Significant recent warming over the northern Tibetan Plateau from ice core $\delta^{18}O$ 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). We reconstructed the temperature series for 1951-2008 from the $\delta^{18}O$ records. In addition, we combined the ZK $\delta^{18}O$ records with those from three other ice cores in the northern TP (Muztagata, Puruogangri and Geladaindong) to reconstruct a regional temperature history for the period 1951-2002 (RTNTP). The RTNTP showed significant warming at $0.51\pm0.07°C(10yr)^{-1}$ since 1970, a higher rate than the trend of instrumental records of the northern TP ($0.43\pm0.08°C(10yr)^{-1}$) and the global temperature trend ($0.27\pm0.03°C(10yr)^{-1}$) at the same time. In addition, the ZK temperature record, with extra length until 2008, seems to suggest that the rapid elevation-dependent warming continued for this region during the last decade, when the mean global temperature showed very little change. This could provide insights into the behavior of the recent warming hiatus at higher elevations, where instrumental climate records are lacking.

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 (Yao et al., 2012). In recent years, the global average surface temperature has experienced relatively little change in recent years (Easterlin and Wehner, 2009), whereas accelerated warming continued on the TP for the same period of time (Yan and Liu, 2014; Duan and Wu, 2015). However, the rapid warming trend over the Plateau was established with data from meteorological stations located at relatively low elevations, and warming trend for higher elevation regions remains uncertain.

In addition, spatial biases also exist in the TP temperature records. Most instrumental records as well as various paleoclimate proxies are located in the eastern and southern Plateau (Thompson et al., 2000; B. Yang et al., 2014; Herzschuh et al., 2010; Pu et al., 2011). There is generally a lack of climate data in the northern, and 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. However, the northern TP (Fig. 1) is a climatologically important region involving complicated interactions between the mid-latitude westerlies and the subtropical Asia monsoon circulation. It may serve as a bridge linking the high and low latitude climatic processes (Y. X. He et al., 2013). It is therefore essential to evaluate the extent and magnitude of regional climate changes over this region without coverage bias.

The ice core $\delta^{18}O$ is an important paleoclimate proxy on the TP (Thompson et al., 2000; Qin et al., 2002), and has been generally considered to be a reliable indicator for past temperatures (Yao et al., 2006; Joswiak et al., 2010). However, great discrepancies still exist among different temperature reconstructions and instrumental records owing to the distinct geographic locations and atmospheric circulation conditions (Liu and Chen, 2000; N. Wang et al., 2003; Y. Q. Wang et al., 2003; Yao et al., 2006). Therefore, it is important to establish more high resolution temperature records on the TP, particularly over such extensive high elevation regions as the northern TP, in order to evaluate the warming trends at high elevations in light of the recent warming hiatus. In this study, we measured the $\delta^{18}O$ 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 $\delta^{18}O$ records made it possible to study the past climate variations over a relatively inaccessible part of the TP, where instrumental records are very limited. In addition, we also established the regional climate change history by combining ZK with the $\delta^{18}O$ records from other ice cores in the northern TP.

2 Methodology and Data

2.1 Research area and ice core dating

The ZK glacier is located in the northwest part of the TP, covering an area of 337.98 km^2 with a volume of 41.70 km^3 (2005 data, Shi, 2008). The snowline is about 5700～5940 m a.s.l.. In the April of 2009, two ice cores to bedrock (127.7m and 126.7m 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 glacier temperature ranged from -15.2°C to -9.2°C, with a mean temperature of -11.7°C, -12.4°C at 10 m depth and a basal temperature of -9.2°C.

These two ice cores were kept frozen and transported to the State Key Laboratory of Cryospheric Sciences, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences 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.

In the northern TP, the annual cycle of $\delta^{18}O$ along the ice core profile is primarily related to temperature variations (Araguás-Araguás et al., 1998; Yao et al., 2013). The $\delta^{18}O$ compositions in modern precipitation samples collected at northern TP show marked seasonal patterns with the high values in summer and low values in winter (Yu et al., 2009). In addition, the major ions (e.g., Mg^{2+} and $SO4^{2-}$) also show clear seasonal cycles with high concentrations in winter/spring and low concentrations in summer (Zheng et al., 2010). They have been used in past studies as complementary tools in ice core dating in the northern TP (Kang et al., 2007). Therefore, the ZK ice core was dated by using the seasonality of $\delta^{18}O$ in conjunction with the seasonal variations of major ions, including Mg^{2+}, Ca^{2+} and $SO4^{2-}$, with a reference layer of β activity peak in 1963 (Fig. 2). The core 1 was dated back to 1951 at 16.38 m depth with uncertainty estimated within 1 year (Fig. 2, Zhang et al., 2016). Based on the dating result and density of the ice core profile, the mean annual net accumulation rate was estimated to be low for ZK glaciers (190 kg H_2O m^{-1} yr^{-1}). This study focused on the $\delta^{18}O$ records in the top 16.38 m of the ice core, corresponding to the time period 1951-2008.

2.2 Climate data

The ZK glacier is located at a transition zone with shifting influences between the westerlies and the Indian summer monsoon (Yao et al., 2013). Based on the climate records from the two nearby meteorological stations, at Gêrzê (32°09′, 84°25′, 4414.9m a.s.l., 1973-2008) and Xainza (30°57′, 88°38′, 4800m a.s.l., 1961-2008) (Fig. 1), the local mean monthly temperature ranges from -10.8°C in January to 10.7°C in July, with an annual average of 0°C. Precipitation averages 257 mm per year, of which 75% falls between June and September (Fig. S1a).

