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Tree-ring evidence of 'Little Ice Age' glacier advances in southern Tibet

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Abstract: The history of late Holocene glacier fluctuations in eastern Tibet was studied by determining the ages of trees growing on glacier deposits. Maximum tree ages yield minimum ages of AD 1760 and 1780 for moraine formation at the maximum extent of the 'Little Ice Age' glacier advances in two glacier forefields. Subsequent moraines could be dated to the beginning of the nineteenth and the beginning of the twentieth century. Larch trees from a third glacier forefield in southeastern Tibet show evidence of glacier activity from 1580 to 1590, from the end of the eighteenth to the beginning of the nineteenth century and from 1860 to 1880. One glacier at Mt Gyalaperi recently advanced in both 1951 and 1987. Periods of glacier advances can partly be correlated with periods of growth reductions in chronologies of total ring width and maximum latewood density derived from trees growing on slopes above the glacier valleys. Correlation functions with meteorological data suggest that maximum latewood density of subalpine *Picea balfouriana*, *Larix griffithii* and *Abies delavayi* var. *motouensis* is positively correlated with summer temperature, while ring width of these species and of subalpine *Juniperus tibetica* is also sensitive to winter conditions prior to the growing season.

Key words: Dendroglaciology, dendroclimatology, glacier history, summer temperatures, 'Little Ice Age', Tibetan Plateau.

Introduction

The Tibetan plateau represents a huge heating surface in subtropical latitudes with an average elevation above 4000 m and a surface of more than 2×10^6 km^2. Therefore, it plays an important role in the development and dynamics of the Asian monsoon system (Murakami, 1987; Prell and Kutzbach, 1992; Webster *et al.*, 1998). However, the strength of the Tibetan low pressure system, which is one of the major driving forces of the Indian and East Asian summer monsoon, is strongly influenced by the extent of the snow cover over northern Eurasia during the preceding winter. After cold winters with extensive snow cover, the heating of the plateau surface is delayed and weakened owing to the loss of reflected solar energy and the amount of sensible heat that is required for warming the soil surface (Barnett *et al.*, 1988; Vernekar *et al.*, 1995; Sankar-Rao *et al.*, 1996). More than 80% of the annual rainfall in this region falls during the summer monsoon season between June and September. Thus, most of the annual accumulation and ablation of glaciers in this region occurs during the summer months (Ageta and Fujita, 1996; Benn and Owen, 1998 and references therein) and makes them an important source of information about monsoon dynamics and changing climatic conditions.

Today, the glaciated area on the Tibetan plateau covers about 57 000 km^2 (Thompson *et al.*, 1995). During recent decades, glaciers and permafrost areas in Tibet and in the Himalayas have experienced tremendous shrinkage (Mayewski *et al.*, 1980; Yamada *et al.*, 1992; Jin *et al.*, 2000; Narama, 2002; Zech *et al.*, 2003). Therefore, prominent lateral and terminal moraine ridges formed during former glacier advances can be observed in many glacier forefields. Their fresh morphology, poor soil development and sparse vegetation cover attest to their relatively recent formation. Detailed records about the younger glacier history on the Tibetan plateau, however, are scarce.

Tree-ring width and $\delta^{13}C$ chronologies from long-living *Juniperus tibetica* from Qamdo in eastern Tibet (Figure 1) have shown that the period from 1150 to 1380 was characterized by higher temperatures than today (Helle *et al.*, 2002; Bao *et al.*, 2003). The succeeding period from 1430 until the late nineteenth century witnessed a series of cold intervals indicated by growth reductions and changes in the isotopic composition in juniper trees (Wu, 1992; Bao *et al.*, 2003). This period, which was not a uniform cold event over the whole Northern Hemisphere (Bradley and Jones, 1993), is here referred to as the 'Little Ice Age' (LIA) period in Tibet. The glaciers under consideration are comparatively small, have a maritime

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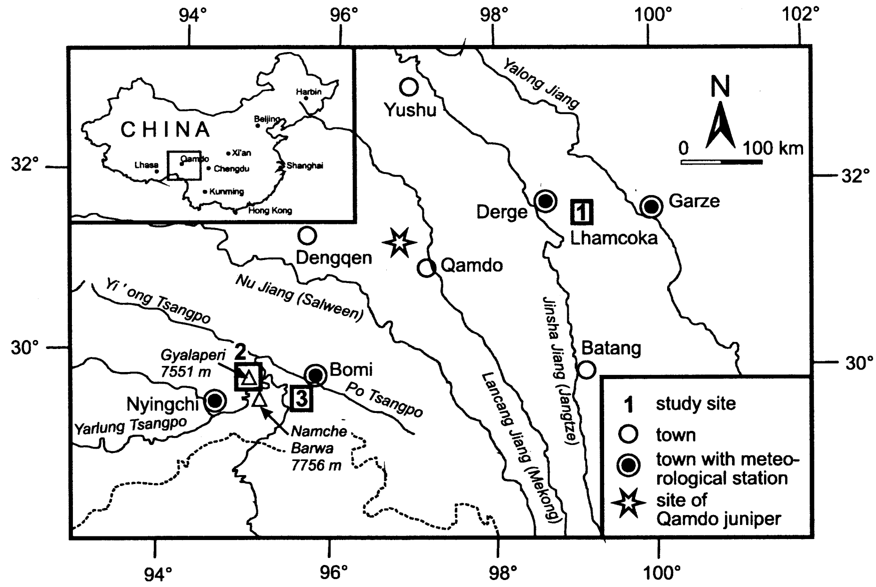

Figure 1 Sketch map of southeastern Tibet and location of the study areas

temperature regime and are thus very sensitive to climatic changes (Zhou *et al.*, 1991). In addition, their present or former glacier terminus lies below the upper tree limit. This provides the opportunity to study the glacier history using dendrochronological methods that have been widely applied to derive precisely dated records of former glacier fluctuations (eg, Heikkinen, 1984; Holzhauser, 1985; Solomina, 1996; Wiles *et al.*, 1999).

In central eastern Tibet, forests on south-facing slopes near the upper tree limit are composed of *J. tibetica* while *Picea balfouriana* is found on all other exposures. The alpine timberline reaches 4500 m on north-facing slopes, but isolated Juniper trees on south-facing slopes may reach an elevation of more than 4600 m (Frenzel, 1998). In the marginal mountain ranges of southeastern Tibet the forest composition is much more complex and includes a large number of tree species. The upper tree line in this very humid area lies at 4350–4250 m. Of special interest for this study are pioneer tree stands of *Larix griffithii* on lateral and terminal moraines. Additional information about climate fluctuations can be derived from studies of *Abies delavayi* var. *motouensis* and *A. georgii* var. *smithii,* which form dense forests on neighbouring slopes.

