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Carbon dynamics of peatlands in China during the Holocene

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ABSTRACT

Understanding the responses of the carbon-rich peatland ecosystems to past climate change is crucial for predicting peat carbon fate in the future. Here we presented a data synthesis of peatland initiation ages, area changes, and peat carbon (C) accumulation rate variations in China since the Holocene, along with total C pool estimates. The data showed different controls of peatland expansion and C accumulation in different regions. The peat C accumulation rates were 32.3 (ranging from 20.7 to 50.2) g C m^{-2} yr^{-1} in the Qinghai-Tibetan Plateau (QTP) and 14.7 (ranging from 7.4 to 36.5) g C m^{-2} yr^{-1} in the Northeast China (NEC). The peaks of peatland expansion and C accumulation in the QTP occurred in the early Holocene in response to high summer insolation and strong summer-winter climate seasonality. The rapid peatland expansion and maximum C accumulation rate in the NEC occurred in the middle-late Holocene. Peatlands scattered in the coastal and lakeside regions of China expanded rapidly at the onset of the Holocene due to large transgression, consistent with the stronger summer insolation and monsoon, and during the middle and late Holocene, as a response to the high and stable sea level and the strong summer monsoon. The carbon storage of peatlands in China was estimated as 2.17 (ranging from 1.16 to 3.18) Pg, among which 1.49 (ranging from 0.58 to 2.40) Pg was contributed by peatlands in the QTP, 0.21 (ranging from 0.11 to 0.31) Pg by those in the NEC, and 0.47 Pg by those scattered in other regions of China. Our comparison of peatlands dynamics among regions in China showed that climate and monsoon are the essential factors in determining the expansion and carbon accumulation patterns of peatlands, although their effects on peatland formation and C accumulation is complex owing to land availability in peatland basins and regional moisture conditions.

1. Introduction

Though covering only 3% of the land surface, northern peatlands currently represent nearly one-third of the global soil carbon (C) pool (Gorham, 1991; Yu et al., 2010). Undisturbed peatlands have acted as persistent carbon sinks over thousands of years, mainly due to their low decomposition rates under the waterlogged and extremely cold soil conditions (Gorham, 1991; Yu, 2011). In this regard, the C sequestration capacity of peatlands is very sensitive to the alterations of temperature and precipitation, with the C balance easily challenged under the projected climate change (Martini et al., 2007). In order to analyze C dynamics and their feedback to global climate change, it is pivotal to have a detailed understanding of large-scale controls over peatlands expansion and C accumulation in different regions. However, several recent syntheses about

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peatland development histories during the Holocene have been confined to northern and tropical peatlands (Turunen et al., 2002; Smith et al., 2004; MacDonald et al., 2006; Page et al., 2006; Tarnocai, 2006; Tarnocai and Stolbovoy, 2006; Yu et al., 2009, 2010; Jones and Yu, 2010). Data of peatlands distributed in other regions are needed to achieve a more comprehensive understanding of peatland histories and climate controls.

Since peatlands in China are affected by Asian summer monsoon, their accumulation histories would provide insight into climate controls of peatland development under monsoon climates (Zhao et al., 2014). Peatlands in China cover an area of about 10,441 km^2 and store about 4.69 Pg peat (Yin, 1991), distributing through the climate domain from −5 to 18 °C in mean annual temperature and from 200 to 1600 mm in mean annual precipitation (Zhao et al., 2014). Peatlands on the Qinghai-Tibetan Plateau (QTP) and in the Northeastern China (NEC) are two representative ones (Chai, 1981; Ma et al., 1987). Most of the peats have been accumulated in the Holocene (Xiang et al., 2009). The thickness of peats ranges from 0.5 to 10 m on the QTP (Sun, 1992; Sun et al., 2001) and from 1 to 2 m in the NEC (Ye et al., 1983; Xia, 1988). Peatlands scattering in the coastal and lakeside regions began to form since the early Pleistocene and expanded rapidly during the Holocene (Ma et al., 1987). Evidences showed that climate in the Holocene varied frequently and significantly in China (Gasse et al., 1991; Zhao and Yu, 2012; Guo et al., 2013), making carbon accumulation of peatlands significantly variable (Yang, 1990). Therefore, studying carbon dynamic of peatlands in China is necessary for understanding their roles in regional carbon cycle for a long-time span.

