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A 6000-year record of changes in drought and precipitation in northeastern China based on a $\delta^{13}C$ time series from peat cellulose

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Abstract

We report a new peat $\delta^{13}C$ proxy record for humidity or precipitation based on C3 plants from northeastern China. The record reveals two times of significant climate shift and eight severe drought periods during the past 6000 years, all of which have the nature of widespread global occurrence. The variability of precipitation shows periodicities of around 70, 80, 90, 107, 110, 123, 134, 141, 162, 198, 205, 249, 278, 324, 389, 467, 584, 834 and 1060 years. The occurrence and persistent times of drought and periodicities of precipitation show good correspondence with solar variability. The remarkable correlations between peat $\delta^{13}C$, peat $\delta^{18}O$ and atmospheric $\Delta^{14}C$ suggest that on timescales of decades to centuries the changes in drought and precipitation are likely caused by variations of atmospheric circulation and atmosphere–ocean interactions in large-scale patterns that seem to be related to solar variability.

Keywords: drought; solar cycles; climate change; peat; maars; monsoons; Holocene; China

1. Introduction

The stable carbon isotopic compositions ($\delta^{13}C$) of peat have been used as a proxy indicator for relative humidity [1], mainly based on the response of the relative composition of C3 and C4 plants in peat of the tropics to a variation in relative humidity or precipitation. Since the occurrence of C4 plants decreases with an increase in both altitude and latitude [2] (bogs in general are damp places where C4 plants occur less), it may restrain application of this kind of proxy record. Here a new peat $\delta^{13}C$ climate proxy record based on C3 plants has been studied in Jinchuan of northeastern China (42°20′N, 126°22′E), which covers the middle of the Hypsithermal through the Medieval Warm Period to the Little Ice Age. A previous paper reported the relation between Jinchuan peat $\delta^{18}O$ and palaeotemperature [3].

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2. Background on peat $\delta^{13}C$ proxy record for humidity or precipitation

Peat is a mixture consisting of different species of plant remains. Plants with different pathways of photosynthetic CO_2 fixation have different ratios of stable carbon isotopes. It has been confirmed [4–6] that C3 plants utilize the Calvin cycle with the first intermediate of photosynthesis being phosphoglyceric acid, a three-carbon molecule, and their $\delta^{13}C$ values show considerable variation. The majority of $\delta^{13}C$ values reported for C3 plants are around −28 to −26‰. At the lower end of the total range, values in the order of −30‰ and less have been reported, while at the other end values of up to −20‰ have been measured [4]. C4 plants fix carbon initially into four-carbon acids and then via the Calvin cycle into three-carbon molecules, and their $\delta^{13}C$ values measured range from about −16 to −9‰ [4–6]. CAM plants either use the C3 pathway or a pathway similar to that found in C4 plants, and their $\delta^{13}C$ values reported in most cases are in the range of −20 to −10‰ [4–6]. Table 1 shows that the $\delta^{13}C$ values of modern dominant plants forming peat in the Jinchuan peatland range from −28.58 to −21.58‰. In addition, the $\delta^{13}C$ values of the Jinchuan peat entire profile (Fig. 1a) range from −30.12 to −22.76‰, all of which fall in the scope of $\delta^{13}C$ values for C3 plants. So it can be considered that Jinchuan peat consists of C3 plants.

Table 1
Isotopic composition of modern plants in Jinchuan peatland (‰)

No.	Plant name	$\delta^{13}C$[a]
J-P1	*Carex schmidtii*	−23.43
J-P2	*Carex lasiocarpa*	−25.54
J-P3	*Pedicularis manshurica*	−24.28
J-P4	*Picea koyamai*	−24.92
J-P5	*Sphagnum oligoporum*	−28.58
J-P6	*Betula fruticosa*	−25.60
J-P7	*Phragmites communis*	−21.58
J-P8	*Sphagnum acutifolium*	−26.16
J-P9	*Typha latifolia*	−23.76
J-P10	*Breidleria arcuaia*	−26.68
J-P11	*Pleuozium schreberi*	−26.32

[a]Standard deviation of the $\delta^{13}C$ value is better than ±0.1‰ (1σ).

