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The disappearance of glaciers in the Tien Shan Mountains in Central Asia at the end of Pleistocene

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A B S T R A C T

Glaciers in Central Asia are among the largest ice masses in the Eurasian continent and have supplied vital water to local inhabitants for thousands of years. The glaciers in this region are generally believed to be remnants of the last deglaciation, however, glacier variability in the central Asian mountains since the Last Glacial Maximum (LGM) has not been well documented. Here, we report an 86.87 m-deep ice core record drilled on an ice cap in the Tien Shan Mountains of Central Asia. Radiocarbon dating of organic soil from the bottom of the ice-core borehole showed that the age of the soil was 12,656 − 12,434 cal years before present, coincident with the beginning of the Younger Dryas cold period (YD). This result indicates that the ice cap did not exist in the Bølling-Allerød period (BA), which was the warm period before the YD, and that the BA climate was significantly warmer than at present. It also indicates that the ice cap has never entirely disappeared in any warm periods throughout the Holocene. We estimated that during the BA its extent was 43% or less of the present glacier coverage in the mountains. Our results suggest that this region at the end of Pleistocene was considerably warmer than at present, and that most of the present glaciers in this region are not relics of the Last Glacial period, but are composed of ice formed during the YD and Holocene.

1. Introduction

The Tien Shan Mountains are one of the major mountain systems in Central Asia. The range extends over 2000 km, from east Uzbekistan (69°E, 42°N) to northeast of the Taklamakan Desert in China (95°E, 43°N). The major Tien Shan peaks rise from over 4000–7000 m a.s.l. in height. Seasonal precipitation in Central Asia is triggered by the interactions between the atmospheric circulation from the Siberian high and the westerlies (Aizen et al., 1997). The Tien Shan Mountain ranges pose a barrier to western and northern air masses moving toward Central Asia, thus playing an important role in determining the local climate.

Glaciers presently exist in mountains mainly above 3500 m a.s.l. and most rivers originating from the Tien Shan glaciers flow down through the steppes and deserts surrounding the mountains to form the Aralo-Caspian, Balkhash, Issyk-Kul, and Tarims endortheic basins. Water from the mountains has sustained the local inhabitants of this arid area for thousands of years and has maintained the Silk Road, a famous ancient trade route between Asia and Europe (Stein, 1925; Yang, 2001; Yang et al., 2006a). Changes in glacier runoff in the Tien Shan Mountains have always been a major cause for the migration of human communities during their long history in the region. Nevertheless, the present Central Asia settlements and their millions of inhabitants still depend very much on water from the glaciers, and any decline in glacial area and volume poses a major concern to the countries of Central Asia (Barnett et al., 2005; Sorg et al., 2012).

Although the change of glacier extent in the Holocene and the late Pleistocene is a great concern in this region, the deglaciation

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process has not been documented well, particularly in the end of Pleistocene. There have been many geomorphological studies for the Pleistocene and the Holocene in this area (e.g. Koppes et al., 2008). For example, glacier chronological studies have revealed that glaciers in the Tien Shan largely extend in the Last Glacial Maximum in 19,000 years before the present (yr BP), gradually retreat after 11,000 yr BP, and that present glaciers are generally remnants of the last deglaciation (Grosswald et al., 1994). However, it has been suggested that the dating the glaciers on the basis of age of moraines in Tien Shan does not give clarity due to picking of samples from the pseudo-moraines, which were formed mainly during mass gravitational lithogenesis as a reaction to the disappearance of glaciers in the late Pleistocene and causes even greater confusion in dating (Melnikova, 1987; Shatravin, 1992, 1994a, 1994b, 2007, 2012; Romanovskiy, 2007). Furthermore, although geomorphic mapping of moraine sequences could reveal advances of glacier extent, it is unable to reconstruct minimal extent or disappearance of glaciers since the evidences are usually beneath the present ice coverage. The model of stadial glacier recession in late Pleistocene and Holocene has been adopted for Tien Shan by Shitnikov (1957) on the basis of the alpine classical model (Penck and Bruckner, 1909). According to this model, the modern glaciation in Tien Shan is a continuous extension of the Pleistocene decaying glaciation. Shitnikov believed that since the maximum glaciation in the late Pleistocene, the glacier recession has formed eight stadial moraines correlated with 1850 yr rhythms of climate variability, i.e. changes in atmospheric moisture flow over the continents (wet/dry periods). Until the 1980s many researchers followed this model defining stadial glacial moraines in Central Asia based on the description of alpine landform and random radiocarbon dating of allochthonous organic matter from the moraine surface soils (Sevastyanov, 1974; Pomortcev, 1980; Maksimov, 1980). Later, Maksimov et al. (1987) has concluded that this method was not correct because the descriptively defined moraines do not match the radiocarbon dating in many cases. A hypothesis about a “quasi-stationary” state of the Tien Shan glaciers in Holocene also scientifically not motivated, because the dendrochronological and lichenometric dating of the Tien Shan glacial stadials reflect only the late Holocene, i.e. not more than the last 1000 years (Solomina, 1999).