So far much work about peatlands in China was focused on the formation and distribution patterns, mainly for the usage of peat resources (Chai, 1981; Ma et al., 1987). With ubiquitous warming and intensified anthropogenic activities, many studies tried to determine the peat carbon density and peat carbon storage (Wang, 2009; Wang et al., 2012; Liu et al., 2012b), and the accumulation rate of peat and carbon (Sun et al., 2001; Bao et al., 2010). They also tried to identify the influencing factors for peat accumulation (Large et al., 2009; Gao et al., 2010; Zhou et al., 2010). Very recently, a synthesis effort by Zhao et al. (2014) did well in providing a systematic knowledge of climate controls of peat expansion in the last 50 ka, the peat C accumulation rates based on only six peat cores, however, could not represent the C accumulation dynamic patterns on either regional or national scale. Therefore, we still lack a fundamental and comprehensive understanding of peat C accumulation and its climate controls in different regions of China. Moreover, the estimates of the size of peat carbon pool are far from accurate, varying from 0.6207 to 2.085 Pg (Xie, 2004; Yu et al., 2005; Wang et al., 2012; Liu et al., 2012a). Thus, re-estimating the peat carbon pool in China is necessary for understanding the role of China's peatlands in regional C cycling. In this paper, we present a synthesis about the inception age and C accumulation data of peatlands in China. We also discuss the broad-scale controls of peatland dynamics in different regions in China since the Holocene. The objectives of this paper are (1) to present the database of peatland initiation and area change of peatlands in China since the Holocene; (2) to document C accumulation variations and associated broad-scale controls in different regions of China including QTP, NEC and other regions; and (3) to estimate the carbon pool of China's peatlands by using a new calculating method based on accumulation rate and peatland area change.

2. Data sources and data analysis

The latest information available of individual sites served as the original data sources for the peatland map (Fig. 1) (see Table S1 of Text S1 in Supplementary material). These peatland data sets are available in raster digital formats as well as text formats. First, we collected the needed information including peat basal date, location site, latitude and longitude from available published sources and our own newly collected, unpublished data. Then, we classified all the peat basal data as different groups of 0–3 ka, 3–6 ka, 6–10 ka, and >10 ka. Finally, we plotted all the peat basal date information on the peatland map of China where circles with different colors represented peatlands with different peat basal dates.

The radiocarbon-dated (^{14}C) ages of basal peat, indicating the onset of peat-accumulating conditions, were derived from original published sources for peatlands of China (see Table S2 of Text S1). We chose 260 and 206 ^{14}C peat basal dates for QTP and NEC, respectively, according to the following selection criteria: (1) the peat depth should exceed 30 cm; (2) peat profiles were mostly continuous peat; (3) only the oldest date from peat base was used from a single peatland basin as initiation age.

For QTP and NEC, all ^{14}C dates were converted to calibrated dates by using the IntCa104 datasets (Reimer et al., 2004). We added the number of dates within 2-σ range (95% probability) of calibrated ages at 10-year intervals to construct the frequency histograms. For other regions of China, the frequency histograms were constructed by adding the number of basal dates of 0–3 ka, 3–6 ka, 6–10 ka and >10 ka. The frequency histograms were then added to calculate cumulative percentages. To estimate the apparent C accumulation rates, we averaged the rates in each 1000-year bin for each region (QTP, Zoige, NEC and Sanjiang Plain). The time-weighted rates for each available site in 1000-year were calculated by the following equation (Lahteenoja et al., 2012) based on raw data including multiple calibrated ages, bulk density and carbon content measurements.

$$\mathrm{CA} = r/1000 \times n \times c$$

where CA is C accumulation rate (g m^{-2} yr^{-1}); r is peat accumulation rate (mm yr^{-1}); n is dry bulk density (g m^{-3}); c is C content (g C g^{-1} dry weight). Among them, peat accumulation rate (mm yr^{-1}) of 1000-year bin for each profile was calculated based on the dated depths of the profile; the dry bulk density and C content, for peatlands in the QTP, were calculated based on the measurements of seven peat profiles (Chen et al., 2014); for peatlands in NEC, the average dry bulk density and C content were 0.20 g m^{-3} (Huang et al., 1988) and 12.7% g C g^{-1} dry weight (Huang, 1998; Liu et al., 2012b) respectively, based on all the values available. The C accumulation rate based on the method might not be accurate for a single peat profile since the bulk density and C content varied significantly in different peat profiles; however, it was proper to estimate the C accumulation rate of the whole region.

The peat apparent C accumulation rates for QTP were based on 20 sites, including 15 sites from Zoige, and for NEC on 14 sites, including 8 sites were from Sanjiang Plain. The rates of area change for each peatland region (Zoige, other sites on QTP, NEC and Sanjiang Plain) were calculated based on the frequency histograms and total area of each region.

To estimate changes in peat C pools at 1000-year intervals in different regions (Zoige, other sites on QTP, NEC and Sanjiang Plain), we multiplied the cumulative peatland area (which was calculated based on cumulative basal age and the total area for each region at that time) by C accumulation rates for each 1000-year bin. The C-pool intervals were added up to obtain cumulative C pools. Only the standard errors of the mean C accumulation rates were taken into consideration to generate the C pool ranges, ignoring other uncertainties in basal ages, possible non-linear peatland expansion, and peatland areas.

