^{10}Be AND MAGNETIC SUSCEPTIBILITY IN CHINESE LOESS

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Abstract. ^{10}Be and magnetic susceptibility have been measured in the top 12 m of a loess profile from Luochuan, Central China. Comparison of the ^{10}Be concentration record with the SPECMAP $\delta^{18}\text{O}$ profile of the last 130 ka (Imbrie et al., 1984) results in a new time scale which is in good agreement with that of Kukla et al. [1990]. The calculated loess accumulation rates can be compared with existing records of dust falls in Pacific deep sea sediments and Antarctic ice cores. A simple model for the ^{10}Be dust flux demonstrates that a significant part of the magnetic susceptibility signal in palaeosol horizons is due to in situ pedogenic production. During times of high loess accumulation (cold palaeoclimate) this contribution is negligible.

Introduction

The Loess Plateau in Central China represents one of the best suited archives on land for the study of long-term climatic variations. It has been built up continuously by dust falls over the last 2.4 million years [Heller and Liu, 1982]. During warm and humid climate conditions loess was subject to weathering and transformed into palaeosol.

Shen et al. [1992] recently measured the concentration of cosmogenic ^{10}Be in loess in order to derive a time scale for the last 730 ka by matching the ^{10}Be variations with the astronomically tuned $\delta^{18}\text{O}$ curve [Imbrie et al., 1984].

Magnetic susceptibility in loess measured earlier [Heller and Liu, 1986] revealed profiles similar to the records of $\delta^{18}\text{O}$

in deep sea sediments and of ^{10}Be in loess. Those profiles have also been used to establish a time scale [Kukla et al., 1988]. Several hypotheses have been forwarded to explain the good agreement between susceptibility and $\delta^{18}\text{O}$. These hypotheses being concerned with the formation of the magnetic signature can now be discussed in view of the ^{10}Be

Fig. 1. Lithology, specific low field magnetic susceptibility and ^{10}Be concentration of the upper 1200 cm of the loess profile from Luochuan. Layer S0 consists of a partly reworked Holocene black loam. L1LL1 and L1LL2 both are composed of yellowish-brown coarse-grained loess. L1SS1 refers to a weathered interlayer within the first order loess unit L1. It consists of loess with slightly darker color indicating a period of increased pedogenesis. S1 is formed of prominent brown soil corresponding to oxygen isotope stage 5. L2 finally represents the beginning of the second loess unit. Lithological nomenclature according to Kukla et al. [1990].

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data obtained from samples in which both parameters, the ^{10}Be isotope and the magnetic susceptibility, have been measured.

A total of 116 samples from the upper 12 m of the Luochuan section ($35^\circ 45' \text{ N}$, $105^\circ 25' \text{ E}$) was analysed for ^{10}Be concentration and magnetic low field susceptibility. In the depth interval 0.4 - 12 m, at every 10 cm a slice of about 2 cm was analysed resulting in a mean time resolution of about 1'000 years. First the magnetic susceptibility was determined using the Bartington susceptibility meter of the Institute of Geophysics of the ETH Zürich. Then ^{10}Be was extracted from the same material following a procedure described by Shen et al. [1992]. All ^{10}Be measurements were carried out using the accelerator mass spectrometer of the Paul Scherrer Institute and the ETH Zürich. Each sample was analysed at least twice with final measurement uncertainties of 3-5%.

^{10}Be Time scale

The comparison of ^{10}Be concentration and magnetic susceptibility in Figure 1 shows a high degree of similarity (correlation factor $R = 0.89$). Generally higher values of both parameters are observed in the weathered layers and soils L1S1 and S1. However, there is a significant difference between ^{10}Be and magnetic susceptibility. The ratios between soil and loess layers amount to 2-3 for ^{10}Be and 4-6 for susceptibility. The shape of the susceptibility curve looks smoother than the one of the ^{10}Be concentration. Before discussing the similarity of ^{10}Be and magnetic susceptibility in more detail a time scale needs to be established.

