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Application of Magnetic Susceptibility Measurements in Paleogeographic Investigations

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Abstract—A series of publications, in which the magnetic susceptibility measurements are used for paleoclimatic interpretations and geological correlation of the loess formation, is analyzed. As is shown, the magnetic susceptibility parameters are of a limited value for the indicated purpose, and their interpretation should be specially substantiated in each concrete case.

Key words: magnetic susceptibility, paleoclimate, correlation, loess, buried soil, oxygen-isotope data.

The magnetic susceptibility parameters (κ) of sedimentary rocks are frequently used in paleogeographic studies, but not all the inferences based on these parameters are correct and convincing. In my previous works (Bol'shakov, 1996, 1997, 2000), I considered methodological and methodical problems of interpreting the rock magnetism, in particular, their magnetic susceptibility (MS) parameters. It was shown that MS of deposits depends on many factors, such as climatic conditions of sedimentation, remoteness of magnetic material provenance, local geochemical and geomorphological conditions, peculiar features of rock lithology, composition of underlying rocks, time, and others. With such a multi-factor dependence, we should establish basic causes responsible for MS variations to understand their implications for particular problems. In other words, it is necessary to know the specific physical mechanism responsible for origin of a magnetic signal, and only after this, one can judge whether the posed problem will be solved using MS measurements. This requirement is a basic one in the methodological approach that allows obtaining of non-controversial results when using methods of rock magnetism and, particularly, the MS measurements (Bol'shakov, 1997, 2000). Unfortunately, when requirements of this methodological approach are inconsistently followed, we obtain paradoxical results. For instance, the spectral analysis of MS parameters carried out by Beget and Hawkins (1989) and by Kukla *et al.* (1990) revealed periodical changes of κ values in loess–soil sequences of Alaska and China, which are comparable with variation cycles of orbital elements. Regretfully, secular changes of sedimentation rate has not been taken into account in the first case, and incorrect physical prerequisites have been used in the second attempt to take this factor into consideration (Bol'shakov, 1997, 2000). To put it another way, time scales in the both cases were primarily inconsistent and incorrect, although the results of the spectral analysis did show that periods of κ variations are correlative with orbital cycles. Accordingly, the important conclusion (correct, ~~in the authors' opinion,~~ but unjustified from my standpoint) has been drawn that orbital variations exerted a decisive effect on loess formation in our planet. My analysis of inferences and conclusions, presented in works mentioned above (Bol'shakov, 2000) and in some others I refer to below, suggests that there is a certain mystic attitude of authors to the possibilities of MS method, although their own empirical data refute universalism of the latter. Such examples of interpreting MS measurements cause the distrust only.

The work by Thouveny *et al.* (1994) is one of few others, in which MS measurements are correctly used in terms of methodology. Examining deposits of volcanic lakes of France, they proposed a simple model of climate-induced κ variations that was supported by data of paleobotanical, granulometric, mineralogical, geochemical, and diatomic analyses. Under cold conditions, freezing–thawing processes in the ground enhanced erosion of volcanic rocks and deposition of clastic material (κ increase). At the time of warming, the accelerated processes of soil and plant formation lead to the intensification of organic sedimentogenesis that increases proportion of nonmagnetic fraction (κ decrease). With the help of radiocarbon dating and methods mentioned above, paleoclimatic variations during the last 140 ka were comprehensively studied, and a "fine structure" of climatic changes during the oxygen-isotope (OI) stage 5*e* of the last interglacial period was confirmed. Thouveny *et al.* (1994, p. 503) pointed out: "The magnetic susceptibility is thus strongly sensitive to variations of the local climate and constitutes an accurate proxy record, particularly useful when paleobiological or organic geochemical information loses some degree of significance (that is, under the coldest conditions)."

The aim of this work is to analyze the concrete examples of using MS measurements as a tool of geological correlation and other works related to different aspects of paleoclimatic studies. The analysis is dedicated to rocks of the loess formation (LF) in different regions of the world, because the MS method was most widely used to study these rocks.

Before the analysis of publications, I should note that the MS measurements are especially effective when κ values exceed 10^{-4} SI units. Such values usually imply presence of ferrimagnetic minerals (magnetite, maghemite, greigite or titanomagnetite) in sediments. If κ values are lower, the interpretation of MS measurements is difficult, requiring additional studies. A good illustration is the work by Hounslow and Maher (1999) who showed, using data of the X-ray study, Mössbauer spectroscopy, electronic and optic microscopy, and sediment dispersion, that κ values lower than 5×10^{-5} SI units are ineffective for acquiring paleoclimatic information from MS measurement results. This example also shows that low magnetic susceptibility of deposits deprives the MS method of its advantages, which are the insignificant time consumption, low cost, and possibility to obtain data without destruction of studied subjects, i.e., to carry out measurements directly in a natural exposure or in a core section of bottom sediments. One more advantage of MS measurements is possibility to obtain data in a digital form suitable for a direct processing and further interpretation.

CRITICAL ANALYSIS OF SOME RESULTS BASED ON MAGNETIC SUSCEPTIBILITY MEASUREMENTS

Geological correlation of Pleistocene deposits and interrelated climatic interpretations are usually based on comparison of MS parameters and OI data. Without this, the MS measurements are much less effective, as the widely used OI time scale is climatochronostratigraphic in essence.

