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Decapitation of high-altitude glaciers 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 to inland Tibetan Plateau. The combined analysis of tritium (^{3}H), ^{210}Pb, mercury tracers, along with other chemical records, revealed that the two coring sites had not received net ice accumulation since at least the 1950s and 1980s, respectively, implying an annual ice loss rate of more than several hundred millimeter water equivalent over these periods. Both mass balance modeling at the sites and in situ data from 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 glacier decapitation (i.e., the loss of the accumulation zone) is a widespread phenomenon from the southern to inland Tibetan Plateau even at the summit regions. This 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 (TP, 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 retreating has been occurring to more than 82 % of the total glaciers in the region (Liu et al., 2006), and that since the 1970s glacier areas have reduced by several percent in the central (Ye et al., 2006) and up to 20 % in the northeastern marginal areas of the TP (Cao et al., 2010; Pan et al., 2012). In situ stake observations have also confirmed a continuously negative

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mass balance during last decade in the region (Yao et al., 2012). 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 TP 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 maker 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 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 glacier decapitation is a wide-occurring phenomenon at summit regions over the TP, here we report the ice accumulation chronology of two glacier ice cores taken from high elevations ($\sim$ 5800 m a.s.l.) at Mt. Nyainqentanglha (NQ) and Mt. Geladaindong (GL) in the TP. In addition to radioisotopes ^{3}H and ^{210}Pb and other geochemical tracers, the depth profile of Hg is used as a new chronological 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 TP is home to the largest volume of glacier ice outside the polar regions (Grinsted, 2013). Climatically the southern and central TP is influenced primarily by the Indian Monsoon and the continental climate of central Asia (Bryson, 1986). The TP blocks mid-latitude westerlies, splitting the jet into two currents that flow the south and north of the plateau, respectively. The plateau is

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also a major forcing factor on the intensity of the Asian monsoons, resulting in a complete reversal of weather patterns occurring from summer to winter (Bryson, 1986; Tang, 1998).

Two ice cores were retrieved as part of the Sino-US Cooperation Expedition (Fig. 1). The NQ ice core (30°24.59′ N, 90°34.29′ E, 5850 m a.s.l.), by drilling to the bedrock with a length of 124 m, was collected in September of 2003 from the Lanong glacier pass on the eastern saddle of Mt. NQ (peak height: 7162 m) in the southern TP. The GL 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. GL (peak height: 6621 m), which is the summit of the Tanggula Mountains in central TP 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. NQ region (Shi et al., 2005) and 5570 m a.s.l. in the Mt. GL region (Zhang, 1981), suggesting that both ice cores could be located in the accumulation area. Snow pits, with a depth of 40 and 70 cm at the NQ and GL coring sites respectively, were also sampled at a 10 cm depth interval.

The NQ ice core was transported frozen to the Climate Change Institute (CCI) at the University of Maine, USA, whereas the GL core to the State Key Laboratory of Cryospheric Sciences (SKLCS) of the Chinese Academy of Sciences (CAS), Lanzhou, China. The cores were sectioned at 3 to 5 cm intervals in a cold (−20 °C) room, with the outer parts being scraped off using a pre-cleaned ceramic knife. The inner parts 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 outer parts 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 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 (TEL), Institute of Tibetan Plateau Research (ITP), CAS, Beijing. Soluble major ions (Na^+, K^+, Mg^{2+}, Ca^{2+},

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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 CCI (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 (UCTEL) 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 TEL, CAS, China. 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 results showed good agreement of differences within 15 % (Loewen et al., 2007; Zhang et al., 2012).

The ^3H was measured by using Quantulus Low-level Liquid Scintillation Counters (LSC) (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., ^3H) 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 ^3H, in increasing amounts, were released into the atmosphere by above-ground thermonuclear tests between AD 1952–1963, resulting in atmospheric levels several orders of magnitude above natural cosmogenic concentrations (Clark

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and Fritz, 1997). Remarkably, this global ^{3}H marker is not present in the NQ ice core (Fig. 2). All samples collected from the NQ core had ^{3}H activity below the detection limit of 0.1 TU with the exceptions of one surface sample, which represents the fresh winter snow, and another sample at 31 m, which is attributed to ^{3}H contamination during sampling. Similar to the ice cores recently taken from the Naimona'nyi Glacier in central Himalaya (Kehrwald et al., 2008), the absence of anthropogenic ^{3}H markers below surface samples implies that the cored glacier ice could have pre-dated the AD 1950s, suggesting that the NQ site might not have received net ice accumulation since at least the AD 1950s.