First attempts to date loess younger than 200 ka have been made using the methods of ^{14}C and thermoluminescence (TL). However, many of the published TL dates on Chinese loess are unreliable due to laboratory problems [Pye and Li, 1989]. The ^{14}C dating range is limited by the half-life of 5730 years to about 40 ka. The most successful dating approach so far was based on the assumption of a constant susceptibility flux [Kukla et al., 1988]. To establish a ^{10}Be time scale over the last 130 ka the high resolution ^{10}Be concentration record (Figure 1) was matched to the dated SPECMAP $\delta^{18}\text{O}$ record

Fig. 2. Specific magnetic susceptibility (a) and ^{10}Be concentration as a function of time after matching the ^{10}Be curve (c) to the SPECMAP $\delta^{18}\text{O}$ time scale (b).

Fig. 3. Comparison of the loess accumulation rate (b) with the dust mass flux in the Pacific sediment core V21-146 [Hovan et al., 1991] and the dust volume flux in the Vostok ice core [Petit et al., 1990].

[Imbrie et al., 1984] (Figure 2). The matching procedure is straightforward during periods of fast signal change. Here the precision of the time scale is as good as that of the SPECMAP time scale provided the climate variations are recorded simultaneously by the $\delta^{18}\text{O}$ variations in the deep sea sediments and by the ^{10}Be concentration changes in loess. During periods of minor climate variations, the matching is more difficult leading to possible uncertainties of 5-10 ka relative to the SPECMAP time scale. Our time scale is in good agreement with that of Kukla et al. [1990] except for the period S1.

Dust flux

Using the new ^{10}Be time scale the accumulation rate can be calculated. Figure 3b shows three distinct periods with clearly enhanced accumulation rates. They correspond basically to the ^{18}O stages 2, 4 and 6 correlated with cold glacial periods.

The dust forming the Chinese Loess Plateau comes from the deserts of northwestern China and Mongolia. The material is transported mainly in the troposphere below about 2 km and is driven by northwesterly winds related to the Mongolian anticyclone [Pye and Li, 1989]. Today such conditions occur mainly during winter times. During cold glacial periods when the northern insolation and the precipitation rates were low, the summer monsoon in southern Asia was reduced leading to more frequent dust falls also in spring and summer. Dust residing higher up in the troposphere is transported over much longer distances and can reach the Pacific Ocean where it becomes incorporated into the sediments. The core V21-146 is located in the northwest Pacific downwind from China ($37^\circ 41' \text{ N}$, $163^\circ 02' \text{ E}$, depth: 3968 m) [Hovan et al., 1991]. Its record of aeolian dust flux (Figure 3a) can be expected to be correlated with the Chinese loess profile (Figure 3b). The Vostok ice core in Antarctica (Figure 3c) on the other hand is far away from the Asian continent and reflects more the global dust flux [Petit et al., 1990].

A comparison of the three records reveals the following:

1. There are three periods of enhanced dust flux. Two of them occur around 20 and 60 ka and are well supported by many ^{10}Be data points. The third one at 130 ka is based only on a few points (Figure 3b). It is missing in Figure 3c due to inaccuracies in the Vostok time scale during that time interval (Lorius, personal communication). The highest dust fluxes correspond to minima of the northern hemispheric summer insolation according to the theory of Milankovitch.

2. The relative amplitudes of the dust peaks at 20 ka and 60 ka are similar for the loess and the Vostok record. During the Holocene, however, the dust flux at Vostok is considerably smaller than at the Loess Plateau.

3. There is a trend of increasing mean dust flux towards the present in all three records, except for the Holocene where the dust flux at Vostok is low. The gradual increase in the rate of aeolian accumulation in core V21-146 takes place over the last 500 ka and is therefore not visible in the short period shown in Figure 3a. It has been attributed to an overall trend towards drier climate in China caused by the uplift of the Tibetan Plateau [Kutzbach et al., 1989].

^{10}Be and magnetic susceptibility fluxes

As discussed by Shen et al. [1992], the ^{10}Be flux can be considered to be composed of two different components: The atmospheric flux, F_A , which is caused by the interaction of cosmic ray particles with the atmosphere, and the dust flux, F_D , with ^{10}Be attached to the grains during their exposure at the surface in the source region of the dust. The total ^{10}Be flux, F_B , has been plotted in Figure 4 as a function of the accumulation rate. The linear regression line shows that F_B can be approximated by

$$F_B = F_A + F_D \quad (1)$$

under the assumption that F_A is constant. This simple linear approach neglects several complications:

1. The atmospheric flux F_A is not constant. There are clear indications that the production rate of ^{10}Be may have changed by as much as 40% due to changes of the geomagnetic dipole moment [Bard et al., 1990]. However, the mean flux F_A of $1.3 \cdot 10^6$ atoms $\text{cm}^{-2} \text{a}^{-1}$ over the last 130 ka deduced from Figure 5 is within the range of today's values [Somayajulu et al., 1984].

