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ORIGINAL PAPER

Combined tree-ring width and $\delta^{13}C$ to reconstruct snowpack depth: a pilot study in the Gongga Mountain, west China

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Abstract Tree-ring width (TRW) and stable carbon isotope ($\delta^{13}C$) in tree-ring cellulose of subalpine fir (*Abies fabri*) were used to develop high-resolution climate proxy data to indicate snow-depth variations in the Gongga Mountain, west China. Tree radial growth- and $\delta^{13}C$-climate response analyses demonstrated that the TRW and $\delta^{13}C$ at the timberline (3,400 m.a.s.l.) are mainly influenced by temperature and precipitation of previous growth seasons and current summer (June to August) under cold and humid conditions. Considering the analogous control factors on both tree growth and carbon isotope discrimination ($\Delta^{13}C$) and snow accumulation, the negative and significant relationships between tree-ring parameters (TRW and $\Delta^{13}C$) and mean monthly snowpack depth were found. Herein, by combining two tree-ring parameters, a primary snow-depth reconstruction (previous October to current May) over the reliable period A.D. 1880–2004 was estimated. The reconstruction explains 58.0% of the variance in the instrumental record, and in particular captures the longer-term fluctuations successfully. Except the period with extreme higher snowpack depth around 1990, the snowpack depth seems to fluctuate in a normal way. The reconstruction agrees with the nearby snowpack depth record in Kangding and the mean observed snowpack-depth variations of the stations on the Tibetan Plateau, particularly at long-term scales. The snowpack depth in low-frequency fluctuations, during the past century, agrees quite well with the Eastern India precipitation covering the period of previous October–current May. Our results suggest that combing tree-ring width and $\delta^{13}C$ in certain subalpine tree species growing on the Tibetan Plateau may be an effective way for reconstructing regional snowpack variations.

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1 Introduction

As a sensitive indicator of climate change, snow is a vital water resource in western China. Previous studies have suggested that an increase in the spring snow cover days on the Tibetan Plateau (TP) is closely associated with the variation in the East Asian summer monsoon (Zhao et al. 2007). Using longer time series, Roboch et al. (2003) documented that strong India summer monsoon precipitation is actually preceded by higher-than-normal TP snow cover in winter and spring. Thus, it is essential to understand snow cover days or snowpack depth characteristics including the natural variability of seasonal and long-term range of extremes. However, such a perspective is still limited by a sparse network of snow measurement sites from the mountains of west China where instrumental records for the past five decades (Qin et al. 2006).

Although instrumental climate records are limited in length, proxy climate data, such as tree rings, have been useful in extending these records. To date, different tree-ring parameters have successfully been used to reconstruct variability of past temperature (Esper et al. 2002; Liu et al. 2005, 2007; Briffa et al. 2007; Büntgen et al. 2008), precipitation (Gray et al. 2004; Shao et al. 2005; Treydte et al. 2006), drought (Cook et al. 1999; Li et al. 2007; Liu et al. 2008), etc. deduced from width, density, stable carbon isotope ($\delta^{13}C$), and stable oxygen isotope ($\delta^{18}O$) (McCarroll and Loader 2004, and the reference therein). To date, only few studies have used tree rings to investigate the response of tree growth to snowpack (Perkins and Swetnam 1996; Peterson and Peterson 2001) and the long-term snowpack variability (Woodhouse 2003). Perhaps more powerful is the potential to produce reconstructions which contain information on both temperature and precipitation variability by combining parameters which response to diverse climate factors in different seasons. Multi-parameters dendroclimatology may give access to bi-variables signal at the same site (Gagen et al. 2006), which would potentially allow reconstructions of complex variations in regional synoptic climate from a single tree population.

The Gongga Mountain, a leading peak in the snow-covered mountains of the western China with an altitude of 7,556 m.a.s.l., is the highest mountain in Sichuan Province. The Hailuogou Valley, lying at the foot of the Gongga Mountain is well-known for the coexistence of lower-altitude modern glaciers and the flourishing forest at the same elevation. In high altitudes, particularly alpine timberlines, trees growth is considered to be highly sensitive to climatic or environmental changes (Smith et al. 2003; Oberhuber 2004; Liu et al. 2005; Shao et al. 2005; Liang et al. 2009). In the Gongga Mountain, Wu and Shao (1995) reported a March–July mean minimum temperature reconstruction by tree-ring width since AD 1800, suggesting some relationship between tree growth and winter environmental conditions. This gives us an idea to explore the potential to reconstruct variations of cryospheric elements (e.g., snowpack depth and glacier mass balance) based on timberline tree-ring parameters in the Gongga Mountain

We present a new tree-ring width and a $\delta^{13}C$ chronologies of *Abies fabri* at the alpine timberline in the Hailuogou valley, Gongga Mountain. The objectives of this study are: (1) to detect the responses of tree growth rates and carbon isotope discrimination ($\Delta^{13}C$) of timberline fir to climate variation in the Gongga Mountain; (2) to estimate the possible linkage between tree-ring parameters and snowpack depth; (3) to reconstruct snowpack depth variations and discuss the possible controlling factors on long-term fluctuations.

2 Data and methods

2.1 Climate of the study area

Located in the south-eastern fringe of the Tibetan Plateau (TP), Mountain Gongga is the highest peak in the eastern part of the TP and Hengduan Mountain region. It is a bordering mountain in the transitional zone between the dry TP and humid Sichuan basin. Its climate is temperate and is characterized by the southwest (India and Bengal monsoons) and southeast monsoon in the wet season and by the westerly and the Qinghai–Tibet monsoon in the dry season.

Meteorological data, including temperature, precipitation and snow depth for 1988–2004 were provided by the Ecological System and Environment Research Station (29°36′N, 101°53′E, 3000 m elevation) (GGS station), Chengdu Institute of Mountain Hazards and Environment, Chinese Academy of Sciences. The climate is characterized by high rainfall, a long winter period of rain and snow (about 9 months; Cheng et al. 2004). The annual mean temperature is 3.8°C. The total annual precipitation is about 2,000 mm (Cheng et al. 2004), with a maximum for June to September accounting for 44.0% of total, and 771.0 mm for the period of previous October to current May which may fall as snow at the high altitude region, resulting in a long period of snow-cover. In addition, the nearby meteorological data from the Luding station (LD) (29°53′N, 102°18′E, 1321 m.a.s.l., period of record: 1960–2004) and the Kangding station (KD; 101°58′ N, 30°03′E, 2615 m.a.s.l., period of record: 1952–2004) were used.

