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Department of Geological Engineering, Mining and Sciences

# PALEOENVIRONMENTAL RECONSTRUCTIONS FROM CAVE SEDIMENTS OF THE MORAVIAN KARST, CZECH REPUBLIC

By

PAVEL SROUBEK

A DISSERTATION

Submitted in partial fulfillment of the requirements

for the degree of

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This dissertation, "PALEOENVIRONMENTAL RECONSTRUCTIONS FROM CAVE SEDIMENTS OF THE MORAVIAN KARST, CZECH REPUBLIC," is hereby approved in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY in the field of Geology.

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# Thesis Abstract

This thesis presents a paleoclimatic/paleoenvironmental study conducted on clastic cave sediments of the Moravian Karst, Czech Republic. The study is based on environmental magnetic techniques, yet a wide range of other scientific methods was used to obtain a clearer picture of the Quaternary climate. My thesis also presents an overview of the significance of cave deposits for paleoclimatic reconstructions, explains basic environmental magnetic techniques and offers background information on the study area – a famous karst region in Central Europe with a rich history.

In Kulna Cave magnetic susceptibility variations and in particular variations in pedogenic susceptibility yield a detailed record of the palaeoenvironmental conditions during the Last Glacial Stage. The Kulna long-term climatic trends agree with the deep-sea SPECMAP record, while the short-term oscillations correlate with rapid changes in the North Atlantic sea surface temperatures. Kulna Cave sediments reflect the intensity of pedogenesis controlled by short-term warmer events and precipitation over the mid-continent and provide a link between continental European climate and sea surface temperatures in the North Atlantic during the Last Glacial Stage. Given the number of independent climate proxies determined from the entrance facies of the cave and their high resolution, Kulna is an extremely important site for studying Late Pleistocene climate.

In the interior of Spiralka Cave, a five meter high section of fine grained sediments deposited during floods yields information on the climatic and environmental conditions of the last millenium. In the upper 1.5 meters of this profile, mineral magnetic and other non-magnetic data indicate that susceptibility variations are controlled by the concentration of magnetite and its magnetic grain size. Comparison of our susceptibility record to the instrumental record of winter temperature anomalies shows a remarkable correlation. This correlation is explained by coupling of the flooding events, cultivation of land and pedogenetic processes in the cave catchment area. A combination of mineral magnetic and geochemical proxies yields a detail picture of the rapidly evolving climate of the near past and tracks both natural and human induced environmental changes taking place in the broader region.

# Contents:

## Chapter 1

# Introduction: Cave Sediments and Environmental Magnetism

## 1.1. Character of Cave Sediments

Caves function as large sediment traps, accumulating samples of clastic, chemical and organic debris mobile in the natural environment during the life of a cave. Cave sediments are one of the most richly varied deposits that form in continental environments and tend to be preserved for long spans of time. They are divided into cave entrance and cave interior facies.

### *1.1.1. Cave entrance and cave interior environment*

The cave entrance is the part of the cave which is within the reach of daylight, but mostly not exposed to direct sunlight. The outside climate is damped to some degree. Vegetation is typically not present or is found in minimal amounts. Due to their protective nature cave entrances were often sites of human and mammalian habitation. The cave interior is the part of the cave which is not within the reach of daylight and is constantly in total darkness. No vegetation is found in the cave interior whatsoever and the cave interior was not a site of human or mammalian habitation, however in certain cases animals and perhaps humans got lost in the cave and their bones are found there. The cave interior contains vast deposits of fluvially deposited sediments and speleothems. Eventhough the cave sediments are well preserved from erosion, in certain cases erosion and weathering may occur (see further), however on a much smaller scale than on the surface. Since daylight reaching the cave is minimal or none, the process of pedogenesis (e.g. Bonifay, 1962 [6]) does not occur.

### *1.1.2. Genesis of cave entrance and cave interior filling*

The material found in the cave entrance either originates from weathering of the cave walls or is transported into the cave from outside. A typical example of communication between cave and surface is an entrance opening horizontally in the slope of a valley above the present day drainage level. At the mouth of the cave, a talus cone typically forms by material sliding down slope and falling into the cave entrance. The talus cone has the same composition as the slope sediments which consist mostly of limestone debris and loam. Since the sediments transported into the cave have to pass through the cone, its development has an effect on the infilling of the cave. If the cave is open, the sedimentation into the cave proceeds uninterrupted and the growth of the talus cone is compensated by transport into the cave or down the other slope of the talus cone. If the talus cone closes the cave, the only sediment brought into the cave is from the higher parts of the talus cone because no access to fresh material from the outside exists. In contrast, when the talus cone is eroded and lowered below the level of previous filling the sedimentation in the cave can stop altogether. An important factor in the genesis of the cave is the retreat of the entrance and the sedimentary content due to erosion. In such cases, the sediments preserved in the present day entrance may have been deposited quite deep in the cave.

The situation in the cave entrance is substantially simpler when clastic deposits are transported and deposited in the cave by fluvial processes, roof collapse and communication with surface through chimneys. The fluvial sediments range from boulders to clay depending on the energy of the medium.

Table 1-1. Rockshelter Processes (modified after Farrand, 1985[24]). Cryoturbation is deformation of sediments by frost action, solifluction is mobilization of frozen sediment by thawing.

| Agent | Deposition | Alteration |
|---|---|---|
| Cryoclastism | Debris and Frost Slabs | Split/Fissured Stones |
| Collapse | Large Blocks and Shattered Fragments | Crushed Debris |
| Solifluction | Hillslope Sludge | Displacement of Earlier Deposits |
| Cryoturbation | | Churning of strata, Rounding of Stones |
| Flowing Water | Flood deposits, Surface Runoff | Erosion, Gullying: Travertine |
| Wind | Well-sorted Sand and Silt | |
| Solution | Hydration Spalls; Grain-by-Grain Disaggregation; Flowstone | Leaching of $CaCO_3$, Rounding of Stones; Cementation |
| Pedogenesis | | Chemical/ Mineralogical Changes; Leaching, $CaCO_3$ Cementation; Root Disturbance |
| Humans and Animals | Artifacts, Garbage, Body Wastes, Imported Stones and Tools | Physical Disturbance; Digging, Burrowing; Chemical Alteration (decreased pH) |

### *1.1.3. Type of cave sediment, agents of deposition and alteration*

The types of sediments found in a cave entrance described as a function of the agent of deposition or alteration are summarized in Table 1-1.

Cave sediments are in general immature. The source and mode of transport and deposition of the sediments typically vary, resulting in a very poor sorting and high angularity of the sediments. Windblown sediments typically accumulate on the leeward side of the talus slope and are mixed with limestone debris. Limestone debris (Laville et al., 1980[45]) use the term eboulis), typically loosened by frost action from the cave walls is mostly fine and sharp-edged and is mixed up with sediments of foreign origin. The presence of the clay constituent loosened from decomposed limestones is indicated by tiny aggregates of idiomorphically limited quartz and/or pyrite. Layers of flowstones (secondary calcium carbonate precipitates) are indicators of cessation of transport of surficial material into the cave. Soliflucted material can interrupt the orderly build-up of deposits within a cave entrance or it can be evenly mixed with autochtonous deposits within the same sedimentary level. Soliflucted deposits can be detected because of their characteristic morphology displaying blunted edges and polished surfaces as a result of transport. Solifluction and cryoturbation can also alter the deposits by resorting and displacing them. These processes can also leave a characteristic relief, roundness and polish due to mechanical abrasion. The distinction between solifluction and runoff is often arbitrary and is based on better sorting of the grains. Sediments deposited by runoff are moderately to coarsely porous, the limestone detritus is coated by clay which penetrates even into the underlying layers. The space between individual particles of debris is filled with fine globules of clay or argillaceous loam. Stream deposits may be found in cave entrances which lie low enough to rivers to be flooded or which communicate with stream cave passages. The fluvial deposits typically range from gravel to silt.

#### *1.1.4. Cave sediments as recorders of Quaternary climatic cycles*

Cave sediments depend significantly on the climate as it impressed characteristic features on them when they were still part of the surface cover. Cave sediments allow us to study traces of climatic changes without regard of its further change due to the "conserving" effect of the protective entrance and interior cave environment. Of the fundamental importance of Quaternary stratigraphy are the following phenomena occurring in cave entrances: deposition of loess, intensity of mechanical weathering, intensity of chemical weathering (soil formation) occuring prior to deposition in the cave entrance and secondary accumulation of calcium carbonate.

Loess in cave is similar to surficial loess except that it contains varying and often significant amount of debris. Similarly to the accumulation of loess on the surface, accumulation of loess in the cave occurs during the colder periods. Deposition is often quite rapid causing the paleonthological content to decrease.

Mechanical weathering can be best observed on limestone detritus. In general, the debris found in a cave can be either *autochtonnous*, i.e. originating from within the cave itself and deriving exclusively from the parent rock in which the cave has formed or *allochtonnous*, i.e. having its origin outside of the cave and being transported inside. Angular debris forms during periods of intense mechanical weathering. More specifically, the composition of a frost-weathered debris deposit is a function of the parent rock, the moisture content, the rate, intensity and duration of the frost conditions, however other factors unique to each particular cave environment may have a strong effect on the nature of the given debris deposit. In general, the degree of frost weathering to be attributed to any cave sediment is at least roughly measurable in terms of its texture. Debris deposits formed during severe conditions when the freeze-thaw cycles were few but of high amplitudes tend to have coarser textures accompanied by relatively little matrix. In contrast, sediment consisting of finer debris is attributed to less severe conditions, when the freeze-thaw cycles were more frequent but of lesser amplitude.

Chemical weathering is most readily apparent on limestone. Originally angular blocks become blunted and their porosity increases. The dissolved carbonate often migrates downward where it precipitates in the form of calcareous concretions. Some of the strongly corroded sediments totally lack fine-grained limestone, however coarser fragments may still be present. The insoluble residue manifests itself in the clay fraction of soils. Severely weathered feldspars require intensive chemical weathering which typically did not occur in the warm periods of the Younger Pleistocene in Central Europe. The amount of organic matter and clay content in the sediments is also a close indicator of soil formation and thus humidity and warmth of climate. Humous soils are often brought into the cave by rain wash during the warm period or by solifluction during episodic thawing which is characteristic of unstable but still cold conditions at the initiations and terminations of major cold phases. Soils forming on strata of quartz and silicates do not change significantly in grain size, however soils forming on fine limestone detritus increase its fine grain size content during the process of weathering. The presence of the clay constituent loosened from decomposed limestones is indicated by tiny, ramified aggregates of idiomorphically limited quartz and/or pyrite.

Secondary carbonate can form in situ in the lower soil horizons, but it can be also brought into the cave from the surface. Sinter detritus loosened from the walls and ceiling can be also a substantial part of the sediments.

### *1.1.5. Review of literature on entrance facies cave sediments*

The principles of cave sedimentology were first stated by Lais (1941)[43]. He based the paleoclimatic reconstruction on the assumption that freezing and thawing of water in pore spaces of entrance cave sediments was the most important factor in rock disaggregation under cold climate. His work was extended in the postwar years by several European scientists (e.g. Bonifay, 1962[6]; Kukla and Lozek, 1958[40]).

Among the more recent studies the most extensive is that by Laville et al. (1985)[46] of the Perigord region of southwestern France. More than 200 rockshelters were excavated in this karst area revealing an enormous wealth of archeological findings from 200 Ky B.P. till the Late Paleolithic. A truly comprehensive stratigraphic approach was utilized which included the sedimentary fill of the cave entrances as well as the archeological findings. Using major unconformities as indicators of interstadials and combining cave sedimentological information with that derived from palynological and paleontological investigations, the authors created a high resolution and complex scheme of climatic history. By means of this scheme shelter sites could be correlated from one to another and an unprecedented regional sequence was created. The work also resolved some of the basic problems of interpreting Paleolithic and cast considerable light upon the general environmental background of the culture history.

Farrand (1979)[23] conducted an extensive study of Abri Pataud in southwestern France where approximately 8m of sediments accumulated during the Late Wurm Glacial. The methods used in this study were measurement of grainsize distribution, roundness, porosity and carbonate content of bulk samples as well as determination of the clay and heavy mineral content of the matrix. Using principally carbonate content and clay mineralogy Farrand distinguished several weathered horizons which formed during moderate and humid climates. In these horizons the carbonate content and the amount of montmorillionite decreased whereas the amount of illite increased. The roundness of clasts as well as their porosity increased as well. The sediments in the cave originated from break-up of the surrounding bedrock or were transported into the cave either by wind or by human inhabitants. Based on the heavy minerals present, the source of the windblown sediments was determined to be the floodplain of the Vezere River. Farrand suggests that cave studies deal with a very specialized niche, where each site must be dealt with as a specialized entity, and where the sedimentation is controlled by numeroues variables. The same sorts of variations found in sites under different climatic regimes may not be interpereted in the same fashion. Only a generalization drawn from the application of a series of techniques to a great number of sites can lead to a valid regional conclusion

In Cantabrian Spain, Butzer (1981)[9] analyzed the sedimentary fill of five caves. The author distinguished external sedimentary sources from the autochtonnous cave sediments by comparing the cave sediments with the surficial deposits. The paleoenvironmental interpretation of the key sequences led to a regional climatostratigraphy that included forty enviromental phases which correlated with the oxygen isotope phases 2-5.

A useful discussion on rock shelter and cave sediments was published by Farrand (1985)[24]. He described the uniqueness of the cave environment and presented the possible modes of deposition and alteration of cave sediments in particular providing altenatives to the frost weathering theory of formation of limestone. On a more critical note he discussed the pitfalls of paleoenvironmental interpretations of grainsize data and of variations in clay mineralogy.

## 1.2. Environmental magnetism and its application to karstic studies

Magnetic properties of iron oxides found in naturally occurring sediments do reflect changes in environmental processes operating at the Earth's surface (e.g., Thompson and Oldfield, 1986[68]; King and Channell, 1991[38]). The extent to which magnetic properties are maintaining their character or are subject to change depends on the nature of the processes to which the magnetic minerals have been subjected. Chemical alterations are most importantly caused by weathering, soil formation and sediment diagenesis. These processes lead to transformation between magnetic mineral types, conversion of paramagnetic iron to ferri or antiferromagnetic minerals or conversion from ferrimagnetic minerals to paramagnetic forms. The alteration processes can also lead to dilution or concentration of magnetic minerals depending on the stability of the surrounding non magnetic minerals and on their authigenic production. The second factor strongly influencing the make up of the sediment are the methods of transport which can lead to an overall decrease in grain size and a change in particle shape and grain-size sorting. Thus the final sediment can differ significantly from the source rock. Lastly, during deposition magnetic minerals can be diluted to various degrees depending on the precipitation of authigenic minerals. The change in the energy of the sedimentary environment can lead to grain size sorting.

### *1.2.1. Mineral magnetic parameters*

In order to characterize the magnetic mineralogy, the concentration of magnetic minerals and their grain size researchers conduct a wide range of magnetic measurements. The commonly measured parameters and interparamteric ratios discussed below and summarized in Table 1-2.

*Magnetic susceptibility (*$\chi$, MS*)*

Magnetic susceptibility is a measure of the ease with which a material can be magnetized. It is a defined as a ratio between magnetization induced in a material and the field which induces the magnetization. Diamagnetic minerals (e.g. carbonates) have values of $\chi$ negative, paramagnetic minerals (e.g. most silicates) have values of $\chi$ positive. In environmental magnetic studies we deal mostly with ferromagnetic minerals (magnetite, maghemite, hematite), where $\chi$ is not a scalar and does not show a linear behaviour between the inducing fields and the induced magnetization. Valuse of $\chi$ for ferromagnetic minerals is much larger than one. Most commonly used is the mass specific susceptibility measured in $m^3$/kg. Susceptibility is a function of magnetic mineralogy, concentration of magnetic minerals as well as grainsize. Among the common minerals present in natural materials are magnetite type ("magnetite") and maghemite ("maghemite") type minerals which have values of $\chi$ several orders of magnitude larger than the rest of the minerals and thus $\chi$ can be used as a quick measure of the concentration of these minerals. The dependance of $\chi$ on grainsize for magnetite has a complex behavior and was extensively studied (e.g. Maher, 1988[48]).

*Frequency dependence of magnetic susceptibility (FDMS,* $\chi_{fd}$*)*

Susceptibility is typically measured at low frequencies when the magnetization remains in phase with the field. However, when the frequency is increased, the relaxation effects become more important leading to a decrease in the in-phase component of magnetic susceptibility. The dominant cause of drop in magnetic susceptibility with increasing frequency is the presence of ferrimagnetic grains lying close to the single domain (SD)/superparamagnetic (SP) boundary. By measuring $\chi$ with changing frequency, researchers one can obtain a grain size distribution of particles near this size boundary. Typically, however, MS is measured only at two frequencies and the factor

$$FDMS = \frac{MS_{low\_freq} - MS_{high\_freq}}{MS_{low\_freq}} * 100[\%]$$

is used as an estimate of the concentration of SP grains.

*Temperature dependance of $\chi$*

The behavior of $\chi$ with changing temperature is typically used for identifying magnetic minerals present in samples. The identification is complicated by the fact that $\chi$ variations with temperature are a function of grain size, internal stress and crystalline anisotropy. At increased temperature, new magnetic phases may form further complicating the interpretation of the $\chi$ vs. temperature curves.

*Anisotropy of $\chi$ (AMS)*

The relative ease with which a material can be magnetized in various directions is expressed by its AMS. The anisotropic behaviour of $\chi$ may have different causes, the shape anisotropy being of major importance for environmental magnetic purposes. Depending on the sedimentary environment, irregular magnetic particles tend to align themselves in various orientations, thus causing the bulk sample to show anisotropic behavior of $\chi$ . Typical for a low energy sedimentary environment is a primary depositional fabric where the foliation plane is horizontal and the lineation does not have preferred orientation. Currents acting on the sedimenting particles tend to align the particles' long axes either in the direction of the flow or in a direction perpendicular to the flow. Also the short axis is tilted from the vertical direction and is dipping in the direction of flow. AMS is also useful for identifying postdepositional motions in the sediment such as slumping or cryoturbation.

*Anhysteretic remanent magnetization (ARM)*

Anhysteretic remanent magnetization is imparted to samples in a laboratory by subjecting the sample to a strong alternating field which is smoothly decreased to zero in the presence of a small steady field. A linear change in remanence with field strength is found for steady fields with magnitudes similar to the Earth's field. This linear rate of change is referred to as the susceptibility of ARM ($\chi_{ARM}$). Susceptibility of ARM for single domain magnetite is 800x$10^{-6}$ $m^3$/kg which is approximately an order of magnitude more than that for multidomain magnetite and two orders of magnitude more than for hematite.

*Isothermal remanent magnetization (IRM) and Saturation IRM (SIRM)*

Isothermal remanent magnetization is the remanent magnetization acquired by deliberate exposure of a material to a steady field at a given temperature. The intensity of IRM depends on the field until saturation is reached. The maximum remanence acquired is denoted SIRM. The laboratory generated fields are often not sufficient to reach true SIRM of the material (e.g. hematite) and thus the highest fields (typically around 1T) are used. SIRM values for magnetite rapidly drop with increasing grainsize from single domain through pseudosingle domain to multidomain grains. The shape of the IRM acquisition curve is often used to distinguish between low and high coercivity magnetic minerals.

*Saturation magnetization ($M_s$)*

Saturation magnetization is the highest magnetization possible induced in a presence of a large laboratory field. The intensity of $M_s$ is solely a function of mineralogy and grain size. The value of $M_s$ for pure magnetite is 480 kA/m whereas for hematite it is only 2.5 kA/m, $M_s$ is thus a good indicator of magnetite concentration.

*Temperature dependance of $M_s$*

The behavior of $M_s$ during heating is a fundamental magnetic property used to identify magnetic minerals present. It is independent of such variables as crystal size, shape and internal stress. The Curie temperature is indicated when the $M_s$ becomes zero and is characteristic of the magnetic mineral present. The shape of the $M_s$ vs. temperature curve can also indicat phase changes during heating.

Table 1-2. Magnetic parameters and interparametric ratios characterizing composition, concentration and grain size of ferromagnetic minerals typically used for detecting environmental change

| Parameter | Used to determine | Comments |
|---|---|---|
| Magnetic susceptibility ($MS_{fe}$) | Concentration of "magnetite" | grain size dependant |
| Susceptibility to anhysteretic remanent magnetization ($MS_{ARM}$) | Concentration of "magnetite" | grain size dependant |
| Saturation isothermal remanent magnetization (SIRM) | Concentration of "magnetite" | grain size dependant |
| Saturation magnetization ($M_s$) | Concentration of "magnetite" | |
| $MS_{ARM}$/MS | grain-size of "magnetite" | sensitive to finer grain sizes, SP and MD grains=similar results |
| $MS_{ARM}$/SIRM | grain-size of "magnetite" | SP and MD grains=similar results |
| $MS_{ARM}$/SIRM vs. FDMS | grain-size of "magnetite" | |
| S ratio $S=-IRM_{-300mT}/IRM_{900mT}$ | ratio between "magnetite" and "hematite" concentration | |
| Hard IRM $HIRM=(-IRM_{-300mT}+IRM_{900mT})/2$ | Concentration of "hematite" | |
| Frequency dependance of MS $FDMS=(MS_{LF}-MS_{HF})/MS_{LF}$ | Concentration of SP magnetite grains | does not detect the full range of SP grains |
| High field MS | Concentration of paramagnetic minerals | |

### *1.2.2. Examples of environmental magnetic studies*

The environments most typically subject to magnetic studies are terrestrial loess deposits and lake and deep sea sediments. Studies of loess deposits, especially from north-central China, have conclusively shown that variations in magnetic susceptibility from the loess and the interbedded paleosols sequences correlate with climatically controlled variations in the marine oxygen isotope record (Heller and Liu, 1982[27]; Kukla et al., 1988[42]; Kukla and An, 1989[41]; Beget et al., 1990[5]; Beget and Hawkins, 1989[4]). Wang et al. (1990)[74] have also shown the presence of Milankovitch cycles in the magnetic susceptibility records from these loess deposits. These studies conclude that variations in magnetic susceptibility indicate climatically controlled processes and provide a means of precisely correlating continental and oceanic climatic change records for the Quaternary (Banerjee, 1994[2]). Zhou et al. (1990)[76] and Maher and Thompson (1995)[51] observed that accumulation rates of magnetic minerals and magnetic parameters sensitive to grain size consistently vary between paleosol and loess layers. The high values of $MS_{ARM}$/SIRM and FDMS in paleosols could only be caused by magnetite grains near the supraparamagnetic(SP)/single domain (SD) size boundary (Maher, 1988[48]). These observations ruled out the possibility that magnetic enhancement is caused only by varying dilution of one type of magnetic minerals with non-magnetic components as suggested by Heller and Liu (1982)[27] and Kukla et al. (1988)[42]. A predominance of evidence now suggest that variations in $\chi$ in the loess/paleosol sections are due to pedogenic enhancement (creation of fine-grained authigenic magnetic minerals) during warmer interglacials and interstadials (Hus and Han, 1992[32]; Maher, 1991[49]; Maher and Thompson, 1995[51]; Zheng et al., 1991[75]; Heller et al., 1991[29]). In fact, FDMS is extremely sensitive to changes in weathering conditions and can indicate mineralogical changes which were not recognized in previous pedological descriptions. In addition to the signal attributed to fine-grained authigenic magnetic minerals, both loess and paleosol carry a background signal attributed to a roughly constant detrital input (Heller et al., 1991[28]; Verosub et al., 1993[72]). Hus and Han (1992)[32] observed that thermal demagnetization of the natural remanent magnetization (NRM), SIRM and ARM showed similar results for loess as for paleosols and concluded that the latter inherited a significant amount of magnetic minerals from the loess. On the other hand, Verosub et al. (1993)[72] conclude that pedogenesis is the dominant source of magnetic minerals not only in paleosols but also in loess. They reached this surprising conclusion after treating the samples with citrate-bicarbonate-dithionite which extracts secondary ferrimagnetic grains. The magnitude of $\chi$ of soils and loess decreased by 87% and 59%, respectively and reached the same values.

The studies in lake and deep sea sediments focus primarily on detecting climatic change, changing source of sediments and, in the case of the youngest sediments, deciphering the human impact.

For example, the interglacial sediments of Lake Baikal (Peck et al., 1994[62]) are characterized by low concentrations (low $\chi$ , $MS_{ARM}$, SIRM and $M_S$) of low coercivity magnetic minerals (high S ratio and low HIRM) whereas sediments from the Late Pleistocene glacial stage have higher concentrations of magnetic minerals with an increased proportion having higher coercivities (low S ratio, high HIRM). This behavior corresponds well to a conceptual model of increased biogenic sedimentation during the warmer periods and increased eolian sedimentation during the arid colder periods. The high degree of coherence between the correlation of the $\chi$ and HIRM and the SPECMAP delta $^{18}O$ curve (Imbrie et al., 1984[34]) for the last 250,000 years suggest that the magnetic properties of the Lake Baikal sediments are tracking global climate change and provide an age estimate beyond the radiocarbon dates.

Banerjee et al. (1981)[3] were able to delineate three different zones of magnetic particle size by measuring the $\chi_{ARM}/\chi$ ratio on cores from Long Lake, Minnesota. These zones correlated with changes in the pollen assemblage caused by climatic deterioration during the Little Ice Age ("Big-Woods") and subsequent large scale cultivation (Post "Ambrosia Rise"). The change in grain size was interpreted to reflect changes in erosional flux.

Oldfield et al. (1979)[58], in their study of magnetic properties in the catchment of Jackmoor Brook, England showed that the ratio of isothermal remanent magnetization (IRM) versus SIRM is different for bedrock, cultivated soil and woodland soil. Measurements of IRM/SIRM on the suspended load of Jackmoor Brook indicated that it originates mostly from the cultivated soil.

Thompson et al. (1975)[69] observed that downcore magnetic susceptibility variations in sediments of Lough Neagh (N. Ireland) correlate positively with chemical and pollen indicators of forest clearance, soil disturbance, extension of farming and accelerated supply of inorganic material in to the lake. Dearing and

Flower (1982)[16] measured $\chi$ in recently accumulating sediments in the same lake and found correlation with rainfall which further supports the link between erosion and concentration of magnetic minerals. Basaltic soils are the major source of titanomagnetites in the Lough Neagh sediments as $\chi$ of suspended load is higher in rivers flowing on basaltic bedrock. Similar measurements were accomplished by Thompson and Morton (1979)[67] on the sediments of Loch Lomond, Scotland who observed correlation between $\chi$ of the sediment and the amount of the fine grained fraction. Variations of $\chi$ were attributed to land use changes leading to differential erosion of a single source - primary magnetite from the bedrock which passed through a soil forming stage. This model explains $\chi$ variations within a stream entering the lake, between streams, within a lake core and between near shore and deep water cores in Loch Lomond.

Ferromagnetic minerals present in sediments are typically controlled by the character of the source region, mode and energy of the transporting medium, and by depositional as well as by post-depositional processes. For example, Walker and Lowe (1990)[73] showed that high concentrations of magnetic minerals together with high quantities of Na, K, Ca and Mg at the base of a sedimentary profile on the Isle of Skye reflected inwash of unweathered, finely comminuted, clastic material from freshly exposed glaciogennic sediments. The subsequent increase in the carbon content indicated gradual stabilization of slopes around the catchment and maturation of soils as the vegetation cover expanded in response to climatic amelioration. In contrast, Oldfield et al. (1985)[59] suggested, based on mineral magnetic analysis of the fine grained fraction of sediments transported from the Rhode River catchment in

Maryland, USA, that sediment source variations rather than climatic or postdepositional processes control ferromagnetic particle size. Dearing et al. (1986)[17] in a study of the Seeswood Pool catchment in Warwickshire, England showed that sediment transport can be traced if various size fractions are analyzed. Studies by Curtis (1987)[14] showed that fluctuations in the type of magnetic minerals, their concentration and grain size in anoxic conditions of lake sediments provide information on the rate of degradation of organic matter through microbial sulphate reduction, detrital iron oxide reduction and methanogenesis. A follow-up study by Jelinowska et al. (1997)[36] on sediments from Polish Bledowo Lake showed that authigenic pyrite appeared only in organic rich layers with low Ca/Mg ratios. The presence of vivianite together with pyrite indicated that bacterial degradation of organic matter occurred first in the sulphate-reducing zone and only later in the methanogenesis zone. In addition, the presence of large magnetic particles of uniform grain size suggested bacterial reduction of detrital iron oxides during both sulphate reduction and methanogenesis.

Magnetic studies of sediments collected from the highly eutrophic Slapton Ley (Foster et al., 1998[25]) also suggested dissolution of ferrimagnetic and canted antiferromagnetic minerals under anoxic conditions. Results indicated that magnetite production was very limited even though Slapton Ley is extremely productive and trends toward hyper-eutrophic conditions.

