Geological Quarterly, 2007, **51** (2): 105–126

High-resolution terrestrial archive of climatic oscillations during Oxygen Isotope Stages 5–2 in the loess-palaeosol sequence at Kolodiiv (East Carpathian Foreland, Ukraine)

Maria ŁANCZONT and Andrij BOGUCKYJ

Łanczont M. and Boguckyj A. (2007) — High-resolution terrestrial archive of climatic oscillations during Oxygen Isotope Stages 5–2 in the loess-palaeosol sequence at Kolodiiv (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 105–126. Warszawa.

The Kolodiiv section is situated in the central part of the East Carpathian Foreland, on the right bank of the Sivka River, the Carpathian tributary of the Dniester River. This paper summarizes investigations on the loess deposit succession representing the Late Pleistocene, which is subdivided by several interstadial palaeosols and is bracketed by the last interglacial soil/organic deposits and Holocene soil. The Kolodiiv loess-palaeosol sequence provides an excellent high-resolution terrestrial archive of changing climate during OIS 5–2 in the East Carpathian Foreland and forms the basis for a regional pedo- and loess stratigraphy. The stratigraphic scheme was constructed on the basis of palaeosol occurrence, lithological variation in the deposits, and also the results of TL dating and palaeomagnetic investigations. The exposure at Kolodiiv contains an archaeological site with Middle Palaeolithic materials. Four types of palaeopedological taxa have been distinguished within the loess deposit at Kolodiiv: an interglacial (Eemian) soil unit, which includes one or two soil-forming episodes; interstadial palaeosols, which include two weaker soil-forming episodes; thin interstadial two-horizon palaeosols; monogenetic incipient palaeosols. Palaeosols from the first and second group form the Horohiv soil unit correlated with OIS 5. Palaeosols from the third and fourth group occur as different soil types within the Dubno 1 and 2 units, which correspond to OIS 3.

Maria Łanczont, Department of Physical Geography and Palaeogeography, Maria Curie-Skłodowska University, al. Kraśnicka 2 CD, PL-20-718 Lublin, Poland, e-mail: lanczont@biotop.umcs.lublin.pl; Andrij Boguckyj, Department of Geography, Lvov University, Dorošenka 41, 290000 Lvov, Ukraine *(received: March 3, 2006; accepted: January 18, 2007).*

Key words: East Carpathian Foreland, last interglacial/glacial period, loess, palaeosol, organic deposits.

INTRODUCTION

From the end of the Neogene, after the retreat of the Paratethys Sea, terrestrial deposits have accumulated on the Ukrainian East Carpathian Foreland, which reaches heights of 200–500 m a.s.l., and is situated between the Carpathians and the Podolian Upland. Pliocene and Quaternary alluvial deposits, and loess accumulating from the Early Pleistocene, are the main types of terrestrial deposits in this region. Loess accumulation occurred under periglacial conditions in the extraglacial zone. The relief of the East Carpathian Foreland is mainly characterized by high, wide and flat interfluve areas covered by impoverished Carpathian gravels, and by deep river valleys with a step-like system of Pleistocene terraces. The higher outer over-valley, and lower inner-valley terraces and slopes are covered by a thick loess mantle with many fossil soils. Complex cultural layers from the Middle and Upper Palaeolithic occur at many loess sites (Łanczont and Boguckyj, 2002).

This paper summarizes the investigations on the unique loess succession at the Kolodiiv section associated with a terrace representing the Late Pleistocene. This succession is subdivided by several interstadial soils and is bracketed by the last interglacial soil/organic deposits and Holocene soil. The Kolodiiv loess-palaeosol sequence provides an excellent terrestrial archive of changing climate during the last 120 ka in the East Carpathian Foreland and forms a basis for the regional pedo- and loess stratigraphy. The Kolodiiv section is situated in a transition zone between the East European periglacial loess province (Veličko, 1990; Bolihovskaja and Bolihovskyj, 2001; Trofimov *et al.*, 2001 and many other authors) and western Europe with isolated and smaller loess patches (Fig. 1). The Vistulian loesses in Ukraine, which are typical of the east European loesses, contain fewer loess-palaeosol units that have been recognized and defined than the west European ones. This phenomenon should be probably related to the evolution of natural processes during the Vistulian Glacial, which may have been modified by continentality of climate (Różycki, 1976; Van Andel, 2002). Temperature gradients and climate severity may