2.2 Sampling and chronology development

A. fabri Craib is a dominant subalpine vegetation between 2,900–3,600 m in the Gongga Mountains. It has ability to growth in snowy environment. Due to the shallow-root distribution of trees, its growth may be influenced by the soil environment variations at the upper soil profile. The trees were chosen in open stands near the upper timerline (29°33.2′N, 101°58.2′E, 3,520 m.a.s.l.; Fig. 1). Totally, 50 cores from 20 trees were sampled with an increment borer (5 mm) at breast height. The third cores were obtained from the dominant trees for isotopic analysis. All the sampled trees grow at the same elevation and topographical aspect. The tree-ring samples were processed following standard dendrochronological practices (Cook and Kairiukstis 1990). After a rigorous cross-dating of the tree-ring cores, ring widths were measured with a resolution of 0.01 mm, and then the quality of the cross-dating was checked using the COFECHA program (Holmes 1983). The tree-ring measurements were standardized to remove the biological growth trend as well as other low-frequency variations due to stand dynamics. In order to keep most of climate-

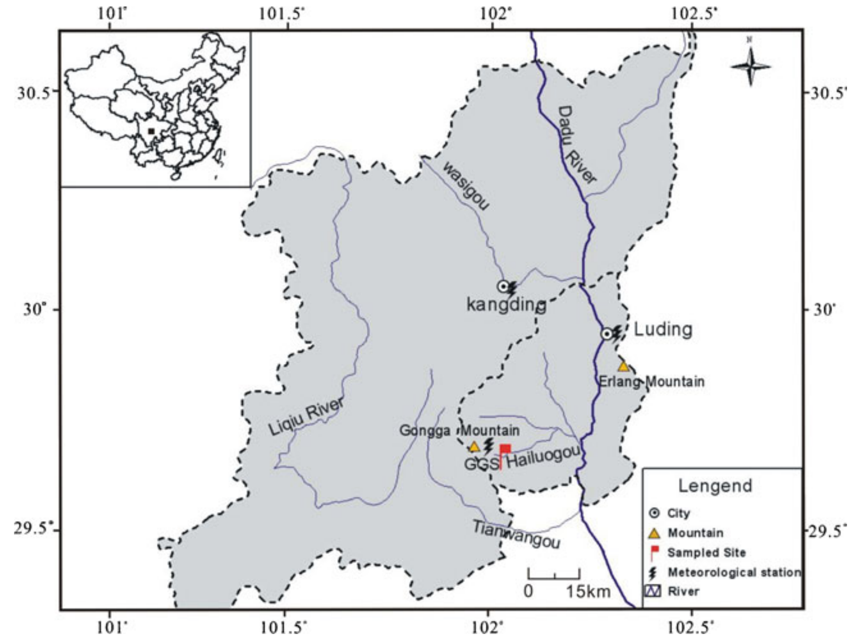

Fig. 1 Map showing the location of tree-ring sample site and meteorological stations

related variance in the data, we used conservative detrending methods by fitting a negative exponential curve to individual records to remove long-term growth trends thought to be induced by non-climatic influences. The detrending and chronology building was done with ARSTAN (Cook and Holmes 1986). The chronologies statistics is listed in Table 1.

2.3 Isotope analysis

Nine trees without obvious damage signs were selected for isotope analysis. The rings were cut into sub-samples with 1-year resolution under a binocular microscope and rings of different trees growing in the same year were pooled. The α-cellulose was extracted in order to avoid different isotopic signatures based purely on changes in the relative abundance of the individual constituents of the wood, having different isotope signatures (Loader et al. 2003). We adapted the delignification procedure described by Green (1963) to facilitate batch processing of the samples (Loader et al. 2003). The cellulose was weighed and combusted to CO_2 for stable carbon isotope analysis with an Isoprime mass spectrometer (Isoprime, Micromass Ltd., Manchester, UK) linked to an element analyzer (EA 3000, Euro Vector, Milan, Italy) in the Key Laboratory of Ecohydrology and Integrated River Basin Science, Cold and Arid Regions Environmental and Engineering Research Institute (Liu et al. 2008). The standard deviation of measurement during combustion and analysis was <0.05‰. The results were expressed as δ-values relative to the Vienna Peedee Belemnite (V-PDB) standard:

$$\delta^{13}C = \left\{R_{(\text{sample})}/R_{(\text{standard})} - 1\right\} \times 10^3 \text{per mil}, \quad \text{where } R = {}^{13}C/{}^{12}C \tag{1}$$

Table 1 Statistical character of the standard (STD) width chronology and $\delta^{13}C$ series

	PC1	SNR	EPS	STD	MS	AC1	$\delta^{13}C$	$\Delta^{13}C$
STD	0.318	4.90	0.831	0.122	0.108	0.355		
Mean							−23.3	16.1
Standard Error							0.0	0.0
Standard deviation							0.5	0.5
Variance							0.3	0.3
Kurtosis							0.4	1.9
Range							3.0	3.4
Minimum							−24.8	14.6
Maximal							−21.9	18.0
Numbers							206.0	206.0

In the table, PC1 refers to the first component, SNR refers to ratio of signal and noise, EPS refers to the expressed population signal, STD refers to standard deviation, MS refers to mean sensitivity, and AC1 refers to first-order autocorrelation

The carbon isotope discrimination ($\Delta^{13}C$, CID) in plant leaves associated with the carbon fixation by C_3 plants was expressed by Farquhar et al. (1989) as follows:

$$\Delta^{13}\mathrm{C} = \frac{\delta^{13}\mathrm{C}_{\mathrm{air}} - \delta^{13}\mathrm{C}_{\mathrm{plant}}}{1 - \frac{\delta^{13}\mathrm{C}_{\mathrm{plant}}}{1{,}000}} \quad (2)$$

where $\Delta^{13}C$ is the carbon isotope discrimination by the plant. $\delta^{13}C_{air}$ and $\delta^{13}C_{plant}$ are the $\delta^{13}C$ values of ambient air and plant cellulose, respectively. Annual resolution of atmospheric $\delta^{13}C$ and atmospheric CO_2 concentration estimated by McCarroll and Loader (2004) were used in Eq. 2.