The $\delta^{13}C$ values in C3 plants are significantly different from those of atmospheric carbon dioxide due to processes of isotope fractionation during the incorporation of CO_2 by plants. A widely used quantitative expression for these processes has been introduced [7]. Previous work has shown that $\delta^{13}C$ of C3 plants is sensitive to variations in humidity or precipitation, and $\delta^{13}C$ values are mainly influenced by soil water content or precipitation [4–8]. Therefore the $\delta^{13}C$ time series of tree-rings, generally C3 plants, has been widely used as the proxy indicator for a variation of soil moisture or precipitation [9–14].

Recent investigation has shown that Jinchuan peat developed from a constant catchment basin being a dried maar lake, which is one of the around 10 maar lakes in Northeast China Volcanic Field [15]. In the Jinchuan peat core the *Carex* species of plants, such as *Carex schmidtii* and *Carex lasiocarpa*, dominate the plant remains [3]. The *Carex* species of plants are annual plants and have shallow rooting systems that make them sensitive to and dependent on the variation of soil moisture or precipitation. Although a statistical relationship between relative humidity and other climate parameters, especially temperature, has been reported for C3 plants, such associations have been considered as to be indirect, as a result of the cross-correlation between these other climate parameters and relative humidity [4,8]. The determination of CO_2 trapped in the ice core at the Antarctic has shown that on timescales of centuries to a millennium the concentration of atmospheric CO_2 increased almost linearly around 15 ppmv, and the $\delta^{13}C$ values of atmospheric CO_2 decreased around 0.2‰ over the past 6000 years [16]. So we consider that as a first approximation the influence of variation of both the concentration of atmospheric CO_2 and its $\delta^{13}C$ on the $\delta^{13}C$ values of C3 plants is negligible in our case. Therefore, Jinchuan peat originated mainly from the sedimentation of dead *Carex* species of plants, just like trees annually store climate information in their rings, can annually keep the palaeoclimate information during the growing season in store. The higher the $\delta^{13}C$ in the peat cellulose, the smaller the soil moisture or precipitation, and vice versa.

Mixed α cellulose samples prepared from a Jinchuan peat specimen consisting of a 2 cm thick slice corresponding to about 20 years [3] were loaded into borosilicate tubes together with preheated copper (II) oxide. The loaded tubes were evacuated to ensure that all traces of absorbed water were removed. They were then sealed with an oxy-gas torch and heated in a muffle furnace at 550°C [17]. The resulting CO_2 gases after purification were measured using a MAT-252 mass spectrometer. Values of $\delta^{13}C$ are expressed relative to VPDB standard [18] and the overall precision was found to be better than $\pm 0.1‰$ (1σ). All dates described in this text are based on the calibrated radiocarbon age [3].

3. Drought and precipitation variability over the last 6000 years

The $\delta^{13}C$ time series of Jinchuan peat cellulose is shown in Fig. 1. Three main climate stages inferred from this proxy record can be identified. In the period from around 4000 BC to 2200 BC nearly all of the $\delta^{13}C$ values are lower than the overall mean values of $\delta^{13}C$. Combined with variation of the $\delta^{18}O$ of Jinchuan peat cellulose in the same period [3], we can interpret this as a relatively wet and cold period. During this period, lakes in Europe had persistently high water levels [19] and glaciers throughout the world advanced [20]. However, within this relatively wet and cold