An increasing number of studies report the occurrence of greigite in lacustrine sediments and suggest that the mineral can remain stable in sediments for a considerable period if sulphate reduction is arrested (e.g. Snowball and Thompson, 1990[64]; Hilton, 1990[30]). However, the precursor for greigite formation appears to be an oscillating oxycline controlled by seasonal, eustatic and isostatic effects or by hydrocarbon seepeage. Greigite formation is often associated with magnetite dissolution in response to anthropogenically induced eutrophication linked to bacterial reduction of sulphates and decomposition of organic matter. The recognition of magnetite dissolution in lake sediments provides evidence that severe reducing conditions once prevailed (e.g. Geiss et al., 2004[26]).

### *1.2.3. Environmental magnetic studies in cave deposits*

One sedimentary environment which has not been exploited adequately by environmental magnetists is the cave environment. Papers which are concerned at least marginally with solving environmental issues in caves are discussed below.

Ellwood (1971)[18] studied a sequence of fine silts and clays in the interior of Climax cave, Georgia. He determined the age and sedimentation rate by comparing the curves of declination and inclination variations with an independent Arizona archeomagnetic curve. The comparison is possible because the Paleosecular variation (PSV) features appear to be persistent up to a distance of 3000km. Ellwood concluded that the sequence was deposited between 700 and 1500 A.D. and that the sedimentation rate was 6.9 cm/1000 years. He concluded that variations in sedimentation rate are caused by climatic change.

The PSV and NRM intensity records were obtained from clastic sediments in three caves in the Mediterrannean region by Creer and Kopper (1976)[13]. These authors used paleoclimatic evidence for dating and determining the sedimentation rate. They compared their PSV records from caves to established PSV records from other regions. A sedimentary section from Jeita cave, covering the time interval from 14,000 years till present was paleomagnetically investigated. Four complete oscillations of the inclination records were observed and were correlated with the Black Sea inclination record. Steeply dipping beds of fluvial sediments from the entrance of Hermits cave, Spain were investigated. Strong deviations of the paleomagnetic directions from the normal field and very large amplitudes in the PSV record were observed. As the paleomagnetic declination was not biassed in the paleowaterflow direction and redeposition would result in a higher directional scatter the anomalies are interpreted as evidence of a geomagnetic excursion. From the supposed 3000 year duration of geomagnetic excursions and a high rate of deposition is determined (50cm/1000 years).

A layer of archeological artifacts related to an Upper Paleolithic cave painting in the cave of Tito Bustillo was dated between 11200 and 11600 B.P. by comparing the magnetic declination, inclination and intensity curves from this site with reference data from Lake Windermere, England. With the aid of SEM it was determined that the sediments were originally aeolian and were transfered in the cave by water action. From the fact that no sediments were deposited on the section studied and from the evidence of a lengthy period overgrowth under the aegis of a long term pH change, the authors concluded that the sands were undisturbed and could preserve the orientation of the magnetic particles. As deep in the cave there is minimum bacterial action and thus limited diagenesis, it is thought that the orientation of the magnetic particles is from the time of deposition and thus the NRM reflects the secular field from that time.

Noel (1986)[54] described a study of the magnetic fabric from Peak Cavern in Derbyshire, England. Measurements of AMS indicated that most of the clays showed primary depositional fabric and were stable recorders of the paleomagnetic field. The only exception was a 30 cm thick layer which showed scattered directions of the principal axes of susceptibility suggesting that the original magnetic fabric has been overprinted by the growth of secondary authigenic magnetic minerals. Further, a method for estimating water current directions based on comparing the direction of the magnetic remanence vector and the direction of maximum $\chi$ was modified. This modified technique allowed Noel to detemine that stream paleoflow had an opposite sense than today.

In Pwll y Gwynt Cavern in Wales, Noel (1986)[55] used AMS to distinguish whether deposition was controlled by hydrodynamic forces or whether it originated from a dispersion of suspended particles undergoing viscous shear. Sediments in the cave were deposited on variously sloping surfaces. Measurements of directions of magnetic lineation allowed Noel to conclude that the transition from a strike oriented to a dip oriented magnetic lineation probably occurred on a slope of about 20 degress.

Ellwood et al. (1996)[20] produced a paleoclimatic record for the time interval between 9000 and 3500 years B.P. from magnetic susceptibility measurements on clastic cave sediments of the interior facies. The two profiles were located in an archeological excavation pit in Konispol Cave, Albania. Susceptibility highs correlated well with periods of warm and humid climate as defined from other European regions. Variations in $\chi$ were attributed to the content of pedogenetically controlled maghemite which formed outside of the cave and was then washed or blown into the cave.

Ellwood (1999)[20] further expanded on his work in Albania by studying several other caves in the Mediterranean region. He correlated the $\chi$ record from one cave to another as well as with the climatic record from the Greenland ice core. His new work stresses the usefulness of the cave environment for studying past climal change.

## 1.3. The Study Area: Moravian Karst

Moravian Karst is a classical karst region with karstic features which are the best developed in the entire Czech Republic. It contains approximately one thousand registered caves, the Amatereurs Cave system being one of the most extensive in Central Europe. The acessibility, beauty of the karstic landscape and the unique scientific findings led to a several centuries long history of excavations and scientific studies of the caves and surficial kartic features. The Moravian Karst contains caves with hundreds of meters of well accessible profiles in clastic deposits, both of the interior and entrance facies and thus offer a suitable location for paleoclimatic studies.

### *1.3.1. Geograpical Outline of the Moravian Karst*

Moravian Karst is a small karst region with an area of 85 $km^2$. It forms a narrow stripe 25 km long and roughly 3-6 km long north of Brno in the southeastern Czech Republic. The surface of the Moravian karst is characterized by plateaus with sinkholes. The plateaus are cut by several canyon shaped valleys. The average elevation is 448 m a.s.l. In the north-western part of the Moravian Karst the Sloup Semi-blind Valley has developed. The Sloup Creek flows through the valley and then sinks into the Sloupske Caverns. In the north-eastern part of the Moravian Karst, the Holstejn Semi-blind Valley exists in which the Bila Voda Creek flows. Both creeks flow underground in the Amateurs Cave, where they join and eventually resurface as the Punkva River. Punkva River flows out of the karstic region via the Punkva Valley. For a general overview of the topography of the Northern part of the Moravian Karst and the locations of the studied caves see the following chapters and in particular Fig. 3-1.

### *1.3.2. Geological Outline of the Moravian Karst*

Most of the Moravian Karst is formed by Devonian and Lower Carboniferous limestones (Musil, 1993)[53] (see Fig. 3-1). The basement of these Paleozoic sediments is the granitoid Brno Massive of Precambrian age, the surface of the granitoid is weathered into a sandy eluvium and is overlain by purple reddish clastic terrestrial sediments. The Lower Devonian lacustrine and lagunar limestones filled depressions on this weathered surface and their present day thickness is more than 1700 meters. Middle Devonian basal clastic sediments, which were transported from the south, reach a maximum thickness of 400 meters. In the Givetian, large scale carbonate sedimention began. These carbonate sediments reach a present day thickness of more than 2000 meters. Different types of limestones belonging to the reef facies form this so called Macosska Formation. The limestones vary from thickly bedded to laminated and are distinguished into separate units based on the dominant fossils. The youngest carbonates of this sedimentary unit belong to the Lower Carboniferous.

The folding of the carbonates occurred during the Sudetian Phase of the Variscan Orogeny. Open and often overturned folds occur together with low angle reversed faults along which the underlying basement is occasionally overthrusted. Overlying the carbonates are grey green, slightly silicified shales. A rapid marine transgression occurred towards the end of the Lower Carboniferous and the area of the Moravina Karst was covered by black grey silty shales and siltstones and fine-grained greywackes. These sediments are preserved in a 1000 meter thickness to the east of the Moravian Karst. During the period between the Upper Carboniferous and Jurassic no sedimentation occurred in the area.

In the Middle Jurassic, a sea once again trangressed over the present day Moravian Karst and white grey micritic and biosparitic limestone was deposited. The sea retreated in the Late Jurassic and intensive weathering in subtropical conditions took place; the Jurassic sediments were mostly eroded and terrestrial sedimentation took place. During this period, another episode of karstification occurred. The conditions again changed at the Lower-Upper Cretaceous boundary when the area was covered by an epicontinental sea in which thick layers of sands were deposited.

The present day character of the Moravian Karst - a fluviokarst - began developing in the Early Paleogene (Hypr, 1981[33]; Panos, 1963[61]; Bosak et al., 1989[8]) by the appearance of valley semi-poljes (semi-interior valleys). During the Paleocene, Eocene and mostly Oligocene as well as in the Lower Miocene tectonic block movement lowered the base level and caused dissection of the karst pediment by valleys. The deepening of valleys occurred in several stages and was associated with formation of differing cave levels. Simultaneous entrenchments of blind valleys dissected the valley semi-poljes. In the Middle Miocene (Badenian), a marine transgression covered a large portion of present day Czech Republic including the Moravian Karst. Caves were probably completely filled and the pre-Badenian relief was fossilized (Adamek, 2005[1]) Due to increased tectonic activity in the Pliocene, base level continuously fell (Bosak and Horacek, 1982[7]). The pre-Badenian river-system (valley network) was partially restored, erosion of the marine sediments took place in several phases and probably continued into the Quaternary (Musil, 1993[53]). During this time period the karstic waters re-established old underground drainage routes (Bosak and Horacek, 1982[7]).

Panos (1963)[61] on the other hand believes that new Plio/Pleistocene (inserted) cave levels developed during removal of the Badenian sediments. Alternatively, the majority of caves could have developed during the Plio/Pleistocene and only terrestrial sediments were ever deposited in caves (Pribyl, 1988[63]; Musil 1993[53]). In either case, the (marine) sediments in caves were replaced by several cycles of terrestrial deposition and erosion during the Quaternary. The sedimentary cycles were controlled by the relative movement of the base level in part due to a periodically changing climate (Musil, 1993[53]).

The age of the sediments in the caves is not well known. Entrance facies sediments were dated by paleontological methods to be from Holocene or the last glacial period (Horacek and Lozek, 1988[31]). Exceptionally continuous deposition occurred since the end of the Riss glacial period as indicated by the results from Kulna Cave (Valoch, 1988[71]; this work). Older sediments are probably also present in the caves and their age may reach as far as the early Pleistocene (Cilek, pers. comm.). Musil (1993)[53] proposes that during the several transitions between cold and warm periods the rivers of the Moravian Karst were overloaded by sediments. This caused accumulation in valleys leading to reactivation of higher level caves and their infilling. As erosion probably occurred in the warmer periods and the process repeated, juxtaposed sediments are expected to be of different age.

### *1.3.3. History and Climatic Conditions of the Bílá Voda Catchment*

The catchment of the Bílá Voda River is situated in a hilly terrain with elevations between 400 and 700 m a.s.l. The area was originally covered by dense forests dominated by beech and fir, including linden, pine, elm and oak (Nožička, 1957[57]). The first settlers who had any significant impact on the forest cover began arriving in the 14th century. The common technique used to clear forests in order to create glades for villages, pastures and fields was to first slash and then burn the trees. By the end of the 16$^{th}$ century, up to fifteen small settlements were found in the catchment of the Bílá Voda River (Černý, 1982[11]). The immediate vicinity of the Bílá Voda River was left untouched except for a 3 km stretch between the villages of Otinoves and Rozstání to the east of Holštejn, which was converted into fields. During the period of unrest and wars that ravaged the 17$^{th}$ century Europe, inhabitants of small villages were constantly facing severe danger. Most of the small settlements in the catchment of the Bílá Voda, distant from the relative safety of larger towns, were thus abandoned and were soon overgrown by forests (Černý, 1970[10]). Towards the end of the 17$^{th}$ century, the catchment area became part of the property of the Liechtenstein family who left detailed records regarding management of their property. According to the 1694 inventory (Materna, 1961[52]), the forests were in very good condition and were providing wood for fuel and building purposes. In the middle of the 18th century several natural disasters struck the area. In 1739, a severe storm destroyed many trees and shortly after, in 1740, an eight week long fire caused further devastation (Knies, 1902[39]). Several shorter fires followed resulting in an average loss of 50% of all trees. According to the so called Plumlov Inventory prepared by De Gesau in 1764 (Nožička, 1957[57]), the dominant species remaining after the calamities were beech and fir accompanied by oak and pine.

The human impact on the forests during the beginning of the 18th century was also significant. Smelters built in the nearby towns of Adamov and Blansko demanded wood for charcoal production. The preferred tree was beech which was needed in large quantities and thus the forests in the area suffered severely. For decades timber harvesting was conducted mercilessly and without any plan. By the year 1732 hardly any beech tree were left in the vicinity of Blansko (Materna, 1961[52]). Only after forest devastation was almost complete did any improvement begin. However, progress in the forest management was not long lasting. By the year 1785, cutting of trees once again surpassed planting of seedlings. An 1810 map of the area shows only very limited forested sections. By the mid eighteen hundreds, beech had virtually vanished. What remained of the forests was destroyed by a cyclone in the year 1868. Between the years 1888 and 1897 the type of vegetation changed dramatically. The need for charcoal diminished and wood was used for heating and building instead. The fast growing spruce became the most wanted tree and was preferentially planted. By the beginning of the 20th century it became the dominant tree in the area of the Moravian Karst and this situation lasts till present day.

## 1.4. Organization of Thesis

The thesis consists of three published papers. Chronologically the first presents preliminary work in the Kulna Cave and is presented in Appendix A as it was later superseded by the fundamental paper summarizing investigations in this cave presented in Chapter 2. A paleoenvironmental study on the interior cave sediments conducted in Spiralka Cave is found in Chapter 3. Chapter 4 contains a conclusion to our environmental magnetic studies in the Moravian Caves as well as suggestions for future work.

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## Chapter 2

# A Late Pleistocene Paleoclimatic Reconstruction Based on Mineral Magnetic Properties of the Entrance Facies Sediments of Kulna Cave, Czech Republic

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# A Late Pleistocene Paleoclimatic Reconstruction Based on Mineral Magnetic Properties of the Entrance Facies Sediments of Kulna Cave, Czech Republic

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SUMMARY

Kulna Cave is located in the Moravian Karst, a well-developed karstic region formed in Devonian limestones in the eastern part of the Czech Republic. The entrance facies sediments in the cave consist of interbedded layers of silts (loess) and clay-rich silts (loam) that were either directly blown into the cave entrance or redeposited in the cave by slope processes during the Last Glacial Stage. The layers of loess and loam overlie fluvial sands and gravels deposited during the Last Interglacial. Previous research at Kulna concentrated on the archaeology, palaeontology and dendrology of these entrance facies sediments. From these data, palaeoenvironmental conditions in the vicinity of the cave were reconstructed. Our results suggest that susceptibility variations and in particular variations in pedogenic susceptibility yield a more detailed record of the palaeoenvironmental conditions at the cave during the Last Glacial Stage. Magnetic susceptibility ($\chi$) was measured on approximately 700 samples collected throughout three well-studied profiles in the cave entrance. The $\chi$ record is well defined and correlates from one profile to another. Mineral magnetic measurements [$F_D$, ARM/SIRM, S-ratio, $\chi(T)$] suggest that $\chi$ variations in the Kulna sediments from the Last Glacial Stage are controlled by the concentration of magnetite and /or maghemite formed during pedogenesis. After the removal of the effects of fine carbonate debris and detrital ferromagnetic minerals on the bulk x record, we obtained a record of pedogenic susceptibility ($\chi_p$) that serves to quantify the concentration of magnetic minerals formed during pedogenesis. Therefore, $\chi_p$ can be thought of as a proxy reflecting the intensity of pedogenesis, which in turn is controlled by climate. Our $\chi_p$ record is also in good agreement with the median grain size record of the Kulna sediments (another proxy for climatic change). We suggest that in the case of Kulna, $\chi_p$ is more sensitive to climate change than bulk $\chi$. The Kulna pedogenic susceptibility record shows variations on both long and short timescales. The long-term trends are in good agreement with the deep-sea SPECMAP record, while the short-term oscillations correlate well with rapid changes in the North Atlantic sea surface temperatures. Our results suggest that Central European climate during the Last Glacial Stage was strongly controlled by the sea surface temperatures in the North Atlantic. Short-term warmer events and perhaps higher precipitation over the mid-continent increased the intensity of pedogenesis. Given the number of independent climate proxies determined from the entrance facies of the cave and their high resolution, Kulna is an extremely important site for studying Late Pleistocene climate.

**Key words**: caves, entrance facies, grain size, magnetic susceptibility, palaeoclimate, pedogenesis.

## 2.1. Introduction

Magnetic properties of iron oxides found in naturally occurring sediments have been shown to reflect changes in environmental processes operating at Earth's surface. The most commonly measured magnetic property, mass specific magnetic susceptibility ($\chi$), is a function of concentration, grain size and mineralogy of magnetic minerals. This dependance has lead researchers to interpret changes in $\chi$ measured in sedimentary sections in terms of environmental changes occurring at the time of deposition. Attributing environmental significance to $\chi$ and other mineral magnetic parameters has spawned a new field of study known as "environmental magnetism". In the last two decades, enormous progress has been made in the successful application of magnetism to paleoenvironmental and paleoclimatological problems. Thompson and Oldfield (1986)[53], King and Channell (1991)[25] and Verosub and Roberts (1997)[58] provide excellent reviews of our current state of knowledge. These reviews demonstrate that measurements of $\chi$ and other magnetic parameters have a growing role to play in many diverse and important areas of the environmental sciences.

Magnetic minerals are affected by physical and chemical weathering (including pedogenesis), intensity of erosion, and by the energy of transport of the sediments. Since these processes produce a wide range of sediments, mineral magnetic methods can be utilized in a broad range of environments. In loess-paleosol sequences, for example, $\chi$ proved to be a sensitive detector of the long-term, large-scale terrestrial climatic change (e.g. Kukla et al., 1988[29]; Banerjee, 1994[2]; Maher et al. 1994[32]; Heller and Evans, 1995[20]). In deep sea sediments, concentration dependant magnetic parameters have been used to decipher the paleoclimatic record in a very simple, rapid and non-destructive matter (e.g. Robinson, 1986[43]; Colin et al., 1998[7]; Arnold et al., 1998[1]). In addition, magnetic properties can be used to discriminate between different fluxes of terrigennous sediments and to detect in situ production (e.g. Tarduno, 1994[51]). The realization that $\chi$ provides not only a useful parameter for high resolution core correlation but is also rapidly affected by human impact, triggered a wide range of lake-watershed studies (e.g. Oldfield et al. 1979[37]; Banerjee et al. 1981[4]; Peck et al. 1994[39]). Lake sediments became increasingly important for short term climatic studies as the need to set present day environmental processes in a longer time perspective became apparent (Thouveny et al., 1994[54]; Snowball, 1995[54]).

One sedimentary environment which has not been exploited adequately by the environmental magnetists is the cave environment. The cave environment can be divided from the sedimentological point of view into an entrance facies and an interior facies. The entrance facies includes fine grained sediments transported from the vicinity of the cave by wind and water and the coarser debris transported into the cave by slope processes. Stratigraphically, the entrance facies is the most valuable section of the cave. The cave entrance contains pollen as well as datable archeological and paleontological remains which are protected from surface erosion, weathering, and biochemical alteration. The interior facies develops in those parts of the cave which are more remote from the surface. Sedimentary sequences here are often extensive, consisting of coarse gravels and sands of fluvial origin overlain by flood deposits of laminar silts and clays. Due to the dynamic environment of the cave interiors, sedimentary sequences often represent a series of depositional and erosional events.

Early magnetic studies conducted in the cave environment were concerned with determining the age of the sediments based on magnetic polarity (Schmidt, 1982[46]) or paleosecular variation (Ellwood, 1971[12]; Creer and Kopper, 1974[8]; 1976[9]; Papamarinopoulos et al. 1992[38]). These studies gave considerable attention also to characterizing the sedimentary fabric and paleoflow directions in the interior facies using anisotropy of magnetic susceptibility (Noel and St. Pierre, 1984[35]; Noel, 1986[34]; Turner and Lyons, 1986[55]). However, they reported very little on environmental magnetic properties of cave sediments. To date, only Ellwood et al. (1996)[13] and Sroubek et al. (1996)[49] have attempted environmental magnetic studies in this environment. Ellwood and co-authors produced a paleoclimatic record for the time interval between 9000 and 3500 years B.P.using magnetic susceptibility measurements on clastic interior facies sediments. Profiles were located in an archeological excavation pit in Konispol Cave, Albania. The authors correlated $\chi$ highs with periods of warm and humid climate as defined from other nearby located European regions. Variations were attributed to the content of pedogenically controlled maghemite which formed outside of the cave and was then washed or blown inside. Ellwood (1999)[14] further expanded on his work in Albania and studied several other caves in the Mediterranean region. He correlated the $\chi$ record from one cave to another as well as with the climatic record from the Greenland ice core. His new work also stresses the usefulness of the cave environment for studying past climate change. Sroubek et al. (1996)[49] reported on preliminary mineral magnetic results from a study of Kulna Cave, Czech Republic. This study yielded a well defined magnetic susceptibility signal, the peaks in $\chi$ being caused by increased concentrations of magnetite and/or maghemite. Climatic change, human occupation, changing source region and depositional processes were suggested as possible candidates controlling the concentration of magnetic minerals.


Fig. 2-1. Location of the Moravian Karst in the eastern part of the Czech Republic.

In this paper we describe a detailed follow up to our preliminary environmental magnetic study of the clastic sediments in the entrance facies of Kulna Cave (Sroubek et al., 1996)[49]. Kulna Cave is located in the Moravian Karst - a well developed karstic region formed in Devonian limestones located in the Eastern part of the Czech Republic (Fig. 2-1). Kulna is an approximately hundred meters long tunnel-shaped cave forming the upper level of the large Sloupsko-Sosuvske Cave System. Kulna contains clastic sediments spanning nearly the entire Upper Pleistocene and has a long history of investigation by traditional archeological methods. The first excavations in the sedimentary filling of the cave began in 1880 when stone tools and bones of extinct animals were found. Systematic research was completed between years 1961 and 1976 by scientists from the Moravian Museum in Brno led by Dr. Karel Valoch. Valoch (1988)[56] subdivided the entrance facies into 14 recognizable stratigraphic units based on archeological artifacts, pollen and charcoal analysis, paleontological findings and several radiocarbon dates. Layers 1-6, were assigned to the Holocene/Late Wurm Glacial Stage, 6a through 9b to the Early Wurm and layers 10 to 13 to the Eemian Interglacial Stage. Layer 14, which originated during the Riss Glacial Stage, is presently buried due to partial collapse in the excavation. Individual layers were found to consist mostly of loam loess-like sediments with varying amount of debris. Layers 1-5 (which have been removed due to industrial activity in the cave) consisted of dark brown sandy silt with varying amount of limestone debris. Layer 6 is a yellow brown loess-like sediment mostly devoid of debris, while layer 6a is a yellow brown loess-like silt with a small amount of fine detritus. Layer 7a consists of interbedded layers of yellow brown loess-like sediment and reddish brown clayey silt, with thin layers of thickness below 1cm. Layer 7b is similar to 7a except that it is totally devoid of debris and laminae thickness is much smaller (1-5 mm). Pellet sand horizons of colluvial origin (Kukla, 1975)[28] are located in the bottom part of this layer. Underlying layers 7c and 7d are both finer grained with layer 7c being richer in limestone debris. Layer 8a is a grey brown, clayey silt with variable amounts of debris. Layer 9 has a similar matrix

| Layer | Fauna | Charcoal | Culture | Stratigraphy | Abs. Age (Ka) | OI stage |
|---|---|---|---|---|---|---|
| 6 | Glacial steppe fauna | Humid climate | Magdalenien | W3, Late Glacial or Young Wurm | 11.5, 7.6, 11.6, 12.3* | 2 |
| 6a | Cold dry steppe | | Micoquien | Stadial or end of Old Wurm | 21.3 | 2 |
| 7a,b | Steppe, colder and warmer | Decidous, present day climate | Micoquien | Stadial or end of Old Wurm | 38.6, 45.7 50± 5✝ | 3,4 |
| 7c | Steppe w/ numerous thermophilous specious | Coniferous forests, mildly cold | Micoquien | Interstadial Kulna or Moershooft | | 5a |
| 7d | Similar as above, forest species dominate | Coniferous forests, mildly cold | Micoquien | Temperate stadial, interstadial | | 5b |
| 8a | Similar as above, forest species rare | Mild to slightly cold | Micoquien | Temperate stadial | | 5c |
| 9 | Forest , warmer | | Micoquien | End of Eemian or Amersfoort | 69± 7✝ | 5d |
| 11b | Steppe and forest | Deciduous forests, colder and more humid climate than present day | Taubachien | Colder oscillation during Eemian | | 5e |
| 12 | steppe | | | Eemian, oscillation 2 | | 5e |
| 13 | warmer | | Taubachine, Middle Paleolithic | Eemian, Oscillation1 | | 5e |
| 14 | Cold, steppe | | Middle Paleolithic w/Levallois technique | Cold oscillation of Eemian or Riss | | 6 |

Table 2-1. Summary of the results from the previous study of the Kulna Cave (modified after Valoch (1988, 1992). The unmarked ages are uncalibrated conventional radiocarbon dates, * indicates an AMS date and ✝ indicates an ESR date by Rink al. (1996).

to 8a but is richer in limestone debris. Layer 10 appeared only infront of the cave and profiles exposing that layer no longer exist. Layer 11 is a complex of grey silty clays

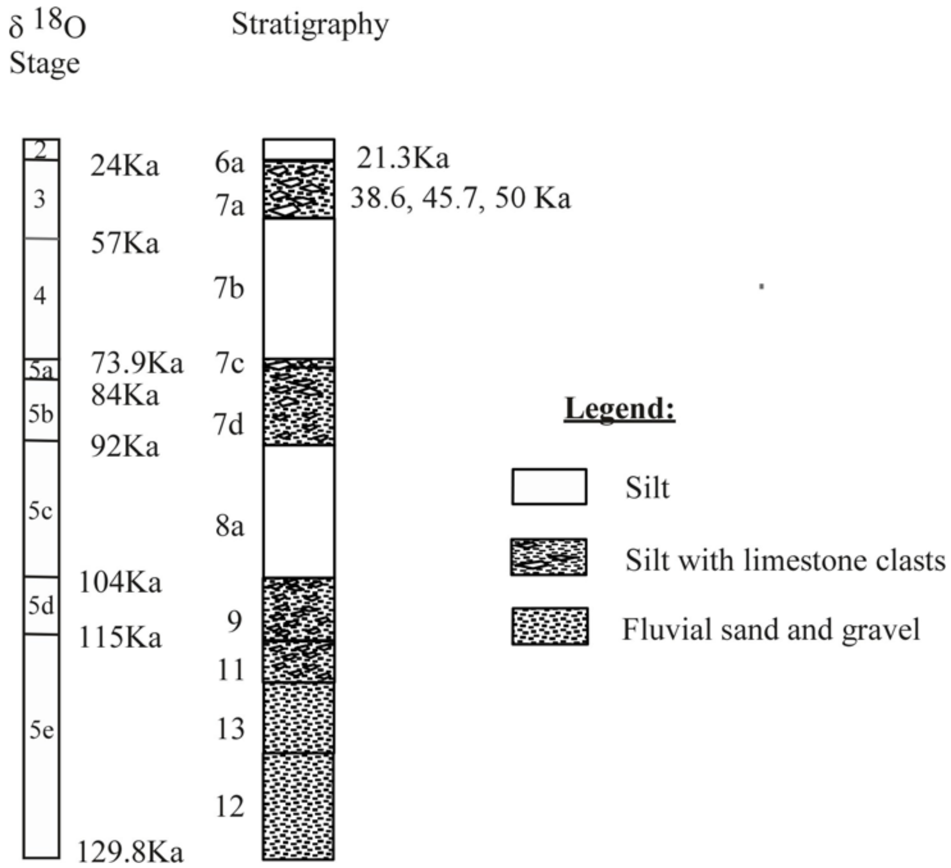



Fig. 2-2. Simplified stratigraphic profile of the Late Pleistocene sequence in the entrance of Kulna Cave. Ages are radiocarbon dates from Valoch (1988), except for the 50 ka age, which is an ESR date from Rink et al. (1996). The correlation with the deep-sea stable oxygen isotope record is from Valoch (1992). Oxygen isotope age model is from Martinson et al. (1987).

rich in limestone debris. Layer 12 is a dark brown fluvial sediment consisting of interbedded layers of clayey silts and sandy gravels. Layer 13 forms a lense of clayey silts within layer 12 and in some locations appears to overlie layer 12. Valoch (1992)[57] also correlated the Kulna stratigraphy with oxygen isotope stages of Labeyrie (1984) thereby tying the Central European climatic record to variations in the global ice volume. Table 2-1 and Fig. 2-2 summarize the findings of Valoch

(1988[56], 1992[57]) in the layers from the Last Glacial and Last Interglacial. Also included in Table 2-1 are the ESR dates on layers 7a and 9 of Rink et al. (1996)[42].

In our preliminary study (Appendix A, Sroubek et al., 1996)[49], we sampled three profiles described by Valoch (Fig. 2-3) located in different parts of the cave entrance (Fig. 2-4) to determine whether the magnetic susceptibility signal showed similar patterns. From each profile, we collected $8cm^3$ samples of the fine grained matrix at a 10 cm vertical spacing. Measurements of mass specific magnetic susceptibility ($\chi$) indicated that sediments in the Kulna Cave did yield a well defined signal correlatable from one profile to another (see Fig. 9 of Sroubek et al., 1996[49] in Appendix A). Additional magnetic parameters suggested that the variation in $\chi$ is controlled mostly by the concentration of magnetite and/or maghemite rather than by mineralogical or grain size changes.