2.4 Data analysis

To detect the influence of climate conditions on tree-ring growth (TRW) and carbon isotope discrimination (CID), climate–tree ring relationships were investigated using simple linear correlation analysis with monthly precipitation and maximum, minimum, and mean temperature for the period from previous June to current September, because the growth of many tree species is affected not only by the climatic conditions of the current year but also by those of previous year (Liu et al. 2005; Takahashi et al. 2005).

Using TRW and $\delta^{13}C$ as predictor, snowpack depth variations were reconstructed back to AD 1890 (EPS>0.85 and cores for isotopic analysis >8) by a linear regression model. The one-leave-out cross-validation method was employed to verify our reconstruction, since the meteorological data set too short to carry out a robust split-sample calibration (Michaelsen 1987). Evaluative statistics include the Person's correlation coefficient (R), reduction of error (RE), sign test (ST) (Fritts 1976). The low-frequency comparison between reconstructed snowpack depth and India precipitation was conducted by fitting the five-order polynomial trends lines.

3 Results

3.1 Regional climate variability

Annual mean temperature of the GGS, LD, and KD show a significant warming trend of 0.28°C, 0.22°C, and 0.23°C per decade during the 1975–2004 periods, respectively (Fig. 2a). Before AD 1975, the slight decreasing trends in yearly temperature of LD and KD were found with the rate of −0.13°C and −0.03°C per decade. The observed precipitation in three meteorological stations all showed the increasing trends of 25.6, 11.1, and 45.6 mm per decade covering the period of 1975–2004, respectively (Fig. 2b). Before AD 1975, the precipitation in LD and KD showed decreasing trends of −128.1 and −75.5 mm per decade, respectively. As a whole, the temperature trends of three stations indicated better coherence than that of precipitation.

The snowpack depth recorded in GGS and KD stations showed a similar trends during their common periods (r=0.64, P=0.005) though with different absolute values caused by different altitude (not shown). However, the high-frequency variations between two series were not concurrent especially in the detailed fluctuations. For example, the higher snowpack depth in 1991 at GGS station was not found in the records at LD station; the relative higher variability is more pronounced in the record of GGS station than in that of KD station.

3.2 Chronology statistics

The subalpine ring width chronology of fir ranged from 1774 to 2004 (Fig. 3). Based on expressed population signal (EPS) statistics, the width chronology met signal strength acceptance after AD 1880. The standard (STD) width chronology showed a low mean sensitivity (Table 1), which is typical for trees in humid monsoonal environments (Fan et al. 2008; Liang et al. 2009). The first principal accounts for 31.8% variations of total in STD chronology, respectively.

The $\delta^{13}C$ series covers the period from 1800–2004 (Fig. 4a). The mean $\delta^{13}C$ value is −23.3‰ with a standard deviation of 0.5‰ (Table 1). The range of tree-ring $\delta^{13}C$ is about 3.0 ‰ from 1800–2004. From AD 1800 to 1920, a slight increasing trend in $\delta^{13}C$ was found, which may be related to the "juvenile effects" even though the first 10 years of each cores were not mixed into the corresponding samples. After AD 1920, the tree-ring $\delta^{13}C$ became more negative gradually due to the increasing atmospheric CO_2 concentration. The $\Delta^{13}C$ series after the correction removing the effects of increasing atmospheric CO_2 concentration, exhibited a stable decreasing trends covering the past 200 years (Fig. 4b).

3.3 Growth/carbon isotope-climate relationships

Temperatures in the autumn season (previous September to November (LD station), previous September to current January and current April to June (GGS station)) have predominant and positive influence on the radial growth (Table 2). The $\Delta^{13}C$ in all trees holds the significant and positive responses to the monthly mean temperature, maximum temperature, and minimum temperature for the period from early spring (April to May). At the higher-latitude sites, minimum temperature covering the period of

