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Constraints for paleoclimate studies

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Patterns and timing of loess-paleosol transitions in Eurasia: Constraints for paleoclimate studies

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ABSTRACT

Loess-paleosol sequences are the most extensive terrestrial paleoclimate records in Europe and Asia documenting atmospheric circulation patterns, vegetation, and sedimentary dynamics in response to glacial-interglacial cyclicity. Between the two sides of the Eurasian continent, differences may exist in response and response times to glacial changes and finding these is essential to understand the climate systems of the northern hemisphere. Therefore, assessment of common patterns and regional differences in loess-paleosol sequences (LPS) is vital, but remains, however, uncertain. Another key to interpret these records is to constrain the mechanisms responsible for the formation and preservation of paleosols and loess layers in these paleoclimate archives. This study therefore compares LPS magnetic susceptibility records as proxies for paleosol formation intensity for selected sites from the central Chinese Loess Plateau and the Carpathian Basin in Europe over the last 440 kyr. Inconsistencies and crucial issues concerning the timing, correlation and paleoclimate potential of selected Eurasian LPS are outlined.

Our comparison of Eurasian LPS shows generally similar patterns of paleosol formation, while highlighting several crucial differences. Especially for paleosols developed around ~200 and ~300 ka, the reported timing of soil formation differs by up to 30 ka. In addition, a drying and cooling trend over the last ~300 ka has been documented in Europe, with no such evidence in the Asian records. The comparison shows that there is still uncertainty in defining the chronostratigraphic framework for these records on glacial-interglacial time scales in the order of 5–30 kyr for the last ~440 ka. We argue that the baseline of the magnetic susceptibility proxy in loess from the Carpathian Basin is the most striking difference between European LPS and the Chinese Loess Plateau. In our opinion, many of the current timing/age differences may be overcome once a comparable stratigraphic interpretation is achieved.

1. Introduction

The recognition of loess-paleosol sequences (LPS) as some of the most spatially extensive terrestrial paleoclimate archives in Europe and China (Heller and Liu, 1984; Kukla, 1977, 1978; Kukla and An, 1989; Liu et al., 1985) led to a wide interest in exploring their potential in paleoenvironmental reconstructions. This seminal scientific achievement led the way to detailed studies on loess stratigraphy, patterns in dust genesis, emission and deposition, lateral variability of loess physical and geochemical characteristics, and their direct comparison with chronologically better constrained marine paleoclimate series, at least in the resolution of orbital time scales (e.g. Bronger, 2003; Rousseau and Puisségur, 1990; Sun et al., 2006a). However, the direct correlation of loess geoarchives between Europe and the

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(Central) Chinese Loess Plateau ((C)CLP) is of major importance for the understanding of Eurasian continental climatic temporal and spatial evolution (e.g. Bronger, 2003; Gendler et al., 2006). Several recent publications attempted at correlating south-eastern European loess proxy data with records from the CCLP. Albeit partly inconsistent among each other from a chronological viewpoint (Basarin et al., 2014; Buggle et al., 2009; Marković et al., 2012), as presumably corresponding loess/soil units between records are bracketed by different starting or ending ages, these studies provide reference records with correlative (magnetostratigraphy, correlation of proxy data to reference data) age control.

Here we focus on the last ~440 ka, because reliable high-resolution (resolution allowing clear interpretation to a ~20 kyr scale facilitating the identification of orbital precession, well-documented age model construction) magnetic susceptibility (MS) records from Europe are limited to this time frame. Further back in time in European sequences, no precession scale variability can be identified, most probably due to lower accumulation rates and insufficient sampling resolution. MS is widely used in assessing the intensity of soil formation, a proxy reflecting sediment/soil moisture, which in turn impacts on the degree of post-sedimentary silicate weathering. Thus, MS might be considered as a proxy for past moisture availability, but may also reflect temperature influence via weathering and evaporation (An et al., 1991; Buggle et al., 2013, 2014; Han et al., 1996; Heller et al., 1993; Maher et al., 1994; Song et al., 2014). During relatively humid (and warm) climate phases, soils developed on the substrate loess, and the magnetic susceptibility is enhanced by weathering of common iron-bearing minerals (e.g. Peng et al., 2014) and subsequent (microbial) neo-formation of iron-oxides including magnetite and maghemite (e.g. Maher et al., 1994). The high MS of these minerals is the main origin of magnetic susceptibility enhancement in the course of pedogenesis and therefore directly reflecting climatically-controlled sediment/soil-moisture variations (e.g. Buggle et al., 2014 and references therein).

