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Key Points:

- Summer temperature and precipitation during the Current Warm Period in China are not unprecedented over the past 22,000 years
- The Holocene maxima of summer temperature and precipitation in different regions of China occur at about the same time (8.1–7.6 ka)
- The variations of summer temperature and precipitation in northern China are mainly determined by insolation

Supporting Information:

- Supporting Information S1
- Data Set S1

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The Position of the Current Warm Period in the Context of the Past 22,000 Years of Summer Climate in China

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Abstract Identifying the position of the Current Warm Period (CWP) in the context of the long-term climatic trend is vital for understanding the impact of human activity on climate change. Reconstructions of summer temperature and precipitation in eight subregions of China over the past 22,000 years show that the CWP summer temperature and precipitation in these subregions are all lower than in the Early to Middle Holocene. The timing of the Holocene temperature and precipitation peaks in northern China (including Northwest China, North China, and Northeast China) is mainly determined by orbital forcing. Greenhouse gas forcing and the land ice-sheet help to fine-tune the timing of the climate maxima. These findings show that the climate since the Last Glacial Maximum in northern China is more sensitive to nonanthropogenic external forcings, whereas the summer precipitation in Southwest China since the early 20th century is controlled more by anthropogenically forced changes.

Plain Language Summary Reconstructions of past climate are essential for understanding current warming trends and reasons under the natural climate and anthropogenic forcings. We provide a climate reconstruction in China over the past 22,000 years using a novel method combining proxy data and model simulations. Our results show that the summer temperature and precipitation during the Current Warm Period in the eight subregions of China are both significantly lower than those in the Early and Middle Holocene, and the long-time summer temperature and precipitation variations are highly sensitive to orbital forcing. These results highlight the regional consistency in the climate response to external forcings.

1. Introduction

A clear warming trend of 0.81°C in the global annual mean temperature spanning the past 100 years (1919–2018 CE) is seen in the HadCRUT4 gridded dataset of observational global historical surface temperatures (Morice et al., 2012); this past 100 years of global warming is referred to here as the Current Warm Period (CWP). Four other instrumental gridded datasets of observational global historical surface temperatures (NOAA, NASA GISTEMP, Cowtan and Way, and Berkeley Earth) also show this warming (Rahmstorf et al., 2017). This indicates that our planet has been experiencing warming since the industrial revolution (Cheng, 2020). Ascertaining the position of the CWP in the context of long-term global climate change is a key challenge for obtaining a better understanding of the contribution of natural variability to postindustrial climate trends (Masson-Delmotte et al., 2013). Paleoclimate proxy records (e.g., pollen) are generally used to generate a composite of regional and global climate sequences (CILIMAP Project Members, 1976; Kaufman et al., 2020; Marcott et al., 2013; Marsicek et al., 2018; Rehfeld et al., 2018; Shakun et al., 2012), which allow for the examination of changes in the global climate system from a long-term perspective.

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However, few proxy records from China are included in these reconstructions, especially regional reconstructions (Cao et al., 2017; Liang et al., 2020; Wu et al., 2019; E. Zhang et al., 2019), despite the fact that this region is home to one fifth of the world's population and the Qinghai-Tibetan Plateau (QTP) and East Asian monsoon play important roles in the global climate system.

Previous studies have reconstructed Holocene temperature and precipitation in China (Herzschuh et al., 2019; Wu et al., 2019; Zheng et al., 2016), but they have quite low resolution that does not allow the position of the CWP to be studied by extending the instrumental data on annual to decadal timescales. Moreover, their climatic signal-to-noise ratios are also small because of the complexity of the response of pollen records to climate, like other types of proxy records, which leads to large uncertainty in the reconstructed climate. Thus, it remains difficult to resolve climatic changes from these proxy records back as far as the Last Glacial Maximum (LGM; 22 ka) and with annual or decadal resolution (Wu et al., 2019; Q. Xu et al., 2016). Climate model experiments (He et al., 2013; Liu et al., 2009; Smith & Gregory, 2012; Timm & Timmermann, 2007) have been used to simulate the East Asian climate since the LGM. However, there is still a distinct mismatch in data-model comparisons mainly due to uncertainty related to the internal variability that has random phase, the scenario uncertainty arising from an imperfect knowledge of external forcings, and the uncertainty in the model itself (Hawkins & Sutton, 2009). This mismatch limits the accurate determination of the position of the CWP with respect to the long-term climate evolution.

