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Variations of eastern Asian monsoon over the last 140,000 years

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Key words. – Loess-soil sequence, Central China, Last climatic cycle, Eastern Asian monsoon.

Abstract. – The Weinan loess-soil sequence in central China provides lithogenic, geochemical and palaeopedological evidence of past changes in the eastern Asian monsoon climate. The chemical weathering index as defined by Nesbitt and Young [1982] and the SiO_2/TiO_2 ratio are found to be sensitive indicators of the strengths of the summer and winter monsoons, respectively. Over the last climatic cycle, strong summer monsoon occurred during six time intervals with an apparent ~ 20,000 years cycle, the major component of the Earth祖 orbital precession. Strengthening of the summer monsoon and weakening of the winter monsoon is approximately in phase. Micromorphological study of the palaeosols reveals a landscape variability from north temperate steppe (Haplic Chernozems) to subtropical forest (Chromic Luvisol) environments for the interstadial and interglacial periods.

Changes in both orbitally produced north summer insolation and glacial age boundary condition are necessary to explain the major shifts of the monsoon climate. Within the constraints of the used time scale, the variations of the monsoon climate are basically coeval with the variations of the global ice volume as indicated by the marine $\delta^{18}O$ record. Moreover, the amplitudes of monsoon variability and the landscape evolution display however striking discrepancies with the global ice volume, which may be better explained by the variations of the summer insolation in the northern hemisphere with an apparent time lag of several thousands years for the loess sequence.

Variabilité de la mousson est-asiatique au cours du dernier cycle climatique

Mots-clés. – Sequence sol-loess, Chine centrale, Dernier cycle climatique, Mousson est-asiatique.

Résumé. – La sequence sols-loess de Weinan en Chine centrale apporte des informations de type lithogenique, geochimique et paléopédologique se rapportant aux changements passes du climat de mousson est-asiatique. L拔ndice d扒lteration chimique défini par Nesbitt et Young [1982] et le rapport SiO_2/TiO_2 apparaissent comme des indicateurs sensibles de l拔ntensite de la mousson d'été et d抒iver, respectivement. Au cours du dernier cycle climatique, la mousson d'été s拔ntensifie au cours de 6 intervals avec un cycle apparent de ~ 20000 ans, correspondant à la precession. Renforcement de la mousson d'été et affaiblissement de la mousson d抒iver sont approximativement en phase. Les etudes micropaleontologiques menées sur les paléosols révèlent la diversite des paysages allant de la steppe nord-tempérée jusqu'à la forêt subtropicale au cours des périodes interstadiaires et interglaciaires.

Les changements dans l拔nsolation d'été de l'hémisphère nord et dans les conditions limites des périodes glaciaires sont nécessaires pour expliquer les modifications majeures du climat de mousson. Tenant compte des contraintes liées à l'échelle chronologique utilisée, les variations de la mousson sont fondamentalement contemporaines de celles du volume global des glaces enregistrees par les isotopes de l'oxygène ($\delta^{18}0$) dans les oceans. Les differences importantes observées cependant entre l扒mplitude de variabilité de la mousson et l'évolution du paysage d抲ne part, le volume global des glaces d扒utre part, peuvent être mieux expliquées par les variations de l'insolation d'été dans l'hémisphère nord avec un décalage apparent de plusieurs milliers d'années pour la sequence des loess.

Version française abrégée

La sequence loess-sols de Chine est l'un des enregistrements continentaux les plus complets et continus du climat des 2,5 derniers millions d扒nnees [Liu, 1985 ; Kukla, 1987]. La partie la plus importante du plateau des loess est située à l拔nterieur de la zone de la mousson asiatique. La section de Weinan, en marge meridionale du plateau des loess est sensible à des processus de sedimentation primaire et de pédogenèse secondaire. Elle fournit les enregistrements lithogeniques, geochimiques et paleopedologiques des changements passes du climat de mousson est-asiatique. La sequence qui recoupe les derniers 140 000 ans consiste en six paleosols avec loess intercales. Le loess Malan L1, recouvert par le sol holocène SO, contient deux paleosols nettement séparés nommés Ll-2 et Ll-4 (depuis le sommet jusqu扒 la base). Le paleosol sous-jacent S 1 datant du dernier interglaciaire, polygenique, consiste en trois unites pedologiques superposees (Sl-1, S l-2 et S l-3) de differente morphologie.

La méthode de datation radiocarbone AMS a été pour la premiere fois appliquee à une sequence loess-sols, en Chine. Les resultats calibres puis combines avec les datations par thermoluminescence montrent que l'âge des limites entre sols et loess correspond approximativement à celui des stades isotopiques. Ceci suggère que les correlations entre stratigraphie des loess et stratigraphie isotopique des enregistrements oceaniques peuvent être considérées comme une méthode d抗laboration finale de l抗chelle chronologique de la sequence loess-sols.

L拔ndice d扒lteration chimique defini par Nesbitt et Young [1982] s抗st avéré un indicateur sensible de l拔ntensite de la mousson d'été. Son application est basée sur des observations anterieures selon lesquelles la temperature et les precipitations, les deux facteurs dominants influençant l扒lteration chimique, sont principalement liées à la mousson d'été [An *et al.,* 199 1]. Au contraire, le rapport moleculaire SiO_2/TiO_2 peut être utilisé pour définir les variations de la mousson d抒iver. Ces deux elements sont en effet stables, donc relativement indépendants de l'altération ; et le renforcement de la mousson d抒iver entraîne l'augmentation de la poussiere eolienne grossière avec pour consequence l'augmentation de la quantité de quartz (c抗st-a-dire des valeurs élevées en SiO_2). Ces nouveaux index peuvent se reveler meilleurs que ceux utilises anterieurement car ils permettent de distinguer de façon plus precise les effets respectifs des moussons d'été et d抒iver. Leur signification climatique peut egalement être clairement expliquee.