Ice core studies provide a means of reconstruction of continuous climate history. However, there have been only two ice cores in Central Asia that cover the entire Holocene and the Pleistocene deglaciation (Thompson et al., 1989, 1997b). Furthermore, their oxygen-isotope records, which are commonly used as air temperature proxies, have some difficulties of interpretation due to effect of melt and multiple moisture sources. The age of the basal section of ice cores can be indicative of ice-free conditions in the mountain when the climate was warmer and/or dryer than present. For example, onset of neoglaciation of 6000 years ago in western Mongolia has been identified by the age of basal ice (Herren et al., 2013). This approach could reveal minimal glacier extent in the past, which cannot be determined by geomorphological studies. Although ice cores have been recovered from some glaciers in the Tien Shan Mountains (e.g. Thompson et al., 1997a; Kreutz et al., 2001; Lee et al., 2003), they were generally shallow, covered only the last decades, and did not reach the bedrock.

The Grigoriev Ice Cap is located in the Teskey Ala-Too Range in Inner Tien Shan, an area running southward from Lake Issyk-Kul, the world's second largest mountain lake (Fig. 1). The ice cap is a simple ice cap covering a small mountain approximately 8 km^2 in area. Most of the glaciers in this region are ice caps or small valley glaciers. Thus, the shape of the Grigoriev Ice Cap is representative in this region in terms of size, location and topography and response of this glacier to climate change is probably typical in the region.

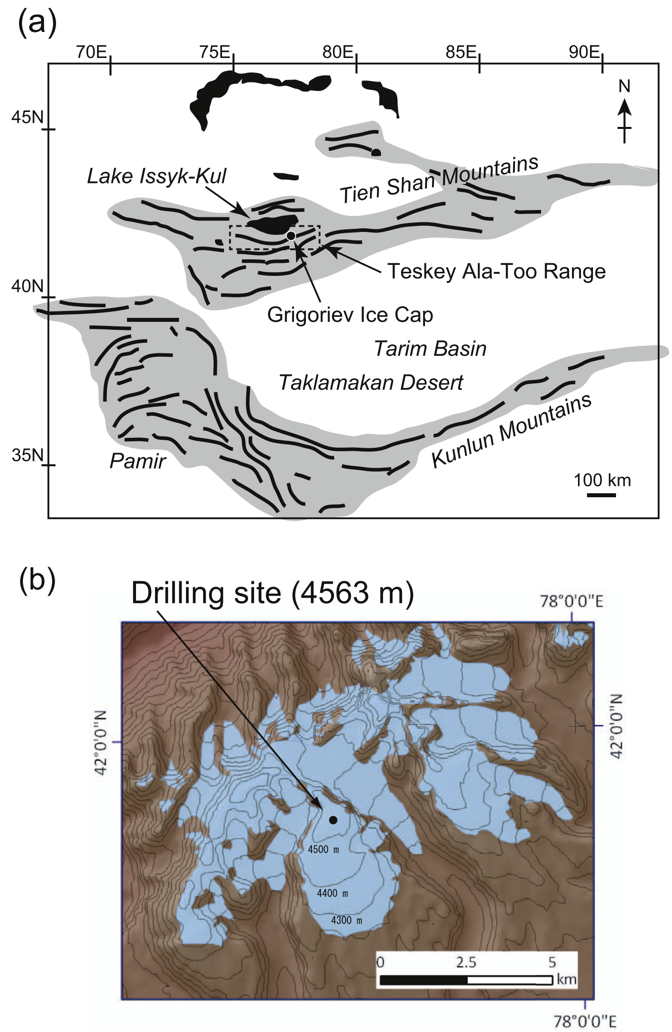

Fig. 1. Location (a) and map (b) of the Grigoriev Ice Cap in Kyrgyzstan, Central Asia, showing the drilling site. Light blue area indicates present glacial coverage based on a satellite image (ASTER) acquired on Aug. 14, 2004. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

The top is a flat snow field at an elevation of 4563 m a.s.l. According to our 2-year instrumental records at the summit of the ice cap (2005–2007), the annual total precipitation is approximately 290 mm (Fujita et al., 2011). Water vapor is mainly supplied by the westerlies (Numaguti, 1999). Shallow ice cores have been recovered in 1990, but they were 16.5 and 20 m in length and covered only the last 50 years (Thompson et al., 1997a). We successfully drilled an ice core from the surface to the bottom at the same site on the ice cap in 2007. This study aims to document the variability of glaciers and climate in Central Asia using the ice core of the Grigoriev Ice Cap in Tien Shan Mountains. In particular, we discussed the climate and glacier extent when the ice cap did not exist based on the dating of bottom of the ice core.