2. During windy periods the grain size distribution is coarser than during calm periods. Since the ^{10}Be concentration is inversely proportional to the grain size of the particles it becomes attached to [Shen et al., 1992], the ^{10}Be flux during loess periods could be slightly overestimated.

Higher susceptibility in palaeosol layers has been observed

Fig. 4. Relationship between the ^{10}Be flux and the accumulation rate.

Fig. 5. Estimated ratio of the in situ formed pedogenic susceptibility component, F_X , to the total susceptibility flux, F_S .

in many profiles [e.g. Kukla et al., 1990]. Three hypotheses have been brought up to explain this observation: (1) Pedogenic in situ formation of ultrafine magnetite by chemical precipitation [Maher and Taylor, 1988]. This process depends on the moisture of the soil. (2) A quasi-constant sedimentation of natural magnetic dust from the higher atmosphere [Kukla et al., 1988]. This model is based on the good correlation between susceptibility and $\delta^{18}\text{O}$. (3) A selective concentration of magnetite in the soil by calcium carbonate dissolution and compaction [Heller and Liu, 1984].

The question arises whether a combined analysis of susceptibility and ^{10}Be may help to clarify the origin of the magnetic susceptibility. A possible approach is to assume that the total magnetic susceptibility flux, F_S , is also composed of two components, a dust flux, F_D' , and an additional unknown flux, F_X .

$$F_S = F_D' + F_X \quad (2)$$

F_X includes additional sources of magnetic susceptibility flux such as a possible atmospheric component or a pedogenic in situ produced component.

F_X can be quantified under the assumption that the magnetic susceptibility dust flux, F_D' , is proportional to the ^{10}Be dust flux, F_D .

$$\text{With } F_D' = \alpha \cdot F_D \quad (3)$$

equation (1) becomes $F_D' = \alpha \cdot (F_B - F_A)$ and thus equation (2) turns into

$$F_X = F_S - \alpha \cdot (F_B - F_A) \quad (4)$$

The relative contribution of F_X to the total flux F_S in percent is given by

$$F_X/F_S = (1 - \alpha \cdot (F_B - F_A)/F_S) \cdot 100 \quad (5)$$

α corresponds to the linear regression factor between F_S and $(F_B - F_A)$ and amounts to $0.41 \cdot 10^{-11} \text{ cm}^3$. F_X/F_S is shown as a five point running mean in Figure 5.

It varies between 0 and about 80% with low values in loess layers and high values in palaeosols and weathered loess. The highest values are observed in palaeosol S1 which corresponds to oxygen stage 5. Although these results represent only first order approximations, they show that F_X is dominated by pedogenic in-situ formed magnetic

susceptibility during warm and humid periods. The lower mean values of F_x during loess periods as shown in Figure 5 relate to higher but also more variable F_D' contributions indicating that a quasi-constant atmospheric susceptibility flux according to hypothesis (2) is, if existent at all, of no importance. This finding agrees well with the observation in Figure 1 that the ratio between soil and loess layers is different for ^{10}Be and susceptibility. The larger increase of susceptibility during soil intervals can only be explained by an additional source such as pedogenic processes which are not correlated with the ^{10}Be concentration.

Conclusions

Matching the ^{10}Be concentration record with the SPECMAP $\delta^{18}\text{O}$ profile provides a new time scale for the last 130 ka. The loess accumulation rate derived from this time scale shows well defined maxima around 20 and 60 ka B.P. in good agreement with dust records observed in deep sea sediments (V21-146) and in Antarctic ice (Vostok).

A simple correlation model shows that the magnetic susceptibility can be divided in a dust related component and an in-situ produced pedogenic component. The latter is dominant with up to 80% in palaeosol horizons and low till negligible during high accumulation periods explaining the different soil to loess ratios of ^{10}Be and susceptibility. Variations of the atmospheric production rate of ^{10}Be and changes in the grain size distribution must be taken into account in future studies.

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