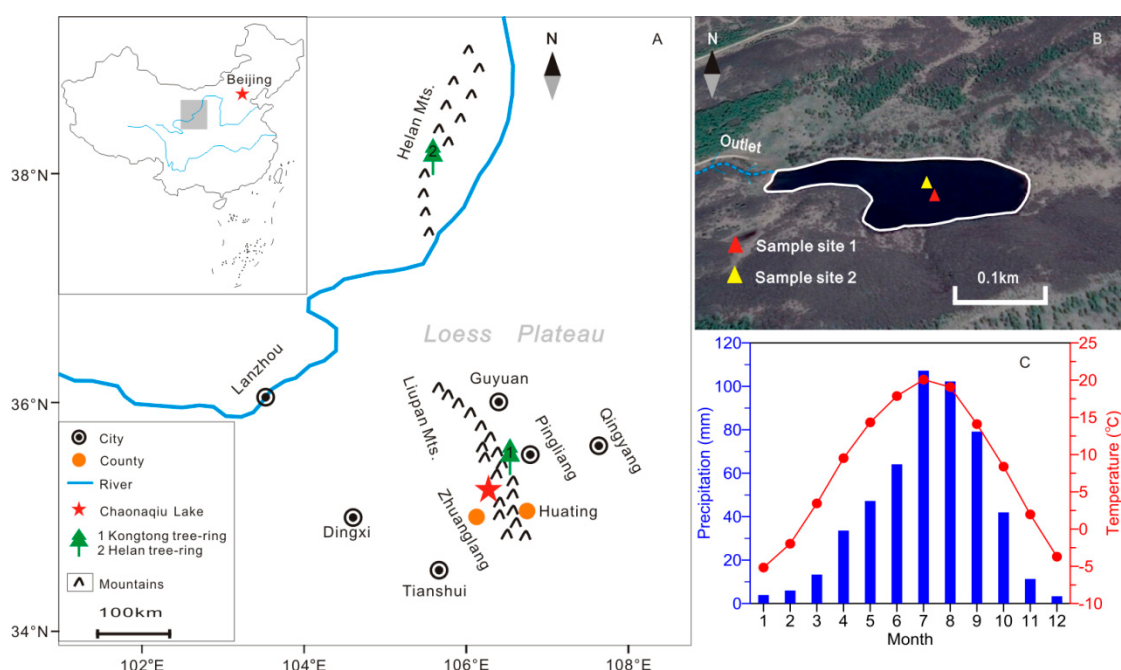


Figure 1. Overview of the study site. (A) Location of the study area on the western Chinese Loess Plateau, and other sites mentioned in the text. (B) Aerial view of Lake Chaonaqiu and sample site locations. (C) Monthly mean precipitation (blue bars) and monthly mean temperature (red dotted line) recorded at Zhuanglang meteorological station since 1960.

We used a UWITEC gravity corer to collect four surface sediment cores at two sample sites located in the center of the lake ($35^{\circ}15'53.08''$ N, $106^{\circ}18'35.99''$ E) during September 2012. The sediment profiles were undisturbed, and the sediment–water interface remained clear. Cores CNQ12-1 and CNQ12-4 were extracted from Sites 1 and 2, respectively (Figure 1B). Core CNQ12-1 (~73 cm long) was subsampled in the field at one-centimeter intervals to quantify the core's mass depth. Core CNQ12-4 was also subsampled at centimeter intervals, but in the laboratory rather than in the field. Owing to minor compaction during transportation and storage, and material loss during subsampling, we were unable to establish an accurate mass depth, and therefore age model, for core CNQ12-4.

2.2. Methods

For both cores CNQ12-1 and CNQ12-4, we measured ^{137}Cs and $^{210}\text{Pb}_{\text{ex}}$ radioactivity via high-resolution, multi-channel gamma-ray spectrometry using an Ortec Hyperpure Germanium (HPGe) well detector (GWL-250-15), with an experimental error of <10% and detection limit of 0.1 Bq kg^{-1} (at 99% confidence; [26]). The bulk carbonate content (carb%) of core CNQ12-1 was determined by titration with diluted perchloric acid (HClO_4 ; 0.1 mol L^{-1}), with an analytical precision better than 0.5% [27,28]; we also measured the bulk carbonate content of core CNQ12-4 to crosscheck our results from the first core. We selected a total of twenty-six representative sediment samples and one surface soil sample for X-ray diffraction (XRD) analysis. Samples were ground to a grain size of $<74 \mu\text{m}$ (<200 mesh) before measurements, after which XRD patterns were obtained using a PANalytical X'Pert Pro MPD diffractometer with $\text{CuK}\alpha$ radiation, and a Ni filter set at 40 kV and 40 mA intensity. Diffraction patterns were scanned from 3° to $70^{\circ} 2\theta$, using a step size of 0.02° [29]. Finally, elemental compositions of odd-numbered samples from core CNQ12-1 were determined using an X-ray fluorescence spectrometer (XRF, Axios advanced (PW4400); [30]). All measurements were performed at the Institute of Earth Environment, Chinese Academy of Sciences (IEECAS), Xi'an, China.

