Dramatic loss of glacier accumulation area on the Tibetan Plateau revealed by ice core tritium and mercury records

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Abstract

Two ice cores were retrieved from high elevations (~5800 m a.s.l.) at Mt. Nyainqentanglha and Mt. Geladaindong in the southern and central Tibetan Plateau region. The combined tracer analysis of tritium (^{3}H), ^{210}Pb and mercury, along with other chemical records, provided multiple lines of evidence supporting that the two coring sites had not received net ice accumulation since at least the 1950s and 1980s, respectively. These results implied an annual ice loss rate of more than several hundred millimeter water equivalent over the past 30-60 years. Both mass balance modeling at the sites and in situ data from the nearby glaciers confirmed a continuously negative mass balance (or mass loss) in the region due to the dramatic warming in the last decades. Along with a recent report on Naimona'nyi Glacier in the Himalaya, the findings suggest that the loss of accumulation area of glacier is a possibility from the southern to central Tibetan Plateau at the high elevations probably up to about 5800 m a.s.l. This mass loss raises concerns over the rapid rate of glacier ice loss and associated changes in surface glacier runoff, water availability, and sea levels.

1. Introduction

Data from remote sensing and in situ observations suggest that glacier shrinking has been prevailing over the Tibetan Plateau (including the Himalaya hereafter) in the past decades (e.g., Liu et al., 2006; Kang et al., 2010; Fujita and Nuimura, 2011; Bolch et al., 2012; Kääb et al., 2012; Yao et al., 2012; Neckel et al., 2014), raising major concerns over their impact on water supplies to some 1.4 billion people in Asia (Immerzeel et al., 2010), and on global sea level rise (Jacob et al., 2012; Gardner et al., 2013; Neckel et al., 2014). It has been estimated that glacier retreat has been occurring to more than 82% of the total glaciers in the region (Liu et al., 2006), and thus since the 1970s glacier areas have reduced by several percent (about 4.8%) in the central Tibetan Plateau (Ye et al., 2006) and up to 20% in the northeastern marginal regions of the Tibetan Plateau (Cao et al., 2010; Pan et al., 2012). In situ stake observations have also confirmed a continuously negative mass balance during the last decade in the region (Yao et al., 2012; Zhang et al., 2014). However, quantitative changes in the glacier ice volume, a key parameter for assessing retreating glaciers' impact on water supply or sea level rise, remain poorly known due to the lack of in situ measurements on glacier thickness through time. Although remote sensing techniques have provided some assessments in glacier thickness globally, especially in the last decade, the application of those techniques to the Tibetan Plateau region is rather limited due to complexity of the regional topography (Jacob et al., 2012; Kääb et al., 2012; Gardner et al., 2013; Neckel et al., 2014).

Based on the lack of distinctive marker horizons of atmospheric thermonuclear bomb testing (e.g., beta radioactivity, ^{36}Cl, and tritium (^{3}H)) in an ice core retrieved from Naimona'nyi (6050 m a.s.l.) in the Himalaya, a recent study suggests that there might not have been a net accumulation of glacier mass at the site since at least the 1950s (Kehrwald et al., 2008). This thinning could be very significant considering that the mass loss occurred at the upper part of the glacier where it is normally considered as the accumulation area (Shi et al., 2005). To test whether such loss of glacier accumulation area is occurring at high elevations over the Tibetan Plateau, here we report the two ice core records taken from high elevations (~5800 m a.s.l.) at Mt. Nyainqentanglha and Mt. Geladaindong in the Tibetan Plateau. In addition to radioisotopes ^{3}H and ^{210}Pb and other geochemical tracers, the depth profile of mercury (Hg) is used as a new marker for the last century based on known atmospheric depositional histories.

2. Methodology

With an average elevation of over 4000 m a.s.l., the Tibetan Plateau is home to the largest volume of glacier ice outside the polar regions (Grinsted, 2013). The Tibetan Plateau blocks mid-latitude Westerlies, splitting the jet into two currents that flow the south and north of the plateau. The plateau is also a major forcing factor on the intensity of the Asian monsoons. The southern and central Tibetan Plateau is climatically influenced primarily by the Indian monsoon during the summer monsoon season and the Westerlies during the non-monsoon season (Bryson, 1986; Tang, 1998).

Two ice cores were retrieved as part of the Sino-US Cooperation Expedition (Fig. 1). The Nyainqentanglha ice core (30°24.59′N, 90°34.29′E, 5850 m a.s.l.), by drilling to the bedrock depth of 124 m, was collected in September of 2003 from the Lanong glacier pass on the eastern saddle of Mt. Nyainqentanglha (peak height: 7162 m a.s.l.) in the southern Tibetan Plateau. The Geladaindong core (33°34.60′N, 91°10.76′E, 5750 m a.s.l.), 147 m in length (did not reach to the bedrock), was collected in October 2005 from the Guoqu glacier on the northern slope of Mt. Geladaindong (peak height: 6621 m a.s.l.), which is the summit of the Tanggula Mts. in the central Tibetan Plateau and the headwater region of the Yangtze river. Elevations of both ice coring sites are higher than the snow line altitudes (close to the equilibrium line altitudes (ELAs)) of around 5700 m a.s.l. in the Mt. Nyainqentanglha region (Shi et al., 2005) and 5570 m. a.s.l. in the Mt. Geladaindong region (Zhang, 1981). However, these ELAs were retrieved from the glacier area data from several decades ago (e.g., data in 1980s and 1970s for the Mt. Nyainqentanglha and Mt. Geladaindong region respectively), and may not reflect present-day ELAs. Snowpits, with a depth of 40 cm and 78 cm at the Nyainqentanglha and Geladaindong coring sites, respectively, were also sampled at a 10-cm depth interval.

The Nyainqentanglha ice core was transported in a frozen state to the Climate Change Institute at the University of Maine, USA, whereas the Geladaindong ice core to the State Key Laboratory of Cryospheric Sciences of the Chinese Academy of Sciences, Lanzhou, China. The cores were sectioned at 3 to 5 cm intervals in a cold (-20 ℃) room, with the outer sections (approximately 1 cm) being scraped off using a pre-cleaned ceramic knife. The inner sections were placed into whirl-pak bags. After being melted at room temperature, the water samples were collected into HDPE vials for subsequent analyses. Ice chips from the outer sections of the cores were collected at an interval of 1 m from the upper 40 m of the cores for the analysis of ^{210}Pb.

All of the samples were measured for $\delta^{18}O$ on a MAT-253 isotope mass spectrometer (±0.1‰ precision) via the standard CO_2 equilibration technique at the Key Laboratory of Tibetan Environment Changes and Land Surface Processes, Institute of

Tibetan Plateau Research, Chinese Academy Sciences, Beijing. Soluble major ions (Na^{+}, K^{+}, Mg^{2+}, Ca^{2+}, SO_4^{2-}, Cl^{-}, and NO_3^{-}) were measured by ion chromatography (Dionex DX-500), and elemental analysis (e.g., Bi, Fe, Al) was done by inductively coupled plasma sector field mass spectrometry at the Climate Change Institute (Kaspari et al., 2009).