In order to establish the representativeness of the ZK ice core $\delta^{18}O$ for the regional climate, we performed correlation analysis, using Pearson's correlation coefficient (r), between the ice core $\delta^{18}O$ time series and temperature records from the nearby meteorological stations (Gêrzê and Xainza), and the instrumental temperature series from a network of 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. It should be noted that most of the stations used in ITNTP time series were located on the eastern part of the northern TP with altitudes ranging from 2767 to 3367 m (Guo and Wang, 2011), whereas this study focused on the higher (> 5700 m) and more extensive western part of the northern TP (Fig. 1). In addition, spatial correlations were carried out between ZK $\delta^{18}O$ and the CRU 4 gridded temperature reanalysis data (Mitchell and Jones, 2005) on the KNMI Climate Explorer (http://climexp.knmi.nl).

In this study, in addition to the ZK series, we also attempted to reconstruct a regional temperature series by combining ZK with other ice core $\delta^{18}O$ records in the northern TP, including Muztagata (Tian et al., 2006), Puruogangri (Yao et al., 2006), Geladaindong (Kang et al., 2007) and Malan (N. Wang et al., 2003) (Fig. 4 and Table 2). We first examined the consistency of these ice core records and excluded Malan from the reconstruction because of its drastically different temporal patterns from the rest of the records. To combine the remaining 4 ice core records, we derived the $\delta^{18}O$ anomalies for each ice core series to eliminate the difference in the absolute values, and calculated their average (Fig. S2), which was then used to reconstruct the regional temperature time series.

3 Results and Discussion

3.1 The ZK ice core $\delta^{18}O$ variation and its relationship with regional meteorological data

The raw $\delta^{18}O$ values throughout the ZK ice core profile from 1951 to 2008 were presented in Figure 2. For this section, the $\delta^{18}O$ values 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 $\delta^{18}O$ values were relatively low in the 1960s, followed by an increasing trend from 1970s to the end of the record.

Stable oxygen isotope in precipitation could be affected by a variety of environmental factors. In addition to temperature, the $\delta^{18}O$ values in ice cores could also be affected by precipitation seasonality and amount (Dansgaard, 1964). To exclude possible influence of precipitation, we first examined whether the seasonal distribution of precipitation experienced any significant changes during the study period by using the precipitation records from the two nearby stations. Results showed weak positive trends for the proportion of precipitation in winter and spring, and no statistically significant trends for the proportions of precipitation in summer and fall (Fig. S1b and c). This suggests that changes in seasonal distribution of precipitation did not exert a major influence on the $\delta^{18}O$ values in ZK ice cores during the period 1961-2008. Besides, we found no significant correlation between the ZK $\delta^{18}O$ record and precipitation amount recorded at the stations (Table S1). Partial correlation analysis showed this to be true even when annual temperature was controlled ($r_{partial} = 0.01$, $p > 0.1$). This suggests that precipitation amount had little influence on the ZK $\delta^{18}O$ values.

On the other hand, the ZK $\delta^{18}O$ time series showed positive correlation with annual temperature measured at each of the nearby stations ($r = 0.31$, $p = 0.07$ for the Gêrzê station; $r = 0.43$, $p = 0.002$ for the Xainza station), the mean 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). 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} (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).

Significant spatial correlation existed between the ZK $\delta^{18}O$ series and the CRU gridded temperature data 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 were consistent with previous results between the ZK $\delta^{18}O$ series and station temperature records (Table 1). The stronger spring temperature signal recorded in ZK $\delta^{18}O$ record may be attributed to the different seasonal moisture sources in this region. At Shiquanhe and Gêrzê, Yu et al. (2009) found that during the non-monsoon period (October–June) when local moisture recycling and the westerlies dominate the moisture sources, air temperature correlates more strongly with $\delta^{18}O$ in precipitation. On the other hand, precipitation $\delta^{18}O$ in monsoon season could be affected by a variety of factors other than temperature, including the convection intensity, distance from moisture sources and amount effect (Y. He et al., 2015; Tang et al., 2015). This could obscure the relationship between $\delta^{18}O$ and air temperatures (Joswiak et al., 2013). In addition, previous studies in the central Himalayas found that high elevation areas (> 3000ma.s.l.) can receive up to 40% of their annual precipitation during cold season because of terrain locked low pressure systems and orographically forced precipitation (Lang and Barros, 2004), a much higher percentage than that of surrounding low altitude areas of the same region (Pang et al., 2014). Therefore, the ZK ice core (located at 6226 m a.s.l.) could have had more cold-season (non-monsoonal) precipitation than that indicated by nearby meteorological stations, located at much lower elevations. Both factors could result in a stronger signal of spring temperature in the ZK ice core $\delta^{18}O$ record.

3.2 Regional temperature reconstruction

Detailed comparisons were made between the ZK $\delta^{18}O$ and the $\delta^{18}O$ time series of four nearby ice cores, including Muztagata, Puruogangri, Geladaindong and Malan (Fig. 4 and Table 2). The cooling around 1960s was present in all ice cores, and this was consistent with the observed cold period during this time over the entire TP (Liu and Chen, 2000). Moreover, the significant increasing trend from 1970s to present was observed in all except Malan ice core $\delta^{18}O$ series. We calculated the Pearson correlation coefficients among these ice core $\delta^{18}O$ series (Table 3). The results showed weak correlations between the annual values of these series. This lack of correlation could result from the differences in location, elevation and hence local climates. It could also arise from uncertainties in ice core dating. In order to reduce the impact of dating uncertainties, we used the 5 year running averages instead of annual values, and these series showed much stronger correlations, suggesting possible common regional climate patterns preserved in these ice core series. This coherence is important when we use the average of multiple sites to develop a regional composite.