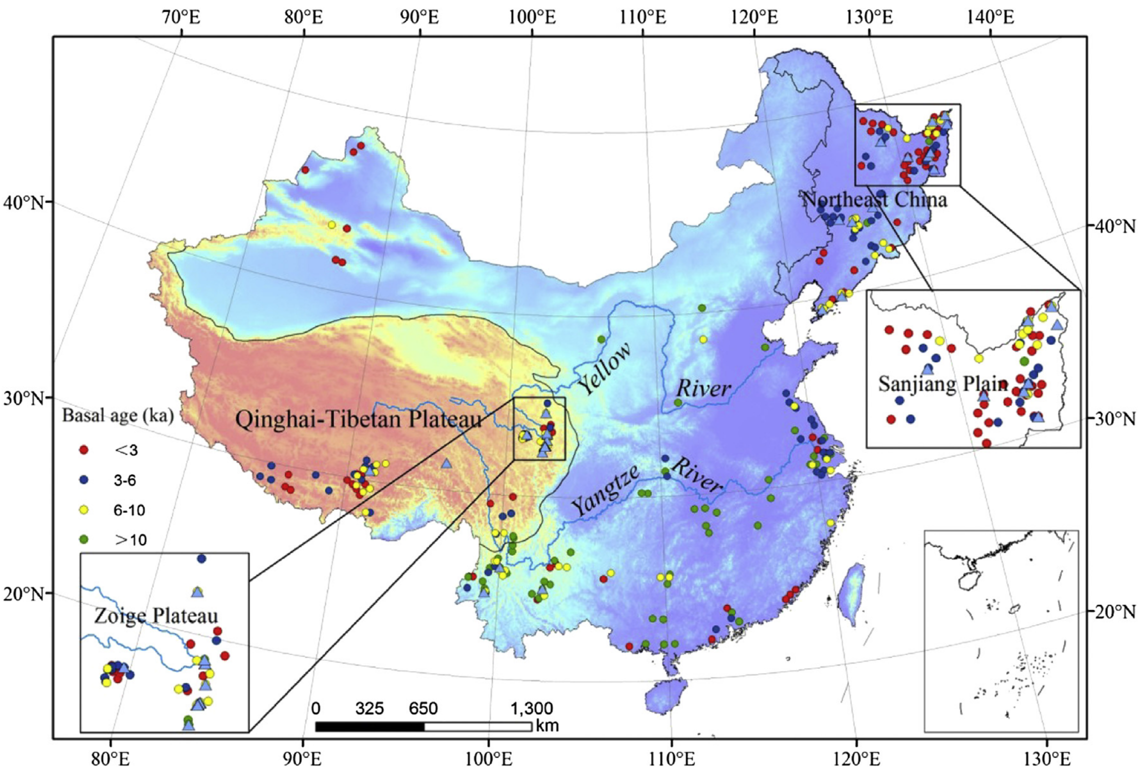

Fig. 1. Chinese map of peatland regions and peatland study sites with basal peat ages. Small dots; colors showing the ages of peatland initiation. Red: <3 ka; blue: 3–6 ka; yellow: 6–10 ka; green: >10 ka). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

The total area of QTP is estimated to be 5086 km^2, among which about 4600 km^2 (90%) distribute in Zoige (Sun and Zhang, 1987) and 486 km^2 scatter in other regions of QTP (Song et al., 1985; Sun et al., 1998; Liu et al., 2012b). We used 5086 km^2 in our calculations of peatland area and peat C pool changes in QTP. We calculated peatland area of NEC as 2050 km^2, including 350 km^2 in the Sanjiang Plain (Huang et al., 1988; Ma, 2013). When calculating peat C pool and area change, we used the peatland areas from literatures and books instead of those from the new peatland map, since the latter ignored those peatlands covering less than 5% of the landmass, causing underestimation of the peatland area (see Table 1 and the Supplementary material).

3. Results

3.1. Peatland initiation patterns and its carbon accumulation rate

Peatlands from the Qinghai-Tibetan Plateau started to form before 17 ka, with the expansion peak and initiation sharp decrease occurring at around 11.5–10 ka and 13–12 ka, respectively (Fig. 2B). The initiation gap appeared at 9.2–9.5 ka BP. The peat C accumulation rates on the QTP ranged from 20.7 to 50.2 g C m^{-2} yr^{-1}, with a mean value of 32.3 g C m^{-2} yr^{-1}; the peat C accumulation rates in Zoige ranged from 20.7 to 52.5 g C m^{-2} yr^{-1}, with a mean value of 33.4 g C m^{-2} yr^{-1}; the peat C accumulation rates of other regions scattered on the QTP ranged from 12.6 to 32.9 g C m^{-2} yr^{-1}, with a mean value of 20.8 g C m^{-2} yr^{-1}. Overall, the peat C accumulation rate on the QTP showed a peak in the early Holocene at 7–9 ka BP (Fig. 2C). Peatlands in the NEC began to form around 17 ka, with about 80% initiating after 10 ka BP (Fig. 2D). Three initiation peaks of peatlands in the NEC were found at around 3–1 ka, 6–4 ka and 11.2–10.5 ka. Before 6 ka, the C accumulation rates were very low, with the lowest rates occurring at 8 ka and then increased gradually from 6 ka to 3 ka, followed by an abrupt increase over the past 2 ka. The observed peak of peat C accumulation rate was 45.1 g C m^{-2} yr^{-1} in Sanjiang Plain and 36.5 g C m^{-2} yr^{-1} in NEC at 1ka, five and three times higher than the average value of other time periods, respectively (Fig. 2E). The overall average rates were 13.3 (ranging from 4.5 to 45.1) g C m^{-2} yr^{-1} in Sanjiang Plain and 14.7 (ranging from 6.6 to 36.5) g C m^{-2}yr^{-1} in NEC (Fig. 2E). Nearly 40% of peatlands scattered in other region of China formed before 10 ka BP and the other 60% initiated after 10 ka BP. Among those formed after 10 ka BP, only 22% of peatlands initiated at 6–10 ka BP and about 39% at 3–6 ka BP and 39% at 0–3 ka BP (Fig. 4A). It is worth mentioning that almost no peatlands expanded in QTP and NEC during the past one or two centuries (Fig. 2B, D).