The effect of climatic conditions on the magnetic susceptibility of soils has long been known (Le Borgne, 1960; Babanin, 1971; Vadyunina and Babanin, 1972; Tite and Linington, 1975; Mullins, 1977; Smirnov, 1978). Paleoclimatic aspects of magnetic susceptibility measurements in loess and soil were also touched upon in these works. In due time, Mullins (1977) warned against hasty attempts to use the κ values of soils (or ratio between these parameters in soils and loess κ_{bs}/κ_l) to solve the reverse problem, i.e., to determine paleoclimatic conditions during the formation time of LF horizons based on results of MS measurements. His was aware of multifactor dependence of κ values on various, not only climatic, conditions (for instance, on composition of underlying rocks), as is mentioned above.

Among the early and well-known publications, there are works by Heller and Liu (1984, 1986), in which the MS variations in LF horizons of China (the Luochuan section) were attributed to influence of paleoclimatic factors. They concluded that magnetic susceptibility could be indicative of paleoclimatic changes based on a regular variation of κ values throughout the section: κ values are higher in soils, formation of which is related to interglacial stages, and lower in loess horizons deposited during glaciation stages. When Heller and Liu correlated MS parameters and OI data, they found them rather consistent with each other. The consistency is, however, not absolute, not "one-to-one" as mentioned in their work published in 1986. The main apparent inconsistency is that the Matuyama–Brunhes (M/B) reversal recorded in the OI Stage 19 in deep-sea sediments is at the level 21st to 23rd excursions in the Luochuan section, which corresponds to the buried soil 8 (BS 8). Admitting the chemical nature of characteristic rock magnetization in this section, Heller and Liu (1986) shift the B/M reversal boundary to the loess bed about 1 m above the BS 8 roof, whereas in earlier publications (Heller and Liu, 1984; Liu *et al.*, 1984) it was established inside and even at the base of BS 8. Displacement of inversion records during secondary magnetization is a well-known phenomenon (Bol'shakov, 1995, 1996), although the displacement magnitude should be assessed in a special manner (Bol'shakov, 1999). The displacement in question is voluntary and insufficiently justified, because loess, in the authors' opinion, accumulated under cold arid conditions, whereas the B/M reversal is established in the warm OI Stage19.

Heller and Liu (1986) also attempted to carry out the spectral analysis of MS parameters. They accepted the time scale based on paleomagnetic time markers and the constant sedimentation rate throughout the section, but did not obtain a significant result. Tuning the κ values to insolation curves estimated by Berger (1978), they revealed however periodical changes approximating cycles of 40 k.y., which characterize variations of the equator plane (obliquity) to the ecliptic plane. Heller and Liu (1986) concluded that cycles of 23 k.y. are missing because of a low effect of precession-related seasonal variations of insolation on the LF origin in China. This inference is doubtful, as the territory of China is under influence monsoons (paleomonsoons) (Banerjee *et al.*, 1993), the intensity of which is interrelated with precession of the earth's axis (Rossignol-Strick, 1983; Prell and Kutzbach, 1987). It is even stranger that periods of 100 k.y. of κ variations, which are clearly seen in Fig. 1 borrowed from the work by Kukla *et al.* (1988), have been missing. I did not plot MS data for the Luochuan section reported by Heller and Liu (1984, 1986), because they used a logarithmic scale that conceals some details, and because the both are co-authors of the work by Kukla *et al.* (1988). The mentioned unsatisfactory results of the spectral analysis are likely a consequence of insufficiently elaborated time scale that was applied to the Luochuan section. In addition, the effect of various factors on the κ value