Data assimilation (DA) can overcome the limitations of both proxy-based reconstructions and model simulations by combining the two approaches. It uses observational data (i.e., proxy-based reconstructions and instrumental records) to constrain climate model simulations, in order to obtain optimal estimations of past climate that both conform to the physical laws of climate and are consistent with proxy-based reconstructions (Bhend et al., 2012; Fang & Li, 2016; Goosse et al., 2006; Hakim et al., 2016; Shi et al., 2019; Steiger et al., 2018; Talagrand, 1997; Widmann et al., 2010). In this study, we investigate the position of the CWP summer temperature and precipitation variations in various regions in China since the LGM using an updated version of the optimal information extraction (OIE) method (a new DA method) to reconcile the differences between the proxy-based reconstructions and model simulations. The reconciled high-resolution records of summer temperatures and extended summer precipitation over the past 22,000 years are then used to explore the regional climate response to the glacial-interglacial transition, and constrain the intensity and magnitude of the CWP.

2. Data and Methods

Several datasets were compiled and processed by the OIE method to produce a temperature and precipitation record in China for the last 22,000 years. We used instrumental data records, proxy-based reconstructions, and data from model simulations. Specifically, the datasets include two instrumental records of temperature and precipitation in China (Zhao & Zhu, 2015); two proxy-based East Asian summer temperature reconstructions (Cook et al., 2013; Shi, Ge et al., 2015a) and the proxy-based precipitation reconstructions in China (Shi et al., 2017); the 13 members of the full-forcing experiment of the global Community Earth System Model-Last Millennium Ensemble (CESM-LME) simulation (Otto-Bliesner et al., 2016); the 70 members of the full-forcing experiment of the global LOch-Vecode-Ec-bilt-CLio-agIsm Model-Large Common Era Ensemble (LOVECLIM-LCE) simulation; and the full-forcing experiment of the global Transient Climate of the Last 21,000 years (TraCE-21ka) simulation using the Community Climate System Model version 3 (CCSM3) model (He et al., 2013; Liu et al., 2009). In addition, the single-forcing experiments of the CESM-LME simulation (including the greenhouse gas, orbital, land change, solar, and volcanic forcings), and the TraCE-21ka simulation (including the greenhouse gas, orbital, meltwater, and land ice-sheet forcings) are used to explore the mechanism of climate variability.

The proxy-based temperature and precipitation reconstructions over the past millennium are shown in Figures S1–S2 in the supporting information. The combination of the instrumental temperature record and the arithmetic mean of the multi-proxy temperature reconstruction (Shi, Ge et al., 2015b) and the tree-ring-based temperature reconstruction (Cook et al., 2013) is used as the temperature reconstruction target. The

combination of the instrumental precipitation record and the multi-proxy precipitation reconstruction (Shi et al., 2017) is used as the precipitation reconstruction target. The credibility of these datasets can be ranked as follows:

Instrumental data > Proxy-based reconstructed data > CESM-LME simulation data > LOVECLIM-LCE simulation data > TraCE-21ka simulation data

Detailed information on these datasets is given in Table S1 and Text S1 of the supporting information.

Before DA, all datasets were filtered by a simple 11-year moving average to exclude the dominant internal El Niño-Southern Oscillation (ENSO) variability and its direct influence, since the random phase of ENSO in these model simulations would affect the correlation between the proxy-based reconstructions and the climate simulations. Eight independent proxy records and a gridded proxy-based precipitation field reconstruction (Table S2) were used to assess the quality of the reconciliation. Specific information is given in Table S2 of the supporting information.

A modified OIE method was applied to these datasets to resolve temperature and precipitation data for the last 22,000 years. We developed OIE version 2.1 to splice the above compiled datasets, to give a dataset that is derived from multi-proxy cross-quantitative reconstructions and thus is also referred to as a multi-proxy cross-quantitative reconstruction model. The OIE method was first designed as a variant of the Composite Plus Scale method (Mann et al., 2005) to reconstruct the South Asian summer monsoon index over the past millennium (Shi et al., 2014). Several revisions of the method (Neukom et al., 2019; Shi et al., 2017; Shi, Yang et al., 2015b; Yang et al., 2016) have incorporated the local method (Christiansen, 2011), the Bayesian method (Tingley & Huybers, 2010), the generalized likelihood uncertainty estimation method (Wang et al., 2017), and ensemble reconstructions (Neukom et al., 2014). A detailed description of the OIE method is given in Text S2 of the supporting information. The main changes in OIE version 2.1 (compared with version 2.0; Neukom et al., 2019) are that the gridded proxy-based temperature and precipitation reconstructions are used to calibrate/assimilate the three model simulations and the OIE method is extended to paleoclimate DA research.