3. Results

3.1. Chronology

Given that anthropogenic radionuclide ^{137}Cs is deposited from the atmosphere within a year, the point of maximum fallout as recorded in our cores provides a time marker for the year 1964 (all dates reported hereon are given in years AD; [17,26,27]). The ^{137}Cs curves for cores CNQ12-1 and CNQ12-4 both exhibit unimodal distributions with pronounced ^{137}Cs peaks (Figure 2a,b; [31]), a pattern that is similar to the classic pattern of global atmospheric ^{137}Cs fallout [26,32]. This close alignment confirms the reliability of the 1964-time marker in core CNQ12-1, where it occurs at a mass depth of $4.09 \text{ g}\cdot\text{cm}^{-2}$, or average geometric depth of 17.5 cm (Figure 2a; [31]).

The $^{210}\text{Pb}_{\text{ex}}$ curves for cores CNQ12-1 and CNQ12-4 exhibit clear subordinate fluctuations superimposed upon long-term logarithmic trends (Figure 2a,b; [31]). Such subordinate fluctuations might imply that biological activity has exerted a washing effect on $^{210}\text{Pb}_{\text{ex}}$ concentrations, similar to results obtained from Lake Chenghai in Yunnan Province [33]. Considering the influence of biological washing on the accuracy of the $^{210}\text{Pb}_{\text{ex}}$ age model, we chose not to employ $^{210}\text{Pb}_{\text{ex}}$ radioactivity as a basis for generating core chronologies [31].

The chronology for core CNQ12-1 is well established, based on a constant mass accumulation rate of $0.0852 \text{ g}\cdot\text{cm}^{-2}\cdot\text{yr}^{-1}$, and spans the period 1743–2012 (268 years; Figure 2c; [31]). Figure 2 also depicts the ^{137}Cs – ^{210}Pb (CRS model) ages of Chen et al. [24]; the age-control point at 39.75 cm was removed owing to the anomalously large dating error. As shown in Figure 2, the age-control points of Chen et al. [24] correlate well with our dating model. In addition, previous work at Lake Chaonaqiu has shown that the calibrated ^{14}C age of 620 cal yr BP in core GSA07 occurs at 162 cm depth [34–36], in agreement with our ^{137}Cs age model. Together, the close alignment of these multiple age constraints confirms the reliability of our chronology.

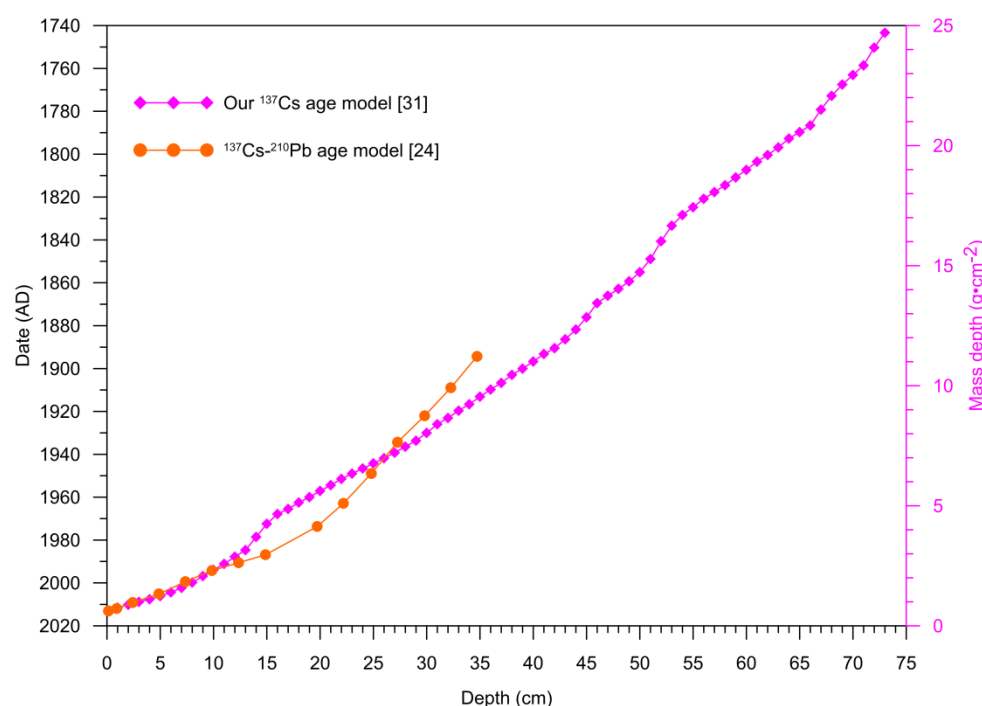


Figure 2. Chronologies for cores CNQ12-1 [31] and GS14A [24] from Lake Chaonaqiu.

3.2. Proxy Indices

As shown in Figure 3, carbonate contents of cores CNQ12-1 and CNQ12-4 range from 0.42% to 6.90%, with average values of 2.88% and 3.38%, respectively. We note the synchronicity of carbonate content between the two cores (Figure 3), which highlights the reliability of the CNQ12-1 carbonate curve. The relatively low carbonate content of the Lake Chaonaqiu sediments likely reflects the influence of two geochemical factors. First, given that precipitation of chemically deposited carbonate is directly affected by salinity [27,37], the relatively low salinity of Lake Chaonaqiu (0.17 g/L; [23]) is not conducive to carbonate precipitation, and thus carbonate sedimentation is minimal. In contrast, Lake Qinghai in central China has a high salinity (14.53 g/L; [38]) due to extensive evaporation, resulting in effective deposition of Ca^{2+} with HCO_3^- and CO_3^{2-} and a correspondingly high (>22%) carbonate content of lake sediments [37]. Second, the bedrock underlying the Lake Chaonaqiu catchment is dominated by red sandstone, which potentially restricts the input of Ca^{2+} and thus limits carbonate precipitation. Where catchments are underlain by limestone, such as Lake Sayram in northwest China [27] and Lake Lugu in southwest China [12,39], runoff supplies Ca^{2+} that is readily deposited with HCO_3^- and CO_3^{2-} , resulting in high (40.4% and 24.62%, respectively) overall sediment-carbonate contents.