We further argue that inconsistent time-scale construction and correlations to different reference datasets (orbital parameters, different deep-sea oxygen isotope records), result in differences in the time scales used for LPS, and thus, no quantitative assessments of loess paleoclimatic proxy data are yet feasible. Nevertheless, the general similarity in the MS-pattern between Europe and China, especially in the structure and amplitude of the soil/pedocomplex record, is striking and provides a unique tool for direct comparison of paleoclimatic trends over the vast Eurasian continent (e.g. Forster et al., 1996; Marković et al., 2015; Song et al., 2017).

Over glacial-interglacial time scales, where loess units (L) and soil complexes (S) appear as stacked in the sedimentary profiles (see Figs. 1, 2, including the S/L classification for soils and loess), the MS records of loess series from the CCLP and south-eastern Europe show rather similar patterns and amplitudes for the considered time interval. However, prior to ~500 ka the MS record in the CCLP shows less amplitude, while in Europe the amplitude remains similar (e.g. Heslop et al., 2000; Marković et al., 2011, 2015; Necula et al., 2015; Song et al., 2017; Sun et al., 2006b). Moreover, it is suggested that the Carpathian Basin was under progressive continentalization over the Middle Pleistocene (gradual weakening of the Mediterranean influence through time; Buggle et al., 2013), which additionally complicates cross-continental correlations before ~450 ka.

Generally, the correlation and 'tuning' of geological datasets to astronomical reference curves can impose cyclic patterns (e.g. Huybers and Aharonson, 2010; Shackleton et al., 1995; Zeeden et al., 2015), and prevent an independent investigation of synchroneity and leads and lags in the environmental response (e.g. Blaauw, 2012). Reference datasets have varying age uncertainties, e.g. Lisiecki and Raymo (2005) suggest a 4 kyr uncertainty over the last million years for the marine $\delta^{18}O$ isotope stack series. Furthermore, selection of tie points is often to some extent arbitrary as patterns in sediments and correlation targets are not identical, leaving the possibility for inconsistency and error. A fundamental problem of LPS paleoclimate archive formation is the temporal development of soils onto previously deposited loess and the resulting overprinting of older deposits by soil formation, directly affecting the assumption of synchronicity of change with reference curves. Similarly, low sediment accumulation rates, syn- and post-depositional alterations of loess/paleosol (LPS) deposits (with bioturbation a major issue, also carbonate dissolution and reprecipitation) can cause signal smoothing (e.g. Stevens et al., 2006, 2011). Such issues limit the general accuracy of individual time scales established by correlation to reference datasets.

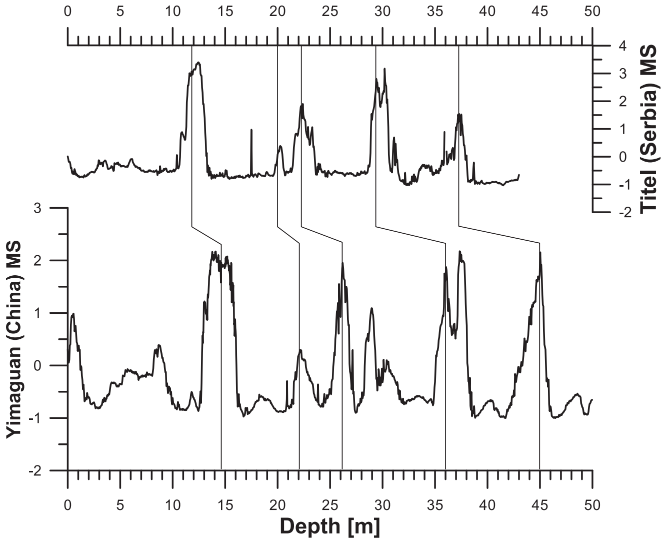

Fig. 1. Comparison of the standardized low-frequency low-field magnetic susceptibility (MS) from Yimaguan (China; here taken as representative for the Central Chinese Loess Plateau; Hao et al., 2012) and Mošorin at the Titel loess Plateau (Serbia; here taken as representative for the Carpathian Basin; Basarin et al., 2014) in stratigraphic depth.