To further test this hypothesis, we analyzed ^{210}Pb activity in the NQ 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 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 NQ site since at least AD 1950s.

In contrast, the GL 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 thermonuclear bomb testing era is still present. We attribute the highest ^{3}H level at 5.74 m to the year AD 1963, 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 established the chronology of the GL ice core by counting annual layers according to the seasonal cycles of δ^{18}O, major ions, and elemental concentrations upward to the top of the core (Fig. 4). The uppermost ice layer was designated as AD 1982 based on annual layer counting above the AD 1963 ^{3}H marker, and is further supported by ^{210}Pb estimation (Fig. 3b).

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To further investigate whether the lack of net ice accumulation at the GL 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 into the atmosphere has been greatly enhanced coinciding with the rise in anthropogenic activities (e.g., mining, burning of fossil fuels). Since Hg is primarily transported through the atmosphere, Hg profiles in ice cores from high (Fain 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 TP, 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 emissions, 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 GL ice core, with the upper-most layer dated to AD 1982, matches well with 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 TP, as well as the history of regional and global Hg production (Hyland and Goodsite, 2006), showing low and stable background levels prior to ∼ AD 1850 , 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 tracks well with the global and regional emission trends. Such declining trends are absent in the GL ice core, supporting our hypothesis that this site has not received net ice accumulation since the 1980s.

The lack of recent deposition of mass (ice) at the NQ and GL glaciers, as well as at the Naimona'nyi Glacier (Kehrwald et al., 2008), suggests that the melting and/or decapitating is a wide-spread phenomenon in at least the southern to central part of the TP and even at the upper regions of high elevation glaciers (5750–6050 m a.s.l.). Although there is a consensus that glaciers in the TP are largely experiencing shrinkage (Yao et al., 2012; Bolch et al., 2012; Neckel et al., 2014), ^{3}H and Hg records reported

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herein provide direct evidence of dramatic thinning occurring at the summit regions of glaciers that had traditionally been considered as net ice accumulation areas where ice cores were retrieved for reconstructing paleoclimate.

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. However, annual precipitation data from local lower elevation meteorological stations, Damxung (444 mm, 50 km southeast of NQ but at an elevation of 4300 m a.s.l.) and Amdo (467 mm, 120 km south of GL but at an elevation of 4600 m a.s.l.) (Fig. 1), suggest that the annual mass losses from upper glacier areas are at least on the order of several hundred millimeter water equivalent (mm w.e.). These estimates are 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).

In situ observed 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 the GL, 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). Since 1995, 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 NQ, Fig. 1) in the southern TP; over the period 2005–2011 mass loss rate at this glacier averaged about 1200 mm w.e. yr^{-1} (Qu et al., 2014). More recently, Neckel et al. (2014) reported the glacier mass changes during 2003–2009 for the eight sub-regions in the TP using ICESat laser altimetry measurements. They estimated that a mass balance at -580 ± 310 mm w.e. yr^{-1} was observed in central TP covering the GL site. The mass losses vary due to different measurements and time periods. Over the entire TP, in situ observed glacier mass losses ranged from 400 to 1100 mm w.e. yr^{-1} during the last decade with an exception of slight mass gain in the

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northwestern of the TP (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 negative mass balance of Zhadang and Xiaodongkemadi glaciers confirm to widespread glacier mass loss from the southern to inland TP during the past decade, which are consistent with our 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 (DDM) was applied to estimate glacier melt at the GL and NQ sites. DDM can determine the daily quantity of snow/ice melt (m_t, mm w.e.) as a function of the mean daily air temperature (T_t, °C) using a factor of proportionality referred to the degree-day factor (DDF, mm °C^{-1} d^{-1}) (Gardner and Sharp, 2009).