Our XRD results indicate that the mineralogic composition of core CNQ12-1 is dominated by quartz, albite, biotite, and calcite (Figure 4). Among the 26 sediment samples tested, the calcite signal exhibits an average strength of 1765 counts, with maximum (2451 counts) and minimum (1522 counts) values occurring at line depths of 1 and 73 cm, respectively (Figure 5). With the notable exception of calcite, the high degree of similarity in primary peaks and mineralogic compositions between lake sediments and surface soils suggests that surface runoff of terrestrial clastic material is the principal source of sediment in Lake Chaonaqiu. By this scenario, the calcite content of the lake sediments reflects subsequent chemical deposition.

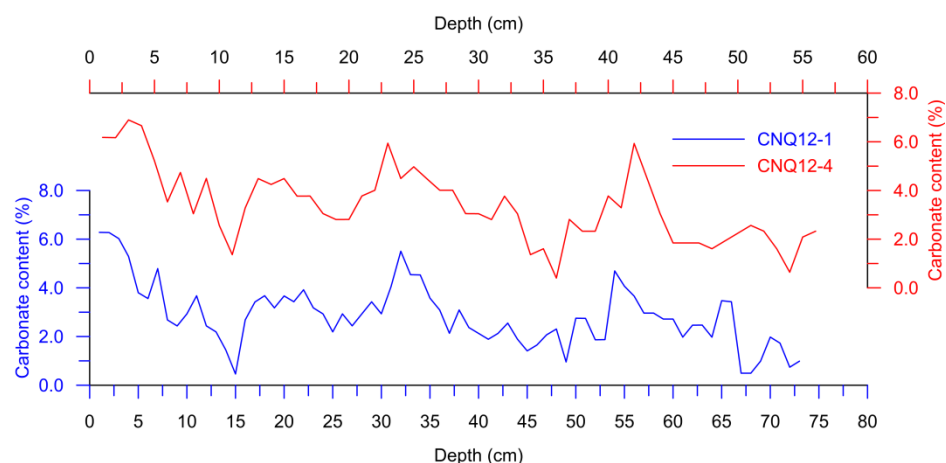


Figure 3. Carbonate contents of the two Lake Chaonaqiu sediment cores.

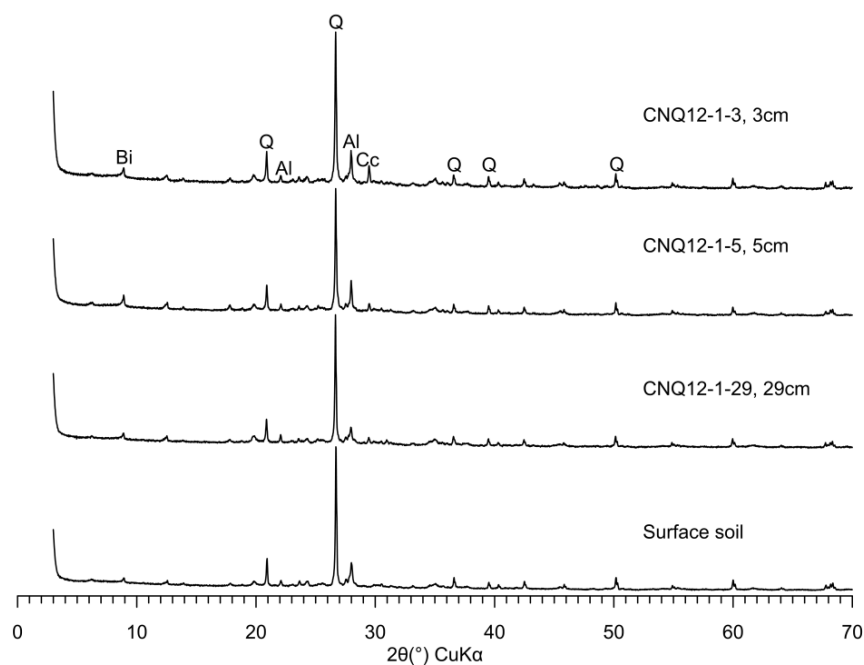


Figure 4. Results of XRD analysis for core CNQ12-1 and surface soils. Q—Quartz; Bi—Biotite; AL—Albite; Cc—Calcite.

The XRF chemical element results reveal an average Ca concentration of 2.44%, with maximum (5.29%) and minimum (1.04%) values occurring at line depths of 1 and 67 cm, respectively (Figure 5). Considering the similarities between Ca concentrations and carbonate content, their shared synchrony with the calcite signal strength (Figure 5), and the absence of calcite from surface soils (Figure 4), we propose that the bulk of carbonate in the Lake Chaonaqiu sediments is authigenic and that the Ca element is derived primarily from calcite.