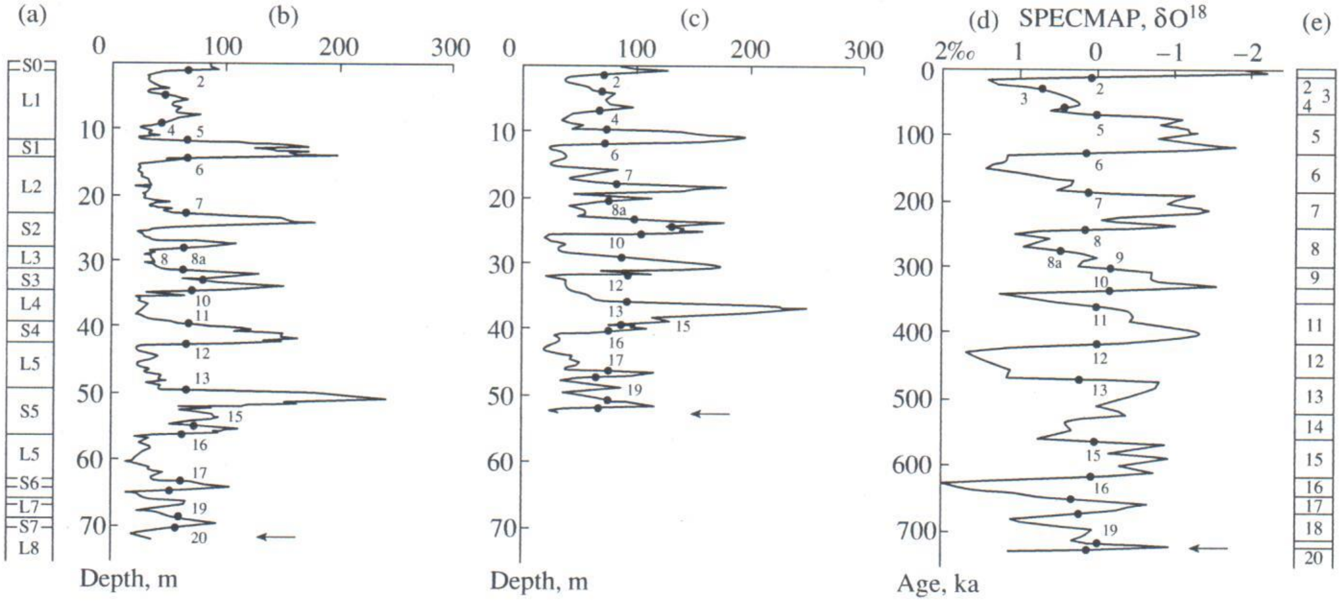

Fig. 1. Lithologic units (a) and magnetic susceptibility (2×10^{-5} SI units) variations in the Xifeng (b) and Luochuan (c) sections correlated with the SPECMAP oxygen-isotope scale (d) and OI stages (e): (L) loess, (S) soil. Large dots mark peaks of OI Stages and their time-equivalents in loess sections. Arrows indicate the position of the Matuyama–Brunhes reversal; all plots after Kukla *et al.* (1988).

variations was inadequately assessed, and interrelations between MS variations and paleoclimatic changes were improperly understood and only declared in works by Heller and Liu (1984, 1986). It is important, however, to substantiate the paleoclimatic nature of each significant minimum and maximum of κ values for the correctness of the spectral analysis.

Similar drawbacks are characteristic of work by Chlachula *et al.* (1998) who studied the Kurtak section in the Ob' River valley (Siberia) and revealed inverse MS parameters characterizing buried soils and loess beds, as κ values turned out to be higher in the latter (Fig. 2), like in the LF sections of Alaska (Beget and Hawkins, 1989; Beget *et al.*, 1990). Chlachula *et al.* (1998) connect κ variations in loess beds with climate-related changes in the wind intensity, giving no proofs in favor of this "Alaskan model" that does explain κ values in loess and soil. As was shown (Bol'shakov, 1997, 2000), the wind-induced MS variations have not been properly justified even by authors of the Alaskan model (Beget and Hawkins, 1989; Beget *et al.*, 1990). In particular, the MS increase in sediments of coarser grain size (Beget *et al.*, 1990) could result not only from eolian, but also from fluvial reworking of sedimentary material. Moreover, the model does not take into account the effect of chemical changes, which are established in loess beds of Alaska (Beget and Hawkins, 1989), on the κ value. That is why the correlation of κ values in the Kurtak section (Siberia) with those in the piston core V21-146 of Pacific sediments seems to be unconvincing. Similarly doubtful are substages distinguished (Fig. 2) in the OI Stage 5 (Chlachula *et al.*, 1998). Substages like 5a and 5b (Fig. 2c) can be distinguished as well in the overlying κ minimum, at the depth of about 10 m, the more so as stratigraphic position and age of BS horizons has not been substantiated in this section.

Thus, data of MS measurements in LF sections of China, Siberia, Alaska, and some other regions imply that *ratio* κ_{bs}/κ_l *characterizing sequences of buried soils and loess beds varies in space and can be equal, greater or less than one*.

In view of frequently exaggerated (for particular situations) paleoclimatic implications of MS measurements in loess and soil sections, of special interest are works, in which κ variations are directly correlated with paleoclimatic curves obtained on the basis of palynological or paleontological data. For instance, Bolikhovskaya *et al.* (1999) studied one of the characteristic LF sequences exposed in the Strelitsa reference section of the central Russian Plain. The correlation of paleoclimatic variations, inferred from palynological data, with data of MS measurements was a main objective of their work. Unfortunately, the interpretation and resulting inferences appeared to be controversial, as it is evident from the following principal conclusion: "In the Strelitsa section, like in other LF sequences, maximums of specific initial magnetic susceptibility (denoted as χ, V.B.) correspond to warm interglacial periods, whereas minimal or lower values of this parameter mark the cold glacial epochs. Paleoclimatic variations appear to be smoothed in initial magnetic susceptibility profiles lacking endothermal cooling phases. The role of gleying that distorts the paleoclimatic record in magnetic profiles is noticeable"

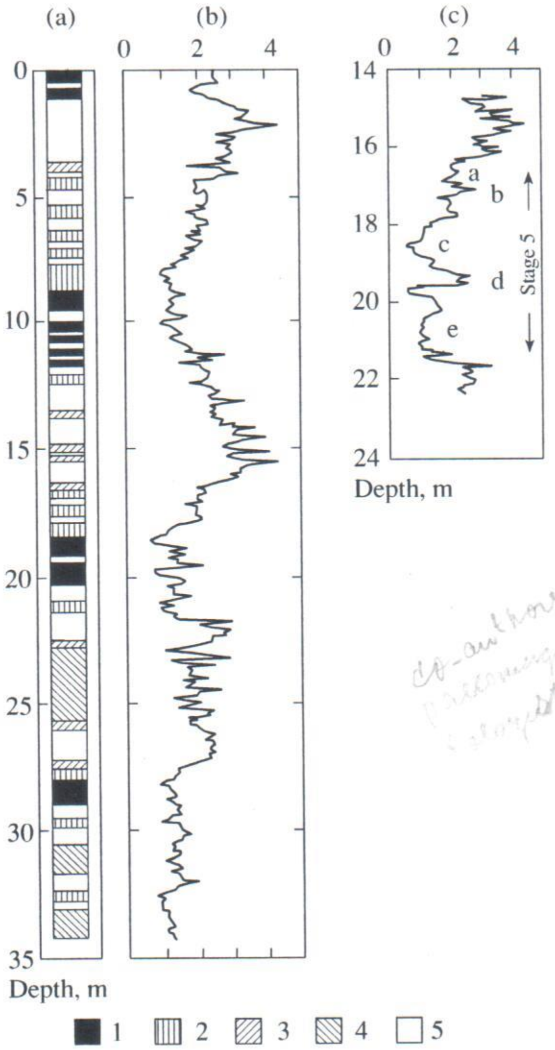

Fig. 2. Lithological column (a), magnetic susceptibility variations in the Kurtak section (b), and OI Stage 5 divided into substages 5a–5f (c) (after Chlachula *et al.*, 1998): (1) chernozem; (2) brown soil; (3) leached gley soil; (4) colluvial loess; (5) eolian loess (specific magnetic susceptibility in 10^{-6} m^3 kg^{-1}).