$$
\begin{aligned}
m_t &= \mathrm{DDF} \cdot T_t \quad && T_t \geq 0 \\
m_t &= 0 && T_t < 0
\end{aligned}
$$

To detect the net mass balance at the NQ and GL coring sites by DDM, we selected daily temperature and precipitation data from the two meteorological stations, Damxung and Amdo, which are the nearest station to the NQ and GL sites (Fig. 1), respectively. Daily temperature and positive accumulated temperature at the two sites were calculated according to the vertical lapse rate (~ 0.6 °C (100 m)$^{-1}$) (Fig. 7). The accumulation rate at each coring site was considered the same as the precipitation amount at the station, although more precipitation may occur at the higher elevations of glacier area compared with those at lower elevation stations (Shen and Liang, 2004; Wang et al., 2009). 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. According to the previous works from the southern to central TP (Wu et al., 2010; Zhang et al., 2006), we selected DDF values between 3–14 mm °C^{-1} d^{-1}.

Figure 7 shows a dramatic increasing trend in positive accumulated temperature at the NQ and GL sites but no clear trend in precipitation. Furthermore, a clear decreas-

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ing trend of the cumulative net mass balance ($r^2 = 0.98$ at NQ and $r^2 = 0.99$ at GL, significant level at 99 % for T test) is presented at the two coring sites, indicating a consistent status of glacier mass losing (Fig. 7). The calculated net mass balance during 1963–2013 was -959 ± 532 mm w.e. yr^{-1} (range: -163–-1756 mm w.e. yr^{-1}; according to the DDF value range 3–14 mm °C^{-1} d^{-1}) at the NQ site and -1225 ± 473 mm w.e. yr^{-1} (range: -296–-2153 mm w.e. yr^{-1}) at the GL site (Fig. 8). Considering that more precipitation (Shen and Liang, 2004) and possibly larger lapse rate in the glacier regions (e.g., annual mean value of 0.72 ± 0.01 °C (100 m)$^{-1}$ in the northern slope of Mt. Everest, Yang et al., 2011), the actual mass loss rates at the two sites might be slightly less than these estimated values.

4 Conclusion

Meteorological data suggest dramatic warming has occurred in the TP 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) on glacier during the last decade ranging from Himalayas to the north of the TP except for the northwestern of the TP (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 the majority of the observed glaciers has risen beyond the highest elevations of the glaciers; that is, there is no more net accumulation area and hence the glaciers have been “decapitated” and subsequently the entire glacier is ablating (Yao et al., 2012). Although glacier mass balance varies depending on climate change and geographical conditions as shown on the TP (e.g. Yao et al., 2012; Bolch et al., 2012), our ^{3}H and Hg ice core records confirm that the upper glacier areas (e.g. ∼ 5750–6000 m a.s.l.) are rapidly transforming into ablation areas in recent years. In particular, extensive ablation has caused substantial mass loss of the NQ and GL glaciers since at least the 1950s in the southern part and the 1980s in central part of the TP, respectively.

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We suggest that the glaciers on the southern to inland TP 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 outburst (Richardson and Reynolds, 2000; Zhang et al., 2009). Further, glacier decapitation 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.

Author contributions. S. Kang was the lead scientist of the entire project and F. Wang was the principal investigator of the mercury sub-project. S. Kaing and F. Wang wrote the first draft of the manuscript, with inputs from all other co-authors. U. Morgenstern did the tritium measurement and interpretation. M. Schwikowski did the ^{210}Pb measurement and interpretation. Y. Zhang, B. Grigholm and S. Kang did the major ions, elements, and stable oxygen isotope measurements and analyzed the data. J. Ren, T. Yao, D. Qin and P. Mayewski conceived and designed the experiments.