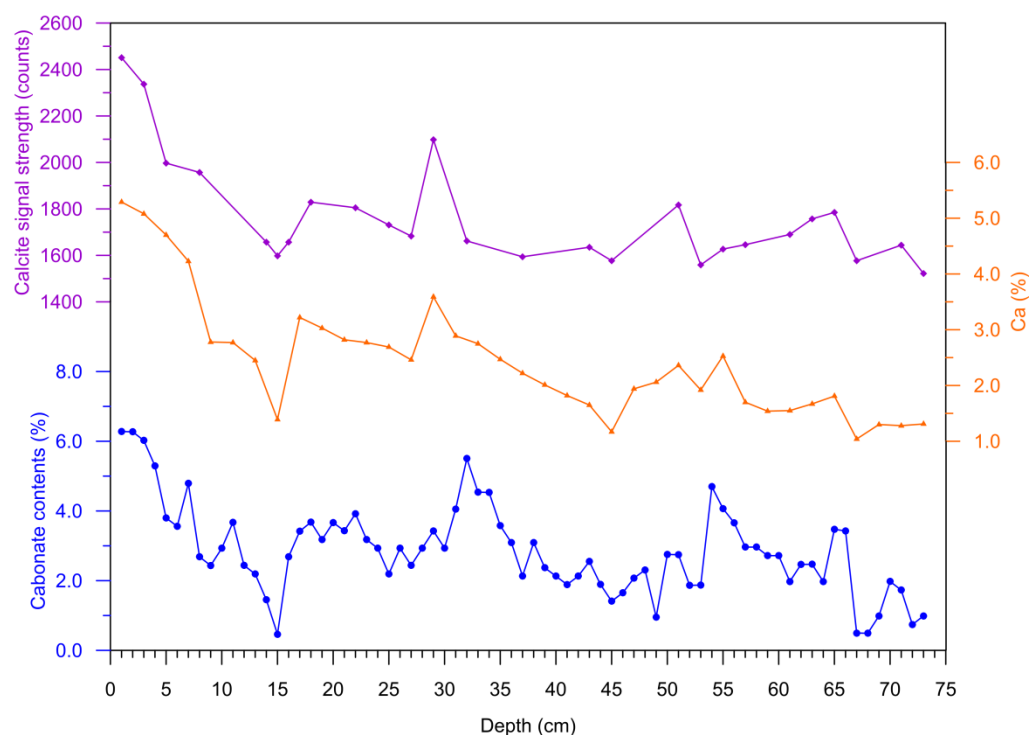


Figure 5. Change curves for carbonate content, elemental Ca concentration, and calcite signal strength for core CNQ12-1.

4. Discussion

4.1. Climatic Significance of the Authigenic Carbonate

The rate of carbonate precipitation in a lake is generally controlled by the ratio of evaporation to precipitation (E/P), with higher E/P resulting in carbonate supersaturation in the water column, elevated carbonate precipitation rates, and a greater overall carbonate content in lake sediments [27,37,40–44]. Moreover, the content of authigenic carbonate in lake sediments can reveal the dominant role of evaporation or precipitation in controlling E/P values. For example, while Lan et al. [27] reported that authigenic carbonate precipitation in Lake Sayram results from extensive summer evaporation, the authors also observed that evaporation should weaken as precipitation increases, resulting in unsaturation of lake water and a decline in carbonate sedimentation. Therefore, carbonate contents in Lake Sayram sediments can be used as an indicator of regional precipitation [27]. Although similar carbonate-derived paleoenvironmental interpretations have also been presented for Lake Bosten [40,44], Lake Dali [42], and Lake Sasikul [45], other studies have explored the role of temperature in controlling E/P values. At Lake Qinghai, for instance, since temperature is the dominant factor that controls the evaporation, and the influence of temperature on the salinity of lake water (primarily influenced by regional E/P) may be stronger than that of precipitation. Therefore, carbonate content in Lake Qinghai sediments can be used as an indicator of regional temperatures [37,43]. Similar inferences have been made for Lake Daihai [41].

Located on the margin of the Asian summer monsoon (ASM) region, our site in the Liupan Mountains experiences diurnal and annual temperature variability; mean annual evaporation (~1102.80 mm; [46]) is 1.8 times the mean annual precipitation (~615 mm; [25]). Because the surface of Lake Chaonaqiu is typically frozen between November and March [46], chemical carbonate precipitation results from extensive summer evaporation. Therefore, authigenic carbonates formed in Lake Chaonaqiu by chemical and biochemical processes are closely associated with the evaporation of the lake water, which primarily controls the chemical composition and salinity of the lake water. Given that the evaporation of lake water is mainly controlled by temperature and atmospheric relative humidity, and humidity

is determined by temperature and atmospheric precipitation, we infer that the formation of authigenic carbonates in Lake Chaonaqiu can be linked predominantly with temperature. In addition, upon comparing the carbonate content with mean annual temperature data from Pingliang and Zhuanglang meteorological stations (Figure 6), we observed a positive correlation between them (Figure 6A,B), reinforcing our view that variations in authigenic carbonate content in Lake Chaonaqiu sediments can be employed as an indicator of regional temperature. When temperature rises, evaporation is enhanced, leading to an increase in Ca^{2+} and HCO_3^- concentrations of lake water and promoting carbonate supersaturation, and thus resulting in higher carbonate contents in Lake Chaonaqiu sediments. Conversely, unsaturation causes the carbonate content of lake sediments to decline when temperature drops. Therefore, we interpret the carbonate content in Lake Chaonaqiu sediments as a proxy index for temperature changes in this region, with increased carbonate content is related to higher temperature, and vice versa.