Total Hg concentration was analyzed following U.S. EPA Method 1631 using a Tekran® 2600 at the Ultra-Clean Trace Elements Laboratory at the University of Manitoba, Canada, or Jena® MERCUR in a metal-free Class 100 laminar flow hood placed in a Class 1000 cleanroom laboratory at the Key Laboratory of Tibetan Environment Changes and Land Surface Processes. Field blank samples were collected during each sampling and their Hg concentrations were always lower than 0.3 ng L^{-1}. Certified reference materials ORMS-2 and ORMS-3 (National Research Council of Canada) were used for QA/QC, and the recoveries were within 5% of their certified values. To further ensure the data quality, samples were measured in both labs and agreed within 15% of each other (Loewen et al., 2007; Zhang et al., 2012).

The ^{3}H was measured by using Quantulus Low-level Liquid Scintillation Counters (Morgenstern and Taylor, 2009) at the Institute of Geological and Nuclear Science, National Isotope Centre, New Zealand. The ^{210}Pb activity was indirectly analyzed by measuring α decay of ^{210}Po at an energy of 5.3 MeV using alpha-spectrometry at Paul Scherrer Institut, Switzerland (Gäggeler et al., 1983).

3. Results and Discussion

3.1. Ice Core Chronology

Some of the most prominent global stratigraphic markers recorded in ice cores over the last century are radionuclides (e.g., ^{3}H) released during nuclear bomb tests (Kotzer et al., 2000; Pinglot et al., 2003; Kehrwald et al., 2008; Van Der Wel et al., 2011). Large amounts of ^{3}H, in increasing amounts, were released into the atmosphere by above-ground thermonuclear bomb tests during 1952-1963 AD, resulting in atmospheric levels several orders of magnitude above natural cosmogenic concentrations (Clark and Fritz, 1997). Remarkably, this global ^{3}H marker is not present in the Nyainqentanglha ice core (Fig. 2). All samples collected from the Nyainqentanglha core had ^{3}H activity below the detection limit of 0.1 TU with the exceptions of two samples, one at the surface (13.4 TU), and the other at a depth of 31 m (29.1 TU). The slightly elevated ^{3}H concentration at 31 m of this core cannot represent the 1963 AD nuclear bomb horizon for several reasons. If this sample would represent the 1963 AD bomb horizon, the ^{3}H concentration would have been much higher (> 200 TU), and there would have been broader ^{3}H spikes in that depth region as the atmospheric nuclear bomb testing occurred over more than a decade. This is clearly not the case (Fig. 2). The absence of anthropogenic ^{3}H markers below surface samples has recently been reported in an ice cores taken from the Naimona'nyi Glacier in the central Himalayas (Kehrwald et al., 2008). Therefore, similar to the Naimona'nyi Glacier, the Nyainqentanglha site might not have received net ice accumulation since at least the 1950s AD.

To further test this hypothesis, we analyzed ^{210}Pb activity in the Nyainqentanglha ice core. Radioactive decay of ^{210}Pb, a product of natural ^{238}U decay series with a half-life of 22.3 yr, has been successfully applied to ice core dating on a century time-scale (Gäggeler et al., 1983; Olivier et al., 2006). The ^{210}Pb activity was 940.5 mBq kg^{-1} at the topmost sampling layer at a depth of 0-0.9 m; however, it decreased sharply to near the background level (7.6 mBq kg^{-1}) in the next sampling layer at 5.6 m depth (Fig. 3a). High ^{210}Pb activity in the upper layer indicates enrichment due to the negative mass balance. Based on the extremely low ^{3}H and ^{210}Pb activities immediately beneath the surface layer, we conclude that there has been no net ice accumulation at the Nyainqentanglha site since at least 1950s AD.

In contrast, the Geladaindong ice core exhibited a classic ^{3}H profile, with a sharp spike of up to 680 TU at a depth of 5.22-6.23 m (Fig. 2), suggesting that the ice accumulated at this site during the 1963 AD thermonuclear bomb testing era is still present. We assign the 5.74 m depth, which shows the highest ^{3}H level, to the year 1963 AD when above-ground nuclear tests peaked just prior to the Nuclear Test Ban Treaty (Van Der Wel et al., 2011). Based on this ^{3}H bomb test horizon, we establish the chronology of the Geladaindong ice core by counting annual layers according to the seasonal cycles of $\delta^{18}O$, major ions, and elemental concentrations upward to the top of the core (Fig. 4). The uppermost ice layer is designated as 1982 AD based on annual layer counting above the 1963 AD ^{3}H marker, and is further constrained by ^{210}Pb estimation of 1982 $\pm$5 years for the surface of the core (Fig. 3b).

The Geladaindong ice core dating by counting annual layers is in agreement with previous work in the region. Based on snowpit records in the Guoqu glacier, Mt.

Geladaindong, Zhang et al. (2007) reported that Ca^{2+} and other major ions in snowpits varied seasonally with higher values during the winter half year. At our coring site, there were firn layers (snowpit) with a depth of 78 cm. The bottom of the snowpit was glacier superimposed ice, indicating one year transferring from snow to ice. Thus, melt could happen during the summer but seasonal signals were still preserved in ice layers. In other words, the melt water (or percolation) should not disturb the layers deposited in previous years as suggested in other studies (Namazawa and Fujita, 2006; Eichler et al., 2001).

One assumption in dating by counting annual layers backwards from the 1963 AD nuclear bomb horizon to 1982 AD is that there was annual net ice accumulation during this period. Uncertainties in the chronology will thus rise should there be no net accumulation in one or some of the years due to ice melt. This does not appear to be the case for the Geladaindong ice core, as the annual variation patterns and amplitudes in the main ion concentrations were similar upward and downward from the 1963 AD layer, suggesting no occurrence of strong melt (Fig. 4). Furthermore, the air temperatures were much lower before 1980s than those in the last three decades according to the data observed from the nearby meteorological stations such as Amdo. Indeed, the continuously deficit mass balance (cumulative negative mass balance) has only been reported since the 1990s in the central Tibetan Plateau (e.g., Xiaodongkemadi glacier, near to the Geladaindong region; Yao et al. 2012), as well as in the northern neighboring region (e.g., Glacier No. 1, Tienshan Mts.; Zhang and others, 2014), due to dramatic warming in recent decades. Therefore, we might suggest that the mass loss of the coring site occurred mainly from the 1990s in the central Tibetan Plateau.