Here, we aim at highlighting these issues, which prevent an accurate and correct comparison of different loess time scales as defined in selected LPS records from Eurasia. In spite of these issues, correlation including magnetic polarity stratigraphy are powerful methods and form an important part of integrated stratigraphy (e.g. Gradstein, 2012; Heslop et al., 2000; Hinnov and Hilgen, 2012). Furthermore, methods circumventing interpretations based on alignment have been established for astrochronology through testing against phase-randomized surrogates (Zeeden et al., 2015), and may be developed specifically for millennial-scale variability in the future.

The Carpathian (Middle Danube) Basin (CB) loess horizons show a clear baseline of low MS values (in the order of $2–3 * 10^{-4}$ SI; $\sim 2 * 10^{-7}$ m^3/kg) in several profiles for at least the last 450 ka (e.g. Basarin et al., 2014; Marković et al., 2012; Fig. 1). For LPS from the CCLP, the chronological time span represented by loess is shorter and the MS baseline is different between records (e.g. Hao et al., 2012; Sun et al., 2006b). This is not surprising when considering the spatial extent of the CLP and the range of the different climatic zones encompassed. This suggests that apparently the south-eastern European loess horizons have a MS baseline more similar to the drier north-western part of the CLP (see e.g. Sun et al., 2006a; Young Jeong et al., 2008), rather than the more humid CCLP. Differences in the MS baseline may be caused by higher sedimentation rates under stadial climates and shifts in dust sources (e.g. Varga et al., 2016), but still remain speculative. Nevertheless, in both regions dust deposition over interglacial/interstadial periods was continuous and provides characteristic fingerprints of MS-records allowing for continent-wide data comparison (e.g. Shi et al., 2003; Yang and Ding, 2014; Zeeden et al., in press).

Synchronicity of change may be expected between Asia and Europe at glacial-interglacial as well as millennial time scales because of the observation of similar millennial scale climate variability in proxy data

