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Significant recent warming over the northern Tibetan Plateau from ice core δ^{18}O records

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Abstract

Stable oxygen isotopic records in ice cores provide valuable information about past temperature, especially for regions with scarce instrumental measurements. This paper presents the $\delta^{18}O$ result of an ice core drilled to bedrock from Mt. Zangser Kangri (ZK), a remote area on the northern Tibetan Plateau (TP). Combining the ZK $\delta^{18}O$ records with those from three other ice cores in the region (Muztagata, Puruogangri and Geladaindong), we reconstructed the regional temperature history covering 1951–2008 for the northern TP. The reconstruction showed significant warming at 1.12 ± 0.08 to $1.31 \pm 0.10\,^{\circ}C(10\,yr)^{-1}$ since 1970, a much higher rate than the trend of instrumental records of the northern TP ($0.45 \pm 0.06\,^{\circ}C(10\,yr)^{-1}$) and the global temperature trend ($0.28 \pm 0.02\,^{\circ}C(10\,yr)^{-1}$) at the same time. Moreover, the rapid warming remained for this region during the last decade, when the mean global temperature showed very little change. Our study suggests that the temperature variations at high elevations could behave differently due to specific topographic and circulation mechanisms.

1 Introduction

With an average elevation over 4000 m a.s.l., the Tibetan Plateau (TP) is the highest and most extensive highland in the world. In recent decades, it has experienced rapid warming and drastic environmental changes such as fast glacier retreat and land deterioration (Cyranoski, 2005; Yao et al., 2012). The rapid warming over the Plateau was established with data from meteorological stations as well as various paleoclimate proxies located mostly in the eastern and southern Plateau (Thompson et al., 2000; B. Yang et al., 2014; Liu et al., 2013; Herzschuh et al., 2010; Pu et al., 2011). The northern part of the Plateau (northern TP, Fig. 1) is a climatologically important region involving complicated interactions between the mid-latitude westerlies and the subtropical Asia monsoon circulation, which may serve as a bridge linking the high and low latitude climatic processes (He et al., 2013). It is therefore essential to evaluate the

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extent and magnitude of regional climate changes over this region without coverage bias. However, there is generally a lack of climate data over this region, particularly in the northwest TP, where meteorological stations were sparse, and long-term high-resolution climate records were difficult to obtain because of the formidable terrain and harsh environment.

The global average surface temperature has experienced relatively little change since it hit a record high in 1998, despite the continued increase in the atmospheric concentration of CO_2 and other greenhouse gases (Easterling and Wehner, 2009; Jones et al., 2009; Foster and Rahmstorf, 2011; Kaufmann et al., 2011). This recent warming hiatus was established from instrumental records of surface temperature around the world, and bias could arise from the uneven spatial coverage (Cowtan and Way, 2014), in particular, the lack of records in crucial high elevation regions. The past decade was the warmest decade on record for the TP with surprisingly high temperature values (Yang et al., 2009; You et al., 2010). It is important to understand the behavior of recent warming hiatus over the high elevation regions by the analysis on the TP.

The ice core δ^{18}O is an unique paleoclimate proxy on the TP (Thompson et al., 2000; Yao et al., 2000; Qin et al., 2002; Kang et al., 2007; Joswiak et al., 2010), and has been generally considered to be a reliable indicator for past temperatures (Yao et al., 2006). However, great discrepancies existed among different temperature reconstructions and observational temperature records owing to the distinct geographic conditions and atmospheric circulation systems (Liu and Chen, 2000; N. Wang et al., 2003; Yao et al., 2006; Y. Q. Wang et al., 2003). Therefore, it is important to establish more high resolution temperature records on the TP, and evaluate the extent of the recent global warming hiatus, particularly over the extensive high elevation regions with sparse instrumental climate records, such as the northern TP. In this study, we measured the δ^{18}O values in an ice core drilled from the Zangser Kangri (ZK) glacier on the northern TP, from which temperature changes in the past decades could be established. The ZK ice core δ^{18}O records made it possible to study the past climate variations over a relatively inaccessible part of the TP, where instrumental records are very limited.

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Combined with the $\delta^{18}O$ time series from other ice cores in adjacent areas, we were able to evaluate the climate change at the regional scale.

2 Methodology and data

The ZK glacier is located in the west-northern part of the TP, covering an area of 337.98 km^2 with a volume of 41.70 km^3 (2005 data). The snowline is about 5700–5940 m a.s.l. In April of 2009, two ice cores to bedrock (127.7 and 126.7 m in length for Core 1 and Core 2, respectively) were recovered from the glacier (34°18′05.8 N, 85°51′14.2 E, 6226 m a.s.l., Fig. 1). The temperature of the ZK glacier was far below 0 °C throughout the borehole. The glacier temperature ranged from −15.2 to −9.2 °C, with a mean temperature of −11.7 °C, 10 m temperature of −12.4 °C and basal temperature of −9.2 °C.