Sample collection and data treatment

Chronology development and dendroclimatological investigations

The study site locations and the number of analysed trees are given in Table 1 and Figure 1. From almost all sampled trees, two increment cores were collected. Maximum latewood density chronologies (MLD) were developed at the Swiss Federal Research Institute in Birmensdorf according to the techniques described by Eschbach *et al.* (1995). Ring width is automatically recorded during this procedure. From trees where MLD was not registered, ring width (RW) was measured with a precision of 0.01 mm using a LINTAB II measuring table. Crossdating between the individual growth curves was accomplished by visual comparison and the statistical parameters *t*-test and sign test calculated by TSAP software (Rinn, 1996). Long-term growth trends in the raw data caused by tree aging were removed. In the case of maximum latewood density (MLD), a linear trend line was subtracted from the original values (Cook *et al.*, 1990; Wilson and Luckman, 2003). RW data were standardized by a two-step procedure. First, the ring-width series were detrended by division through an exponential or linear growth curve (Fritts, 1976). Second, the series were filtered with a cubic smoothing spline (Cook and Peters, 1981) with a 50% frequency response function cut-off related to 67% of the length of each series. A major constraint for the climatological calibration of the tree-ring series is the low spatial density and short timescale for the available climate data (Table 2, Figure 1). For calibrating climate–growth relationships, regional series of monthly temperatures and precipitation were created from pairs of the meteorological stations Garze/Derge and Nyingchi/Bomi by applying the methods described by Jones and Hulme (1996). Pearson correlation coefficients were calculated between tree-ring data and monthly series of temperature and rainfall for a 15 month period ranging from July of the summer prior to growth until September of the growth year. In addition, seasonal means were calculated. The highest correlation coefficients found are shown in Table 1. Temperature reconstructions were derived by linear regressions between tree-ring chronologies and the climatic factors that showed the strongest impact on growth in the preceding correlation analyses.

Dendrochronological dating of glacial deposits

The maximum age of trees growing on glacial deposits gives a minimum estimate of the date of sediment formation. The timespan between the retreat of the glacier and the establishment of the first seedlings depends on the nature of the substrate, climatic conditions, altitude, distance of seed dispersal and distance of seed-trees (Sigafoos and Hendricks, 1969; Heikkinen, 1984). Estimates of this so-called 'ecesis' period on glacial sediments vary widely from only a few years (Lawrence and Lawrence, 1959; Sigafoos and Hendricks, 1969, 1972; Heikkinen, 1984; Luckman, 1993; Wiles and Calkin, 1994; Wiles *et al.*, 1999) up to several decades or even one century (Mercer, 1970; Burbank, 1981; Luckman, 1988; Villalba *et al.*, 1990; Harrison and Winchester, 2000). Usually, however, ecesis dates range from 10 to 30 years (Heusser, 1956; Bray and Struik, 1963; Viereck, 1967; McCarthy and Luckman, 1993; Harrison and Winchester, 2000).

In all glacier forefields examined, the glacial deposits of the LIA advances are situated several hundreds of metres below the upper tree limit. Seed-providing trees cover the nearby slopes, so a quick recolonization of the glacial deposits after a

Table 1 Characteristics of ARSTAN standard tree-ring chronologies and best correlations with climate data

	Lhamcoka BD		Lhamcoka E	Gyalaperi LA		Gyalaperi AB	
	RW	*MLD*	RW	RW	*MLD*	RW	*MLD*
Latitude / longitude	31°49′N/99°07′E		31°49′N/99°06′E	29°54′N /94°53′E		29°53′N/94°53′E	
Elevation	4150–4370		4250–4350	3780–3820		4000	
Tree species	*Picea balfouriana*		*Juniperus tibetica*	*Larix griffithii*		*Abies delavayi* var. *motouensis*	
Chronology period	1630–1993		1187–1995	1782–1993		1774–1993	
Chronology length	364		809	212		220	
No. of trees / radii	20/39		16/32	20/37		9/19	
Calibration period (1957–1980): r / R^2	$T_{(Nov/py-Jan)}$ 0.66***/0.44	$T_{(Aug-Sep)}$ *0.72***/0.52*					$T_{(Aug-Sep)}$ *0.79***/0.62*[a]
Verification period (1981–1993): r / R^2	0.61*/0.37	*0.61*/0.37*					*0.57*/0.33*
RE	0.01	*0.01*					*0.14*
CE	0.02	*0.40*					*0.41*
Highest r with temperature (T) or precipitation (N) data			$T_{(Nov/py-Jan)}$: 0.54** $N_{(Feb-Apr)}$: 0.53***	$T_{(Dez/py-Jan)}$ 0.46**	$T_{(Aug)}$ *0.55****	$T_{(Jan)}$ 0.58***	

*$p < 0.05$; **$p < 0.01$; ***$p < 0.001$.
[a]Calibration period 1961–1981.
Data for MLD chronologies are in italics. RW, ring width; MLD, maximum latewood density.

glacier retreat is ensured. Hence, an ecesis of only 5 years was added to the date of the oldest tree to yield a minimum age of the respective moraine stage. Since moraines are usually not covered by trees homogeneously (Villalba *et al.*, 1990), it is particularly important to find the oldest tree of each respective stage (Lawrence, 1950; Heikkinen, 1984). This can not be guaranteed, but is strongly supported by the finding of continuously decreasing maximum tree ages from the outer to the inner moraine stages. The oldest tree ages found are far below the potential maximum age of the respective species, which ensures that the studied trees belong to the first tree generation in the natural succession process.

The trees were cored 1 m above the ground to guarantee a high core quality for dendroclimatological analyses. As a consequence, a timespan between germination and growth to the coring height must be added to the counted tree age to estimate the germination date. Since limited time in the field in the very remote study areas did not allow the sampling of material to establish site-specific age–height growth functions, an estimate of this date must be obtained from studies of comparable environments published in the literature. In the Lys glacier forefield of the Italian Alps, the time interval enclosing ecesis and growth to 1-m tree height varies from 33 to 42 years (Strumia and Schweingruber, 1996). Sigafoos and Hendricks (1972) gave an estimate of 10 years for this interval for 200-yr-old trees with a diameter of about 60 cm. Since this coincides closely with the trees used in this study, 10 years were added to the pith date. While Heusser (1956) found an underestimation of the germination date of spruce trees of up to 30 years when samples were taken at the 30 cm level, McCarthy *et al.* (1991) ascribe this value only to trees with suppressed growth and determined an average time span of 6 to 19 years required to reach a height of 50 cm. Thus, only 5 years were added to the pith date of the younger trees from the inner moraine stages, which were cored at 30 cm height (McCarthy *et al.*, 1991). In those cases where the pith of the tree was missed by the borer, the number of missing rings was estimated as a function of the width and the curvature of the innermost rings preserved in the core and added to the number of countable rings to provide the absolute age of the tree (McCarthy and Luckman, 1993; Villalba and Veblen, 1997).