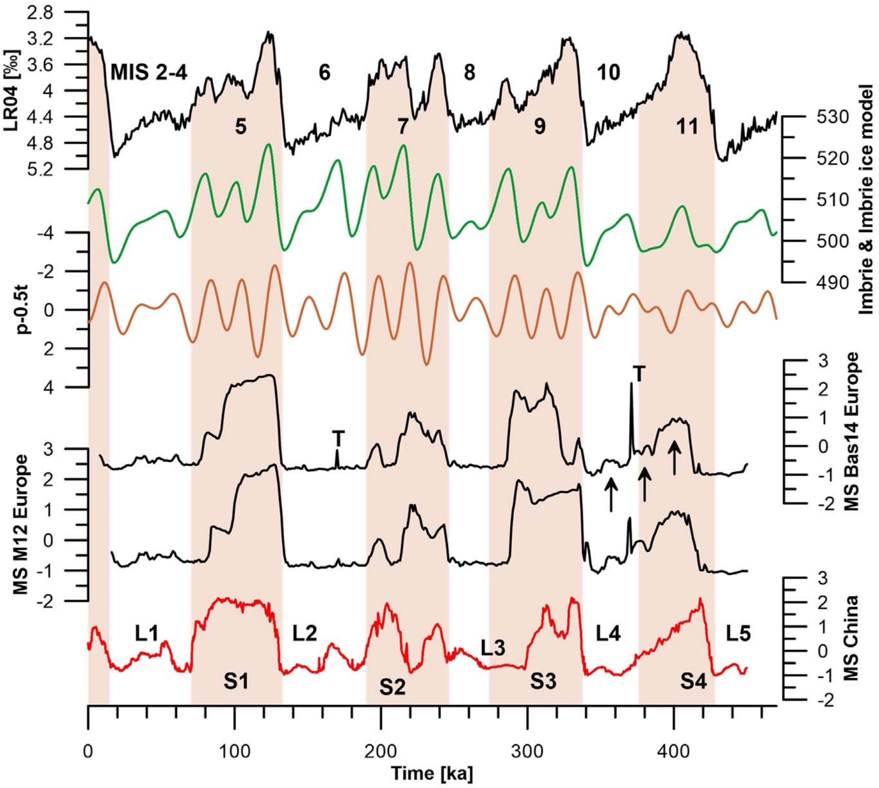

Fig. 2. Comparison of a marine benthic $\delta^{18}O$ isotope stack (Lisiecki and Raymo, 2005; top, black), an ice volume model (Imbrie and Imbrie, 1980; green), p − 0.5t (standardized precession minus 0.5 times standardized obliquity/tilt) as representing Northern Hemisphere insolation (Laskar et al., 2004; Lourens et al., 1996, orange), and the standardized south-eastern European magnetic susceptibility (MS) records from Titel by Basarin et al. (2014, black) and Marković et al. (2012, black), compared to a MS record from Yimaguan in China (Hao et al., 2012; red). 'T' denotes tephra layers in European Loess. Note the differences in patterns between European and Chinese MS records especially between 160 and 260 ka. Also, note the differences in timing of Terminations and onsets of glacial phases on the two parts of the Eurasian continent. Discrepancies between the marine stack, ice model and Loess MS data are apparent. Dark shading represents possible paleosol phases, but can only be indicative as time scale inconsistencies prevent clear statements on the timing. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

over the last ca. 40 kyr (both in frequency and timing; Hambach, 2010; Hambach et al., 2008; Obreht et al., 2017; Shi et al., 2003; Sümegi et al., 2012; Yang and Ding, 2014; Zeeden et al., in press). However, at decadal to centennial time scales, synchronous timing may not even be expected between Asian and European loess as reaction times of soil formation to forcing mechanisms and even forcing may be different (e.g. Marković et al., 2014). Here, we perform a comparison of MS records in selected sites from the CCLP (Hao et al., 2012; Sun et al., 2006b) and the CB (Basarin et al., 2014; Marković et al., 2012; Song et al., 2017). The CB was chosen because all high-resolution data from Europe spanning longer time periods come from this region, and although several LPS from the Black Sea region are reported (e.g. Jordanova et al., 2008; Necula et al., 2013, 2015; Gendler et al., 2006), and have magnetostratigraphic age control, their temporal resolution is in general not as high as resolution from the CLP and CB. These sites within the CCLP and CB lie at two extremities of the dry loess fields of the Eurasian continent. We therefore do not discuss the rather humid and periglacially overprinted LPS records from Europe along the 50° latitude closer to the ice margins (Antoine et al., 2013; Baumgart et al., 2013; Fischer et al., 2017; Hošek et al., 2015, 2017; Lehmkuhl et al., 2016; Taylor et al., 2014; Zens et al., 2017) and the strongly monsoon- and sea level-influenced records of coastal south-east Asia (e.g. Li et al., 1992; Shujian and Tao, 2011; Yingyong et al., 2008).

Grain-size records (from bulk sediment (GS) and quartz particles (QGS)), which form a common orbital-tuning target for loess besides the MS records (e.g. Heslop et al., 2000; Sun et al., 2006b), are unfortunately unavailable for European loess beyond the last glacial cycle (Újvári et al., 2016a) in sufficiently high resolution. Although GS data show a strong resemblance to QGS data from Chinese loess (Sun et al., 2006c), for still unclear reasons only a weak similarity between GS data (Necula et al., 2013; Obreht et al., 2016; Vandenberghe et al., 2014; Zeeden et al., 2016) and the QGS data (Újvári et al., 2016a) is observed for European records. The Sr/Ca data of microcodium, a proxy of past precipitation intensity (Li et al., 2017), show a dominant obliquity imprint in Asian loess not linked to glacial-interglacial cyclicity over the last 700 ka, but as yet such data are only available for the CLP.

