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The Palaeoclimate Variations in the Central Plains Since the Interstade of the Last Glacial Stage

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Abstract Through the study of a high–resolution loess record in the Central Plains, the short–term palaeoclimatic variations since the interstade of the last glacial stage have been discussed in this paper. The palaeoclimate in the East Asian monsoon areas shows different variation patterns in summer and winter. A correlation of the palaeomonsoon records of loess with the $\delta^{18}O$ records of the ice core and deep sea is made, and some of the causes for their differences are also discussed.

Key words: palaeoclimate changes, the Central Plains, last glacial stage

1 Introduction

The high–resolution loess record is one of the most important data for the study of global changes. Situated in the Central Plains, at the transitional zone of the North China plain and the loess plateau, the Mangshan *yuan* (loess platform) is formed from loess strata. The upper part of the loess section here, the Malan loess, represents the fastest deposition even found, and has provided valuable geological evidence for palaeoclimatic changes in the East Asian monsoon areas.

2 Stratigraphical Sequence

We have made a thorough investigation of the loess stratigraphy in the area east of the Sanmenxia Gorge. The Mangshan loess lies northwest of Zhengzhou City, on the south bank of the Yellow River (Fig. 1). The stratigraphical division of the Malan loess of the last glacial stage at Taohuayu has been performed with the detection of the signature of the Heinrich events, and the variations and interrelations of the summer and winter monsoons have been studied (Wu et al., 1995; Xiao et al, 1996; Xiao et al., 1997).

The Zhaoxiayu section in the Mangshan *yuan* (34° 58′N, 113° 22′E) discussed in this paper is about 15 km west of the Taohuayu section, with similar stratigraphic features. Samples from the section have been taken at an interval of 5 cm for magnetic susceptibility

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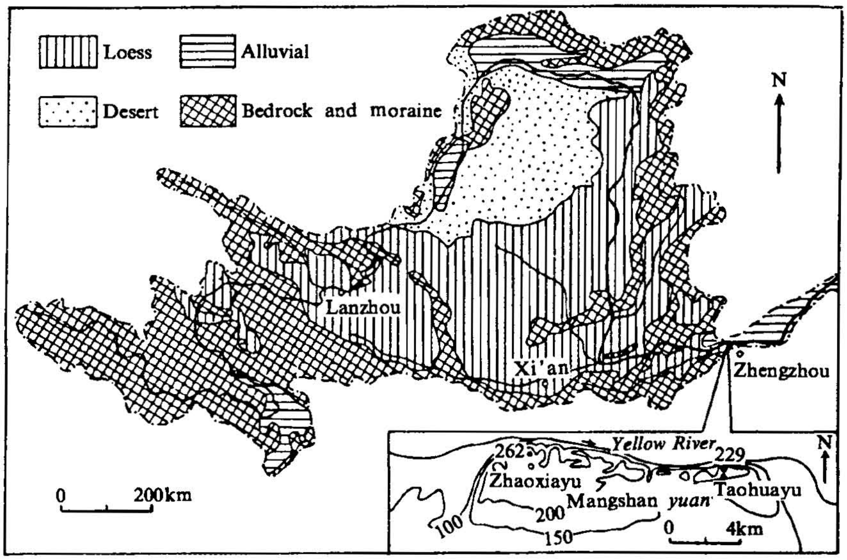

Fig. 1. Distribution of loess in the Yellow River drainage area.

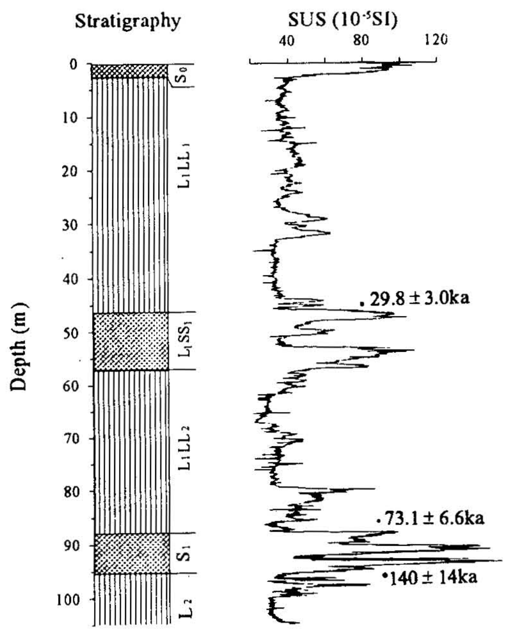

Fig. 2. The loess sequence and magnetic susceptibility through the upper part of the Zhaoxiayu section. OSL ages are given with the susceptibility data.

and grain size distribution analysis, and for optically stimulated luminescence dating (OSL). Fig. 2 shows the results.

The loess stratigraphical units in this section are quite clear, in which the loess L_2 of the penultimate glacial stage is 10.8 m thick, the palaeosol S_1 of the last interglacial stage is 7.4 m thick, and the loess L_1 of the last glacial stage has a total thickness of 85.6 m, including L_1LL_2 of the early stade 30.8 m, L_1SS_1 of the interstade 11 m and L_1LL_1 of the late stade 43.8 m, and the palaeosol S_0 of the postglacial stage is 2.4 m thick.