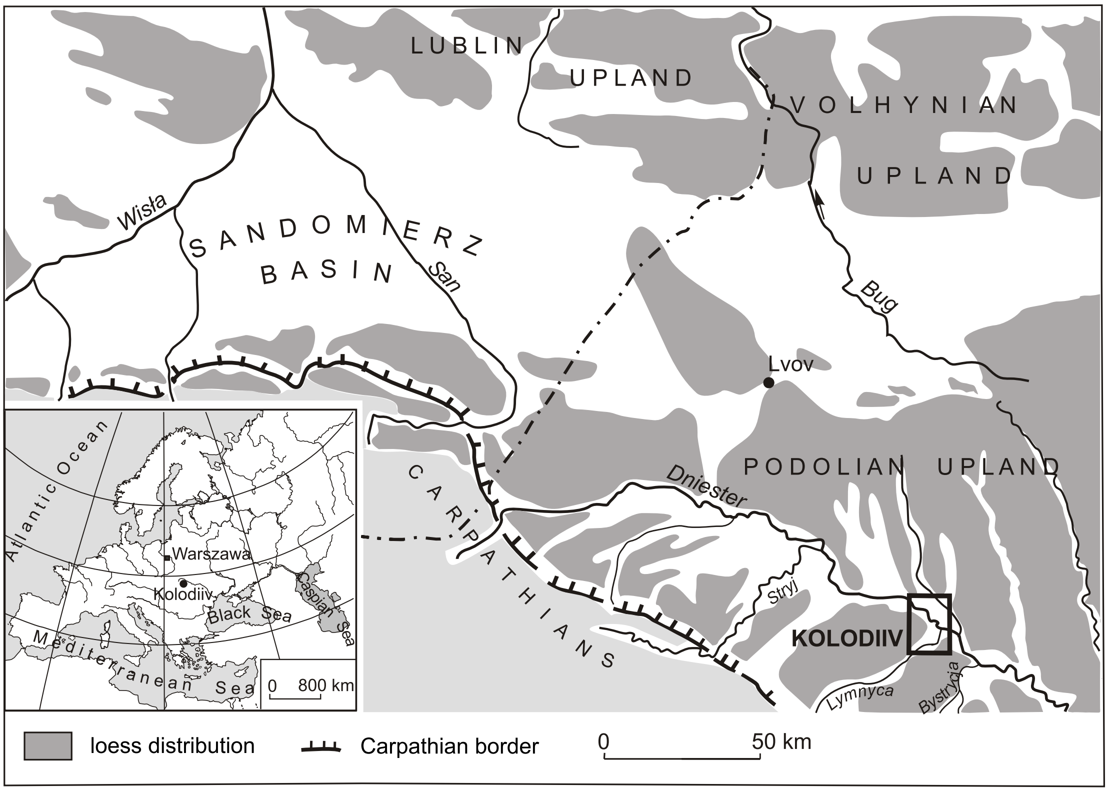

Fig. 1. Sketch map of loess regional distribution showing location of the Kolodiiv site

The extent of the map with shadow relief (Fig. 2) is marked on the map

have increased to the north much faster than in the regions with a typical temperate climate (Emontspohl, 1995). Such circumstances could have influenced the type, intensity and frequency of pedogenetic processes (which were also subjected to local conditions, especially topoclimate). Therefore, the transitional situation of Kolodiiv could provide an important basis for the comparison and correlation of loess stratigraphy between western and eastern Europe.

LOCATION

The loess section examined is situated near the village of Kolodiiv (former Polish name Kołodziejów), in the central part of the East Carpathian Foreland, in the region commonly named the Halyč Prydnistrov'ja, the Dniester River playing a dominant role in the landscape, and Halyč being the centrally located town. The Kolodiiv section is situated on the right bank of the Sivka River, the Carpathian tributary of the Dniester River, near their confluence, about 15 km to the NW of Halyč.

GEOLOGICAL SETTING

The East Carpathian Foreland arose as a peripheral trough/foredeep, lying along the front of the Outer Carpathians, associated with the last stage of orogenesis. This sedimentary basin was filled with molasse deposits of Early to Late Miocene age (Ney *et al*., 1974; Olszewska, 1999; Oszczypko, 2001). The Ukrainian part of the Carpathian Foredeep is founded both on Cretaceous-Paleogene Carpathian Flysch basement, and on a Palaeozoic-Mesozoic platform (Petryczenko *et al*., 1994).

In the right-bank part of the Halyč Prydnistrov'ja region Upper Badenian dark grey calcareous clays, mudstones and sandstones occur directly under the Quaternary deposits. Older deposits (middle Badenian) occur only as narrow strips along the Dniester River valley. They are composed of different types of gypsum, anhydrite with a terrigenous-carbonate admixture and limestones of chemical origin (Šakyn, 1976; Petryczenko *et al*., 1994; Peryt and Peryt, 1994). The Neogene deposits are absent in a wide (7–20 km) belt along both banks of the Dniester River where the sub-Quaternary basement is composed of Upper Cretaceous rocks: marls, sandstones, sands, siltstones and silty mudstones (Šakyn, 1976).

GEOMORPHOLOGIC SETTING

The relief of the East Carpathian Foreland is influenced by the main structural grain of a "Carpathian" trend. Mountainous relief characterizes the part corresponding to the inner epigeosynclinal zone of the foredeep; flat, terrace-accumulation relief predominates in the part corresponding to the epiplatform marginal zone (Kravčuk, 1999).

A transverse relief pattern is typical of the East Carpathian Foreland. Deep and wide valleys of the Carpathian tributaries of the Dniester River run mainly along NE–SW tectonic lines of the basement (Cys', 1962). They divide the upland plain into separate, high, rather flat and wide interfluve areas (Fig. 2).