2. Study site and methods

The ice core was drilled in September 2007 at the top of the Grigoriev Ice Cap (41°58′33″N, 77°54′48″E) in Kyrgyzstan (Fig. 1), with an electro-mechanical drill manufactured by Geo Tecs Co., Japan. The barrel size of the drill was 135 cm in length and 9.5 cm in inner diameter. The detailed specification of the drill was described in Takeuchi et al. (2004). The drill reached bottom of the ice cap at a depth of 86.87 m and penetrated into frozen organic-rich soil layer, which gave us an opportunity to collect approximately 100 g of soil material (Fig. 2a).

The frozen ice core and soil were transported to a cold laboratory at the Research Institute for Humanity and Nature in Kyoto, Japan. In the laboratory, the core was cut every 4–5 cm for the upper 50 m and 1–4 cm below 50 m, which was adapted to layer thinning with depth. The total number of samples was 2067. Each sample was used for analyses of oxygen stable isotopes and particle concentration. Oxygen stable isotope ratios ($\delta^{18}O$) were analyzed with a liquid-water isotope analyzer (DLT-100, Los Gatos Research) in Chiba University. The analytical precision of $\delta^{18}O$ measurement was 0.1‰. Particle concentrations were analyzed with a laser particle counter (System 8103-MC05, HIAC Royco, size range: 2.0–350 μm) in the Research Institute for Humanity and Nature.

In order to identify a reference horizon of the nuclear weapons testing in 1963, the tritium radioactivity was measured every 20 cm between 15 and 25 m of the ice core by liquid scintillation counting in National Institute of Polar Research in Japan.

The bottom soil and three dust samples collected from prominent dusty layers of the ice core (79.55, 81.79, and 85.54 m, Fig. 2b) were analyzed for radiocarbon. The quantities of dust obtained from the ice core were 135.65 mg, 92.19 mg, and 65.41 mg, respectively. The samples were dried and acidified to remove carbonates. The radiocarbon was measured by accelerated mass spectrometers at the University of Tokyo and the Institute of Accelerator Analysis Ltd. (Japan). The obtained radiocarbon ages were calibrated to calendar ages with OxCal version 4.1, using the IntCal09 calibration curve (Reimer et al., 2009).

Using a geographical information-system (GIS) tool, we simulated the past glacier extent in the Teskey Ala-Too Range (area: approximately 6000 km^2), which includes the Grigoriev Ice Cap, and compared it with the present glacier extent. The present glacier extent was obtained from satellite images acquired on Sep. 17, 2007 by the Advanced Land Observing Satellite (ALOS) and on Aug. 14, 2004 by the Advanced Spaceborne Thermal Emission and Reflection radiometer (ASTER). The past glacier extent when the Grigoriev Ice Cap disappeared was estimated using the 90-m digital elevation model (DEM) of the Shuttle Radar Topography Mission (SRTM) with a GIS application of ArcGIS. We obtained the altitudinal distribution of glacier area in every 10 m interval for each catchment, and then we applied the elevation change of glacier terminus in the time of glacier retreat.

3. Results

3.1. Ice core

Ice and firn layers were readily observed in the upper 20 m of the core and were well-demonstrated by the density profile (Fig. 3). The ice layers (thickness) accounted for approximately 60% of the upper 20 m of the core. The density generally increased from ca 250 to ca 800 kg m^{-3} in the firn layers above 6 m, varied from ca 600–850 kg m^{-3} in the alternate firn and ice layers between 6 and 22 m, and showed ca 800–910 kg m^{-3} in the continuous ice layers below 22 m. Borehole temperature measurements revealed that the ice temperature was −2.68 °C at 10 m deep and generally decreased with depth down to −3.92 °C at the bottom. The negative temperature at the bottom indicates basal ice has not melted (Fig. 3).

The tritium measurement showed a maximum at a depth of 18.4 m (Fig. 4), which was attributed to the maximum fallout from nuclear weapons testing in 1963. The total mass of snow and ice above the tritium peak layer was 12,000 mm water equivalent (w.e.), which corresponds to the net accumulation during the period from 1963 to 2007. Thus, the estimated mean net accumulation on the ice cap between 1963 and 2007 is 273 mm w.e. per year.