## 2.2. Sampling and Laboratory Methods

Our present more detailed work in Kulna Cave focuses on obtaining a high resolution mineral magnetic record and on interpreting the paleoclimatic significance of this record. To this end, we collected close to 700 unoriented $8cm^3$ (2 cm cube) from the same profiles (Fig. 2-4) sampled for our preliminary work (Appendix A, Sroubek et al., 1996[49]). All the layers were sampled continuously (distance between center of samples on average 2.5 cm) except for layers 9 and 11 which were sampled at a 10 cm interval in order to preserve the integrity of the profile and to avoid frequent limestone debris fragments. Note that in the sampled profile 7 layer

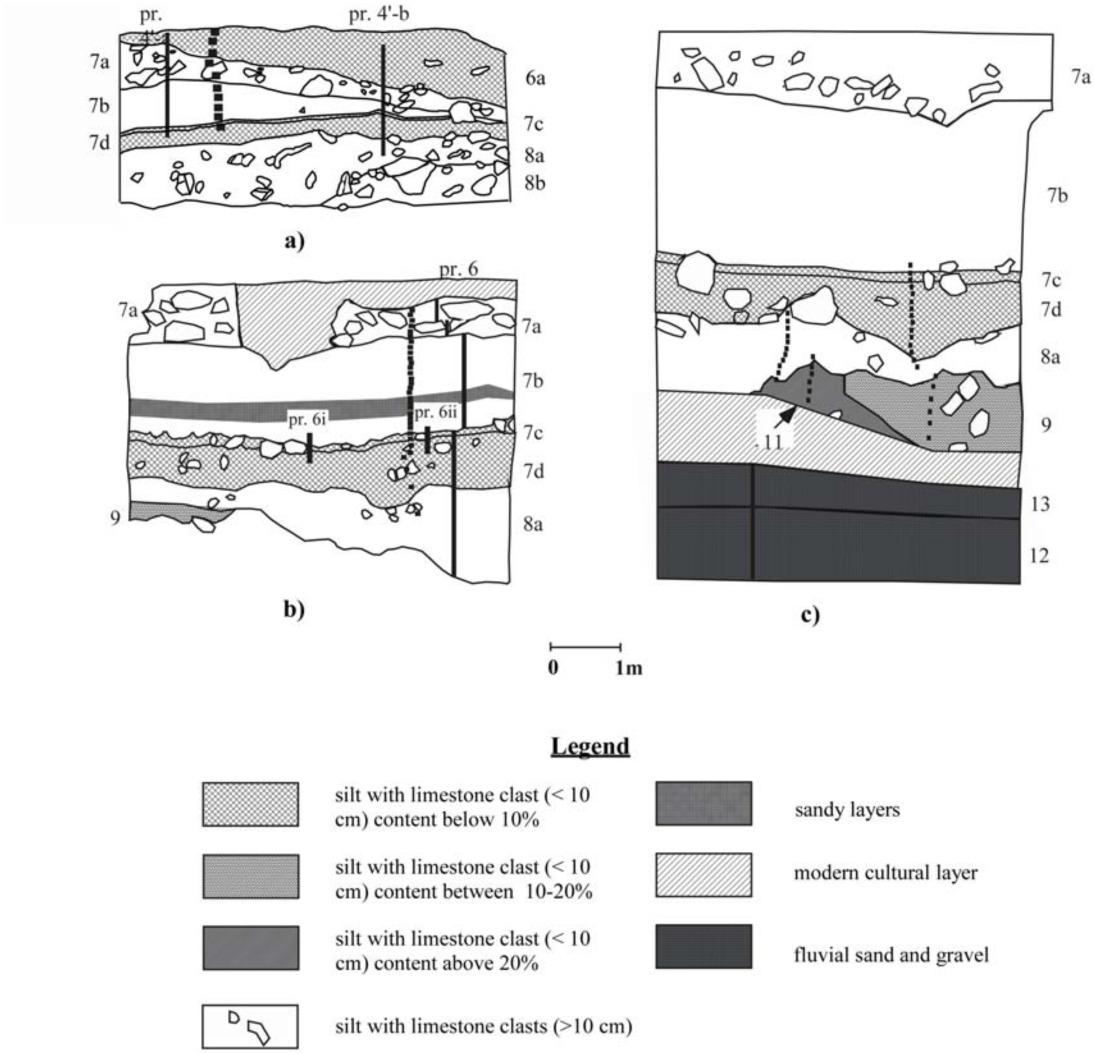



Fig. 2-3. Sketch of (a) profile 4k, (b) profile 6, and (c) profile 7 in the entrance of Kulna Cave. Squares indicate locations of preliminary samples; bars indicate locations of detailed sampling (sample spacing 2.7 cm) as discussed in the text. The modern cultural layer shown in profile 7 (see legend) includes a support and a metal ramp constructed for easier access to the profile.

13 was overlying layer 12 and therefore appears above it also in the stratigraphic column.

$\chi$ of all samples from each of the profiles was measured on the Kappabridge KLY-2 instrument. The magnetic susceptibility measurements were complemented by the following rock magnetic parameters: Paramagnetic susceptibility ($\chi_{para}$) was determined on 3-4 samples from each layer from the high field segments (above 300 mT) of the hysteresis loops as measured on the Variable Frequency Translation Balance (VFTB). Samples from certain layers contained abundant carbonate debris.


Fig. 2-4. Sketch of the southern entrance and central part of Kulna Cave. Pits excavated by Valoch and co-workers are represented by hatched polygons. Thicker lines (labelled 4k, 6 and 7) represent locations of sample profiles.

To remove the effect of carbonates on magnetic susceptibility, all samples were treated by acetic acid which dissolved the carbonate. Frequency dependence of magnetic susceptibility ($F_D$), was calculated using measurements obtained on a Bartington Instruments MS-2 susceptibility meter at 470 Hz and 4700 Hz. Saturation isothermal remanent magnetization was acquired for every fourth sample in a profile at 1T (SIRM, $IRM_{1000mT}$) and at 300 mT ($IRM_{300mT}$) on a Sapphire Instruments SI-6 pulse magnetizer. The ratio $IRM_{300mT}/IRM_{1000mT}$ was used to obtain the S-parameter. The anhysteretic remanent magnetization (ARM), acquired for every fourth sample in a profile using the Sapphire Instruments SI-4 AF demagnetizer (biasing field 0.05 mT the alternating field 1T), was used to calculate the magnetic grain size dependant ratio ARM/SIRM. All remanence parameters (IRMs and ARM) were measured on a Schonstedt SSM-2 spinner magnetometer. The behavior of magnetic susceptibility as a function of temperature was measured on the Kappabridge KLY –2 equipped with a furnace CS-2 on two samples from every layer. Heating (from 20 to 700 °C, step 3 °C) in argon atmosphere prevented oxidation of the sample. Anisotropy of magnetic susceptibility (AMS) was measured on 15 oriented cylindrical samples ($6cm^3$) for each of the layers except for layer 6a for which 3 samples were measured. This layer was no longer accessible after the manager of Kulna (Moravsky Kras Co.) opened the cave to the public. The AMS tensor for each sample was calculated, using 15 individual measurements on each sample on the Kappabridge KLY-2, as prescribed by methods developed by Jelinek (1978)[24].

To complement our magnetic measurements we measured the grain size of the fine grained matrix of Kulna Cave. Grain size distributions (0.7-700 μm) were measured on 150 mg subsamples using a 20 channel Leeds and Northrup MICROTRAC II laser particle size analyzer. Morphology of guartz grains (Quartz exoscopy) was studied in order to determine the mode of transport and post-depositional changes. Five grains from each layer were analyzed under high magnification using the JOEL jsm-35 C electron microscope.

To compare our work in Kulna with other similar surficial deposits we collected 10 samples on loess and 10 samples on paleosols from the Cerveny Kopec (Red Hill) outcrop in Brno and 10 samples from the fluvially redeposited loess-like sediments in river terrace of Bila Voda Stream near Holstejn village in the Moravian Karst. Magnetic susceptibility, frequency dependance of magnetic susceptibility and grain size distribution were measured on these samples.

## 2.3. Laboratory results

The detailed susceptibility records from profiles in the entrance of the Kulna Cave as expected, show good correlation (Fig. 2-5) as was also the case with our preliminary work. In the two exposures of layer 6a, $\chi$ shows low values between 2-3 $x10^{-7}$ $m^3/kg$. A rise in susceptibility up to 4 x $10^{-7}$ $m^3/kg$ can be observed in all 3 exposures of layer 7a. Layer 7b contains two distinct peaks in the lower half of the layer reaching up to 6 x $10^{-7}$ $m^3/kg$. Except for profile 4'-a $\chi$ drops in all profiles across layer 7c and the top of layer 7d to $3x10^{-7}$ $m^3/kg$. In the lower part of layer 7d $\chi$ again increases up to 5 x $10^{-7}$ $m^3/kg$. Layers 8a, 9 and 11 show fairly rapid changes in $\chi$ values. Finally, the sediments of layers 12 and 13 at the base of the profile have distinctly lowest $\chi$ values around $1x10^{-7}$ $m^3/kg$.

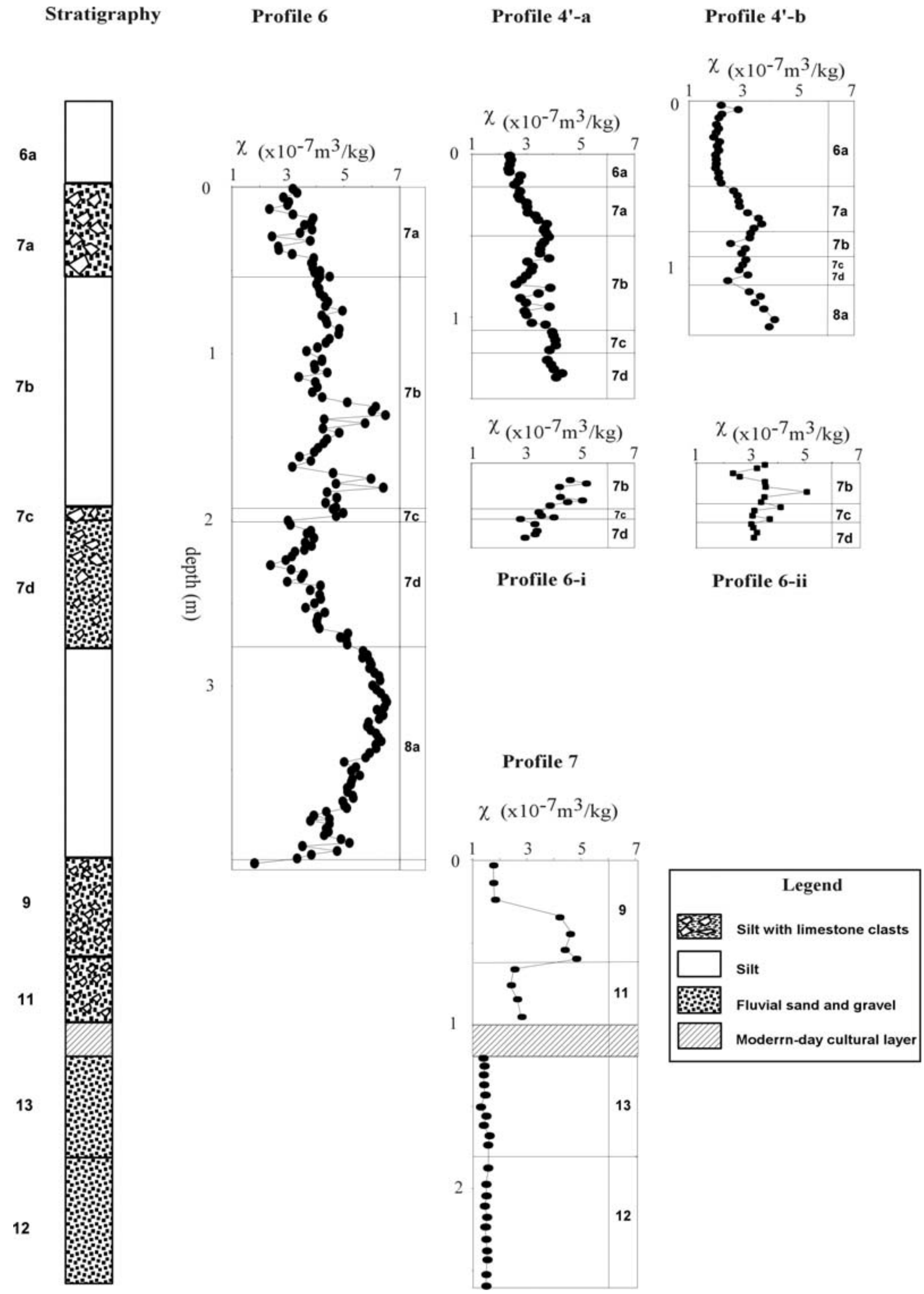



Fig. 2-5. Detailed record of the magnetic susceptibility measured throughout the main entrance of Kulna Cave. Profile and sampling locations are shown in Fig. 2-4.

The $\chi$ trends in the same layers sampled in different profiles are similar, therefore we have created a composite record (Fig. 2-6a) summarizing the $\chi$ variations in layers 6a-11 of the entrance facies in Kulna Cave. The $\chi$ data in this composite record are typically from those outcrops where each layer reached its maximum thickness: layer 6a in profile 4, layers 7a, 7b, 7c, 7d and 8a in profile 6 and layers 9 and 11 in profile 7.

The composite record of $\chi$ shows in general a good agreement with the climatic interpretation of Valoch (1988)[56] and the conclusions of Ellwood et al. (1996, fig. 7[13]), i.e. the highest $\chi$ values are associated with layers interpreted to have been deposited during warmer climate. However, several discrepancis are apparent. For example, layers 7c and 7d, which according to Valoch (1992)[57] originated during relatively warmer oxygen isotope stages (OIS) 5a and 5b have lower $\chi$ values than the overlying layer 7b deposited during stadial conditions (OIS 4). Similarly layers 9, 11, 12 and 13, deposited during the near optimal conditions of OIS 5d and 5e as suggested by Valoch (1992)[57], show generally very low $\chi$ values.

To explain the discrepancies between the $\chi$ record and the temperature data discussed above we have undertaken the following experiments to determine the cause of $\chi$ variations: a) measurements of frequency dependance of $\chi$ ($F_D$, Fig. 2-6b) in order to estimate the concentration of ferromagnetic grains near the SD/SP boundary, b) measurements of the ARM/SIRM ratio (Fig. 2-6c) in order to characterize variations in grain size of the remanence carrying grains (SD and larger), larger values of ARM/SIRM indicate smaller magnetic grain size, c) measurements of the S-ratio (Fig. 2-6d) in order to quantify the ratio between high

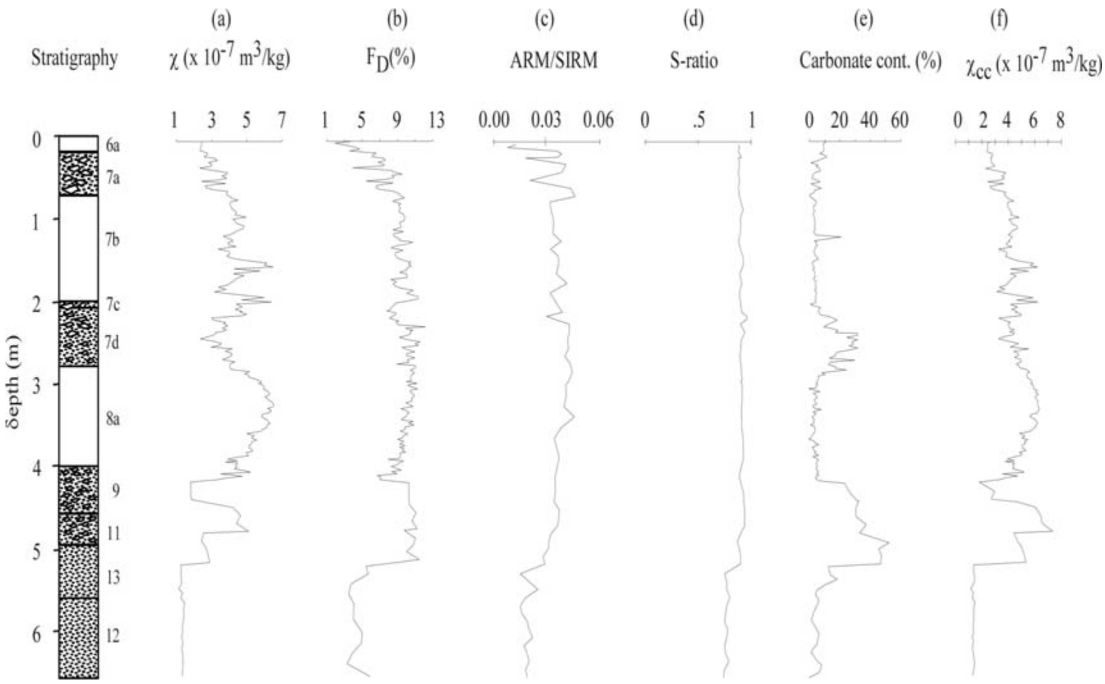



Fig. 2-6. Magnetic parameter data from the composite profile constructed from the individual profiles sampled in the entrance facies of Kulna Cave. (a) $\chi$, (b) $F_D$, (c) ARM/SIRM, (d) S-ratio, (e) carbonate content in weight per cent, and (f) $\chi_{cc}$.

and low coercivity minerals and d) measurements of paramagnetic susceptibility quantifying the contribution of paramagnetic minerals to the bulk $\chi$, and e) mesurements of $\chi$ versus temperature (Fig. 2-7) in order to determine the magnetic minerals present.

The record of $F_D$ shows a gradual increase in the upper part of the profile (layers 6a, 7a) from values as low as 2% up to approximately 9%. Throughout most of the profile (layers 8a to 11) the $F_D$ values are between 9 and 11%. In two lowest most layers (13 and 12) $F_D$ values are once again quite low, typically below 5%. The ARM/SIRM record has a similar trend to the $F_D$ data. The ARM/SIRM shows a rather scattered increase throughout layers 6a, indicating a decrease in magnetic grain size. Throughout layers 8a to 11 the ARM/SIRM values shows minimum variability, but decreases towards the base of the profile in layers 13 and 12. These layers thus contain larger magnetic grains. The S-ratio is nearly constant (between 0.85-0.95) in layers 6a-11, then drops rapidly at the boundary between layers 11 and 13 and is again nearly constant throughout layer 13 and 12 (between 0.75-0.8). The S-ratio results indicate that the upper layers contain an increased concentration of low coercivity magnetic minerals as compared to the bottom two layers. Paramagnetic susceptibility (data not presented) contributes approximately 20% to the bulk $\chi$ signal in layers 6a to 11 and by nearly 40% in layers 13 and 12. High temperature susceptibility curves for layers 7b and 7d show a very similar trend. Both curves show nearly constant values of $\chi$ to about 200 °C, after which they increase to a maximum slightly below 300 °C and then decrease to about 400 °C. Above 400°C $\chi$ rises rapidly till 450 °C and then drops to a constant background value. The first rise in the record probably indicates the transition of Fe-hydroxides into maghemite, the $\chi$ loss between 300 and 400 $^{o}$C is attributed to maghemite converting to hematite, the second $\chi$ rise suggests reduction of hematite to magnetite and the $\chi$ loss around 580 $^{o}$C confirms the presence of magnetite and/or a phase converting to magnetite. The sample from layer 6a shows a similar trend to the previously discussed two samples except for the presence of the first peak at approximately 300 °C. The second peak is present in this sample, but is less pronounced. Layer 6a thus probably contains no (or minimum) amount of Fe-hydroxides and less hematite. Our preliminary results on IRM acquisition and on

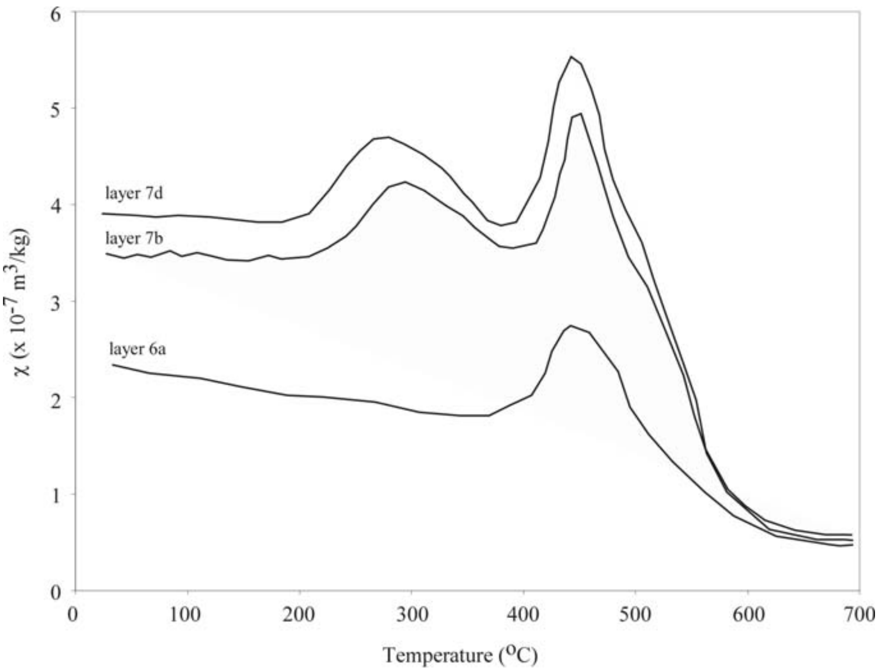



Fig. 2-7. High-temperature susceptibility measurements of samples from the Kulna Cave sediments. Samples were heated in argon atmosphere; only the heating curves are shown.

thermal demagnetization of SIRM presented in Figures 5 and 6 of Sroubek et al. (1996)[49] also suggest the presence of maghemite or low coercivity magnetite in all the layers of the entrance facies of Kulna.

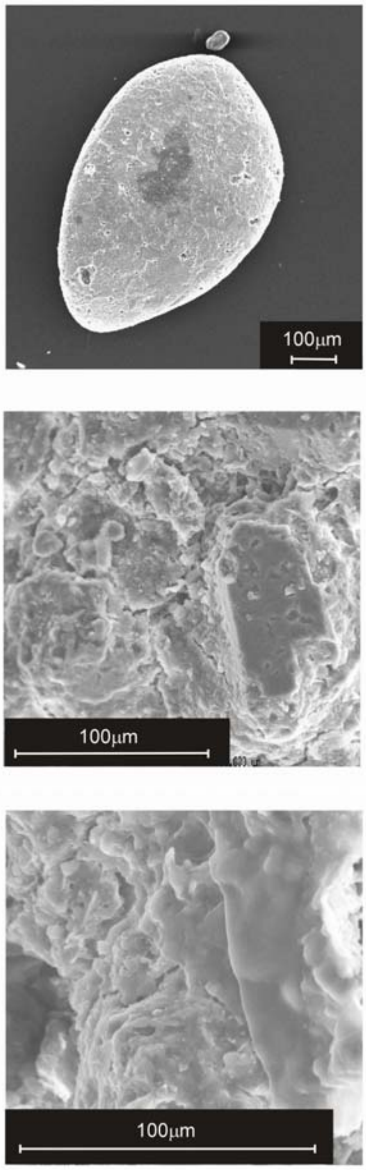
The mineral magnetic properties of loess in layer 6a suggest that the low $\chi$ values are caused by a lack of grains near the SP/SD boundary. In loam-rich layers

7b, 7c, 7d, 8a, 9 and 11 the content of the grains near the SP/SD boundary is higer and $\chi$ variations are probably controlled by changes in concentration of these fine grains. Layer 7a appears to be a transitional layer. In its upper part the magnetic grain size and concentration of grains near the SP/SD boundary is low and similar to layer 6a. Throughout layer 7a the magnetic grain size decreases and at its base is similar to the underlying layer 7b. Magnetic mineralogy throughout layers 6a to 11 does not show significant variations except for the lack of Fe hydroxides in layer 6a. The fluvial deposits of layer 13 and 12, show a distinctly different magnetic signature as indicated by the increased presence of paramagnetic and high coercivity ferromagnetic minerals as well as by the increase in magnetic grain size.

Mineral magnetic properties of layers 6a to 11 are very similar to those of surficial loess/paleosol sequences. Low values of $F_D$ in layers 6a and 7a are similar to those we measured on unweathered loess in the Cerveny Kopec profile ($F_D$ = 3-5%), whereas the high values in layers 7b to 11 correspond well with paleosol samples from Cerveny Kopec ($F_D$ = 8-11%). Similar elevated values of $F_D$ are common in paleosols interbedded in loess sequences around the world (e.g. Evans and Heller, 1994[15]; Eyre and Shaw, 1994[17]). Also the high temperature susceptibilitity curve for layer 6a is similar to the results of Oches and Banerjee (1996)[36] measured on loess samples from the classic profile at Dolni Vestonice whereas the high temperature susceptibilitity curves for layer 7b and 7d are similar to results on the parabraunerde soils from Dolni Vestonice.

The above discussion suggesting that sediments in layers 7a to 11 underwent pedogenic weathering is supported by the analysis of the surface of quartz grains from these layers. Pedogenic dissolution as well as in-situ production of new phases is apparent from the morphology of quartz grain surfaces. Fig. 2-8 shows both dissolution features as well as newly precipitated silica on the surface of a wind blown silt sized quartz grain. The pedogenic weathering of the sediments probably occurred either in the source area or more probably in the near vicinity of the cave immediately prior to transport into the cave entrance. AMS data for layers 7b to 11 presented in Fig. 2-9 suggest that settling velocities were low as the foliation plane for these sediments is near horizontal and the degree of anisotropy is minimal (below 3 %). The Kulna data are typical for an environment in which the anisotropy is controlled by the compaction process rather than by settling energy or surface topography (Tarling and Hrouda, 1983[52]). The direction of the mean principal anisotropy axis, pointing northwest-southeast, suggest the sediments were transported into the cave through the southern cave entrance. Directional AMS data for each layer indicate that at a 95% confidence level there is no difference in the direction of transport or sedimentary fabric.

The similarity between the magnetic properties of the sediments deposited in layers 6a to 11 with those of the surficial loess/paleosol sequences suggests that the $\chi$ variations throughout layers 6a to 11 should be a good proxy for climate during the Last Glacial. The reason why we observe discrepancies between the climatic record of Valoch (1988, 1992)[56][57] and our data is probably due to the fact that magnetic




a) The smooth surface of this round sand sized grain is an evidence of aeolian transport.

b) Detail of one indentation on the surface of the quartz grain from figure a). The surface is significantly disturbed by corrosion occuring during the process of pedogenetic weathering . Note the original smooth surface in the lower right hand side of the image.

c) Another detail of an indentation on the surface of the quartz grain from figure a). The smooth surface in the right hand side of the image is amorphous silica which precipitated during the process of pedogenic weathering.

Fig. 2-8. Electron microscope images of quartz grains from layer 7b of the entrance facies sediments of Kulna Cave. The surface features yield evidence of the mode of transport and of post-sedimentary processes.

susceptibility measured in Kulna is not only a measure of the concentration of pedogenic ferromagnetic minerals, but is also affected by detrital ferromagnetic grains and both paramagnetic (mostly clays) and diamagnetic (quartz and carbonates) minerals. These contributions, if significant, must be subtracted from the bulk $\chi$ to

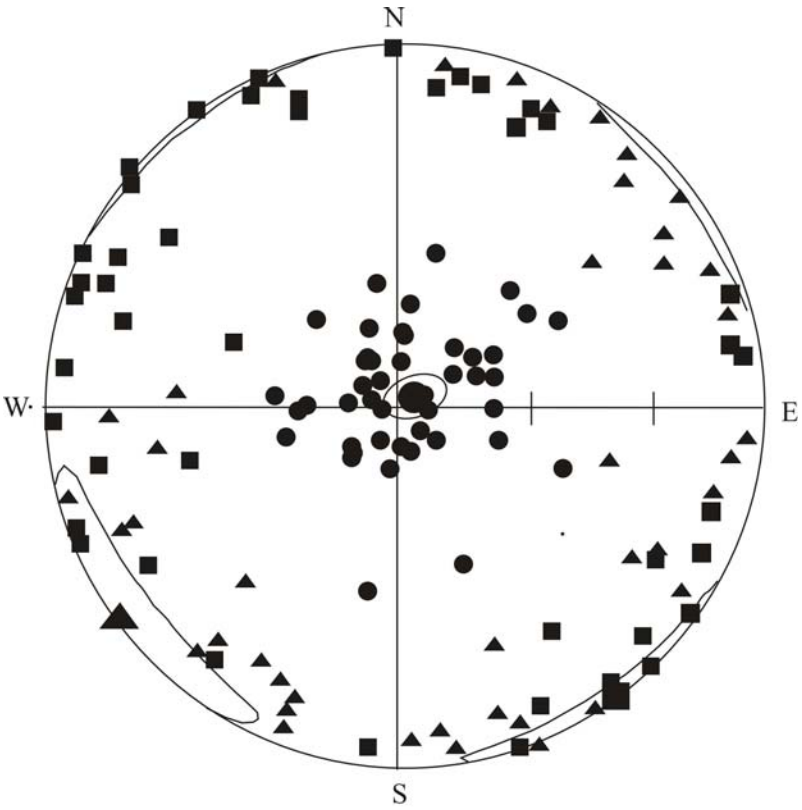



Fig. 2-9. Directions of the three principal axes of magnetic susceptibility for sediments from layers 7b-11 of Kulna Cave. Squares represent maximum axes, triangles intermediate axes and circles minimum axes. The larger symbols represent the means and the ellipses indicate the 95 per cent level of confidence of the mean. The orientations of the mean directions are as follows. Maximum: D=143°, I=0°; intermediate: D=233°, I=4°; minimum: D=52°, I=86°.

obtain a true pedogenic signal. The concentration of paramagnetic minerals in Kulna Cave is fairly constant (~ 20%) throughout the layers and thus does not contribute significantly to susceptiblity variations shown in Fig. 2-6a. On the other hand, the variable amount of carbonate debris (Fig. 2-6f) affects the $\chi$ signal. In layer 6a the carbonate content is approximately 10%, and in layers 7d, 9 and 11 varies between 10 to 50%. In all the other layers carbonate debris is mostly below 5%. The magnetic susceptibility signal after the subtraction of the diamagnetic carbonate contribution ($\chi_{cc}$, Fig. 2-6f) has a similar character to the bulk $\chi$ record on Fig. 2-6a but shows significantly higher values than the $\chi$ record in the carbonate rich layers.