Therefore, lithostratigraphy, biological climate proxies (e.g. Kukla, 1977) and mainly MS records (Basarin et al., 2014; Buggle et al., 2009; Marković et al., 2011; Necula et al., 2015; Song et al., 2017) have been used for establishing correlative time scales for European loess. These are supported by magnetostratigraphy and amino-acid-racemization stratigraphies in several places (e.g. Fink and Kukla, 1977; Marković et al., 2011, 2015). For the last glacial cycle ^{14}C and luminescence dating have also been employed in deriving age models (e.g. Bösken et al., 2017; Fitzsimmons et al., 2012; Lang et al., 2003; Schmidt et al., 2010; Song et al., 2015; Stevens et al., 2011; Újvári et al., 2014, 2016b), though the dating range (for ^{14}C) and precision (for luminescence) limits their applicability in some cases.

Comparison of the Titel and Mošorin (south-eastern Europe) MS records on the timescales compiled by Marković et al. (2012) and Basarin et al. (2014) reveals several intervals where different timing is suggested for almost the same stratigraphic unit (Fig. 2). Most obvious are the differences in timing of up to ca. 30 kyr for the S2 and S3 soil complexes related to marine isotope stages (MIS, see Fig. 2) 7 and 9, and a different pattern and timing in the order of 5 kyr for S1 and S4. Also, the existing time scales for LPS from the CLP are not always consistent (e.g. Song et al., 2007; and Sun et al., 2006b). For example, different temporal interpretations have been proposed for paleosols S1–S4, where the offsets in the onset or the end of S4, S3 and S1 differ by ca. 30 ka. Our aim is to highlight the as yet poorly understood common paleoclimatic features and differences between dust deposition and soil formation phases in Europe and Asia.

2. Comparison of loess records from Europe and Asia

Here, a two representative and high-resolution datasets are used for a simplified comparison (Fig. 1). In several European sections and datasets, a transition from Mediterranean type (rubified) soils to black-grey steppic soils around 400–300 ka is observed (Buggle et al., 2014; Marković et al., 2009). A progressive continentalization expressed in cooling and moisture decrease reflected in soil formation was postulated for the CB (Batajnica/Stari Slankamen) and the Lower Danube

Basin (Mircea Voda; Buggle et al., 2013, 2014). In the CLP, no such trend over the last 500 ka can be observed (e.g. Ding et al., 2002b). This trend in European aridization does not influence the onset and the transitions of glacial/interglacials, but rather determines the amplitude and pattern in the MS fluctuations.

The S4 paleosol MS pattern is similar between Europe and the CCLP, although the drop in MS towards the L4 unit has more variability and three local maxima in Europe (see arrows in Fig. 2). The exact timing of the S4 unit in both Europe and the CCLP is yet uncertain. The exact age of the distinct features of the S4 paleosol is challenging to determine, because neither MS nor QGS records show high similarity to marine records, ice volume models, and insolation (Fig. 2; e.g. Basarin et al., 2014; Hao et al., 2012; Zeeden et al., 2016).

The L4 unit shows low MS values in its upper (younger) part in the CCLP and Europe, which gradually increase towards the lower boundary of S4. This increase comprises two weak MS maxima in the CB and a tephra layer (Marković et al., 2015), while less distinctive patterns and a more gradual change can be observed in the CCLP.

The S3 pattern in the CCLP and Europe is clearly similar and shows two distinctive MS peaks separated by relatively low MS values (Fig. 2), in several loess records three maxima are present (Basarin et al., 2014; Sun et al., 2006b). Different timing of the onset and end of pedogenesis has been suggested for both the CCLP and Europe (compare e.g. Basarin et al., 2014; Hao et al., 2012; Marković et al., 2012; Sun et al., 2006b). A rather weak MS peak at the onset of the S3 is more prominent in European than in the CCLP loess (Basarin et al., 2014; Marković et al., 2012), and its relation to (orbital/ice) forcing is yet unclear at least for Europe. This horizon marks a clearly independent paleosol in Europe, although the interbedded thin loess unit might be overprinted by pedogenesis in most, but not in all, European sections. The European L3 loess shows a clear baseline of low MS values, whereas similar baseline values are only briefly reached in the CCLP at the onset of the L3 loess (Fig. 2). Furthermore, its temporal duration is not yet well constrained, with estimates ranging from e.g. ~20–25 kyr (Sun et al., 2006b, not shown) or ~40 kyr (Ding et al., 2002a, 2002b, not shown) in the CCLP, whereas it was suggested to have lasted ~40 ka in Europe (see Fig. 2; Basarin et al., 2014; Marković et al., 2012). The L3 unit and its timing is interpreted differently within Europe and Asia, and shows more variability in the CCLP than in Europe (Fig. 1, 2), an issue that might also reflect the limited amount of data available from Europe.