The specific implementation of the method is as follows.

Step 1: Locating the predictors

Outputs (temperature and precipitation) from the three climate models in each region that are significantly and positively correlated with the combined proxy and instrumental climate data in the same region are selected as the predictors.

Step 2: Assigning the weight of predictors

The selected model outputs are averaged, weighted by the correlation coefficients between the model outputs and the combined proxy and instrumental climate data in each region. This step is the same as in the Composite Plus Scale method used in Mann et al. (2005).

Step 3: Regressing the averaged predictor

The averaged model outputs are calibrated against the combined proxy and instrumental climate data in each region using the Ensemble local regression method (Shi et al., 2012). This assumes that the regression coefficients are random variables that obey a uniform distribution with a range between the traditional classical calibration (direct regression) and the inverse calibration (indirect regression).

Here a random 75%/25% partition on a set of the compiled proxy and instrumental temperature/precipitation (900–1999 CE/1470–2000 CE) is used to train/test the statistical model using cross-validation, since the calibration period influences the regression model (McShane & Wyner, 2011). The random sampling was applied 30 times.

Step 4: Estimating the uncertainty

General likelihood uncertainty estimation (Blasone et al., 2008; Wang et al., 2017) is used to estimate the reconstructed uncertainty. Specifically, the root mean square error (RMSE) is used to evaluate the weight

of each scenario. These weights are used to generate new members. The updated distribution of the new reconstruction members is used to allow a probabilistic assessment of uncertainties.

Moreover, the traditional verification variables (the reduction of error [RE], the coefficient of efficiency [CE], the squared Pearson correlation coefficient, the RMSE, and the RMSE skill score [RMSE-SS]) are used to test the accuracy of the reconstructions (verification skill) for each member, and the independent proxy reconstructions are used to further estimate the uncertainty of the reconstructions.

3. Results

In the second China National Climate Change Assessment Report, China is divided into eight subregions (Figure 1), each with distinctive climatic and geographical characteristics (Zhou et al., 2014). We splice the proxy-based reconstructions, the CESM-LME simulation, the LOVECLIM-LCE simulation and the TraCE-21ka simulation using the OIE method (Figure S3). Figure 1 compares the three model outputs before and after DA with the proxy reconstructions. There is no significant correlation between climate simulations and proxy-based reconstructions before DA. The OIE-based climate simulations agree much better with the proxy-based reconstructions (Figure 1) after DA. The relationships between the proxy-based climate reconstructions and the OIE-based climate reconstructions from the three climate models over the eight subregions of China are all significant at the 90% confidence level. The significant differences between them in trends and amplitudes before DA were corrected after DA (Figure 1). Moreover, the OIE-based reconstructions also broadly passed the test using the traditional verification variables in Table S3 and the independent proxy reconstructions in Figure S4 (see Texts S3–S4 of the supporting information for detailed information). We therefore conclude that our updated OIE method has produced newly resolved datasets that resolve the amplitude of the CWP.

Figure 2 shows that the CWP does not correspond to the warmest interval for summer temperature in China, with the Early and Middle Holocene all warmer than the CWP. The summer temperatures during the Bølling-Allerød warm period (14.6–12.9 ka) in most of China are also higher than in the CWP except in the QTP and Southwest China (SWC). The summer temperature (Figure 2) reaches its maximum in the Middle Holocene followed by a long-term decrease. Figure 2 also shows that the CWP is not the wettest interval in China. In the eight subregions precipitation gradually increased after the LGM, reached a maximum in the Middle Holocene, and then gradually decreased. The most obvious regional feature is that the low level of precipitation during the CWP in SWC is unprecedented since the Bølling-Allerød warm period and is equivalent to values in the last glacial period.