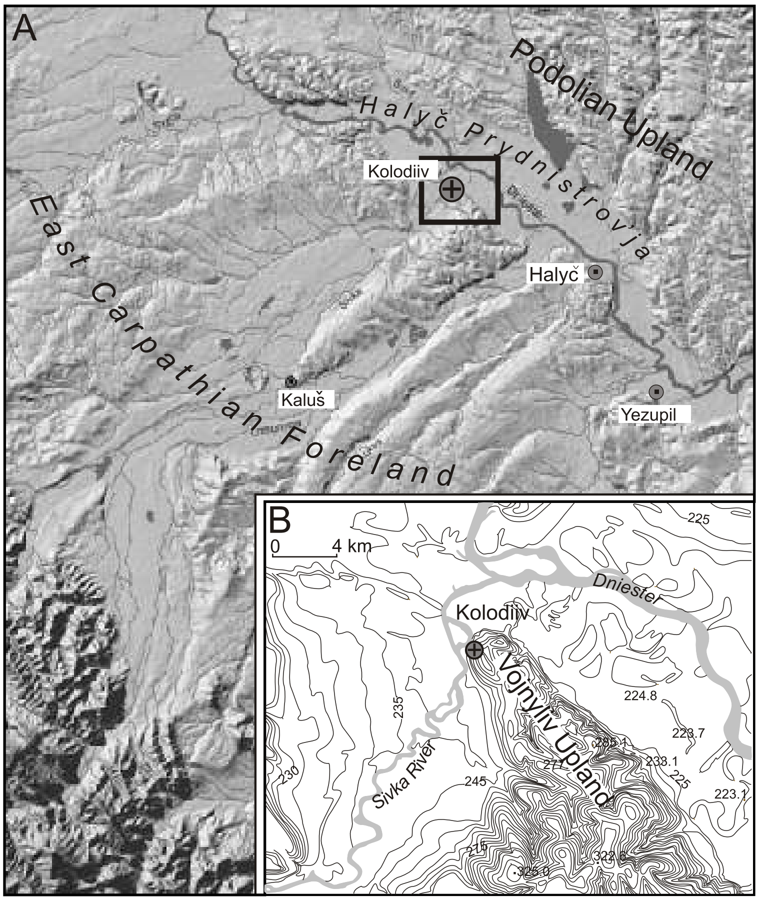

Fig. 2. A — shadow relief map indicating the loess section studied; B — hypsometric sketch of the Kolodiiv area

Two main morphological surfaces with a complex history extend at different heights along the interfluve areas. These are denudation surfaces of pediment type, and denudation-accumulation surfaces with a residual alluvial cover (Hofštejn, 1964). The Lojeva surface, related to the Early Quaternary, is a common and predominant type of landscape. It is composed of several steps (Łanczont and Boguckyj, 2002), which occur from 70–80 to 110–130 m above the river bottoms. Patches of impoverished Carpathian gravels found on this surface are up to several metres thick (Kravčuk, 1999). The Krasna surface is older (Upper Pliocene), and occurs about 30–50 m over the Lojeva surface. It is preserved only partially, on the top parts of high elevations near the Carpathian margin, and on the Prut and Dniester interfluve.

A well-developed system of Pleistocene terraces occurs in the larger river valleys. A step-like pattern of flat terraces is characteristic of the landscape of the entire Prydnistrov'ja region. The stratigraphic scheme of the Dniester River terraces is based on morphological and hypsometric criteria. The highest terraces (VII and VI) are correlated with the Krasna and Lojeva surfaces. Successive terraces (V–II) are typical intra-valley forms developed along rivers. The lowest terrace I, which fills the bottoms of valleys, is a complex form, partly or wholly of Holocene age (Demedyuk, 1966).

Kolodiiv is situated in the northwestern part of the Vojnyliv Upland (a region of the East Carpathian Foreland), which is on interfluve of the Sivka and Lymnyca rivers (Kravčuk, 1999). That region is an elongated rectangle, the longer sides of which are marked by the valleys of the above-mentioned rivers, and the shorter sides being defined by the Kaluš Basin (probably of tectonic origin) in the south-west and the Dniester River valley in the north-east. The Lojeva Plateau occupies the Vojnyliv Upland, which slopes gradually (from 365 to 320 m a.s.l.) from the mountains towards the Dniester River. Its topographic range reaches 100–110 m. The Lojeva surface is strongly dissected near the confluence of the Dniester and Sivka rivers, where the height differences are the greatest (up to 110 m). The Vojnyliv Upland forms steep sides of the valleys of the Dniester and Lymnyca rivers, so the upland block is asymmetrical. The high and steep scarps are dissected by many gullies and short but rather deep (20–25 m), usually dry small valleys.