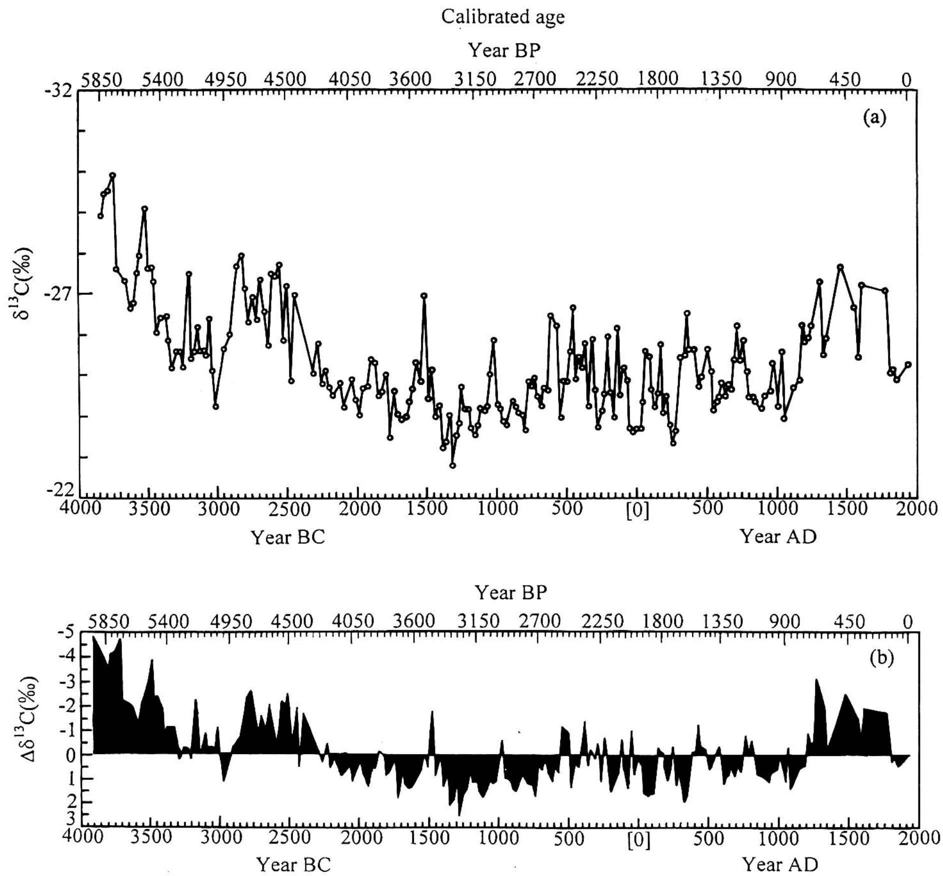

Fig. 1. (a) The $\delta^{13}C$ profile of peat cellulose for the Jinchuan core. (b) It's deviation from the mean of the peat $\delta^{13}C$ time series. The zero line indicates the mean of the peat cellulose $\delta^{13}C$ values for the entire profile. Positive shifts denote lower (than average) soil moisture or precipitation and negative shifts higher soil moisture or precipitation.

interval the $\delta^{13}C$ values also show rapid fluctuations. From around 2800 BC onwards the $\delta^{13}C$ values increase continuously, which may indicate a gradual drying of the regional climate and eventually the occurrence of a drought period since around 2200 BC. Climate shift likely has the nature of widespread global occurrence. It was in the period of climate shift that the lake levels in Europe also decreased [19]. In particular, in the north African and Arabian region the climate changed from a wet to arid phase which led to Saharan desertification, the largest change in land cover during the last 6000 years [21–23].

In the following 3400 years most of the $\delta^{13}C$ values are higher than the overall mean values of peat $\delta^{13}C$, which shows that the history of the local climate variation enters into a relatively dry and warm period because Jinchuan peat $\delta^{18}O$ in the same period is also clearly higher [3]. During the period from around 2200 BC to 1200 AD about eight drought periods can be identified (Fig. 1b). The distribution of them is roughly half before and half after the Christian era. The four drought episodes after the Christian era occurred around 50–150 AD, 250–400 AD, 600–750 AD and 820–1200 AD, respectively. They can also be observed both in the Great Plain and the western USA [24–25], and in Europe [19].

For the four drought periods before the Christian era a limited number of proxy records has been discussed. According to long-term persistence and intensity of drought inferred from peat $\delta^{13}C$, the four drought episodes seem to be the most severe drought periods over the last 6000 years in northeastern China. The first three drought periods (2200–1850 BC, 1820–1550 BC and 1450–1000 BC) correspond to the Xia Dynasty (about 2100–1600 BC) and the Shang Dynasty (about 1600–1000 BC) in the Chinese history, respectively. The dry climate conditions seemed to influence the ancient social-economic activities so much that they stimulated the development of ancient astronomical and meteorological observations. They, together with the act of praying for rainfall, became the most common contents recorded in characters on bones or tortoise shells, one of the oldest scripts in the world. The study of oracle inscriptions discovered in the Yin Xu, the capital of the Shang Dynasty, has shown that in the total 151 pieces of oracle inscriptions of praying for rainfall marked by a definite month, 137 pieces just described the act of praying for rain but without any record of actual rain. Only 14 pieces described rain as an actual result [26]. The fourth drought period from around 950 BC to 550 BC corresponds to the West Zhou Dynasty (about 1000–771 BC) in Chinese history. Severe drought events in this period have often been noted in Chinese historical records [27], for instance, "no rainfall from the second year to the sixth year in King Xuan's time of West Zhou Dynasty" (826–820 BC) and "Luo River and Wei River (two large tributaries of the Yellow River near the Xian city) dried up in the third year of King Mu's time of West Zhou Dynasty" (974 BC). This is also verified by the low lake levels in Europe at that time [19].