4.2. Temperature Variations at Lake Chaonaqiu over the Past 300 Years

As shown in Figure 6, the carbonate content of lake sediments is relatively low for the periods 1743–1750, 1770–1780, 1792–1803, 1834–1898, 1930–1946, and 1970–1995, which we interpret as reflecting cooler temperatures at Lake Chaonaqiu. It is noteworthy that low values in the period of 1834–1898 coincided with the final cold stage (1830–1890) of the Little Ice Age (LIA) in China [9,47,48]. According to Wen [49], several extremely cold events in Zhuanglang County have been described in historical documents (No. 1 and 18 in Table 1). For example, “On 1 October, the 7th year of the region of Emperor Tongzhi, Qing Dynasty (the traditional Chinese calendar, equivalent to 14 November 1868), Zhuanglang County was seriously impacted by a snowstorm, which buried roads and crushed vegetation.”

The carbonate record also exhibits elevated values in 1813–1822, 1910–1928, and since 2000, which reflect periods of relatively higher temperatures in the Lake Chaonaqiu region. According to Wen [49], extremely warm events in Huating County (~40 km southeast of Lake Chaonaqiu) were also reported in historical accounts (No. 19 in Table 1): “In Autumn, the 3rd year of the reign of Emperor Xuantong, Qing Dynasty (1911), vegetation bloomed again in Huating”.

4.3. Temperature Variations on the Western Loess Plateau over the Past 300 Years

The pattern of temperature variability at Lake Chaonaqiu over the last few centuries is similar to those reconstructed for other regions in China [7,50]. For example, the comparison of our carbonate dataset to tree-ring records from Kongtong Mountain [22], Helan Mountain [20], and the mid-eastern Tibetan Plateau [51] reveals a considerable degree of convergence among the various datasets over the past 300 years (Figure 6). Specifically, variations in carbonate content at Lake Chaonaqiu are broadly synchronous with fluctuations in tree-ring width, confirming that our record is a robust indicator of regional temperature. Due to dating uncertainties, sampling resolution, and site characteristics, the cold periods 1743–1750, 1770–1780, and 1970–1995 are not represented in the tree-ring records. Nonetheless, three cold periods (1792–1803, 1834–1898, and 1930–1946) and two warm intervals (1813–1822 and 1910–1928) are clearly documented in tree-ring and lake carbonate records alike (Figure 6). Although tree-ring-inferred temperatures exhibit subordinate fluctuations during the 1834–1898 cold episode, most likely owing to site-specific factors, the majority of regional cold extremes occurred during the cold intervals (Figure 6). This pattern suggests that the thermal signature of the LIA in China [9,47,48] was prevalent throughout the WLP, and indicates that this regional variability is captured in the Lake Chaonaqiu sedimentary record on a decadal scale.

Table 1. Extreme cold and warm events on the WLP identified in historical literature [49,52].

No.	Solar Calendar Dates	Description
1	31 July 1744	The 9th year of the reign of Emperor Qianlong, Qing Dynasty: On 25 July, uncovered oats in Haiyuan, Guyuan, and Huating Counties were damaged by frost.
2	24 April 1748	The 13th year of the reign of Emperor Qianlong, Qing Dynasty: On the evening of 1 March, seedlings were killed by frost in Gangu, while on the same night, heavy snow fell on the suburbs in Guyuan.
3	1749	The 14th year of the reign of Emperor Qianlong, Qing Dynasty: Longde, Guyuan, and other counties were affected by summertime and autumn frosts.
4	December 1773	The 38th year of the reign of Emperor Qianlong, Qing Dynasty: In November, relief aid was provided to refugees of frost and famine in Jingchuan.
5	1776	The 41st year of the reign of Emperor Qianlong, Qing Dynasty: Chongxin, Jingchuan, and Lingtai Counties were affected by frost.
6	4 June 1777	The 42nd year of the reign of Emperor Qianlong, Qing Dynasty: On 2 May, Dingxi experienced frost.
7	December 1783	The 48th year of the reign of Emperor Qianlong, Qing Dynasty: In November, wildflowers were in full bloom in Zhenyuan. ★
8	November 1817	The 22nd year of the reign of Emperor Jiaqing, Qing Dynasty: In November, peach trees were blooming in Zhenyuan. ★
9	December 1818	The 23rd year of the reign of Emperor Jiaqing, Qing Dynasty: In November, peaches and winter jasmine were blooming in Zhenyuan. ★
10	1837	The 18th year of the reign of Emperor Daoguang, Qing Dynasty: On 12 January (equivalent to 5 March 1838), grain rations and seeds were provided to frost refugees in Guyuan and Longde.
11	19 December 1840	The 20th year of the reign of Emperor Daoguang, Qing Dynasty: On 26 November, the old and new taxes were postponed for frost refugees in Longde.
12	December 1841	The 21st year of the reign of Emperor Daoguang, Qing Dynasty: In November, the old taxes were postponed due to frost in Guyuan and Lingtai.
13	14 November 1868	The 7th year of the reign of Emperor Tongzhi, Qing Dynasty: On 1 October, Zhuanglang County was seriously impacted by a snowstorm, which buried roads and crushed vegetation.
14	April 1871	The 10th year of the reign of Emperor Tongzhi, Qing Dynasty: In April, Lingtai County suffered heavy frost, resulting in serious damage to seedlings.
15	1873	The 12th year of the reign of Emperor Tongzhi, Qing Dynasty: During summer, Lingtai County suffered heavy frost, resulting in the loss of seedlings and crops.
16	30 May 1884	The 10th year of the reign of Emperor Guangxu, Qing Dynasty: On 8 April heavy sleet fell in Jingchuan.
17	May 1890	The 16th year of the reign of Emperor Guangxu, Qing Dynasty: In April, crops in Jingyuan were killed by frost.
18	September 1902	The 28th year of the reign of Emperor Guangxu, Qing Dynasty: In August, seedlings in Guangleli (an ancient placename), Zhuanglang County, were damaged by heavy frost.
19	1911	The 3rd year of the reign of Emperor Xuantong, Qing Dynasty: In Autumn, vegetation bloomed again in Huating. ★
20	1915	The 4th year of the Republic of China: In Autumn, vegetation bloomed again in Huating. ★

Table 1. Cont.