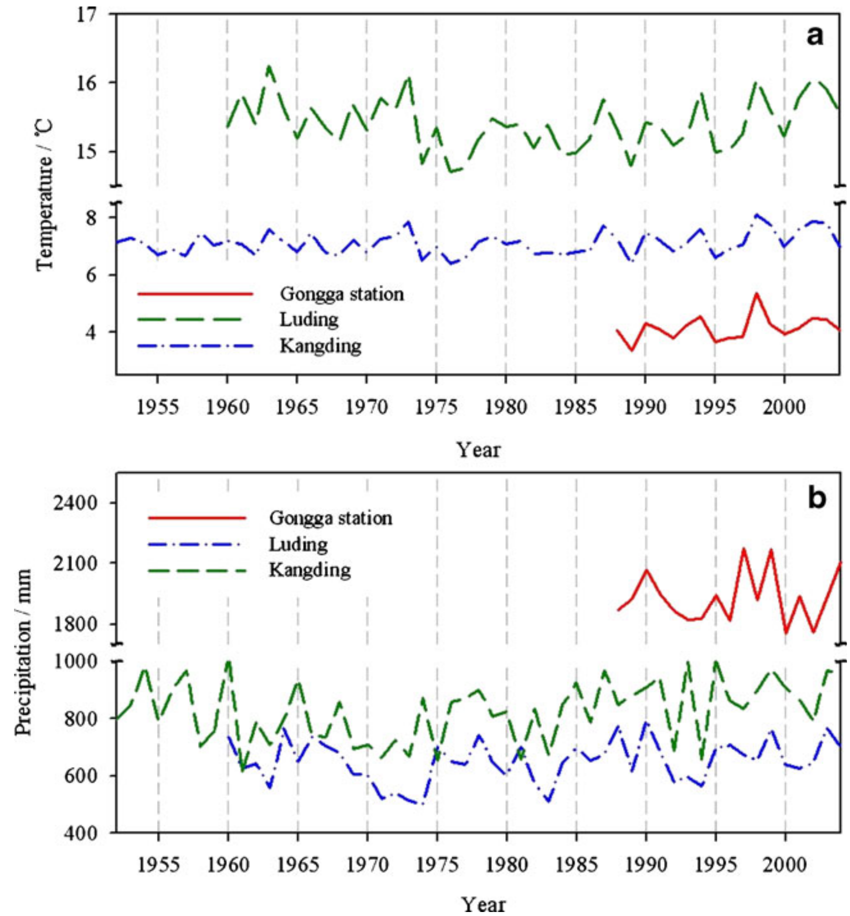

Fig. 2 Temporal changes of the recorded mean temperature and total annual precipitation in near-by meteorological stations (Gongga Mountain *GGS*, Luding *LD*, Kangding *KD*)

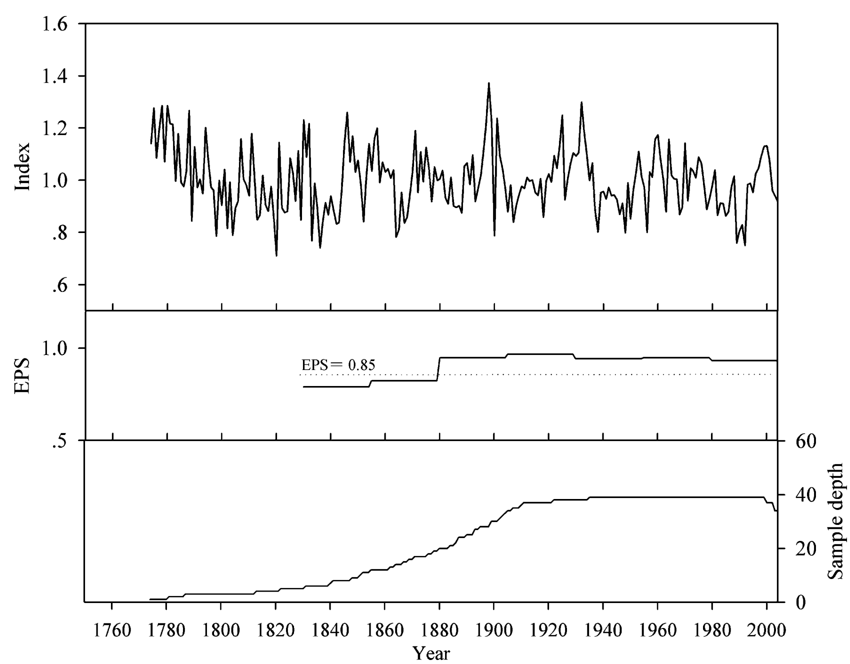

Fig. 3 Tree-ring width standard chronology of fir and sample depth (number of cores). The expressed population signal (EPS) statistic also shown

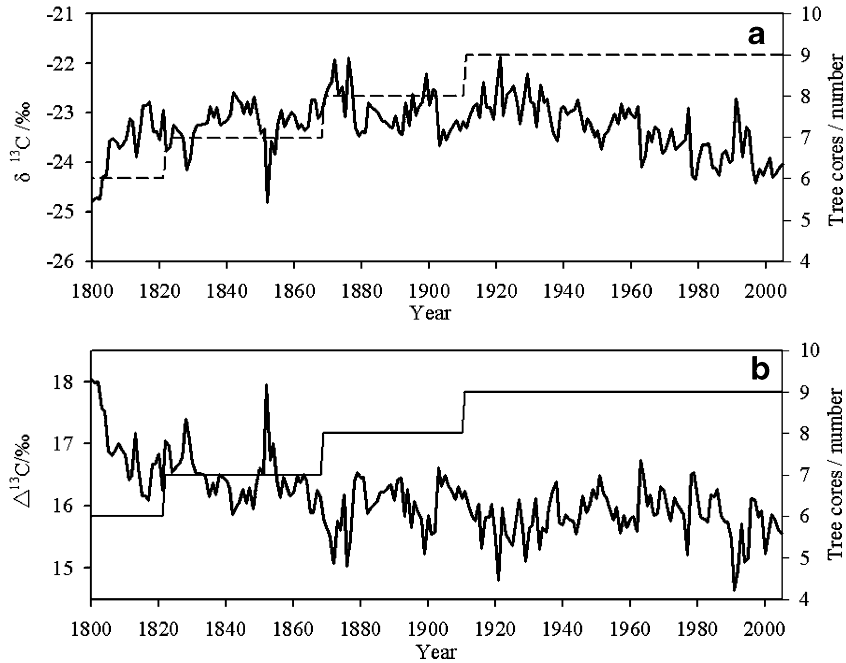

Fig. 4 **a** The stable carbon isotope $\delta P^{13P}C$ series of tree rings, along with the sample depth, **b** carbon isotope discrimination ($\Delta^{13}C$, CID) series which removed the effects of increasing CO_2 concentration

January–May has strong influence on carbon isotope discrimination in trees.

Interestingly, the previous summer monsoonal rainfall and current August precipitation had positive effect on tree growth. Precipitation in previous December to March and current May, falling mainly as snow, exhibits negative effects on tree growth. Positive influences of precipitation from October to December of the previous year on carbon isotope discrimination in trees were found. Although the monthly signal varies somewhat between tree-ring width and climate data in the stations with different altitude, tree growth is generally favored by warm and dry winters and springs, and wet growing seasons especially in current August. The carbon isotope discrimination in subalpine trees was controlled by the climate conditions of previous growth seasons (e.g., positive effects of temperature and precipitation), suggesting that climate conditions at the onset of growth are critical to carbon isotope discrimination during the growth season.

In general, responses of tree-proxy to climatic data of different meteorological stations had some differences in months and intensity. However, the common information of climatic controls on tree growth and carbon isotopic discrimination were identified. For much more reliability on climatic reconstruction, the climatic data of GGS was just used in further analysis.