Oxygen stable isotope of the ice core (Fig. 4) showed higher values near the surface (mean above the tritium peak (18.4 m): −9.66‰), generally stable between 18.4 and 80.3 m (mean: −10.85‰), and very low values below a depth of 80.3 m (mean: −13.93‰). Oxygen isotope values abruptly changed from ca −10.6‰ to −20.32‰, which is the minimum value of the ice core, at a depth between 80.3 and 83.3 m. Considerable variations of the stable isotope values along the core probably reflect seasonal signals, however, their counting for dating purpose seems to be useless because of our suggested reduction of the core due to its synsedimentary melting.

The particle concentration curve shows small spikes from the surface down to a depth of 80.5 m, and significantly higher spikes below that depth (Fig. 4). The mean concentration below the depth was 6-fold greater than that of the upper part (76 versus 13 × 10^4 numbers ml^{-1}). As with the oxygen isotope curve, the spikes of particle concentration are unlikely to represent annual features.

Microscopy of the bottom soil revealed that it consisted mostly of amorphous dark-colored materials (soil organic matter) and mineral particles, as well as small amounts of plant fragments, pollen grains, and microbes (e.g., algae and bacteria). The content of organic matter was 3.1% in dry weight. The mineral part included abundant gravel-size particles more than 2 mm in diameter, which

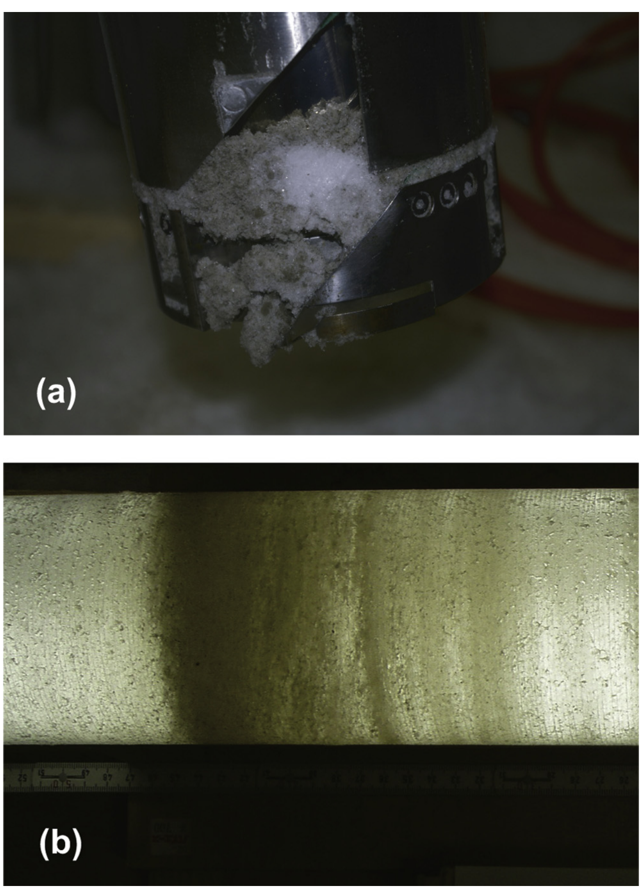

Fig. 2. Samples for radiocarbon measurements in this study: (a) Organic soil beneath the Grigoriev Ice Cap adhered to the ice-drill cutters during the last run. (b) Prominent dust layer is shown in an ice core section at 85.54 m below the surface (1.33 m above the bottom).

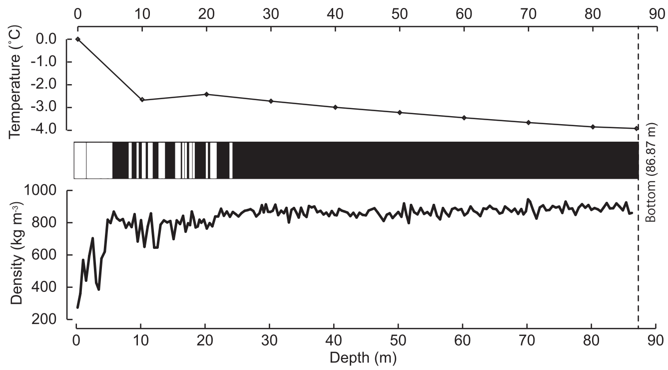

Fig. 3. Records of borehole temperature, ice core visible stratigraphy, and density. For visible stratigraphy, white and black mean firn and ice layers, respectively.

were significantly larger than the aeolian dust particles in the glacial ice. This suggests local rocks as the source of the soil mineral component; similarly local vegetation was the source of the organic matter. Because the soil was found at the mountain summit, it was unlikely to be transported by a glacier.