In contrast to the rest of the ice cores, the Malan $\delta^{18}O$ record showed a cooling trend since 1970s (Fig. 4e). Such continuous low level of $\delta^{18}O$ could be caused by the change of local climate conditions (Y. Q. Wang et al., 2003), but could also result from post-depositional processes on the chemical profiles, such as summer melting, evaporation and condensation, all of which could modify the relationship between ice core $\delta^{18}O$ 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.

Moreover, the correlations between Geladaindong and three other ice cores, i.e. ZK, Muztagata, Puruogangri were relatively low even after 5 year running averages (Table 3). The lack of correlation could be attributed to its local climate conditions (Table 3), such as the influence of local convective vapor due to its more northern location (Kang et al., 2007). However, the ice cores of ZK, Muztagata, Puruogangri and Geladaindong shared similar patterns of $\delta^{18}O$ variations, especially their increasing trends since 1970s (Fig. 4). Moreover, regional composite with Geladaindong records correlates very strongly with that without Geladaindong (r = 0.95, 1951-2002, $p < 0.0001$), and two series showed very similar temporal patterns (Fig. S2). Therefore, we decided to include the Geladaindong ice core $\delta^{18}O$, so that the final regional reconstruction could have larger spatial coverage to better represent the regional climate of the northern TP. The regional temperature series was reconstructed for 1951-2002, the common period covered by the four ice core $\delta^{18}O$ records. Meanwhile, a temperature reconstruction based solely on ZK ice core $\delta^{18}O$ record was constructed for 1951-2008 to investigate the temperature variations since the late 1990s.

Before establishing the temperature reconstructions, it was necessary to derive the $\delta^{18}O$-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 $\delta^{18}O$ 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 average isotope sensitivity of 0.53‰ °C^{-1} for the present-day 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} to convert the $\delta^{18}O$ values to temperature for the Muztagata ice core. The isotope sensitivity usually increases with elevation as indicated by Rayleigh-type equilibrium fractionation model (Rowley et al., 2001). Kang et al. (2007) obtained 1.40‰ °C^{-1} $\delta^{18}O$-temperature relationship from the linear regression between the 5 year running average of Geladaindong $\delta^{18}O$ records and regional instrumental temperature records. In our study, the strongest correlation was found between the 5 year running average of the regional $\delta^{18}O$ record and ITNTP ($r = 0.89$, $p < 0.001$) (Fig. S3). The ZK $\delta^{18}O$ correlates most strongly with the 5 year running average of the mean temperature from two nearby stations (Gêrzê and Xainza, $r = 0.60$, $p < 0.001$) (Table 1). Based on these significant relationships, the isotope sensitivities were determined as 1.46‰ °C^{-1} for the regional $\delta^{18}O$ series and 1.18‰ °C^{-1} for ZK $\delta^{18}O$ series, and were used to reconstruct regional temperature series for the northern TP (RTNTP) and the ZK temperature series respectively. Additional analysis showed that as isotope sensitivity value increases, the response of decadal warming rate decreases, especially for the isotope sensitivity values greater than 1.0 (Fig. S4).

The reconstructed regional temperature for the northern TP (RTNTP) was presented in Figure 5a together with the temperature reconstruction for the ZK ice core (Fig. 5b), ITNTP (Fig. 5c) and the global temperature series (Fig. 5d) for comparison. We first compared the RTNTP with the ITNTP, and found strong correlation between the two temperature series (r = 0.65, p < 0.001). Spatially, significant correlations also existed between the CRU gridded surface temperatures and the ITNTP (r = 0.50 to 0.60, n = 42, p < 0.01), as well as between CRU and the RTNTP (r = 0.40 to 0.60, n = 52, p < 0.01) over a large region (Fig. 6). The study area had the strongest correlations (r > 0.50, p < 0.01). This suggested that the regional reconstruction adequately captured temperature variation on the northern TP.

3.3 Recent rapid warming trend over the northern TP

The regional reconstruction was compared with the global annual temperature series (Fig. 5d) and the ITNTP (Fig. 5c) in order to investigate the recent warming trend since 1970s. LOESS regression was used to smooth the data and estimate the general trend. The reconstruction captured the cooling period during 1960s, as well as the prominent warming since the 1970s to the end of the record, with the highest rate of increase in the late 1990s (Fig. 5). For the period from 1970 to 2002, the RTNTP showed more rapid warming trend at the rate of 0.51±0.07°C$(10yr)^{-1}$ than that of the global temperature (0.27±0.03°C$(10yr)^{-1}$). The RTNTP rate was also higher than the ITNTP rate of increase at 0.43±0.08°C(10yr)$^{-1}$ for the same time period. From 1990 to 2002, the warming accelerated on the northern TP with rates of temperature increase at 0.95±0.21°C(10yr)$^{-1}$ for the RTNTP and 0.90±0.29°C(10yr)$^{-1}$ for the ITNTP, much higher than the warming rate of the global temperature (0.37±0.13°C(10yr)$^{-1}$). These results seemed to indicate enhanced warming at the high elevation regions on the northern TP.