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Au cours du dernier cycle climatique, la mousson d'été s'intensifie pendant 6 intervalles de temps (129 000- 118 000 B.P. ; 100000-90000 B.P. ; 85 000-74000 B.P. ; 63 000-49 000 B.P. ; 37 000-24000 B.P. et 12000 B.P.) avec un cycle apparent de 20000 ans, composant majeur de la precession de l'orbite terrestre. Intensification de la mousson d'été et affaiblissement de la mousson d'hiver sont approximativement en phase. Tenant compte des facteurs micromorphologiques et des propriétés chimiques des sols et mineralogiques des argiles, les sols SO et S l-2 peuvent être classes en *Luvic Phaeozems,* Ll-2, Ll-4 et S l-l comme *Haplic Chernozems* et S l-3 comme *Chromic Luvisol* dans le système FAO [FAO-UNESCO, 1974]. Ceci suggere une grande variété d'environnements allant de la steppe nord tempérée a la forêt subtropicale pendant les periodes interstadiaires et interglaciaires. La position de ces sols dans les systèmes de classification des sols chinois [STCRG et CRGCSTC, 1991], français [Duchaufour, 1983] et nord-americain [soil survey staff, 1975] est egalement proposée dans l'article. La succession des paléo-pédogenèses, cohérente avec les variations de l'index de mousson d'été, peut être utilisee comme indicateur de correlations stratigraphiques regionales ou même globales.

Les changements dans l'insolation d'été de l'hémisphère nord et dans les conditions limites des periodes glaciaires sont necessaires pour expliquer les modifications majeures du climat de mousson. Tenant compte des contraintes liées à l'échelle chronologique utilisee, les variations de la mousson sont fondamentalement contemporaines de celles du volume global des glaces enregistrees par les isotopes de l'oxygene ($\delta^{18}O$) dans les océans. Des differences, apparaissent cependant: (1) le stade isotopique 1 et le sous-stade 5e enregistrent des valeurs isotopiques cornparables, suggerant un volume de glace equivalent. Le renforcement de la mousson d'été et l'affaiblissement de la mousson d'hiver durant l'Holocène sont significativement moins importants que durant le dernier interglaciaire ; (2) la périodicité de ~ 20000 ans est significativement plus marquee dans la sequence loess-sols de Chine que dans les enregistrements isotopiques dans les oceans ; et (3) les sous-stades isotopiques 5a et 5c enregistrent des valeurs isotopiques similaires alors que la mousson d'été est beaucoup plus active durant le stade 5c que durant le stade 5a. Ces differences peuvent être mieux expliquees par les variations de l'insolation d'été dans l'hémisphère nord avec un décalage apparent de plusieurs milliers d'annees pour la sequence des loess. Ceci suggere que l'accumulation loessique est liée au volume global des glaces alors que la formation des sols est plus certainement en relation avec l'insolation au travers des variations de la mousson d'été.

INTRODUCTION

The eastern Asian monsoonal system is one of the most dynamic components of the global atmospheric circulation. The modern monsoon climate is driven by two seasonally alternative mechanisms. In summer, the differential heating between the continents and the surrounding oceans results in a low surface pressure over northern Africa and southern Asia and two high pressure cells respectively over the southern Indian Ocean and the western subtropical Pacific. Interaction of these surface cells gives rise to the warm-humid southwestern and southeastern winds, i.e. the southwestern and the southeastern monsoons. The climate in summer in the eastern and central parts of China (including the Loess Plateau) is mainly controlled by the southeastern monsoon [Zhang and Lin, 1987; Liu and Ding, 1993]. In winter, the strong contrast of surface temperature between the cold Eurasian continent and the relative warm adjacent oceans leads to the development of the Siberia-Mongolian high pressure cell and the Aleutian low pressure cell over the North Pacific. The interaction of the two surface cells gives rise to cold-dry northerly winds, i.e. the winter monsoon [Zhang and Lin, 1987; Liu and Ding, 1993].

According to the numerical experiments of the general circulation [Manabe and Terpstra, 1974; Manabe and Broccoli, 1990; Kutzbach et *al.*, 1989], the initiation of a monsoon-like climatic pattern in eastern Asia is associated with the orography uplift which could be traced back to the late Pliocene [CSEQXP, 198 1]. The result is in agreement with the geological records in the Chinese Loess Plateau within the monsoon zone. The thickness and the average grain size of loess decrease from the northwest to the southeast through the plateau [Liu, 1966; 1985], indicating that the sources of the aeolian dust are primarily related to the winter monsoon activities. This is also confirmed by the recent meteorological studies on modern dust storms [Liu et *al., 198 1;* Liu, 1985]. The palaeosols in the loess sequence have been interpreted as a result of the climates with strongly contrasted seasons and high summer evapotranspiration, i.e. a monsoon-like climate. The degree of development of the soils decreases from the southeast to the northwest, in consistence with the direction of the summer monsoon weakening [Liu, 1985; Bronger and Heinkele, 1989; Guo, 1990; Guo et *al.,* 199 1; Guo *et al.,* 1993]. The loess-soil sequence in China can therefore be regarded as a good record of the variations of the eastern Asian palaeomonsoon.

In recent years, a number of studies have been carried out to assess the relationship between the loess-soil sequence and the monsoon climate [An *et al.*, 1991; 199 la; Liu and Ding, 1993; Guo *et al.*, 1993]. In spite of wealth of the acquired results, the following aspects remain to be further addressed.

(1) A number of physical or chemical indexes (e.g. carbonate content, magnetic susceptibility, grain size, etc.) has been used to assess the long-term variations of palaeomonsoon. However, new indexes with clearer climatic significance are needed.