(Bolikhovskaya *et al.*, 1999, p. 690). It is evident that the initial inference (first sentence) is considerably weakened by conclusions following it. Moreover, the second and the third sentences are also controversial, conflicting with each other. The second one suggests that χ variations are paleoclimatic in origin, whereas the third sentence means that the paleoclimatic record in magnetic profiles is distorted by gleying. It would be logical to speak in this case about a distorted MS record of paleoclimatic events that actually implies defects of the record. My conclusion can be convincingly confirmed, probably for the first time, by concrete data presented in the work by Bolikhovskaya *et al.* (1999) (Fig. 3). For instance, transitional χ values correspond to soil horizons Romenka AC (interglacial phase) and Kamenka AI[1] (glaciation phase), and the average χ value of the AI[1] horizon is not lower than that of the AC horizon, as one can seen from the plotted χ variations. (I use the term transitional values instead of higher or lower values used in the original work, because the latter are misleading if the reference value is not indicated. For instance, values lower than a maximum may be easily greater than values specified as higher than a minimum.) Further, minimal χ values of the horizon B of the Inzhava soil correspond to the interglacial period, and transitional χ values characterizing the Korostel loess and AI[1] horizon of the Voron soil mark glaciation episode. Moreover, the χ maximum of AI horizon has no "lag behind the climatic optimum," as is stated in the work, but corresponds to the endothermal cooling phase of the Muchkap Interglacial, the beginning and climatic optimum of which are characterized by the minimal and transitional χ values (B and AB horizons of the Voron soil, Fig. 3). The ultimate refutation of the first sentence in above conclusion is a merit of experts in paleomagnetology (Virina *et al.*, 1998). They argued that ratio between χ values of loess and BS could be greater or less than one in different regions (I already mentioned this above) and pointed out that the χ value minimum is recorded in the podzolic horizon of the Mezin soil correlated with an interglacial stage. Another important fact is also presented in the work by Virina *et al.* (1998). Low χ values are typical of gley soils in the loess–soil sequence below morainal deposits that has not been considered in the work by Bolikhovskaya *et al.* (1999).

co-authors palaeomagnetologists

Thus, the analysis of two publications (Bolikhovskaya *et al.*, 1999; Virina *et al.*, 1998) reveals diverse relations between κ values of loess and soil horizons in the Strelitsa section and climatic events recorded here. In contrast to authors who insufficiently substantiated their controversial inferences with respect to paleoclimatic implications of κ values, I arrived at two following conclusions: (1) ratio κ_{bs}/κ_l can substantially vary in space and with time (might be greater of less than one); (2) variations of κ values in the LF sections *not always reflect* climatic changes ~~recorded in the latter~~ and could be controlled by nonclimatic factors.

Clearly, the second conclusion does not deny that the κ value *may reflect* paleoclimatic changes. For instance, this is evident from work by Pospelova and Levkovskaya (1994), in which concurrent variations of MS parameters and palynological data in a soil bed the Pogrebya section (Moldavia) are established.

The work by Maher *et al.* (1994) is another example of the comprehensive analysis used to establish κ variations in soils and corresponding climatic changes, and to apply the obtained data for a concrete paleoclimatic interpretation. In this work, the magnetic susceptibility of recent soils of the Loess Plateau (China) was correlated in a mathematical form with amount of precipitation. Both parameters increase southeastwards. Their quantitative connections were used to assess variations of precipitation in the past. As the authors themselves