Since the late 1990s, the global temperature showed very little change and even decreasing trend since 2005 (Fig. 5d). The relatively flat warming trend was also recorded in the ITNTP (Fig. 5b). However, the ZK series revealed a continued warming trend in recent years after a brief pause during the early 2000s (Fig. 5b). We calculated mean decadal annual temperature change based on the LOESS regression model for all three time series (Fig. 7). For both the global temperature and ITNTP series, the highest average warming rates occurred during 1990s, and then decreased significantly since 1999 (Fig. 7c and d). The reduction of warming rate in the ITNTP series was consistent with results by Duan and Xiao (2015), who found weaker warming trend during the period 1998-2013 in the northern TP based on the instrumental temperature records. However, the rates of increase remained high for the temperature records in the ZK series since 1999 (Fig. 7b), in contrast to the slowdown of climate warming observed for the global mean and ITNTP temperature records since 1999 (Fig. 7d). The persistent high warming rates derived from our regional reconstructions seem to suggest that the elevation-dependent warming is still evident over the high elevations of the northern TP despite the reduced warming rates observed at lower stations in ITNTP (Fig. S5).

The persistent rapid warming in the northern TP could have been caused by the regional radiative and energy budget changes (K. Yang et al., 2014; Yan and Liu, 2014; Duan and Xiao, 2015). Many studies show that the snow/ice-albedo feedback is an important mechanism for enhanced warming at high elevation regions (Liu and Chen, 2000; Pepin and Lundquist, 2008; Rangwala and Miller, 2012). Ghatak et al. (2014) found that the surface albedo decreases more at higher elevations than lower elevations over the TP in recent years. Qu et al. (2013) observed a decreasing trend for the snow/ice albedo at the Nyainquentanglha glacier region, central TP, for the period 2000 to 2010. It has been found that the glacier albedo for the nine glaciers in western China has decreased during the period 2000-2011, especially for the central TP (J. Wang et al., 2014). For example, the glacial albedo of Dongkemadi and Puruogangri glaciers decreased at a rate of 0.0043-0.0059 yr^{-1} and 0.001-0.004 yr^{-1} respectively. Reduced surface albedo increases the surface absorption of solar radiation, and may have contributed to the continued warming over the high elevation regions of the northern TP. Further research is needed to identify and quantify the exact mechanisms accounting for the temperature variations over the Plateau.

4 Conclusions

This study presented a $\delta^{18}O$ time series of the ZK ice core from the northern TP, based on which a temperature record was reconstructed for the period 1951-2008. Moreover, by combining the ZK $\delta^{18}O$ with three other ice cores from the northern TP, a regional temperature history was established from 1951 to 2002. These temperature reconstructions captured the rapid warming trend since 1970, and showed continued warming since 1999 at much higher rates than those of the global average temperature and the instrumental temperature records for the northern TP.

Possible explanations for this continued warming might lie in the regional radiative and energy changes at higher elevations over the northern TP. However, the exact physical mechanisms responsible for the consistently significant warming at higher elevations remain unclear, partly due to the scarcity of available observations. Further studies are needed to understand the specific characteristics of this warming trend on the TP, as well as the response mechanisms of high elevations regions to global changes.

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Figures

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), Geladaindong (Kang et al., 2007) and Malan (Wang et al., 2003) over the northern TP. The inset shows the relative location of the northern TP to the entire TP. The black rectangle indicates the study area. Red and blue arrows represent the circulation patterns for the study region. Red arrows indicate the direction of the Indian monsoon (near surface) in summer, and blue arrows indicate the dominant westerlies (mid to upper troposphere) in winter.

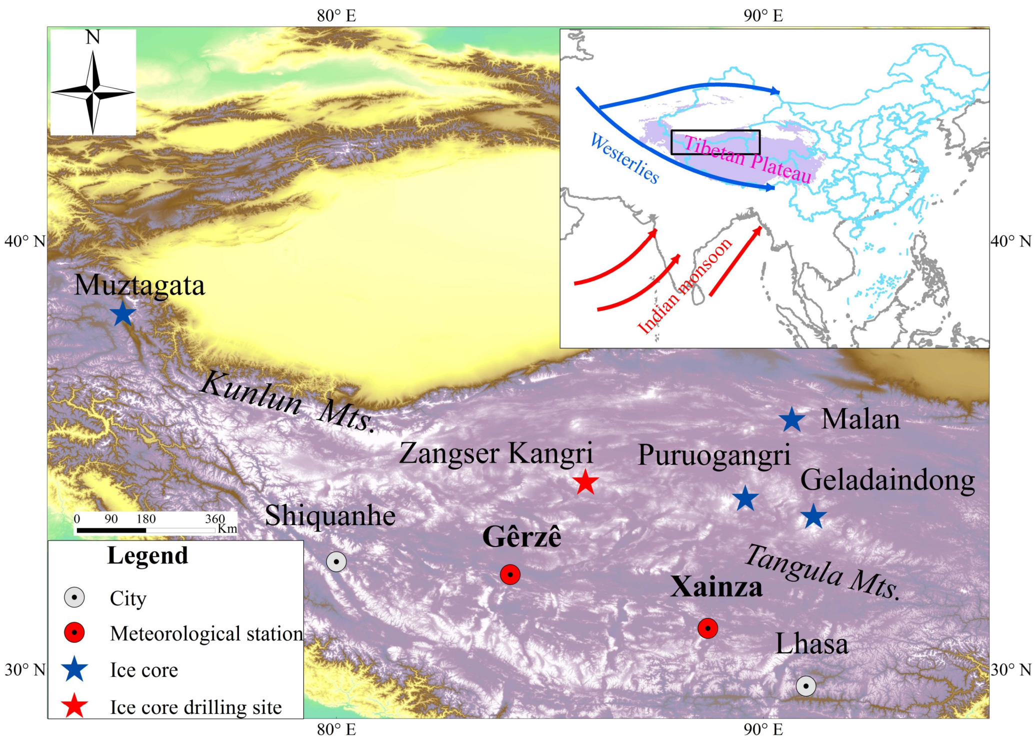

Figure 2. Variations of $\delta^{18}O$ in the ZK ice core and other data used for dating, including beta activity and major ion concentrations. We calculated the logarithm to the base 10 for the concentrations of the Ca^{2+} and Mg^{2+} to facilitate dating.