Measurements of radiocarbon showed that the radiocarbon age of the bottom soil was 10,640 ± 90 ^{14}C years before the present (yr BP), which corresponded to 12,656–12,434 cal yr BP (probability: 68.2%, Table 1). Three samples collected from dust layers in the ice core, which were likely derived from windblown dust from the surrounding ground surface, also contained enough organic carbon for radiocarbon measurements. Their ages were 6174–5919 cal yr BP at 79.55 m deep (7.30 m above the bottom), 8017–7969 cal yr BP at 81.79 m deep (5.07 m above the bottom) and 8155–8002 cal yr BP at 85.54 m deep from the surface (1.33 m above the bottom, Table 1).

3.2. GIS simulation of glacier extents

Satellite images showed that the 710 glaciers presently existing in the study area of the Teskey Ala-Too Range have total area of

Table 1
Result of radiocabon measurements for the Grigoriev ice core samples. For the calibrated calender year, ranges are given with 68.2% probalility.

Material	Depth, m	Sample weight, mg of dryweight	^{14}C age, yrs BP	Calibrated age, cal. yrs BP	Lab code
Dust layer	79.55	135.65	5235 ± 60	6174–5919	MTC-13476
Dust layer	81.79	92.19	7189 ± 33	8017–7969	IAAA121389
Dust layer	85.54	65.41	7240 ± 40	8155–8002	IAAA72749
Bottom soil	86.87	182.41	10,640 ± 90	12,656–12,434	MTC-13477

596 km^2 (Fig. 5). We identified 12 catchments (Fig. S1) and calculated hypsometry of glacier extent in each catchment (Table S1 and Fig. S2). The elevation difference between the present terminus (4163 m) and the bedrock at the drilling site (4476 m) is 313 m. We therefore assume that glacier termini retreated as much as 313 m in elevation in all catchments in order to keep their individual gradients of area–elevation relationships (Fig. S2). We finally estimate that the past glacier extent when our studied ice cap disappeared was 254 km^2, which is 43% of the present glacial coverage (Fig. 5 and Table S1).

4. Discussion

4.1. Reasons for stable isotope variations of the ice core

The visible stratigraphy of the ice core suggests that summer surface melting has occurred during the most recent decades. Because of the sub-zero temperature of the ice, surface melt water could refreeze in lower layers, thus the average value of the oxygen stable isotope could be preserved in the ice core. Although stable isotopes in central Asian precipitation vary for many reasons, including source of moisture, amount effect, and temperature, the observations have revealed that stable isotopes of precipitation in the Tien Shan Mountains are closely correlated with the temporal variations of air temperature (e.g. Tian et al., 2007). Therefore, the higher values above the tritium peak are likely due to climate warming in the last 44 years, which is consistent with the previous

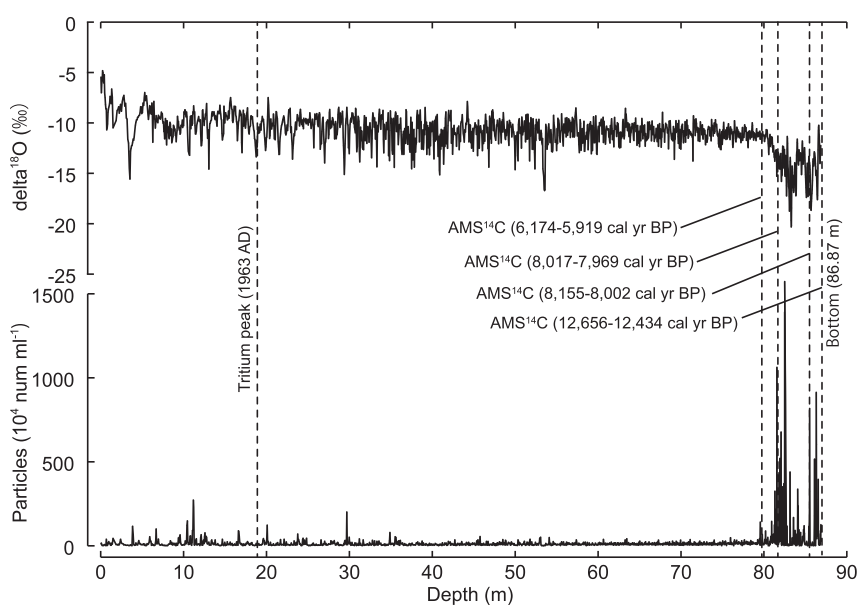

Fig. 4. Records of oxygen stable isotope ratios (δ^{18}O) and total particle concentration (size range: 2.0–350 μm) in the ice core of the Grigoriev Ice Cap. Dashed lines show the reference horizons revealed by the tritium peak and radiocarbon ages of the dust layers.