(2) The impact of monsoon on landscape is crucial to the understanding of monsoon variability and the possible forcing mechanisms on a global scale. However, studies on the landscape evolution (especially the palaeovegetation) of the Loess Plateau is puzzled by the uncertain origin and small amount of pollen in loess.

We present here some results from a loess section at Weinan, Shaanxi. The paper aims to (a) identify the climatic events during the last 140,000 years; (b) access the variability of eastern Asian monsoon climate based on two monsoon indexes; (c) interpret the impact of the palaeomonsoon on the landscape evolution of the Loess Plateau through a, palaeopedological approach; and (d) discuss the dynamic relationship of the monsoon climate with the variations of the global ice volume and the summer solar insolation in the northern hemisphere.

I. – MATERIALS AND METHODS

A) Weinan loess section

The newly found Weinan loess section, with a total thickness of 174 m, is located at the southern-most part (38°20'N, 109°30'E) of the Loess Plateau, about 55 km east to the city of Xian (fig. 1). The modern climate in this region is semi-arid, with the annual mean temperature of 13.3 °C and mean annual precipitation of 580 mm in Xian [Sun, 1989]. The section contains more than 40 palaeosols and the stratigraphy can be roughly correlated to that of the Baoji section reported by Ding et al. [1990] and Rutter et *al.* [1991] although the polygenesis of the palaeosols were not studied by these authors. We study here the uppermost part, 12.84 m in thickness, spanning about the last 140,000 years. The major advantage of the section is that the palaeosols are well preserved and clearly expressed (fig. 2). The Malan Loess L1, overlain by the Holocene soil SO, contains two clearly separated palaeosols named as Ll-2 and Ll-4. The interglacial soil S1 is polygenetic consisting of three superposed soil units (S 1-1, S 1-2 and S 1-3) with different morphology, favorable to define the climatic stages within the Last Interglacial (deep-sea oxygen isotope stage 5).

B) Monsoon indexes

To establish the summer and winter monsoon proxy indices, one hundred and twenty-six samples were taken at 10 cm interval from the section. Total chemical composition was analyzed using X-ray fluorescence method on a Philips PW-1400 unit by heating the sample at 1,000°C for two hours. We used the chemical weathering index defined by Nesbitt and Young [1982] as the summer monsoon index (SMI). This application was based on the previous observation that the temperature and precipitation on the Loess Plateau, the dominant factors for chemical weathering, are mainly related to the summer monsoon intensity [An et *al.*, 1991]. The formula of the index is as follows.

$$SMI=[Al_2O_3/(Al_2O_3+CaO+Na_2O+K_2O)] \times 100 \text{ (molecular proportion)}$$

Decalcification is the first soil-forming process which may occur even at the beginning stage of soil formation [Birkeland, 1984]. As the above formula takes account of the CaO content in the samples, decalcification of the soils profiles and accumulation of secondary carbonate in the underlying loess may lead to overstated contrasts of the weathering index between the soils and the interbedded loess beds. In order to eliminate this effect, $CaCO_3$ contents were measured gasometrically and were used to rectify the primary X-ray fluorescence data. The corrected results represent the chemical composition of non-calcareous material and were used to calculate the final SMI.

We used the molecular ratio of SiO_2/TiO_2 as the winter monsoon index (WMI), because: (1) these two elements are the most stable components, thus relatively independent of weathering; and (2) stronger winter monsoon intensity gives rise to coarser aeolian dust and consequently, higher quartz content (high SiO_2 content) [Liu, 1985]. In comparison to the indexes used in earlier studies, our summer and winter indexes have clearer climatic significance.

C) Depth-age conversion

According to the astronomical theory of palaeoclimate, the long-term variations in the geometry of the Earth's orbit and rotation are the fundamental causes of the climate changes over the past 2 or 3 Ma [Berger, 1989]. The calculated variations of the orbital parameters or the orbitally forced solar insolation [Berger et *al.*, 1991] can be used to date the geological record of palaeoclimate [Imbrie et *al.*, 1984; Raymo *et al.*, 1989; Ruddiman *et al.*, 1989].

In this study, the depth-age transformation is based on (1) the AMS-^{14}C dating of humins made by the ^{14}C laboratory of Peking University; the radiocarbon ages were transformed into calendar ages following the calculation method A of Stuiver and Reimer [1993] before being compared with the chronology of SPECMAP oxygen isotope record [Imbrie *et al.*, 1984] and that of the insolation variations (tab. I); (2) the thermoluminecence (TL) dating. made by the TL laboratory of Peking University (tab. I); and (3) the correlation of the SMI to the variations of the summer insolation in the northern hemisphere [Prell and Kutzbach, 1987]. We assume an average time lag of 6,000 years for loess because (1) this time lag allows us to establish a time scale well fitted with the absolute dating results (fig. 2); and (2) geological and absolute dating in a great number of studies indicate that the coldest and driest conditions prevailed the

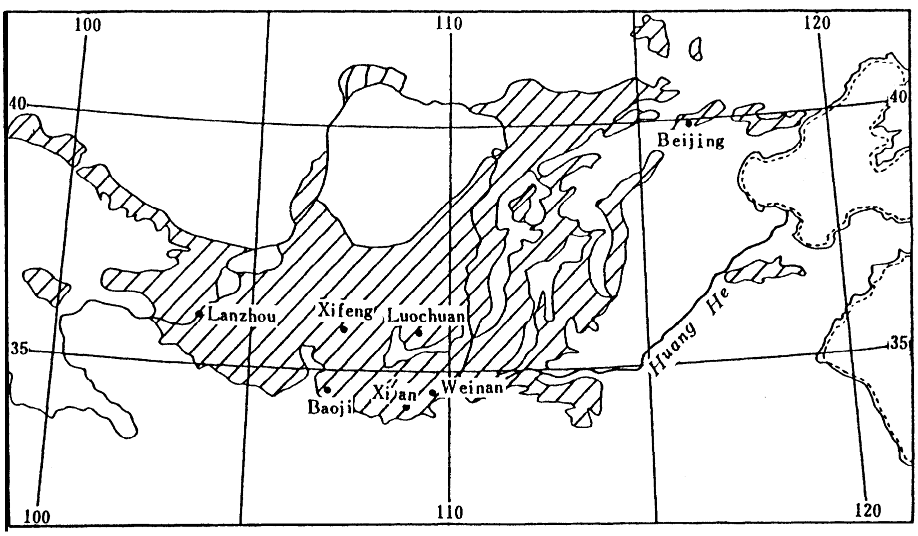

FIG. 1 Location of the Weinan loess section [after Kukla and An., 1989].
FIG. 1. – *Localisation de la section de Weinan [d'après Kukla et An., 1989].*