The Quaternary system of the terraces younger than terrace VI (Lojeva surface) is represented by the Pleistocene terraces V and II, and the Holocene terraces (Figs. 3 and 4). Terrace V (55–65 m) is preserved only as fragments and forms a triangular salient in the Dniester and Sivka interfluve. Slopes of terrace V are covered with deluvial loess deposits enriched in gravels derived from an eroded alluvial cover. Terrace II is well developed (Fig. 4). It occurs as large plains on the left bank of the Sivka River, and on the right bank it forms a shelf 1.0–0.4 km wide, which becomes gradually narrower towards the Dniester River valley, and is less and less distinguishable as an individual element of relief. The relative height of this terrace varies, from about 20 m (in places even 15 m) to 25–27 m in the northern salient of the terrace extending towards the Dniester River valley. The terrace edge, especially in the sections undercut by the Sivka River, is a high SSW–NNE scarp. The terrace forms a sharply outlined morphological surface due to the loess cover, which hides smaller elements of palaeorelief, e.g. rock bars separating individual terrace patches. Therefore, the terrace structure is more complex than it appears from the modern relief.

The bottom of the Sivka River valley is about 1 km wide. The Holocene terrace, 3–5 m high, covers almost the whole valley bottom. A narrow proximal floodplain about 0.5–1 m high occurs just near the river channel (Fig. 3). Around Kolodiiv the rock channel is covered with a thin layer of well-rounded, disc-shaped pebbles, which have been transported by the modern river (Fig. 5).

HISTORY AND SCOPE OF RESEARCH

In 1971 the Ukrainian geologist Demedyuk published the first description of the Kolodiiv section. He discovered a layer of fossil peat 1 m thick separating the Pleistocene alluvia of the

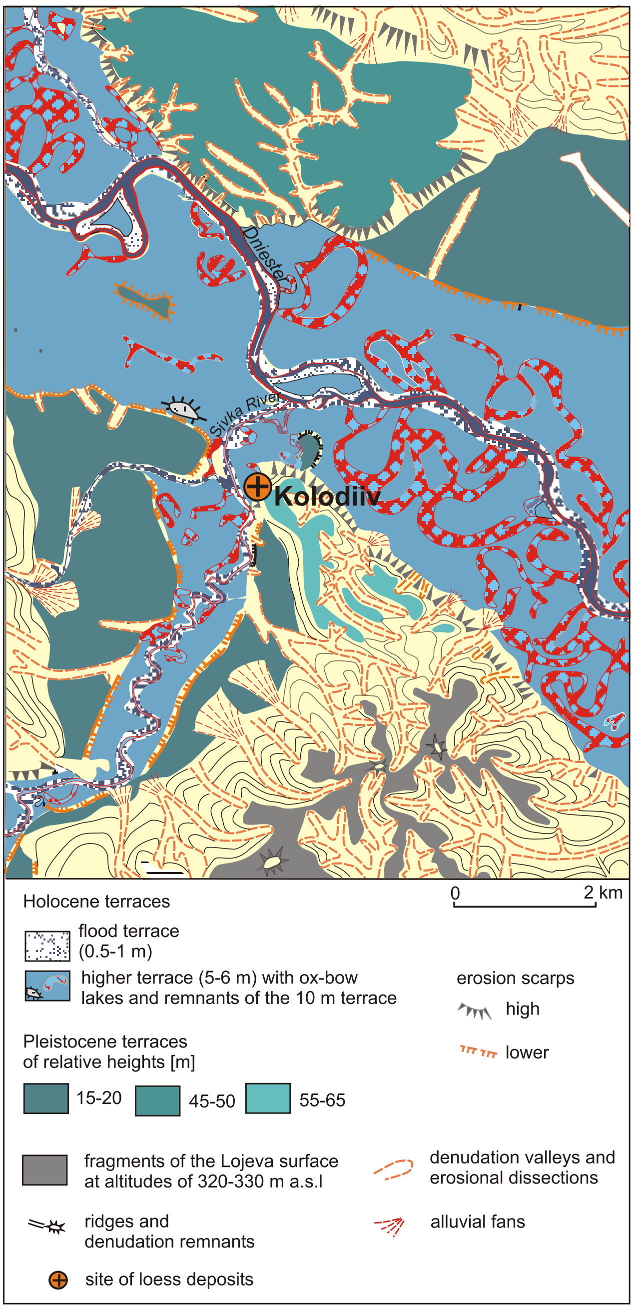

Fig. 3. Geomorphological map of the Sivka River valley in the Kolodiiv area

Fig. 4. Terrace II of the Sivka River and terrace VI of the Dniestr River

Fig. 5. Rock bed of the Sivka River; original weathering forms of the Cretaceous sandstones

Sivka River from the subaerial loess cover. As Pleistocene interglacial organic deposits are almost absent in the East Carpathian Foreland, that discovery was and remains very important. Pollen analysis (Demedyuk and Khrystoforova, 1975; Gurtovaya, 1983; Kalinovych, 2002; Bezus'ko and Bezus'ko, 2003; Sytnyk *et al*., 2003) and macrofossil analysis (unpublished results of the analysis made by Veličkevič in 1977) unequivocally indicate that the peat was formed in the Mikulino (= Eemian) Interglacial.