These two ice cores were kept frozen and transported to the State Key Laboratory of Cryospheric Science (SKLCS) for processing. This study was based upon the analysis of Core 1. A total of 2884 samples were taken from Core 1 at a resolution of 4–6 cm. The outer ~ 2 cm of each sample was removed for stable oxygen isotope analysis. The inner portion of the ice core was collected in pre-cleaned polyethylene sample containers for chemical and dust particle analyses. Stable oxygen isotope ratio ($\delta^{18}O$) was determined using a Picarro Wavelength Scanned Cavity Ring-Down Spectrometer (WS-CRDS, model L2120i). Major cations and anions were analyzed using a Dionex-600 and ICS-2500 ion chromatograph respectively. Ice core dating was performed based on the seasonality of $\delta^{18}O$ and major ions with a reference layer of β activity peak in 1963 (Fig. 2a). This study focused on the measured time series of $\delta^{18}O$ in the top 16 m of the ice core, corresponding to the time period 1951–2008.

The annual $\delta^{18}O$ values in ice cores are not only affected by temperature but also the seasonality of precipitation. Therefore, a significant change in seasonal distribution of precipitation could change the $\delta^{18}O$ values even when temperature remains constant. In order to make sure that the ZK $\delta^{18}O$ values mainly reflect temperature

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change, we first examined whether precipitation seasonality experienced any significant changes during the study period by using the precipitation records from the two closest meteorological stations, at Gêrzê (32°09′, 84°25′, 4414.9 m a.s.l., 1973–2008) and Xainza (30°57′, 88°38′, 4800 m a.s.l., 1961–2008) (Fig. 1). We established the time series for the proportions of both summer and winter precipitation from 1961 to 2008, and both of them showed no statistically significant trends (Supplement Fig. S1). Therefore, changes in precipitation distribution should have very little impact on the δ^{18}O values in ZK ice cores, and that they mainly record temperature signals.

In order to establish the representativeness of the ZK ice core δ^{18}O for the regional climate, we performed correlation analysis, using Pearson's correlation coefficient (r), between the ice core δ^{18}O time series and temperature records from the nearby meteorological stations (Gêrzê and Xainza), and the instrumental temperature series from meteorological stations in the northern TP (hereafter, ITNTP). The ITNTP time series was derived from 14 climate stations used in Guo and Wang (2011), and was extended to 2014 based on the data provided by the Data and Information Center, China Meteorological Administration. Most of the stations used in ITNTP time series were located on the eastern part of the northern TP with altitudes ranging from 2110.5 to 4700 m (Guo and Wang, 2011), whereas this study focused on the higher (> 5700 m) and more extensive part of the northern TP (Fig. 1). Spatial correlations were also carried out between ZK δ^{18}O and the CRU 4 gridded temperature reanalysis data (Mitchell and Jones, 2005) on the KNMI Climate Explorer (http://climexp.knmi.nl). Linear regressions with time series were performed to quantify the changes of ZK δ^{18}O and the δ^{18}O time series of four nearby ice cores. We employ non-parametric LOESS regression technique (Cohen, 1999) to estimate the trend of the global temperature, the instrumental temperature record of the northern TP (ITNTP), and the regional temperature reconstruction.

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3 Results

3.1 The $\delta^{18}O$ variations in ZK ice core

The raw $\delta^{18}O$ values from 1951 to 2008 showed distinctive seasonal variations that could be clearly observed throughout the profile. The $\delta^{18}O$ values in the top 16 m of this core ranged from −17.65 ‰ at 13.8 m to −3.79 ‰ at 6.85 m, with an average value of −10.97 ‰ (Fig. 2).

The ZK $\delta^{18}O$ time series showed positive correlation with annual temperature measured at nearby stations ($r = 0.31$, $p = 0.07$ for the Gêrzê station; $r = 0.43$, $p = 0.002$ for the Xainza station), as well as the averaged annual temperature of the two stations ($r = 0.34$, $p = 0.01$) and ITNTP ($r = 0.35$, $p = 0.02$) (Table 1). Stronger correlation existed between the ZK $\delta^{18}O$ and spring (March–May) temperature of the stations (Table 1), suggesting more influence of spring temperature on the ZK $\delta^{18}O$ values. Linear regressions led to a mean $\delta^{18}O$-temperature slope of 0.85 ‰ °C^{-1} with values ranging from 0.67 to 0.98 ‰ °C^{-1} for annual values (Table 1). This is consistent with the published $\delta^{18}O$-temperature relationships derived from ice cores over the northern TP (X. X. Yang et al., 2014). Spatial correlation analysis between the ZK $\delta^{18}O$ series and the CRU gridded temperature data also revealed significantly positive correlations in the region surrounding the drilling site. The ZK $\delta^{18}O$ series showed positive correlations with annual mean and minimum temperatures for most part of the northern TP (Fig. 3). The most significant and spatially extensive correlations were found between the ZK $\delta^{18}O$ and spring temperatures (Fig. 3c and d), which was consistent with the correlation results between the ZK $\delta^{18}O$ series and station temperature records (Table 1).