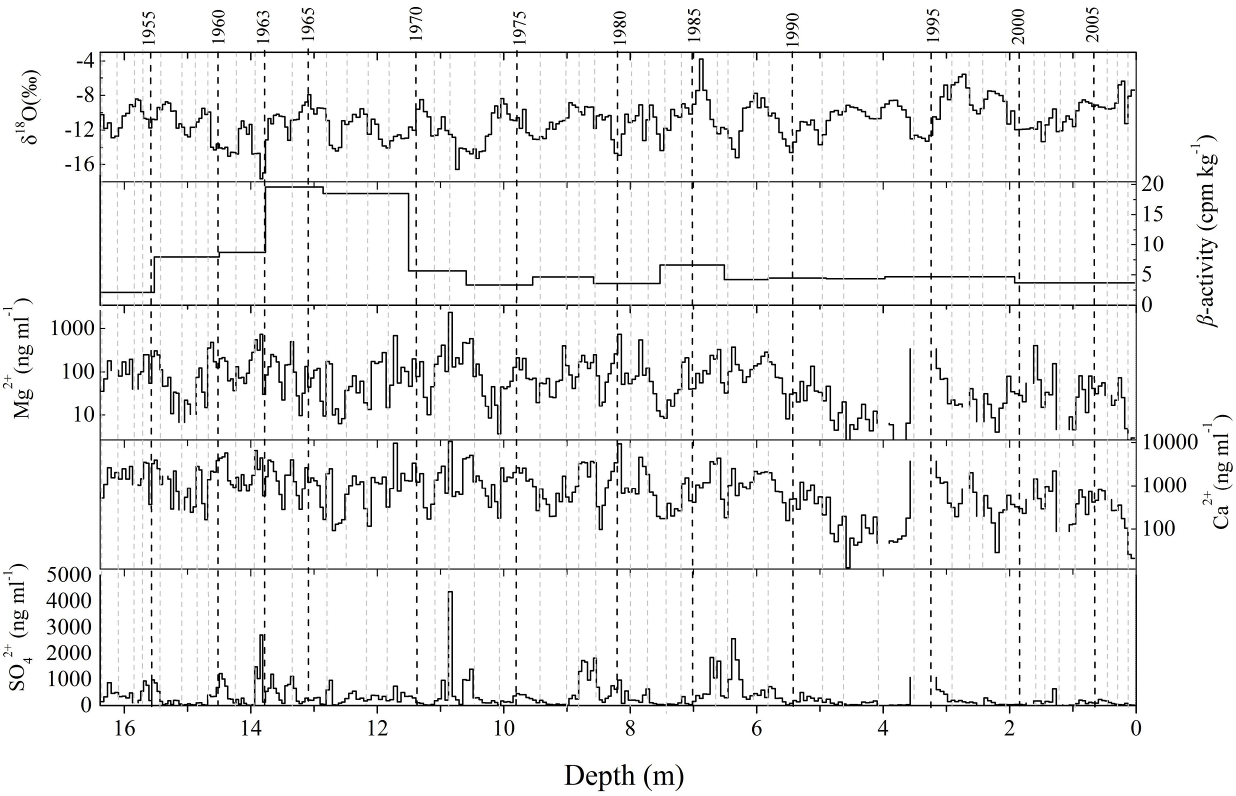

Figure 3. Spatial correlations of ZK ice core $\delta^{18}O$ record with CRU-gridded (Mitchell and Jones, 2005) annual mean temperature (a), annual minimum temperature (b), spring mean temperature (c), and spring minimum temperature (d) for the period 1951-2008. Only correlation coefficients significant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