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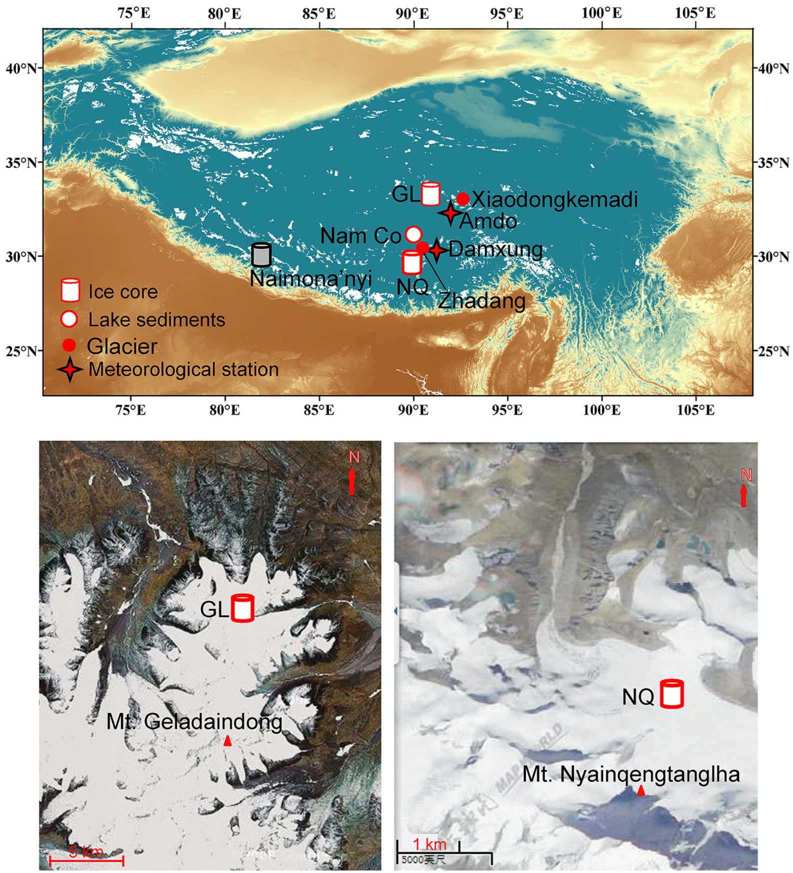

Figure 1. Location map of ice cores from Mts. Geladaindong (GL) and Nyainqengtanglha (NQ) on the Tibetan Plateau. Also shown are the locations of the Naimona'nyi ice core by Kehrwald et al. (2008), Dongkemadi glacier, Zhadang glacier, the meteorological stations and Lake Nam Co.

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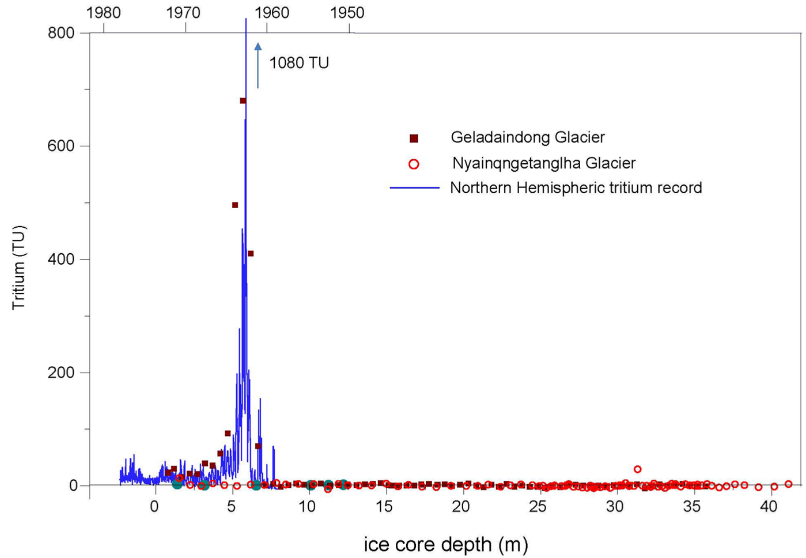

Figure 2. The tritium profiles of the Geladiandong (GL) and Nyainqengtanglha (NQ) ice cores compared with tritium in Northern Hemispheric (NH) precipitation. Error bars for the ice core samples are shown, but in most cases are only about half of the symbol size. To enable direct comparison, both the GL and precipitation tritium records are decay-corrected to the date of GL 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 NH precipitation record is scaled to match the maximum tritium concentration in the ice core to mid-1963 (time of highest tritium concentration in the NH meteoric water), and to start with 1982 (date of the surface ice, see text). To match the tritium concentrations in the GL core, the Ottawa precipitation record had to be multiplied by a factor of two. This indicates that the tritium concentration on the TP is about twice of that of Ottawa, due to a more direct input of stratospheric air, which is the main atmospheric tritium reservoir.