The standardized values of $\delta^{18}O$ time series from 1951 to 2008 of the ZK ice core is presented in Fig. 4a. The $\delta^{18}O$ values were relatively low in the 1960s, followed by an increasing trend from 1970s to the end of our record (Fig. 4a). Further comparisons were made between the ZK $\delta^{18}O$ and the $\delta^{18}O$ time series of four nearby ice cores, including Muztagata (Tian et al., 2006), Puruogangri (Yao et al., 2006), Geladaindong

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(Kang et al., 2007) and Malan (N. Wang et al., 2003) (Fig. 4 and Table 2). The cooling around 1960s was present for all ice cores, and this was consistent with the observed cold period during this period over the entire TP (Yang et al., 2009). Moreover, the significant increasing trend from 1970s to present was observed in the ZK, Muztagata, Puruogangri and Geladaindong ice core δ^{18}O time series. We calculated the Pearson correlation coefficients among these ice core δ^{18}O series (Table 3). The results showed weak correlations between the annual values of these series due to the differences in location, elevation and uncertainties in ice core dating. However, the 5 year running averages of these series showed stronger correlation, suggesting possible common forcings preserved in the ZK, Muztagata and Puruogangri series. This coherence is important when we use the average of multiple sites to develop a regional composite. The lack of correlation between Geladaindong and the other three ice cores could reflect its unique local climate conditions (Table 3). We compared the regional average ice core δ^{18}O series with and without Geladaindong (Fig. S2). The two series were highly correlated ($r = 0.95$, $p < 0.0001$), with little differences in variations and trend (Fig. S2). Therefore, despite the difference, we decided to include the Geladaindong ice core δ^{18}O so that the final regional reconstruction could better represent the regional climate of the northern TP.

The Malan δ^{18}O record, on the other hand, revealed a cooling trend since 1970s (Fig. 4e). Such continuous low level of δ^{18}O could be caused by the change of local climate conditions (Y.Q. Wang et al., 2003), but could also be caused by post-depositional processes on the chemical profiles, such as summer melting, evaporation and condensation, all of which could modify the relationship between ice core δ^{18}O and temperature (Hou et al., 2006). Furthermore, the correlation analysis showed that the Malan time series was negatively correlated with other four time series, and the negative relationships were more significant after 5 year running averaging (Table 3). Therefore, we excluded the Malan record from further analysis.

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3.2 Regional temperature reconstruction

The ice cores of Muztagata, ZK, Puruogangri and Geladaindong shared similar patterns of δ^{18}O variations, especially the increasing trends since 1970s (Fig. 4). Therefore, it was reasonable to reconstruct a regional temperature series based on the integration of the four ice core δ^{18}O records. First of all, it was necessary to derive the δ^{18}O-temperature relationship to understand the magnitude of the temperature variation over the northern TP. Yu et al. (2009) calculated the isotope sensitivity between monthly mean δ^{18}O values in precipitation and the monthly mean temperatures at Gêrzê and Shiquanhe (Fig. 1) as 0.33 and 0.37 ‰ $°C^{-1}$ respectively. State of the art atmospheric models with integrated water isotopes modeling suggested an isotope sensitivity of 0.53 ‰ $°C^{-1}$ for the present precipitation falling at the grid where the ZK core was recovered (Risi et al., 2010). Tian et al. (2006) used the range of 0.6 to 0.7 ‰ $°C^{-1}$ δ^{18}O-temperature relationship to transform the δ^{18}O notation of the Muztagata ice core. Moreover, the isotope sensitivity usually increases with elevation as indicated by Rayleigh-type equilibrium fractionation model (Rowley et al., 2001). Therefore, in this study, we also applied the range of 0.60 to 0.70 ‰ $°C^{-1}$ for the regional temperature reconstruction for the northern TP. This temperature reconstruction was likely to represent the low end of the range due to the lower elevations of other three ice cores (Table 2) than the Muztagata site (7010 m a.s.l.).

Signal noise in the ice core records and abnormal climate fluctuations including the extreme temperature and precipitation events, could lead to bias in δ^{18}O values (X. X. Yang et al., 2014). Moreover, the four ice cores used in this study cover an extensive region from west to east, and the location difference may result in the difference in moisture origin for the four ice core sites hence the difference in δ^{18}O variations. Therefore, we first established the δ^{18}O anomalies for each ice core series to eliminate the bias present in the absolute values, particularly for years with missing data because of the difference in the length of the record. We then took the average of the δ^{18}O anomalies (Fig. S2), and used the mean values to reconstruct the regional temperature anomalies time series using 0.60 to 0.70‰°C^{-1} δ^{18}O-temperature relationship. This regional temperature reconstruction (Fig. 5a) was first compared with the ITNTP (Fig. 5b). The comparison showed that the regional reconstruction was closely correlated with the ITNTP temperature record ($r = 0.72$, $p < 0.001$). Moreover, significant correlations were found between the CRU gridded surface temperatures and both the ITNTP ($r = 0.50$ to 0.70, $n = 47$, $p < 0.01$) and the reconstructed temperatures ($r = 0.40$ to 0.70, $n = 58$, $p < 0.01$) over a large region. The study area had the strongest correlations ($r > 0.50$, $p < 0.01$) (Fig. 6). This suggested that the reconstruction adequately captured temperature variation on the northern TP.