Table 2 Correlation coefficients calculated between the residual (RES) and standard chronologies (STD), carbon isotopic discrimination series (CID) and climatic series of monthly mean temperature (mean-T), monthly maximum temperature (max-T), monthly minimum temperature (min-T), precipitation (PRE), and relative humidity (RH) of the significant months

	Months/STD	Months/RES	Months/CID
Mean-T			LD: Apr–May, r=0.39, P<0.01 KD: Apr–May, r=0.38, P<0.01
Max-T	LD: Sep/p-Nov/p (r=0.44, P=0.003); Jun (r=0.29, P<0.05)		LD: Apr–May, r=0.38, P<0.01
Min-T		GGS: Sep/p-Jan, r=0.57, P=0.02 Apr–Jun, r=0.51, P=0.04	GGS: Jan–May, r=0.58, P<0.05 LD: Apr–May, r=0.33, P<0.05
PRE	GGS: Jun/p-Sep/p, r=0.51, P=0.04 LD: Oct/p (r=0.35, P<0.05); Dec/p-Mar (r=-0.33, P =0.03); May (r=-0.30, P <0.05); Aug (r=0.43, P <0.01)	LD: Dec/p-Mar (r=-0.31, P=0.04), Aug (r=0.42, P<0.01)	KD: Oct/p-Dec/p, r=0.36, P<0.01
RH	GGS: May, r=0.51, P=0.04		

3.4 Tree-ring width/$\Delta^{13}C$–snowpack depth relationships

The seasonal snowpack depth distribution indicates that snowfall generally begins in October and snow cover lasts until mid May to early June (Fig. 5). This period can be divided into three sub-periods as post-monsoon (Oct/p-Nov/p), winter (Dec/p-Feb), and pre-monsoon (Mar–May) periods (Kothawale et al. 2008). There is a relative high snow accumulation in post-monsoon period and a lower snowpack depth values in winter. The mean highest snow depth recorded at the mountain station (GGS) was found during the spring and early summer (February–May).

Generally, the higher snowpack accumulation in high-altitude mountain areas is determined by the lower temperature and higher precipitation (falling as snow) during late autumn, winter, and early spring. The TRW and $\Delta^{13}C$ of trees at the timberline in this study were mostly controlled by the temperature and precipitation during this period. An association between TRW/$\Delta^{13}C$ and snowpack accumulation, therefore, is possible.

Based on prior assumption, a transfer function was calculated based on a simple linear regression model, using the TRW (STD) and $\Delta^{13}C$ as the dependent variable and the total accumulation of snow depth covering the period of previous October to current May as the independent variable. Significant negative relationships were found among tree-ring parameters and snowpack depth variations (Fig. 6a and b) in the GGS stations. When combing the two tree-ring proxies together, more robust interconnection has been found (Fig. 6c). The reconstruction accounts for 58.0% of the actual snow-depth variances, agreeing well with variations in the actual snow depth during their common period 1988–2004, especially in low-frequency fluctuations (Fig. 6c). The leave-one-out cross-validation method was employed to verify our reconstruction (Michaelsen 1987; Fritts 1976). The Person's correlation coefficient (r) value is 0.61. The cross-validation test yielded a positive RE (0.35), indicating predictive skill of the regression model. The sign test (ST) value is 10+/6−.

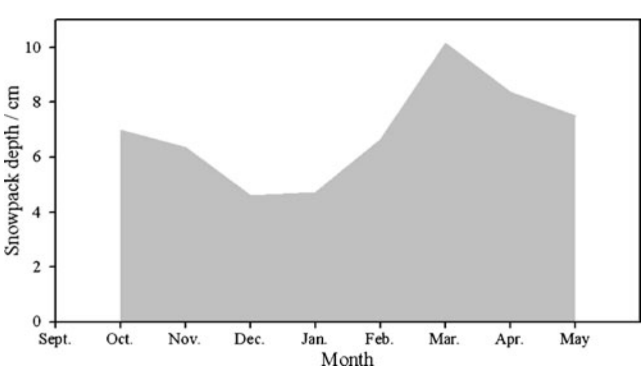

Fig. 5 Seasonal snow distributions in the study area covering the winter and spring seasons (from previous October–current May)

We also test the tree-ring proxies and the long snow-depth record of KD station. The significant and negative correlation between tree-ring width and snow depth was also found ($r=-0.33$, $P=0.02$); figure not shown). However, the correlation between tree-ring $\Delta^{13}C$ and snow depth were not significant. These distinct responses between tree-ring proxies and snow-depth record in different sites may depend on location within the Plateau, resulting in spatially heterogeneous snow accumulations.

3.5 Pilot snowpack-depth reconstruction

On the basis of the model, the reconstructed snowpack depth series covered the period A.D. 1880–2004 of reliable internal signal strength (Fig. 7), based on the commonly acceptance EPS statistics for tree-ring width chronology (>0.85, Wigley et al. 1984) and a sufficient sample depth for isotopic analysis ($n>8$). The most notable features of the reconstruction are the higher snow accumulation around 1990. The reconstruction also shows that the Gongga Mountain experienced some lower snow-depth episodes during 1910s, 1930s, 1950–1980, and later 1990s. The higher snow-depth intervals occurred during 1910s, 1940s, and the period around 1990 with the highest values during past 100 years. We defined extreme snowpack depth years as those years with values more than one standard deviation (plus or minus) from the average. Although there are several clusters of extreme years over the past century, the century is notable for the long period of snow-depth variations in a normal way except the higher snowpack depth values around 1990. There are 15 higher and nine lower extreme snowpack depth years totally.