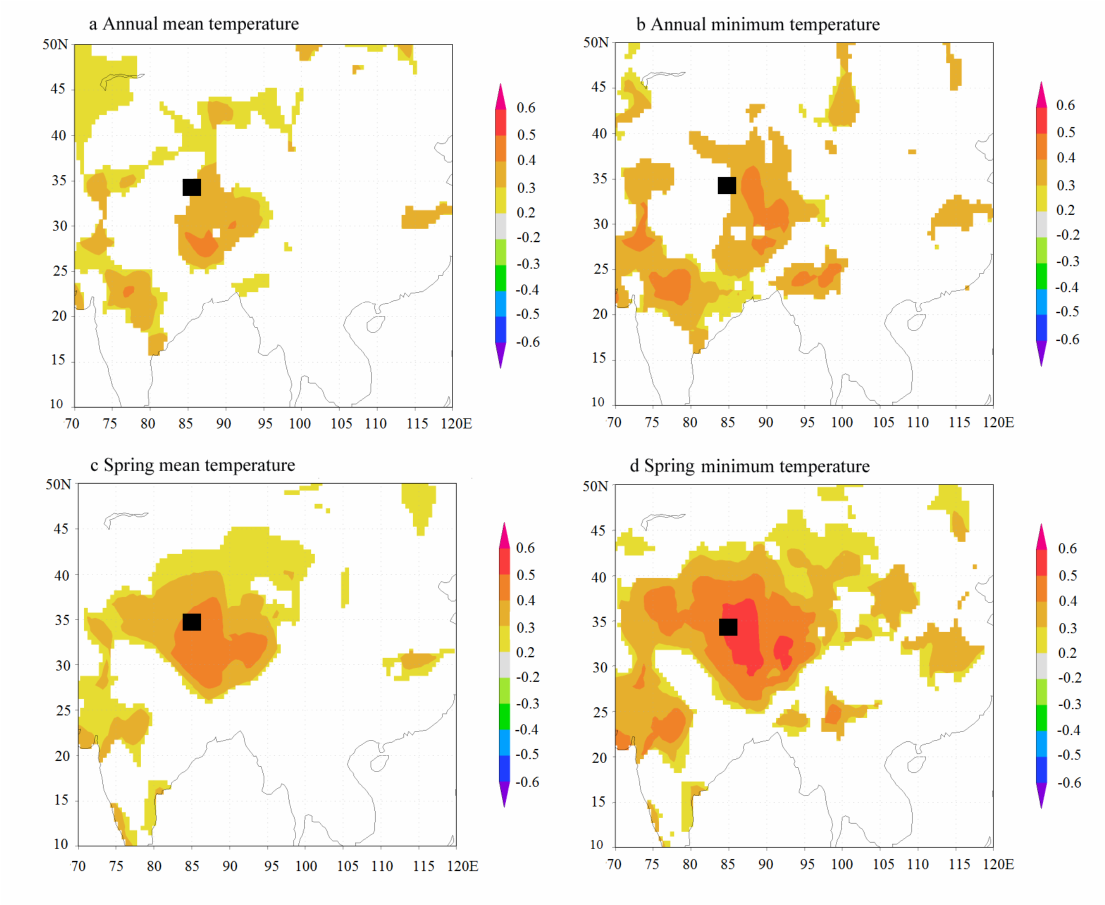

Figure 4. Comparisons of the anomalies of $\delta^{18}O$ records in the ZK ice core (a) with $\delta^{18}O$ records from Muztagata (b), Puruogangri (c), Geladaindong (d) and Malan ice cores (e). Thin lines represent annual values, thick lines the 5-year running averages, and the dotted lines the linear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