The S2 pedocomplex is clearly divided into two soil horizons in the CCLP with the younger soil showing the strongest magnetic enhancement (e.g. Hao et al., 2012; Heslop et al., 2000; Sun et al., 2006b). In Europe, the patterns are more heterogeneous (Zöller, 2010 and references therein), and in the CB and Lower Danube this doubling is not always well expressed (e.g. Buggle et al., 2009; Fitzsimmons et al., 2012; Necula et al., 2015). Based on low- to medium-resolution datasets (Buggle et al., 2009), the entire S2 pedocomplex in Europe was related to its counterpart in the Chinese loess stratigraphy. Other studies (Basarin et al., 2014; Marković et al., 2012) correlate the three pedogenic phases of the S2 (and early L2) to the penultimate interglacial (S2) and the older part of the L2 in the CLP. While the chronological framework of the base of the S2 is consistent, the timing of its upper boundary is less well established. In our opinion, the MS (and GS data from the CLP) allow for different interpretations regarding the timing and duration of the S2, and the timing of the MS enhancement interval in the CCLP and Europe. Recent data are in favour of the interpretation from the CLP (Li et al., 2017; Song et al., 2017). Additional independent dating will be necessary for an unambiguous solution. Also, automatic pattern matching and statistical evaluations (Kotov et al., 2016; Necula and Panaiotu, 2008) may help solve this standing stratigraphic issue.

The younger part of the L2 loess lacks variability in most of south-eastern Europe records, while the older part shows the presence of a weak soil. A weak magnetic enhancement and finer sediment grain sizes (see e.g. Hao et al., 2012; Sun et al., 2006b; Fig. 2) occur in the L2 in the CCLP. The weakly developed soil at ca. 160–180 ka (in the age model of Hao et al., 2012; Fig. 2) is strictly in accordance with MIS 6 (MIS 6d; Railsback et al., 2015). This gives difficulties comparing the MIS 7/MIS 6 boundary in the marine realm with the L and S (loess/soil) classification in terrestrial systems without ambiguity, especially where the MIS 7 and also early MIS 6 is amalgamated to one soil formation phase (e.g. Zeeden et al., 2016). Marković et al. (2015) deliberately placed a question-mark at the correlation between Luochuoan in China (Hao et al., 2012; Lu et al., 1999) and their record from Mošorin/Stari Slankamen in the (CB), highlighting the uncertainty in similarity in pattern and timing. In the L2, at least one yet undated tephra occurs in several south-east European records, traceable from the Mediterranean to the Black Sea shores (Marković et al., 2015), raising an opportunity for future timescale improvements, as was already achieved for L1 loess (Veres et al., 2013; Zeeden et al., in press; Obreht et al., 2017).

A rather uniform last interglacial soil complex (S1) is roughly correlated to MIS 5 in the CB. Whether this soil represents only MIS 5e, the warmest and oldest part of the MIS-related soil complex, and whether a correlation to MIS 5 or MIS 5e can be applied in low- and high-sedimentation rate areas is yet unclear (Chen et al., 1999). In China, a threefold division of the MIS 5 soil can readily be observed in the drier western CLP (e.g. Chen et al., 1999; Ding et al., 1998; Peng et al., 2014) and soils are related to MIS 5a, 5c, 5e (Porter, 2001; also see Fig. 2).