4 Discussion

4.1 Tree growth- and $\Delta^{13}C$-climate relationships

Forest timberlines are affected by the most severe climatic conditions, including low temperature, strong winds, short growth season, and more snow accumulation. Generally, TRW responds to temperature variations at high altitudes as reported from other sites at cold high-elevation sites in northwest China (Liu et al. 2005; Gou et al. 2007; Liang et al. 2008). The radial growth of timberline *A. fabri* was correlated to the temperature of previous September to January and April to June. It is possible that a warm autumn is favorable for fir growth in the next year at the timberline by synthesizing non-structural carbohydrates and other organic substances, although cambial activity has already stopped in October. Such climate response was also reported for timberline forests growing in the cold-moist environment on the east, northeast and southeast Tibetan

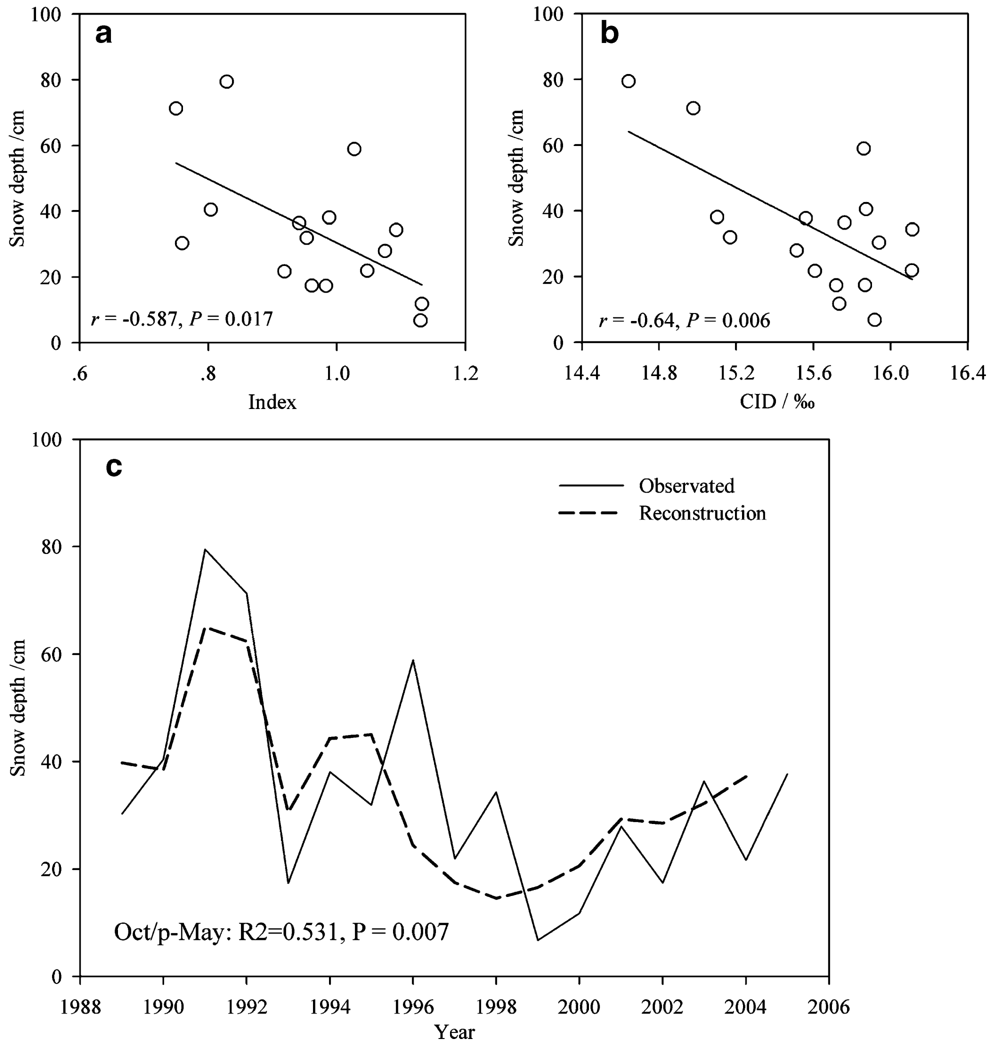

Fig. 6 **a** Correlation between standard tree-ring index and snow depth; **b** correlation between carbon isotope discrimination and snow depth; **c** comparison between observed and reconstructed snow depth. The final calibration model accounts for 58.0% (P=0.006) of the total variance of the snow depth over the calibration period 1988–2004

Plateau and Qilian Mountains (Shao and Fan 1999; Brauning 2001; Liu et al. 2005; Gou et al. 2007; Liang et al. 2009), and in other areas of Northern Hemisphere (Wilson et al. 2007; Büntgen et al. 2008). For carbon isotopic discrimination in trees, the dominant period on tree carbon fixation was mainly found in April to May or January–May at the alpine zones. This period is crucial and determines the onset of tree growth in the current year. If the previous winter and current spring temperature is very low, the annually frozen soil layer could be thicker and the thawing time could be delayed in the coming growth season. High temperatures in early summer are effective for tree growth by prolonging the duration of growth season (Camarero et al. 1998). Such a climatic response was also found in the previous study in the Qilian Mountains, northwest China (Liu et al. 2007). Anyway, the significant correlation

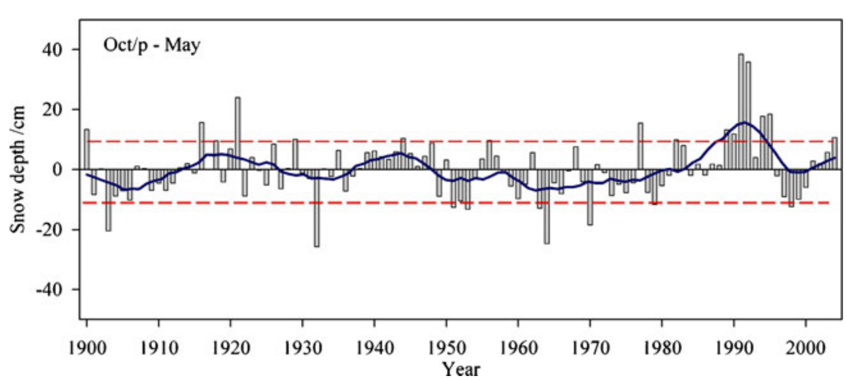

Fig. 7 Reconstructed snow depths in the Gongga Mountain over the period of 1890–2004. The values of snow depth were transformed into the departure against the mean of 1890–2004. The thick line was smoothed with an 11-year FFT-filter (fast Fourier transform) to emphasize long-term fluctuations

periods between the two tree-ring parameters and temperature mainly focused on snow cover or snow melt seasons in subalpine region.