Table 2 Name and location of meteorological stations

Station	Location	Altitude (m a.s.l.)	Period (years)
Garze	31°38′N / 99°59′E	3393	1951–2000 (51)
Derge	31°44′N / 98°34′E	3201	1957–2000 (44)
Bomi	29°52′N / 95°46′E	2736	1961–1997 (37)
Nyingchi	29°34′N / 94°28′E	3000	1951–2000 (49)

Results

Site Lhamcoka

Lhamcoka (Chinese: Xinluhai) is the Tibetan name of a glacial lake situated in the Chola Shan (shan = chinese for 'mountain'), a mountain range running NW–SE rising to 6168 m a.s.l. Tree-ring samples were collected in five locations within the area (sites Lham A–E, Table 1, Figures 2–4). Sites Lham A and Lham C are situated in two neighbouring glacier forefields formed by outwash plains and moraine ridges. On the basis of the minimum ages of the trees growing on these moraines, a provisional chronology of the glacial history for the last 200 years was established by Bräuning and Lehmkuhl (1996). Meanwhile, the site was revisited and recollected to gain material for wood density analyses and to improve the sample depth and hence the dating accuracy. Since the trees in the glacier forefields showed impact of human activity by chopping, additional trees were sampled in two undisturbed localities (Lham B and Lham D). These sites are located high above the former glacier forefields and had not been impacted by glacier advances during the LIA. Since the tree-ring chronologies of sites Lham B and D are extremely similar ($r = 0.73$; 1700–1994), they were combined to one single chronology (Lham BD) composed of 20 trees. A ring-width chronology with a length of 809 years was developed from *J. tibetica* (site Lham E, Table 1, Figures 3 and 4) from a site located above the lateral moraine of the LIA maximum stage.

Tables 3 and 4 list the ascertained tree ages and the derived minimum ages of the corresponding moraine stages (compare Figure 2). In both glacier forefields, minimum ages between equivalent moraine stages differ by about 10 years only. The

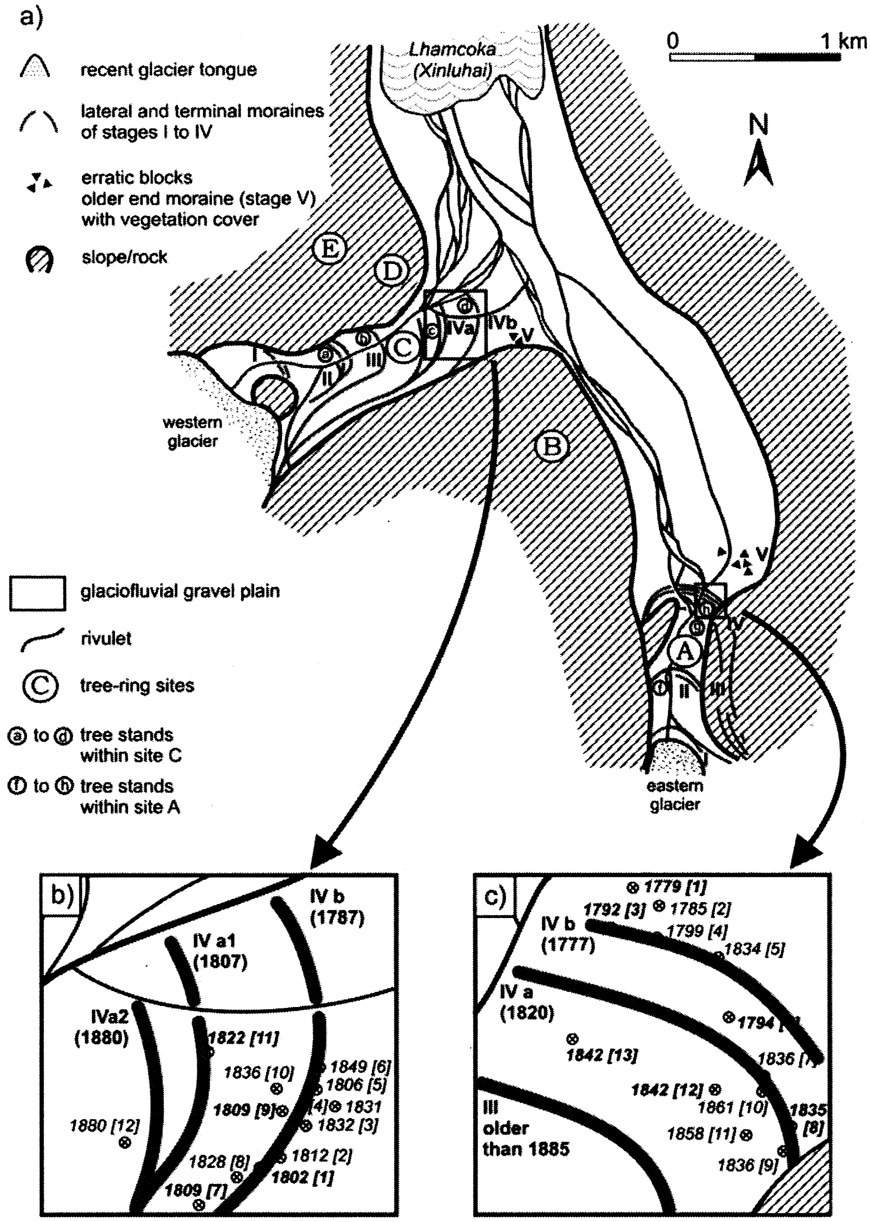

Figure 2 Map of the extent of recent glaciations and positions of the sampled trees from a glaciated area on the northern slope of the Chola Shan, south of the lake Lhamcoka (Xinluhai). (a) Overview and location of the tree-ring sites, (b) detail from the forefield of the western glacier, (c) detail from the forefield of the western glacier. Dates in (b) and (c) give the position of the trees and the absolute tree ages, bold italics give the age of the oldest tree of the corresponding moraine stage, numbers in brackets are the tree numbers referred to in Tables 3 and 4. Base map from Bräuning and Lehmkuhl (1996)