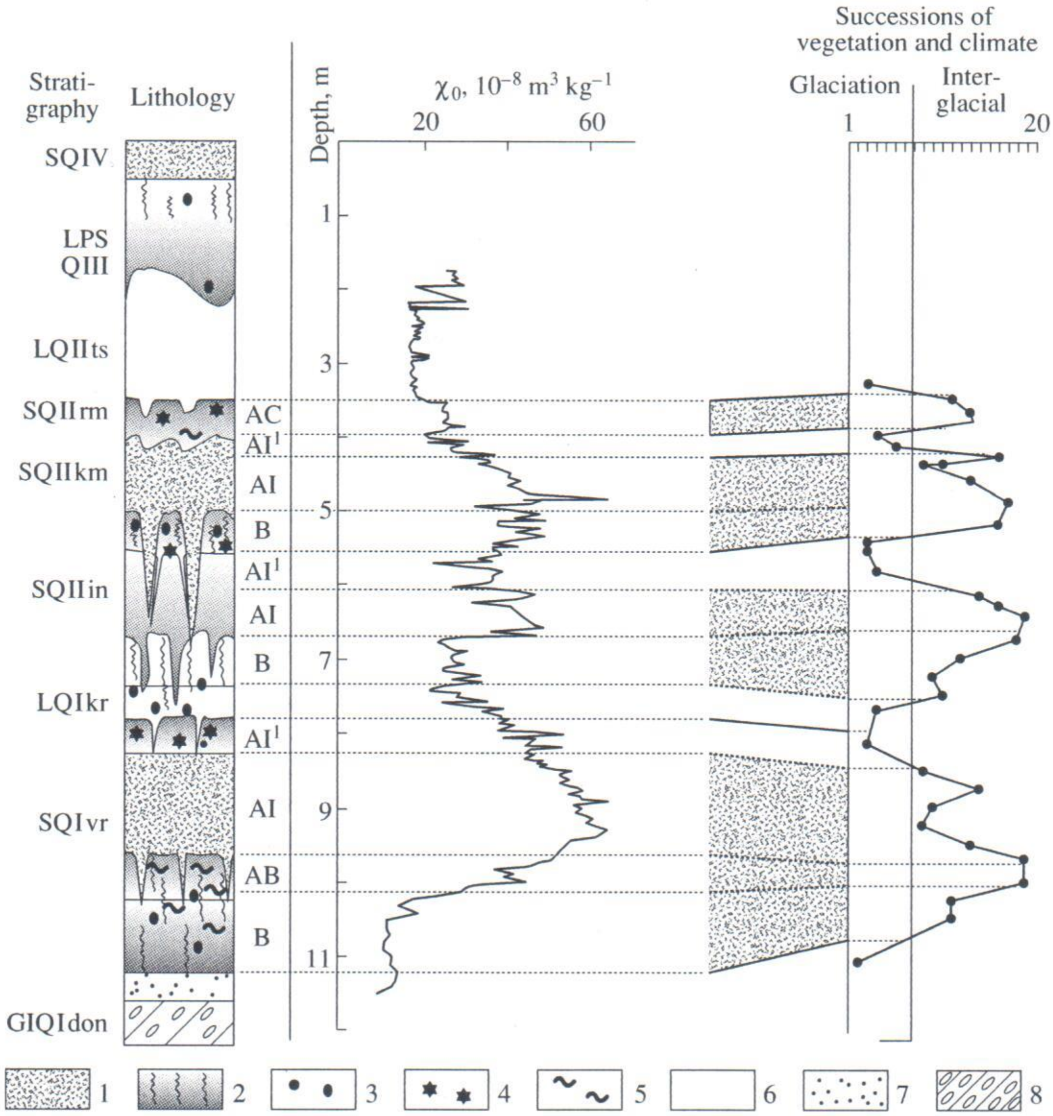

Fig. 3. Stratigraphy, lithology, κ variations, and paleoclimatic curve characterizing the Strelitsa section (after Bolikhovskaya *et al.*, 1999): (1) horizons A of buried soils; (2) horizons B of buried soils; (3) mole runs; (4) ferruginate patches; (5) gleying; (6) loess; (7) sand; (8) Don morainal bed. Letter symbols: (ts), Tyasma, (rm) Romenka, (km) Kamenka, (in) Inzhava, (kr) Korostel, and (vr) Voron loess and soil horizons. Shadowed areas near paleoclimatic curve correspond to interglacial stages.

note, the precipitation value they estimated for the last interglacial stage substantially exceed those obtained on the basis of other methods, in particular, by using the model of the general circulation in the atmosphere. In my opinion, this is a consequence of two essential drawbacks of the used approach (Maher *et al.*, 1994). First, assuming that κ values in soils are controlled by pedogenesis, they admit the soil formation to be dependent on the amount of precipitation only. It is known, however, that pedogenesis depends, for instance, on the sedimentation rate. Under a higher rate of eolian material sedimentation, the soil formation will be suppressed, and the κ value in soils will be lower. This inference agrees well with data on the Loess Plateau of China: κ values increase, as thickness of loess–soil horizons decrease here in the direction from Lanzhou to Luochuan (Burbank and Li, 1985; Kukla *et al.*, 1988; Hus and Han, 1992; Evans and Heller, 1994). Hence, formula describing dependence of κ value in soils on the amount of precipitation should include parameter of sedimentation rate as well. Accordingly, the κ value dependence on the amount of atmospheric precipitation remains ambiguous, since we do not know what kind of relations, if any, connects sedimentation rate and atmospheric precipitation, and how constant they are. One more complexity is a possible effect of temperature variations within the Loess Plateau, which likely take place and may influence pedogenesis as well, and, hence, the κ value of soils.