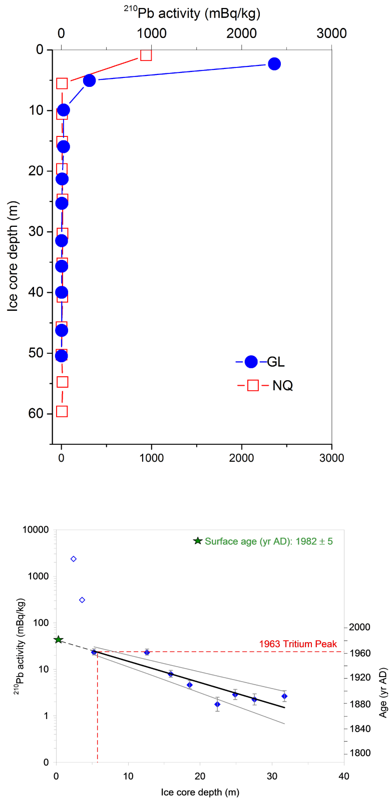

Figure 3. (a) ^{210}Pb activity profiles of the GL and NQ ice cores; **(b)** ^{210}Pb activity vs. depth for the GL core. The age–depth relationship was derived from an exponential regression of ^{210}Pb activity against depth. The uppermost two samples (open diamonds) were excluded (enrichment due to 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.

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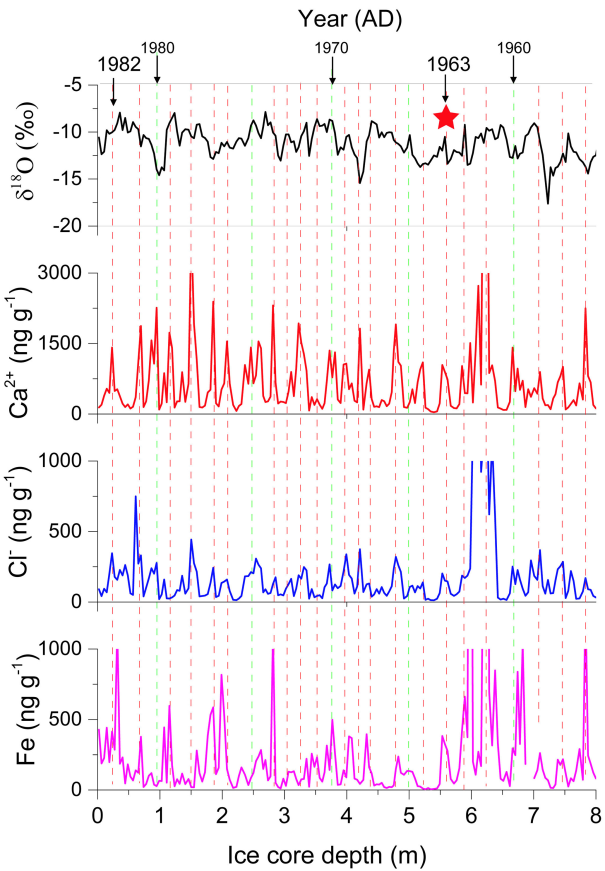


Figure 4. Dating of the GL ice core by annual layer counting based on the seasonal cycles of $\delta^{18}O$, Ca^{2+}, Cl^- and Fe according to the anchor of 1963 tritium peak (red star) (dashed lines represent the annual boundaries).