maximum LIA advance stage IVb has a minimum age of 1787 at the western versus 1777 at the eastern glacier. On the other hand, the positions and ages of trees 1 and 2 at the eastern glacier indicate that the surface of the outwash plain must have been stabilized around 1765 at a distance of 15 m and in 1770 at a distance of 5 m from the moraine. Hence, the drainage of the glacial meltwater did not occur diffusively over the whole outwash plain, but was already concentrated in clearly defined drainage channels probably using the same outlet positions through the moraine ridge as today. Therefore, the position of the glacier front must already have retreated from the terminal position, which is confirmed by the estimate of a minimum age of 1777 given by the oldest tree growing directly on the moraine crest (No. 3 in Figure 2c). Moraine stage IVa has a minimum age of 1807 at Lham C versus 1820 at Lham A. Difficulties arise for the precise dating of the younger moraine stages III and II owing to the small number of trees available. According to the tree data, stage III at the eastern glacier must be older than 1885. The date from the western glacier, which is older than 1918, is derived from one tree only and seems far too young. Stage II, which partly splits into a multiple ridge, is older than 1907. Stage I lies closer to the present glacier terminus and is bare of trees.

Calibration and verification statistics between the composite data set of the climate stations Garze and Derge and tree-ring series of *P. balfouriana* are given in Table 1. MLD is highly correlated with the temperature sum of August and September ($r = 0.72$), whereas no significant positive correlations exist with precipitation. RW is positively correlated with temperatures of early summer (June and July) and especially with November to January preceding the growth season ($r = 0.66$). A strong influence of winter temperature on tree growth has also been found by Scuderi (1987) in the Sierra Nevada (California) and is explained by the use of stored photosynthates in long and cold winters that are missing for the formation of earlywood in the next growth season.

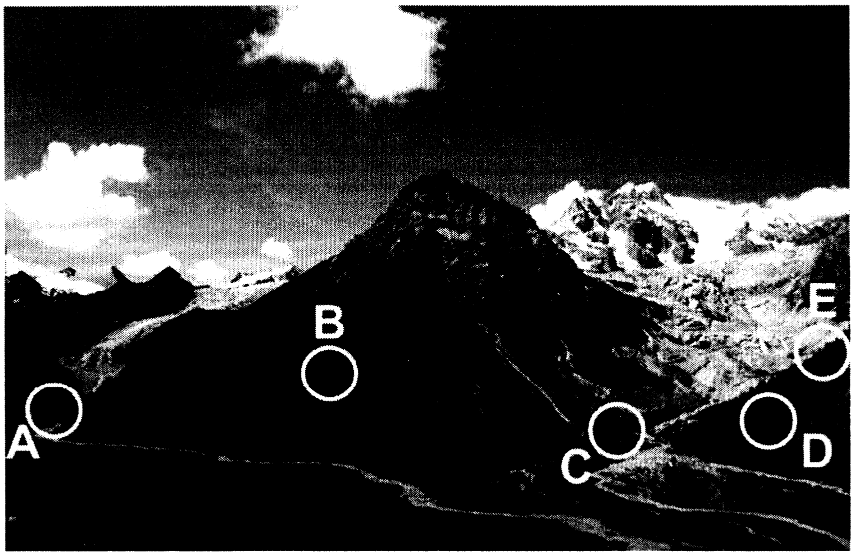

Figure 3 Overview of the forefields of the eastern and western glaciers south of Lhamcoka lake in the Chola Shan. White circles mark the tree stands investigated

Reconstructions of summer and winter temperatures were calculated based on linear regressions of the Lham BD MLD chronology with summer temperature and Lham BD RW chronology with winter temperature (Figure 5a, b). The 'Reduction of Error' (RE) and 'Coefficient of Efficiency' statistics are positive in all cases, indicating that the derived temperature reconstructions are reliable (Fritts, 1976).

The RW chronology of Lham E juniper (Figure 5c) is the only chronology showing significant positive correlations to precipitation during spring (February to April, 1957–1993: $r = 0.53$; $p < 0.001$). Since the absolute amount of precipitation is only low during this period (Garze: 38 mm, Derge: 39 mm), this could be interpreted in the sense that dry and cold winters with a prevalence of continental air masses can lead to growth reductions in juniper. Juniper grows at the upper timberline on southern aspects and is therefore exposed to frost dryness in late winter, when solar radiation on south slopes is already high. This interpretation is confirmed by the high positive correlation with winter temperature (Table 1). Besides, summer temperature is also positively correlated with ring width. The correlation coefficient with June to August temperature is 0.39 (1951–1993; $p < 0.05$). However, the juniper chronology is influenced by the effect of forest fires in the nineteenth and twentieth century, whereas the older parts of the chronology seem to be undisturbed (Figure 5c). Since the stand disturbances affect the calibration period, no climate reconstruction was calculated. In general, RW of juniper (Lham E) can be cautiously interpreted as a proxy for winter temperature and dry-cold versus warm-humid conditions.

As shown in Figure 5, the variability of winter temperature during the past 300 years was greater than the variability of summer temperature. This is probably related to periods of varying strength of the Siberian High. Although no direct linkage can be made between periods of low tree growth near the upper tree line and glacier advances, the high-altitude chronologies that cover the whole LIA period could indicate climate fluctuations that could have triggered the LIA glacier advances, as ring-width chronologies often show inverse trends to periods of glacier advance (LaMarche and Fritts, 1971; Scuderi, 1987). The maximum moraine stage IVb follows a period that is characterized by low index values in all three chronologies during most times in the period 1700–1770. (Figure 5). Low winter temperatures prevailed during 1730–1750 and 1760–1770. This long period of cold climatic