Table 1
Summary Results of peatlands in China.

		Area (km^2)	C Pool (Pg) (Range)	Holocene C rate (g C m^{-2} yr^{-1})
QTP	Zoige	4600	1.42(0.56–2.27)	33.4
	Other region	486	0.07(0.02–0.13)	20.8
		5086	1.49(0.58–2.40)	32.3
NEC	Sanjiang Plain	350	0.03 (0.02–0.05)	13.3
	Other region	1700	0.18 (0.09–0.26)	15.0
		2050	0.21(0.11–0.31)	14.7
Other region of China		3290	0.47	

3.2. Carbon pool and area change rate

During early Holocene, the rate of area change pattern of peatlands in the QTP was very similar to that in the NEC. Although the overall rate of area change was higher than that of peatlands in the NEC, there appeared an opposite trend of area change rate between the two sites at the period of 3–3.5 ka BP and 8.5–9 ka BP (Fig. 3B). Based on the area change over time from basal age frequencies and average C accumulation rates, we estimated peat C pools as 1.49 Pg (ranging from 0.58 to 2.40 Pg) for QTP peatlands, including 1.42 Pg

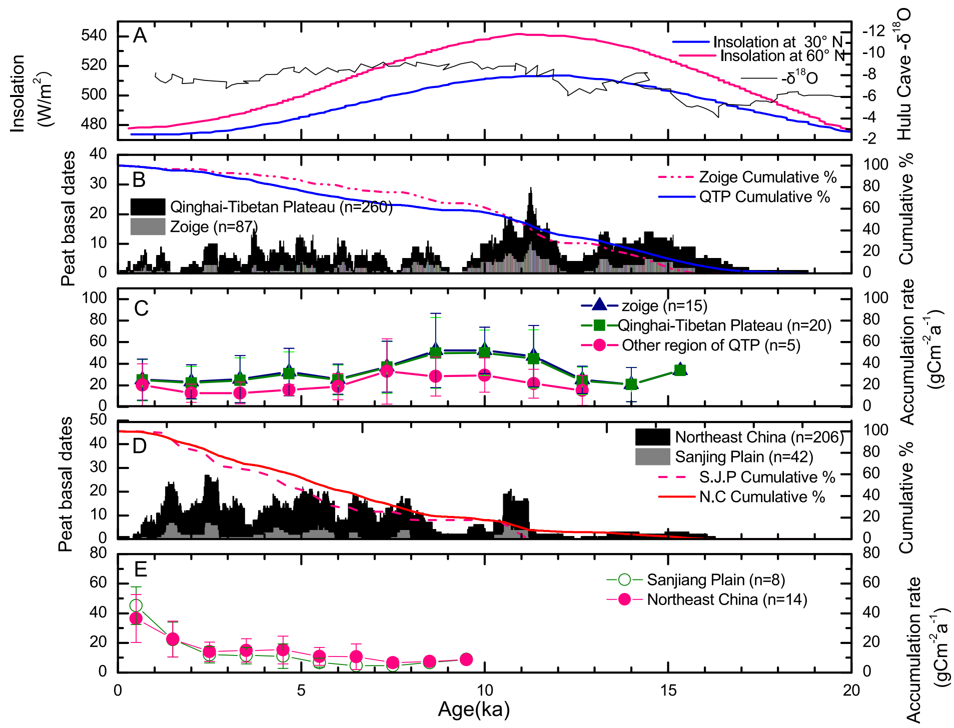

Fig. 2. Chinese peatland records since the Holocene. (A) Summer insolation at 30 °N (black curve) and at 60°N (red curve) (Berger and Loutre, 1991b), and the oxygen isotope record at Hulu Cave (Wang et al., 2001). (B) Peat basal ages plotted as calibrated age frequency (bars) and cumulative percentage (smooth curve) for peatlands in the Qinghai-Tibetan Plateau ($n = 173$) and Zoige ($n = 87$); (D) for peatland in the whole Northeast China ($n = 212$) and Sanjiang Plain ($n = 48$). (C) Average peat carbon accumulation rates (g C m^{-2} yr^{-1}) at 1000-year bins (showing means of the bins from various sites in a region and standard errors of the means) for peatlands in the Qinghai-Tibetan Plateau ($n = 20$) and Zoige ($n = 15$); (E) all peatlands from Northeast China ($n = 14$) and Sanjiang Plain ($n = 8$). See Text S1 for the sources and references of peatland data. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