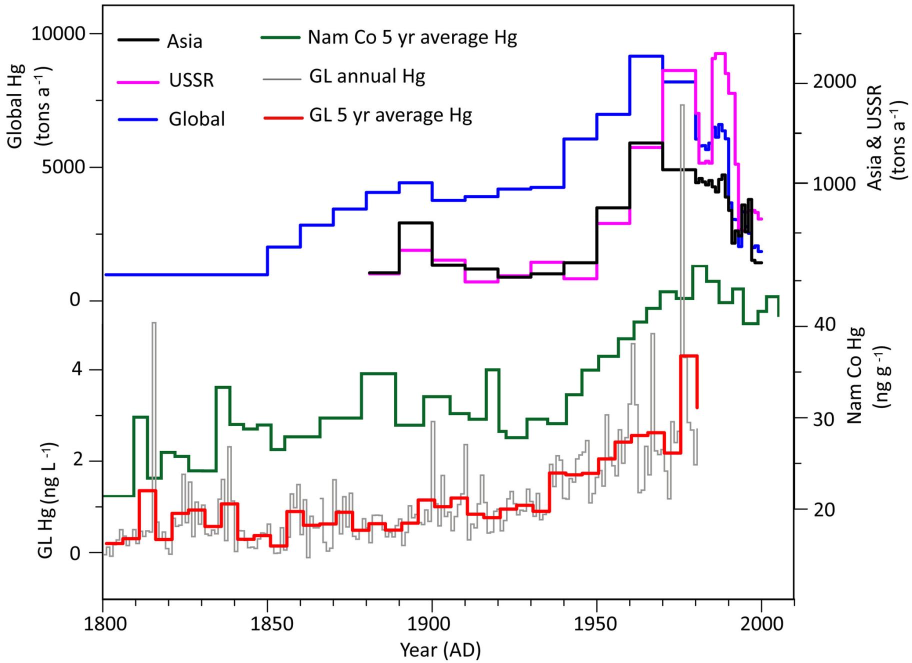
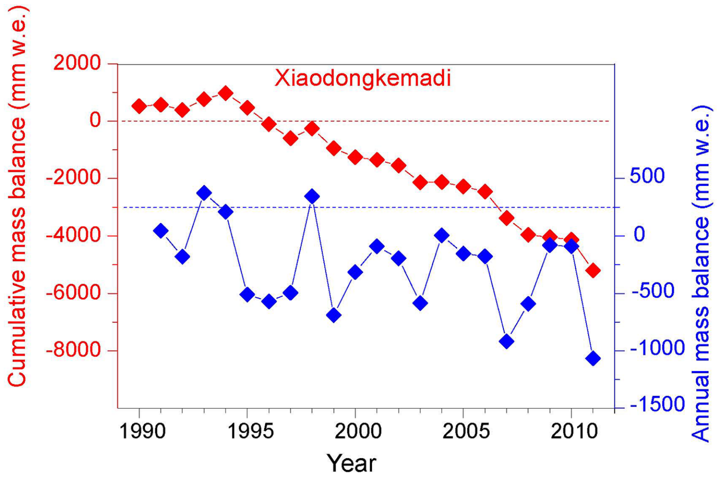


Figure 5. Comparisons of Hg records from the GL ice cores with those from Lake Nam Co sediments (Li, 2011), as well as with known history of the regional (Asia and USSR) and global Hg production (Hylander and Goodsite, 2006).



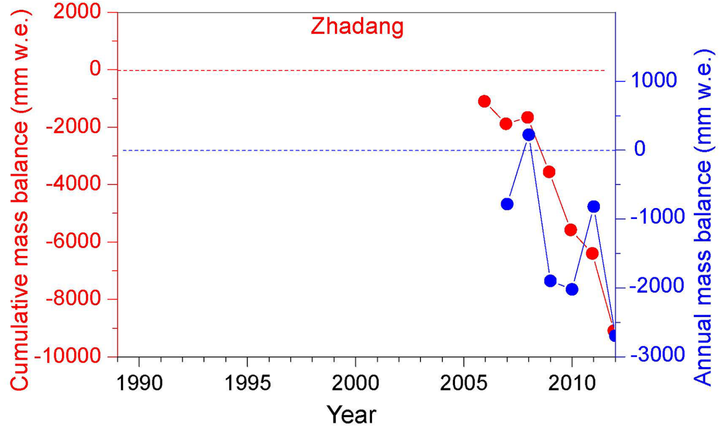


Figure 6. Annual and cumulative mass balance for Zhadang glacier in Mt. Nyainqengtanglha (Qu et al., 2014) and Xiaodongkemadi glaciers in Tanggula Mts. (Yao et al., 2012).