To further investigate whether the lack of net ice accumulation at the Geladaindong site occurred since the 1980s, we examined the profile of Hg in the ice core. Although naturally occurring in the Earth's crust, Hg emission (especially the gaseous elemental mercury) into the atmosphere has been greatly enhanced coinciding with the rise in anthropogenic activities (e.g., mining, burning of fossil fuels). Mercury has a lifetime of approximately 0.5-2 years (Holmes et al. 2010) and can be transported globally via atmospheric circulation. Hg profiles in ice cores from high (Faïn, et al. 2008) and mid-latitude (Schuster et al., 2002) regions have matched the general chronological trends of global atmospheric Hg emissions or global industrial Hg use. As atmospheric transport is essentially the only transport pathway for anthropogenic Hg to the Tibetan Plateau, due to the region's high altitudes and minimal to nonexistent local industrial activities (Loewen et al., 2007), ice cores from the region could provide a useful indicator for atmospheric Hg concentrations, as demonstrated by Hg profiles in snowpacks overlying the glaciers across the plateau (Loewen et al., 2007; Zhang et al., 2012).

As shown in Fig. 5, the Hg profile in the Geladaindong ice core, with the upper-most layer dated to around 1982 AD, matches the atmospheric Hg depositional chronology established from sediment records in Nam Co (Li, 2011), a large alpine lake (4710 m a.s.l.) on the Tibetan Plateau, as well as the history of regional and global Hg production (Hylander and Goodsite, 2006), showing low and stable background levels prior to ~1850 AD, with a steady concentration increase from the mid-20th century to the 1980s.

Beyond the 1980s, the Nam Co sediment record shows a decline in Hg concentrations, which matches the global and regional emission trends. Such declining trends are absent in the Geladaindong ice core, supporting that this site has not received net ice accumulation since the 1980s. There are some secondary timing differences of the Hg trend between the lake sediment and the ice core (e.g., during the 1970s), which might be attributed to the lower resolution of the lake sediment record (about 5 yrs) compared to that of the ice core record (1 yr).

The lack of recent deposition of mass (ice) at the Nyainqentanglha and Geladaindong glaciers, as well as at the Naimona'nyi glacier (Kehrwald et al., 2008), suggests that the melting and/or loss of the accumulation area of glacier occurred in at least these three ice coring regions of the Tibetan Plateau. Although there is a consensus that glaciers in the Tibetan Plateau are largely retreating (Yao et al., 2004, 2012; Bolch et al., 2012; Neckel et al., 2014), ^{3}H and Hg records reported herein provide direct evidence of dramatic thinning occurring at the upper regions of glaciers (probably up to about 5800 m a.s.l.) that had traditionally been considered as net accumulation areas (Zhang et al., 1981; Shi et al., 2005).

3.2. Observed and Modeled Mass Balance

Due to a lack of precipitation data at the coring sites, we cannot directly quantify the annual ice loss in these high-altitude glaciers. The annual precipitation data from local lower elevation meteorological stations are 444 mm at Damxung (50 km southeast of Nyainqentanglha but at an elevation of 4300 m a.s.l.) and 467 mm at Amdo (120 km south of Geladaindong but at an elevation of 4800 m a.s.l.) (Fig. 1). These data suggest that the glaciers have experienced a net loss of at least several hundred millimeters each year (mm w.e. yr^{-1}). The estimate is considered as the lower limit as glacier areas in high mountainous regions generally receive more precipitation (accumulation) than at lower elevation stations (Shen and Liang, 2004; Wang et al., 2009; Liu et al., 2011). Although no observational data are available for the central Tibetan Plateau region, precipitation has been shown to increase 0.87 to 11 mm with every 100 m increase in elevation in the neighboring Qilian and Tienshan Mts. (Liu et al., 2011).

In situ observational data using mass balance stakes close to our coring sites are available only for a short time period in the recent past (Kang et al., 2009; Yao et al., 2012; Qu et al., 2014). Mass balance measurements of Xiaodongkemadi glacier (80 km south of Mt. Geladaindong, Fig. 1), started in 1989, showed slightly positive mass accumulation until the mid-1990s, then changed to a net mass loss over time (Yao et al., 2012) (Fig. 6). During the period 1995-2010 AD, the cumulative mass loss reached 5000 mm with an annual mass loss rate of about 300 mm w.e. A much higher mass loss rate was observed in situ at Zhadang glacier (5 km east of the Mt. Nyainqentanglha, Fig. 1) in the southern Tibetan Plateau; over the period 2005-2011 AD mass loss rate at this glacier averaged approximately at1200 mm w.e. yr^{-1} (Qu et al., 2014). More recently, Neckel et al. (2014) reported the glacier mass changes during 2003-2009 AD for the eight sub-regions in the Tibetan Plateau using ICESat laser altimetry measurements. These authors estimated that a regional average mass balance of -580$\pm$ 310 mm w.e. yr^{-1} was observed in the central Tibetan Plateau sub-region covering the

Geladaindong coring site. The mass balance of glaciers varies due to different measurements and time periods. Over the entire Tibetan Plateau, in situ observed glacier mass balances ranged from -400 mm w.e. yr^{-1} to -1100 mm w.e. yr^{-1} during the last decade with an exception of slight mass gain in the northwestern of the Tibetan Plateau (e.g. western Kunlun Mts. and Karakoram regions) (Yao et al., 2012; Bolch et al., 2012; Gardelle et al., 2012; Neckel et al., 2014). The clear deficit mass balances of Zhadang and Xiaodongkemadi glaciers are consistent with our findings from the two ice core records.

In order to further assess whether the intensive melting could happen in the high elevations of the coring sites, a degree-day model (DDMs) was applied to estimate glacier melt at the two ice core sites. DDMs can determine the daily quantity of snow/ice melt (m_t, mm w.e.) as a function of the mean daily air temperature (Tt, ℃) using a factor of proportionality referred to the degree-day factor (DDF, mm ℃$^{-1}$ d^{-1}) (Gardner and Sharp, 2009).

$$m_t = \mathrm{DDF} \times T_t \qquad T_t \geq 0$$

$$m_t = 0 \qquad T_t < 0$$

To detect the net mass balance at the Nyainqentanglha and Geladaindong coring sites by DDMs, we selected daily temperature and precipitation data from the two meteorological stations, Damxung and Amdo, which are the nearest stations to the Nyainqentanglha and Geladaindong sites (Fig. 1), respectively. Daily temperature and positive cumulative temperature at the two sites were calculated based on the minimum (0.5 ℃/100 m) and maximum (0.72 ℃/100 m) temperature lapse rate reported by Li and Xie (2006) and Yang et al.(2011), respectively, for the Tibetan Plateau (Tables 1 and 2, Fig. 7). The medium value was set as the global average of 0.6 ℃/100 m. The accumulation rate at each coring site was considered the same as the precipitation amount at the stations nearby, although more precipitation is likely to occur at the higher elevations as discussed before. Due to the differences in the surface energy-balance characteristics of snow and ice (including albedo, shortwave penetration, thermal conductivity and surface roughness), reported DDFs vary greatly among regions and times. Based on previous work in the southern and central Tibetan Plateau (Wu et al., 2010; Zhang et al., 2006), we selected DDF values of 3.0 (minimum for snow), 5.3 (medium for snow), 9.2 (medium for ice) and 14 (maximum for ice) mm ℃$^{-1}$ d^{-1}, respectively (Tables 1 and 2).