Moreover, there are two other dominant features in Figure 2. First, the extended summer (May to September) precipitation anomalies in SWC during the CWP is equivalent to that in the LGM period (Figure 2). Second, there is an upward trend of the extended summer (May to September) precipitation anomalies in Northwest China (NWC) over the past 549 years (1470–2018 CE; Figures 2 and S2), which is consistent with the tree-ring width record (B. Yang et al., 2014) and the tree-ring $\delta^{18}\text{O}$ record (G Xu et al., 2015). The possible mechanism for these two features is explored in Section 4.

After standardizing the nonlinear long-term trends of summer temperatures for the eight subregions in China (Figure 3a), we see that summer temperatures increased at first, reached a maximum in the Middle Holocene, and then gradually decreased. The maximum values of the nonlinear temperature trends occur in broadly the same period (8.1–7.9 ka) in the eight subregions. The slowly increasing and then decreasing trends in precipitation (Figure 3b) are similar to those for temperature. The maximum value of precipitation with a nonlinear trend falls broadly in the same period (7.9–7.6 ka) in the eight subregions (Figure 3b). In addition, the magnitude of temperature variations differs among the eight subregions (Figure 3a), and the temperature variations in the QTP and SWC are smaller than in the other regions, and the precipitation variation in SWC is also less than in the other regions. Overall, the co-varying long-term trends of temperature and precipitation anomalies show peaks that occur at a similar time (8.1–7.6 ka), which is independent of the choice of filter and how the filters are applied (see Figure S5 and Text S5 in the supporting information), and is also within the range of the thermal maxima of the Holocene epoch (10,000 to 6,000 years ago) from the marine sediment cores (Bova et al., 2021).