3 Variations of Summer and Winter Monsoons and Their Interrelations

The climate of China is characterized by seasonal alternations of the summer and winter monsoons of East Asia and its variation history is well documented by the loess–palaeosol sequence (An et al., 1990; 1991b; 1991c). It has been revealed that the magnetic susceptibility which reflects the strength of the southerly warm and wet summer monsoon from the low latitude ocean can serve as a good proxy index of the summer monsoon climate (An et al., 1990; 1991a), showing that the higher the susceptibility, the more rainfall and higher temperature in summer. The coarse size fraction of the loess and palaeosol, which depends on the cold and dry northerly winter monsoon from the high latitudes of the North Hemisphere, can be regarded as an effective proxy index of the winter monsoon climate (Porter and An, 1995; Xiao et al, 1992).

It is shown in Fig. 3 that the variation of summer monsoon climate indicated by the susceptibility curve is opposite to that of winter monsoon climate indicated by the >45 μm size fraction on the 10^4–year scale since the interstade of the last glacial stage, that is, the enhanced summer monsoon corresponded to the declined winter monsoon in the last interstade and the postglacial stage, and the declined summer monsoon corresponded to the enhanced winter monsoon in the late stade. However, the amplitude of summer monsoon changes was evidently smaller than that of the winter monsoon, especially in the late stade of the last glacial stage.

It can also be seen from Fig. 3 that at the close of the early stade of the last glacial stage when the global climate began to turn into the interstadian stage, there was a great difference in the variation of the winter monsoon and the summer monsoon. In other words, the summer monsoon indicated by susceptibility was slowly strengthening, but the winter monsoon indicated by the >45 μm size fraction was declining so rapidly that it equaled or even exceeded the subsequent interstade, and then a slow strengthening followed. One of the possible causes for the inconsistency that the winter monsoon changes took place earlier than the summer monsoon changes is related to the fact that the pedogenesis affecting the strength of susceptibility signal lags behind the variation of the summer monsoon, but the >45 μm size fraction coincides with the variation of the winter monsoon. Another cause worthy of attention is that during this period the weakening of winter monsoon occurred earlier than the strengthening of summer monsoon, because in the case of regression in the marginal sea as a part of the summer monsoon source regions there was a process of transgression into the shelf and land. During the interstade of the last glacial stage the strengthening of summer monsoon corresponded to the weakening of winter monsoon, and vice versa, which shows that they varied almost synchronously. At the turn of the interstade to the late stade the strengthening of summer monsoon occurred simultaneously with the weakening of winter monsoon, but there was a great difference in their amplitudes. At the beginning of the late stade the summer monsoon withered rapidly with a high–amplitude fluctuation, but the winter monsoon gradually intensified with a low amplitude of fluctuation. In the early phase of the late stade the declined summer monsoon was limited to a narrow range of fluctuation as the changes of susceptibility values are less than 10×10^{-5} SI, but the enhanced winter monsoon varied more widely because the fraction of grains of over 45 μm is over 30%. In the middle of the late stade the summer monsoon became stronger, but the fluctuation was not as large as the winter monsoon that was almost close to that in the interstade. In the late phase of the late stade the winter monsoon showed much more evident changes than the summer monsoon with high–amplitude oscillations on a 10^3–10^2–year scale. From the late stade to the postglacial stage there was a period of evidently enhanced winter monsoon, the strength of which equaled or exceeded that of the early phase of the late stade, which is generally considered as the Younger Dryas; but no remarkable changes were seen in the summer monsoon during this period. After the beginning of the postglacial stage the summer monsoon strengthened rapidly and the winter monsoon weakened accordingly. In the Holocene optimum stage the winter and summer monsoons were still unstable and a declining summer monsoon and an enhancing winter monsoon will ensue.