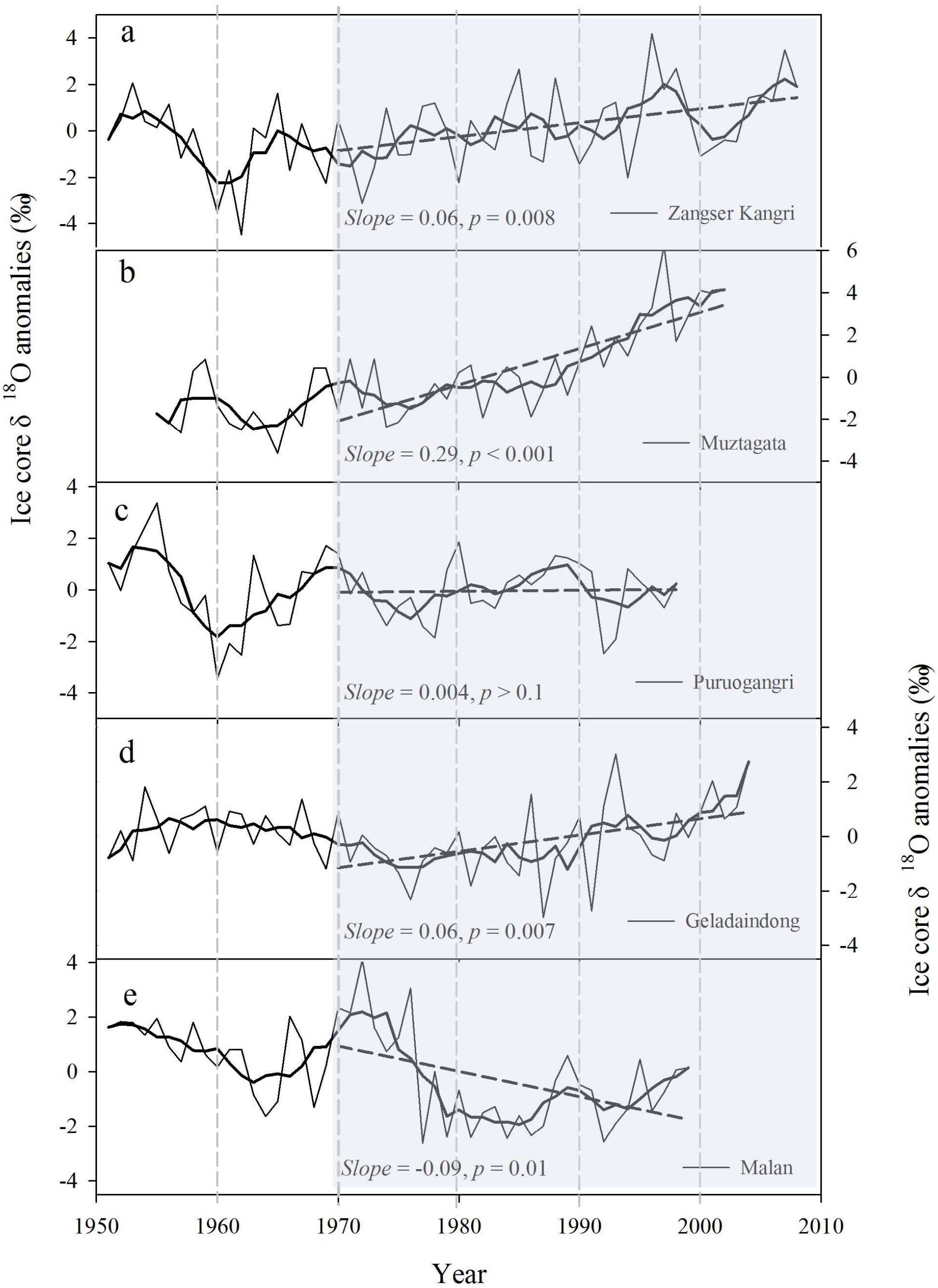

Figure 5. The reconstructed regional temperature series for northern Tibetan Plateau (RTNTP) from ZK, Muztagata, Puruogangri and Geladaindong ice core $\delta^{18}O$ records (a), the reconstructed temperature series from ZK ice core $\delta^{18}O$ record (b), the instrumental temperature record for the northern TP (ITNTP) (c), and global average temperature (d). Black trend lines were estimated using the non-parametric LOESS regression technique with a span of 0.4; the dots indicate the raw values of corresponding temperature series; shading represents the 95% confidence intervals of the estimated trends.

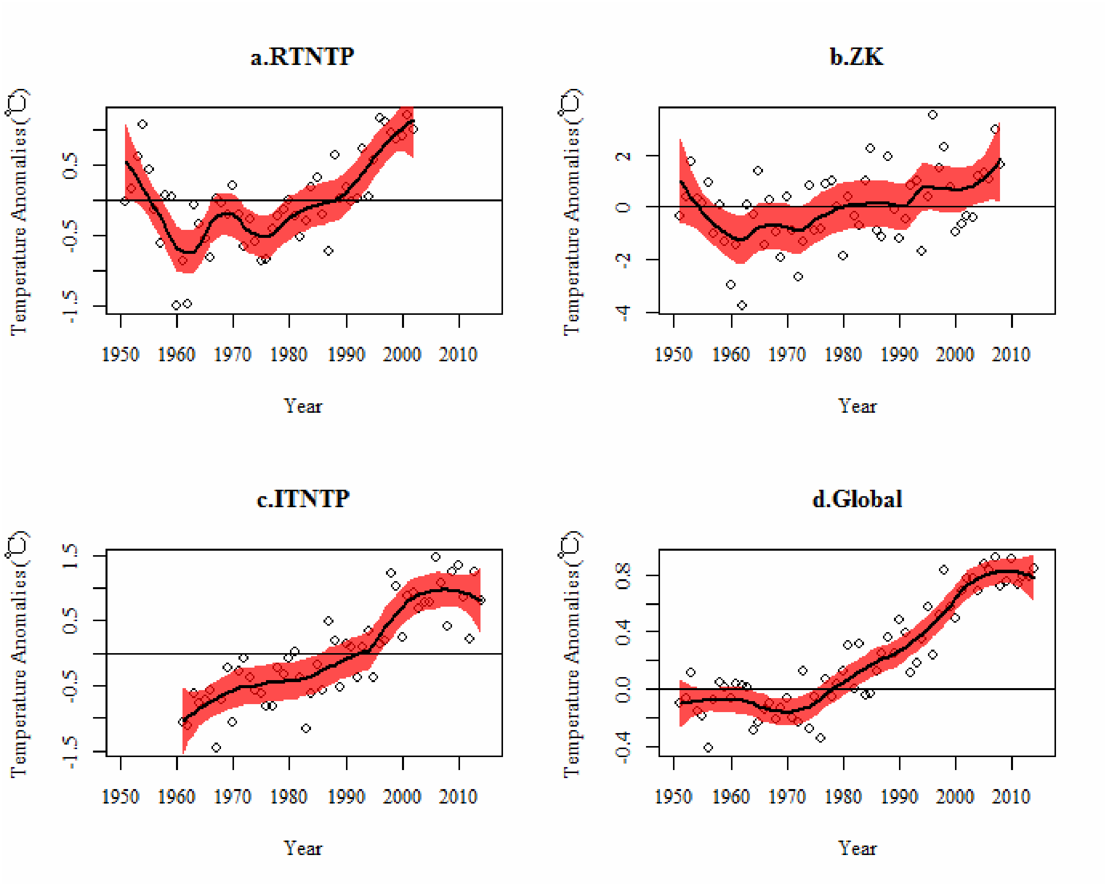

Figure 6. Spatial correlations (r values in color, $p < 0.01$) between the gridded annual mean temperature data (the CRU 4 temperature time series, 0.5° × 0.5° resolution, Mitchell and Jones, 2005) and the instrumental temperature record of the northern TP (ITNTP) (Guo and Wang, 2011) for the period 1961-2002 (a), and the regional temperature reconstruction series for the period 1961-2002 (b). The black rectangle indicates the study area and the blue rectangle indicates the region covered by ITNTP.