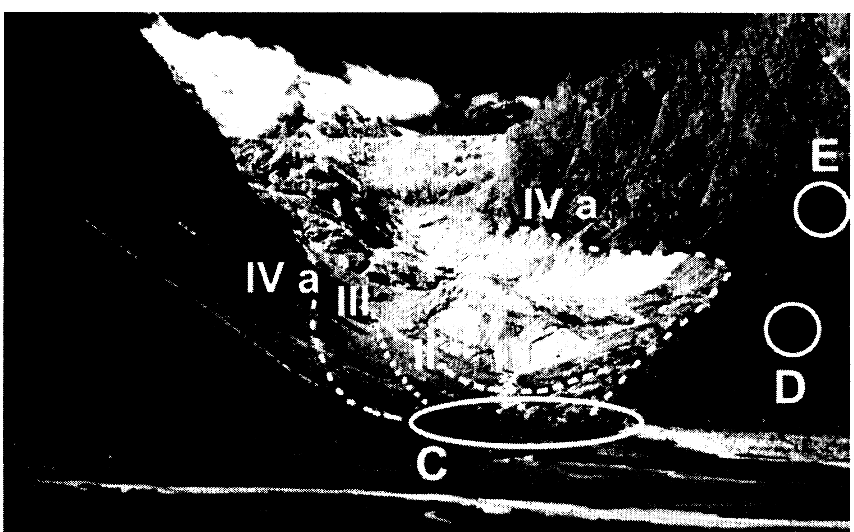

Figure 4 Forefield of the western glacier with the position of the tree-ring sites and moraine ridges

Table 3 Tree ages of site Lham A (*Picea balfouriana*) and estimated minimum ages of corresponding moraine ridges of the eastern glacier of Lhamcoka. The bold dates indicate the minimum pith dates of the oldest trees and the minimum age estimates for the corresponding moraine stages. The position of the tree stands and the trees Nos 1 to 13 is shown in Figure 2c

Tree number	Measured rings	Date of first measured ring	Missing rings to pith	Pith date	Tree stand	Minimum ages of moraine stages
1	211	1784	5	**1779**	h	outwash plain in front
2	202	1793	8	1785		of **IV b: 1765**
3	197	1798	6	**1792**		**IV b**
4	189	1806	7	1799	h	**1777**
5	157	1839	4	1834		
6	196	1799	5	**1794**		between IV a and IV b: 1780
7	153	1842	6	1836		
8	151	1844	9	1835	h	**IV a**
9	147	1848	12	1836		**1820**
10	125	1869	8	1861		
11	131	1864	6	1858		outwash plain
12	138	1852	7	1845	h	between stages
13	146	1849	7	**1842**		**III and IV a: 1830**
14	81	1911	9	1902		outwash plain
15	82	1910	10	**1900**	g	between stages **II and III: 1885**
16	40	1952	5	1947		
17	39	1953	11	1942		**II**
18	51	1941	9	1932	f	**1907**
19	54	1938	6	1936		
20	71	1921	4	**1917**		

conditions obviously was the triggering event for a major glacier advance culminating before 1780. Periods with cold summers occurred around 1700, 1725–1770, 1810–1820, 1900–1920, in the 1960s and after 1975. The retreating glacier stages IVa, III and II might be correlated with cooler periods at 1790–1800 and 1810–1820, around 1870 and at the beginning of the twentieth century.

Site Gyalaperi

Mt Gyalaperi (7151 m a.s.l.), north of the deep gorge of the Yarlung Tsangpo, is one of the easternmost high peaks of the Himalaya. At its northern end, six local glaciers merge into one huge valley glacier that reaches about 800 m below the alpine tree limit formed by *A. delavayi* var. *motouensis* at *c.* 4200 m a.s.l. The lateral moraine of the valley glacier is covered by a dense forest of *L. griffithii*, which is underlain by scattered dying junipers. Younger fir trees form a second tree layer below the dominant larches. Maximum tree ages were determined along a transect through the moraine ridge (Figure 6). On the distal moraine slope, the oldest germination date of larch is 1763, while the oldest trees on the crest and on the inner slope of the moraine show maximum ages of 1855 and 1865, respectively. The measured rings of a dying juniper on the crest of the moraine date from 1852 to 1994, the innermost *c.* 5 rings of the core are missing. Tree ages of firs on the outer moraine slope are 42 and 150 years. It can be concluded that the light-demanding pioneer larch forest mixed with juniper represents the first tree generation in a primary succession after the deposition of the moraine. This forest is now invaded by late-successional and shade-tolerant firs. According to the tree ages, the minimum age of the moraine is older than 1760. The inner slope of the moraine was still covered by glacier ice while the outer slope already carried open pioneer tree stands. The

Table 4 Tree ages of site Lham C (*Picea balfouriana*) and estimated minimum ages of corresponding moraine ridges of the western glacier of Lhamcoka. The bold dates indicate the minimum pith dates of the oldest trees and the minimum age estimates for the corresponding moraine stages. The position of the tree stands and the trees Nos 1 to 12 is shown in Figure 2b

Tree number	Measured rings	Date of first measured ring	Missing rings to pith	Pith date	Tree stand	Minimum ages of moraine stages
1	165	1830	28	**1802**		
2	169	1826	14	1812		
3	155	1837	5	1832		**IV b**
4	149	1846	15	1831	d	**1787**
5	171	1811	5	**1806**		
6	142	1853	4	1849		
7	176	1818	9	**1809**		outwash plain
8	161	1834	6	1828	d	between stages
9	177	1815	6	**1809**		**IV a and IV b:**
10	145	1847	11	1836		**1795**
11	166	1826	4	**1822**	c	**IV a**
12	107	1885	5	1880		**1807**
13	64	1928	0	**1928**	b	**III: 1918**
14	64	1928	7	**1921**	a	**II**
15	58	1934	0	1934		**1912**

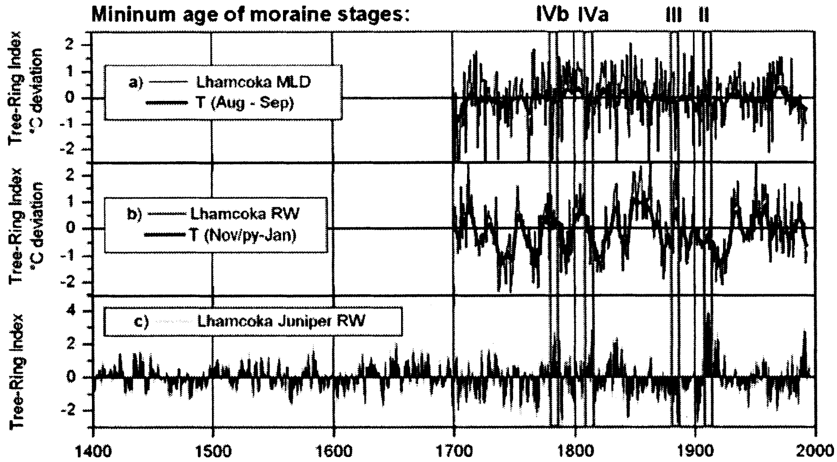

Figure 5 (a), (b) Maximum latewood density and ring width chronologies of Picea balfouriana from Lhamcoka BD and derived reconstructions of summer and winter temperatures. The temperature reconstructions were smoothed with a 5-year moving average. (c) Ring-width chronology Lham E from Juniperus tibetica. Glacial stages refer to Figure 2 and Tables 3 and 4

lifespan of one larch that was killed during a recent glacier advance dates from 1901 to 1987 (see asterisk in Figure 6), proving that the contemporary position of the glacier after 1987 is the highest since the middle of the nineteenth century.