Detailed, interdisciplinary investigations of this section were undertaken in 1998. They were made by a team of Polish and Ukrainian scientists, and supported by KBN Grant 6 P043E 031 15. During the field expedition in 1999, a layer of fossil gyttja was found near the fossil peat, at similar hypsometric level. Its palaeobotanical content supports an Eemian age (Kalinovych, 2001, 2002).

The exposure at Kolodiiv also contains a Middle Palaeolithic site. The archaeological materials were found in 1991 (Cyrek and Sytnyk, 2002). Unfortunately, extensive excavations made in 1999, 2000 and 2003 yielded on limited results.

The investigations of the Pleistocene loess deposits in the Kolodiiv profile were made in 1998–2003, along the over 800 m long section of the straight edge of terrace II. In the southern part of the section examined the river has undercut the terrace (Fig. 6). A cut-off occurring at the foot of the northern part was an active river channel in the last part of the 19th century (Teisseyre, 1896). The terrace edge is strongly affected by landslide processes, and the bottom parts of the loess scarp are inaccessible for exploration in many places as they are buried under a thick layer of slope wash deposits. Therefore, good exposures need be sought along the whole terrace edge. The spatial variability of the deposits forming the terrace was studied, especially palaeosols of different stratigraphic rank, from interglacial to interphase ones, and layers of organic accumulation.

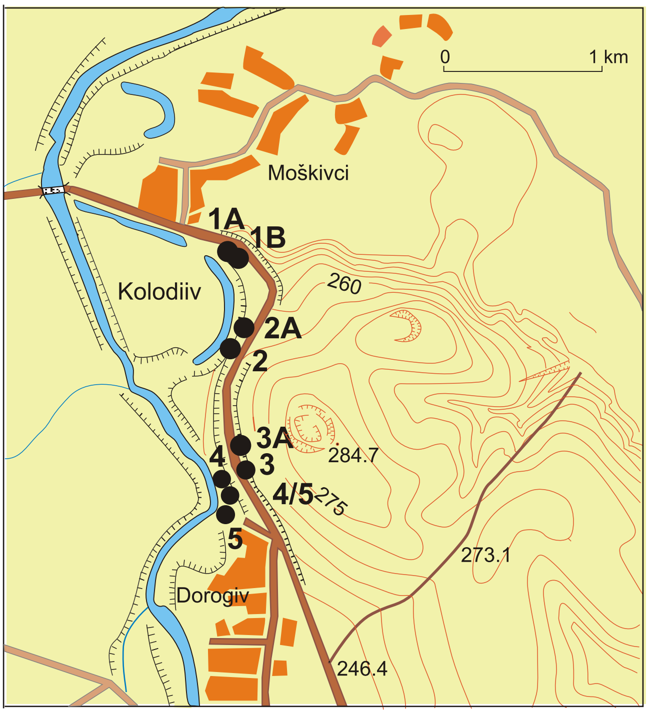

Fig. 6. Situation sketch of the set of Late Pleistocene profiles at Kolodiiv

The nine exposures examined were denoted with combined number-letter symbols increasing up-stream (Fig. 7). Full profiles of the deposits building the terrace, from the rock basement to the Holocene soil, were exposed at three locations (Fig. 7; 2, 2A, 3 profiles) and elsewhere only selected parts were revealed.

Detailed field investigations of exposures were the main method of study of the Quaternary deposits at Kolodiiv. These stratigraphic-palaeogeographic works concerned the lithological and facies characteristics of deposits, their origin and stratigraphic variation. Sketch and photographic documentation was made in the field, as well as sampling for laboratory analyses. Samples were collected from all layers distinguishable by their lithology, and every 0.5 m from the massive parts of the profiles examined. Commonly analyses conventional for loess research were carried out, together with more specialist studies. Grain-size analysis was carried out applying the aerometric method, but with modifications (Seul, 2007). Granulometric indices were calculated on the basis of grain diameters expressed on the phi scale following Folk and Ward. Humus content was estimated using the method of Tiurin, carbonate content using the volumetric method of Scheibler, and the content of iron oxides by means of a colorimetric method (Łanczont and Boguckyj, 2002). Thermoluminescence ages of selected samples were also ob-