After this long and relatively dry and warm period the local climate pattern eventually changed. From around 1200 AD onwards both peat $\delta^{13}C$ and $\delta^{18}O$ [3] decreased rapidly. During the following 600 years all of the $\delta^{13}C$ and $\delta^{18}O$ values are lower than the overall mean values of peat $\delta^{13}C$ and $\delta^{18}O$, respectively, which shows graphically the relatively wet and cold climate during the 'Little Ice Age' (Fig. 1b). The transition of the climate pattern can also be considered as being widespread globally because similar climate changes can also be observed in North America [24–25], Europe [19] and equatorial east Africa [28].

4. Implication for solar variability

The widespread global drought variability is consistent with the assumption of an external global force such as solar force. The history of solar variability can be derived from the ^{14}C content in tree rings [29]. We can see from Fig. 2 that most of the dry and warm periods over the last 6000 years correspond well with stronger solar activity (small $\Delta^{14}C$ values), and the relatively wet and cold periods both before 2200 BC and after 1200 AD correspond well with relatively weaker solar activity (large $\Delta^{14}C$ values). The du-

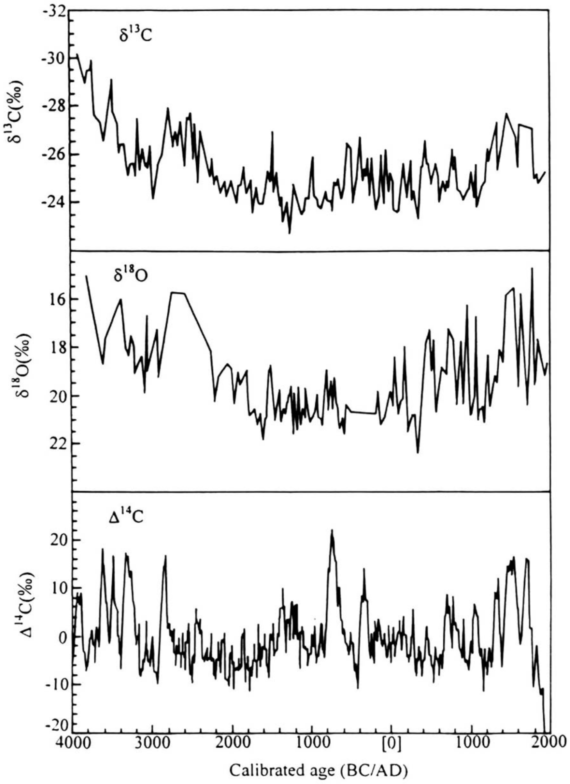

Fig. 2. A comparison between $\delta^{13}C$ and $\delta^{18}O$ of peat cellulose (this study) and $\Delta^{14}C$ of the tree-rings ([31]) plotted against the same timescale.

ration times for both drought and wet periods were broadly coeval with stronger or weaker periods of solar activity, respectively. When solar activity changed from a weak period at around 2200 BC, which was characterized by around six strong peaks of $\Delta^{14}C$, to a stronger period, the climate inferred from peat $\delta^{13}C$ and $\delta^{18}O$ correspondingly completed the transition from the relatively wet and cold to the relatively dry and warm phase. On the contrary, when solar activity changed back from a stronger period at around AD 1200 to a weak period, the climate correspondingly completed the transition from a dry and warm to wet and cold phase.