The indices of RW and MLD of larch show positive correlations to winter and summer temperature, respectively (Table 1). However, two dramatic growth suppression events occur in 1951 and 1987 (Figure 7b, c). There is no periodicity in growth reductions that point to insect infestations like that of larch bud moth in the Swiss Alps (Weber, 1995). Since the extreme growth reduction after 1987 corresponds to the death year of one larch killed by a glacier advance, it can be assumed that the period of extremely reduced growth after 1951 also indicates an event of strong glacier movement. Ring width of larch has proved to be sensitive to summer temperature and to reflect climatic changes on a local scale also in Canada (Luckman, 1996). Beside larch, fir trees that grow on the steep rocky slope above the glacier were cored. Since these firs are not directly influenced by the glacier, and since MLD of fir revealed satisfactory calibration and verification statistics (Table 1), a summer temperature reconstruction was derived from fir, although MLD of larch shows similar growth variations (Figure 7). The young age of the firs does not provide information about climatic conditions during the LIA maximum glacier advance. Periods of cooler summers from 1850 to 1870 and in the middle of the twentieth century differ from the summer temperatures reconstructed for the Lhamcoka region (Figure 5a) and emphasize the local character of climate in the very humid glacier valley of southern Tibet. The cool summers around 1950 and after 1985 correspond to advance periods of Gyalaperi glacier.

Site Bomi

The study site is situated south of Bomi (29°46.5′N/ 95°42′E; Figure 1) in the same mountain system as site Gyalaperi. Two cores from each of four *L.* cf. *griffithii* were collected. The trees grew on the distal slope of a terminal moraine at *c.* 4000 m a.s.l. (Figure 8). One tree only had 117 rings and was excluded from further analysis, since dating was not possible. The substrate of the moraine consists of loamy sand containing blocks of about 0.5 m in diameter overlain by blocks of 1.5–2 m in diameter. It seems possible that the moraine is

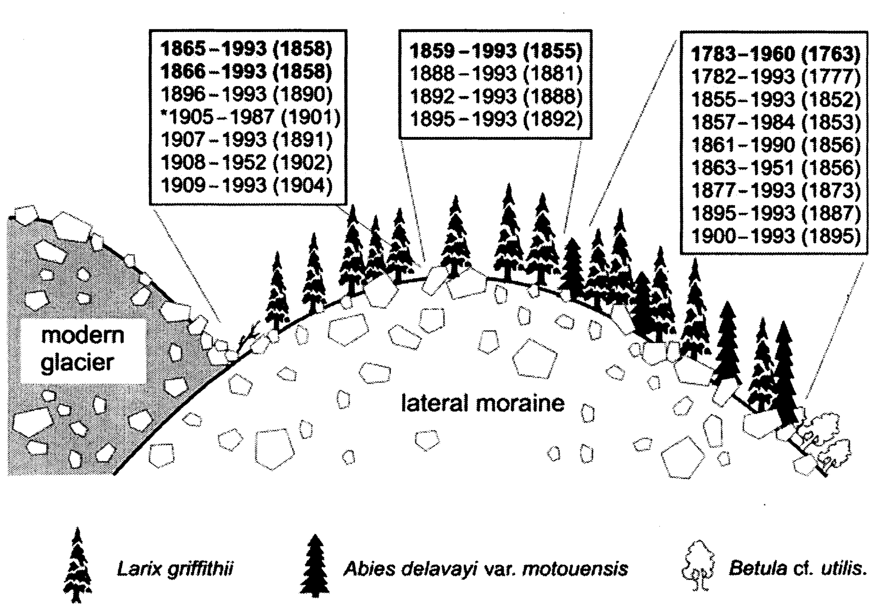

Figure 6 Cross-section through the lateral moraine of Gyalaperi glacier with tree ages. Bold numbers give the dates of the oldest trees in each section

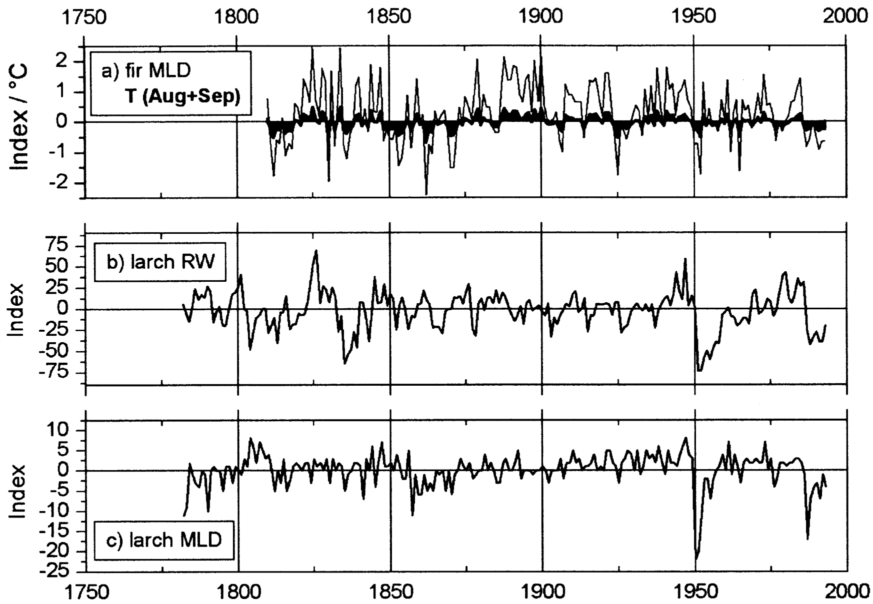

Figure 7 (a) MLD chronology of Gyalaperi fir and derived summer temperature reconstruction (b) Gyalaperi larch RW index chronology (c) Gyalaperi larch MLD chronology

composed of an old core, probably of Lateglacial or Neoglacial age on which the large blocks were deposited during younger glacier advances (B. Frenzel, personal communication, 2003).