3.3 Recent rapid warming trend over the northern TP

The regional reconstruction was further compared with the global annual temperature series derived from HadCRU4v (http://www.cru.uea.ac.uk/cru/data/temperature/) (Fig. 5c) and the ITNTP (Fig. 5b) in order to investigate the recent warming trend since 1970s. The LOESS regression was exploited to smooth the data and estimate the general trend. The reconstruction captured the cooling period during 1960s, as well as the prominent warming from the 1970s to the end of the record with the highest rates of increase in the late 1990s (Fig. 5). Compared with the warming rate of global temperature (0.28 ± 0.02 °C(10 yr)$^{-1}$) from 1970 to 2008, the regional reconstruction showed more rapid warming trend at the rate of 1.12 ± 0.08 to 1.31 ± 0.10 °C(10 yr)$^{-1}$ (corresponding to 0.60 to 0.70‰°C^{-1} δ^{18}O-temperature relationship respectively). These rates were also higher than the ITNTP rate of increase at 0.45 ± 0.06 °C(10 yr)$^{-1}$ (1970–2008). From 1990 to 2008, the warming accelerated on the northern TP with much larger rates at 1.57 ± 0.22 to 1.83 ± 0.26 °C(10 yr)$^{-1}$ for the regional reconstruction, and 0.64 ± 0.16 °C(10 yr)$^{-1}$ for the ITNTP, much higher than the global average (0.34 ± 0.06 °C(10 yr)$^{-1}$). These results demonstrated that elevation dependency of the climatic warming is evident for the high elevation regions on the northern TP.

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The global mean temperature reached a record high in 1998 and experienced relatively little change since then despite the continued increase in the atmospheric concentration of CO_2 and other greenhouse gases (Easterling and Wehner, 2009). The regional temperature reconstruction recorded a high value in 1996, two years earlier the high temperature recorded in ITNTP and the global temperature in 1998. This time discrepancy could be caused by the uncertainties in the ice core dating and did not influence the results of trend analysis. In Fig. 5c, the global temperature showed very little change since 2000 and even decreasing trend since 2005. However, the regional temperature reconstruction revealed a continuous significant increasing trend from 1980 until the end of the record in 2008 (Fig. 5a). The continuous warming trend was also recorded in the ITNTP (Fig. 5b). We calculated mean decadal annual temperature change based on the LOESS regression model for all three time series (Fig. 7). The global temperature had the highest average warming rate during 1990s, and then showed much reduced warming rates since 1999 (Fig. 7c). Meanwhile, the rates of increase remained high for the temperature records in the reconstruction and the ITNTP since 1999 (Fig. 7a and b). This seemed to suggest that recent global warming hiatus was not observed at high elevation regions in the northern TP. In fact, the observation of the past decade with little or no surface temperature warming since 1998 was based on global instrumental measurements with potential biases due to the lack of coverage for high latitudes and altitudes (Cowtan and Way, 2014). Our results suggest that this phenomenon should be evaluated in more depth at high elevations.

4 Discussion

Many mechanisms have been proposed to explain the observed warming hiatus since 1998, including prolonged solar minimum (Kaufmann et al., 2011), changes in atmospheric water vapor and aerosols (Solomon et al., 2011), volcanic forcing on tropospheric temperature (Santer et al., 2014) and increased ocean heat uptake (Meehl et al., 2011). However, the above factors may account for the reduced warming rates at the global scale, but could not explain the spatial and temporal discrepancies at the regional scale, such as the rapid temperature rise on the northern TP during the recent global warming hiatus period. Several studies suggest that the recent cooling in the equatorial Pacific associated with the strengthening of trade winds (England et al., 2014) may play a central role in the slowdown of the rate of global surface warming over the last decade (Trenberth and Fasullo, 2013; Kosaka and Xie, 2014; Tollefson, 2014). Chen and Tung (2014) proposed that the slowdown was mainly caused by the transport of heat into deeper layers of the Atlantic and the Southern oceans. Generally, it was believed that the natural variability in the Atlantic and Pacific multidecadal oscillations contributed to recent warming hiatus (Watanabe et al., 2014; Steinman et al., 2015). If these ocean processes played a central role in the recent global warming hiatus, it is possible that the relatively large distance between the high elevation northern TP and the equatorial ocean processes could reduce or delay their influences, hence caused the lack of reduced warming on the northern TP.