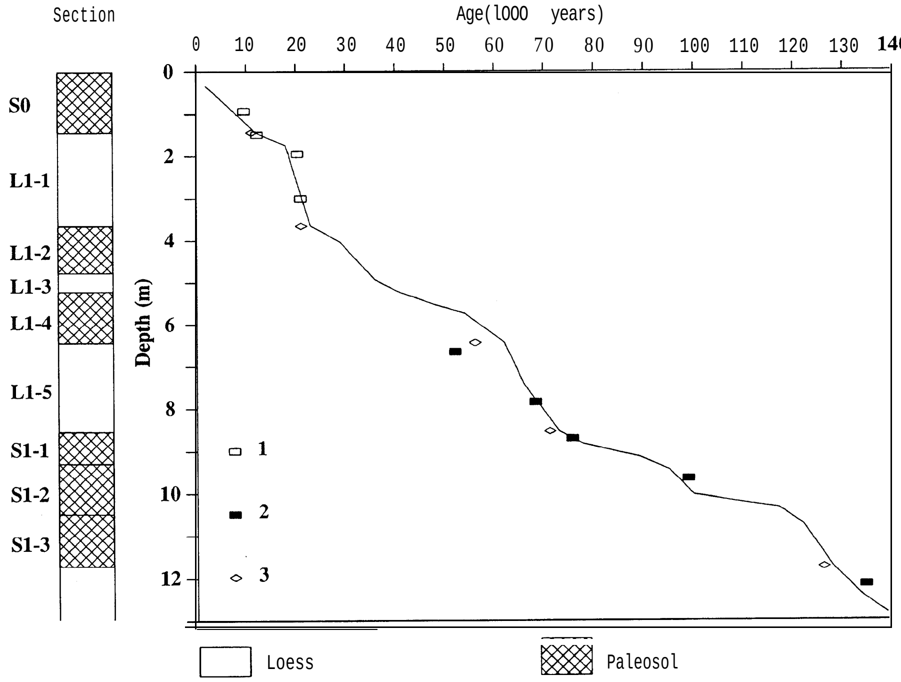

FIG. 2. – Stratigraphy and depth-age transformation curve of the Weinan loess section. The depth-age conversion is made by correlating the summer monsoon index (SMI) to the average northern hemisphere summer solar insolation from Prell and Kutzbach [1987] assuming a time lag of 6,000 years for the loess sequence. 1: AMS-^{14}C dating; 2: thermoluminecence dating; 3: stratigraphic boundary ages provided by the magnetic susceptibility age model of Kukla *et al.* [1988].

FIG. 2. – *Stratigraphie et courbe transformée profondeur-ages de la section des loess de Weinan. La conversion profondeur-ages est faite par correlation de l拔ndex de mousson d'été* (SMI) *avec 1 慖nsolation moyenne d'été dans l'hémisphère nord de Prell et Kutzbach [1987] tenant compte d抧n décalage de 6000 ans pour la sequence des loess.* 1: *datations AMS^{14}C ; 2: datations par thermoluminescence; 3: âges des limites stratigraphiques provenant du modèle de Kukla* et al. *[1988] base sur la susceptibilité magnétique.*

TABLE I. – Absolute dating results of the Weinan section.
TABL. I. – *Datations absolues de la section de Weinan.*

Depth (m)	Radiocarbon and TL age (year B.P.)	Dating method	Calibrated age according to Stuiver and Reimer (1993) (cal. year B.P.)
0.93	8640± 190	AMS- 14c	9532
1.49	10270±380	AMS-14C	12101
1.95	17130±260	AMS-14c	20296
3.00	17570±180	AMS- 14c	2092 1
6.65	51910±210	TL	
7.85	68100±3780	TL	
8.70	75620± 10540	TL	
9.65	98760± 12350	TL	
12.16	134480±17680	TL	

Loess Plateau at 18,000 years B.P. [Liu, 1985], which has a time lag of about 6,000 years to the minimum insolation (24,000 years B.P.). The obtained depth-age transformation is also in good agreement with the stratigraphic boundary ages provided by the magnetic susceptibility age model [Kukla et al., 1988; Kukla and An, 1989] (fig. 2) although the last one is based on some assumptions being still contentious points [Zhou et al., 1990; Han et *al.*, 199 1].

D) Pedological analyses

To assess the impact of the monsoon on the landscape of the Loess Plateau, the palaeosols were studied in using macro- and micromorphological methods combined with selected soil chemical analyses. The micromorphological descriptions were made according to Bullock et al. [1985]. pH in water (1 :5 earth/water ratio), cation exchangeable capacity (CEC, NH_4OAc method) and exchangeable cations were analyzed by the methods outlined in McKeague [1978]. Free iron content was determined by the extraction procedures (citrate-bicarbonate-dithionite or CBD) established by Mehra and Jackson [1960]. Total chemical analyses of the clay fraction were made by X-ray fluorescence method to determine the SiO_2/Al_2O_3 molecular ratio.