No.	Solar Calendar Dates	Description
21	1930	The 19th year of the Republic of China: Longde was affected by black frost, resulting in famine refugees.
22	1932	The 21st year of the Republic of China: Piangliang County was affected by frost during the first half of the year. Black frost during the autumn damaged seedlings in Lingtai, impacting an area spanning >100 miles from north to south.
23	2 May 1933	The 22nd year of the Republic of China: On 8 April, Jingyuan County was impacted by a blizzard and intense cold, with >5 feet (equivalent to 155 cm) of snow falling in the Longshan Mountains.
24	1938	The 27th year of the Republic of China: It took quite a long time for the heavy snow in Huating to melt.
25	1940	The 29th year of the Republic of China: Black frost killed seedlings in Lingtai, Huating, Chongxin, Pingliang, Zhenyuan, Zhuanglang, and Jingning Counties.
26	1941	The 30th year of the Republic of China: Black frost killed seedlings in Pingliang, Huating, Chongxin, Jingning, and Longde Counties.
27	1942	The 31st year of the Republic of China: Black frost killed crops in Huating, Zhuanglang, Pingliang, and Jingning Counties.
28	1943	The 32nd year of the Republic of China: Frost caused extensive damage in Baishui (and thirteen other towns in Pingliang County), Jingfchuan and Shuiluo (and eleven other towns in Zhuanglang County), and Jingning.
29	1945	The 34th year of the Republic of China: Pingliang suffered frost. On 4 October, early frost in Zhenyuan resulted in extensive ice cover (frost of 0.3 cm depth) and killed >80% of the buckwheat crop. In May, black frost occurred in Chengguan and five other towns in Huating County, damaging the majority of the sprouting wheat crop.
30	1947	The 36th year of the Republic of China: On 9 and 10 October heavy frost occurred in Longpan town in Pingliang County, killing the foxtail millet and buckwheat crops. Black frost occurred in Ankou and Longyan towns, Huating County, resulting in the destruction of sprouting wheat and other crops. On 15 May frost damaged crops in Shenshangu town, Chongxin County.
31	1971	On 25 September early frost occurred in Pingliang and other regions, with minimum ground temperatures of -4°C to -1°C .
32	1972	Between 13 and 15 May late frosts occurred in Dingxi, Pingliang, and Qingyang, with minimum ground temperatures of -5°C and widespread cotton, corn, and crop failures.
33	1974	Late frost ends on 9–10 May in Gansu Province, and is delayed for 15–22 days in Tianshui and Longdong. Such an occurrence is rare in recent decades.
34	1976	The average temperature for July in Pingliang dropped by $0.1\text{--}1.0^{\circ}\text{C}$.
35	1979	Extreme cold events occurred in 1979, 1987, 1993, and 1995 in Ningxia Province (1949–2000).
36	1987	
37	1993	
38	1995	

Notes: black stars indicate extreme warm events.