Some studies addressed the link between the TP temperature variations and the variability of the equatorial central-eastern Pacific SST through the Asian monsoon system (Zhao et al., 2009; Nan et al., 2009). It is suggested that the marine air masses transported from the southwest monsoon cannot directly reach the northern TP, where continental recycling intensified by evaporation over the land surface is evident and influences the precipitation δ^{18}O, hence the ice core δ^{18}O (Tian et al., 2001; Yao et al., 2013). The indirect influence of southwestern monsoon could delay the response of northern TP to the ocean circulation. Moreover, the weakening relationships between Asian monsoon and Pacific SST since the 1970s (Chang et al., 2001) could imply a lag in the response of TP temperature to global ocean circulation, causing the discrepancy in temperature trends between the northern TP and the global average.

The lack of obvious warming hiatus over the northern TP could also be caused by regional processes in the recent decade. Previous studies suggest that annual and seasonal mean wind speeds show statistically decreasing trends during 1980–2005 over the TP (You et al., 2014), and that sensible heat flux also exhibited significant decline since the mid-1980s induced by decreased surface wind speed (Duan and Wu, 2009). The wind stilling may have favored the continuous rapid warming over the study region by decreasing the efficiency of the energy exchange (K. Yang et al., 2014). In addition, Zhong et al. (2010) showed that vegetation density in the TP increased from 1998 to 2006, a possible response to precipitation increase in this region. The increased vegetation density may also have contributed to the continuous warming by reducing albedo and heat loss. Furthermore, the surface albedo decreases more at higher elevations than at lower elevations owing to the retreat of the 0 °C isotherm and the associated retreat of the snow line (Ghatak et al., 2014). The decreasing surface albedo could lead to enhanced warming at high-elevation regions. Detailed studies are needed to examine temperature variations of other regions over the Plateau, as well as other high elevation regions in order to provide a comprehensive evaluation.

The elevation dependency of climate warming trends over the TP has been reported in previous studies (Mountain Research Initiative EDW Working Group, 2015). The instrumental temperature records showed that elevation-dependent warming was enhanced over the TP since 2001 (Yan and Liu, 2014). However, the decadal temperature trends calculated from meteorological stations in the northern TP failed to follow the elevation-dependent warming pattern over the two periods of 1961–2014 and 1970–2014 (Fig. S3). Guo and Wang (2011) suggested that the radiative and dynamical heating induced by pronounced stratospheric ozone depletion may have contributed to the significant warming over the northern TP rather than the elevation dependency. However, the elevations of stations in the northern TP range from 2767 to 3367 m (with an average of 3011 m), which were lower than the average 4000 m for the entire plateau. The consistently high warming rates derived from our regional reconstruction demonstrate that the elevation-dependent warming is still evident on high elevation regions on the northern TP.

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5 Conclusions

This study presented a $\delta^{18}O$ time series of the ZK ice core from the northern TP. The ice core $\delta^{18}O$ record reflected local temperature variations. A regional temperature history from 1951 to 2008 was reconstructed based on a composite series derived from this ice core $\delta^{18}O$ and three other cores from the northern TP. The regional temperature reconstruction captured the rapid warming trend since 1970, and the continuous warming since 1998, which was different from the established warming hiatus for global temperature anomalies. The warming rate since 1998 was much higher in northern TP than that of the global average temperature. It indicated that the warming slowdown seen in global and hemispheric average temperature may be different in high elevation regions.

Possible explanations for this continued warming over the northern TP might lie in its relative distant location from the equatorial oceans, where processes such as strengthening of the trade wind and the heat transport into the deeper ocean were believed to account for the recent warming hiatus. In addition, the persistent wind stilling and the increasing vegetation density may also contribute to the continuous warming. This provided insights to understanding the behavior of the recent warming hiatus on high elevations with limited instrumental climate records. Further studies are still needed to understand the specific characteristics of this warming hiatus on the TP, as well as the response mechanisms of high elevations regions to global changes.

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Table 1. Correlation coefficients and linear slopes between $\delta^{18}O$ values in the ZK ice core and instrumental spring (March–May) and annual temperature from closest Gêrzê (1973–2008) and Xainza stations (1961–2008), the averaging records of the two stations (1961–2008), and the ITNTP series (1961–2008).

	Gêrzê		Xainza		Stations averaging		ITNTP
	Mar–May	Annual	Mar–May	Annual	Mar–May	Annual	Annual
Correlation coefficients	0.52[c]	0.34[a]	0.45[c]	0.34[a]	0.48[c]	0.34[a]	0.35[a]
Slope	0.93[b]	0.67[a]	0.93[b]	0.98[a]	1.00[c]	0.88[a]	0.87[a]

[a] $p < 0.05$; [b] $p < 0.01$; [c] $p < 0.001$.

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Table 2. Basic information of ice cores from the northern TP.