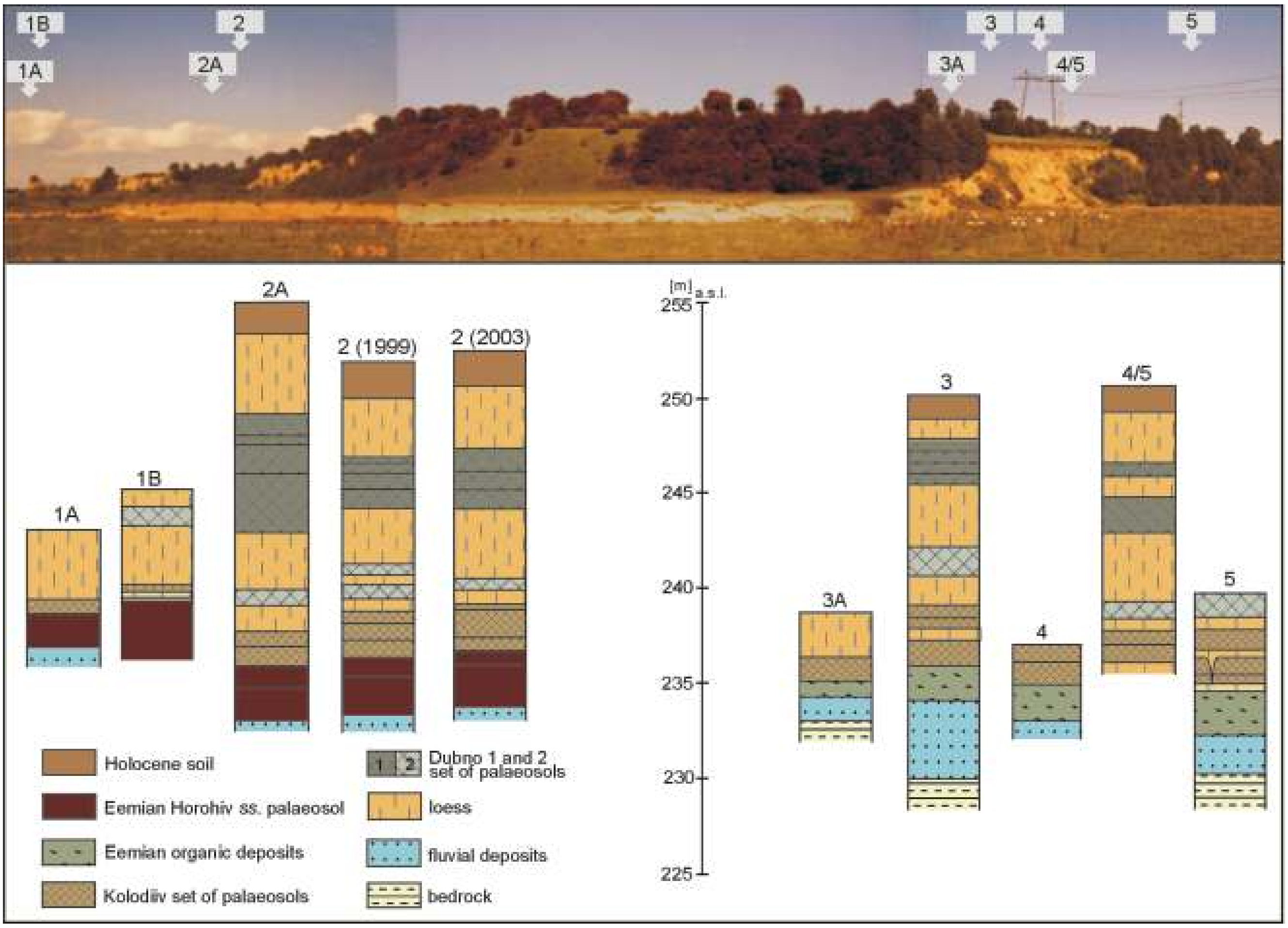

Fig. 7. Set of the Late Pleistocene profiles at Kolodiiv

tained (Kusiak, 2007). Heavy mineral analysis was made (Racinowski, 2007) of the Kolodiiv 2A profile. The contents of transparent minerals were determined, and two weathering indices were calculated: O/S+N, i.e. the index of deposit maturity, which gives numerical information about the relation between transparent minerals resistant (O) to weathering, those moderately resistant (S), and those non-resistant (N); C/G+A, i.e. the index giving the ratio of zircon grains to the sum of garnet and amphibole grains. These sensitive indices are useful when drawing conclusions about hypergenetic processes occurring in loess covers, and estimating the dynamics of the sedimentary environment.

The detailed descriptions of all the profiles examined together with their stratigraphic interpretation were published by Łanczont and Boguckyj (2002), so in this paper we give only shortened versions. The results obtained and other data are shown in the diagrams compiled for individual profiles.

The stratigraphic scheme of loess and loess-like deposits at Kolodiiv was constructed on the basis of palaeosol occurrence, lithological variation of the deposits, archaeological criteria, and also the results of palaeomagnetic investigations (Nawrocki *et al*., 2007). The palaeocryogenic studies were of little use stratigraphically as cryogenic structures (especially ice wedge casts) are very weakly developed in the area examined. When resolving detailed stratigraphic problems, the results of TL dating were difficult to interpret due to the large amount of slope material redeposited from the older units. Therefore, some of the TL results are overestimates with respect to the ages expected on the basis of geological and palaeopedological interpretation, as well as by comparison with palynological data. The stratigraphic position of the Eemian biogenic deposits is clear as the flora is characteristic.