We have performed spectral analysis on the peat $\delta^{13}C$ time series using the Scargle method for non-equispaced data [30]. The power spectrum shows the periodicities ranging from 70 to 1061 years (Fig. 3). These power spectra are not only very similar to the periodicities of temperature inferred from Jinchuan peat $\delta^{18}O$ [3], but also very similar to the periodicities of solar variability derived from atmospheric $\Delta^{14}C$ [31]. This unified structure in the climate variability provides evidence for a close linkage between solar variability and climate change.

5. Discussion

At the present stage it is difficult to select enough mono-species samples from Jinchuan peat for the determination of isotopes. As a first approximation, however, the influence of different species of plants on the climate signal of $\delta^{13}C$ of mixed peat cellulose seems to be negligible in the case of Jinchuan peat. For instance, at around 3800 BC (corresponding to the 5 m depth sample of the Jinchuan peat core [3]) the $\delta^{13}C$ values of mixed peat cellulose reach a minimum (Fig. 1) though the relative composition of *Phragmites communis* with the highest $\delta^{13}C$ value in peat reaches a maximum at the same time. During the period from around 3800 BC to 3400 BC the $\delta^{13}C$ values of mixed peat cellulose increase with a decrease in the relative composition of *P. communis* in peat. These seem to show that the variation of $\delta^{13}C$ of mixed peat cellulose depends mainly on the variation of $\delta^{13}C$ of the dominating plants in peat, such as *Carex* species of plants, which was caused by a variation in climate. In addition, the investigation of tree-rings showed that a carbon isotope series composed of the $\delta^{13}C$ average of different species of C3 plants seems to display more significant correlations with the climate parameters than any of the individual species series [32]. The question of whether the carbon isotope series of peat mixed cellulose consisting of different species of C3 plants has a similar advantage remains to be further confirmed.

There has been strong evidence that the thermal state of the oceans can lead to drought conditions by inducing perturbations in patterns of atmospheric circulation and the transport of moisture [25,33,34]. The research area is located in the East Asian monsoon area. General circulation models have simulated impacts of the thermal state of the

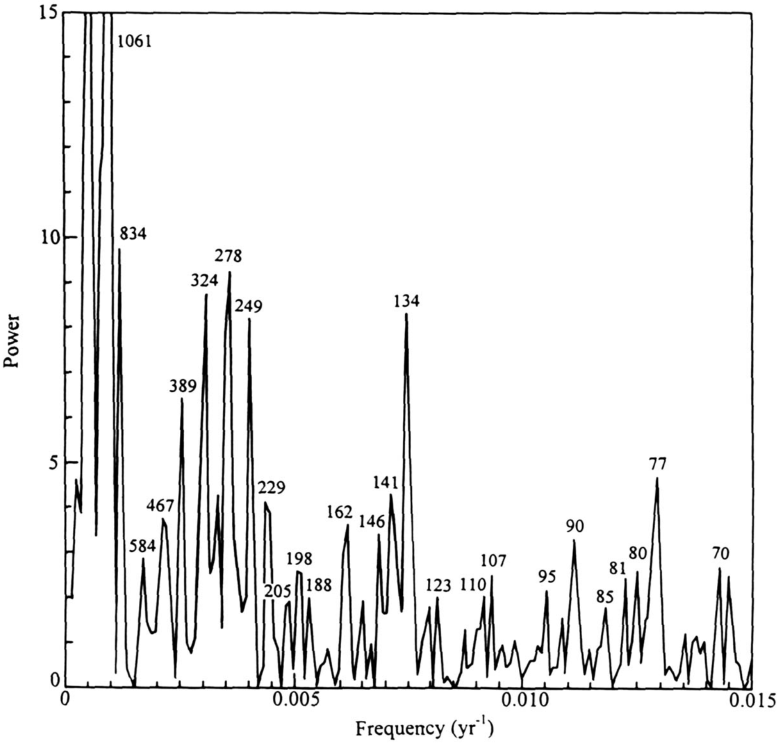

Fig. 3. Power spectrum of the $\delta^{13}C$ time series of peat cellulose from Jinchuan. Numbers above peaks indicate the corresponding periodicities (years).