Ice core	Muztagata	ZK	Purogangri	Geladaindong	Malan
Latitude (N)	38°17′ N	34°18′5.8″ N	33°54′ N	33°34′37.8″ N	35°50′ N
Longitude (E)	75°06′ E	85°51′14.2″ E	86°06′ E	91°10′35.3″ E	90°40′ E
Altitude (m)	7010	6226	6200	5720	5680

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Table 3. Correlation coefficients $\delta^{18}O$ values in the ZK (1951–2008), Muztagata (1955–2002), Puruogangri (1951–1998), Geladaindong (1951–2004) and Malan (1951–1999) ice cores, and the regional $\delta^{18}O$ values averaged from ZK, Muztagata, Puruogangri and Geladaindong ice cores. The values in bold are the correlation coefficients of annual values, and the values in italic are the correlation coefficients of 5 year running average values.

	ZK	Muztagata	Puruogangri	Geladaindong	Malan	Regional average
ZK		**0.26**	**0.14**	**–0.02**	**–0.27**	**0.68**[c]
Muztagata	*0.68*[c]		**0.09**	**0.04**	**–0.2**	**0.79**[c]
Puruogangri	*0.46*[c]	*0.28*		**–0.08**	**0.17**	**0.51**[c]
Geladaindong	*–0.07*	*0.24*	*–0.12*		**0.05**	**0.38**[c]
Malan	*–0.40*[b]	*–0.33*	*0.14*	*0.18*		

[a] $p < 0.05$; [b] $p < 0.01$; [c] $p < 0.001$.

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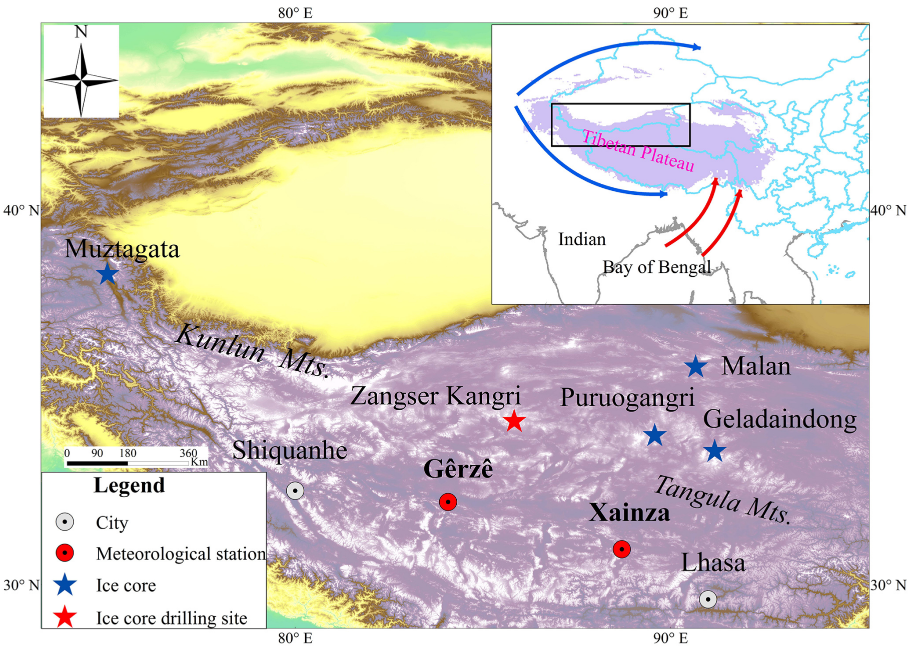

Figure 1. Location of the ice core drilling site of ZK, two nearby meteorological station sites, and the location of other ice cores described in the text: Muztagata (Tian et al., 2006), Puruogangri (Yao et al., 2006), Geladaidong (Kang et al., 2007b) and Malan (Wang et al., 2003) over the northern TP. The inset shows the relative location of the northern TP to the entire TP. Red arrows indicate the direction of the southwest monsoon in summer and blue ones indicate the dominant westerlies in winter. The black rectangle indicates the study region.

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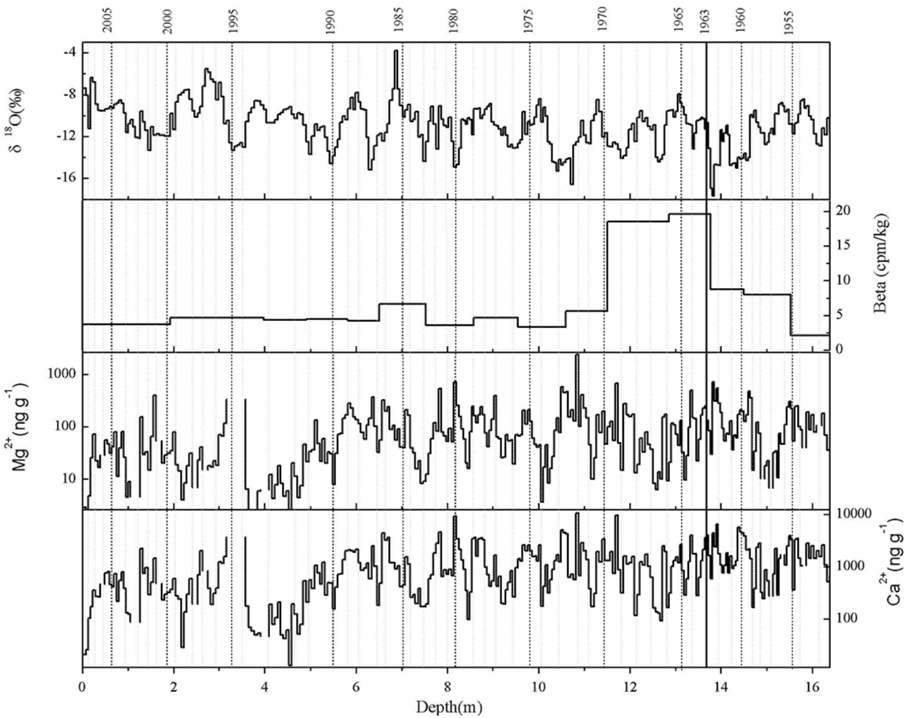