II. – MONSOON VARIABILITY OVER THE LAST CLIMATIC CYCLE

The SMI and WMI depicted in figures 3a and 3b mainly reflect variations of the summer and winter monsoon strength, respectively. Over the past 140,000 years, strong summer monsoon occurred during six time intervals (129,000 – 118,000 yrs B.P.; 100,000 – 90,000 yrs B.P.; 85,000 – 74,000 yrs B.P.; 63,000 – 49,000 yrs B.P.; 37,000 – 24,000 yrs B.P. and 12,000 – 0 yrs B.P.) with a dominant frequency of about 20,000 years, a major component of the variations of Earth抯 orbital precession [Berger, 1989]. Although the Weinan depth-age transformation is based on the correlation with the variation of solar insolation in the

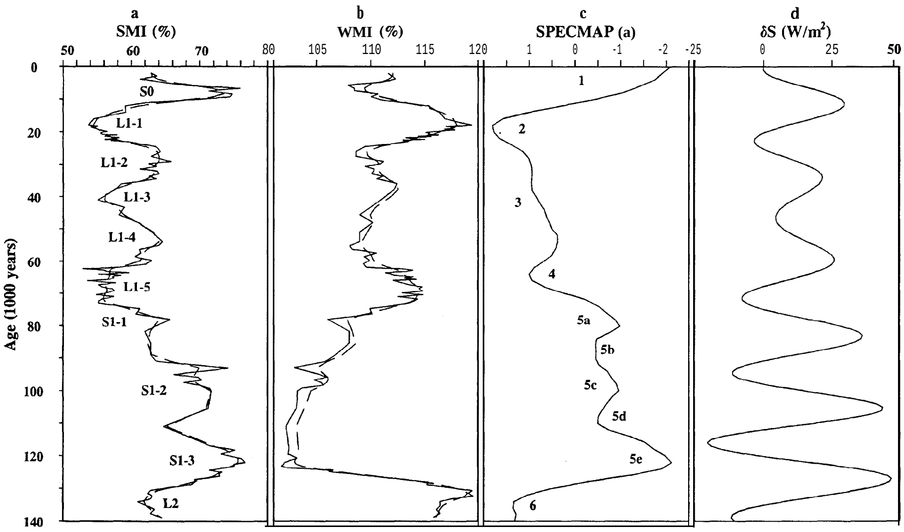

FIG. 3. – Comparison of monsoon variability with variations of the global ice volume and summer insolation in the northern hemisphere. (a) Summer monsoon index (SMI), the dashed line is the smoothed SMI curve; (b) winter monsoon index (WMI), the dashed line is the smoothed WMI curve; (c) SPECMAP composite oxygen isotope record from Imbrie *et al.* [1984] as the indication of the global ice volume variations; (d) the average northern hemisphere summer solar insolation from Prell and Kutzbach [1987].

FIG. 3. – *Comparaison de la variabilité de la mousson avec les variations du volume global des glaces et de l拔nsolation d'été de l 感emisphere nord.* (a) *index de mousson d'été* (SMI), *la ligne discontinue représente la courbe* SMI *lissée ;* (b) *index de mousson d抙iver* (WMI), *la ligne discontinue représente la courbe* WMI *lissée;* (c) *enregistrement isotopique composite SPECMAP de Imbrie et al. [1984] utilisée comme indication des variations du volume global des glaces;* (d) *insolation moyenne d'été de l 感emisphere nord [Prell et Kutzbach, 1987].*

northern hemisphere, strongly driven by precession, the consistency of the established time scale to the orbitally independent magnetic susceptibility age model [Kukla et al., 1988; Kukla and An, 1989] and the absolute dating (fig. 2) strongly support the ~ 20,000 years frequency of the monsoon climate.

Strongest summer monsoon occurred near 123,000 years B.P. during the formation of S l-3 soil. In figure 3, comparable monsoon strength is displayed for Ll-2, Ll-4 and S 1-1 soils. The Holocene soil SO peak is slightly lower than that of S l-3 and higher than that of S l-2. The last aspect is somewhat inconsistent with the genesis of the paleosols because the SO and S l-2 soil is genetically comparable (discussed below). We observe a significant increase in Al_2O_3 and a decrease in SiO_2 in the Holocene profile (which can lead to higher SMI), which may be attributable to human-induced modification of the soil. Apart from the removing of plant-building elements by agriculture, the habitants on the Loess Plateau are accustomed to transport a mixture of all types of waste to the field as soil fertilizer and considerable modification of soil composition can be expected.

Study on the phase relationship between the summer and winter monsoons needs more work. Nevertheless, a visual comparison of the two monsoon indexes reveals several important relationships: (1) stronger summer monsoon is ap-徳roximately in phase with weaker winter monsoon with an exception that the intensification of the summer monsoon leads slightly the drop of the winter monsoon at the beginning of the Last Interglacial; (2) during the deposition of Ll-1 and Ll-5, the SMI shows equal values while much higher WMI values are observed near 18,000 years B.P.; and (3) the WMI of the previous glacial (upper part of L2) dropped to about the same level as 18,000 years B.P. whereas the SMI is higher during the previous glacial. The anomaly of the WMI in the Holocene profile showing similar values to Ll-2 and Ll-4 seems to support the above interpretation of the results of Holocene soil.

III. – PALAEOSOLS AND IMPACT OF MONSOON ON LANDSCAPE

The major chemical properties of the palaeosols and loess are shown in figure 4 and figure 5. The pH values vary between 7.4 and 8.5, suggesting slightly alkaline environments. For all the studied soils, the exchangeable cations are dominated by Ca^{2+} and Mg^{2+} and the adsorb complex is saturated. The SiO_2/Al_2O_3 ratio of the clay fraction range from 3.65 to 4.23 and the free iron content varies from 1.02 to 2.22, indicating limited weathering degree.