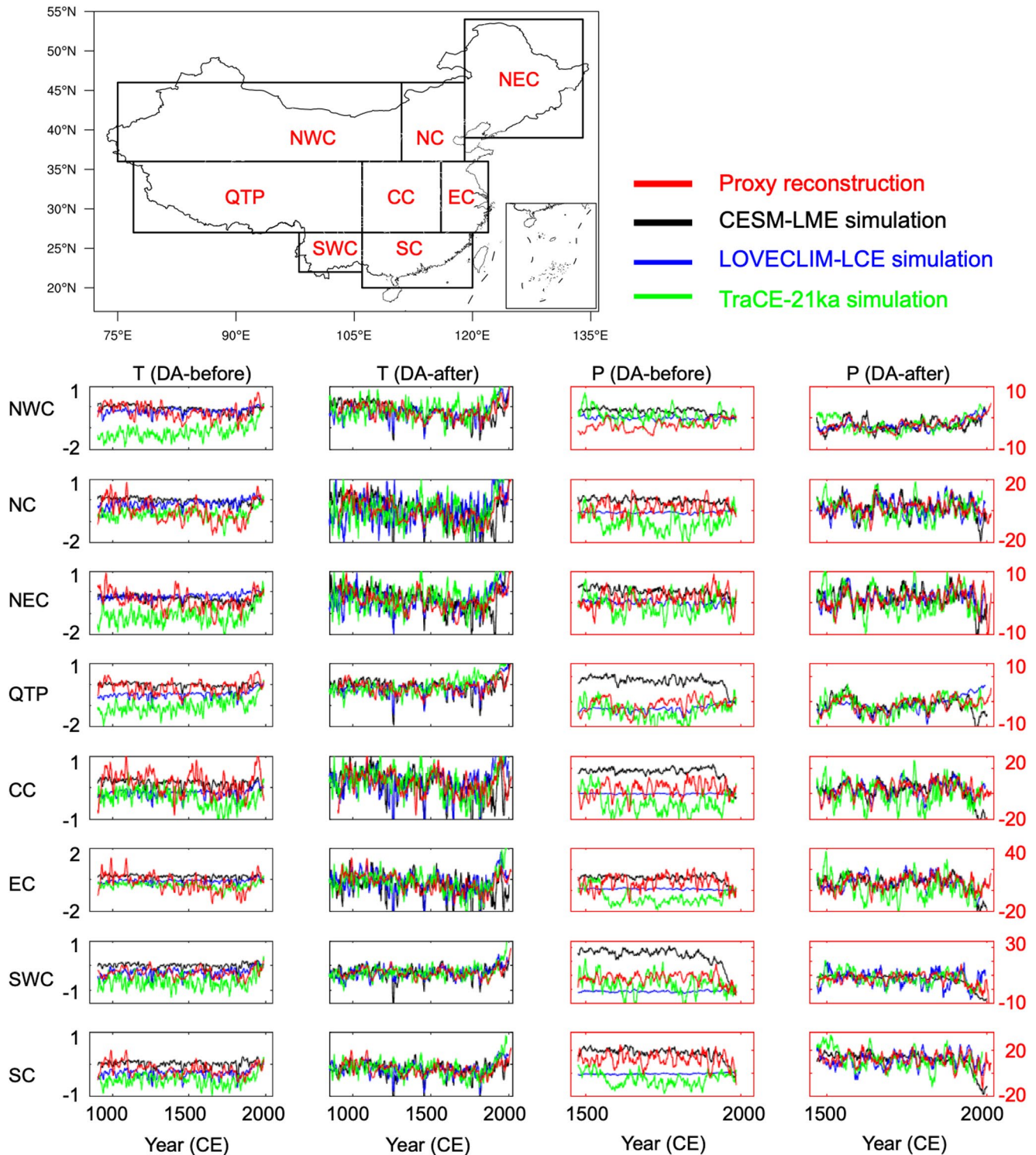


Figure 1. Comparison of the three model climate simulation results before and after data assimilation (DA) with the proxy reconstructions for the eight subregions of China shown in the map. Left panels show summer (June to August) temperature anomalies (black y-axes; units: °C) and right panels show extended summer (May to September) precipitation anomalies (red y-axes; mm/month) with respect to 1961–1990 CE. Red line: proxy reconstructions; black line: CESM-LME simulation; blue line: LOVECLIM-LCE simulation; green line: TraCE-21ka simulation. Northeast China (NEC; 39°N–54°N, 119°E–134°E); North China (NC; 36°N–46°N, 111°E–119°E); East China (EC; 27°N–36°N, 116°E–122°E); Central China (CC; 27°N–36°N, 106°E–116°E); South China (SC; 20°N–27°N, 106°E–120°E); Qinghai-Tibetan Plateau (QTP; 27°N–36°N, 77°E–106°E); Southwest China (SWC; 22°N–27°N, 98°E–106°E); and Northwest China (NWC; 36°N–46°N, 75°E–111°E). CESM-LME, Community Earth System Model-Last Millennium Ensemble; LOVECLIM-LCE, LOch-Vecode-Ecbilt-CLio-agIsm Model-Large Common Era Ensemble.