Figure 9 shows living ages and growth changes of larch 1 to 3 as raw measured values in the upper part and as detrended index curves in the lower part of the diagram. Since the innermost rings of larch 2 were not preserved, the date of germination is unknown. The extreme growth reductions between 1580 and 1590 are more pronounced on one side of the tree trunk, where partly missing rings occur. Larch 1 dates back to 1264, thus providing a minimum age for the moraine. Beside the 1580–1590 growth depression, the tree shows periods of extremely narrow rings from 1780 to 1790 and after 1865, when missing rings are numerous. Extreme growth reductions and corresponding wood anatomical features were often related to direct contact or extreme nearness of a glacier (Lawrence, 1950; Sharp, 1958; Oeschger and Röthlisberger, 1961; Bray and Struik, 1963; Holzhauser, 1984, 1985; Luckman, 1988; Villalba *et al.*, 1990; Heikkinen, 1984). On the other hand, distinct growth reductions can also be the result of damage to the root system from geomorphologic activity triggered by glacier movement.

In summary, after a period of undisturbed tree development since at least AD 1250, growth suppression pointing to phases of glacier activity occurred from 1580 to 1590 and from the end of the eighteenth to the beginning of the nineteenth century. A third period of activity from 1860 to the 1880s is indicated by growth suppressions, missing rings and the death of larch 1, by the occurrence of a resin duct ring in larch 2 and by the formation of reaction wood caused by tilting in larch 3 (Figure 9).

Discussion

At all investigated sites, the glacier advances during the LIA did not exceed those of the so-called 'Neoglacial' period, which occurred between 4000–3000 BP and 1500 BP in Tibet (Wang

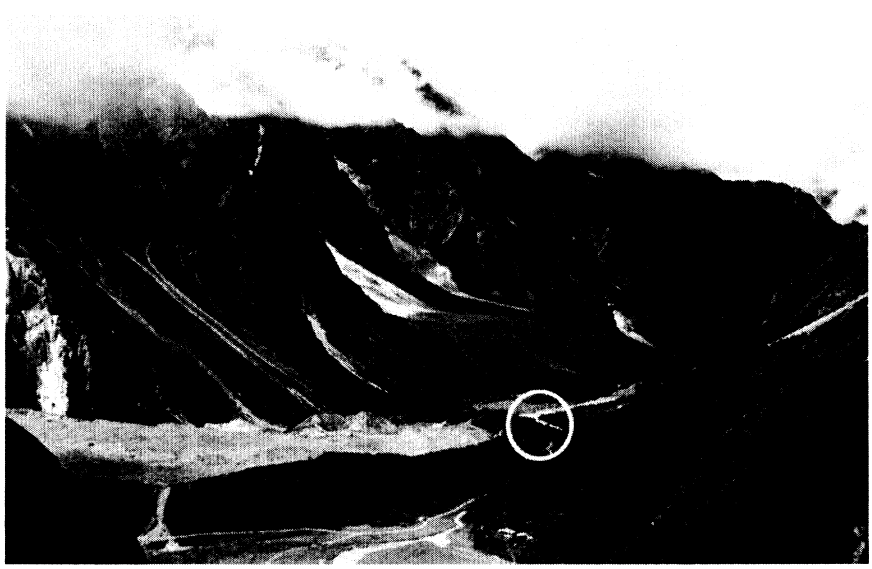

Figure 8 Glacier south of Bomi with location of the tree-ring site (circle)

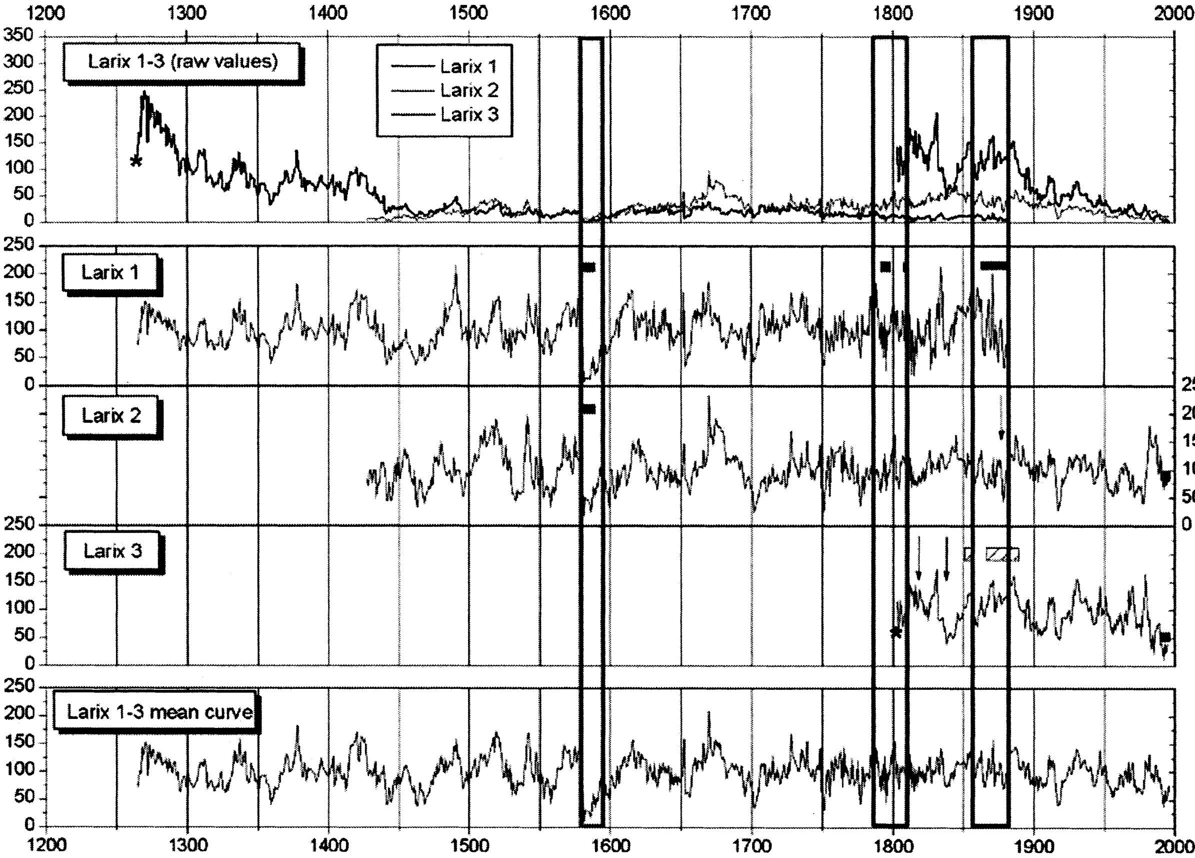

Figure 9 Raw values (upper part) and indices (middle part) of ring width of Larch 1–3 in a glacier forefield south of Bomi. The lower part shows the mean curve of the three trees. Black bars mark periods with missing rings, hatched bars mark the occurrence of compression wood. Black squares at the end of the ring-width curve indicate that the terminal ring next to the cambium was present; an asterisk symbolizes the presence of the pith. Periods of deduced glacier advances are symbolized by bold vertical frames