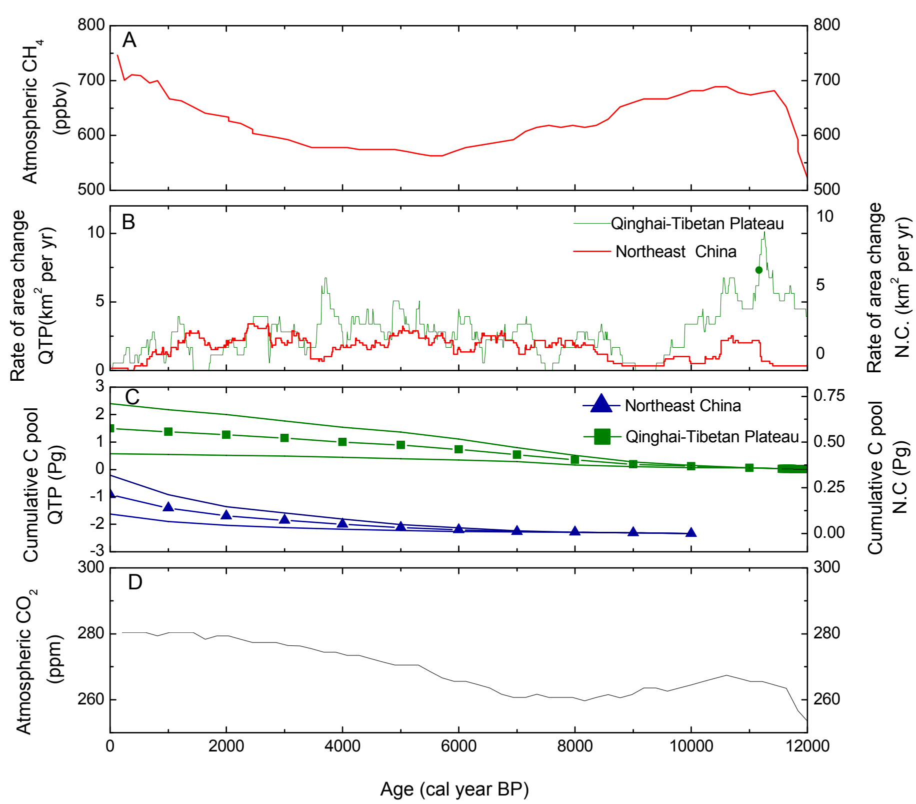

Fig. 3. Implications of Chinese peatlands for the global carbon cycle during the Holocene. (A) Atmospheric CH_4 concentration (Dällenbach et al., 2000); (B) rate change of peatland area increase (km^2 per year) for peatlands in the Qinghai-Tibetan Plateau and Northeast China, based on cumulative basal age frequencies as in Fig. 2, assuming that individual peatlands expanded in their area linearly since their formation; (C) temporal change in (observed) cumulative carbon pools of peatlands from the Qinghai-Tibetan Plateau and Northeast China, based on peatland area estimates (Fig. 3B) and carbon accumulation rates (Fig. 2C, E) for different regions; and (D) atmospheric CO_2 concentration (Schilt et al., 2010).

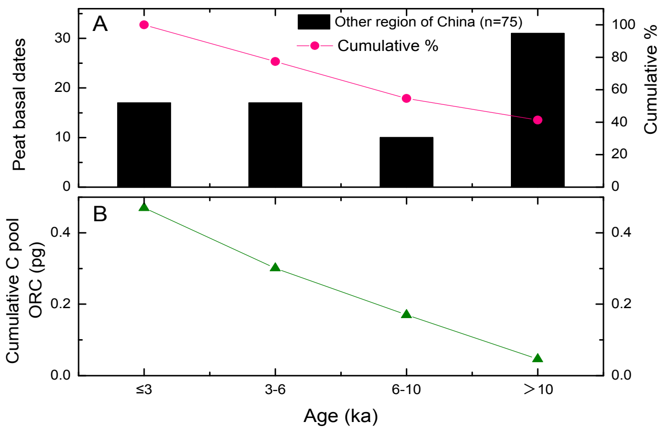

Fig. 4. Peatland initiation and cumulative C pool of other region of China. (A) Peat basal ages plotted as age frequency (bars) and cumulative percentage (smooth curve) for peatlands in other region of China ($n = 75$) and its temporal change in (observed) cumulative carbon pools (B).

(ranging from 0.56 to 2.27 Pg) in Zoige peatlands, and 0.21 Pg (0.11–0.31) for NEC, with 0.03 Pg (0.02–0.05) stored in Sanjiang Plain peatlands. The amount of C pool in other region of China was estimated to be 0.47 Pg based on previous studies (Liu et al., 2012a; Ma, 2013). Therefore, the total carbon pool of peatlands in China is 2.17 Pg, ranging from 1.16 to 3.18 Pg.