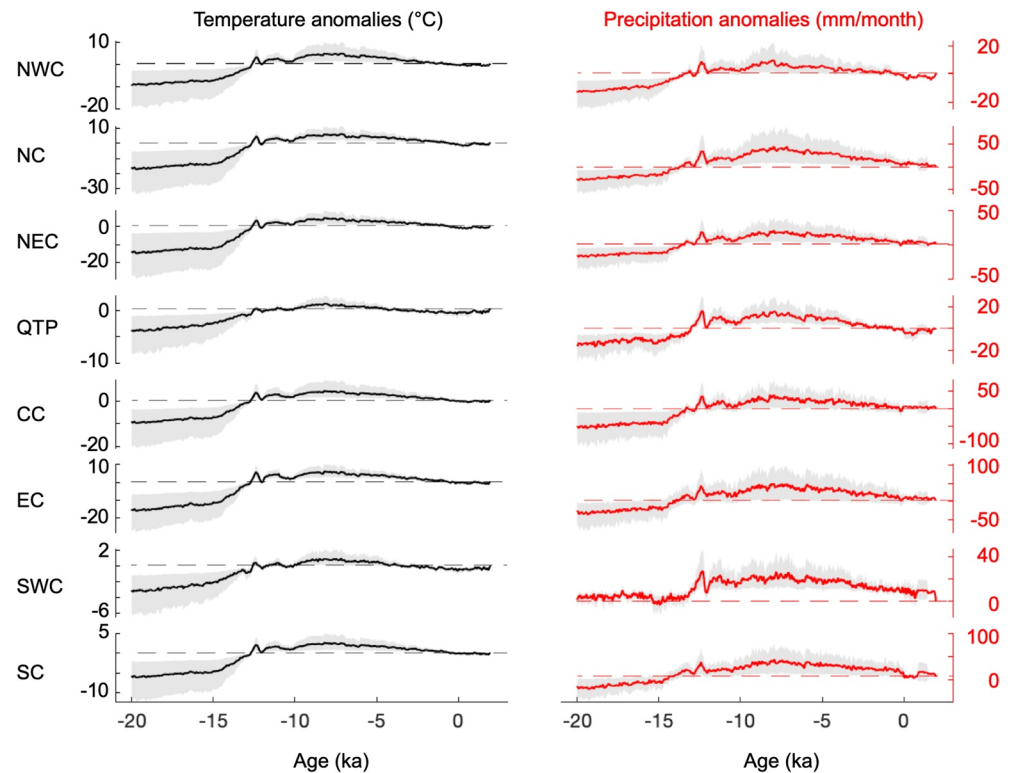


Figure 2. Comparison of summer (June to August) temperature anomalies (black line; units: °C) and extended summer (May to September) precipitation anomalies (red line; units: mm/month; with respect to 1961–1990 CE) from reconciliations for the interval 19,999 BCE to 1969 CE with uncertainty (gray shading represents the range from the 10th to 90th percentiles of the ensemble members). The dotted line is the average temperature and precipitation during the Current Warm Period (1919–2018 CE). Data have been smoothed using a 99-year moving average.

4. Discussion

The TraCE-21ka model output also shows no difference between the timing of the Holocene peak summer temperature over all of China and of summer precipitation in northern China (Liu, Wen et al., 2014c; Liu, Zhu et al., 2014b). This is because the summer temperature and precipitation changes are both dominated

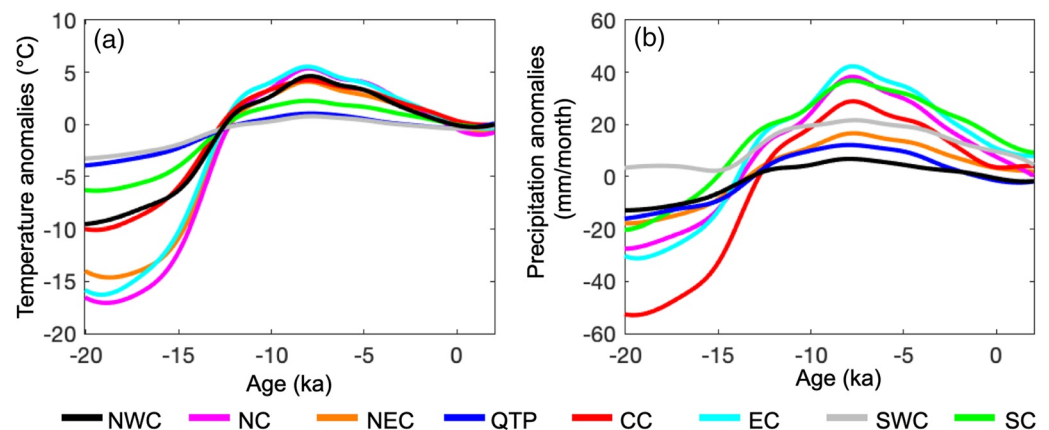


Figure 3. Nonlinear long-term trends in summer (June to August) temperature anomalies (units: °C) (a) and extended summer (May to September) precipitation anomalies (units: mm/month) (b) (with respect to 1961–1990 CE) in eight subregions in China. Orange line: NEC; magenta line: NC; cyan line: EC; red line: CC; green line: SC; blue line: QTP; gray line: SWC; and black line: NWC. CC, Central China; EC, East China; NC, North China; NEC, Northeast China; NWC, Northwest China; QTP, Qinghai-Tibetan Plateau; SC, South China; SWC, Southwest China.