The $\chi_{cc}$ record is composed of a pedogenic component created by authigenic grains near the SP/SD boundary and a background component originating from larger detrital ferromagnetic grains. Forster et al. (1994)[18] suggest that pedogenic and background susceptibilities can be separated utilizing the del$\chi$ vs.$\chi_{cc}$ plot, where del$\chi$ is the difference in magnetic susceptibility measured at two frequencies. The assumption behind this test is that pedogenic susceptibility ($\chi_p$) is carried by grains near the SD/SP boundary and that the larger grains, which show no frequency dependance of $\chi$, are responsible for the background susceptibilitity ($\chi_b$). $\chi_b$ is defined by the intercept between the best-fit-line to the data points on the del$\chi$ vs.$\chi_{cc}$ plot and the $\chi_{cc}$ axis. The pedogenic component ($\chi_p$) is the difference $\chi_{cc} - \chi_b$, $F_c$ (%), the "collective true frequency dependence" characterizing the grainsize distribution of the SP fraction, is the slope of the best-fit-line to the del$\chi$ vs.$\chi_{cc}$ plot. For determining $\chi_b$ and $\chi_p$ in the Kulna sediments we followed the approach of Forster et al. (1994)[18] and constructed del$\chi$ vs.$\chi_{cc}$ plots for each layer in Kulna (Fig. 2-10).

The best-fit-lines suggest that layers 7b, 7c, 7d and 8a have both very similar intercepts or $\chi_b$ (between 1 and 1.4 x$10^{-7}$ $m^3/kg$) as well as the slopes or collective $F_c$ (around 12%). Layers 9 and 11 have a somewhat higher $\chi_b$=2.41 x$10^{-7}$ $m^3/kg$ but exhibit the same $F_c$. Layer 6a is distinctly different with an intermediate $\chi_b$=2.04 x$10^{-7}$ $m^3/kg$) and a very high $F_c$ (21%). Layer 7a appears to be a transition between 6a and the underlying layers with $\chi_b$ =1.63 x$10^{-7}$ $m^3/kg$ and $F_c$ = 16%. The results from

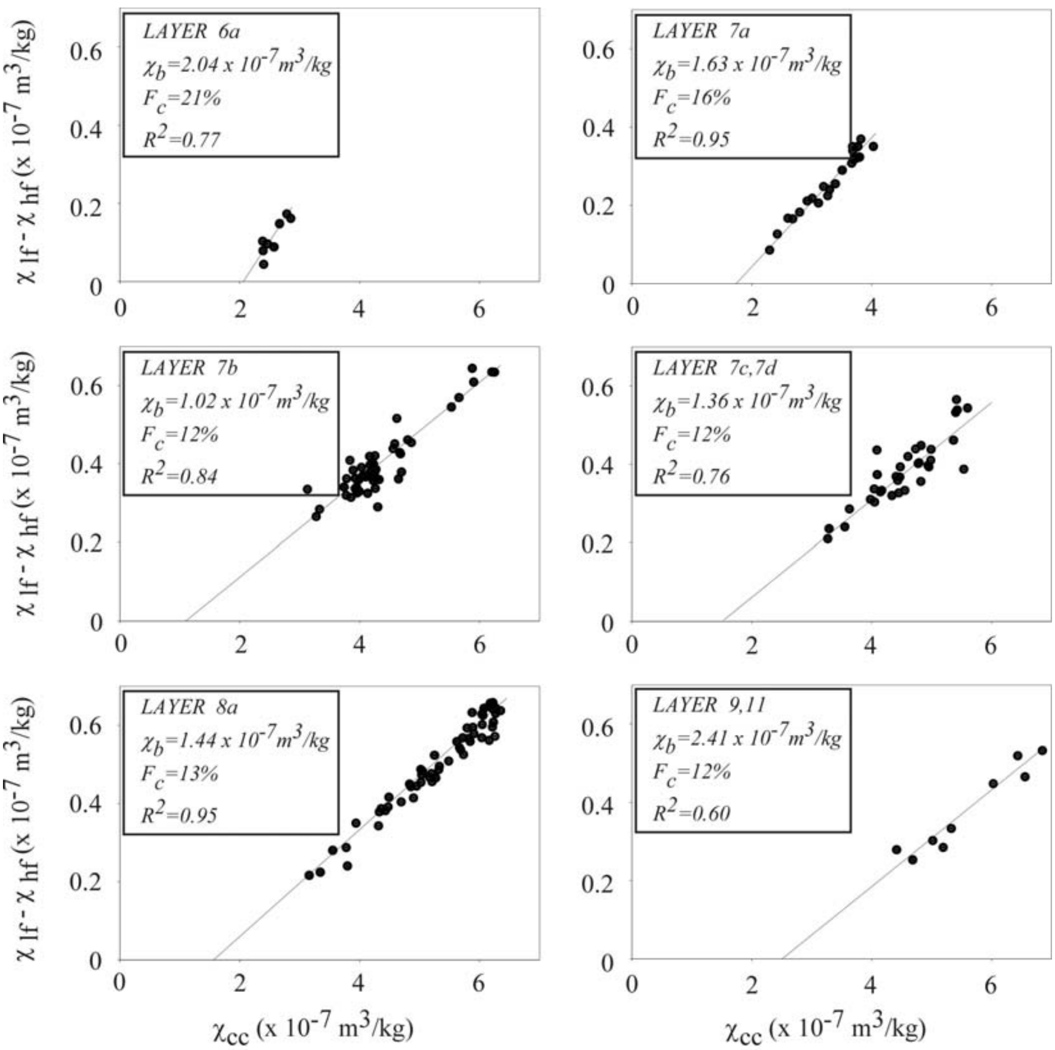



Fig. 2-10. Plot of magnetic susceptibility after removal of carbonate debris ($\chi_{cc}$) versus the difference between low- and high-frequency magnetic susceptibility ($\chi_{lf}$-$\chi_{hf}$) for individual sedimentary layers deposited in the entrance of Kulna Cave. The best-fit line through the data points yields the background susceptibility ($\chi_b$), equal to the intercept $\chi_{cc}$ axis and the collective true frequency dependence ($F_C$) given by the slope of the best-fit line. $R^2$ is the coefficient of determination measuring the quality of the fit of the line to the data points. For further explanation see text.

Kulna's layer 6a are similar to unweathered loess beds around Brno which yield $\chi_b$ values between 1.5-2.2 x$10^{-7}$ $m^3$/kg (Forster et al., 1996[19]). The rest of layers in Kulna have characteristics which correspond well to the Late Pleistocene loess/paleosol sequencies of Southern Moravia ($\chi_b$ = 0.8-1.1 x$10^{-7}$ $m^3$/kg and $F_c$ around 13%, Forster et al., 1996[19]).

Utilizing the $\chi_b$ values for each layer we calculated $\chi_p$ for each sample in the composite profile (Fig. 2-11). The record of $\chi_p$ shows lower values than the $\chi_{cc}$ record (Fig. 2-6f). The difference between $\chi_p$ and $\chi_{cc}$ is most significant for layers 6a and 7a, which consist mostly of unweathered loess where the pedogenic component is minimal. Also in layer 9 and 11 the $\chi_p$ values are significantly lower than in the $\chi_{cc}$ record. One possible explanation is that during intensive pedogenesis the newly formed magnetic grains can grow to larger sizes than the SP/SD boundary and then behave as if they were part of the background signal. If this explanation is correct the method of Forster et al. (1996)[19] must be applied and interpreted with greater care.

Grain size measurements on the sedimentary matrix in Kulna were conducted in order to provide independent climatic proxy-data, which could be compared with the $\chi_p$ record. Pure loess is typically well sorted, with up to 70% of the grains in the range between 10-50 microns and has a positively skewed size distibution (Kukal, 1971[27]; Pecsi, 1990[40]). During pedogenesis, however, grain size distribution of weathered loess becomes bimodal and the mean grain size decreases with the formation of clay minerals (Martini and Chesworth, 1992[33]). Variations in mean grain size and other parameters characterizing grain size distribution thus can be used as climate proxies. For example, Ding et al. (1988)[11] correlated grain size variations in Chinese loess with several short-lived episodes of climatic ameliorations occurring during the Last Glacial (Dansgaard-Oeschger Events) and

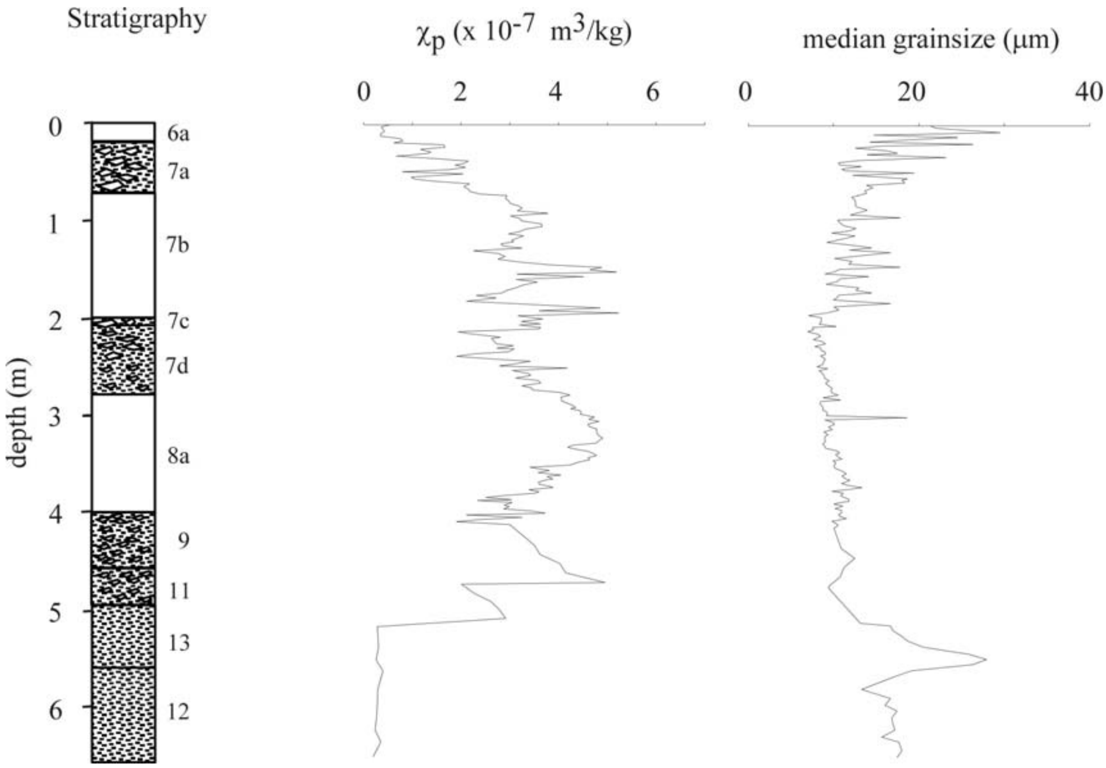



Fig. 2-11. Comparison of pedogenic susceptibility ($\chi_p$) and median grain size measured through our composite profile in he entrance of Kulna Cave. For details on obtaining ($\chi_p$) see text and Fig. 10.

Chen et al. (1988)[6] used the silt/clay ratio as a measure of the intensity of the monsoons. On a more local scale, our grain size measurements from the loess/paleosol complex at Cerveny Kopec **(**Fig. 2-12a,b) thirty kilometers SW from Kulna, show changes in the character of the grain size distribution which can also be attributed to climatic change. Pristine loess from stadials (Fig. 2-12a) is positively skewed with median grain size varying between 20–35 µm and the paleosols formed during interstadials (Fig. 2-12b) are bimodal with median grain sizes as low as 9-14 µm.

The grain size distribution of the matrix of most Kulna sediments (Fig. 2-12d-j) is in all cases bimodal with the silt peak between 30-60 μm and the clay peak between 3-5 μm. We attribute the silt peak to the presence of windblown particles and the clay peak to grains formed during the pedogenic weathering. The ratio of the clay peak to the silt peak therefore reflects the intensity of the pedogenic process prior to deposition in cave. In layer 6a (Fig. 2-12d), the grain size distribution is very similar to the loess (Fig. 2-12a), the silt peak strongly dominates suggesting the presence of mostly eolian particles. In layer 7a the clay peak becomes more prominent as loam laminae are interbedded with the loess layers. In layer 7b samples with both the clay peak dominating and the silt peak dominating can be observed. Layer 7b consists of thin laminae of strongly weathered soil sediments, responsible for the high relative concentration of the clay particles, and of loess laminae which are composed mostly of silt sized grains. In layers 7d, 8a, 9 and 11 the clay peak dominates over the silt peak suggesting intensive weathering. In some cases the silt peak flattens into a shoulder probably due to the presence of fine carbonate detritus. The size distribution of layers 12 and 13 is very similar to surficial fluvial deposits (Fig. 2-12c) from the nearby Bila Voda River.

The ratio between the silt peak (loess) and the clay peak (weathered loams) is well reflected in the median grain size record (Md, Fig. 2-11). The nearly pristine loess in layer 6a thus has maximum grainsize, with the rapid variations in the Md probably recording wind intensities. In layer 7a the Md overall decreases representing a more intense weathering process and the increased presence of loam.

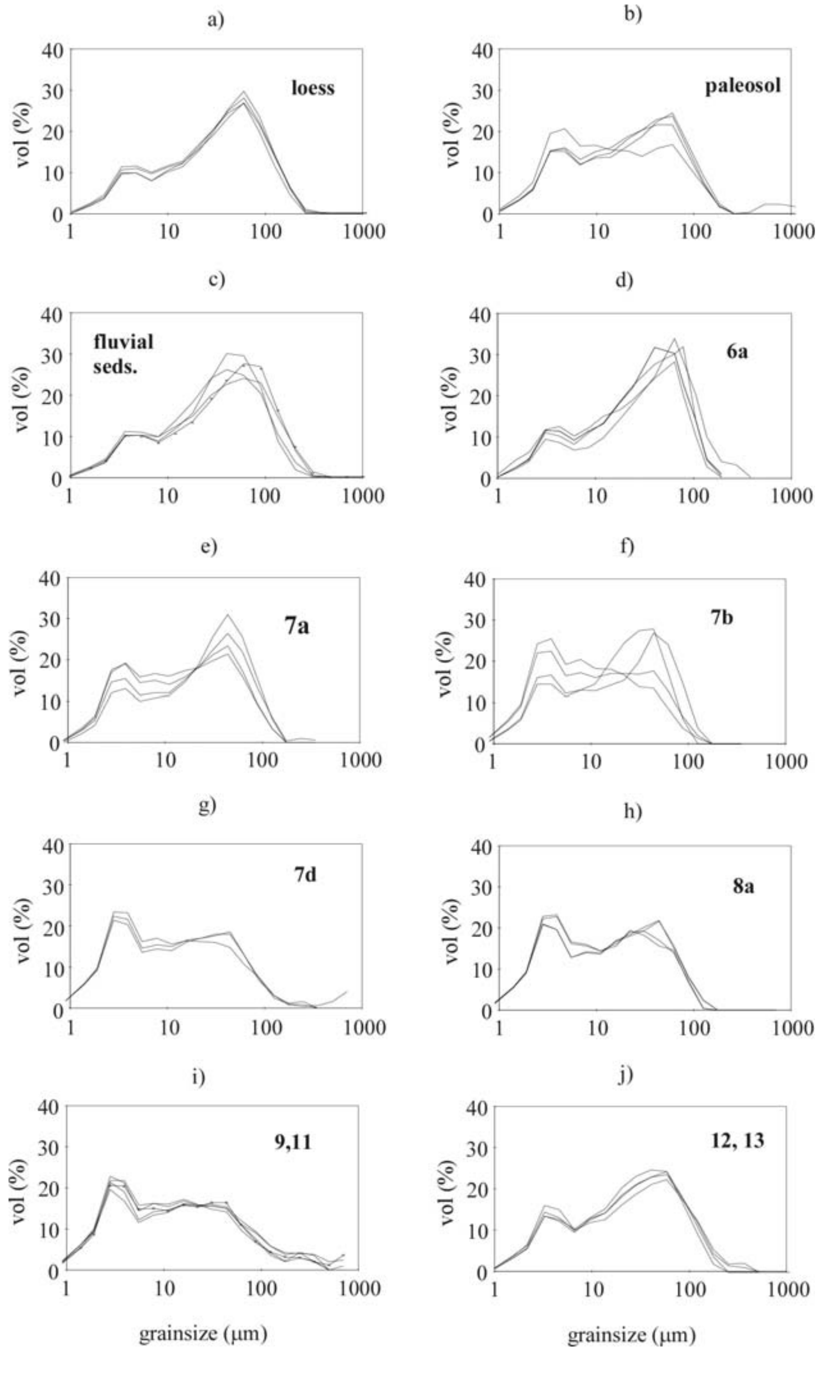



Fig. 2-12. Size distribution of the fine-grained matrix in the entrance facies of Kulna Cave compared to sediments of similar origin. (a) Loess samples from Red Hill (Cerveny Kopec); (b) palaeosol sediments from Red Hill (Cerveny Kopec); (c) samples of fluvial sediments deposited by Bila Voda River; (d–j) samples from layers in the entrance of Kulna Cave.

The rapid variation in Md in layer 7b reflects the presence of alternating loam (low Md) and loess (high Md) laminae. In layers 7d, 8a, 9 and 11 Md stays nearly constant at its minimum values, suggesting that only loams are present and no pure loess remains. Finally, the fluvial sediments in layers 13 and 12 have overall slightly higher Md. The rapid rise and drop in layer 13 reflects the presence of a sandy lens, reflecting a local increase in the energy of the depositional environment. A general inverse correlation between the two paleoclimatic proxies Md and $\chi_p$ is apparent in layers 6a to 7c. The variations in Md and $\chi_p$ in the massive layer 6a and thickly laminated layer 7a, show a high degree of inverse correlation which can also be seen in the fine structure of both records. In layer 7b, this detailed correlation is not apparent mostly because of the different sampling methods used for magnetic and grain size measurements. In layer 7a the thickness of laminae is roughly equal to the size of the sampling box, but in the finely laminated layer 7b, the magnetic susceptibility sample reflects an averaged signal of several laminae, whereas the much smaller grain size sample is lithologically homogeneous. However, when samples from individual laminae are compared, the correlation between magnetic susceptibiliy and median grainsize becomes apparent. For example in the laminae of layer 7b we can see a direct correlation between the median grain size and $\chi$. The yellow brown laminae of eolian origin have very low $\chi$ (2-3 x $10^{-7}$ $m^3$/kg) and very fine median grain sizes (~7 μm), whereas the adjacent reddish laminae show significantly higher $\chi$ (~6 x $10^{-7}$ $m^3$/kg), coarser median grain sizes (~ 14 μm), and a bimodal grain size distribution.

In layers 7d, 9 and 11 the correlation between Md and $\chi_p$ is not as pronounced as in the upper part of the profile. The Md values are nearly constant, while $\chi_p$ shows distinct variations. The likely explanation is that the measured Md is not as sensitive to the intensity of the pedogenic process as $\chi_p$ is. The grain size decrease probably reaches a certain minimum threshold as the lower limit of the grain size measuring instrument is 0.7 μm. $\chi_p$, on the other hand, is extremely sensitive to changes in concentration of particles around 0.03 μm which cannot be detected by the grain size measurements. This conclusion is supported by the good inverse correlation between Md and the magnetic grain size of remance bearing magnetic minerals (ARM/SIRM). The ARM/SIRM which is sensitive to grain size changes in a comparable range to Md also shows a nearly constant progression throughout layers 7d-11 and increases significantly only in layer 6a, 7a and 13, 12.

## 2.4. Discussion

Results from the sediments in Kulna Cave suggest that pedogenic magnetic susceptibility ($\chi_p$) and grain size variations respond sensitively to the intensity of the pedogenetic process controlled by  principal soil forming factors: climate and vegetation. Figure 15 compares the $\chi_p$ record with the SPECMAP oxygen isotope record (Imbrie et al., 1984) and with the North Atlantic sea surface temperature data obtained from core K-708-1 at the latitude 50°N (Ruddiman, 1987[44]). The SPECMAP yields general climatic trends as it was created by averaging data from world-wide data. On the other hand, the single core temperature data come from a localized region which, due to the ocean atmosphere interactions, has a dominant impact on the climate of Central Europe (Bradley, 1999[4]).

The correlation between the Kulna Cave $\chi_p$ record and the established Late Pleistocene climatic records is based on limited age control for the Kulna sediments. Radiocarbon age dates are available for layers 6a (21 Ka) and 7a (39 and 46 Ka). The transition from fluvial sediments in layers 12 and 13 to loess like sediments in the overlying layers was interpreted by Valoch (1992)[57] as the climatic boundary between OIS 5e and OIS 5d. The ESR date of Rink et al. (1996)[42] for layer 7a (50 Ka) is in good agreement with the radiocarbon dates from this layer whereas the 69 Ka ESR date for layer 9 appears to be anomalously young. A total of 6 layers were dated by the ESR technique, out of which only two did not violate the principal of superposition. The possibility of a systematic error is thus also probable for the ages reported for layer 7a and 9. Therefore, we did not put any weight on the ESR dates. Age control over the central part of the profile (layers 7b to 11) is based on matching the $\chi_p$ record with the K-708-1 sea surface temperature data.

Beginning with the onset of the Last Glacial, the $\chi_p$ measured on loam and loess in Kulna (layers 6-11) shows an overall good agreement with the SPECMAP record and a very high degree of correlation with the North Atlantic sea surface temperatures. This correlation does not exist at the base of the section (layers 13, 12) because of a different lithological character of these sediments. Fluvial sediments deposited in layers 13 and 12 show a totally different mineral magnetic signature, yielding low $\chi_p$ values related to the change in source area of these deposits.

Based on our results and on the correlation of the $\chi_p$ record with the SPECMAP and the North Atlantic Sea Surface temperature records we propose the following reconstruction of the climatic conditions in the vicinity of the Moravian Karst: During the climatic optimum (OIS 5e) a stream flowing into the cave deposited sands and gravels that make up layers 12 and 13. The source of these sediments were Lower Carboniferous Greywackes which crop out only 2km north and upstream of the cave. After the stream ceased flowing into the cave, loams consisting of weathered loess together with limestone debris were being transported into the cave from the southern entrance. The heavy mineral assemblage of the loams suggests their origin as the granitoids of the Brno Massif (Kvitkova, 1999[26]), which crop out only 5 km west of the Cave. The sand and silt sized particles were transported mostly by the eastward blowing winds (Kukla, 1975[28]) and mantled the slopes of the Sloup Valley. Conditions at the beginning of the Last Glacial were mild leading to intensive pedogenesis of the sediments in the vicinity of the cave, prior to their redeposition into the entrance. The varying $\chi_p$ throughout layers 7d, 8a, 9 and 11 reflects the slight climatic oscillations during the OIS substages 5a, 5b, 5c, 5d (see Fig. 2-13). Climatic deterioration and lower productivity of pedogennic ferromagnetic minerals during OIS 4 lead to lower values of $\chi_p$ near the top of layer 7d and in layer 7c. Slight climatic improvement during the Pleniglacial (OIS 3) is reflected by more intensive pedogenesis during deposition of layer 7b. In fact several short lived episodes of significant climatic amelioration are represented by the three $\chi_p$ peaks in this layer. As the climatic conditions during the Late Glacial (OIS 2) were deteriorating the pedogenic activity was decreasing and windblown sediments were deposited directly into the cave entrance. At first the deposition of loam and loess alternated as reflected by the rapidly changing $\chi_p$ in layer 7a, but eventually (in layer 6a) loam deposition ceased all together. What limited pedogenesis occurred resulted in a narrow size distribution of grains around the SD/SP boundary (Eyre, 1997[16]) yielding extreme values of $F_c$ (Fig. 2-10).

Our Kulna record as well as the single core North Atlantic record show significantly more variability than the SPECMAP record. Also many other marine oxygen isotopic records as well as other marine proxies (e.g. foraminifera, debris content) point to significant higher degree of environmental change (Rudimann, 1987[44]; Heinrich 1988[21]) than a three-fold sequence (OIS 2,3,4) of the SPECMAP. Periods of relatively short-termed warm episodes during the OIS 4 are not restricted only to the deep sea record. Well-defined intervals of relatively mild climatic conditions during the middle and later parts of the Last Glacial have been identified in the oxygen isotope profiles from the Greenland ice sheet (Dansgaard et al., 1982[10]). The century to millenia scale "Dansgaard-Oeschger" interstadial events appear to have involved a shift in climate between warm and cold stages of around 7 °C. The close correspondence between the ocean and ice core records suggests that the atmosphere and ocean surface were a coupled system repeatedly undergoing a massive reorganisation. However, relating these major climatic changes to the terrestrial record has proven to be more difficult because a) the atmospheric circulation providing the link betweens oceans and continents is not yet fully understood, and b) many terrestrial proxies are not sufficiently sensitive to record such low amplitude and short lived climatic events (Lowe and Walker, 1997[31]).

Global Cirulation Models (GCM) (e.g. Wright et al., 1993[60]; Broecker et al., 1990[5]) yield more insight into the first question, attempting to establish dynamic links between atmospheric circulation and forcing mechanisms such as melting of ice sheets, circulation, vegetation cover or presence of dust in the atmosphere.

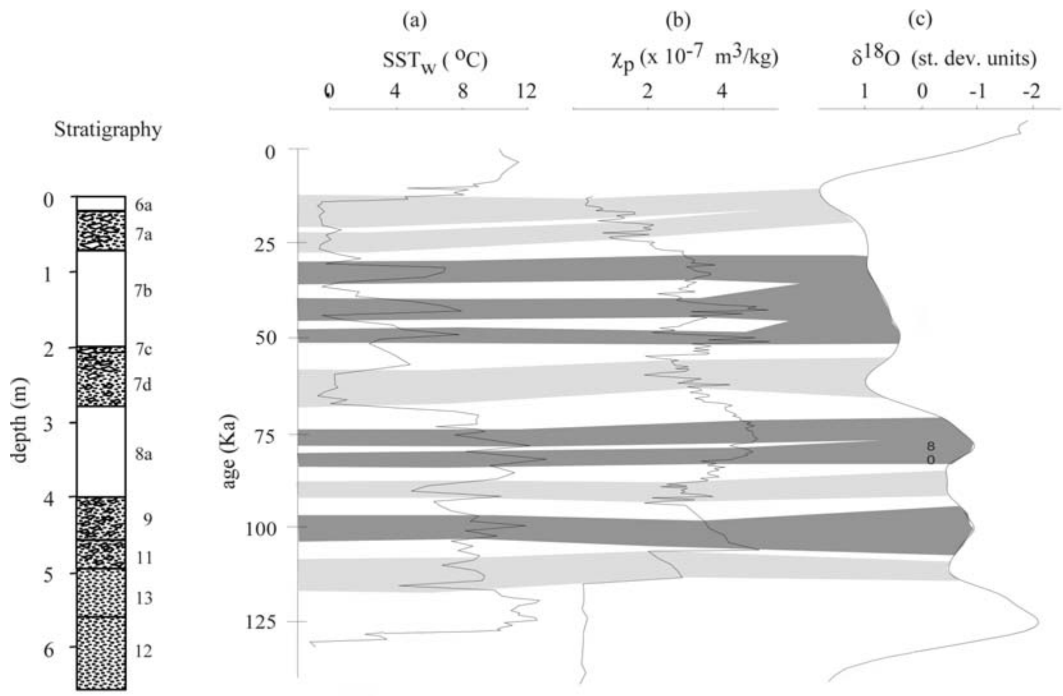



Fig. 2-13. Comparison of the North Atlantic sea surface temperature record with the pedogenic susceptibility measured in the sediments of Kulna Cave and the average deep-sea oxygen isotope record. Light-shaded bands indicate correlation of cold episodes; dark-shaded bands indicate correlation of warm episodes. (a) $SST_w$: winter sea surface temperature recorded from core K708-1 located approximately 50N, 24W (after Ruddiman 1987[44]); (b) $\chi_p$ pedogenic magnetic susceptibility measured through a composite profile in the entrance of the cave; (c) smoothed and stacked record of $d^{18}O$ variations obtained from five cores and measured on shallow-dwelling planktonic foraminifera (Imbrie et al. 1984[34]).