4. Discussion

4.1. Patterns and controls of peatland expansions in China since the Holocene

The rapid expansion of peatlands occurred at 11.5–10 ka on the QTP and 11.2–10.5ka in the NEC (Fig. 2B, D), in line with the result of a previous study that northern peatlands in China expanded rapidly during the early Holocene (Zhao et al., 2014). It was suggested that the expansion peak of global northern peatlands in the early Holocene was mainly attributable to the peak summer insolation and strong climate seasonality (Zhao and Yu, 2012), since high plant production and low peat decomposition could be achieved in this climate of warm summer and cold winter (Jones and Yu, 2010). This mechanism can also explain the early-Holocene initiation peak of peatlands in this study. The general decreases in peatland expansion and C accumulation on the QTP in the mid- and late-Holocene, also similar to the trend of global northern peatlands, were probably a response to the decrease in summer insolation and the subsequent cooling climate after the Holocene Thermal Maximum (MacDonald et al., 2000; Kaufman et al., 2004; Zhao et al., 2011), since low temperature inhibits plant growth and subsequently limits organic matter input (Herzschuh et al., 2006). However, inconsistent with the summer insolation variation (Fig. 2A), peatlands in the NEC expanded rapidly in the middle and late Holocene, with additional initiation peaks occurring at 3–1 ka and 6–4 ka, indicating that the effect of cooling induced by decrease in summer insolation on peatland expansion was complex (Korhola et al., 2010). Moreover, despite the continuous increase in summer insolation from 20 to 10 ka, an obvious decrease in peatland expansion occurred at 13–12 ka in the QTP and the NEC. This was probably due to the sharp decrease in summer monsoon as indicated by Hulu Cave record (Fig.2A), suggesting that besides insolation, monsoon is another important factor in determining the expansion patterns of peatlands. A large portion of peatlands scattered in China besides the QTP and the NEC, mainly distributed in coastal and lakeside regions, formed before 10 ka [Fig. 4; (Ma et al., 1987)], corresponding to large transgression in the comparatively high and increasing insolation (Berger and Loutre, 1991) and strong summer monsoon as indicated by theδ^{18}O record at Hulu cave [Fig.2A; (Wang et al., 2001)]. Moreover, peatlands also expanded rapidly during the middle and late Holocene (Fig. 5), probably in response to the high and stabilized sea level and strong summer monsoon (Chai, 1981; Zhong and Zhang, 1981; Ma et al., 1987). Stable sea level can favor the growth of plants and subsequent organic matter input (Moy et al., 2002), which is very important for peatland development. Similarly, it was reported that lower summer insolation in the mid-and late Holocene corresponds with relatively higher frequency of peatlands initiation in subtropical China (Zhao et al., 2014).

4.2. Carbon accumulation dynamics during the Holocene

Different from the peatland initiation peak at 11.5–10 ka, the C accumulation rate peak occurred at 9–7 ka on the QTP, indicating that the mechanisms controlling peatland initiation and peat C accumulation were probably different. The observed peak of peat C accumulation rate at 1 ka BP in the NEC (Fig. 2D), with five and three times higher than the average value of other time periods in Sanjiang Plain and in the NEC, respectively. It would be partly explained by less decomposition of the newly formed peats in the past 1ka, because younger peat layers have been subjected to peat mineralization for a shorter time than their older counterparts (Clymo, 1984). According to the study of Korhola et al. (1995), the true rate of C accumulation was usually only 70% of the apparent long-term rate (Korhola et al., 1995). In addition, although the climate became drier and colder in the late-Holocene, it was still characteristic with wet and cold condition in the NEC (Zu, 1973; Xia, 1988; Yang, 1990). The cold and wet climate resulted in low decomposition rate of peat organic matters, thereby benefiting peat accumulation. Besides, the stable sea level resulting from the relatively cold and dry climate in the late-Holocene was another important contributor to the rapid C accumulation (Xia and Wang, 2000).

For the QTP, we estimated its peat C accumulation rate as 32.3 g C m^{-2} yr^{-1}, higher than that of 20.4 g m^{-2} yr^{-1} in our recent paper (Chen et al., 2014), which was based on multiple dating data from seven peat sites. The difference among results of different researches indicated that uncertainty still exists in the estimated peat C accumulation rate on the QTP. To our knowledge, our estimation of peat C accumulation rate of 14.7 g C m^{-2} yr^{-1}is the first for peatlands in the NEC. Compared with other places, the peat C accumulation rate of the QTP and NEC, respectively, was much higher and comparable to that of Northern peatlands (18.6 g C m^{-2} yr^{-1}), tropical peatlands (12.8 g C m^{-2} yr^{-1}) and Southern peatlands (22.0 g C m^{-2} yr^{-1}), suggesting a strong carbon accumulation in the alpine peatlands on the QTP (Yu et al., 2010).