Figure 2. Variations of $\delta^{18}O$ in the ZK ice core and data used in dating: beta activity and major ion concentrations. We calculated the logarithm to the base 10 of the concentrations of the ions to facilitate dating.

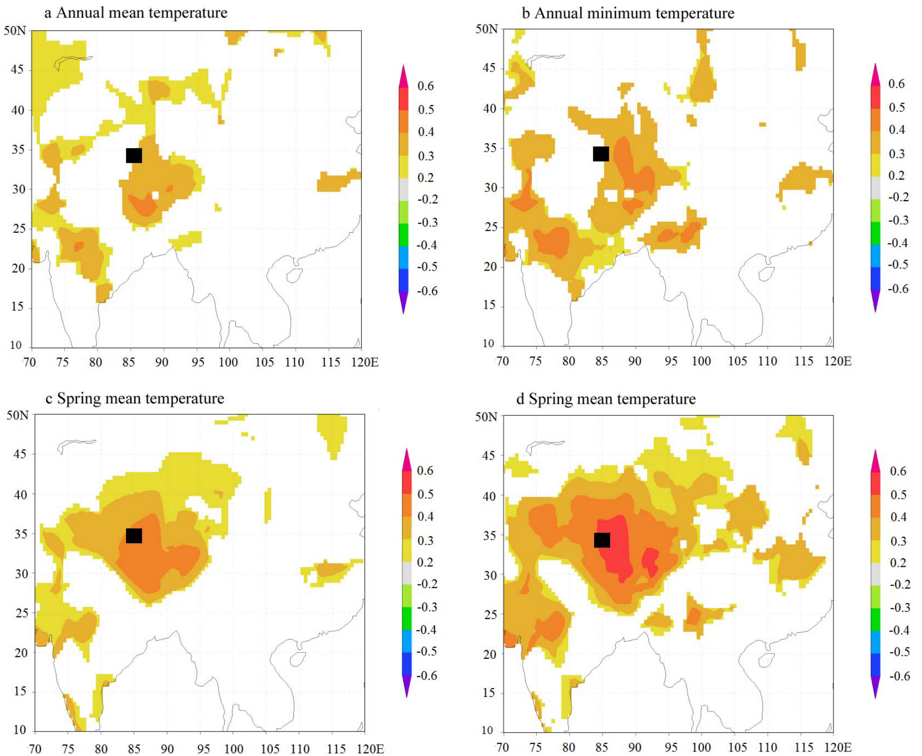

Figure 3. Spatial correlations of ZK ice core δ^{18}O record with CRU-gridded (Mitchell and Jones, 2005) annual mean temperature **(a)**, annual minimum temperature **(b)**, spring mean temperature **(c)**, and spring minimum temperature **(d)** for the period 1951–2008. Only coefficients of correlation significant at $p < 0.01$ are shown. The black rectangle indicates the ZK ice core site.

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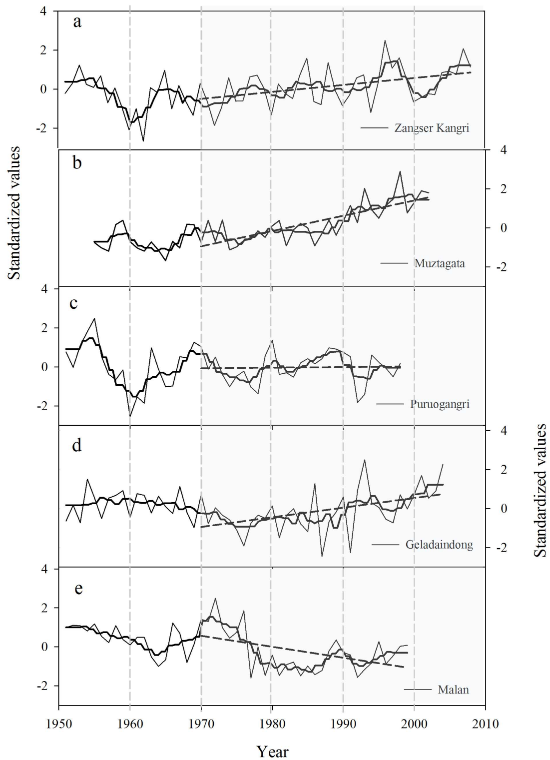

Figure 4. Comparisons of the standardized δ^{18}O records in the ZK ice core **(a)** with δ^{18}O records from Muztagata **(b)**, Puruogangri **(c)**, Geladaindong **(d)** and Malan ice cores **(e)**. Thin lines represent annual values, thick lines the 5 year running averages, and the dotted lines the linear trends since 1970.