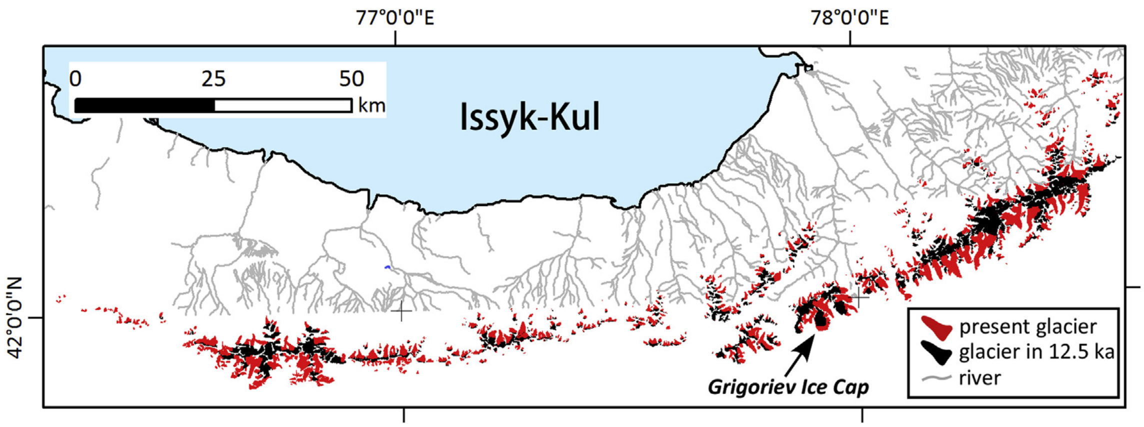

Fig. 5. The GIS-based comparison of the glacier extents in the Teskey Ala-Too Range of the Tien Shan Mountains in Bølling-Allerød (black) and at present (red). We assumed the 313 m difference between glacial termini in BA and now. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

shallow ice cores from the same site (Thompson et al., 1997a). The lower values below 80.5 m suggest very cold climate. The difference of the mean values between below and above 80.5 m was 3.42‰, which is comparable to the difference between the values of Holocene and the Last Glacial period in a Greenland ice core (4‰, Stuiver et al., 1995).

4.2. Initiation of the Grigoriev Ice Cap: geochronological and paleoclimatic constraints

The carbon in the bottom soil was likely derived from biological processes on the ground, including flora and microbes, before the glacier existed. Since the soil had been preserved at the bottom of an ice cap at a temperature below freezing, contamination of the soil by other carbon was unlikely. Therefore, the carbon age would indicate the approximate time of the ice cap's initiation. Age offset of decades to centuries towards old ages may exist due to mixture of old carbon to the dust (Sigl et al., 2009), however, the offset could be smaller since most of dust was likely to be derived from ground soil surrounding the glacier. These ages indicate that the present ice cap did not contain ice that formed in the Last Glacial period before 12.5 cal ka BP, but mostly that which was formed during the Holocene. The basal ice of low stable isotope value is likely to be deposited during a very cold climate event between about 12.5 and 8.0 cal ka BP, preceded the Holocene climate optimum in the Tien Shan area (7–4 cal ka BP, Feng et al., 2006). The cold climate event can be the Younger Dryas (YD, 12.9–11.7 cal ka BP) or the 8.2 ka cooling event, which can be seen in Greenland ice cores and other proxies in the Northern Hemisphere (e.g. Alley et al., 1997). The ice between the upper two dust layers (6.0–8.0 cal ka BP) is likely to correspond to the Holocene climate optimum and was relatively thinner (2.24 m for a period of about 2000 years) compared with the ice between the lower two dust layers (3.75 m for a period of about 150 years). The ice also did not show more positive values than those of the ice below or above. These facts suggest the glacier partly melted during the mid-Holocene, and there was hiatus of ice layers in this period.

Since the age of the bottom soil coincides with the beginning of the YD, the Grigoriev Ice Cap was not present at that time. Studies of glacier chronology in the Inner Tien Shan area showed significant glacial expansions during MIS2 (29–21 ka; Narama et al., 2007, 21 and 15 ka; Zech, 2012). In addition, several glacial advances during the lateglacial have been relatively well defined in Muztag Ata and Kongur Shan of western Pamir (Seong et al., 2009). If any glacier existed on the place of the Grigoriev Ice Cap before, it disappeared sometime between the Last Glacial Maximum (LGM, 22 cal ka BP) and the YD. The GISP2 Greenland ice cores confirm that the climate became warmer at the onset of the Bølling-Allerød warm period (BA) ca 15.0–14.3 cal ka BP (Stuiver et al., 1995). It continued for approximately 1800 years, then terminated at the onset of the YD ca 13.2–12.6 cal yr BP. Therefore, the organic soil was most likely formed during the BA warm period.