Recently, Ruddiman et al. (1989)[45] and Raymo et al. (1990)[41] discovered major feedback mechanisms between melting of ice sheets, fluxes of freshwater into the North Atlantic and the North Atlantic Deep Water (NADW) circulation. According to Huntley and Prentice (1993)[22] these processes are tied to the continental climate in Europe and directly affect the biomass both in the ocean and on the land.

Banerjee (1997)[3] approaches the question of recording short term changes in the terrestrial record and suggests that ferromagnetic minerals forming in soils typically respond faster to climatic change than most of the other proxies. However the paleoclimatic record from the loess/paleosol sequences is often lost due to later processes in particular due to diagenetic changes occurring in polygenetic soils (Oches and Banerjee, 1996[36]). One has to bear in mind that mineral magnetic properties reflect not only processes creating, but also destroying the ferromagnetic minerals. We suggest, that the $\chi_p$ measured in Kulna Cave records rapid climatic variations in OIS 3 because of the protective character of the cave. The soil weathered on the surface was transported into the cave where it was not exposed to the weathering agents. The cave thus fossilizes any impact acting on the sediment on the surface.

Our interpretation of the climatic conditons around Kulna, show an overall good agreement with that of Valoch (1988, 1992)[56][57]. The only major discrepancy exists in the interpretation of layer 7c. According to Valoch, the thin layer 7c originated during milder interstadial climate and proposes a correlation with OIS 5a (see Table 2-1). Our results, on the contrary, suggest that layer 7c and the top of layer 7d witness a cooling episode correlatable with OIS 4. Valoch's intepretation is based solely on large and small fauna, because gastropods and plant remains were found in a limited amount and could not contribute to the entire picture. This interpretation is however complicated by the possible multiple origin of the fossil material, yielding a biased information about the large fauna in the vicinity of the cave. In addition, newer research indicates that some species found in layer 7c (e.g. Cervus elaphus – red deer) have been found associating with a variety of environmental conditions ranging from oakwoods to full glacial tundra (Stuart, 1982[50]).

## 2.5. Conclusions

1. The results from Kulna suggest that cave sediments can yield a detailed climatic record of changing conditions during the Last Glacial/Interglacial. However, the transport, sedimentary and post-depositional processes in the cave environment are complex and lead to lithological and textural variations in the strata. Prior to attempting a climatic reconstruction, a high degree of understanding of the depositional history must be obtained.
2. We suggest that the pedogenic susceptibility $\chi_p$ measured in the Kulna Cave sediments is a better climatic proxy than $\chi$. $\chi_p$ is a sensitive indicator of the intensity of pedogenic weathering and thus records changing climatic conditions in the vicinity of the cave. $\chi_p$ correlates well with another proxy for pedogenetic weathering - the median grain size of the matrix of Kulna sediments.

3. The $\chi_p$ record from Kulna shows a general agreement with the SPECMAP record over the time period 15-110 Ka B.P.. The detailed structure of the $\chi_p$ record for this time interval is nearly identical to the sea surface temperature data from the North Atlantic. The high degree of correlation suggests that during the Last Glacial Stage Central European climate was strongly controlled by the sea-surface temperatures in the North Atlantic. Short-termed warmer events and perhaps higher precipitation over the mid-continent increased the intensity of pedogenetic weathering.

4. The proposed climatic interpretation is in a broad agreement with the interpretation of Valoch (1988[56], 1992[57]) which was based archeological, fyto- and zoopaleontological. The large amount of data collected from Kulna allows a multifaceted view of the conditions around the cave as well as of the processes that lead to sediment deposition. Kulna is the first site in the Czech Republic where one single profile yielded such a profusion of data. The amount of independent climatic information and high resolution of the record makes Kulna an extremely important site for studying the Late Pleistocene climate.

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## Chapter 3

# Historical Climatic Record from Flood Sediments Deposited in the Interior of Spirálka Cave, Czech Republic

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# Historical Climatic Record from Flood Sediments Deposited in the Interior of Spirálka Cave, Czech Republic

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## ABSTRACT

Magnetic susceptibility ($\chi$) was measured on more than four hundred samples collected from a five meter high section of fine grained sediments deposited during flood events in the interior of Spiralka Cave. Spiralka Cave is located in the northeastern portion of the Moravian Karst, Czech Republic. In the upper 1.5 meters of this profile, mineral magnetic ($\chi_{fe}$, ARM/SIRM, S-ratio, $\chi$(T)) and other non-magnetic measurements, heavy mineral concentration, loss on ignition and particle grain size) indicate that $\chi$ variations are controlled by the concentration of magnetite and by magnetic grain size, i.e. increased magnetic susceptibility results from increased concentration of coarse grained magnetite. A positive correlation with Ti and Zr concentrations in this part of the profile with our magnetic susceptibility record is a detrital signal responding to changing environmental conditions in the catchment area. Furthermore, a comparison of our susceptibility record to the record of winter temperature anomalies constructed from both instrumental and historical

records collected at the Klementinum Observatory in Prague shows a remarkable correlation. The most probable explanation of these correlations is that during years with warmer winters (positive winter temperature anomalies) and less snow cover the floods were less intensive but probably had access to larger tracts of cultivated land as agriculture tended to expand during these warmer periods. Cultivation of the land provided flood waters with greater access to coarser grained magnetite-like materials exposed by tilling of soils. Lower in the profile, interpreting the environmental significance of magnetic susceptibility variations is more complex as remobilization of iron has occurred. Nevertheless the magnetic susceptibility record when coupled with non-magnetic measurements can be shown to correlate to known environmental conditions present in Central Europe during the deposition of the lower portion of the profile.

Keywords: magnetic susceptibility, environmental magnetism, paleoclimate record, interior facies cave sediments

## 3.1. Introduction

### Cave Sediments and Paleoenvironmental Reconstructions

Cave sediments can be divided into entrance facies and interior facies. Entrance facies sediments either form directly in the cave (e.g. roof collapse) or are transported into the cave from the Earth's surface (wind, mass wasting, fluvial processes). The cave entrance acts as a trap in which the sediments are well protected from erosion. The entrance facies provides a wealth of material suitable for paleoclimatic/paleoenvironmental interpretation including its fossil and archeological artifacts (e.g. Sutcliffe, 1985[52]; Campy and Chaline, 1993[8], Goodfriend and Mitterer, 1993[21]; Horáček and Ložek, 1988[23] and Valoch, 1988[55]), clastic detritus (Kukla and Ložek, 1958[29]; Trudgill, 1985[54]; Campy et al., 1992[9] and Farrand, 2001[19]) and paleosols (e.g. Courty et al., 1991[11]; Wattez et al., 1989[58] and Maggi et al., 1991[33]). Interior facies cave sediments found deeper in the cave are genetically divided into chemogenic and clastic. The latter, which typically range from clays to gravel, are transported into the cave interior by fluvial processes while chemogenic sediments, i.e. speleothems, form as precipitates from drip-water. Chemogenic and clastic sediments can be found "sandwiched" between each other. This alteration is often climatically controlled; clastic material being deposited during periods of increased availability of sediments during cold periods and chemogenic sediments typically forming during periods of climatic amelioration (Lauritzen, 1995[31]). Many researchers have attempted to create climatic reconstructions from interior facies cave sediments. These attempts were based principally on oxygen isotope records obtained from datable stalagmites and flowstones (e.g. Schwarcz, 1989[46]; Lauritzen, 1990[30]; 1995[31]; Atkinson et al., 1986[1]; Sturchio et al., 1994[51]; Kashiwaya et al., 1991[24] and Gascoyne, 1992[20]).

## Environmental Magnetism and its Application to the Cave Environment

Environmental magnetism is the study of environmental (including paleoenvironmental) processes based on measurement of magnetic properties of the studied materials. Iron-containing minerals are extremely sensitive to environmental processes occurring on Earth's surface which modify their phase, concentration, and grain size. Laboratory measurements of the magnetic signature of sediment samples allow rapid characterization of ferromagnetic minerals and subsequent determination of environmental processes operating on the Earth's surface. Excellent reviews illustrating the wide range of environmental magnetic applications used to solve paleoclimatic/environmental problems can be found in e.g. Thompson and Oldfield (1986)[53], Maher and Thompson (1999)[35] and Evans and Heller (2003)[18].

Environmental magnetic techniques which have been successfully used for climatic reconstruction not only from lake sediments but also from loess deposits and deep sea sediments, have been scarcely applied in the cave environment. Only Ellwood et al. (1996)[17] and Šroubek et al. (1996, 2001)[49][50] attempted climatic reconstructions from mineral magnetic properties of clastic sediments deposited in caves. Ellwood and co-authors produced a paleoclimatic record for the Mediterranean region spanning the time interval between 9000 and 3500 years B.P. using magnetic susceptibility highs as indicators of climatic improvement and susceptibility lows suggesting deterioration. Šroubek et al. (2001)[50] used a similar approach for their climatic reconstruction obtained from the sediments of Kůlna Cave, Czech Republic, spanning the entire Last Glacial Stage. Similar to Ellwood, these authors measured magnetic susceptibility throughout several profiles in fine-grained clastic sediments found in the cave entrance. Variations in the magnetic susceptibility record were explained by growth of extremely fine-grained magnetite and/or maghemite formed by pedogenic processes during warm and humid climatic periods and redeposition into the cave.

## Importance of Paleoenvironmental Studies

Unforced natural variability of the climate system is very important on multi decadal and century time scales. Knowing both the spatial and temporal variations of climate change over the past several centuries remains a key to assessing possible human impact on the post-industrial climate. A more confident estimate of this impact is possible, if a faithful empirical description of climate variability over several past centuries can be obtained (Grove, 1988[22]; Kukla and Gavin, 1992[28]; Oeschger, 1992[41] and Rind and Overpeck, 1993[44]). Because widespread climate data determined from instrumentation are available for only about one century we must use available climate proxy indicators combined with instrumental records to obtain an empirical description of large scale climate variability during past centuries. Currently, there is a lack of available data not only from oceans and remote parts of the world but also from areas with a long history of inhabitance and richness of climatic information from instrumental measurements and historical written records (Bradley, 1985[3]).

To test the suitability of cave deposits to record a high resolution environmental magnetic signal we chose to study the sediments of Spirálka Cave. Spirálka Cave is located in the eastern part of the Czech Republic; an area where there are only few suitable sedimentary environments from which paleoclimatic reconstructions are possible. At our sampling locality in the cave, the deposits consist of young sediments with suitable material for radiocarbon dating, show only one depositional hiatus, no obvious erosional events and are of significant thickness. The deposition at the location is currently ongoing and thus the mineral magnetic environmental proxy record can be calibrated by comparison with instrumental data. The character of recent flood deposition can also be directly assessed by in-situ observations. Finally, the catchment of the Bílá Voda River providing sediments to the cave is lithologically uniform and thus the changing character of sediments should reflect climate variability and human impact rather than a changing source region.

## 3.2. Description of the studied area

## Description of the Spirálka Cave

Spirálka Cave is located in the northeastern segment of the Moravian Karst (Fig. 3-1), which is a part of a larger geomorphological complex of the Drahany Upland. The entrance of the cave is situated 471 m a.s.l. approximately 1 km south of the HolštejnVillage in the Hradský Žleb Valley. Spirálka Cave (Fig. 3-2) consists of a series of shafts totaling a depth of 40 m from which it continues both northward and eastward. The northern continuation is terminated by a sump that connects Spirálka Cave with Piková D ama (Queen of Spades) Cave.To the east a narrow passage named Tunelová (Tunnel) Passage continues for 60 m where it terminates in the Odtokový (Outflow) Sump. Spirálka Cave continues for five hundred meters past the Odtokový Sump where it ends in the Macošský Sump. Passages between the Odtokový and Macošský Sumps are the riverbed of the Bílá Voda (White Water) River. The overall length of the cave is 1691 m. Hydrology and Topography of the Studied Area

Spirálka Cave System belongs hydrographically to the Northern Part of the Moravian Karst, an area drained by the Bílá Voda River (Fig. 3-1), which is a

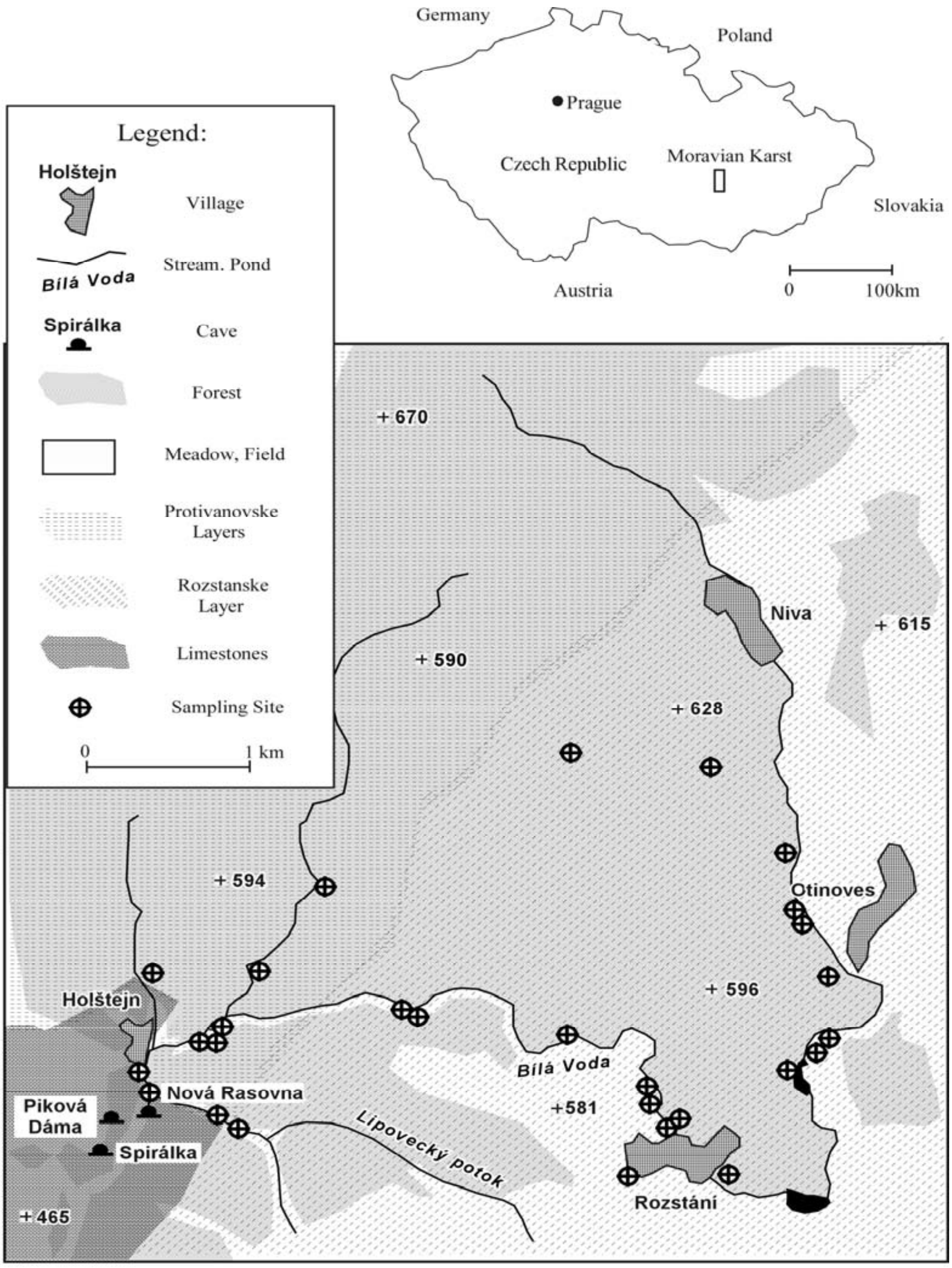


Fig. 3-1 Location of the Moravian Karst and superimposed topographical and geological plan showing the catchment of the Bílá Voda River which supplies sediments to the Spirálka Cave. Crosses indicate position of samples collected in the catchment area

tributary to the Punkva River. The Bílá Voda River flows 22.2 km on the surface until it reaches the Holštejn half-blind valley, where it sinks in the Nová Rasovna (Knacker's House) Cave (443 m a.s.l) and continues underground until it enters Spirálka Cave through Macošský Sump and several other minor sumps. The altitude of Bílá Voda in Spirálka Cave during normal-flow conditions is approximately 400m

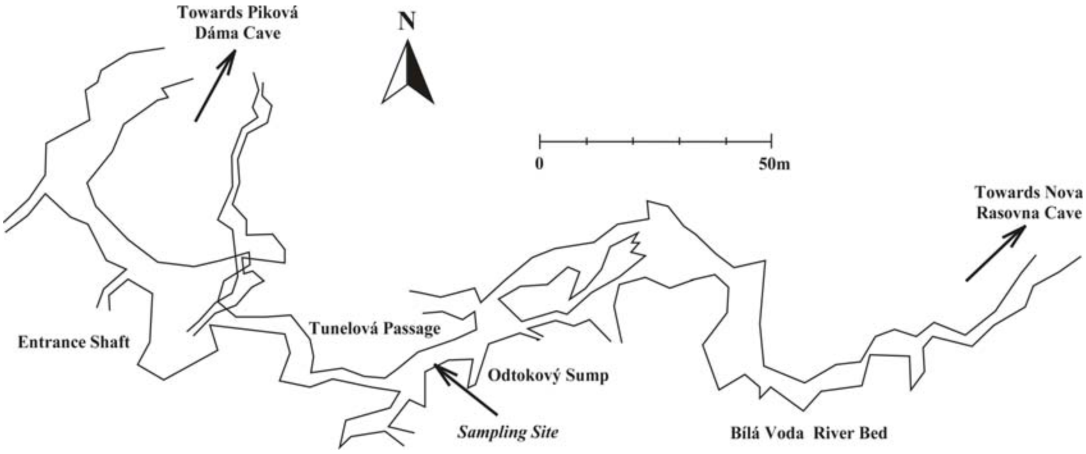


Fig. 3-2. Map of the Spirálka Cave showing the location of the sampled sedimentary profile

a.s.l. Throughout most of its lower flow, between the villages of Rozstání and Holštejn, the Bílá Voda Stream flows through a broad valley with gentle slopes. The valley's width ranges from 50 to 150 meters and its depth is no larger than 60 m. The valley floor, i.e. the floodplain of the stream, is covered by meadows whereas the slopes are typically forested. The undulating terrain of the upland in the vicinity of Holštejn is forested, higher upstream cultivated fields dominate. Fields also typically cover the gently sloping landscape along the stream north of the village of Rozstání, where the maximum elevation reaches 670 m.

### Clastic Sediments in the Spirálka Cave

Spirálka Cave is rich in clastic and chemogenic sediments most of which are recent, however older (probably Eemian) deposits are also present (Vít, 1998[56]). One of the more extensive sedimentary profiles includes a four meter thick gravel bed located in the vicinity of the Macošský Sump. The upper part of this profile was deposited between years 1845-1955 A.D. Our sampling site was located in the most impressive sedimentary section found in the cave, situated at the east end of the Tunelová Passage (Fig. 3-2) near the large flowstone formation “Varhany” (Organ). Here, floodwaters have been depositing layers of fine grained sediments reaching a total thickness of 5 m. The studied profile (Fig. 3-3) consists of sandy silt with layers and lenses of sand with maximum thickness of 10 cm. In the lower part are two 50 cm thick layers of gray green organic rich clayey silt. The top of the bottom layer is coated with a 2-3 mm thick flowstone layer. The base of the profile consists of coarse sands and fine gravels. Isolated wood fragments can be found throughout the profile, the upper part has abundant charcoal fragments. Presently, floodwaters reach this profile only during exceptionally high surface runoff. However, in the past, the conditions might have been different. Only in the last four decades that the cave has been known, speleologists have observed reactivation of certain passages and sumps. Vít (1998)[56] suggests that in the historical past the Odtokový Sump was clogged and a permanent underground lake was connecting the Tunelová Passage with parts of the cave lying south of the entrance shaft and continuing further into the Piková Dáma Cave.

### Ecological History of the Bílá Voda Catchment

The catchment of the Bílá Voda River was originally densely forested by beach and fir until the arrival of settlers in the 14th century (Nožička, 1957[39]). However, the first human impact was minimal. A 1694 Liechtenstein inventory describes the forest being in very good condition. By the beginning of early 18$^{th}$ century beach wood was intensively harvested for nearby smelters and by the year 1810 hardly any beech tree was left in the vicinity of Blansko (Materna, 1961[36]). The human induced forest calamities of 18$^{th}$ and 19$^{th}$ centuries were enhanced by forest fires and cyclones occasionally destroying up to 50% of all trees (Knies, 1902[26]). By the end of the 19$^{th}$ century the need for charcoal diminished and wood was used for heating and building instead. The fast growing spruce became the most wanted wood species and was preferentially planted. By the beginning of the 20$^{th}$ century it became the dominant tree in the area of the Moravian Karst and this situation lasts till present day.

## 3.3. Sampling and Laboratory Methods

### Sampling of the Cave Sediments

Samples for the mineral magnetic investigation, grain size analysis, trace element geochemistry and organic matter analysis were collected from a profile near the flowstone formation “Varhany”. In the upper part of the section (0-180 cm), large plexi-glass boxes (15 cm x 45 cm x 12 cm) were cut into a smoothed face of the profile. Each box, resampled in the lab, yielded a total of 15 by 3 tightly spaced cube samples (8 $cm^3$). We collected 5 boxes, placed vertically one above another, providing us with 225 samples from 75 horizons. Samples were taken from the interval 185 – 410 cm by pushing small sampling boxes (8 $cm^3$) directly into the smoothed face of the profile. In this fashion we collected a total of 180 samples from 90 horizons. Finally, the 28 bottom most samples from the depth interval 410 – 475 cm were collected from a manually drilled sedimentary core with a diameter of 2.7 cm.

## Mineral Magnetic Measurements on the Cave Sediments

Mass specific magnetic susceptibility ($\chi$) on all samples was measured using a Kappabridge KLY-2 instrument. The remaining mineral magnetic measurements were determined on one sample for each horizon. Paramagnetic ($\chi_{para}$) and ferromagnetic ($\chi_{fe}$) susceptibilities were distinguished from hysteresis loops measured on Variable Frequency Translation Balance (VFTB). In order to calculate the S-ratio, saturation isothermal remanent magnetization was acquired at 1T (SIRM, $IRM_{1000mT}$) and at 300 mT ($IRM_{300mT}$) using a Sapphire Instruments SI-6 pulse magnetizer. Anhysteretic remanent magnetization (ARM), acquired using the Sapphire Instruments SI-4 AF demagnetizer (biasing field 0.05 mT, alternating field 1T) was used to calculate the magnetic grain size dependant ratio ARM/SIRM. All remanence parameters (IRMs and ARM) were measured on a Schonstedt SSM-2 spinner magnetometer. Saturation magnetization ($J_s$) and hysteresis parameters were measured on a EG&G PAR Vibrating Sample Magnetometer (VSM) (maximum field 1T). Frequency dependence of magnetic susceptibility ($F_D$) was calculated using measurements taken on a Bartington Instruments MS-2 susceptibility meter at two different frequencies, 470 Hz and 4700 Hz. And lastly, to complete our investigation of the magnetic phases present in the samples, saturation magnetization versus temperature experiments were performed on 30 representative samples from throughout the profile. A modified Variable Field Translation Balance (VFTB) (50 mg specimens) was used to monitor the behavior of saturation magnetization from 20°C to 700°C and a Quantum Design Magnetic Properties Measurement System (MPMS) (100-200 mg specimens) instrument was used on samples subjected to low-temperature demagnetization (70$^o$ K to 200$^o$ K). These measurements allowed us to determine the Curie temperature and stoichiometry of magnetic mineral phases present. Collinson (1983)[10], O'Reilly (1984)[40] or Dunlop and Özdemir (1997)[16] present detail descriptions of measuring procedures and significance of measured parameters discussed above.

## Non-magnetic Measurements on the Cave Sediments

To complement magnetic measurements and tie the mineral magnetic data to other paleoclimatic indicators we determined several other characteristics of the sediments in Spirálka Cave. Grain size distributions (0.7-700 µm) were determined on 150 mg sub-samples using a 20-channel Leeds and Northrup MICROTRAC II laser particle size analyzer. The amount of organic matter present in samples was characterized by loss on ignition (LOI). The concentration of trace elements, namely Zr, Ti and Fe, was measured on an XRF spectrophotometer EX310 from Jordan Valley AR, Inc. Details of these methods and their significance can be found in Robertson et al. (2004)[45] or Pansu and Gantheyron (2006)[43], for example.

Wood and charcoal fragments ranging in size from several milligrams to tens of grams were collected throughout the sedimentary profile in Spirálka Cave. A total of approximately 30 samples were collected from which we selected 9 specimens from 4 different depth horizons which were sent to Beta Analytic Laboratory, Florida for radiocarbon dating. Four samples had sufficient mass to be dated by the conventional method; the remaining five samples were analyzed using the accelerator mass spectrometry (AMS) technique.

## Sampling and Analysis of the Surface Sediments

Samples from the catchment of Bílá Voda Stream were gathered in order to compare their mineral magnetic properties with those of the cave deposits. A total of 91 hand samples were collected along a 20 km stretch of the stream from the four sedimentary environments providing a potential source for the cave deposits. We obtained 46 samples of floodplain deposits, 9 samples of field topsoil, 16 samples of forest topsoil, and 20 samples of weathered bedrock. The coordinates of each sample were located using a GPS system; the locations are shown in Fig. 3-1. The hand samples were re-sampled in the lab into 8 $cm^3$ specimens, on which mass specific susceptibility and its frequency dependence were measured.

## 3.4. Laboratory Results

### Results from Mineral Magnetic Measurements on Spirálka Cave Sediments

*Magnetic Susceptibility Results*

The values of magnetic susceptibility ($\chi$) of the three samples collected from each horizon of the cave profile showed consistent values and on the average did not differ by more than 5 percent. Therefore, the magnetic susceptibility values shown in Fig. 3-3a represent an average susceptibility determined from three samples from the same horizon. The record of magnetic susceptibility ($\chi$) shows a well-defined oscillatory pattern in the upper 1.5 m of the profile with the most notable oscillations occurring in the upper most 0.5 m. The values of $\chi$ vary between 10 and 16 x $10^{-8}$ $m^3/kg$ in this depth interval. Lower in the section the $\chi$ record shows two prominent zones between 170 -190 cm and 200-225 cm where $\chi$ values rapidly drop to between 6 and 8 x $10^{-8}$ $m^3/kg$. Below 225 cm $\chi$ values decrease more or less linearly to a low value of 5 x $10^{-8}$ $m^3/kg$ at 390 cm with the exception of two zones (285-305 cm and 320-390 cm) where again there appears to be a rapid decrease in $\chi$ values. At 390 cm depth $\chi$ begins to rise steeply with several rapid oscillations that have magnitudes up to 6 x $10^{-8}$ $m^3/kg$. The highest measured values of $\chi$ in the entire section are at its base and reach 20 x $10^{-8}$ $m^3/kg$.

In the upper 320 cm of the profile, ferromagnetic susceptibility ($\chi_{fe}$) values (Fig. 3-3b), in general, mimic the patterns seen in the total magnetic susceptibility

($\chi$) record as paramagnetic susceptibility is rather constant (~ 6 x $10^{-8}$ $m^3/kg$) in these sediments. However, between 320 and 390 cm, the magnetic susceptibility signal of the sediments is almost entirely paramagnetic and therefore $\chi_{fe}$ signal in these sediments is minor. Below 390 cm, the sediments again show a $\chi_{fe}$ signal. In fact, two prominent peaks centered at 400 cm and 450 cm have magnitudes that are similar to the samples measured from upper portion of the section.