Precipitation of the previous growth season plays a positive role for tree growth of the current year (Table 2). In contrast, more precipitation received in the period when tree growth stopped showed negative effects on current-year tree growth, suggesting that more precipitation of previous winter and spring, which falls as snow, may cause a latter tree growth onset in current year. The similar influence of heavy snow on the growth of trees was also found at a high-altitude forest in Japan (Takahashi et al. 2005). In a detailed study of growth response to climate in the Olympic Mountains, Ettl and Peterson (1995a, b) pointed out that winter precipitation (snowfall) was negatively correlated with tree growth, and the strength of correlations was highest at wet mid-elevation sites than that at dry sites. This suggests that duration of snowpack depth dominates growth response to climate by affecting the length of the growth season. The positive effects of current August precipitation on tree growth may be a result of water stress due to more transpiration caused by plant and soil evaporation by higher temperature in summer (Ettl and Peterson 1995b). Carbon isotope fixation in tree rings is less controlled by precipitation of the current growth season, but the significant positive influence of precipitation in previous October to December was displayed.

In the subalpine zone just below the timberline, late-persisting snow is one of the primary factors restricting conifer establishment and growth in the subalpine parkland (Peterson and Peterson 1994). Snow has both positive and negative effects on trees, including protection from winter desiccation (Kajimoto et al. 2002; Oberhuber 2004) and damage by snow movement (Shen et al. 2001; Seki et al. 2005), respectively. Tree growth in high-snowfall environments (under a marine climate and near the treeline) is generally limited more by precipitation than by temperature, with growth being negatively correlated with snowpack depth (Peterson and Peterson 1994). This supports the tree growth–snowpack depth correlations in our study. The high snow accumulation (about 771.0 mm precipitation occurs as snowfall) can bring a low temperature and a prolonging snow-melting period at the beginning of tree growth, resulting in a short growth season. The negative response of $\delta^{13}C$ (positive response of $\Delta^{13}C$) in tree rings to winter temperature in relation to the depth of winter freezing was also reported in a study in northern Finland (McCarroll et al. 2003). However, the effects of the heavy snowfall can induce a durative tree growth reduction for many years (Kajimoto et al. 2002). This makes the responses of tree growth and CID to snow-depth fluctuations becoming complicated and partially associating in low-frequency signals.

4.2 Validation of snowpack reconstruction

Because of the short snowpack depth record at the high-elevation station (GGS), our snow-depth construction needs further comparison with other longer observed records to validate its regional behavior. The snowpack-depth record (with lower absolute values) covers the 1959–2004 period in the nearby Kangding meteorological station. At annual scale, our reconstructed snow-depth shows a general agreement with the record in Kangding except for the heavy and high variability in snow depth around 1985 (Fig. 8). When the abnormal values in 1984–1986 were excluded, the two series showed a coherent trend to a certain degree (r=0.27, P=0.07). The correlation was improved significantly when the two series were treated by 3-year running mean (r=0.45, P=0.04). If the reduced freedom was considered, the correlation was still significant, indicating the common low-frequency variations in both series. Furthermore, in comparison to the observed average record of the stations on the Tibetan Plateau (Ma 2008), the slight increasing trends in snowpack depth changes of two series were concurrent wonderfully, although the inter-annual variations were not corresponding well (Fig. 8).

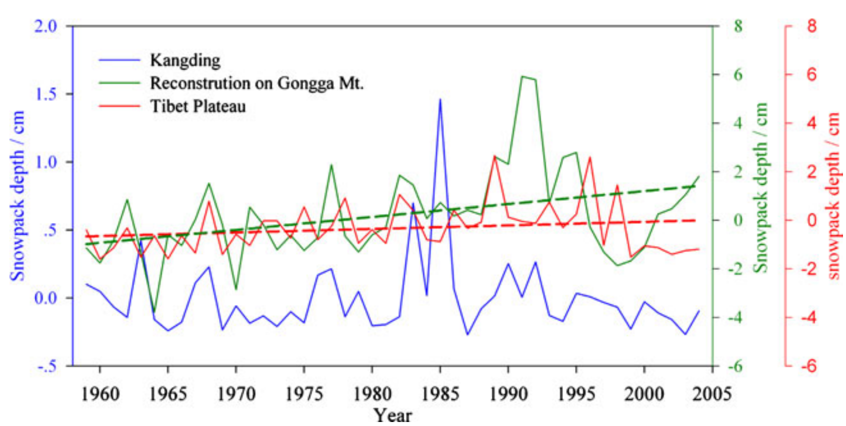

Fig. 8 Comparison between reconstruction snow depth and the neighboring records in Kangding stations and the mean snow depth recorded by the meteorological station above 2,000 m.a.s.l covering the Tibet Plateau. The *dashed lines* are the long-term increasing trends by linear regression