This deglaciation event in the Tien Shan Mountains has also been suggested by previous investigators. For example, palynological data from Teskey Ala-Too Range (Inner Tien Shan), close to the Grigoriev Ice Cap, have determined an interglacial period at the end of the late Pleistocene (Melnikova, 1987; Osmonov, 1991). This interglacial has been also discovered from palynological and radiocarbon analyses of the Lake Chatyr Kul (3530 m a.s.l.) sediments and lacustrine terraces (Romanovskiy, 2007).

Our results indicate that the ice cap did not disappear during warm periods of the Holocene, such as the Holocene climate optimum 7–4 cal ka BP (Feng et al., 2006) and the Medieval Warm Period 1.2–0.8 cal ka BP (Yang et al., 2009), which are commonly recorded in climate proxies in Central Asia. Therefore, the BA was likely to be the glacial minimum in the period from the LGM to the present in this region. In other regions, like the Mongolian Altai Mountains, located ca 1000 km northeastward of the Grigoriev Ice Cap, glaciers disappeared during the Holocene Climate Optimum (Herren et al., 2013). This suggests the glacier response to warming events was asynchronous over the Asian high mountains.

Another ice cap probably occupied the place of the present Grigoriev Ice Cap during the LGM. The disappearance of this ice cap indicates that the BA climate was either much warmer or much drier compared with the present climate. Although proxies of the Pleistocene climate in Central Asia are still limited, there are evidences of wetter conditions in western Chinese deserts during the LGM (Yang and Scuderi, 2010). Studies of lake sediments in this region showed that the LGM was characterized by dry climate conditions, whereas the BA climate was wet or moderately wet (Herzschuh, 2006). Furthermore, since the formation of organic soil requires wet and warm conditions, we assume that the BA glacier recession occurred due to a wet and warm climate rather than an arid one.

Stable isotope records of the Guliya ice core on the Tibetan Plateau (approximately 800 km south from the Grigoriev Ice Cap) revealed that the temperature fluctuations during the last 125 ka are consistent with glacial cycles recorded in ice cores from Greenland (Thompson et al., 1997b). It should be noted that $\delta^{18}O$ of the Guliya ice core during the BA show a higher values (−10 to −12‰) compared with those of the LGM (−18.5‰) and the present (−13 to −15‰). If $\delta^{18}O$ is indicative of the air temperature in the region, this is consistent with our statement that the BA climate was warmer than the present one. In contrast, higher isotope values in the BA are not apparent in the Greenland ice cores (Dansgaard et al., 1993; Stuiver et al., 1995). Although the records showed a significant increase in $\delta^{18}O$ during the last deglaciation, the isotopic values during the BA did not exceed the present level. This suggests that the BA climate in Greenland may be comparable to the present climate.

4.3. Broader implication of the disappeared pre-Grigoriev ice cap in the BA

Our estimation shows that the glacier extent when the pre-Grigoriev ice cap disappeared in the BA to be 254 km^2, which is 43% of the present glacial coverage in the Teskey Ala-Too Range (Fig. 5). Behavior of the glaciers might vary depending on their size, location, and/or local climate regime. However, the Grigoriev Ice Cap has typical shape for the glaciers of this area and flows on smooth mountain slopes without any special or distinct features such as debris-covers or confluences. We therefore assume that the retreat of the glacier termini by 313 m in elevation is reasonably applicable for other catchments of the Teskey Ala-Too Range. Since this retreat is a minimal estimate based on the assumption that the glacier terminus in the BA is located at the elevation of the bedrock at the drilling site (4476 m), it is possible that the terminus in the BA was higher than the bedrock elevation. Thus, the actual extent might be smaller than the estimation. This result shows that most of present glaciers in the region are not remnants of the Last Glacial period.