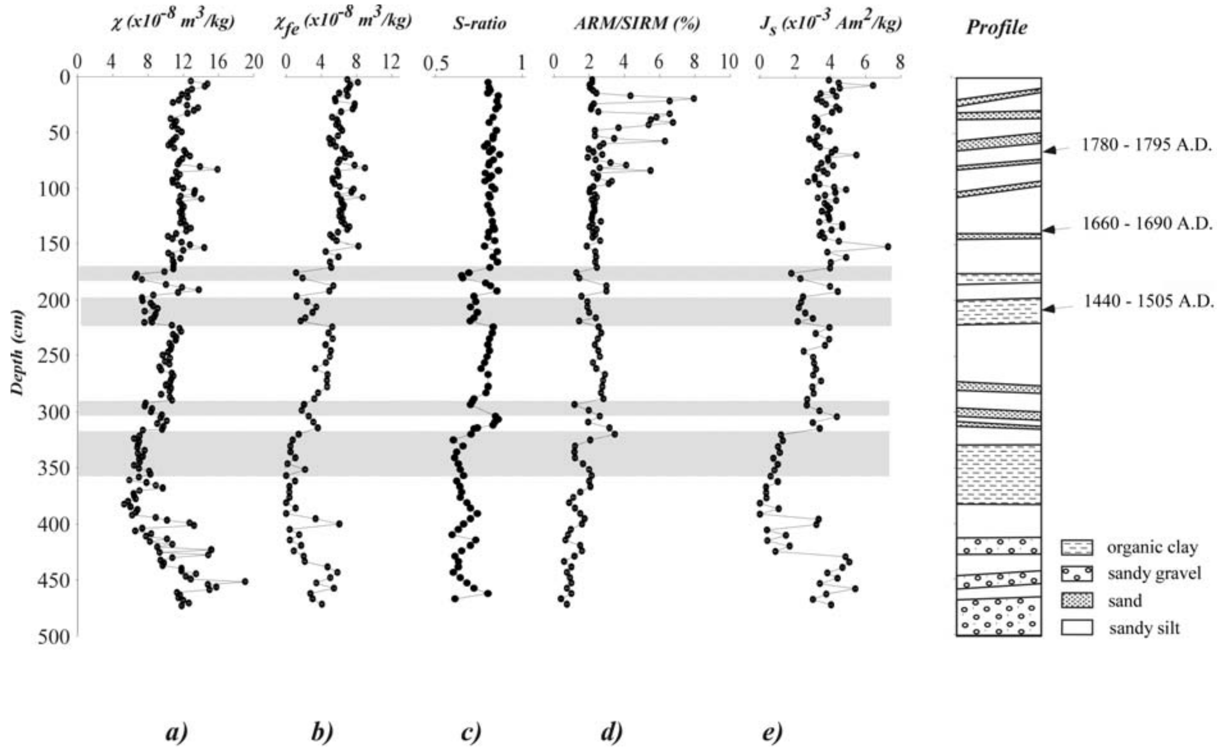


Fig. 3-3 Mineral magnetic parameters measured throughout the sedimentary profile in Spirálka Cave – a) magnetic susceptibility, b) ferromagnetic susceptibility, c) S-ratio (see text), d) ratio of anhysteretic remanent magnetization and saturation isothermal magnetization and e) saturation magnetization. Highlighted are the zones with anomalously low magnetic susceptibility (see text). Also shown is a sketch of the sampled profile showing position of the radiocarbon dated charcoal and wood samples.

*Other Mineral Magnetic Results*

The S-ratio (Fig. 3-3c) of the sediments in the upper 320 cm of the profile in Spirálka Cave is fairly constant with values ranging between 0.8 and 0.85 with exceptions occurring at depth intervals between 170 and 190 cm, 200 and 225 cm, and 285 and 305 cm. Below the 320 cm depth, the S-ratio record drops to values between 0.6-0.65 with little variation downward.

The record of the magnetic grain size parameter ARM/SIRM (Fig. 3-3d) is fairly constant throughout the upper 350 cm of the profile. However, in the uppermost meter of the profile there are four distinct horizons where the ARM/SIRM values increase indicating a relative *decrease* in the magnetic grain size in these horizons. These zones of magnetic grain size decrease correlate with lows in the $\chi$ and $\chi_{fe}$ records (Fig. 3-3a,b). On the other hand, rapid drops in $\chi$ and $\chi_{fe}$ occurring in horizons between 170 and 390 cm, appear to correlate with grain size *increase* (decrease in ARM/SIRM values). Finally, in the lowermost portion values of ARM/SIRM decrease to the lowest values in the profile.

The saturation magnetization ($J_s$, Fig. 3-3e) of these sediments varies between 0 to 7 x $10^{-3}$ $Am^2/kg$ showing variations that are positively correlated to variations seen in both the $\chi$ and $\chi_{fe}$ records. The detail of correlation between the three records is quite striking.

Frequency dependence of magnetic susceptibility (not shown) is generally fairly low with values between 3-5 %. In general, variations in frequency dependence of magnetic susceptibility down profile mimic the variations seen in the $\chi$ and $\chi_{fe}$ records, i.e. where $\chi$ and $\chi_{fe}$ decrease frequency dependence decreases and vice versa and the smallest values of frequency dependence occur at the same horizons where $\chi$ and $\chi_{fe}$ are the smallest.

Lastly, several important observations can be made based on the data described above. Firstly, the data suggest that the susceptibility variations are caused by changes in the content of ferromagnetic minerals as $\chi_{fe}$ record mirrors the $\chi$ record. The paramagnetic contribution to the susceptibility record is more or less constant throughout the upper 320 cm of the profile becoming dominate contributor to the susceptibility record between 320 and 390 cm. Secondly, the ferromagnetic susceptibility variations are caused by changes in the concentration of the ferromagnetic content as the $\chi_{fe}$ record correlates well with the record of saturation magnetization ($J_s$). However, the concentration variations are not the sole factor influencing the magnetic susceptibility changes. The values of S-ratio show several significant drops suggesting that the magnetic mineralogy is not uniform throughout the studied section. Similarly, the ARM/SIRM record shows good correlation with the $\chi$ record suggesting that the magnetic grain size also influences the $\chi$ record especially in the upper meter of the profile. In the *upper meter*, $\chi$ *increase* is mimicked by an *increase in the average magnetic grain size* and vice versa. Similarly, in the *lower meter* of the profile the overall *increasing trend in $\chi$* is also followed by an *increase in average magnetic grain size*. However, in the *four intervals between 170 and 390 cm*, the rapid decrease in $\chi$ values appear correlated to a similar decrease in the ARM/SIRM values suggest the *drops in $\chi$* are accompanied by an *increase in the average magnetic grain size* and not a decrease, but then again S-ratio values also suggest a change in mineralogy.

### *Results from Additional Mineral Magnetic Measurements*

The mineral magnetic parameters described above yield only qualitative information about the magnetic properties of the cave sediments and cannot be independently interpreted since a non-linear combination of mineral magnetic properties typically affects the measured parameters and inter-parametric ratios. Further mineral magnetic measurements are necessary to distinguish the degree to which $\chi$ is influenced by changes in concentration of magnetic minerals, by grain size changes, and by variations in magnetic mineralogy.

Based on the measurements of the S-ratio the sediments can be divided into two groups. Samples with higher S-ratios between 0.80 - 0.85 have steep IRM acquisition curves (not shown) and saturate typically by 300 mT. These IRM acquisition curves are characteristic of minerals with low magnetic coercivity and are labeled “magnetite” type minerals. On the other hand, samples from horizons where the S-ratio values of the sediments show significant drops to 0.60 - 0.65 have IRM acquisition curves (not shown) which are less steep and which do not saturate by the maximum field of 900 mT. Such IRM acquisition curves are characteristic of minerals with higher magnetic coercivity, typically associated with “hematite” type minerals (Dunlop, 1972[15]).

Measurements of saturation magnetization as a function of increasing temperature on the “magnetite” type samples (Fig. 3-4a) show a gradual decrease towards 580°C. Above this temperature the saturation magnetization is virtually zero. The shape of the high temperature magnetization curve suggests that mostly multidomain magnetite and/or maghemite are present in the “magnetite” type samples. The presence of multidomain magnetite in these samples is also confirmed

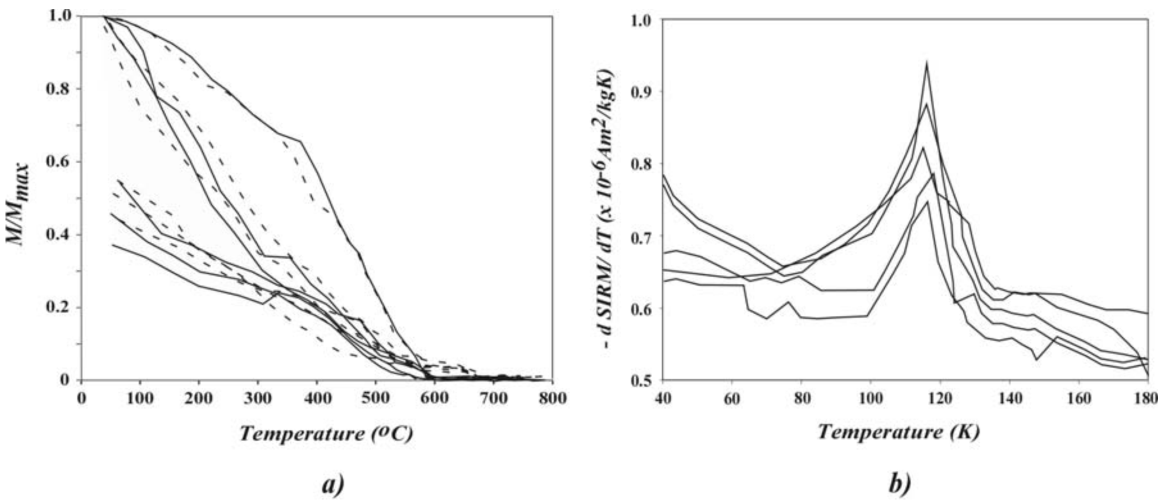


Fig. 3-4. Measurement of saturation magnetization with changing temperature a) from 0 to 800 °C, where solid lines represent “ magnetite” type minerals showing a Curie temperature around 580 °C and dashed lines represent “hematite” type minerals loosing all magnetization around 690°C; b) negative first derivative of saturation remanent magnetization over temperature as a function of increasing temperature for five representative “magnetite” type samples from Spirálka Cave. The peak in the curves indicates position of the Verwey transition

by the position of the Verwey transition (Özdemir et al., 1993[42]) which appears as a rapid drop in remanent magnetization observed at low temperatures (Fig. 3-4b). Magnetite present in the Spirálka sediments is mostly non-stoichiometric, as suggested by the position of the Verwey transition around 115 K (Kletetschka and Banerjee, 1995[25]). Measurement of hysteresis parameters show that most of the grains (Day plot, not shown, Day et al. 1977[13]) fall in the pseudo single domain region which is in a good agreement with the results obtained from measurements of the inter-parametric ratio ARM/SIRM. The samples yield values of ARM/SIRM ratio around 50x $10^{-5}$ m/A, which according to experiments of Maher (1988[34]) correspond to grain sizes around 0.1 μm. Finally, low values of the ARM/SIRM ratio as well as low values of the frequency dependence of $F_D$ between 3-5 % (e.g. Dearing et al., 1996[14]) suggest that the content of superparamagnetic (SP) grains in the Spirálka Cave sediments is minimal.

Samples with a higher content of "hematite" type minerals show different mineral magnetic behavior. Thermomagnetic curves (Fig. 3-4a) show a gradual decrease towards 580 °C; however magnetization is still present above this temperature. This result suggests that hematite, with a Curie temperature of 680 °C, is also present in the samples.

Finally, the difference in mineralogy throughout the sedimentary section is well demonstrated by a plot of SIRM values against the S-ratio (Fig. 3-5) (Snowball, 1993[48]). The trend from high SIRM and S-ratio values toward low values of both parameters is characteristic of the trend from magnetite dominated samples towards samples with decreasing magnetite content and increasing hematite content. In the four above discussed intervals (170-190, 200-225, 285-305 and 320-390 cm) the decreasing values of both SIRM and S-ratio are apparent. The above described mineral magnetic results together with the variable character of the $J_s$ record suggest that in these intervals the magnetic grain size decreases and that this decrease is accompanied by mineralogical changes, in particular a drop in the magnetite content and a relative increase in the hematite content.

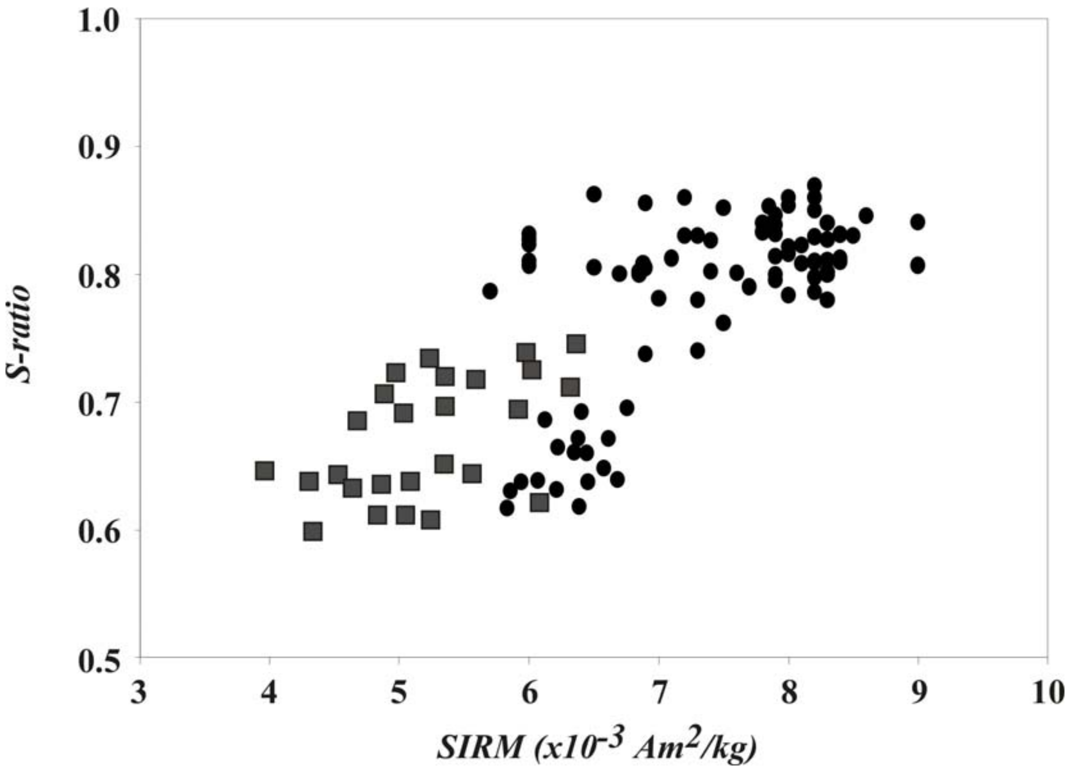


Fig. 3-5. Plot of S-ratio versus saturation isothermal remanent magnetization for representative samples from the Spirálka Cave characterizing their magnetic grain size. Squares indicates samples which showed anomalously low magnetic susceptibility

## Results from Non-magnetic Measurements on Spirálka Cave Sediments

### *Sediment Grain Size*

Median grain size (Fig. 3-6b) characterizing the overall grain size of the samples varied from 20 to 300 μm throughout the sedimentary profile in the Spirálka Cave. In the upper 150 cm of the profile the median grain size shows rapid variations between 60 and 200 μm, a certain degree of negative correlation with the $\chi$ record (Fig. 3-6a) and no correlation with the ARM/SIRM record (Fig. 3-3d). Deeper in the section between 170 and 390 cm the median grain size varies on a much smaller scale and shows an overall decrease from 70 to 30 μm. The large variations in grain size the dominant the upper 150 cm of the profile are missing in this interval. Finally, in the very bottom part of the section the median grain size shows once again a rapid gradual increase to values reaching 300 μm and a good correlation with the $\chi$ record.

### *Organic Matter Content*

Loss on ignition (LOI, Fig. 3-6c) measured throughout the sedimentary profile in Spirálka yields generally low values between 3-8 %. The only exceptions are two intervals at depths of 170-195 and 200-225 cm, where the LOI values increase to more than 10%. In these intervals the otherwise dark brown sediments turn lighter in color and are characterized by a distinct odor. In the upper part of the profile the LOI values are nearly constant around 5%, at the base of the profile where the content of gravel is increasing the LOI values drop to the minimum values of 3 %.

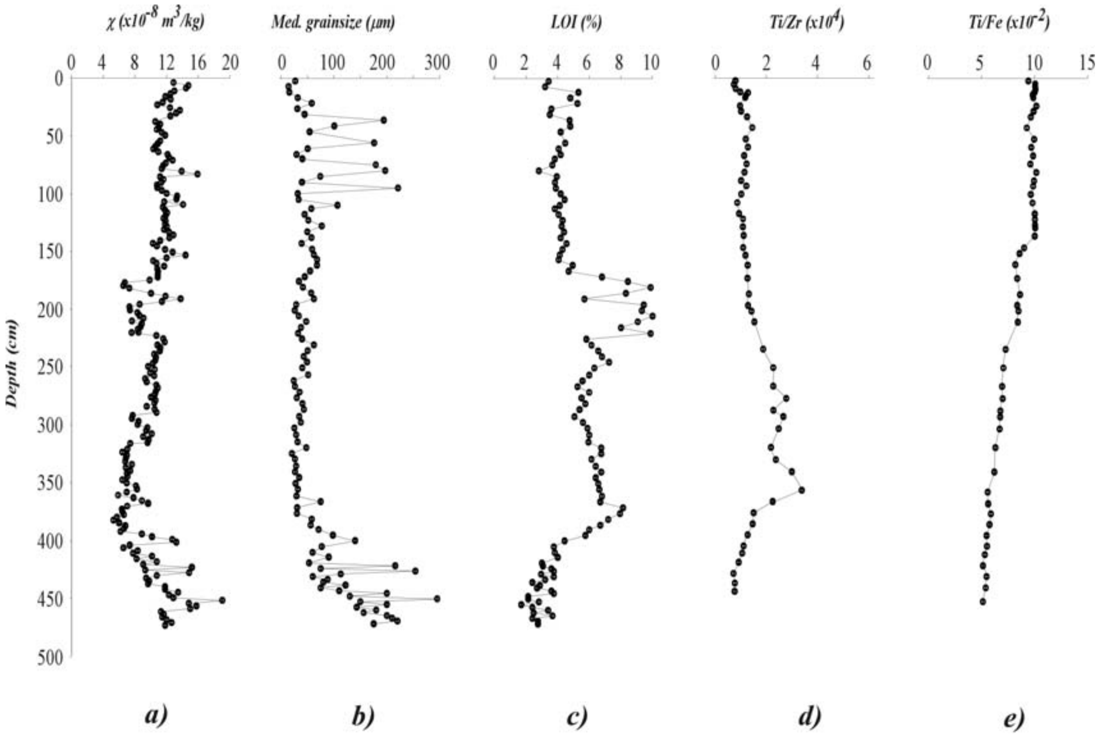


Fig. 3-6. Comparison of non-magnetic parameters with the record of ferromagnetic susceptibility measured throughout the sedimentary profile in the Spirálka Cave – a) ferromagnetic susceptibility, b) sediments median grainsize, c) loss on ignition, d) Ti/Zr concentration ratio e) Ti/Fe concentration ration.

### *Heavy Mineral Content*

Trace element geochemistry was utilized to detect whether post-depositional dissolution of magnetic minerals has affected the magnitude of magnetic properties.

Ti and Zr occur in heavy minerals and are immobile under most post-depositional conditions (Winchester and Floyd, 1977[59]). Variations in Ti and Zr therefore provide proxies for detrital heavy mineral content at the time of deposition and changes in the Ti/Zr ratio may indicate changes in the source of detritus (Muhs et al., 1990[37]; 1995[38]). In contrast, magnetic Fe oxide minerals, which represent nearly 100 % of the total Fe in the samples, are relatively mobile under common post-depositional conditions. For these reasons, comparison between magnetic properties and Fe, Ti and Zr content illustrate both the detrital magnetic signal and post-depositional alteration. The Ti/Zr record (Fig. 6d) shows a gradual rise from values around $1x10^4$ to $4 \times 10^4$ at a depth of about 380 cm, where a sharp drop occurs. The Ti/Fe (Fig. 3-6e) ratio shows an opposite general trend, i.e. a gradual decrease from the top of the profile ($10 \times 10^{-2}$) to approximately $5 \times 10^{-2}$ at the base.

## Results from the Spirálka Cave Catchment Area

$\chi$ and $F_D$ data were determined from materials cropping out in the catchment of the Bílá Voda River in order to determine the source of the sediments in Spirálka Cave and to investigate how environmental processes affecting the catchment were reflected in the cave sediments. The data presented in Fig. 3-7a show that the majority of the samples collected from the floodplain (35 out of 46 samples) yield $\chi$ values between 8 and $17 \times 10^{-8}$ $m^3/kg$ and $F_D$ values between 3 and 6 %. When compared with $\chi$ and $F_D$ data from the cave sediments, the floodplain samples show a much larger range yet the cave sediments appear to be in the central part of this range. On the other hand the $\chi$ and $F_D$ data measured on samples from both field

(Fig. 3-7b) as well as forest topsoil (Fig. 3-7c) show significantly higher values of $\chi$ and $F_D$ than the cave sediment samples. In the case of the field topsoil the majority of the samples (7 out of 9 samples) fall in the range between 24 and 28 x $10^{-8}$ $m^3/kg$ for $\chi$ and 4 to 8 % for $F_D$. The majority of the samples collected from the forest topsoil (9 out of 16 samples) yield values of $\chi$ between 17 to 25 x $10^{-8}$ $m^3/kg$ and values of $F_D$ between 5 to 8 %. $\chi$ and $F_D$ measurements of samples collected from the greywacke bedrock of the Bílá Voda River catchment are quite different. Both the $\chi$ values as well as $F_D$ values of these samples are quite similar, yet slightly lower than the cave sediment samples (Fig. 3-7d). The majority of the bedrock samples (17 out of 20 samples) fall into two ranges. One range is between 6-8 x $10^{-8}$ $m^3/kg$ for $\chi$ and between 2-5 % for $F_D$. The other range yields $\chi$ values between 11 to 13 x $10^{-8}$ $m^3/kg$ and $F_D$ between 2 to 3 %. Interestingly, the first range yields values of $\chi$ which are lower than those from the cave sediments, yet the values of $F_D$ are comparable. On the other hand the second range contains samples which have similar values of $\chi$ to the cave sediments, yet their $F_D$ values are somewhat lower. Comparison of $\chi$ and $F_D$ data from the cave and its possible sources suggests that cave sediments in Spirálka are most probably re-deposited sediments of the Bílá Voda River floodplain. The transport and deposition processes into the cave provided additional sediment filtering which resulted in narrowing the range of $\chi$ and $F_D$. Sediments of the floodplain yield $\chi$ and $F_D$ values which appear to be an average between those of the soil on the one hand and of the bedrock on the other. The perception of the floodplain sediments as a mixture of material provided by these environments is feasible and is supported by our data as sorting and oxidation of primarily one source is less probable because coarser grained sediments coming from different sources are observed at other sites in the cave.

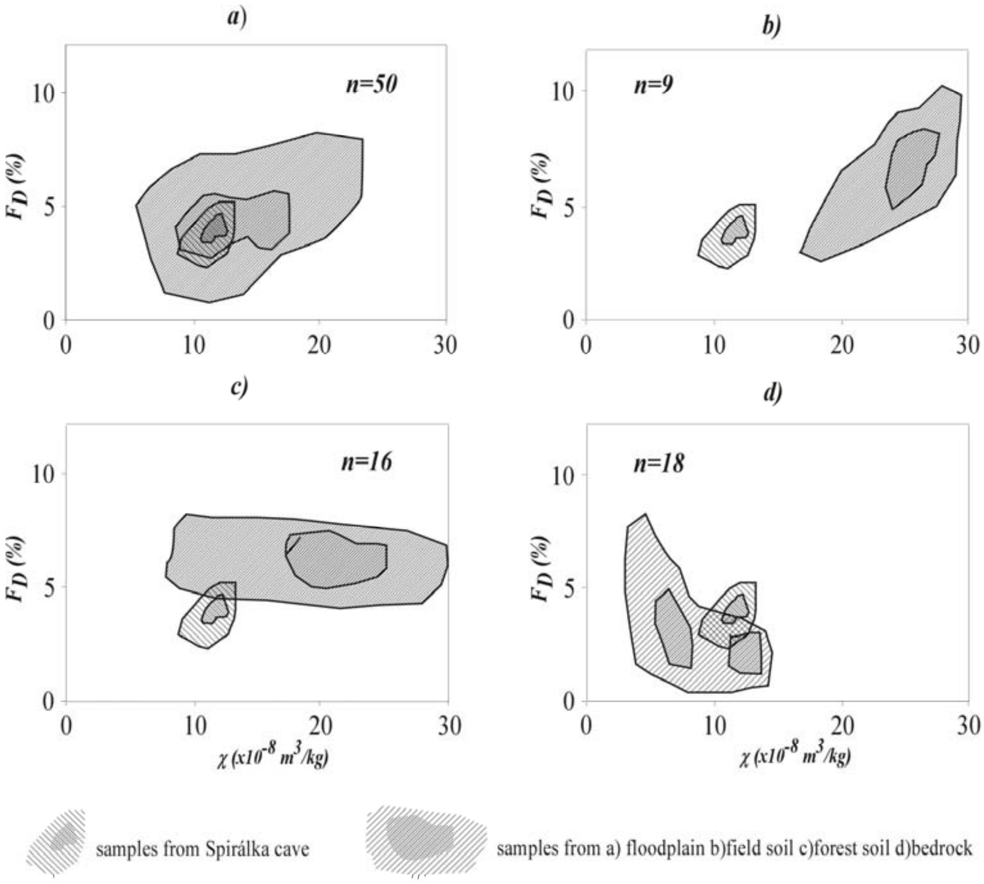


Fig. 3-7. Comparison of two magnetic parameters (magnetic susceptibility and frequency dependence of magnetic susceptibility) measured on samples collected in the Spirálka Cave and in different environments of the catchment of the Bílá Voda River : a) floodplain, b) field, c) forest and d) rock outcrops). The direction of hatching indicates the environment (see legend) and its density the amount of samples which fall in the given area of the plot (dense hatching – 66 % of samples, thin hatching – 95 % of samples). Finally, N indicates the number of samples analyzed in the given environment.

### Age of the Spirálka Cave Sediments

Conventional radiocarbon ages based on the ratio of $^{14}C/^{12}C$ from the wood and charcoal samples collected from the Spirálka profile are listed in Table 3-1 including their one sigma standard deviation. Since the relative concentration of radiocarbon in global reservoirs varies significantly over time, the age determined directly from the $^{14}C/^{12}C$ ratio (the radiocarbon age) has to be corrected by comparing the radiocarbon scale with other scales that use independent dating techniques. In this manner the radiocarbon time scale is calibrated against a chronology of calendar years obtained from tree-rings, varves or uranium series methods (e.g. Vogel et al., 1993[57]). Calendar ages based on the Pretoria Calibration Procedure (Vogel et al., 1993[57]) are also shown in Table 3-1.. In many cases, a single conventional radiocarbon age resulted in several calendar year ages due to the young age of the charcoals.

Comparing the one sigma age intervals from Table 3-1 to the depths of the dated samples and using the principle of superposition, an age model for our profile was developed. Data from the three samples collected 70 cm below the present day surface allow for only two possible age intervals, namely 1650-1670 A.D. and 1780-1795 A.D. The near present day age (1945-1950 A.D.) is very improbable as the cave was discovered in the year 1958 and the current depositional horizon was marked by the original cave discoverers. Since that time only a few centimeters of sediment have been deposited. The exact depositional rate and regularity of the

| Depth (cm) | Sample Name | Conventional Radiocarbon Age (B.P.) | Intercept with calibration curve (years A.D.) | Age intervals of one sigma calibrated results (years A.D.) | Age (years A.D.) | Sed. Rate (cm/100 yr) |
|---|---|---|---|---|---|---|
| 70 | 97SU1 | 150±40 | 1680, 1740, 1805, 1930 | 1670-1820, 1835-1880, 1915-1950 | 1780-1795 | 35 |
| 70 | DSESU | 150±40 | 1680, 1740, 1805, 1930 | 1670-1820, 1835-1880, 1915-1950 | | |
| 70 | 97SU2 | 240±40 | 1660 | 1650-1670, 1780-1795, 1945-1950 | | |
| 140 | U14 | 190±60 | 1675, 1775, 1800, 1945 | 1655-1695, 1725-1815, 1920-1950 | 1660-1690 | 43 |
| 140 | 97SU4 | 190±50 | 1675, 1775, 1800, 1945 | 1660-1690, 1735-1815, 1925-1950 | | |
| 220 | 97SD1 | 130±50 | 1695, 1725, 1815, 1920 | 1675-1770, 1800-1940 | 1440-1505 | 40 |
| 220 | 97SU5 | 410±50 | 1460 | 1440-1505, 1595-1620 | | |
| 220 | 95D2 | 440±80 | 1450 | 1420-1505, 1595-1620 | | |
| 300 | 95U2 | 220±60 | 1665 | 1650-1680, 1745-1805, 1935-1950 | | |

Table 3-1. Summary of radiocarbon dating on wood and charcoal samples from Spirálka Cave. Indicated is depth of sample below present profile surface, sample name, conventional radiocarbon age with one sigma standard deviation, intercept with Pretoria calibration curve (calendar age), one sigma interval of calendar age, most probable calendar age (see text) and average sedimentation rate for given interval.

sedimentary process since the discovery of the cave cannot be exactly estimated as there was not any instrumental monitoring of the sedimentation process installed in the cave. A comparison with age data from the depth of 140 cm suggests that the age interval 1780–1795 A.D. is more probable probable for the 70 cm depth. If the older age interval (1650-1670 A.D.) of the horizon 70 cm was correct then sediments between the depth of 70 and 140 cm would have to be deposited in less than 15 years which is extremely unlikely. There is no evidence for such rapid sedimentation rate.