As shown in Fig. 7, there is a statistically significant ($p<0.01$) increase trend in annual positive cumulative temperatures at the Geladaindong ($r=0.5$) and Nyainqentanglha ($r=0.6$) coring sites during 1966-2013 AD, but not in precipitation. DDM modeling shows clear decrease trends in the cumulative net mass balance at both sites under most of the scenarios except when the maximum of temperature lapse rate and the minimum DDF were applied (Fig. 7). The calculated averaged annual net mass balance (medium) during 1966-2013 AD was -925±576 mm w.e. yr^{-1} (range from 132 ±157 to -3441±944 mm w.e. yr^{-1}) at the Geladaindong coring site (Table 1) and -671 ±538 mm w.e. yr^{-1} (range from 336±150 to -3912±992 mm w.e. yr^{-1}) at the Nyaingentanglha site (Table 2). Since there is likely more precipitation (Shen and Liang, 2004) in the glacier regions, the actual mass loss rates at the two sites might be slightly less than these estimated values. Nevertheless, the mostly deficit mass balances suggest that mass loss most likely occurred at both coring sites during the last decades, which is consistent with the two ice core records. Furthermore, in situ observed mass balance of the central Tibetan Plateau (e.g., the Xiaodongkemadi glacier) shows a continuous deficit mass balance (or cumulative negative mass balance) since the 1990s (Fig. 6) (Yao et al., 2012), which agrees with a dramatic warming as shown in Fig.7 (positive cumulative temperature) in the same period.

4. Conclusion

Meteorological data suggest dramatic warming has occurred in the Tibetan Plateau since the late 1980s and that the magnitude of warming is much greater than that in the low-elevation regions (Kang et al., 2010). This warming has resulted in a continuous negative mass balance (or mass loss) of glaciers during the last decade ranging from Himalayas to the north of the Tibetan Plateau except for the northwestern Tibetan Plateau (e.g., Yao et al., 2012; Bolch et al., 2012; Gardelle et al., 2012; Neckel et al., 2014). In recent years, the altitude of the equilibrium line for some of the observed glaciers has risen beyond the highest elevations of the glaciers; that is, there is no more net accumulation area and subsequently the entire glacier is becoming ablation area (Yao et al., 2012). Although glacier mass balance varies depending on climate change and geographical conditions as shown on the Tibetan Plateau (e.g. Yao et al., 2012; Bolch et al., 2012), our ^3H and Hg ice core records confirm that the upper glacier areas (e.g. about 5750-6000 m a.s.l.) are rapidly transforming into ablation areas in recent decades. In particular, extensive ablation has caused substantial mass loss of the Nyainqentanglha and Geladaindong glaciers since at least the 1950s in the southern part and the 1980s in the central part of the Tibetan Plateau, respectively.

We suggest that the glaciers on the southern to the central Tibetan Plateau might be melting faster than previous data show (Liu et al., 2006; Jacob et al., 2012; Gardner et al., 2013). Ice losses on such a large scale and at such a fast rate could have substantial impacts on regional hydrology and water availability (Immerzeel et al., 2010), as well as causing possible floods due to glacier lake outbursts (Richardson and Reynolds, 2000; Zhang et al., 2009). Further, the loss of glacier accumulation area warns us that recent climatic and environmental information archived in the ice cores is threatened and rapidly disappearing in the mid and low latitudes. As such, there is an urgent need to collect and study these valuable ice core records before they are gone forever.

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Tables

Table 1 Calculated annual net mass balance (mm w.e. yr^{-1}) during 1966-2013 AD based on various degree-day factor (DDF) values (mm ℃$^{-1}$ d^{-1}) and temperature lapse rates (℃/100 m) at the Geladaindong ice core site. Negative values represent deficit mass balances.

Table 2 Calculated annual net mass balance (mm w.e. yr^{-1}) during 1966-2013 AD based on various degree-day factor (DDF) values (mm ℃$^{-1}$ d^{-1}) and temperature lapse rates (℃/100 m) at the Nyainqentanglha ice core site. Negative values represent deficit mass balances.

Table 1

DDF[a, b] / Temperature lapse rate	Minimum 3.0 (snow)	Medium 5.3 (snow) 9.2 (ice)	Maximum 14.0 (ice)
Minimum (Tr1)[c] 0.5	-386 ±220	-1025 ±369 -2108 ±625	-3441 ±944
Medium (Tr2) 0.6	-121 ±192	-925 ±576	-2203 ±811
Maximum (Tr3)[d] 0.72	132 ±157	-109 ±247 -518 ±408	-1021 ±610

a: Wu et al., 2010; b: Zhang et al., 2006; c: Li and Xie, 2006; d: Yang et al., 2011.

Table 2

DDF[a, b] / Temperature lapse rate	Minimum 3.0 (snow)	Medium 5.3 (snow) 9.2 (ice)	Maximum 14.0 (ice)
Minimum (Tr1)[c] 0.5	-469 ±249	-1189 ±400 -2410 ±663	-3912 ±992
Medium (Tr2) 0.6	-2.57 ±212	-671 ±538	-1733 ±791
Maximum (Tr3)[d] 0.72	336 ±150	234 ±207 60.4 ±313	-153 ±450

a: Wu et al., 2010; b: Zhang et al., 2006; c: Li and Xie, 2006; d: Yang et al., 2011.

Figure Captions

Figure 1. Location map of the glacier ice cores Geladiandong (GL) and Nyainqengtanglha (NQ) on the Tibetan Plateau. Also shown are the locations of the Naimona'nyi ice core by Kehrwald et al. (2008), Xiaodongkemadi glacier, Zhadang glacier, the meteorological stations and Lake Nam Co.

Figure 2. The tritium profiles of the Geladiandong and Nyainqengtanglha ice cores compared with tritium in the Northern Hemispheric precipitation. Error bars for the ice core samples are shown, but in most cases are only about half of the symbol size. To enable direct comparison, both the Geladaindong and precipitation tritium records are decay-corrected to the date of Geladaindong ice core drilling (October 2005). The record of tritium in precipitation (upper axis) shows the Ottawa precipitation record (International Atomic Energy Agency, 2013, WISER database: http://www-naweb.iaea.org/napc/ih/IHS_resources_isohis. html) between 1982 and 1953. Tritium from before 1953 has now decayed to zero. The time of the Northern Hemispheric precipitation record is scaled to match the maximum tritium concentration in the ice core to mid-1963, and to start with 1982 (date of the surface ice, see text). To match the tritium concentrations in the Geladaindong ice core, the Ottawa precipitation record had to be multiplied by a factor of two. This indicates that the tritium concentration on the Tibetan Plateau is about twice of that of Ottawa, due to a more direct input of stratospheric air, which is the main

atmospheric tritium reservoir.