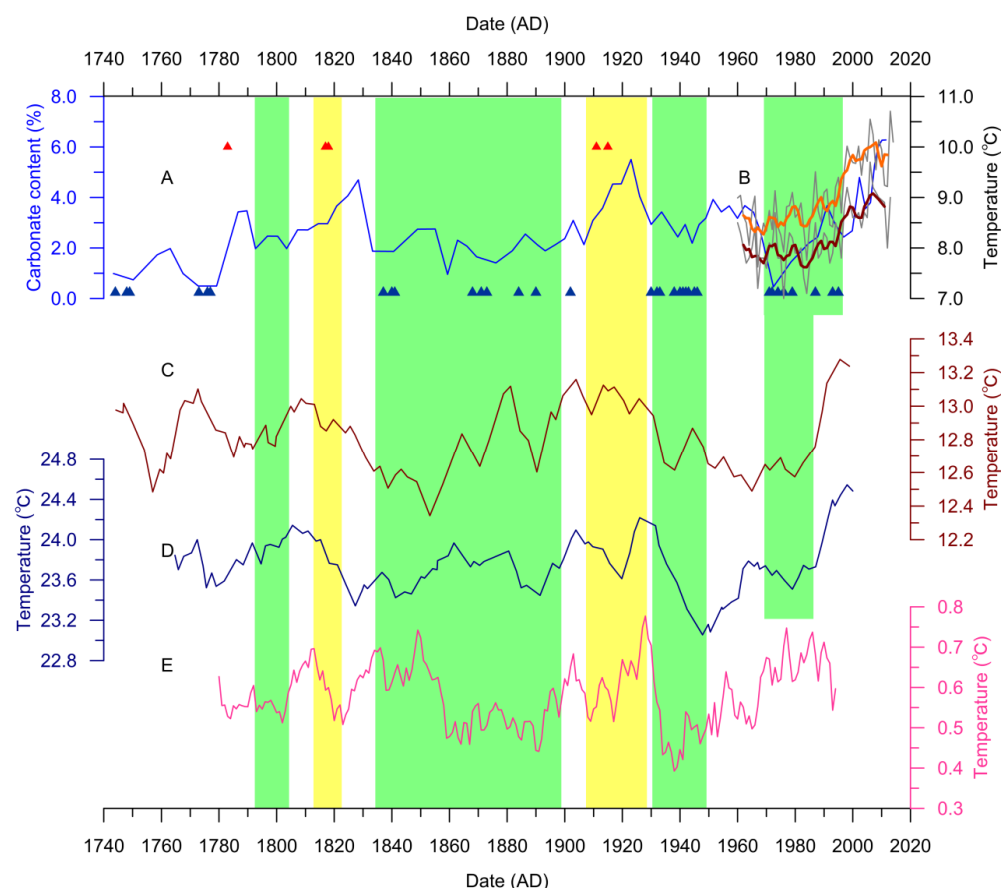


Figure 6. Comparison of the core CNQ12-1 carbonate record and tree-ring-inferred temperatures on the WLP. A. Carbonate content (*this paper*), extreme cold events (blue triangles) and warm events (red triangles) [49,52]. B. Annual mean temperature from Pingliang (orange line; 11-year running average) and Zhuanglang (dark red line; 11-year running average) meteorological stations since 1960. C. February–September air temperatures reconstructed from the Kongtong tree-ring record [22]. D. May–July air temperatures reconstructed from the Dulan–Wulan tree-ring record [51]. E. January–August air temperatures reconstructed from the Helan tree-ring record [20]. The green and yellow shadings indicate cold intervals and warm intervals, respectively.

A total of 33 extreme cold or frost events (details see Table 1) on the WLP recorded in historical literature [49,52] occurred during the six cold periods described above. For example, one of them stated (No. 23 in Table 1): “On 8 April the 22nd year of the Republic of China (the traditional Chinese calendar, equivalent to 2 May 1933), Jingyuan County was impacted by a blizzard and intense cold, with more than five feet (equivalent to 155 cm) of snow falling in the Longshan Mountains.”

In addition to cold events, five episodes of extreme warmth (details see No. 7, 8, 9, 19, 20 in Table 1) on the WLP were also recorded in the historical literature [49], specifically during the intervals of 1813–1822 and 1910–1928. For example, one of them stated (No. 8 in Table 1): “In November, the 22nd year of the reign of Emperor Jiaqing, Qing Dynasty (November, 1817), peach trees were blooming in Zhenyuan”. In general, peach trees bloom in March or April and lie dormant in November. However, when the temperature rises suddenly in November (a meteorological phenomenon known as *Daochunyang*), dormancy is interrupted and the peach trees can bloom again as in the spring. This out-of-season blooming of peach trees, along with the rejuvenation of vegetation (details see No. 20 in Table 1) that is typically dying off during the autumn, is indicative of a warm climate.

4.4. Possible Forcing Mechanisms of WLP Temperature Variations over the Past 300 Years

It is well known that the sun is the ultimate source of energy for Earth's climate [53–57], and myriad studies have demonstrated the close link between solar irradiance and terrestrial temperature variability [9,10,12,43]. For example, the low temperatures over the northern Tibetan Plateau (NTP) are broadly synchronized with the classical solar minima during the past several hundred years [9,10,12]. Moreover, atmospheric circulation, such as the “El Niño–Southern Oscillation” (ENSO) [8,58,59] and Pacific Decadal Oscillation (PDO) [22,60–62] might also influence regional temperature variations. For instance, ENSO-induced changes in regional hydrological cycles are expected to alter patterns of latent heating, thereby impacting temperatures on the southeastern margin of the Tibetan Plateau (S-ETP; [59]). Tollefson [61] argued that warm PDO phases coincide with periods of rapid global warming (e.g., 1920s to 1940s; 1980s and 1990s). As shown in Figure 7A, we note similar trends in low-frequency variability between the Chaonaqiu temperature record and total solar irradiance (TSI; [55]) curves, potentially indicating that TSI is a key driver of WLP temperatures on centennial timescales. This is not the case, however, on decadal/multi-decadal timescales. We speculate that the low altitude of the WLP relative to the TP makes the former less sensitive to changes in solar radiation, despite the weak influence of the ASM in this region. If so, atmospheric circulation might be the dominant factor impacting temperatures on decadal timescales on the WLP.

The PDO has been described by some as a long-lived El Niño-like pattern of North Pacific climate variability [63], which influences climate throughout the Pacific basin [60,63]. Indeed, 50 years of statistical data demonstrates a clear PDO signature both in atmospheric circulation over East Asia and decadal climate fluctuations in China, whereby the warm PDO phase coincides with elevated temperatures and reduced precipitation in northern China, and vice versa [60]. Based on monthly average precipitation and temperature data, Ma [62] also identified elevated precipitation and depressed temperatures in northern China during cold PDO phases, with the opposite being true during warm phases. To further evaluate the role of PDO on temperature changes in Lake Chaonaqiu, we compared our record with (i) temperature data from northern China [64], (ii) temperature anomalies in northwestern/northern China [65], (iii) Northern Hemisphere temperature [8], and (iv) the PDO index [66] for the past 300 years. As illustrated in Figure 7, the cold periods in the Lake Chaonaqiu record correspond to depressed temperatures in northern China and the Northern Hemisphere in general, with negative temperature anomalies in northwestern/northern China and cold PDO phases and vice versa (Figure 7). This suggests that the temperature variations inferred from Lake Chaonaqiu can represent the patterns of climate variations over the past 300 years in northern China, possibly even on a hemispheric scale. More importantly, it reveals that the PDO is closely related to decadal climate change in northern China. This result aligns with earlier studies [22,60] that suggest that a developing El Niño event during the cold PDO phase is related to more precipitation and lower temperatures in northern, northeastern, northwestern [60], and central China [22].