Out of the two possible age intervals (1660-1690 A.D. and 1735-1815 A.D.) of the 140 cm horizon only the older is feasible as dendrochronological dating on a P. alba wood sample (R. Réh, personal comm.) confirms an age date prior to 1700 A.D. Two of the samples (97SU5 and 95D2) from the depth horizon 220 cm yield consistent age data yet the remaining third sample 97SD1 has an age which is not compatible with the previous two. We assume that this result is in error as it is not compatible with the other two dated and that the age of the horizon 220 cm is either between the years 1595-1620 A.D. or more probably between the years 1440-1505 A.D.

Finally, the age of sample 95U2 from the bottom most horizon 300 cm below the present day surface is obviously erroneous as even the maximum possible age of this sample (1650 A.D.) is younger than the minimum possible age (1620 A.D.) of the overlying dated horizon.

Table 3-1 also presents our preferred radiocarbon ages base on the above mentioned arguments and the corresponding approximate average sedimentation rates between dated horizons. Other interpretations of age based on the multiple dates listed in Table 1 rather than our preferred model are naturally possible, but yield fairly unrealistic sedimentation rates.

### 3.5. Discussion

The upper most 1.5 m of the profile (Fig. 3-8) is the only part of the section where sediments did not undergo major post-depositional alteration as the Ti/Fe and Ti/Zr remain fairly constant (Fig. 3-6d,e). A comparison of the $\chi_{fe}$ and ARM/SIRM records (Fig. 3-8a,b) indicates that larger magnetic grain size (low ARM/SIRM values) typically corresponds to larger $\chi_{fe}$ values suggesting that the magnetic grains are predominantly multi-domain (Maher, 1988[34]) while small magnetic grain size (high ARM/SIRM values) is associated with low $\chi_{fe}$ values. Therefore, the oscillatory swings in $\chi_{fe}$ are caused by varying input of "magnetite type" minerals (concentration) and changes in their grain size. The fact that $\chi_{fe}$ and $J_s$ (Fig. 3-8c) show a close positive correlation to the Ti and Zr records (Fig. 3-8d,e) from this portion of the profile suggests that both the variation in magnetite concentration as well as grain size are directly controlled by the rate of erosion in the catchment of the Bílá Voda River.

Comparison of the cave sediment properties with the different catchment environments (including the floodplain, fields and forests) allowed us to place the cave sediment record in historical context of environmental processes occurring in the catchment of the Bílá Voda River. We found that sediments in the cave have very similar properties to sediments of the floodplain and were therefore most probably re-deposited from this environment into the cave during high stands of the Bílá Voda River.

However, water flow conditions in the cave are indeed extremely variable. Just in the last several decades after the cave discovery, significant changes including relocation of the streambed and removal of vast amounts of sediments have been observed. In addition, variable water flow in the Moravian Karst during the last

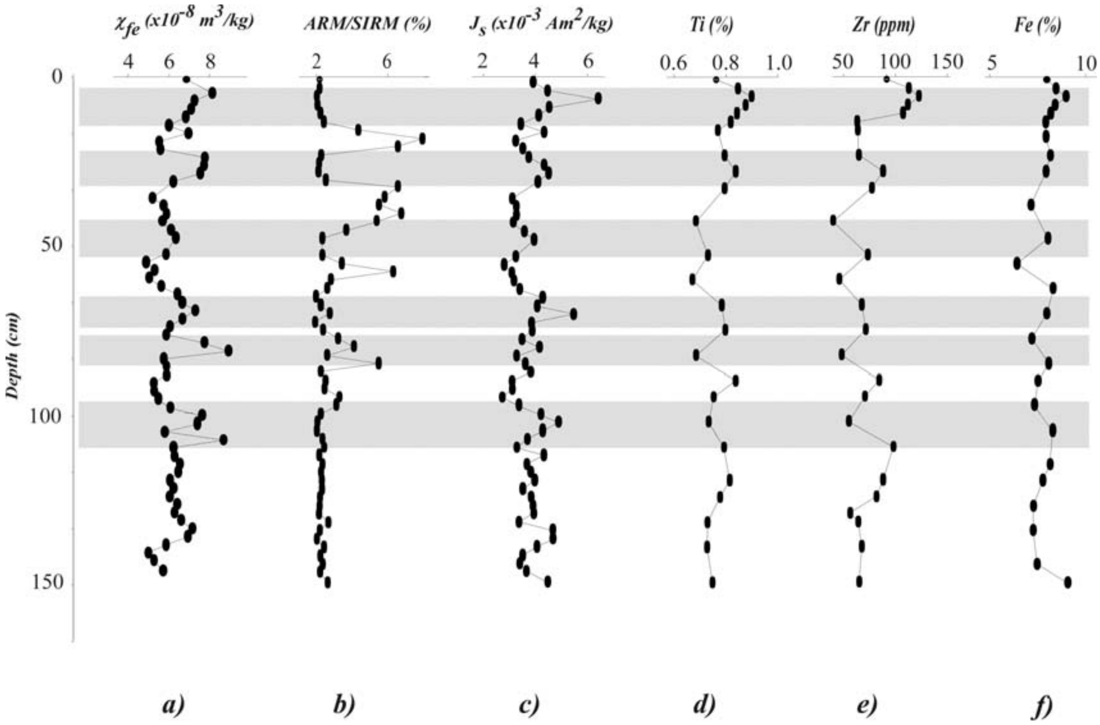


Fig. 3-8. Comparison of parameters providing paleoenvironmental information from the most important upper 150 cm of the sedimentary profile in Spirálka Cave: a)ferromagnetic susceptibility, b) ratio of anhysteretic remanent magnetization and saturation isothermal magnetization (ferromagnetic grain size), c) saturation magnetization, d) Ti concentration, e) Zr concentration and f) Fe concentration. Highlighted are environmentally significant zones in the ferromagnetic susceptibility record and their reflection in the other displayed parameters.

centuries appears quite common. Burkhardt (1952)[7] concludes from historical written records that flooding in the Punkva River cave system was very common in the past and much more severe than in the present years. Knies (1911)[27] also collected historical information on flooding in the Holštejn valley since early 1800's and suggests that flooding in the last century was more extensive than the present.

Brázdil and Dobrovolný (1992)[6] compiled the most detailed and most reliable record of winter temperatures from the region. Their data consist of a ten year running average of winter temperature anomalies calculated from instrumental measurements conducted in the Prague's Klementinum Observatory since the year 1771 and reconstructed from historical written records since the year 1500. Their data set is one of the longest instrumental climatic records in Europe. The close proximity of Spiralka Cave in latitude (~ half a degree) to the observatory allows us the luxury of using this climatic record as a proxy for winter temperatures in the study area. Even though deposition probably occurred only during discrete flood events, our box samples incorporate several flood events resulting in a smoothed signal. Therefore, a reasonable approach is to compare our susceptibility data to the winter temperature data of Brázdil and Dobrovolný (1992)[6]. The filtering of the temperature record is in fact comparable to the degree of smoothing of the cave record produced by the 2 cm sample size applied on discrete flood events. The match between the winter temperature anomalies and $\chi_{fe}$ (Fig. 3-9) was obtained by using the estimates of the sedimentation rates shown in Table 3-1 (adjusted for the last several decades) and calculating the position of year 1771, i.e. where the instrumental temperature record begins. This puts the year 1771 at the depth of approximately 75 cm below the surface of the profile. If our age model is correct then both the $\chi_{fe}$ and medium grain size (Fig. 3-6b) records for the uppermost 75 cm show a strong correlation with the averaged winter temperature anomalies for the entire time interval 1771 – present. The $\chi_{fe}$ record shows a positive correlation with the winter temperature anomalies with minor discrepancies attributed to changes in the sedimentation rate resulting from the varying intensity of floods and/or from their altering frequency. The most probable explanation of the correlation of the records is that during years with warmer winters (positive winter temperature anomalies) and less snow cover the floods were less intensive (smaller grain size) but probably had access to larger tracts of cultivated land (discussed below) as agriculture tended to expand during these warmer periods. Cultivation of the land provided flood waters with greater access to coarser grained magnetite-like materials exposed by tilling of the soil. Banerjee et al. (1981)[2] also suggested development of large-scaled cultivation being responsible for the increase in magnetite grain size distribution in post Little Ice Age sediments from Long Lake, Minnesota. Erosion of these cultivated soils resulted in a greater influx of coarser magnetic grains into Spirálka Cave leading to increased values of magnetic susceptibility even though the physical grain size of the sediments transported into the cave was smaller. During years with colder winters the situation was the opposite. Cultivated lands were not as plentiful and melting of a larger snow cover resulted in more erosive floods. These floodwaters were more energetic, thus carried larger grain sediments, yet with lesser ferromagnetic content and therefore lower magnetic susceptibilities.

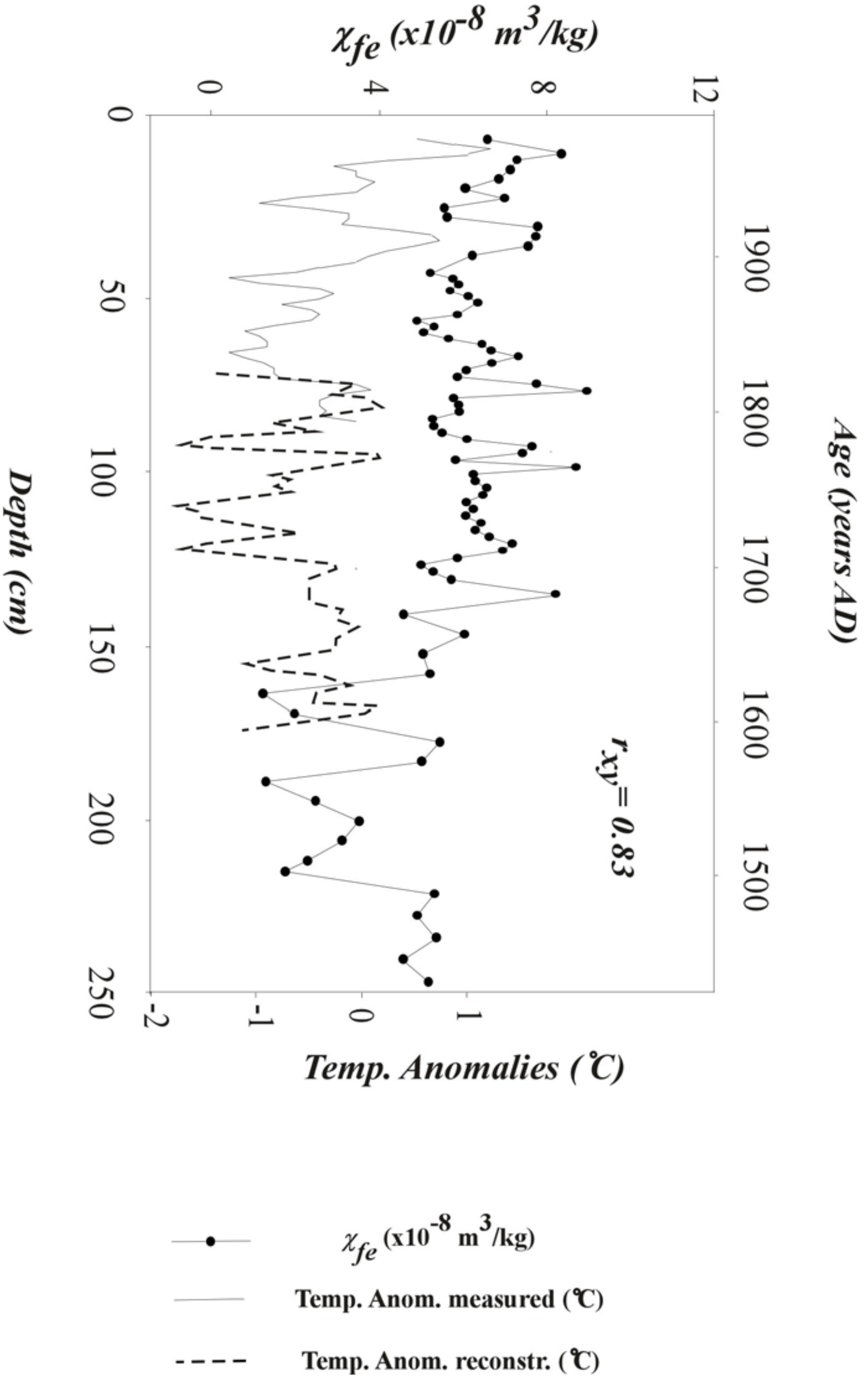


Fig. 3-9. Correlation of the upper 250 cm of the ferromagnetic susceptibility record obtained from the Spirálka Cave with measured and reconstructed winter temperature anomalies for the last four centuries. The correlation coefficient for the measured data is listed

During the last two hundred years, a broader region including the Bílá Voda catchment underwent considerable environmental changes affecting namely the forest cover caused primarily by technological progress in nearby iron industry centers followed by increased volume of construction work.

We suggest that during warmer periods when forest clearance and cultivation were more prominent, sediments/soils containing abundant ferromagnetic materials were more easily eroded and transported during snow melting events. This process explains the correlation between increased ferromagnetic susceptibility of the cave sediments with positive winter temperature anomalies (Fig. 3-9) and correlation of ferromagnetic susceptibility to increased concentrations of ferromagnetic materials with larger grain sizes.

The section of the profile in the depth interval between 160 and 225 cm is affected by post-depositional dissolution of magnetite and is easily detected by a significant decrease in the $\chi_{fe}$ values and in the S-ratio. The color of the sediment in this interval is clearly greener and is rich in organic matter as suggested by two distinct peaks representing two events in the loss of ignition record. Results of the radiocarbon dating imply the age of this organic rich horizon between 1450 and 1600 A.D. According to historical records (Bradley, 1999[4]) the period after the middle of the 15$^{th}$ century was characterized in Europe by a marked climatic deterioration from the conditions of the Medieval Climatic Optimum into the Little Ice Age. In the specific setting of Central Europe this deterioration was accompanied by an increased rate of precipitation (Brázdil, 1988[5]) leading to devastating floods and following increased depositional rates. According to the record from Spirálka Cave the increased flooding led to redeposition of the surficial weathering products into the cave. The green-organic rich layers are most likely sediments from the floodplain covered by rich vegetation, which were being rapidly re-deposited into the interior of the Spirálka Cave during floods. Therefore sediments characterizing seemingly milder and more humid conditions were in fact deposited in the cave during periods of apparent climatic deterioration. The susceptibility lows, however, are not in this case indicative of temperature changes but are a result of reducing conditions in the organic rich layers leading to magnetite dissolution.

The magnetic and non magnetic parameters deeper in the section below the depth of 225 cm are fairly monotonous and indicate warmer and more humid conditions corresponding to the Medieval Climatic Optimum. The values of LOI are fairly high; the median grain size and other erosion indicators are low thus suggesting more humidity and less snow cover. The data paint a picture of a gradual climatic improvement toward the depth of 380 cm i.e. toward ages of approximately 1050-1100 A.D. This climatic improvement is apparent namely in the LOI record (gradual increase) and perhaps also in the $\chi_{fe}$ record (minor gradual decrease).

Finally, in the bottom most 1m of the section we witness once again a definite climatic deterioration. All the parameters show abrupt changes indicative of increased erosion (grain size, Ti/Fe, Ti/Zr, $\chi_{fe}$) and less vegetation cover (LOI).

### 3.6. Conclusion

(1) Results from Spirálka Cave suggest that deposits in the interior of this cavern yield a detailed record of changing environmental conditions during the last millennium. However, transport, sedimentary, and post-depositional processes in the cave environment are complicated and may obscure the primary paleoenvironmental signal. Prior to attempting a climatic reconstruction, a thorough understanding of the above mentioned processes had to be reached.

(2) We have shown that ferromagnetic susceptibility variations in the upper 1 meter of the Spiralka Cave profile are caused by changes in concentration and grain size of “magnetite type“ minerals. Since ferromagnetic susceptibility and concentration variations also show a positive correlation to the erosional parameters from this portion of the profile and to changes in the 10-year running average of winter temperature anomalies for the last 200 years calculated from instrumental measurements at Prague’s Klementinum Observatory (Brázdil and Dobrovolný, 1992[6]), these variations are depositionally controlled by factors affecting the catchment area of the Bílá Voda River, i.e. namely spring floods induced by melting of the snow cover, forest clearance and cultivation episodes, and forest fires. Therefore, ferromagnetic susceptibility in the upper 1 meter of our profile is a

sensitive indicator of environmental events (interplay of man and changing climate) occurring in the catchment area.

(3) The investigation of the conditions in the catchment of the Bíla Voda River also suggest that this area was affected by climatic changes observed in other parts of Europe. The actual events such as the Little Ice Age, Medieval Climatic Optimum, etc. are identified in the data at well defined time/depth intervals. Due to the exclusive conditions operating in the cave which are described above in great detail the sediments responded with a unique feedback mirrored in the specific response of the mineral magnetic parameters.

(4) Radiocarbon age dating for this time interval remains problematic due to the rapid variation of the carbon isotopic content in the atmosphere. Yet, when ages are assigned to the record based on the best age estimate obtained from the radiocarbon dating the correlation between the record and the winter temperature anomalies is remarkable.

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## Chapter 4

# Conclusion, Main Results and Future Work

Caves function as large sediment traps, accumulating clastic, chemical and organic debris mobile in the natural environment during the life of a cave. Cave sediments are one of the most richly varied deposits that form in continental environments and tend to be preserved for long spans of time. The material found in caves either originates from weathering of cave walls, is transported into the cave from outside or forms in situ due to chemogennic processes. Cave sediments depend significantly on the climate as it impresses characteristic features on the material before being transported into the cave. Cave sediments allow us to study traces of climatic changes without regard of its further change due to the “conserving” effect of the protective cave environment and minimal weathering as compared to the surface. Of fundamental importance for Quaternary stratigraphy are the following phenomena occurring in caves: deposition and erosion of sediments, intensity of mechanical weathering, intensity of chemical weathering occurring outside of the cave (soil formation) and secondary accumulation of calcium carbonate.

Magnetic properties of iron oxides found in naturally occurring sediments do reflect changes in environmental processes operating at the Earth's surface. The extent to which magnetic properties are conservative or subject to change depends on the nature of the processes to which the magnetic minerals have been subjected. Chemical transformation is most importantly caused by weathering, soil formation and sediment diagenesis. These processes lead to transformation between magnetic mineral types, conversion of paramagnetic iron to ferri or antiferromagnetic minerals or conversion from ferrimagnetic minerals to paramagnetic forms. The above mentioned processes can also lead to dilution or concentration of magnetic minerals depending on the stability of the surrounding non magnetic minerals and on their authigenic production. The second factor strongly influencing the make up of the sediment is the method of transport which can lead to an overall decrease in grain size and change in particle shape as well as to grain size sorting. Thus the final sediment can significantly differ from the source rock. Lastly, during deposition magnetic minerals can be diluted to various degrees depending on the precipitation of authigenic minerals. The change in the energy of the sedimentary environment can lead to grain size sorting.

Deposits typically investigated by environmental magnetic techniques are terrestrial loess, lake and deep sea sediments. Studies have conclusively shown that variations in magnetic parameters from these environments correlate well with climatically controlled variations in the marine oxygen isotope record and show the presence of Milankovitch cycles. The cave environment, both the entrance as well as the interior, yields an enormous richness of well preserved clastic and chemogennic sediments spanning extensive time intervals and perfectly suitable for environmental magnetic studies, yet researchers have so far rarely availed themselves of this opportunity.

Our environmental magnetic study focused on sediments deposited in caves of the Moravian Karst, SE Czech Republic. The Moravian Karst is a classical karst region formed in Devonian limestones. The karstification process occured during the Late Tertiary and Quaternary and resulted in extensive and well developed surface and underground karstic features. The area contains approximately one thousand registered caves, the Amateurs Cave system being one of the longest in Central Europe. The accessibility, beauty of the karstic landscape and the unique scientific findings lead to a several centuries long history of excavations and researching in the caves and other kartic features. From the vast array of sedimentary sections suitable for paleonvironmental studies we chose two, one in Kulna Cave and one in Spiralka Cave.

Kulna Cave is an approximately one hundred meters long tunnel-shaped cave forming the upper level of the large Sloupsko-Sosuvske Cave System. The entrance facies sediments in Kulna consist of interbedded layers of loess and loam which were directly blown into the cave entrance and/or were redeposited by slope processes from the vicinity of the cave during the last glacial period. The layers of loess and loam overlie fluvial sands and gravels deposited during the last interglacial. Previous research at Kulna concentrated on the archeology, paleontology, and dendrology of these entrance facies sediments. From these data the paleoenvironmental conditions in the vicinity of the cave were reconstructed. Our results suggest that susceptibility variations and, in particular, variations in pedogenic susceptibility yield a more detailed climatic record of the paleoenvironmental conditions at the cave during the last glacial stage.

Magnetic susceptibility ($\chi$) was measured on approximately seven hundred samples collected throughout three well-studied profiles in the cave entrance. The $\chi$ record is well defined and is correlatable from one profile to another. Mineral magnetic measurements (frequency and temperature dependence of $\chi$, magnetic grainsize, coercivity characteristics) suggest that $\chi$ variations in the Kulna sediments from the last glacial stage are controlled by the concentration of pedogenically formed magnetite and/or maghemite. After removal of the effects of fine carbonate debris and detrital ferromagnetic minerals on the bulk $\chi$ record, we obtained a record of pedogenic susceptibility ($\chi_p$). The magnitude of $\chi_p$ quantifies the concentration of pedogenically formed magnetic minerals and reflects the intensity of pedogenesis. Therefore $\chi_p$ is a suitable proxy for those factors that control soil formation: climate and vegetation. Our $\chi_p$ record is also in good agreement with the median grain size record (another proxy for climatic change) of the Kulna sediments. We suggest that in the case of Kulna $\chi_p$ is more sensitive to climate change than bulk $\chi$.

The Kulna pedogenic susceptibility record shows variations both on long and short time scales. The long term trends are in good agreement with the deep sea SPECMAP record while the short term oscillations correlate well with rapid climatic events recorded by changes in the North Atlantic sea surface temperatures. Our results suggest that Central European climate during the last glacial stage was strongly controlled by the sea-surface temperatures in the North Atlantic. Short-term warmer events and perhaps higher precipitation over the mid-continent increased the intensity of pedogenetic weathering. The amount of independent climatic information and high resolution of the record makes Kulna an extremely important site for studying the Late Pleistocene climate.

Spiralka Cave is a more than one kilometer long cave which consists of a 60 meters deep entrance shaft and extensive horizontal passages. The lower part of the cave is occasionally flooded by the Bila Voda Stream which links it with the largest cave in the region, the Amateurs cave, lying further downstream. The five meter high sedimentary profile is located in an intermittently flooded chamber deep in the interior of the cave. It is formed by interbedded layers of silt, sand and gravel with occasional flowstone laminae and organic rich lenses. Radiocarbon dating of charcoal found within the sediments indicates they were deposited over the last several centuries during anomalously high stands of the Bila Voda Stream. The cave has been known for less than fifty years and no previous paleoclimatic studies were conducted in its interior nor in its immediate vicinity.

Magnetic susceptibility ($\chi$) was measured on more than four hundred samples collected throughout the cave section and surface catchment of the Bila Voda. Mineral magnetic measurements (ferromagnetic $\chi$, frequency and temperature dependence of $\chi$, magnetic grainsize, coercivity characteristics) suggest that $\chi$ variations in the Spiralka sediment are controlled by concentration of detrital magnetite and its grainsize. In intervals where the sediments did not undergo post-sedimentary iron reduction magnetic susceptibility proved to be a sensitive climatic indicator, which could be correlated with a temperature record spanning the last four centuries. Comparing other sedimentary parameters (heavy mineral content, loss on ignition and particle size) with the mineral magnetic record from the cave and the surface allowed us to reconstruct important environmental events affecting the Moravian Karst. Comparison of our results with hydrometeorological observations suggests that dramatic flooding events, occurring more frequently in the past than in the present, were responsible for transporting vast amounts of floodplain deposits into the cave interior. Such events had the most significant effect after cold winters with high cumulative precipitation rates. Our results also provide evidence for historically documented events which had strongly influenced the catchment area.

Combination of the mineral magnetic and sedimentological evidence yields data for changes in the wood harvesting system and for devastating fires, which burnt nearly 50% of the forests.

***Future Work***

Both Kulna and Spiralka cave sediments yield extensive datasets containing crucial information about the Earth's Quaternary past. So far, we took advantage of only a very small fraction compared to what further analysis of the measured data can offer. In the following paragraphs I propose an outline of the directions in which future research of these caves should lead.

The boundary between the last interglacial and the last glacial probably had a more dramatic character than originally assumed. According to some authors the short lived cold stadial C26 (Chapman et al., 1999) can be observed also on the continent as a late Eemian aridity pulse (Sirocko et al., 2005). Kulna Cave sediments cover this transition and mineral magnetic proxies are available to investigate the processes that led to the end of the last interglacial. I believe that by studying the end of the last glacial in Kulna Cave we could obtain crucial information on the much debated question of how will our ongoing interglacial terminate.

Oxygen isotope records from Greenland ice cores and North Atlantic marine sediments show numerous rapid climatic fluctuations during the last glacial (e.g. Grootes et al., 1993; Bond et al., 1992) however only very few comparable records from the European continent are available (Allen et al., 1999). The Kulna Cave mineral magnetic record spanning the entire last glacial shows very high temporal resolution and significant environmental sensitivity much needed for understanding the influence of the coupled ocean-atmosphere system on the continent. The high variability of the Kulna record (namely from layer 7b (MIS 4)) and its correlation with the North Atlantic temperatures reflect the extent and effects of the oceanic system on the mid continent. I am convinced that a detail analysis of the Kulna record could provide an important step towards understanding the mechanisms of the millenial scale climatic variability.

Since the Kulna Cave offers a high resolution record of the climatic oscillations occurring during the last glacial it is suitable for studying the physical processes driving these frequent changes. Several concepts including the thermal bipolar seesaw (e.g. Stocker, 1998; Shackleton et al., 2000) were suggested to explain the effect of the ocean thermohaline circulation on the polar climate and the temporal relationship between the temperature shifts in the North Atlantic region. However, the lack of detail terrestrial records prevent us from creating a consistent framework that would answer such questions as the temporal shifts, amplitude modulation and spatial extent of the Dansgaard-Oeschger and Heinrich Events over the continent. Studying the rapid climatic oscillations preserved in the Kulna sediments will allow us to assess the mechanism of continental response to the abrupt Atlantic climatic signal during the last glacial.

The temporal length of the Kulna Cave paleoclimatic proxy record allows us to study climate variability at different scales. Because of the coupling of climatic processes, understanding variability at any scale requires understanding the entire system. At interannual and decadal time scales surface temperature exhibits a strong land sea contrast (e.g. Fraedrich and Blender, 2003; Blender and Fraedrich, 2003) but the relationship at the millenial and longer time scales is not clear for the lack of sufficient temperature sequences from the continents. Kulna could provide information allowing distinction between the magnitude of changes at different scales having implications for understanding future climate development.

In Spiralka Cave more accurate dating is required, especially in the lower part of the profile. Further, more detailed comparison with historical events occuring in the area (such as plagues, forest fires, etc.) could be accomplished in order to offer other age markers useful for improving correlation between the temperature proxy obtained from the cave and the historical temperature data.

***Reference for future work:***

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# Appendix A

## Preliminary Study on the Mineral Magnetic Properties of Sediments from Kulna Cave (Czech Republic)

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# PRELIMINARY STUDY ON THE MINERAL MAGNETIC PROPERTIES OF SEDIMENTS FROM KŮLNA CAVE (MORAVIAN KARST), CZECH REPUBLIC

P. ŠROUBEK[1], J.F. DIEHL[1], J. KADLEC[2], K. VALOCH[3]

Summary: *Magnetic property variations in marine, lacustrine and loess-paleosol sequences have proved to be useful proxies in climate change studies. However in order to correctly interpret the record of the magnetic property variations it is absolutely necessary to have a good understanding of the cause of the observed variations. Most of the ambiguity in loess-paleosol studies is in distinguishing the role of pedogenesis from other climatic factors. Studying the mineral magnetic properties of the protected cave sediments which have not undergone pedogenesis allows us to determine the degree to which detrital input is climatically driven. These results will help us better understand the variations observed in the surficial loess-paleosol sequences.*

*This study reports mineral magnetic data collected from entrance facies sediments deposited during the early Wurmian glacial stage in the Kůlna Cave. The entrance facies sediments consist of loess-like silts with varying amount of talus. The magnetic susceptibility record from these sediments shows higher values in layers originating in colder climates which is different to that commonly observed in surficial loess deposits. Higher values of magnetic susceptibility in Kůlna sediments are probably due to higher concentrations of ferromagnetic minerals (magnetite and maghemite) and due to an increased proportion of superparamagnetic grains. The magnetic mineralogy and the grainsize distribution (grains larger than superparamagnetic) appear not to change throughout the studied profiles. Higher magnetic susceptibility accompanied by an increase in the superparamagnetic fraction observed in the sediments deposited during colder periods can be explained by an increased input from a pedogenic source when the vegetation cover was reduced and the erosion rate increased.*

Keywords: climate, paleoenvironment, magnetic grainsize, magnetic mineralogy, last glacial, eolian sediments.