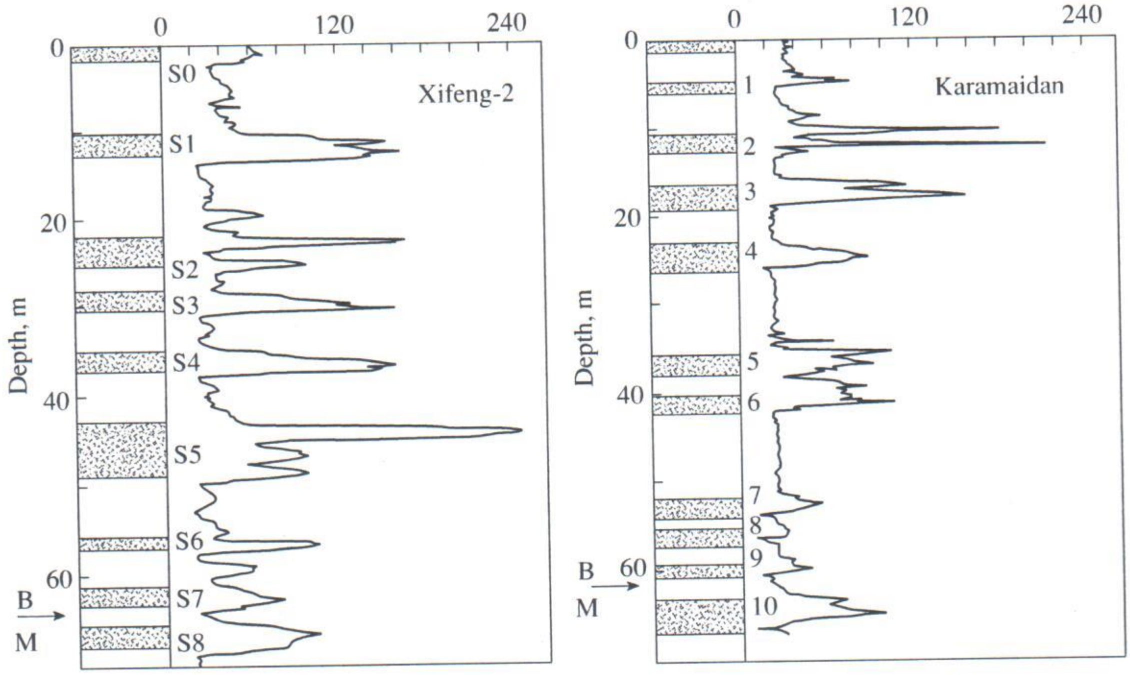

Fig. 4. Magnetic susceptibility (10^{-8} m^3 kg^{-1}) variations in the Xifeng-2 and Karamaidan sections; soils enumerated from the top downward are shown in black, and arrows show position of the Matuyama–Brunhes reversal (after Shackleton *et al.*, 1995, Fig. 1).

Second, Maher *et al.* (1994) ignored possible transformations of magnetic minerals in soils after their burial. I reported on such (at least, qualitative) transformations in buried soils of Bulgaria, similar in magnetic characteristics to soils of China (Bol'shakov, 1996). The possibility of secular changes in BS magnetic properties is implied by obvious diagenetic effects resulting, for instance, by the post-burial decrease of humus content in BS that is progressing with time (Kriger, 1965). Such a process, if it takes place, eliminates possibility of paleoclimatic interpretations based on the comparative analysis of κ value variations in buried and recent soils.

Kukla *et al.* (1988) initiated consideration of MS parameters as chronostratigraphic indicators. Assuming that magnetite concentration, i.e., the κ value in loess or soil, depend on the loess accumulation rate, they proposed a model, according to which the natural magnetite dust from the high atmospheric layers accumulated with a *constant* rate on the Earth's surface. Consequently, if the κ value and thickness of sedimentary layer are known, one may determine the time interval of sedimentation, provided the continuous character of sedimentation. Another necessary condition is presence of two dated levels in the section (usually the beginning of the Holocene and the M/B reversal), and age of layer m can be determined then by formula:

$$T_m = T_1 + \left(\sum_{i=1}^{m} a_i \chi_i\right)(T_2 - T_1)\left(\sum_{i=1}^{n} a_i \chi_i\right)^{-1}, \qquad (1)$$

where T_2 and T_1 are maximal and minimal ages of levels dated in the section; n is amount of layers between dated levels; a_i and χ_i are thickness and magnetic susceptibility of i-layer. Using this model, Kukla *et al.* chronologically correlated the Luochuan and Xifeng sections of the Loess Plateau in China. Their correlation with the SPECMAP oxygen-isotope scale was also adequate (Fig. 1). Later on, it was shown, however (Bol'shakov, 2000, and references therein), that the model under consideration is invalid because of κ variations in response to conditions of pedogenesis. Nevertheless, Shackleton *et al.* (1995, p. 2) attempted to revive the considered model admitting the following mechanism of κ variations: "To a first order approximation, the flux (formation and/or deposition) of the susceptibility carrier is considered constant…". It is evident that the assumed constant "formation" of susceptibility carrier is as much unsubstantiated as the previous condition of constant subaerial deposition. In the last work, chronological correlation between the Xifeng-2 section of China and the Karamaidan section of Tajikistan was based on the MS measurements (Fig. 4). Figure 4 well illustrates the impossibility of correlating these two sections located far apart by means of κ variations as such, although this is admissible for the Loess Plateau section in China (Fig. 1).

The above discussion means that, despite the originality, the MS time scale elaborated by Kukla and his colleagues is invalid, since it is based on incorrect physical prerequisites. According to postulated mechanism, the κ value depends only on sedimentation rate: the lower the rate (in soils) the greater the κ value. According to given comments, this mechanism may work only in regions with well-drained soils (for instance, in China, but not in Alaska or Siberia). Nevertheless, the mechanism is inappropriate even for China, because κ values depend, according to data mentioned above, on one more parameter—on precipitation value (Maher *et al.*, 1994). Then why the MS time scale yields reasonable results (Kukla *et al.*, 1988, 1990; Shackleton *et al.*, 1995)? Here, we should take into consideration two important moments. First, all the age values are originally constrained by strict time boundaries: by the M/B reversal and by the Holocene base or by the Mikulino soil complex (SC 1 in Fig. 4). Provided the absence of sedimentation breaks, these limitations ensure a reasonable dating of loess and soil horizons within the corresponding time intervals. The dates will be even more precise if variability of the sedimentation rate can be taken into account. This is just the second moment conditioning the reasonable dating by formula (1), which puts the κ value into correspondence with the formation time of the given sequence. The correspondence is solely qualitative, as I should repeat that we have no reason to state that soils with low κ values (for instance, lower soils in Figs. 1 and 4) accumulated faster than soils with greater values of this parameter. Similarly founded, or to be more exact, unfounded, one may suppose that sedimentation rate by accumulation of loess horizons is greater by factor of two, three, four, etc., than soil formation. In this case, the age can be calculated from the next formula:

$$T_m = T_1 + \left(\sum_{i=1}^{m} a_i k_i\right)(T_2 - T_1)\left(\sum_{i=1}^{n} a_i k_i\right)^{-1}, \qquad (2)$$

where k_i equals 1 for even-numbered i, or 2, 3 and 4 for odd-numbered i, if we begin to count from the loess layer (accordingly, loess corresponds to odd-numbered I). It is also possible to specify various k_i values for soils with different profiles. Then we can precise again ages of horizons as compared to the alternative approach implying the constant sedimentation rate. In fact, dating by formula (2) is more general than by formula (1) and can be applied to both the automorphic and hydromorphic soils.

Reverting to work by Shackleton *et al.* (1995), I should point out that the MS time scale yielded poor correlation of the Xifeng-2 and Karamaidan sections with the OI scale. Therefore, they had to adjust the MS variation profile to the OI scale. The unavoidable consequence of adjustment, which is not discussed in above publication, is the necessity to displace the M/B reversal from the loess horizon 7 to overlying soil horizon S 7 (Fig. 4).

Closing the discussion of results obtained by Shackleton *et al.* (1995), let us consider their main inference. It implies that of eolian processes in China and Tajikistan depended on dynamics of the global ice volume changes, which is recorded by OI variations in deep-sea sediments, to a greater extent than on theoretical variations of insolation. This inference is based not on interpretation of measured MS parameters, but on adjustment of climatic cycles recorded in the sections to the OI scale. The inference seems reasonable from theoretical side as well. The point is that, as shown in (Bol'shakov V. and Bol'shakov P., 1999), Milankovitch theory has some essential drawbacks, and insolation curves calculated on the basis of this theory disagree with OI data and cannot be used for paleoclimatic interpretations. The last work *directly* demonstrated a decisive effect of insolation, controlled by variations of the Earth's orbital elements, on the global ice volume changes during the Pleistocene. These changes determined, in turn, alternation of glacial and interglacial periods detectable in OI records data, and accompanying reorganizations in the system of atmospheric and oceanic circulation. During glacial periods, for instance, the atmospheric circulation should be intensified because of the greater latitudinal gradient of temperature. That is why the loess formation indicative of the intense atmospheric circulation is directly correlated with glaciations (if we leave aside changes in humidity of a given region).

DISCUSSION

Some works, in which the MS measurements are used to reveal paleoclimatic conditions during loess and soil formation, are considered above. It is possible to state *a priori* that global climatic changes, which influence sedimentation conditions, should be reflected in variations of magnetic properties of LF sequences. These variations are also characteristic of recent soils. For instance, Babanin (1971) established that soil horizons A and C have variable magnetic susceptibility: ratio κ_A/κ_C varies from one or lesser value in sod–podzolic soils to three–four in chernozem and chestnut soil. However, not only climatic but many other factors independent of climate control the κ value of soils. I should again mention the works by Bolikhovskaya *et al.* (1999) and Virina *et al.* (1998), according to which gleying may substantially decrease the κ value in soils, thus reducing to zero the global climatic signal that could be recorded otherwise. On the other hand, global climatic changes themselves are very diverse and complex, variably affecting the environmental parameters, such as temperature, humidity, and others. These changes are nonuniform in different regions of the planet. They all are responsible for the space-and-time variations of κ values in loess and soil.

Hence, the effect of paleoclimate on magnetic properties of sediments cannot be refuted. It is difficult to predict, however, the quantitative changes of magnetic susceptibility in particular LF sections in response to climatic factors, if we do not know the concrete relationships between κ value variations and changes in environmental conditions. It is even more difficult to solve the reverse problem, i.e. to reconstruct paleoclimatic changes based on MS variations. It is pertinent to ask here what the MS measurements mean in terms of paleoclimatic reconstructions, which are under discussion in many works? The magnetic susceptibility parameter represents a particular value and nothing else. Hence, *the application of MS parameters for paleoclimatic reconstructions is a procedure independent of space and time that means the direct correlation of particular κ value intervals with corresponding climatic parameters* (temperature, humidity, direction and intensity of wind, which affect sedimentation rate). Works considered above show that this problem cannot be solved at the global scale, since κ values of loess and soil, as well as ratios κ_{bs}/κ_l depend on local and regional climatic factors. Could it be solved within a separate region, in which the climate-induced κ variations are uniform, as for instance, in sections of the Loess Plateau of China? It is a common practice to consider paleoclimatic changes in generalized qualitative terms like glaciation or interglacial stage, but the problem I pose now is much more complicated. To solve it, it is necessary to establish first the concrete dependence of κ values on every significant climatic factor. Even in the case of a considerable progress in this direction, there is no way to ensure the unambiguous correlation between climatic changes and MS variations and, hence, the unambiguous solution of the reverse problem. The κ value depends, at least, on three parameters, and the unequivocal solution is possible only if these climatic parameters (temperature, moisture, wind intensity) are interrelated in a definite manner. The question whether or not such interrelation is definite and invariant with time is not useless. In the work by Banerjee *et al.* (1993), it is assumed that variations in magnetic properties of Holocene soils, in contrast to soils of earlier interglacial stages, are caused by concurrent changes in the paleomonsoon circulation. Clearly, this should change the regional humidity and eolian transport of sedimentary material in the region, and, hence, the new variations of κ values.