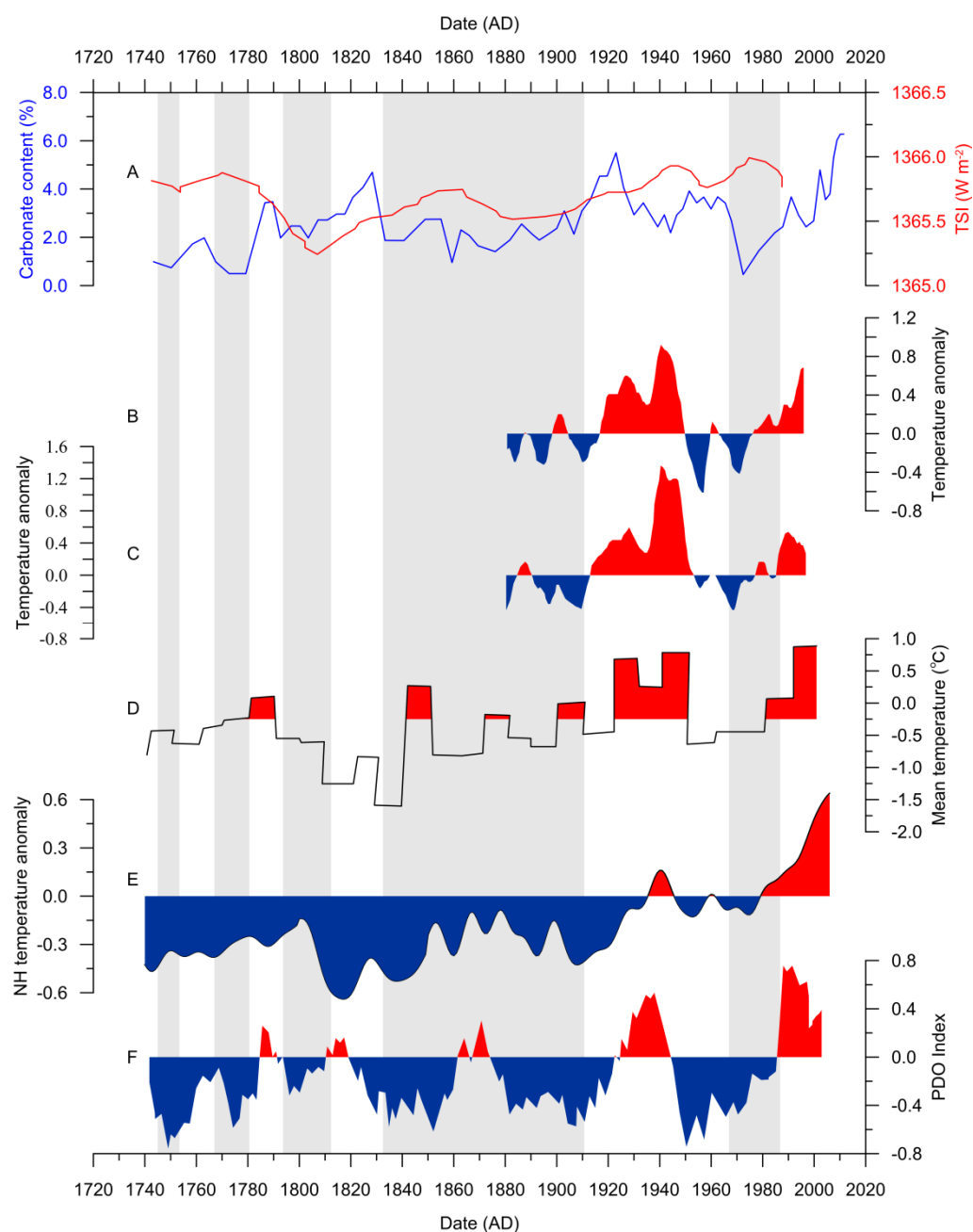


Figure 7. Temperature variations on the WLP and potential drivers. **(A)** Authigenic carbonate content in Lake Chaonaqiu sediments (blue line; this study) and total solar irradiance (TSI) reconstructed from polar ice ^{10}Be (red line; [55]). **(B)** Temperature anomalies in northern China [65]. **(C)** Temperature anomalies in northwestern China [65]. **(D)** Temperature in northern China [64]. **(E)** Northern Hemisphere (NH) temperature anomalies [8]. **(F)** PDO index [66]. Gray shadings highlight intervals of decreased temperature and cold PDO phase.

5. Conclusions

This study employed the content of authigenic carbonate in Lake Chaonaqiu sediments to reconstruct regional temperature variability on decadal timescales, and to investigate possible forcing mechanisms for such variability over the past 300 years. Our results revealed six cold and three warm periods that are consistent with other paleoclimate records from throughout the WLP and northern China, and more broadly across the Northern Hemisphere. This close agreement suggests that temperature variations recorded in the Lake Chaonaqiu region are representative not only of northern China, but also

the Northern Hemisphere. We propose that the PDO is the dominant factor influencing temperature variability on the WLP on decadal timescales.

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