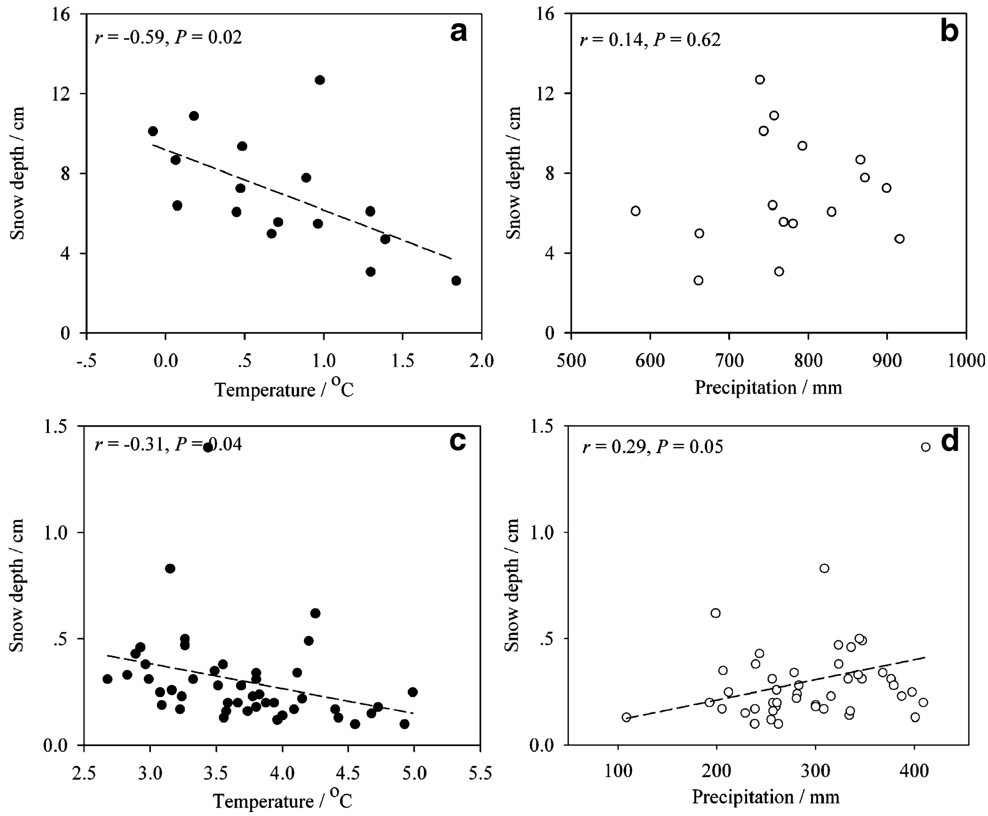

Fig. 9 Effects of temperature and precipitation on snow cover recorded in Gongga Mountain and Kangding stations. The panes **a** and **b** and **c** and **d** represent the correlations of climate parameters and snow depth in meteorological stations of Gongga Mountain and Kangding during the common periods

Comparison with reconstructed snow-depth variations and glacier retreat or advancing in Hailuogou (Li et al. 2009) indicated that during the periods of 1930–1960 and 1980–present, the retreat of glacier is corresponding to the snow-depth decreasing trend. Although without the long-term downward trend in snow-depth variations, a good agreement exists between snow depth and ice-core accumulation in Dasuopu (Duan et al. 2006) during the periods of 1920s, 1940s, and 1990s. These comparisons provided strong evidence of the reliability of our reconstruction, especially in the long-term variations. However, we should be cautious in the uncertainty and potential error of reconstructed snowpack depth. For example, the short snowpack observed record would give bias on evaluation on the correlations between tree-ring proxies and environmental variables. The controlling factors on snow accumulation in mountains areas are more complexity.

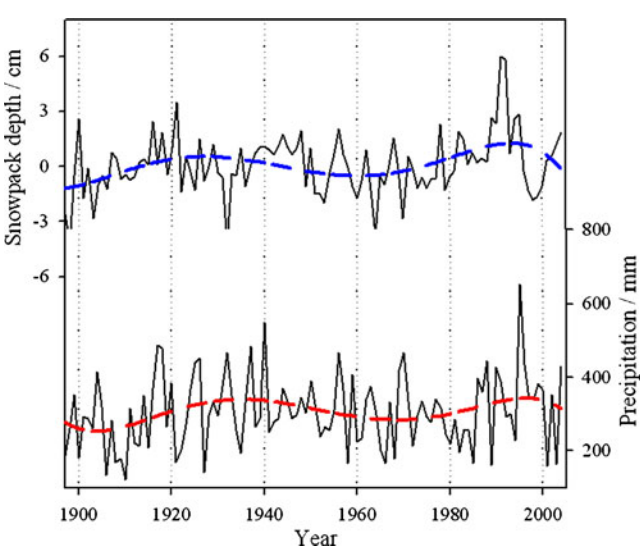

Fig. 10 Comparison of reconstructed snow depths in the Gongga Mountain and the East India precipitation coverings the period (Oct/p-May). The *thick lines* are the five-order polynomial trends lines to emphasize the long-term fluctuations

4.3 Possible factors on long-term snow-depth variations

In general, higher precipitation and lower temperatures are requisite to bring and preserve high snow accumulation on the Tibetan Plateau (Ke et al. 1997; Wei et al. 2005). The correlation between snowpack depth and temperature/precipitation in both the Gongga Mountain and Kangding meteorological stations (Fig. 9) showed that winter temperature had negative and significant effects on snow accumulation, in contrast with a minor effect from precipitation. This implies that the prevailing East Asian winter monsoon will affect snow accumulation (Guo 1994) when the warm and wet air from ocean flows to the plateau by the South Asian winter monsoon through the great and deep valley in the Hengduan Mountains. During the post-monsoon periods, the India (or south Asian) monsoon became weaker. But the warm and wet air mass can invade into the Plateau at given periods with the increasing thermal contrast between the Tibetan Plateau and the tropical Indian Ocean. Thus, we compared the variations of Indian rainfall in east India (Sontakke et al. 2008) and our snowpack-depth reconstruction covering the common period (previous October–current May). Even though there was no good interconnection at the entire period between the two series at annual scale, the long-term fluctuations agree well (Fig. 10). These results imply that precipitation, falling as snow during the period of the previous October to current May in the Gongga Mountain, may be a result of the wet and warm water vapor from India Ocean though great and deep valleys in the Hengduan Mountains (He et al. 2003; Liu et al. 2010) or a feedback of snow cover in the TP and India monsoon. Unfortunately, it is difficult to prove this source-and-pool relationship based on this study.

5 Conclusions

We have described a pilot reconstruction of the snow depth for the Gongga Mountain based on the subalpine fir ring-width and $\delta^{13}C$ chronologies at the timberlines, indicating inter-annual to multidecadal scale snow-depth variability over the past century. This reconstruction sheds new light on cryospheric variability and change in a region where past climate history data is lacking. Our results suggest that combing tree-ring width and $\delta^{13}C$ in certain subalpine tree species growing on the Tibetan Plateau may be an effective way for reconstructing regional snowpack variations. Considerable efforts should be made to explore the detailed physiological response of tree-ring parameters to snow variability in further work.

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