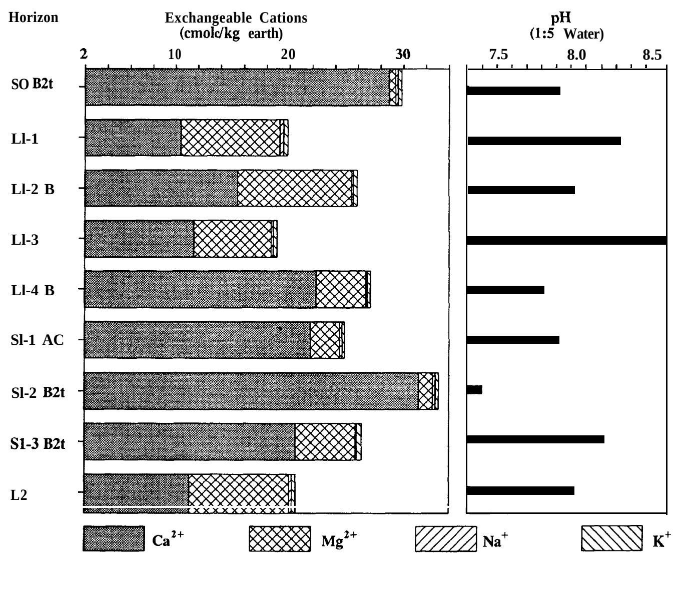

FIG. 4. – Exchangeable cations and pH for selected horizons.

FIG. 4. – *Cations échangeables et pH pour certains horizons.*

The decrease of SiO_2/Al_2O_3 ratio and the increase of free iron content in soils suggest weak pedological enrichment of Fe-Al oxides and hydroxides.

Earlier study [Guo et al., 1993] on the Holocene soil SO and S1 palaeosol in the nearby Xian loess section (55 km west to Weinan) revealed that the clay mineralogy is dominated by 2: 1 type. The soils contain illites, kaolinite, chlorites, smectites and interstratified minerals. The major difference from the underlying loess layers is a slight increase of smectites and the decrease of the crystallinity of illite in soils, suggesting that the weathering is in the stage of *bisiallitisation* [for term see Pedro, 1979].

The micromorphological features allow us to distinguish the studied soils into three groups. The positions of these soils in different classification systems [Fao-Unesco, 1974; Soil survey staff, 1974; Duchaufour, 1983; STCRG and CRGCSTC, 1991] are proposed in table II.

(1) SO and Sl-2 soils. The upper part (Ap, 33 cm) of the Holocene soil SO was strongly reworked by human activity as indicated by the abundant charcoals and brick pieces. An argillic horizon (Bt) is well preserved at the lower part, morphologically similar to that of the S l-2 soil (dark brown, 7.5-5 YR 4/6). They have dense incomplete excremental infilling microstructure with abundant biopores (12-15 %), typical of steppe dominant soil with mollic epipedon [Pawluck and Bal, 1985]. The dense excremental infillings imply that the soil profile can be temporarily water-saturated [Courty and Fedoroff, 1985]. This is in agreement with the Fe-Mn hypocoatings (1-2%) around the biopores. The fine fraction (< 5 μm) is dark reddish brown and humus-riched, containing some reddish brown iron particles, indicating brunification and weak rubification of the fine fraction. The clay coating (about 5%) are dark brown (humus-riched), slightly dusty and non-laminated with moderate birefringence. According to Fedoroff and Goldberg [1982], this type of humus-riched clay coatings are usually associated to vegetations containing some species providing fulvic acid in relative abundance (e.g. resinous). The above features, combined with the chemical properties suggest a ustic soil water regime and a mesic temperature regime [Soil survey staff, 1975]. The vegetation was dominated by dense steppe containing probably some resinous species (fig. 6).

(2) Ll-2, Ll-4 and Sl-1 soils. The AC horizon of these soils are dark brown (10-7.5 YR 4/6) with fine granular structure, underlain by a Ck horizon with abundant calcitic

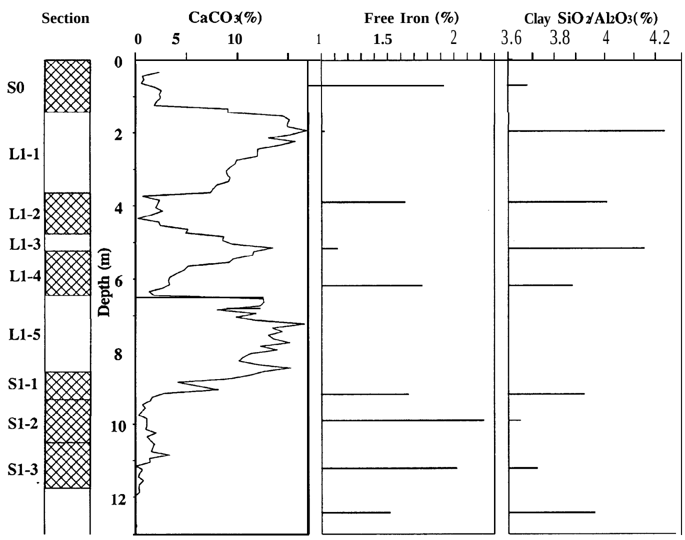

FIG. 5. – Carbonate content profile, free iron content and clay SiO_2/Al_2O_3 ratio for selected horizons. Legend same as in fig. 2.