4.3. Carbon dynamics of peatlands in China and their role in the carbon cycle of terrestrial China

In this research using a new approach based on peatland initiation ages and C accumulation rates, our estimate of C pool in peatlands of China was 2.2 Pg, comparable with that of 2.1 Pg (Wang et al., 2003), but higher than that of 0.6 Pg (Xie, 2004), 0.8 Pg (Yu et al., 2005) in previous studies. One main reason of the relatively low value in the previous studies was probably that they ignored the carbon stored in buried peatlands and used 1 m peat soil depth in calculations (Xie, 2004; Yu et al., 2005), causing underestimation of the C storage in peatlands. Moreover, our estimation was also higher than that of 1.5 Pg which was based on

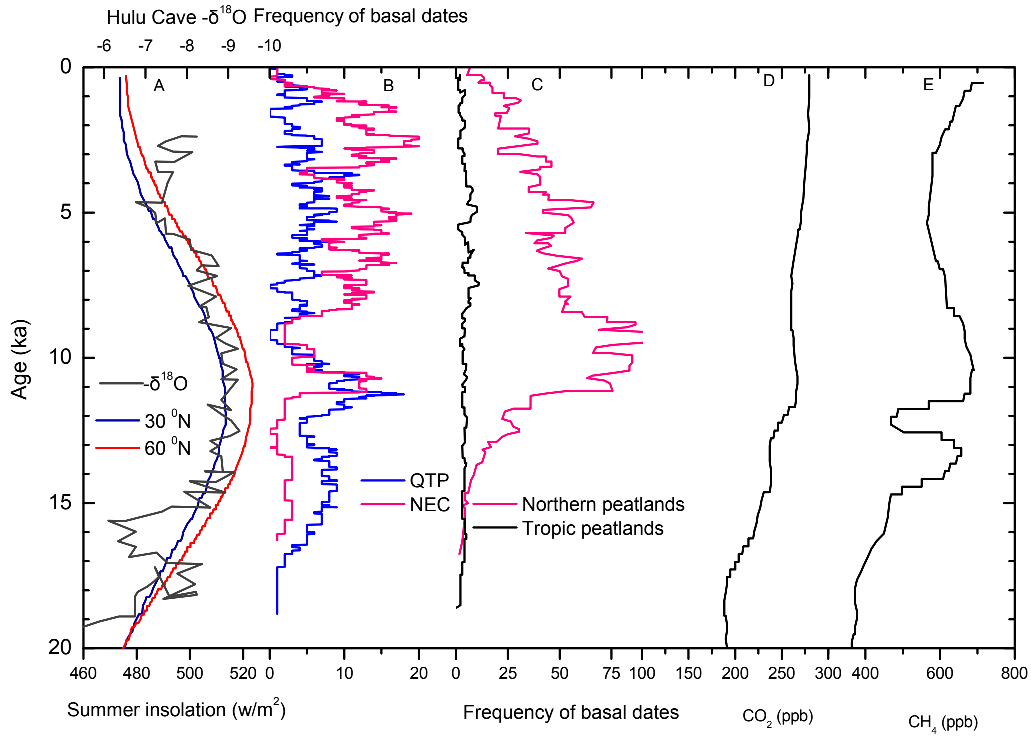

Fig. 5. China's peatland initiation frequency, climate controls and correlations with other palaeorecords. A. Summer insolation at 30 °N and 60 °N (Berger and Loutre, 1991b), and the oxygen isotope record at Hulu Cave (Wang et al., 2001). B. Peat basal ages plotted as calibrated age frequency (bars) and cumulative percentage (smooth curve) for peatlands in the Qinghai-Tibetan Plateau ($n = 173$) and the northeast China ($n = 212$). C. Frequency of basal dates from northern peatlands (MacDonald et al., 2006) and tropical peatlands (Yu et al., 2010); D. Atmospheric CH_4 concentration (Dällenbach et al., 2000); E. Atmospheric CO_2 concentration (Schilt et al., 2010).

peatland area, mean peat depth and mean bulk density (Liu et al., 2012b). The difference mainly stemmed from the higher estimation of 1.3 Pg C for Zoige peatlands in our study, much higher than that of 0.6 Pg C (Liu et al., 2012b) and 0.5 Pg C (Chen et al., 2014).Similarly, Yu et al. (2010) used this new method to estimate northern peat C pools as 547 Pg C, higher than previous estimates of 270 Pg C (Turunen et al., 2002) and 450 Pg C (Gorham, 1991). In calculation of peat C accumulation only that during the Holocene was considered, since the C accumulation rate before 12 ka was not available. Assuming the peat C accumulation rate before 12 ka was the same with that of the Holocene, the peat C storage should be 0.12 Pg C more on the QTP and 0.007 Pg C more in the NEC.