## 1. INTRODUCTION

Mineral magnetic properties of windblown sediments have been a focus of many paleoclimatic studies because of the detailed signal they yield. Studies on loess deposits especially from north-central China have conclusively shown that variations in magnetic susceptibility (*MS*) from the loess and the interbedded paleosol sequences correlate with climatically controlled variations in the marine oxygen isotope record (e.g. *Heller and Liu, 1984*). However, the problem has become one of understanding the origin of magnetic susceptibility variations in the loess deposits. *Zhou et al. (1990)* and *Maher and Thompson (1991)* observed that the concentration of magnetic minerals and magnetic parameters that are sensitive to grain size consistently vary between paleosol and loess

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layers. These observations ruled out the possibility that magnetic enhancement is caused only by varying dilution of one type of magnetic minerals by non-magnetic components as suggested by *Heller and Liu (1984)* and *Kukla et al. (1988)*. The model that pedogenic processes are responsible for the MS enhancement is supported by other investigations (*Hus and Han, 1992; Beer et al., 1993, Heller et al., 1993*). Besides the signal attributed to fine grained authigenic magnetic minerals both loess and paleosols carry a background signal attributed to a roughly constant detrital input (*Heller et al., 1991; Beer at al., 1993*). On the other hand *Verosub et al. (1993)* assume that pedogenesis is the dominant source of magnetic minerals not only in paloesols but also in loess. Recently *Kletetschka and Banerjee (1995)* suggested that enhancement of MS in paleosols may be caused by naturally occurring fires. The ongoing discussion on the importance of pedogenesis as compared to variations in detrital input inspired us to study silty sediments deposited in cave entrances. The sediments accumulating in the protective environment of the cave do not undergo pedogenesis, and may thus yield new information on the origin of the climatic signal in loess/paleosol complexes.

## 2. SITE DESCRIPTION AND PREVIOUS WORK

Kůlna Cave is located in the northern part of the Moravian Karst (Fig. 1) in the Sloup Valley near the boundary between the Devonian limestones and the surrounding Lower Carboniferous non-karstic marine sediments. The cave has a tunnel like shape and is 87 m long, 8 m high with a maximum width of 25 m (Fig. 2). The cave is infilled by approximately 15 m of silt and sandy silt with varying amount of fine to coarse talus (Fig. 3); fluvial sediments are also present. The sediments range in color from brownish yellow to blackish grey, however most of the layers are of various shades of brown. A systematic study of the sediments was undertaken between years 1961 and 1976 by scientists from the Moravian Museum in Brno led by Dr. Karel Valoch (*Valoch, 1988*). Several shafts up to 14 m deep were excavated in the sediments. Using primarily archeological, paleontological and absolute dating methods the researchers have determined that the entrance facies of Kůlna represent almost the entire Upper Pleistocene.


Fig. 1. Location of the Moravian Karst.

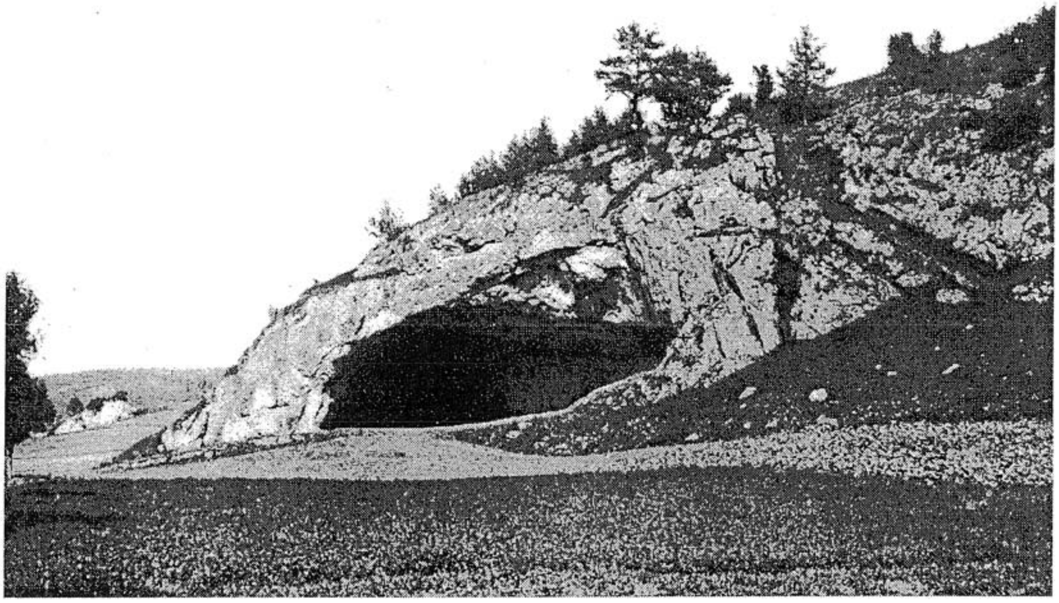

Fig. 2. Entrance to the Kůlna Cave as it appeared at the beginning of this century (after *Absolon (1905)*)

Furthermore, the deposits contain evidence for several cultural layers, including the bones of Neanderthal Man *(Jelínek, 1991)*.

*Valoch (1988)* has subdivided the sediments into 14 principal layers (Table 1). As seen in Table 1 layers 1-6 were assigned to the Holocene/Late Wurm glacial stage, layers 6a–9b to the Early Wurm, layers 10–13a to the last interglacial stage (Eemian) and layers 13b and 14 to the end of the Riss glacial stage. The stratigraphy and paleoenvironmental reconstruction of Kůlna sediments was based on archeological artifacts, pollen and dendrological charcoal analysis, paleontological findings and several radiocarbon dates *(Valoch, 1988)*. Most recently *Valoch (1992)* has correlated the Kůlna stratigraphy with oxygen isotope stages of *Labeyrie (1984)*, thereby tying the Central European climatic record to variations in the global ice volume. Absolute dates obtained using the electron spin resonance (ESR) technique *(Rink et al., in press)* which are not fully in accordance with those of *Valoch (1988)* are also available, however they do not invalidate *Valoch's (1992)* climatic interpretation.

## 3. FIELD AND LABORATORY PROCEDURES

Samples for magnetic susceptibility (*MS*) measurements and mineral magnetic analysis were collected from Early Wurm layers 6a, 7a, 7b, 7c, 7d, 8a and 11b and from layers 12a and 12b from the last interglacial (Eemian). Presented in this paper are only data from layers 6a to 8a. We used a vertical sampling interval of approximately 10 cm and collected two samples per horizon. Samples consisted of 8 cm$^3$ of sediment housed in plastic boxes.

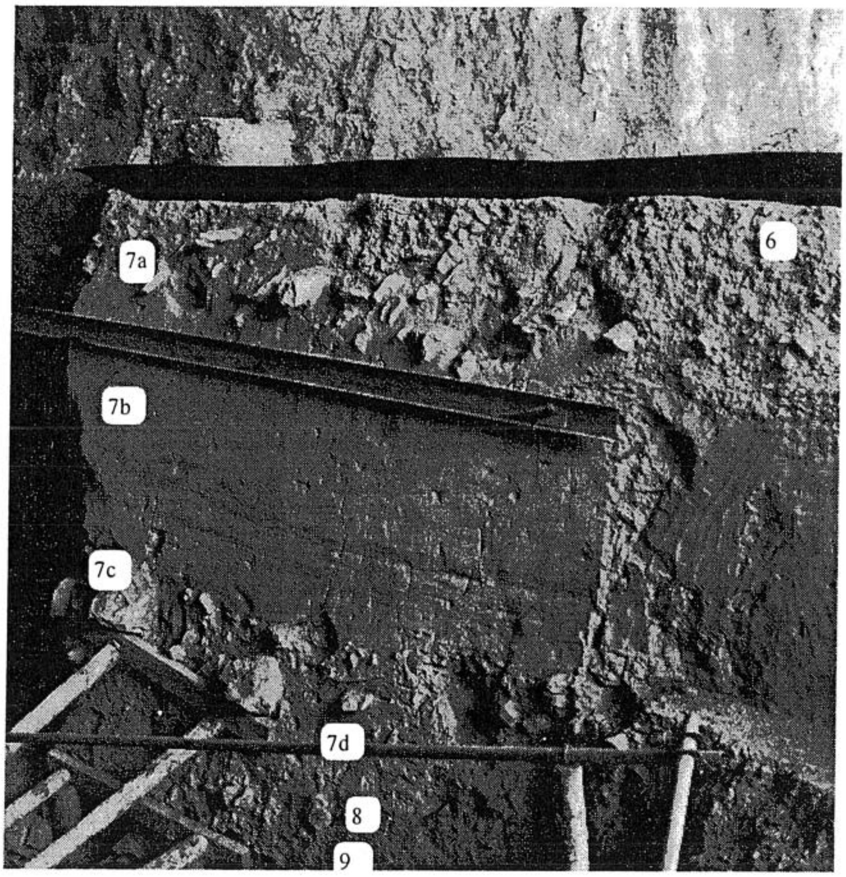


Fig. 3. One of the sampled profiles located in the entrance of the Kůlna Cave. The talus rich layers originated during warmer climate, whereas the silty layers during colder periods. Numbers with letters indicate location of layers determined by *Valoch (1988)*. (photo I. Audy)

Layers 7a, 7b, 7c and 8a were sampled at several different locations in the cave (2–15 m separation) to determine the repeatability of *MS* from one profile to another. A total of 185 samples were collected.

To determine the *MS* and its frequency dependance (*FDMS*) the samples were measured on a Bartington MS-2 magnetic susceptibility meter at two frequencies (470 Hz and 4700 Hz). Isothermal remanent magnetization (*IRM*) was stepwise induced (up to 1 T max.) using a Sapphire Instruments SI-6 pulse magnetizer and the resulting magnetization measured on a Schonstedt SSM-2 spinner magnetometer. A home made furnace with a field of less than 10 nT in the cooling chamber was used to thermally demagnetize the saturation *IRM*. Saturation magnetization ($J_s$) was measured using the Princeton Applied Research vibrating sample magnetometer. Anhysteretic remanent magnetization (*ARM*) was acquired using the Sapphire Instruments SI-4 AF demagnetizer with the *ARM* option. The biasing field was 0·05 mT, the alternating field was 1 T.

Table 1. Ecological, structural and paleocultural interpretation of the Kůlna deposits including correlation with marine oxygen isotope stages and radiocarbon dates. The temperature intensity was determined from several climatic indicators (after *Valoch, 1988,1992*).

| layer no. | Climate and environment | Stratigraphy (IS=interstadial, IG=interglacial) | culture | Oxygen stage (*Labeyrie, 1984*) | $^{14}C$ date |
|---|---|---|---|---|---|
| 1 | forest | | bronze- present | 1 | |
| 2 | forest | | linear ceramic | 1 | |
| 3 | forest | Boreal | epimagdalenien | 1 | 5,510; 10,070 |
| 4 | steppe, cold forest | Preboreal or Dryas III | epimagdalenien | 1 | 11,470; 2,135 |
| 5 | steppe, glacial climate+ warm forest | Allerod, Dryas II Bolling, Dryas I | Magdalenien | 1 | 17,480 |
| 6 | steppe, glacial climate | | Magdalenien | 1 | 11,450; 7,550 11,590; 21,750 |
| 6b | steppe, cold | Young Wurm | Gravettien | 2 | 22,990; 21,630 |
| 6a | steppe, cold | before Podhradem IS | Micoquien | | 21,260 |
| 7a | steppe, warm. cold | | Micoquien | 3 | 38,600; 45,660 |
| 7b | steppe, warm, cold | | Micoquien | 4 | |
| 7c | steppe, warm | Kůlna -IS | Micoquien | 5a | |
| 7d | indifferent, warmer than 8a | | Micoquien | 5b | |
| 8a | indifferent | | Micoquien | 5b | |
| 8b | warm | End of Eem IG | Micoquien | 5c | |
| 9a | steppe, warm+cold | End of Eem IG | Micoquien | 5d | |
| 9b | steppe, warm+cold | End of Eem IG | Micoquien, Taubachien | 5d | |
| 10 | | | Taubachien | | |
| 11a,b | steppe | | Taubachien | | |
| 11c | warm | Eem | Taubachien | 5e | |
| 11d | warm | Eem | Taubachien | 5e | |
| 12a,b | steppe | | Taubachien | 5e | |
| 13a | | Eem | Taubachien | 5e | |
| 13b | | | Early middle Paleolithic | 6 | |
| 14 | steppe, cold | Eem or Riss | Early middle Paleolithic | | |

## 4. LABORATORY RESULTS

### 4.1 Magnetic susceptibility

Variation in mass specific *MS* through layers 6a, 7a, 7b, 7c, 7d, 8a is shown in Figure 4 where each point represents the average value of *MS* from duplicate samples. The average difference of *MS* between duplicate samples was only 6·5 %. As seen in Figure 4, *MS* values from the same layer but from different profiles within the cave show very similar trends. For example, in all three exposures of layer 7b there is a apparent decrease in magnetic susceptibility with increasing depth.

*MS* values in layers 6a and 7a are highly variable (between about 20 and $35 \times \times 10^{-8}$ $m^3/kg$) and no systematic trend can be observed. The boundary between layers 7a and 7b is characterized (in two out of three profiles) by a significant increase in *MS*. *MS*

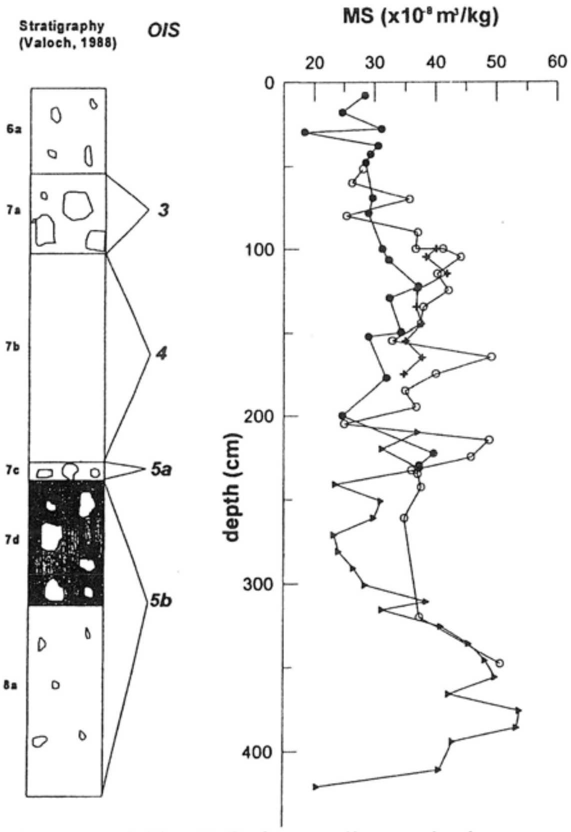


Fig. 4. Log of magnetic susceptibility (*MS*) from sediments in the entrance of the Kůlna Cave. Each symbol represents an average value of *MS* of two samples. Different symbols (circle, triangle, plus) represent samples from various profiles (2 – 15 m separation) within the cave. Also included is the stratigraphy as determined by *Valoch (1988)* and its correlation with the oxygen isotope stages (OIS) (*Valoch, 1992*). Even numbered OIS represent colder climatic periods in this and subsequent figures. Lithology and cultures (in this and subsequent figures): 6a -silt with fine talus (Micoquien), 7a - silt with a large amount of coarse talus (Micoquien), 7b - silt with sandy layers (no archeological findings), 7c - silt with coarse talus (Micoquien), 7d - sandy silt with coarse talus (Micoquien), 8a - silt with fine talus (Micoquien).

then decreases throughout the upper half of layer 7b becoming highly scattered in the lower half of layer 7b and in layer 7c. *MS* gradually increases from values of about 25 to $35 \times 10^{-8} m^3/kg$ at the top of layer 7d to approximately $50 \times 10^{-8} m^3/kg$ in the middle of 8a. Toward the bottom of layer 8a, *MS* drops rapidly to about $20 \times 10^{-8} m^3/kg$.

### 4.2 Magnetic mineralogy

To determine the magnetic mineralogy and its effect on the *MS* variations we measured *IRM* acquisition, thermally demagnetized *SIRMs*, and calculated the *S*-ratios. Stepwise *IRM* was applied to 2–3 samples from each layer. The *IRM* acquisition curves shown in Figure 5 all have similar shapes. Each sample displays a steep initial rise acquiring 90% of its magnetization by 200 – 300 mT. This behavior is attributed to the presence of a low-coercivity mineral such as magnetite and/or maghemite. However, since most samples are not saturated by 1 T, either hematite or goethite is also present in these samples.

Thermal demagnetization of *SIRM* of six samples from layers 6a, 7a and 7b features similar but complex behavior (Fig. 6). Each sample shows a rapid decrease in its magnetization between room temperature and 300 – 400°C with a minor break in slope near 180°C. After 400°C the magnetization decreases gradually showing a noticeable drop at approximately 580°C before being totally removed by 690°C. We suggest that the break in slope at 180°C indicates goethite (Néel temperature 120°C). The change in slope at 400°C was attributed to maghemite although a low unblocking temperature magnetite can

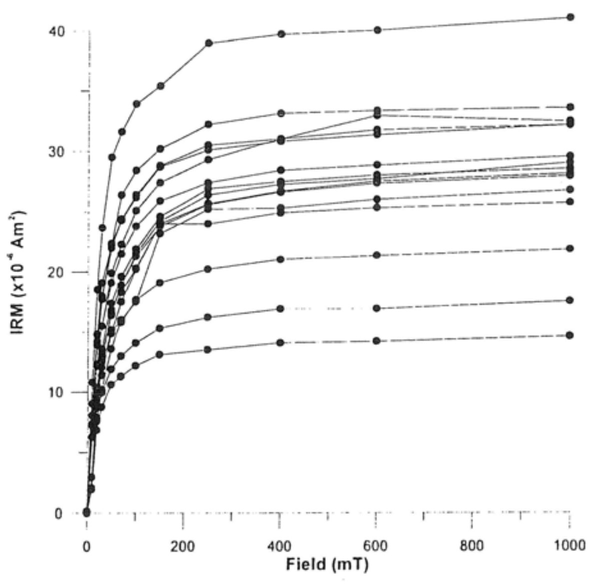


Fig. 5. *IRM* acquisition curves on samples collected from layers 6a, 7a, 7b, 7c, 7d and 8a. The shape of all the curves is very similar and is characteristic of magnetite and/or maghemite together with a high coercivity mineral.

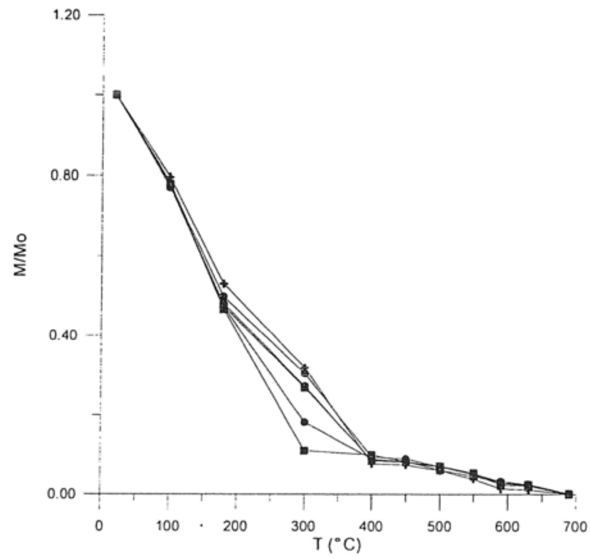


Fig. 6. Thermal demagnetization of *SIRM* of samples from layers 6a (squares), 7a (circles) and 7b (plus).

not be ruled out. *MS* measurements monitored during the thermal demagnetization process could not be used to distinguish between maghemite and a low unblocking temperature magnetite as magnetic material was being generated during heating. The remanence remaining above 580°C is carried by hematite which is either present naturally in the sediments or which formed due to dehydration of goethite between 300 and 400°C (e.g. *Thompson and Oldfield, 1986*) and/or due to the inversion of maghemite as stated above.

The values of the *S* ratio (e.g. *Bloemendal et al., 1988*) (not plotted), which is typically used as an indicator of the relative concentration of high vs. low coercivity minerals is fairly high. It is nearly constant (approx. 0·9) throughout the profile suggesting a high and constant concentration of low coercivity magnetic minerals.

### 4.3 Grain-size dependent magnetic parameters

*MS* variations are also a function of grainsize. To determine the degree to which grainsize influences our *MS* variations grainsize dependant magnetic parameters $MS/J_s$, *FDMS*, $MS_{arm}/MS$ and $MS_{arm}/SIRM$ were measured on one sample per horizon in single outcrops of layers 6a, 7a, 7b, 7d and 8a.

Fig. 7 compares the *MS* record with $MS/J_s$ and *FDMS*. The latter two parameters yield information about the concentration of superparamagnetic grains. All three records show similar trends, namely higher values for layers 7b and 8a and lower values in layer 6a, 7a and 7d. The decrease in $MS/J_s$ in layer 7d is only minor. Values of *FDMS* as high as 10% suggest a relatively high concentration of SP grains *(e.g. Thompson and Oldfield, 1986)*,

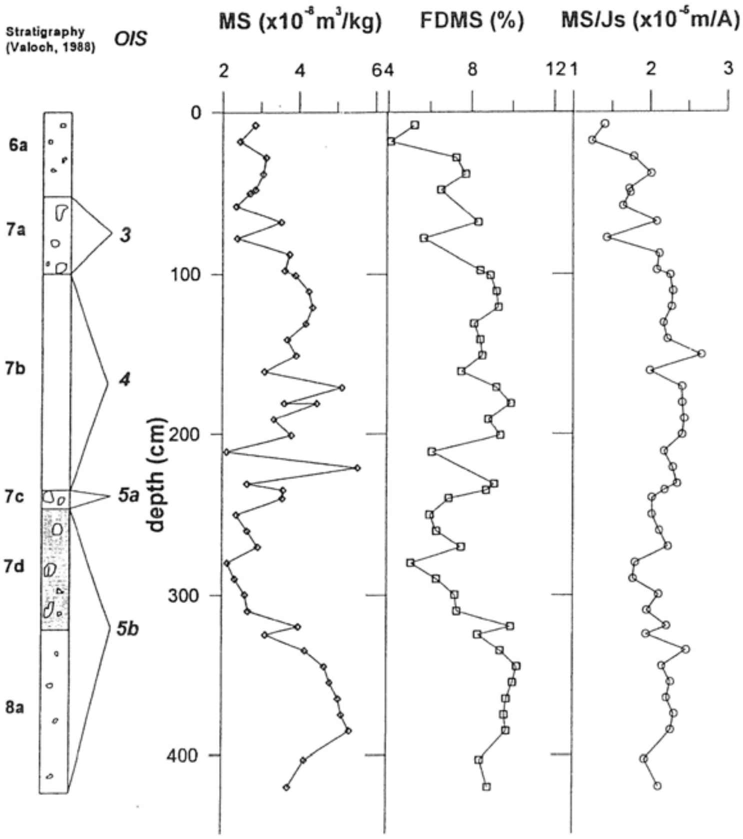


Fig. 7. Comparison of *MS*, *FDMS* and *MS*/$J_S$ logs from one profile in the Kůlna Cave. The correlation between *MS* and *MS*/$J_S$ indicates that higher values of *MS* are partially caused by increased concentration of SP grains.

however the variations in the values of *FDMS* probably do not carry a qualitative information about variations in concentration of SP grains *(Forster et al., 1994)*. On the other hand, *Banerjee et al. (1993)* showed that values of *MS*/$J_S$ give a good quantitative estimate of the concentration of SP grains. The fact that the decrease in concentration of SP grains is partly responsible for the decrease in *MS* is supported by the comparison of *MS* and *SIRM* records (not plotted). Both records are nearly identical except for layers 6a, 7a and 7d where *MS* shows a larger relative decrease.

Fig. 8 compares the *MS* record with the $MS_{arm}/MS$ and $MS_{arm}/SIRM$. The latter two parameters yield information about grainsize changes in the realm above SP. The higher the values of both parameters the smaller the grainsize *(e.g. Banerjee et al. 1981; Maher, 1988)*. Both the record of $MS_{arm}/MS$ and $MS_{arm}/SIRM$ show an increase through layers 6a and 7a and fairly constant but higher values throughout the rest of the profile. This result suggests that the variations in *MS* (except in the youngest layers) are not controlled by changes in the grainsize distribution of grains larger than SP.

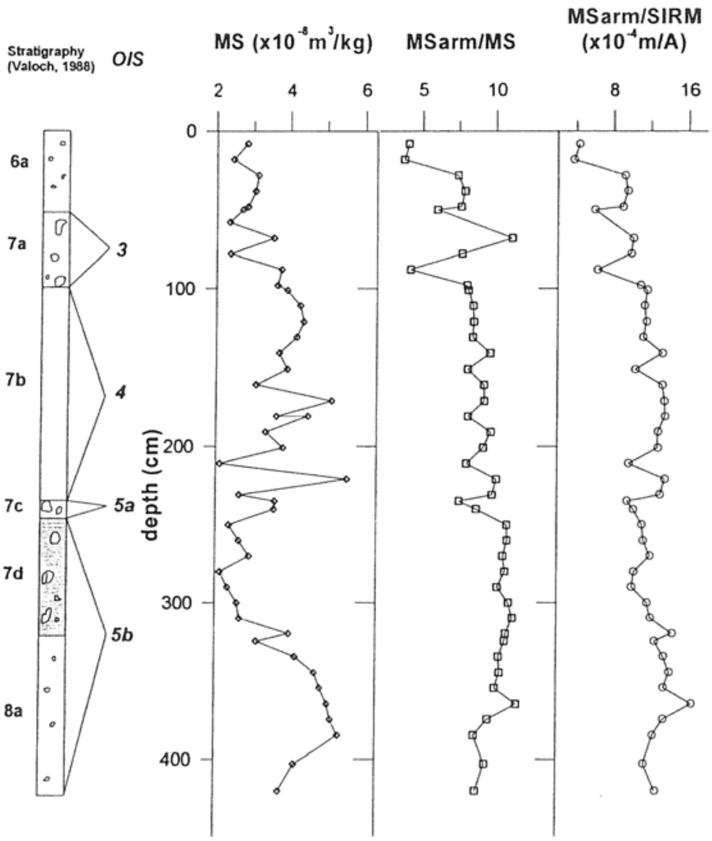


Fig. 8. Comparison of *MS*, $MS_{arm}/MS$ and $MS_{arm}/SIRM$ logs from one profile in the Kůlna Cave. The nearly flat character of the grainsize dependant parameters suggests that variations in *MS* are not caused by grainsize changes (except layers 6a and 7a).

## 5. DISCUSSION

The preliminary results indicate that sediments in Kůlna Cave do yield a well defined mineral magnetic signal. The simplest interpretation of our measurements is that peaks in the *MS* record are caused by increased concentration of ferromagnetic minerals (magnetite and maghemite) and by an increased concentration of SP grains. The mineralogy does not significantly change throughout the studied profile.

Layers 7b and 8a, which originated during colder climates *(Valoch, 1988)*, show significantly higher values of *MS* which is opposite to that typically seen in surficial loess. However the picture of high *MS* corresponding to cooler climates is not totally consistent throughout the profile. For instance, both layer 6a and 7a have low susceptibility values, even though layer 6a has been interpreted as originating in a cold climate. Interestingly, cultural artifacts were found both in layers 6a and 7a. Lower values of *MS* of cave sediments during periods of heavy occupation were observed *by Papamarinopoulos and Creer (1983)* and therefore the low *MS* values in both 6a and 7a may also be the result of human activity rather than climate change. Probably the most significant *MS* change, which is observed between layers 7d (low *MS*) and 8a (high *MS*), surprisingly corresponds to only a minor climatic variation *(Valoch, 1988)*. The low *MS* values of layer 7d may be related to the coarser grainsize of this unit. Therefore, the change in *MS* between 7d and 8a may be controlled by lithology rather than climate. Lastly, *Valoch (1988)* showed that the lower half of layer 7b was affected by cryoturbation. This process could have concentrated the heavy magnetic minerals leading to the high scatter of *MS* values in this layer.

The higher *MS* accompanied by an increase in the SP fraction observed during colder periods could be explained by an increased input from a pedogenic source when the vegetation cover was reduced and the erosion rate increased. The pedogenicaly formed minerals could have been transported from the immediate vicinity of the cave or from further away. The source of the sediments will be a subject of our further studies.

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**Notes:**