We also estimated summer temperatures during the BA based on the difference in the BA and present terminus elevations. An energy-mass balance model for the Grigoriev Ice Cap shows that the annual precipitation and summer mean temperature are major factors to influence mass balance variability (Fujita et al., 2011). The summer (June to August) lapse rate in this area is 7.25 ± 0.17 °C km^{-1}, which was obtained from daily temperatures and geopotential heights at 500 hPa and 600 hPa from 1989 to 2007 of the NCEP/NCAR reanalysis dataset (Kalnay et al., 1996). Based on the lapse rate, the 313-m difference in glacier termini between BA and present is equivalent to approximately 2.7 °C of mean summer temperature, if to assume that the BA moisture conditions were similar to the present ones. This suggests at least 2.7 °C higher mean summer temperature in the BA than now. Geomorphological evidence of the LGM and present positions of the equilibrium line altitude (ELA) suggests 600–700 m difference (Narama et al., 2007) which is equal to 4.4–5.1 °C change in temperature. Because the highest elevation of lateral moraine position should be adapted for the ELA in the past, the temperature rise during deglaciation can be roughly estimated as 7.1 °C (ELA: 600 m). This estimate is higher than the temperature rise reported in other studies. For example, the NCAR CCSM3 general circulation model suggests 1–5 °C temperature increase from the LGM to BA in Central Asia, whereas a more significant increase occurred in the North Atlantic region (Liu et al., 2009). Proxy records also showed the glacial-interglacial temperature changes ranged from 2 to 7 °C in the middle latitudes (30–50°N) (Shakun and Carlson, 2010), although sediment records at the southern margin of the Taklamakan Desert imply that the LGM temperature is ca 13–18 °C lower than the present temperature (Yang et al., 2006b). Reasons of inconsistencies in warming of Central Asia in BA remain uncertain. The climate variability in Central Asia during the Holocene and late Pleistocene has been explained by changes in circulation systems, including the Indian Monsoon, Southeast Asian Monsoon, and westerlies (Herzschuh, 2006). Climate proxies demonstrate intensification of the Asian monsoon circulation at the beginning of the BA (Herzschuh, 2006). This change may have caused significant warming in the Inner Tien Shan area. Therefore, our estimate probably shows upper limits of the BA warming because decrease in precipitation could also cause degradation of glaciers. The energy-mass balance model of the Grigoriev Ice Cap suggests that, if aridity increases, glaciers become more sensitive to changes in precipitation than in temperature (Fujita et al., 2011). This model also reveals an influence of dust deposition since the albedo reduction largely promotes ablation of glaciers. The greater particle concentrations observed in our ice core below 80 m suggest suppression of glacier expansion in that period. It is still difficult to evaluate impact of dust supply on glacier mass balance quantitatively because seasonal timing of dust deposition could significantly alter the extent of its impact (Fujita, 2007). Relation has to be established between the concentration of dust particles in the ice core and surface albedo when the dust was deposited at the glacier surface in order to evaluate the influence of dust.

5. Conclusions

The 86.87 m long ice core recovered from the Grigoriev Ice Cap opened its structure from the surface to the bedrock. Its annual layering is shaded due to disturbance of seasonal signals of oxygen stable isotopes and particle concentration by melt. However, the stable isotopes show very cold climate near the bottom of the ice core and general tendencies of warming upward.

Radiocarbon dating showed the deepest part of the ice core dated from the early Holocene, and the age of the underlain soil coincides the beginning of the Younger Dryas cold period (YD, 12.9–11.7 cal ka BP). This fact indicates that the present ice cap could not have existed before the YD. We suggest that an initial ice cap developed during the Last Glacial disappeared sometime before the YD. This event probably corresponds to the Bølling-Allerød warm period (BA, 14.7–12.9 cal ka BP), which characterizes the BA climate as either much warmer or much drier than now. Since formation of the organic soil found beneath the ice cap requires wet and warm conditions, we can assume that the BA glacier recession occurred due to a wetter and warmer climate rather than a drier one. Our results also indicate that the ice cap has never disappeared during any known warm period of the Holocene. Therefore, the BA was likely to be the glacial minimum in the Teskey Ala-Too Range since the LGM.

Examination of glacier shrinkage during the BA period based on the topography and present glacial coverage in this region revealed that the glacier extent during the BA was 254 km^2, which is 43% of the present glacial coverage. This result shows that most of the present glaciers in this region are not remnants of the Last Glacial period. A difference of 313 m between the terminus elevations in the BA and at present suggests that the BA mean summer temperatures were approximately 2.7 °C higher than now.

Our results suggest drastic deglaciation in this region when compared to other mountain regions, and this significantly influenced the environment. Such a change may have had a significant impact on natural environments in this region. It might also have affected migrations and settlements of the inhabitants of Central Asia during the deglaciation. The above results also suggest that glaciers in this region might respond more sensitively and quickly to the present climate warming than we had thought, which may significantly affect the water resources in this region. Although our data are based on only one glacier in the Tien Shan Mountain, this significant event of glacial retreat and climate warming are likely to have occured at least in the area of Tien Shan Mountain since the synchronicity of glacial change in this region is suggested by many geomorphological studies. However, futher investigations would be necessary to corroborate this conclusion and to better understand the climate system in Central Asia during both this deglaciation and the present time.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.quascirev.2014.09.006.

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