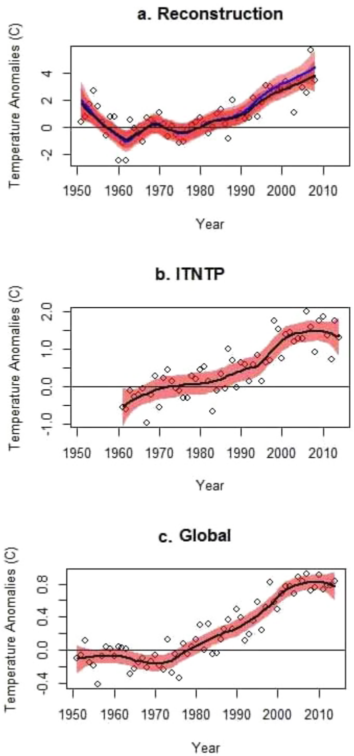

Figure 5. Temperature anomalies series of the regional temperature reconstruction series (the black and blue lines indicate reconstructed temperature anomalies derived from the δ^{18}O/T slope of 0.6 and 0.7 respectively) **(a)**, the instrumental temperature record of the northern TP (ITNTP) **(b)**, and global average temperature **(c)**. Trend lines were estimated using the non-parametric LOESS regression technique with a span of 0.4; shading represents the 95 % confidence intervals of the estimated trends.

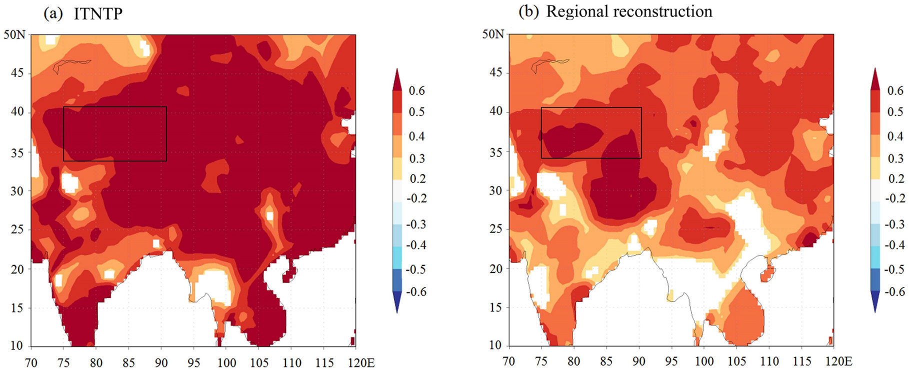

Figure 6. Spatial correlations (r values in color, $p < 0.01$) between the gridded annual mean temperature data (the CRU 4 temperature time series, $0.5^{\circ} \times 0.5^{\circ}$ resolution, Mitchell and Jones, 2005) and the instrumental temperature record of the northern TP (ITNTP) (Guo and Wang, 2011) for the period 1961–2007 **(a)**, and the regional temperature reconstruction series for the period 1951–2008 **(b)**. The analyses were conducted with the KNMI Climate Explorer software (Royal Netherlands Meteorological Institute, http://climexp.knmi.nl). The black rectangle indicates the study region in this study.

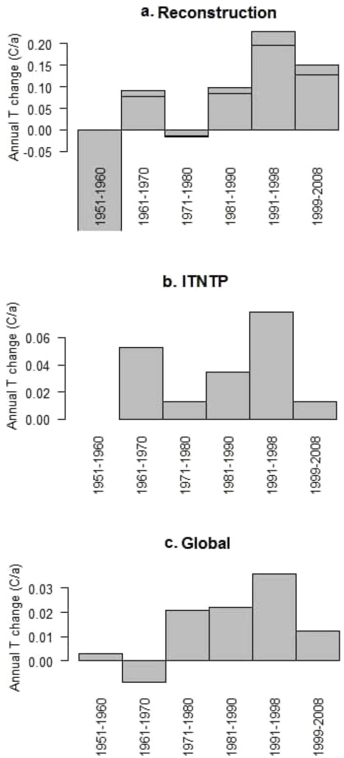

Figure 7. Decadal mean annual change rates derived from LOESS regression model for the regional temperature reconstruction (with the $\delta^{18}O/T$ slope of 0.6 and 0.7 respectively) **(a)**, the instrumental temperature record of the northern TP (ITNTP, Guo and Wang, 2011) **(b)**, and the global temperature **(c)**.

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