Fossil soils and MS enhancement occurs in the last glacial loess mainly related to MIS 3, and is more prominent in the CCLP than in the CB, where phases of pedogenesis phases are visually hardly noticeable (Figs. 1, 2; Marković et al., 2008). Furthermore, the timing independently from tuning and temporal pattern of European MIS 3 soils are especially elusive, this may also be because of dating shortcomings (e.g. Constantin et al., 2015) and spatial heterogeneity of such short phases of pedogenesis (Buggle et al., 2009; Obreht et al., 2017). However, in the Lower Danube loess, where independent age control is provided by the Campanian Ignimbrite tephra layer (ca. 40 ka) at least two peaks of magnetic enhancement are observed. These peaks represent the climatic optimum with MIS 3 interstadial soils separated by the CI tephra (Obreht et al., 2017).

3. Summary and conclusions

Patterns in paleoclimate proxy-data derived from loess-paleosol sequences appear rather similar between Europe and Asia at glacial-interglacial time scales. However, differences in the details have become apparent with a growing amount of higher resolution data from Europe. Most differences can be assigned to different time scales employed for defining the timing of the same patterns. However, also real differences exist. The origin of dissimilarities in patterns between these records is yet unclear, and in our opinion can only be assessed once remaining chronological issues reviewed here are clarified. The correlation between Europe and the CCLP is yet unresolved, and age control beyond correlation to marine geoarchives is required for more conclusive results. In this sense, developments in tephra stratigraphy in European loess hopefully aid in the development of precise loess chronologies and an improved paleoclimatic understanding (Fitzsimmons et al., 2013; Marković et al., 2015; Obreht et al., 2016, 2017; Veres et al., 2013; Zeeden et al., in press). Also, high-resolution palaeomagnetic investigations including the interpretation of relative palaeointensities (Fink and Kukla, 1977; Hambach et al., 2008; Jordanova et al., 2008; Marković et al., 2011; Rolf et al., 2014; Zeeden et al., 2009, 2011) may help in better constrained correlations. Further detailed paleoclimatic information from high-resolution records showing an unambiguous imprint of millennial scale climate variability (e.g. Chen et al., 1999; Obreht et al., 2017; Shi et al., 2003; Yang and Ding, 2014) can also lead to improvements in loess chronologies. It is crucial to understand the forcing mechanisms behind dust deposition and loess and soil formation (Li et al., 2017). No monsoon forcing is expected in central and south-eastern Europe, and pattern similarity between Europe and the CCLP must therefore have a different climatological origin. A common cause like the Siberian High atmospheric pressure system (Dodonov and Baiguzina, 1995; Obreht et al., 2016, 2017), or a synchronous north/southward migration of the Subtropical Jet as the boundary between the Hadley and Ferrell circulations, may play a role. Loess-paleosol sequences consistently show glacial-interglacial patterns, and also an insolation component. No full explanation for the clear dissimilarities to ice volume models (Fig. 2; e.g. Hao et al., 2012) and the marine records is available, but a combination of direct insolation forcing and northern hemisphere climate trends likely contributes. The unclear origin and timing of several patterns (e.g. S2, L2) limit a firmer correlation; high-resolution datasets showing similarities to models (Barker et al., 2011) or well-dated reference datasets (e.g. Kaboth et al., 2017; Martrat et al., 2007) have the potential for improvements. Differences in environmental records between the loess areas may arise from different responses of regional climate systems to the global climate and different contributions of local insolation, differences in dust availability, and local environmental conditions, including vegetation cover and soil formation. Here, we deliberately exclude several spatial and local heterogeneities from this analysis, e.g. differences between different parts of the CLP and the CB, and differences between western and central European loess deposits. This is done to focus on the remaining issues in Eurasian loess in general and beyond local differences which clearly add complexity.

Though a detailed comparison of loess-paleosol sequences over Eurasia is expected to improve our paleoclimatic understanding of this large continental area substantially, as yet chronological issues prevent such approaches. This is a major task for Eurasian and Northern Hemisphere palaeoclimatologists to be undertaken at glacial-interglacial and also millennial time scales. Application of (paleo)climate models, their comparison with more complex proxy data (e.g. QGS) and improved dating can be expected to shed more light on the forcing mechanisms in the future.

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