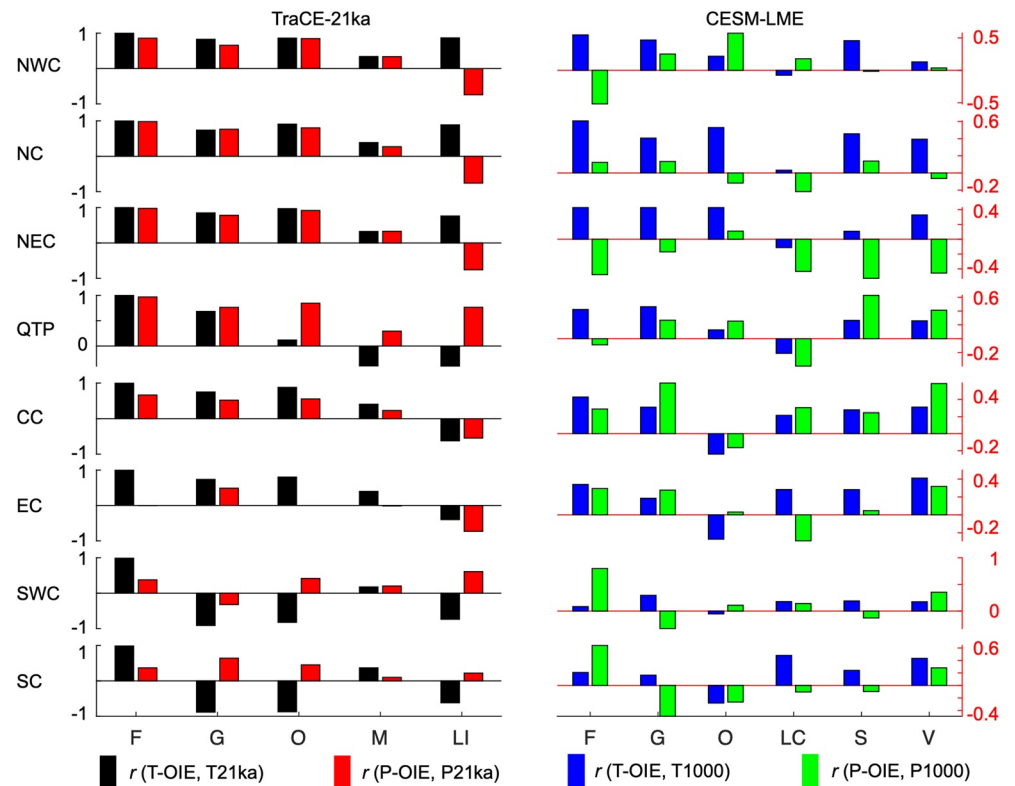


Figure 4. Correlation between the OIE-based summer (June to August) temperature anomalies (T-OIE) and extended summer (May to September) precipitation anomalies (P-OIE) in eight subregions in China and the TraCE-21ka simulations with different forcings as well as the CESM-LME simulation. F: ensemble mean of the full forcings; G: ensemble mean of greenhouse gas forcing only; O: ensemble mean of orbital forcing only; M: ensemble mean of meltwater forcing only; LI: ensemble mean of land ice-sheet only; LC: ensemble mean of land change only; S: ensemble mean of solar forcing only; and V: ensemble mean of volcanic forcing only. T21ka and P21ka are temperature and precipitation of the TraCE-21ka simulations, respectively; T1000 and P1000 are for the CESM-LME simulations. The “ r ” means the correlation coefficient between the simulation and data assimilation. CESM-LME, Community Earth System Model-Last Millennium Ensemble; OIE, optimal information extraction; TraCE-21ka, Transient Climate of the Last 21,000 years.

by direct insolation heating over land (Liu et al., 2003), leading to synchronous precipitation and temperature maxima. However, the summer precipitation changes in three subregions (East China [EC], South China [SC], and SWC) have different trends (Liu, Wen et al., 2014c). In addition, more complex variations were found in proxy records. Some studies suggest asynchronous Holocene optima for proxy records in the East Asian monsoon region (An et al., 2000), whereas others suggest the opposite (Xiao et al., 2002). Our results also show similar trends in NWC, SWC, and SC ($r > 0.85$) but opposite trends in Central China (CC) and EC ($r < 0.70$; Figure S6), compared with a recent integrated gridded proxy-based annual precipitation dataset (Herzschuh et al., 2019). This indicates that the DA provides brand new information to understand the mismatch of model-data comparison since the LGM, based on the accurate match of model-data comparison during the last millennium.

Comparing the OIE-reconstructed summer climate in eight subregions in China and the different model outputs (including the full-forcing TraCE-21ka experiment with the single-forcing CESM-LME experiment; Figures 4 and S7) allows us to explore the mechanism of the summer climate variability since the LGM. The influences of greenhouse gas and orbital forcing on temperature and precipitation are significant (p-value < 0.10), mainly in northern China (NWC, North China [NC], and Northeast China [NEC]; Figure 4, “G” and “O”). The climate variations of the full-forcing experiment have similar trends to the single solar forcing experiments (Figure S7), indicating that orbital forcing plays a dominant role in the temperature and precipitation variations in northern China. Over the past 1,146 years (855–2000 CE), greenhouse gas

forcing plays a significant role (p -value < 0.1) not only in northern China, but also in the QTP and SWC (Figure 4, “G”).