The stratigraphy of the last glacial loesses at Kolodiiv was based on the stratigraphic division of loesses from the Volhyno-Podolian Upland (Bogucki, 1986, 1987), but is much more detailed. Numerous palaeosols of different rank (interstadial and interphase), and loesses associated with them were the basis for distinguishing subordinate stratigraphic units. A distinctive feature of the site is the occurrence of three palaeosols developed in the Early Vistulian, which were named Kolodiiv 1, Kolodiiv 2, and Kolodiiv 3 of the Vistulian (OIS 5a and 5c1, 5c3). This provides a new approach to interpreting the Horohiv palaeosols (Eemian–Early Vistulian), recognized in Western Ukraine (Boguckyj and Łanczont, 2002). The complex letter-number signs used to denote the stratigraphic units distinguished in the western part of Ukraine (Boguckyj and Łanczont, 2002) were borrowed from the Chinese loess terminology (Liu *et al*., 1985; Kukla, 1987; Ding *et al*., 1994; Feng *et al*., 1998; Heslop *et al*., 2000; Wu *et al*., 2002), with our own modifications (Fig. 8). The Horohiv multilayer soil succession representing OIS 5 is denoted as S1 sign, and loess representing OIS 4–2 as L1. Minor loess units within the Vistulian loesses are numbered with increasing age within the main unit, e.g. L1-ll. Intra-loess soils of lower interstadial rank are marked with lower case, and also are numbered with increasing age within the main unit, e.g. L1-s1 (= Dubno 1 horizon).

The stratigraphic units distinguished are compared (Fig. 8) with the stratigraphic succession of Polish loesses published by Maruszczak (1994, 1996, 2001), and with the succession of Ukrainian loesses and palaeosols (Veklič, 1968; Gozhik *et al*., 1995, 2001).

THE DEPOSIT SEQUENCES, PEDOSEDIMENTARY EVOLUTION AND PALAEOENVIRONMENTAL INTERPRETATION

INTRODUCTION

Alluvium of the Kolodiiv terrace is overlain by Eemian deposits, which are overlain in turn by Vistulian loesses and loess-like deposits about 11–17 m thick. A cut-off meander 1–2.5 m deep, filled with gyttja and peat, was present on the terrace in the Eemian Interglacial, and an authigenic soil cover developed on its exposed higher parts. A considerable number of typologically variegated palaeosols is the most important feature of the subaerial deposit series. They reflect great variability of the terrestrial climate during the last glacial, and their spatial pattern indicates that initially the sedimentary environment was a flat and gently inclined surface of gently variable microrelief. The relative heights in this area increased with time as loess deposits accumulated. The narrow terrace was covered not only by falling dust by also by slope material moved by gravitational processes. Gravels of fluvial origin dispersed within younger loess layers indicate that the terrace adjoined the contemporary slope. Pebbles were dislodged from the higher parts of slope, and more precisely from the eroded old alluvial deposits of the Pleistocene higher terrace V. The microrelief of the surface that was gradually covered with loess varied along profiles transverse and parallel to the valley axis; various transverse convex and concave elements of second rank can be reconstructed on the basis of the spatial relationships between the stratigraphic units distinguished in the Vistulian loess. These local topographic conditions probably favoured the preservation of loess representing the entire time interval.

Of course, both more and less favourable places were present in these areas of loess accumulation as regards the possibility of preserving a continuous record of natural processes in the deposits (Mojski, 1993). In some places not only older deposits were preserved but also younger covers accumulated. Kukla and Ložek (1961), and many other authors (e.g. Łanczont, 1995; Schirmer, 2000; Gozhik *et al*., 2001) stressed that the slope loess sequence (i.e. that associated with an inclined surface of valley-side type) was more sensitive to climate fluctuations than loesses on plateaux where the accumulation rate was lower, and pedogenesis of successive cycles (interstadial, interphase) may have overlapped. Climatic fluctuations are recorded in detail, better, and more completely in deposits that accumulated on slopes. In comparison to the subhorizontal deposits of plateaux, layers that accumulated on inclined relief elements were formed (independently of aeolian dustfall intensity) by soil-forming processes of different intensities, and other processes such as water transport and soil dislocation, e.g. solifluction. Soils developing in such loess sequences in warmer periods stabilized the slope surface. In cold periods they could have been eroded but may also have been protected,