Figure 3. (a) ^{210}Pb activity profiles of the Geladaindong (GL) and Nyainqentanglha (NQ) ice cores; (b) ^{210}Pb activity versus depth for the GL core. The age-depth relationship was derived from an exponential regression of ^{210}Pb activity against depth. The uppermost two samples (open diamonds) were excluded (enrichment due to melt). This age-depth relation was anchored using the known age and depth of the tritium horizon. Extrapolation of the age fit to the surface allows estimating the surface age (green star). Error bars and the fine grey lines indicate the 1 sigma uncertainty of the given ages.

Figure 4. Dating of the Geladaindong ice core by annual layer counting based on the seasonal cycles of $\delta^{18}O$, Ca^{2+}, Cl^{-} and Fe according to the anchor of 1963 tritium peak (red star) (dashed lines represent the annual boundaries).

Figure 5. Comparisons of Hg records from the Geladaindong (GL) ice cores with those from Lake Nam Co sediments (Li, 2011), as well as with known history of the regional (Asia and USSR) and global Hg production (Hylander and Goodsite, 2006).

Figure 6. In situ observed annual and cumulative mass balance for the Xiaodongkemadi glacier in the Tanggula Mts. (Yao et al., 2012) and the Zhadang glacier in the Nyainqengtanglha Mts. (Qu et al., 2014).

Figure 7. Variations of annual positive cumulative temperature at the Nyainqentanglha and Geladaindong ice core sites, annual precipitation amount at Damxung and Amdo station, and the estimated cumulative net mass balance based on a degree-day model (DDM) at the two ice core sites during 1966-2013 AD. (Tr1, Tr2 and Tr3 as listed in Tables 1 and 2; dashed lines represent the average of the annual positive cumulative temperature before and after 1990 AD).

Figure 1

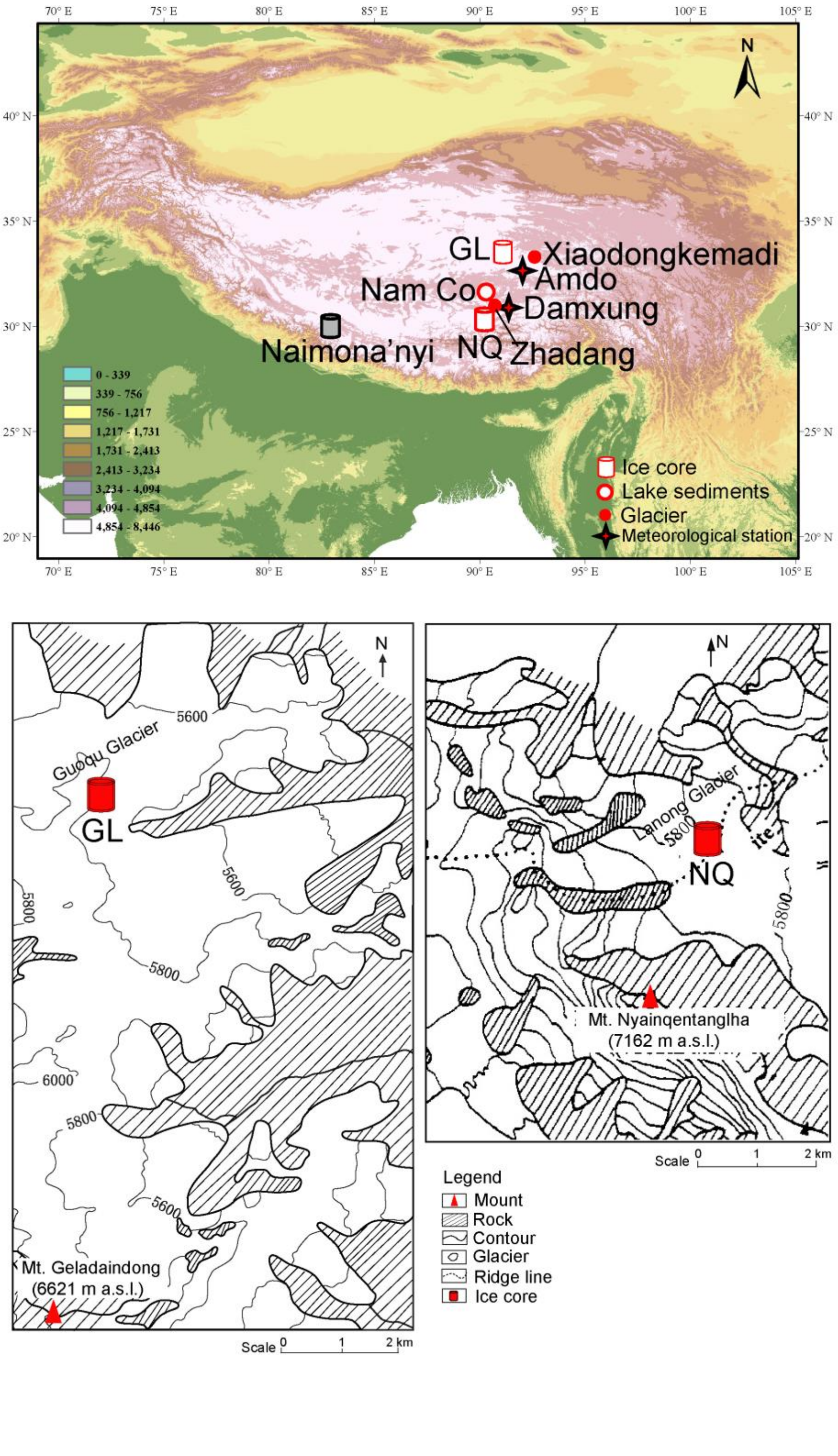

70° E
75° E
80° E
85° E
90° E
95° E
100° E
105° E
40° N
35° N
30° N
25° N
20° N
N
GL
Xiaodongkemadi
Amdo
Nam Co
Damxung
NQ
Zhadang
Naimona'nyi
0 - 339
339 - 756
756 - 1,217
1,217 - 1,731
1,731 - 2,413
2,413 - 3,234
3,234 - 4,094
4,094 - 4,854
4,854 - 8,446
Ice core
Lake sediments
Glacier
Meteorological station
Guoqu Glacier
5600
5800
6000
Mt. Geladaindong
(6621 m a.s.l.)
Scale 0 1 2 km
Lanong Glacier
Mt. Nyainqentanglha
(7162 m a.s.l.)
Legend
Mount
Rock
Contour
Glacier
Ridge line
Ice core