FIG. 5. – *Profil du contenu en carbonate, fer libre et rapport des argiles SiO_2/Al_2O_3 pour certains horizons. La légende est identique à celle de la figure 2.*

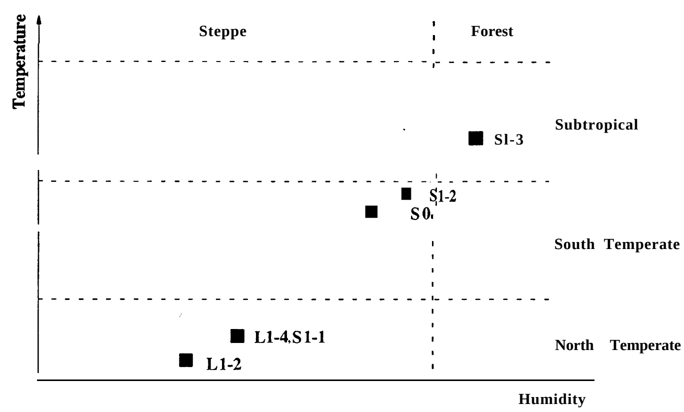

FIG. 6. – Scheme illustration of climatic and landscape interpretations of the palaeosols.

FIG. 6. – *Illustration schématique des interprétations tirées des paléosols concernant le climat et les paysages.*

TABLE II. – Comparative classification of Weinan palaeosols using different soil classification systems.

TABL. II. – *Classification comparative des paléosols de Weinan utilisant différents systèmes de classification des sols.*

Soil	F.A.O. (Fao-Unesco, 1974)	U.S.A. (Soil survey staff, 1975)	France (Duchaufour, 1983)	China (STCRG and CRGCSTC, 1991)
S0	Luvic Phaeozem	Argiustoll	Brunizem Lessive	Helu Soil - Luvic Thermo-black Soil transition
L1-2	Haplic Chernozem	Vermustoll	Chernozem Humifere	Chernozem
L1-4	Haplic Chernozem	Vermustoll	Chernozem Humifere	Chernozem
S1-1	Haplic Chernozem	Vermustoll	Chernozem Humifere	Chernozem
S1-2	Luvic Phaeozem	Argiustoll	Brunizem Lessive	Helu Soil - Luvic Thermo-black Soil transition
S1-3	Chromic Luvisol	Rhodustalf	Sol Brun Fersialtitique Lessive	Argillic Eutrophic Brown-red Soil

pseudomycelia. The fine fraction of these soils is humus-riched, containing a great amount of dark charcoal particles. These soils have typical spongy microstructure, characteristic of steppe soils with mollic epipedon [Pawluk and Bal, 1985]. The well preserved spongy microstructure, consistent with the absence of Fe-Mn and textural features, indicates that the water regime has been ustic and the profiles have rarely been water-saturated [Courty and Fedoroff, 1985]. The lack of rubification and the significant humus accumulation suggest a temperature regime ranging from mesic to frigid type [Soil survey staff, 1975]. These soils can be classified as Vermustolls in U.S.A. system and Haplic Chernozems in FAO system [Fao-Unesco, 1974] with steppe vegetation (tab. II, fig. **6).**

(3) Sl-3 soil. It differs from others in three aspects : (1) weak humus accumulation and strong rubification (reddish brown, 5 YR 4/6-8), indicating that the soil temperature regime is significantly hotter. A thermic regime can be suggested as high summer temperature and strong seasonal contrast are the necessary conditions for rubification [Bresson, 1976; Schwertmann et al., 1982; Duchaufour, 1983]; (2) the moderate Fe-Mn features (3-4%), clay coatings(5-6%) and the presence of calcareous horizon imply that the water regime is still ustic, but significantly wetter than in the younger soils. The same inference can be made from the fissure microstructure observed; and (3) a well developed Bt horizon with prismatic structure. The clay coatings are reddish brown, limpid and microlaminated with high birefringence. Such texture features are commonly interpreted as typical of broadleaf forest soils [Fedoroff and Goldberg, 1982; Avery, 1985; Bullock, 1985]. The morphological features in combination with the chemical properties suggest that the Sl-3 soil was formed under subtropical semi-humid conditions (fig. 6). It can be classified as Rhodustalf in USA system and Chromic Luvisol in FAO system (tab. II).

Our results thus suggest a landscape variability from north temperate steppe to subtropical forest for the interglacial and interstadial intervals (fig. 6), which is in good agreement with the fluctuations of the summer monsoon index.

IV – COMPARISON OF MONSOON CLIMATE TO THE VARIATIONS OF GLOBAL ICE-VOLUME AND NORTH SUMMER INSOLATION

The forcing effects of the global ice-volume and the orbitally forced solar insolation were considered in explaining the variability of eastern Asian monsoon climate [An et *al.*, 1991a; 1991b; Liu and Ding, 1993]. Initial results have shown that the variations of the monsoon climate over the last climatic cycle match both the northern summer insolation and the global ice-volume [An et *al.*, 1991b], while long-term variations of the magnetic susceptibility and grain size back to 2.5 Ma B.P. display a stepwise coupling to the global ice volume which cannot be fully explained by the orbital forcing [Liu and Ding, 1993]. In this study, the monsoon indexes and the results on landscape evolution allow us to further evaluate the extent and the manner of the coupling of the monsoon climate with the ice volume for the last 140,000 years.

We use the SPECMAP stacked oxygen isotope record [Imbrie et *al.*, 1984] as indication of the global ice volume variations (fig. 3c). Visual comparison reveals that major shifts in the eastern Asian monsoon climate over the past 140,000 years are coeval with the deep-sea oxygen isotope record, indicating that the monsoon climate is linked, to some extent, to glacial climatic conditions associated with global ice volume. This relationship may be attributed to the influence of polar ice sheets on the strength of winter monsoon or/and that of the glacial forced sea level changes on the summer monsoon [An et *al.*, 1991b; Liu and Ding, 1993]. During glacial periods, the enlarged polar ice sheets may intensify the Siberia-Mongolian high pressure cell through the down-stream cooling effect, leading to stronger winter monsoon [Liu and Ding, 1993]. Moreover, the ice induced sea level changes can lead to greater distance between the coastal line and the inland of China, thus weakening the influence of summer monsoon on the Loess Plateau [An *et al.*, 1991b].