tropical Pacific Ocean on the Asian summer monsoon and the water cycle in East Asia. The result shows that when the warm seawater is accumulated in the western Pacific warm pool, then the convective activities are intensified from the Indo-China Peninsula to the area around the Philippines. Thus, the western Pacific subtropical high may shift unusually northward, and the East Asian summer monsoon rainfall may be below normal. On the contrary, when the warm seawater extends from the warm pool along the equatorial western Pacific, then the convective activities are weak around the Philippines and are intensified over the equatorial central Pacific near the dateline. Thus, the western Pacific subtropical high may shift southward, and the East Asian summer monsoon rainfall may be above normal [35]. It is the ocean–atmosphere interaction that mainly controls drought and rainfall in the research area. Therefore, based on the variabilities and periodicities of both precipitation and temperature inferred from peat $\delta^{13}C$ and $\delta^{18}O$, respectively, and based on their close relations with solar variability (Fig. 2), there are reasons to suggest that this large-scale pattern of atmosphere–ocean interaction would also have similar variabilities and periodicities with solar activity. Lack of long-term records of sea surface temperature (SST) in the western Pacific Ocean has hindered our efforts to identify physical mechanisms for the correlation, though it has been observed that both variations of SST in Pacific, Atlantic and Indian oceans and global mean SST anomalies all match closely with the envelope of the 11 years cycle of solar activity [36,37].

In addition, there is a complex and nonlinear relation between peat $\delta^{13}C$, $\delta^{18}O$ and atmospheric $\Delta^{14}C$ (Fig. 2). In particular at around 800 BC while the $\Delta^{14}C$ shows as a strong peak, the $\delta^{13}C$ indicates only a small variability of precipitation, although this seems to allow the small decrease of temperature inferred from peat $\delta^{18}O$ at the same time to be explained. This implies that the climate response to solar forcing may also depend on other feedback mechanisms connected with clouds, water vapor, ice-cover, albedo, etc, besides atmosphere–ocean interaction [38–42].

By around 1800 AD peat $\delta^{13}C$ values change from around −27 to −25‰ although anthropogenic activities have introduced more isotopically light CO_2 into the atmosphere, which indicates that the local climate seems to enter again into a dry and warm period. There has been abundant evidence from the climate and hydrological records, as well as information on environmental change, to show a distinct aridity and warm trend in the continent of China since around 1880 AD, and to show an abrupt entrance into the current dry regime around the early 1920s [43]. However, up to 1950 AD the increases in both $\delta^{13}C$ (Fig. 1) and $\delta^{18}O$ [3] have still not surpassed the levels of $\delta^{13}C$ and $\delta^{18}O$ in historical drought periods covering the 'Medieval Warm Period', although the influence of the current drought on public life and economy seems to be enlarged due to an expanded population with stronger industrial and agricultural activities than ever. It seems to show that the current aridity and warm trends will persist likely for a considerably long time which should be taken into account, in particular, when working out strategies for the use and development of water resources in arid and sub-arid areas for an increasing population.

6. Conclusion

1. The $\delta^{13}C$ time series of Jinchuan peat cellulose consisting of C3 plants has been considered as a new climate proxy record which is based on the response of carbon isotope fractionation of C3 plants in peat to the variation in relative humidity or precipitation.
2. The climate proxy record shows on timescales of decades to centuries two times of significant climate shift and eight severe drought periods during the past 6000 years in northeastern China. The first significant climate shift occurred at around 2200 BC when the climate changed from a long-term wet and cold period to a long-term dry and warm period. The second occurred at around 1200 AD when the climate changed from a long-term dry and warm to a long-term wet and cold period. The eight drought episodes are occurred around 2200–1850 BC, 1820–1550 BC, 1450–1000 BC, 950–550 BC, 50–150 AD, 250–400 AD, 600–750 AD and 820–1200 AD, respectively. These climate events have the nature of widespread global occurrence.
3. On timescales of decades to centuries, the climate events including two times of significant climate shift and a series of drought episodes are strikingly correlative to the changes in solar irradiation and solar periodicities. This good correlation between precipitation, temperature and solar activity provides further evidence for solar forcing of the climate variability during the Holocene.
4. Since around 1800 AD the climate seems to enter again into a dry and warm period although up to 1950 AD the drought strength inferred from peat $\delta^{13}C$ has not surpassed the levels in historical drought periods. It seems to show that the current aridity and warm trends will persist likely for a considerably long time, which should be taken into account.

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