Figure 2

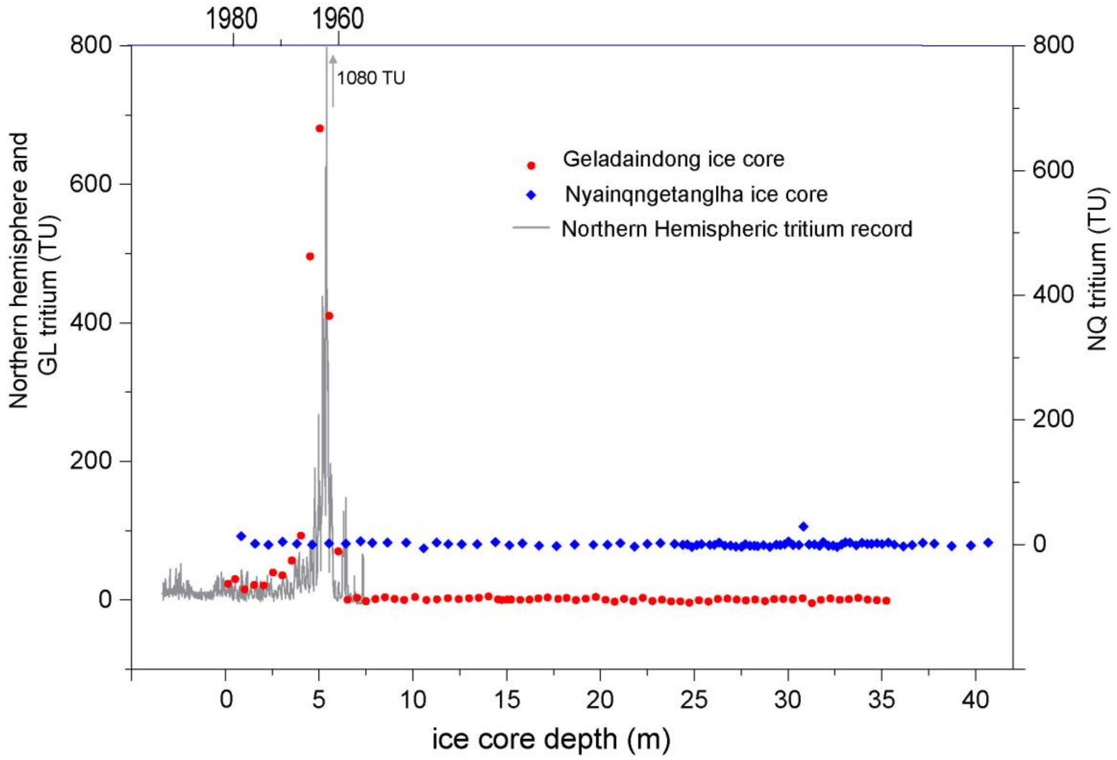
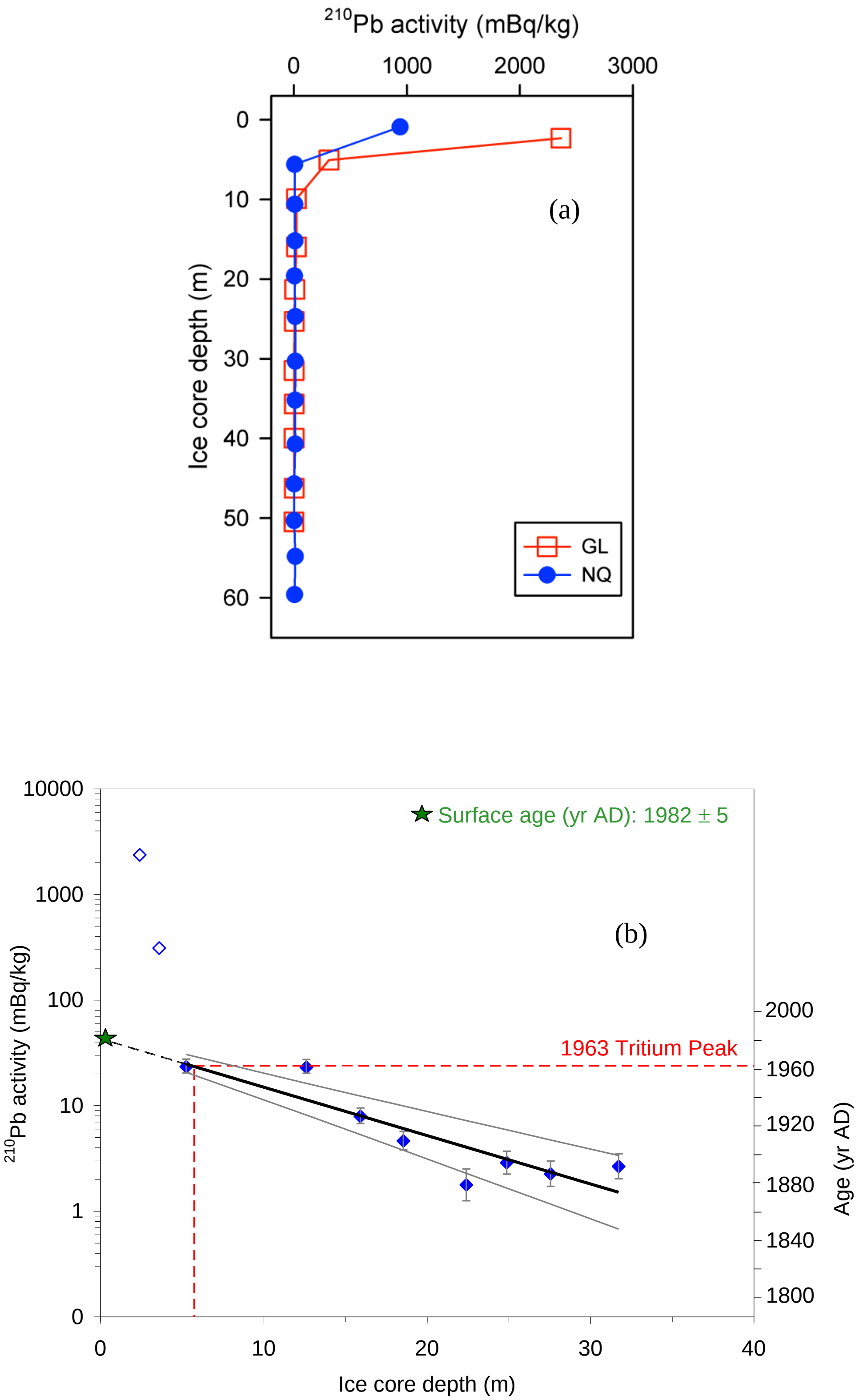
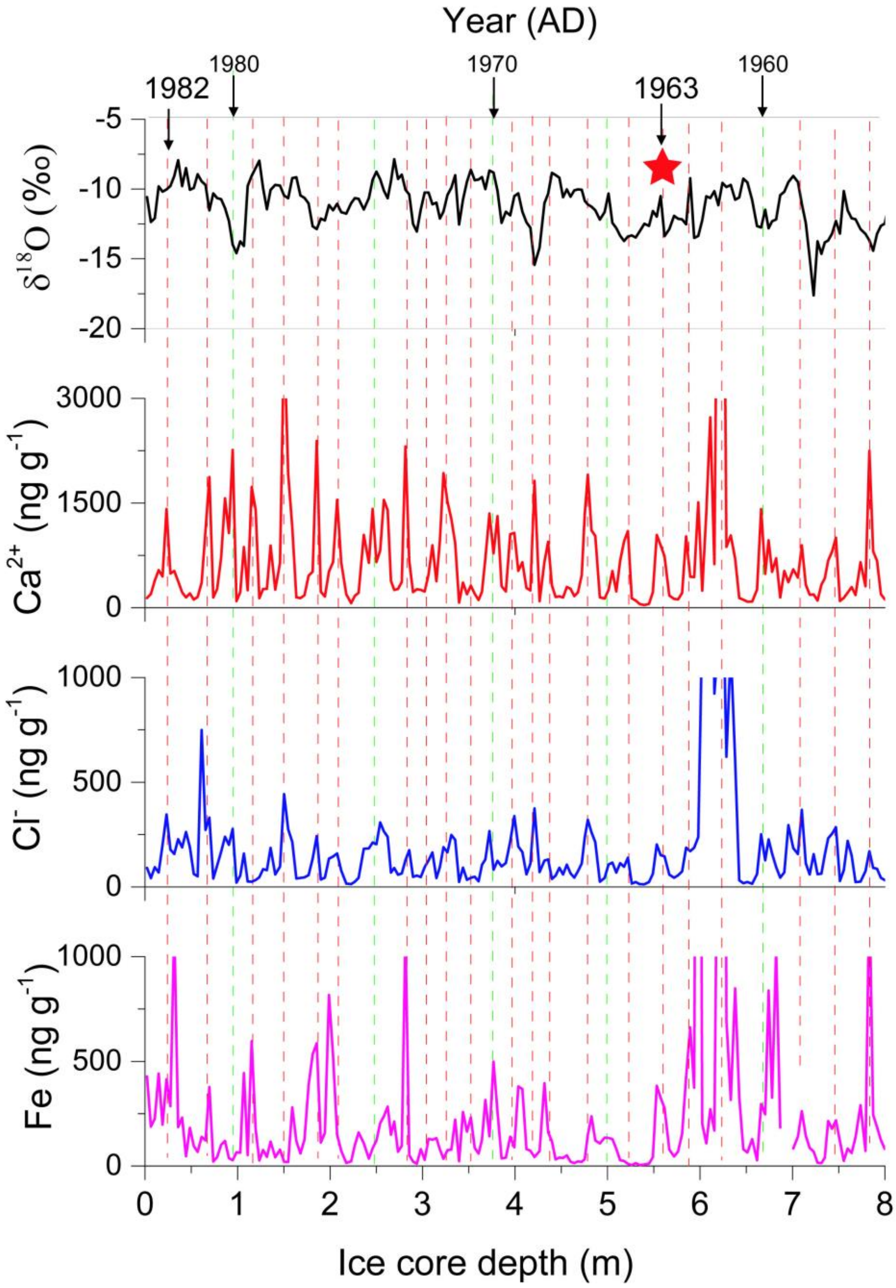