However, striking disagreements exist between the monsoon and oxygen isotope record in the following three aspects.

(1) The oxygen isotope stage 1 and substage 5e show comparable isotope values, suggesting similar ice volumes. The strengthening of the summer monsoon and the weakening of the winter monsoon during the Holocene are however significantly constrained in comparison to the Last Interglacial. The difference is also clearly expressed by the genetic properties of the palaeosols *: SO* is a temperate steppe dominant Luvic Phaeozem [Fao-Unesco, 1974] while Sl-3 shows evidence of subtropical sub-humid forest conditions (Chromic Luvisol).

(2) The oxygen isotope stage 3 is a interstadial period with minor ice-volume fluctuations while two clearly separated intervals with strong summer monsoon are recorded in the Weinan loess-soil sequence. The palaeosols (Ll-2 and Ll-4, Haplic Chernozems) indicate the steppe landscape in Weinan, comparable to that of the S l-l soil. Our results suggest that the ~ 20,000 period, the major component of the orbital precession, is significantly stronger in the loess-soil sequence in China than in the oxygen isotope record, at least over the past 140,000 years.

(3) The oxygen isotope substage 5a and 5c have similar isotopic values while the SMI for S l-2 show much higher values than in S 1- 1. This is in good agreement with the palaeosols : S l-l is a Haplic Chernozem while Sl-2 is a Luvic Phaeozem.

These significant discrepancies between the two records indicate that the global ice volume variations alone cannot explain the variability of the eastern Asian monsoon climate over the last climatic cycle.

It should be mentioned that in Europe and North American, the Holocene soils (Orthic Luvisols) are genetically similar to that of the Last Interglacial, suggesting comparable climatic conditions [Fedoroff and Goldberg, 1982; Fedoroff and Courty, 1988; Guo *et al.*, 1993]. The similarity of climate was also reported in some European pollen records [Guiot *et al.*, 1989]. These results suggest that the glacial forcing may be more important in modulating the palaeoenvironments in Europe and North America than for the Loess Plateau in China.

The variations of the summer insolation in the northern hemisphere [Prell and Kutzbach, 1987] are depicted in figure 3d and compared with the monsoon indexes (figs. 3a and 3b). Strong summer monsoon can be well matched to the high insolation values with a time lag of several thousands years (~ 6,000 years) for the soil-loess record. Although this lag has been assumed in the depth-age transformation, the consistency of the obtained time scale to the absolute dating and to the magnetic susceptibility age model [Kukla *et al.*, 1988; Kukla and An, 1989] strongly support the interpretation. The most interesting relationship is the great similarity in amplitudes and frequency between the SMI and the northern summer insolation curve (fig 3a, 3d). The difference between the monsoon climate and the oxygen isotope record, including the stronger ~ 20,000 years frequency in SMI, can be better explained by the variations of the northern summer insolation. These results suggest that the eastern monsoon climate is dynamically linked to the solar insolation driven by the variations of Earth担 orbital parameters, especially those of the precession. The impact of solar insolation on the eastern monsoon climate, especially the conditions of the interglacial and interstadial periods, is likely to be greater than that of the global ice volume. The relationship may be attributed to the influence of low and midlatitude solar insolation to the strength of the summer monsoon. However, the time lag of the monsoon record suggests a delayed response of the monsoon to the changes in solar insolation.

In ***summary,*** both orbitally produced solar insolation changes and glacial age boundary condition changes are necessary to explain the major shifts of eastern Asian monsoon climate over the last climatic cycle. The northern summer insolation appears to have greater influence on the climatic conditions of interglacial and interstadial periods than the global ice volume. However, the. important discrepancies between the monsoon climate and the global ice-volume variations and the time lag of several thousands years to the insolation changes suggest that other factors (or mechanisms) may have operated in modulating the variability of the eastern Asian monsoon climate. These factors (or mechanisms) should be associated with both the solar insolation and the glacial boundary conditions related to the global ice volume.

CONCLUSIONS

Detailed information of the variations of the eastern Asian monsoon system was preserved in the Weinan loess-soil sequence, which is sensitive to both (primary) depositional and (secondary) pedogenic processes. The time series of weathering index reflects mainly the past changes in precipitation and temperature, and thus the summer monsoon intensity. On the contrary, the changes in SiO_2/TiO_2 molecular ratio, primarily related to quartz content, can be regarded as an indication of the winier monsoon strength as the quartz content is closely associated with the grain-size in the aeolian dust [Liu, 1966; 1985].

Over the past 140,000 years, strong summer monsoon occurred during six time intervals with strong ~ 20,000 years frequency, the major component of the Earth担 orbital precession. The variations of the summer and winter monsoons are basically in phase. The palaeosols revealed that the landscape of different interstadial and interglacial periods in Weinan varies from temperate steppe (Haplic Chernozems and Luvic Phaeozems) to subtropical forest (Chromic Luvisol). This variability of palaeo-pedogenesis can serve as an indicator for regional, even global stratigraphic correlations.

Comparison of the monsoon variability with the global ice volume as indicated by marine $\delta^{18}O$ record, and with the orbitally produced solar insolation changes suggests that both factors are necessary to explain major shifts of the monsoon climate in the past. The evolution of the monsoon climate is coeval with the oxygen isotope record within the constraints of the obtained time scale. However, the amplitude of the monsoon variability and the landscape evolution displayed striking discrepancies to the global ice volume, which may be better explained by the variations of the summer insolation in the northern hemisphere with an apparent time lag of several thousands years for monsoon. These results suggest that other factors (or mechanisms), associated to both the global ice volume and orbitally forced solar insolation, may have operated in modulating the variability of the eastern Asian monsoon climate over the past 140,000 years.

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