and Fan, 1987; Lin and Wu, 1984; Zhou *et al.*, 1991; Lehmkuhl, 1995). Although each glacier reacts individually to climatic fluctuations, the three investigated areas show remarkable similarities in their younger glacial history. The maximum glacier advances during the LIA derived from the minimum ages of the corresponding moraines date roughly from before 1770 at the site Lhamcoka and approximately 1760 at the site Gyalaperi. At site Bomi, a glacier advance at around 1780 can be detected. Furthermore, growth suppressions and missing rings in larch trees give hints of an earlier advance around 1580–1590. Since trees of this age were not found at the other study sites, the glacier advance of 1760–1780 seems to be the largest LIA advance at Lhamcoka and Gyalaperi. Recessional stages of standstill or short readvances can be dated at Lhamcoka to the beginning of the nineteenth century (between 1807 and 1820) and the beginning of the twentieth century (older than 1907 to 1920).

Mayewski *et al.* (1980) point out that the general tendency of the majority of Tibetan glaciers is a more or less steady retreat from the maximum position of the LIA, although it is well known that even neighbouring glaciers in High Asia can vary strongly concerning phases of retreat and advance depending on topographic conditions, exposure to moisture-bringing winds and mass balance. Thus, changes in the terminal position of individual glaciers do not necessarily occur synchronously with climate change. According to old reports of western travellers in the western Himalaya, several large glaciers had been retreating long before 1850 because of the appearance of prominent lateral moraines (Kick, 1985), which points to an earlier maximum ice extent and is in agreement with our findings in eastern Tibet. After the maximum stage of the LIA, glacier retreat was interrupted by periods of readvances or standstill stages in the 1850s, 1910s, 1930s and 1970s (Zheng, 1989). From 1890 to 1910, glaciers in the northern Karakorum, Batura-Mustagh and Rakaposhi-Haramosh west of the study area were advancing as a consequence of a period of strengthened Indian summer monsoon activity (Mayewski *et al.*, 1980). During 1974–1984, 90% of small glaciers with a length less than 10 km were readvancing, such as glaciers around Mt Xixabangma and west of Yamdrok Tso in southern Tibet (Wang, 1988). At site Gyalaperi, two glacier advances after 1951 and 1987 can be detected. However, owing to the special topographic conditions in the cirque of Mt Gyalaperi that lead to nourishment of the glacier by avalanches, the glacier tongue might be sensitive to local events rather than to regional climate changes. It should be borne in mind that not in every case might a reaction of a glacier to climate change be reflected by a retreat or advance of the glacier snout. The thick debris cover of Gyalaperi glacier shelters the ice from direct insolation, so a change in the ice budget of a glacier could be reflected by thinning rather than by recession (Wang, 1988).

Wang and Fan (1987) report from lateral moraines of Arza glacier (29°10′N/96°04′E, southeast of site Bomi) dated by minimum ages of trees on the moraines, growth depressions of trees growing in the vicinity of the moraines and fossil logs dated by ^{14}C. They distinguish moraine stages before 1960, before 1931 (probably 1884–1908), 1813–1852 and 1820$\pm$100. Other glacier advances north of Arza glacier are mentioned around 1150$\pm$80 ^{14}C yr BP and 390–290 BP. The latter period could correspond to the glacier advance at Bomi glacier from 1590 to 1599. In the European Alps, glacier advances occurred at this time, too (Nicolussi, 1990; Holzhauser and Zumbühl, 1999). Some of the other phases mentioned can be compared with moraine stages at Lhamcoka (Tables 3 and 4). However, owing to the broad timespan of the dates suggested by Wang and Fan (1987), it is difficult to decide which of the younger stages at Arza and Lhamcoka correspond with each other. Also in the Tien Shan (Russia), dendrochronological evidence suggests glacier advances in the late sixteenth, late seventeenth, late seventeenth to early eighteenth, late eighteenth and first half of the nineteenth centuries (Serebryanny and Solomina, 1989; Kotlyakov *et al.*, 1991).

The context between lowered RW or MLD-indices and glacier activity has been confirmed by many dendroglaciological studies in mountain regions of the Northern Hemisphere (eg, LaMarche and Fritts, 1971; Bircher, 1982; Renner, 1982; Scuderi, 1987; Furrer and Holzhauser, 1989; Kotlyakov *et al.*, 1991; Luckman, 1993; Nicolussi and Patzelt, 1996). The correlation of periods with reduced tree growth and glacier advances is also dependent on the position of the glacier at the beginning of the climatic fluctuation: after a longer period of cool and moist conditions, a glacier might react much faster on a further short climate deterioration than after a period of strong recession during a warm phase (Furrer and Holzhauser, 1989). Unfortunately, no historic documents about former glacier positions such as in the European Alps (Nicolussi, 1990; Holzhauser and Zumbühl, 1999), in Scandinavia (Karlén and Denton, 1976), in the Andes (Villalba *et al.*, 1990) or in North America (Carrara and McGimsey, 1981; Heikkinen, 1984) are available from the Tibetan plateau. $\delta^{18}O$ ratios from Tibetan ice core records indicate cold periods in the middle of the seventeenth century and around 1820 (Yao *et al.*, 1996a) and a recent warming during the twentieth century (Thompson *et al.*, 2000). However, ice cores from different locations over the Tibetan plateau often show no agreement in their isotope variations (Yao *et al.*, 1996b), which might point to different regional sources of precipitation or climate variability. Although the present paper adds new evidence of glacier history in eastern Tibet, the present spatial coverage of high-resolution climate proxy records during the last millennium in Tibet is far too scanty to draw a general picture of glacier variability during the LIA period. This, however, would be highly important for a reliable estimation of changing environments in high mountain ecosystems and a critical evaluation of scenarios of increasing temperatures and summer monsoon intensity predicted by climate circulation models (Meehl and Washington, 1993; Uchijima and Ohta, 1996; Böhner and Lehmkuhl, 2005).

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