1980
1960
1080 TU
Geladaindong ice core
Nyainqngetanglha ice core
Northern Hemispheric tritium record
Northern hemisphere and GL tritium (TU)
NQ tritium (TU)
ice core depth (m)
800
600
400
200
0
5
10
15
20
25
30
35
40

Figure 3



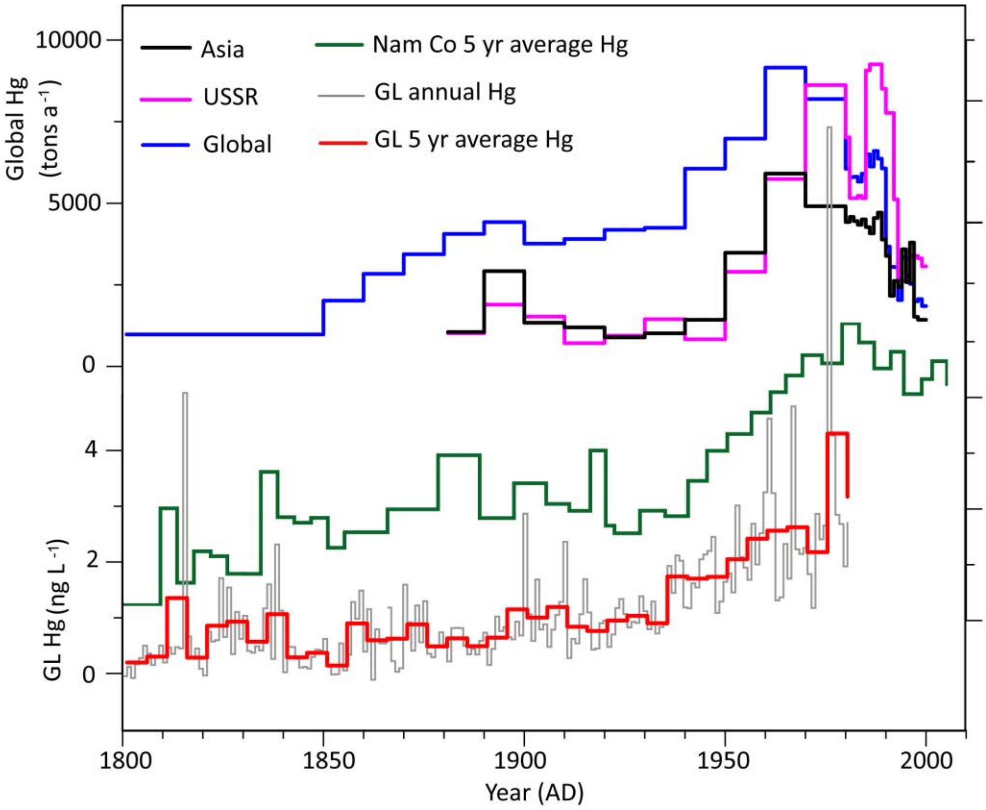
^{210}Pb activity (mBq/kg)
0
1000
2000
3000
0
10
20
30
40
50
60
Ice core depth (m)
(a)
GL
NQ
10000
1000
100
10
1
0
^{210}Pb activity (mBq/kg)
Surface age (yr AD): 1982 ± 5
(b)
1963 Tritium Peak
2000
1960
1920
1880
1840
1800
Age (yr AD)
0
10
20
30
40
Ice core depth (m)

Figure 4



Year (AD)
1982
1980
1970
1963
1960
-5
-10
-15
-20
δ18O (‰)
3000
1500
0
Ca2+ (ng g-1)
1000
500
0
Cl- (ng g-1)
1000
500
0
Fe (ng g-1)
0
1
2
3
4
5
6
7
8
Ice core depth (m)