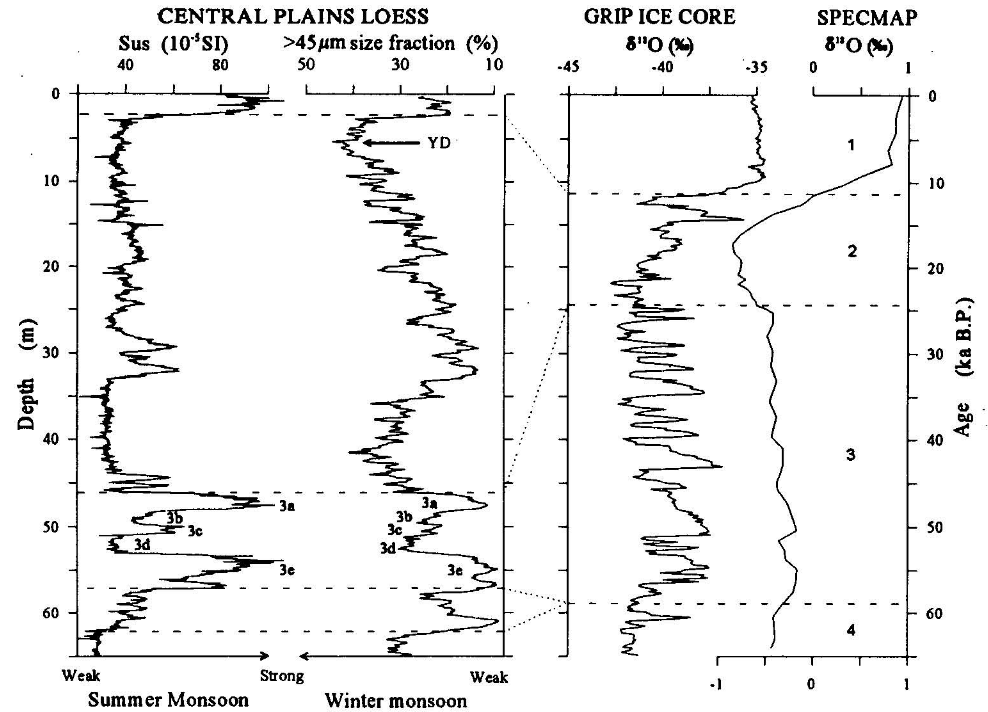

Fig. 3. Magnetic susceptibility and grain size in correlation with the oxygen isotope records from SPECMAP and GRIP ice core.

4 Correlations with Oxygen Isotope Records of the Deep Sea and Ice Core

From the deep sea and ice core records with a timescale in the right part of Fig. 3, it can be seen that the SPECMAP oxygen isotope curve (Martinson et al., 1987) shows no significant climatic change in stage 3, the last interstade, and no large fluctuation in stage 2, the late stade of the last glacial stage. However, high–frequency, high–amplitude fluctuations were present in the oxygen isotope curve of the GRIP ice core (Dansgaard et al., 1993) on the 10^3–year scale in the corresponding stages, which bear some resemblance to the susceptibility curve, especially that of the >45 μm size fraction of the loess record of the Central Plains. Therefore the deep–sea palaeoclimate record can only be correlated with the loess palaeoclimate record on the 10^4–year scale, and the ice core record can match the loess record on the 10^3–year scale to a certain extent.

It has been mentioned above that at the beginning of the last interstade the variation of magnetic susceptibility was inconsistent with that of the >45 μm size fraction. Is it the turning point of the magnetic susceptibility curve or that of the coarse fraction curve corresponding to the boundary between stage 3 and stage 4 of the SPECMAP curve? It appears that the boundary of stages 3–4 (58.9 ka B.P.) corresponds to neither the turn of summer monsoon strengthening reflected by the susceptibility value nor that of winter monsoon weakening reflected by the coarse fraction, but just in the middle of them. For the boundary of stages 2–3 (24.11 ka B.P.), the turning points of the susceptibility values and >45 μm size fraction are concurrent, which seems to illustrate the rapid change in the climate from warm to cold. The boundary of stages 1–2 (12.05 ka B.P.) should be placed at the end of the Younger Dryas event recorded by oxygen isotopes in the ice core, the exact calendar age of 11600 a B.P. (Kerr, 1993).

5 Discussion and Conclusions

The variations of the East Asian palaeomonsoon climate since the last interstade were recorded by the Mangshan loess in the Central Plains, in which the enhanced summer monsoon corresponded to the declined winter monsoon and the declined summer monsoon corresponded to the enhanced winter monsoon on a 10^4–year scale, but on the 10^3–year scale the winter monsoon varied much more intensely than the summer monsoon.

The palaeoclimatic changes recorded by the loess can match the deep–sea oxygen isotope record on the 10^4–year scale, and the variation of the latter is far less intense than that of the former. Moreover, the frequency and amplitude of palaeoclimatic changes on the 10^3–year scale recorded by the Greenland ice core bear a resemblance to that of the winter monsoon recorded by the loess. It demonstrates that the strength of the East Asian winter monsoon is affected to a great extent by the temperature changes in the high–latitude areas including the ice sheet of the North Hemisphere.

There is no evident variation of the $\delta^{18}O$ curves of the deep sea and ice core in the postglacial stage, the Holocene, but there are oscillations on the 10^3–10^2–year scales of both the summer and winter monsoons recorded by the loess in the Central Plains (further study is underway). It is pointed out that the short–term climatic changes bear regional characters and are not applicable for correlations on a global scale.

The palaeoclimate in the East Asian monsoon areas shows different patterns of variation for summer and winter; even on the 10^4–year scale there are great differences in the general pattern that when one is strong the other is weak. Because the palaeoclimatic changes recorded by the ice core and marine deposits are reflections of changes of the average temperature, they cannot be considered identical with those recorded by the loess. Therefore correlations cannot be made in a narrow sense. It is necessary to set up an independent timescale for Chinese loess deposits, which is not to be achieved by using the boundary age of oxygen isotope stages of deep–sea deposits for calibrating the loess–palaeosol sequences. By comparing the independent timescale of Chinese loess records with the ice core and deep sea records it will be possible for us to find more problems. The discussion and understanding of these problems may help us to reveal the distinctive palaeoclimatic changes in the East Asian monsoon areas, and provide new data and clues for the study of the past, present and future global changes.

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