Methane (CH_4), with a global warming potential 25 times that of carbon dioxide on a per mole basis, has a significant impact on the earth's climate system (Solomon, 2007). The total emissions of CH_4 from rice paddies, natural wetlands and lakes in China are 11.25 Tg CH_4 yr^{-1}, nearly 24% of which is contributed by natural wetlands (Chen et al., 2013). In China, peatland is one important component of natural wetland and has a relatively high CH_4 emission rate of 6.46 ± 6.60 mg CH_4 m^{-2} h^{-1}(Chen et al., 2013). Therefore, the dynamics of peatlands may partly explain the trend of CH_4 concentration in China. In particular, changes in peatlands area should be a primary index for peatland CH_4 emission potential. The high atmospheric CH_4 concentration during the early Holocene (Yao and Xu, 2005), coinciding with the rapid expansion of peatlands on the QTP and in the NEC induced by summer insolation, indicated that peatlands in northern China plays an important role in the CH_4 budget during the early Holocene (Fig. 3B). Moreover, a modeling study indicated that enhanced temperature alone could double the CH_4 fluxes from northern wetlands during interstadials (Van Huissteden, 2004). This suggested that warming induced by summer insolation can increase atmospheric CH_4 concentrations directly or through promoting peatland expansion. Global synthesis suggested that northern peatlands played a foremost role in CH_4 budget during the early Holocene, but their tropical counterparts became more important in the mid-Holocene around 8–4 ka (Yu et al., 2010). The lowest atmospheric CH_4 concentration around 5 ka was due to the decrease in peatland expansion from both northern and tropical regions. The general declining trend of CH_4 during the period of 11–5 ka might be caused by the monsoon intensity in China, as inferred from the Hulu cave record (Wang et al., 2001). During the late-Holocene, the atmospheric CH_4 concentration has risen abruptly and anomaly, despite the decreasing expansion of peatlands in China, probably due to the increase of livestock and human waste emissions, large emissions from early rice irrigation and climate feedback. Similarly, people observed a slowdown in the rate of expansion of global natural peatlands after 4 ka, but an increase in atmospheric CH_4 concentration during that period (Yu et al., 2010), supporting that human activities account for the late-Holocene increase in CH_4 concentration (Ruddiman, 2007).

Global peatlands stored a large amount of carbon which can significantly affect the global C cycling (Joos et al., 2004; Elsig et al., 2009). However, the contribution of peatlands to atmospheric CO_2 concentration still cannot be assessed based on the cumulative peat C pool (Yu et al., 2010). A further analysis of partitioning the time-variant net C uptake and C release would provide insight to assessing the role of peatlands in global carbon cycling during the Holocene (Yu, 2011). In China, the peat C pool (2.2 Pg) was much less than that in grasslands (44.09 Pg) (Dai et al., 2009) and forests (28.12 Pg) (Zhou et al., 2000), which may be not large enough to have obvious impact on the regional C budget. On an area basis, the C accumulation of peatlands in China (193 kg C m^{-2}) was much higher than that of grasslands (12 kg C m^{-2}) (Ni, 2002; Li et al., 2004; Xie et al., 2007; Fang and Chen, 2011) and forests (11 kg C m^{-2}) (Liu et al., 2011). In this regard, peatlands in China are indeed an important factor in modifying and alleviating the regional carbon cycle. Despite great variations of peat C sink in China over time, our analysis here shows that peatlands have accumulated 2.2 Pg C over the Holocene, serving as a long-term C sink of nearly 20 Teragram (1 Tg = 10^{12} g) per century. However, because of their fragility and sensitivity, peatlands may turn from long-time carbon sink to carbon source in the context of global warming and intensive human activities.

5. Uncertainties and conclusions

Our synthesis for China indicated that peatlands initiated mostly during the Holocene, although some peatlands did start to form in the MIS3 (mostly at 45–30 ka) and the Bolling-Allerod period in subtropical China (Zhao et al., 2014). The initiation patterns of peatlands were different among different regions. On the QTP, the expansion peak of peatlands and maximum C accumulation occurred at the early Holocene, followed by a general decrease in the middle-late Holocene, suggesting that insolation and monsoon may have played an important role in peat C accumulation on the QTP. For the NEC and other regions in China, however, the rapid formation and C accumulation of peatlands occurred at the middle and late Holocene, which seemed inconsistent with summer insolation but partly in line with summer monsoon variations. The carbon storage of peatlands in the QTP was nearly 7 times higher than that in the NEC, not only due to the larger peatland area but also their higher peat C accumulation rate, indicating that peatlands in the QTP are the most important component of peatlands in China.

The new calculation method used in this research to estimate the peat C pool in China could avoid the uncertainties resulting from the sampling peat depth in previous studies. However, our own estimation still had some uncertainties. Firstly, the dating data used to calculate peat accumulation rate was limited and in poor quality. In calculation of peat C accumulation rate, we used 20 and 14 peat cores dates on the QTP and NEC respectively. However, some of these peat cores only had three dating dates, potentially lacking detailed information. Secondly, we did not use the direct bulk density and C content to calculate the C accumulation rate due to limited data available. Due to variation of bulk density in different peat sites, this method could potentially lead to either overestimation or underestimation of C accumulation rate for a single peat core. However, the calculation uncertainty could be diminished at regional level since overestimation for some peat cores could be offset by those underestimated. Overall, this study can provide insights into the climate control over peatlands development and carbon accumulation dynamics in monsoon-influenced regions.

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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.2014.06.004.

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