The significant influences of land ice-sheet forcing on temperature and precipitation in northern China are opposite (Figure 4, “LI”). The reason is that Figure 4 shows the correlation coefficients (LI) of temperature and precipitation in northern China (NWC, NC, and NEC) are all opposite). The meltwater forcing is responsible for extreme climate events (e.g., the Younger Dryas event; Liu, Lu et al., 2014a), but it has no significant effect ($r < 0.20$) on the long-term climate trends in the eight subregions of China since the LGM (Figure 4, “M”). Moreover, there are no consistent relationships between the OIE-based reconstruction and the different forcing simulations since the LGM in southern China, especially for the past millennium. The OIE-based reconstruction shows that orbital forcing plays a decisive role in the climate transition from glacial to interglacial (Figure 4, “O”). However, the internal variability plays a major role in the climate over the past millennium, although it cannot be represented by a single-member simulation in the TraCE-21ka simulation. This implies that more member simulations are needed to represent the internal variability over centennial to millennial scales. More information on the influences of different forcings is given in Text S6 of the supporting information.

Moreover, the extended summer (May to September) precipitation anomalies in SWC during the CWP is extremely small (Figure 2). According to the instrumental climate research community, some drought events (the summer of 2006, the autumn of 2009 to the spring of 2010, and the late summer of 2011, and the spring of 2013) have occurred in recent decades (Wang et al., 2015). The mechanism is related to the internal variability (including the Interdecadal Pacific Oscillation [Wang et al., 2018], the North Atlantic Oscillation [Feng et al., 2014], and the Central Pacific El Niño [Zhang et al., 2013]). However, there is a distinct downward trend over the centennial scale in the extended summer precipitation in SWC since the early 20th century (Figure 2), which cannot be attributed to the aforementioned internal variability, because the timescales are different. This downward trend may be related to anthropogenic-forced changes in ozone and aerosols, as indicated by the comparison between the OIE-based precipitation reconstruction and the CESM-LME single-forcing and full-forcing simulations in SWC (Figure S8).

In contrast, there is an upward trend of the extended summer (May to September) precipitation in NWC over the past 549 years (1470–2018 CE; Figures 2 and S2). The instrumental record also shows a significant wetting trend in summer during 1961–2010 CE, which is related to the changes of evaporation in response to global warming (Peng & Zhou, 2017). The comparison between the OIE-based precipitation reconstruction and the CESM-LME single-forcing and full-forcing simulations (Figure S9) indicates that greenhouse gas forcing is the main contributor to this upward trend of precipitation in NWC.

5. Conclusions

Our results show that the CWP in eight subregions of China is neither the warmest nor the wettest over the past 22,000 years. Variations in summer temperature in these regions are very similar. The magnitude of variations in precipitation during the CWP in the eight subregions is also very similar to that seen during the LGM in these regions. The summer climate since the LGM in northern China is mainly determined by orbital forcing, while greenhouse gases and the land ice-sheet have less influence and slightly modulate the timing and magnitude of the Holocene optimum climate in northern China. This indicates that the climate since the LGM in northern China is more sensitive to nonanthropogenic external forcings, but the summer precipitation in SWC since the early 20th century may be linked with anthropogenic-forced changes in ozone and aerosols, and the upward trend of precipitation in NWC over the half past millennium is also significantly affected by the role of greenhouse gas forcing. Our newly developed reconstructions provide a basis to test other proxy-only-based climate reconstructions and climate model simulations.

Data Availability Statement

The TraCE-21ka and the CESM-LME datasets are available at: <https://www.earthsystemgrid.org/project/trace.html> and https://www.earthsystemgrid.org/dataset/ucar.cgd.cesm4.CESM_CAM5_LME.html (with a free register). The reconstructed summer temperature reconciliations and extended summer precipitation

reconciliations in the eight regions during the period 20,045 BCE to 2013 CE in Figure S3 are available at: <https://www.ncdc.noaa.gov/paleo/study/32492>. The other climate reconstructions are available at: <https://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/climate-reconstruction> (with a name search).

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