Thus, we are far away from the moment, when correlation between κ values and global climatic changes determining specific dependences like κ–temperature or κ–humidity could be established for particular sections, and when the concrete physical mechanisms enabling the correct interrelation could be understood (if it is possible to found them). Only after positive results obtained on this way of comprehensive study, we may pass from speculations to the correct interpretation of MS measurements in paleoclimatic terms. Only such a strict scientific approach may facilitate solution of the discussed reverse problem, and only it may create a basis for the correct determination of paleotemperatures or paleohumidity with the help of MS measurements, in contrast to immature solutions presented in some works (Maher *et al.*, 1994; Heller, 1998). One should remember therewith the possible diagenetic transformations of magnetic minerals after their burial in loess and soil horizons. If this and another non-climatic factors have a decisive effect on magnetic susceptibility, then the correct solution of the reverse problem appears to be impossible.

Taking into account all complications, which appear when we try to correlate the continental LF (MS variation curve) and marine (OI curve) sedimentary sequences, I advise to use instead either the reference loess (glacial) and soil (interglacial) horizons, or traditional paleoclimatic curves (arid and cold versus humid and warm intervals). Correlations of this kind were carried out in some works (Liu *et al.*, 1984; Ding *et al.*, 1991; Dodonov and Baiguzina, 1995), and the results obtained are as good as those based on interpretation of MS measurements. For instance (Heller and Liu, 1984, 1986; Kukla *et al.*, 1988; Liu *et al.*, 1992), the number of κ-stages distinguished within the Brunhes Chron and below the Olduvai Subchron of the Loess Plateau sections is variable. At the same time, the number of OI Stages correlative with loess and soil horizons below the Olduvai Subchron (Ding *et al.*, 1991) is closer to that presently accepted in contrast to correlations based on MS measurements. The imperfect correlation between deep-sea and continental sedimentary sequences is not only a consequence of mistakenly distinguished stages based on MS measurements. Some inconsistencies can be associated with incompleteness of continental geological records deposits and with ambiguous subdivision of soil complexes (SC). The comparative analysis of works (Dodonov, 1986; Dodonov and Baiguzina, 1995; Shackleton *et al.*, 1995; Liu *et al.*, 1984; Kukla *et al.*, 1988) suggests that the distinguishing of SC ranks (interglacial, interstadial) and individual soil horizons is a subjective procedure, on which the number of MS stages or soil complexes used in correlation depends.

One more problem is positioning of the M/B reversal in LF sequences. The OI scale and paleomagnetic data on deep-sea sediments are very helpful in this case, but the last word should be that of experts in paleomagnetism and magnetism of rocks.

CONCLUSION

Works analyzed earlier (Bol'shakov, 2000) and above lead to the conclusion that interrelations between MS variations in LF sequences and global climatic changes are ambiguous. Climatic impacts can (or cannot) be manifested in different manner not only in LF sequences of various regions, but also in different parts of one section. Two main reasons responsible for this are as follows: (a) MS variations in rocks sequences are controlled by various, frequently independent factors, some of which have no direct relation to global climatic changes; (b) global climatic cycles of the Pleistocene dissimilarly manifested themselves in space and with time.

Summing up the available attempts to use the MS parameters of LF sequences for paleoclimatic reconstructions, I should point out the following. The MS variations are ineffective for establishing qualitative climatic changes of the "glaciation–interglacial stage" type, since a more simple indicator of these events is color of corresponding horizons stable in space and with time. The light color of loess usually indicates a glaciation, whereas the dark color of soil points to an interglacial stage. Attempts to use quantitative MS parameters as indicators of paleoclimatic changes (in temperature, humidity, and duration of climatic cycles), have no proper methodological substantiation so far, since we do not know in detail how the κ value depends on all factors controlling sedimentation conditions. Even in the case of automorphic soils of China, it is difficult to interpret the MS measurements, because it is still unknown what mineral, magnetite or maghemite, is responsible for the magnetic susceptibility increase in buried soils (*Pedogenesis and paleoclimate*..., 1994).

Hence, it is clear that the idea of special importance of MS measurements in LF sequences that is declared by some researchers, and has turned in their minds into a material power forcing them to make unjustified statements, is not a real knowledge as yet. Attempts to use the MS measurements even in those cases, when they were unnecessary, imply that ~~we have no~~ there is a lack of adequate method to study the Pleistocene LF formations.

Application of MS measurements for geological correlation of different sections can be successful in regions with rather uniform sedimentation conditions. This concerns not only LF sequences but other sedimentary rocks as well (Thompson *et al.*, 1980; Verosub and Roberts, 1995). The deviation of κ values from those typical of the given region can be regarded as an express-signal informing us that regional sedimentation conditions changed in the studied locality. Such an express-signal, if available, implies that we have to give special attention to horizons "anomalous" in terms of the MS parameters, and that they deserve a comprehensive study (application of new research methods or more detailed sampling). The κ value is a very sensitive indicator characterizing the lithological peculiarities of deposits, which are often invisible. That is why the MS measurements can be successfully used for revealing lithological distinctions of sedimentary rocks. In particular, the MS method has been widely applied to study the LS horizons and sedimentary core sections from sea bottom. For instance, it helps to distinguish horizons of poorly developed buried soils during the field work.

Thus, only the methodologically correct scientific approach to investigation of magnetic susceptibility in sedimentary rocks can turn the MS method into a useful tool complementing traditional methods of paleogeographic studies.

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