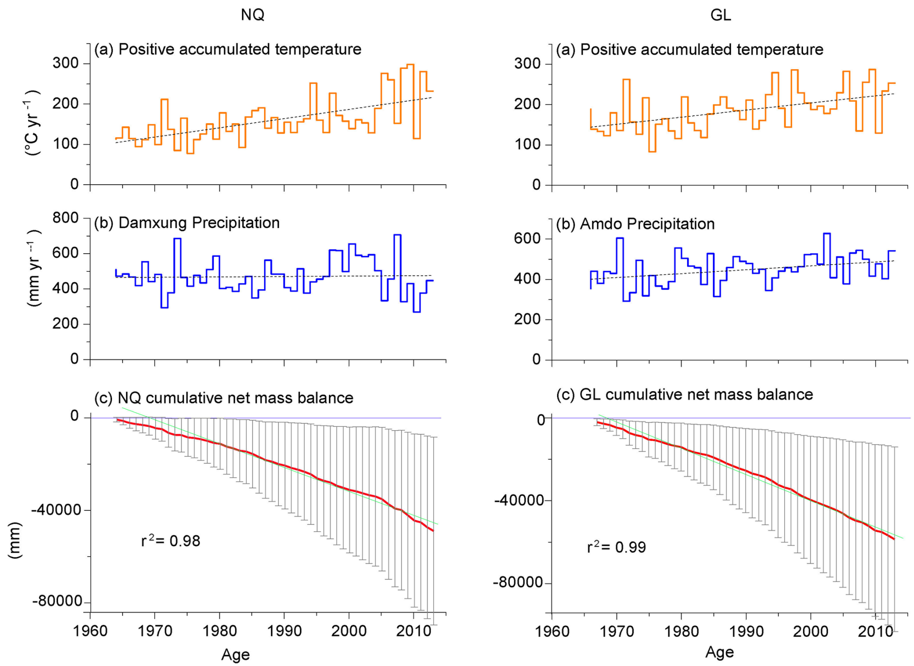


Figure 7. Variations of positive accumulated temperature at NQ and GL sites **(a)**, precipitation amount at Damxung and Amdo station **(b)**, and cumulative net mass balance based on a degree-day model (DDM) at the NQ and GL sites **(c)**. Dashed lines represent linear regression trends in (**a** and **b**). Grey, red and green lines represent annual ranges, averages, and linear regression trends, respectively, in **(c)**.

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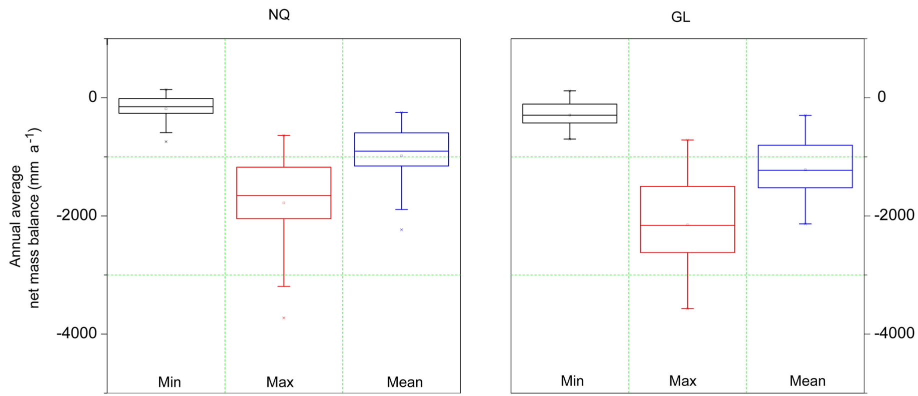

Figure 8. The box chart of the minimum, maximum and mean annual net mass balance at the NQ and GL sites.

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