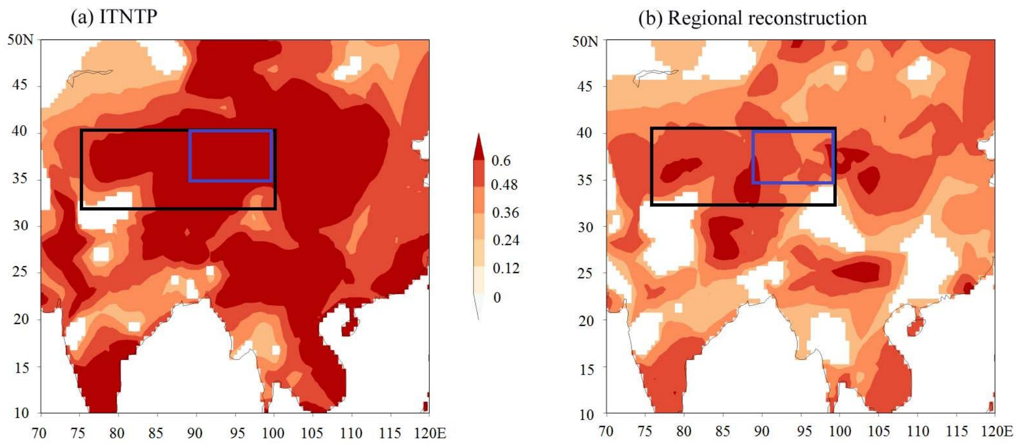

Figure 7. Decadal mean annual change rates for the regional temperature reconstruction series for northern TP (RTNTP) (a), the temperature reconstruction from ZK ice core $\delta^{18}O$ record (ZK) (b), the instrumental temperature record of the northern TP (ITNTP) (c), and global average temperature (d). The decadal mean annual change rates were estimated using the non-parametric LOESS regression model with a span of 0.4.

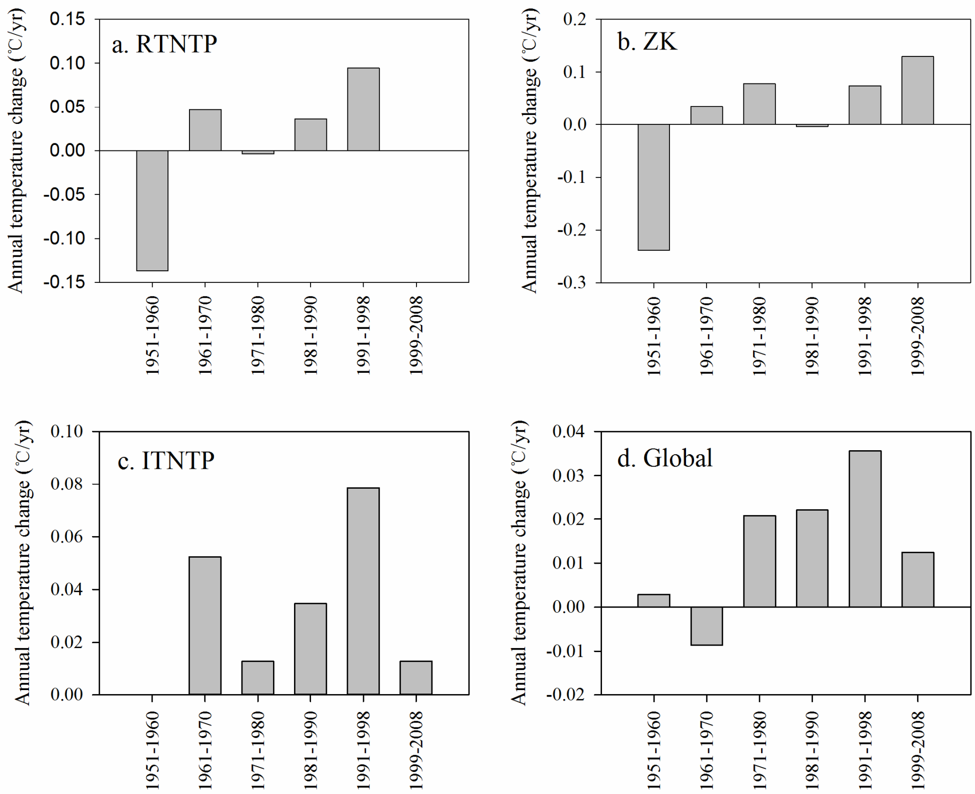

Table 1. Correlation coefficients and linear slopes between the $\delta^{18}O$ values in the ZK ice core and instrumental spring (March–May) and annual temperature from closest Gêrzê (1973–2008) and Xainza stations (1961–2008), the averaged records of the two stations (1961–2008), and the ITNTP series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

		Gêrzê		Xainza		Stations averaging		ITNTP
		March-May	Annual	March-May	Annual	March-May	Annual	Annual
Correlation coefficients	Annual	0.52^{c}	0.34^{a}	0.45^{c}	0.34^{a}	0.48^{c}	0.34^{a}	0.35^{a}
	5 year running average	0.63^{c}	0.53^{c}	0.73^{c}	0.60^{c}	0.73^{c}	0.60^{c}	0.61^{c}
Slope	Annual	0.93^{b}	0.67^{a}	0.93^{b}	0.98^{a}	1.00^{c}	0.88^{a}	0.87^{a}
	5 year running average	0.87^{c}	0.76^{c}	1.54^{c}	1.32^{c}	1.37^{c}	1.18^{c}	0.40^{c}

^{a} $p< 0.05$; ^{b} $p< 0.01$; ^{c} $p< 0.001$.

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′05.8″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

Table 3. Correlation coefficients between the $\delta^{18}O$ 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 (1951-2002) 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**	**0.57**[c]
Muztagata	*0.68*[c]		**0.09**	**0.04**	**-0.20**	**0.80**[c]
Puruogangri	*0.46*[c]	*0.28*		**-0.08**	**0.17**	**0.53**[c]
Geladaindong	*-0.07*	*0.24*	*-0.12*		**0.05**	**0.30**[a]
Malan	*-0.40*[b]	*-0.33*	*0.14*	*0.18*		
Regional average	*0.79*[c]	*0.95*[c]	*0.54*[c]	*0.31*[a]		

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