Figure 5



Asia
USSR
Global
Nam Co 5 yr average Hg
GL annual Hg
GL 5 yr average Hg
Global Hg (tons a-1)
10000
5000
0
Asia & USSR (tons a-1)
2000
1000
Nam Co Hg (ng g-1)
40
30
20
GL Hg (ng L-1)
4
2
0
1800
1850
1900
1950
2000
Year (AD)

Figure 6

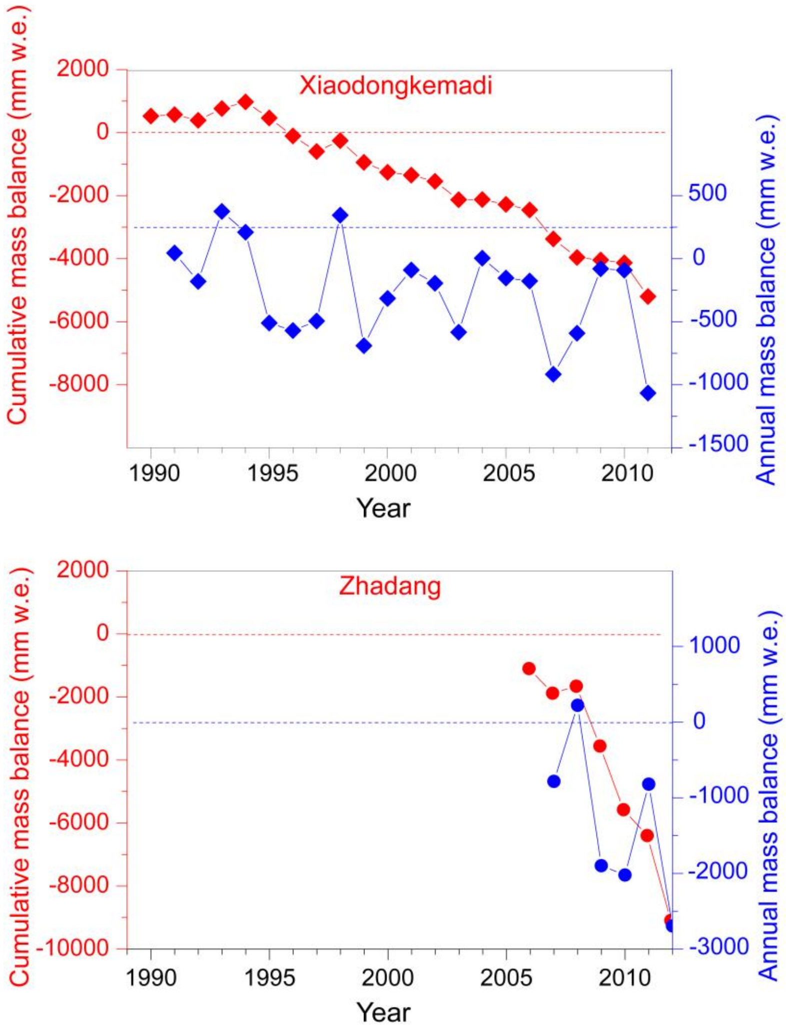
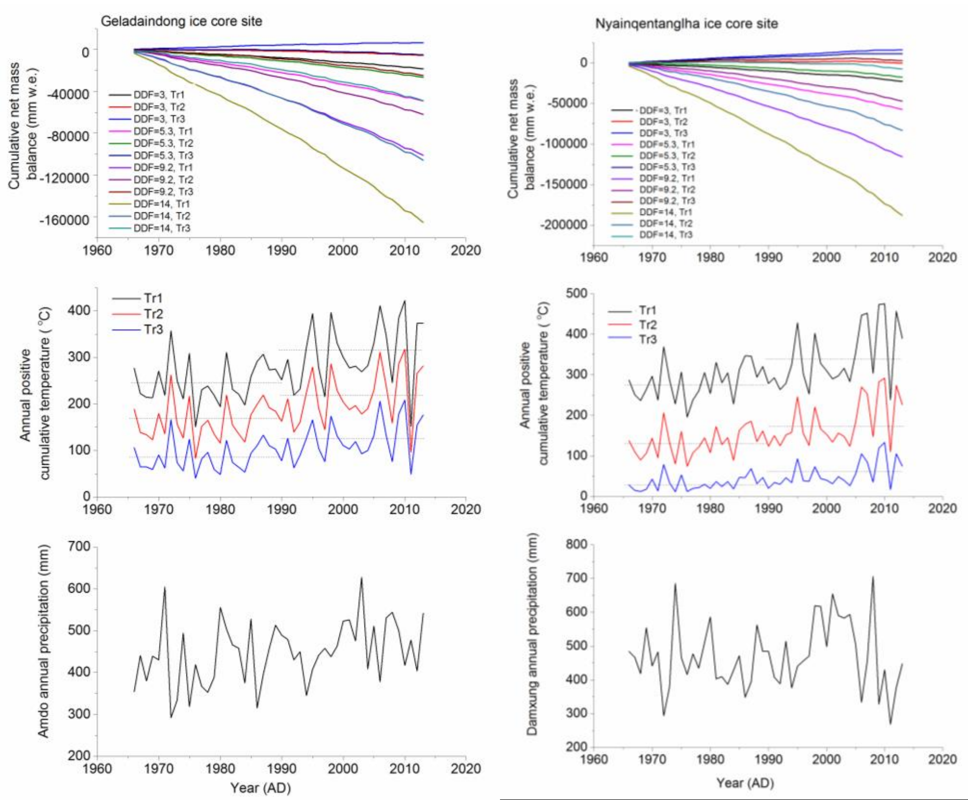

Xiaodongkemadi
Cumulative mass balance (mm w.e.)
Annual mass balance (mm w.e.)
2000
0
-2000
-4000
-6000
-8000
500
0
-500
-1000
-1500
1990
1995
2000
2005
2010
Year
Zhadang
Cumulative mass balance (mm w.e.)
Annual mass balance (mm w.e.)
2000
0
-2000
-4000
-6000
-8000
-10000
1000
0
-1000
-2000
-3000
1990
1995
2000
2005
2010
Year

Figure 7



Geladaindong ice core site
Nyainqentanglha ice core site
Cumulative net mass balance (mm w.e.)
DDF=3, Tr1
DDF=3, Tr2
DDF=3, Tr3
DDF=5.3, Tr1
DDF=5.3, Tr2
DDF=5.3, Tr3
DDF=9.2, Tr1
DDF=9.2, Tr2
DDF=9.2, Tr3
DDF=14, Tr1
DDF=14, Tr2
DDF=14, Tr3
Annual positive cumulative temperature (°C)
Tr1
Tr2
Tr3
Amdo annual precipitation (mm)
Damxung annual precipitation (mm)
Year (AD)