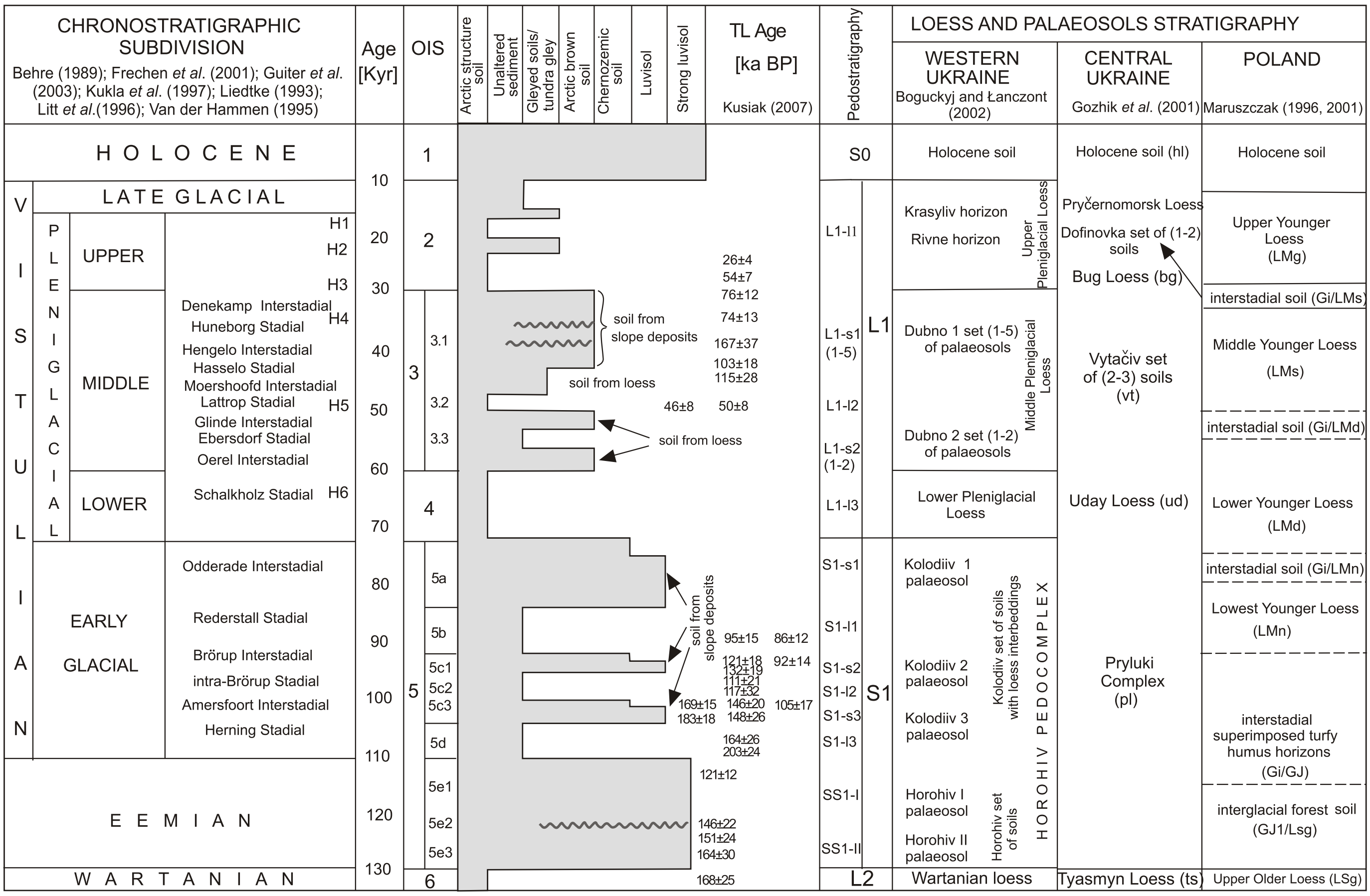

Fig. 8. Sequence of palaeosols in the Eemian and Vistulian of the Kolodiiv site correlated with the stratigraphic divisions of loesses in Ukraine (Boguckyj, 1986; Gozhik *et al*., 1995, 2001) and in Poland (Maruszczak, 1996, 2001)

Detailed division of the last glacial in western Europe after Behre and Lade (1986), Behre (1989), Catt (1991), Liedtke (1993), Van der Hammen (1995) H1–H6 — Heinrich events after Bond *et al*. (1992, 1993)

e.g. by deluvial or solifluction covers. Sometimes para-autochthonous soils could have developed as a result of successive accretion of soil material redeposited from higher parts of the slope (Smolikowa, 1971). In consequence, a single soil profile developed on a plateau can correspond to a complex sequence of soil and sediment interlayers, partially of aeolian origin, and partially composed of silts redeposited by slope processes. Fully developed loess packages (i.e. complete stratigraphic units) can be also preserved between soils. Slope deposits are also represented by material deposited in open or closed depressions of secondary rank (e.g. small dry tributary valleys, closed depressions on interfluve areas), in which quiet, continuous sedimentation occurred till these forms (as sediment traps) were completely filled.

WARTANIAN DEPOSITS (OIS 6)

The rock basement of the Kolodiiv terrace rises from 1.5 to over 2 m above mean water level in the river channel. It is composed of massive, thick-bedded and strongly fissured, yellowish-grey and pale grey sandstones of the Campanian (Šakyn, 1976). The medium-grained sandstones that are carbonate cemented are very weathered and disintegrated near their surface, i.e. just under the Pleistocene alluvium; in places they have transformed into loose sand. A moderately irregular erosion surface of the rock basement, with relative heights reaching 1 m, is covered with alluvial deposits of variable thickness (1–4 m). It contains mainly gravel deposits. The pebbles are irregularly scattered within a sandy and sandy-loamy matrix. They are well rounded; the longer axes of flat, disc-shaped pebbles are up to10–20 cm long. The alluvial deposits are carbonate-free, with a very small content of humus (Fig. 9). In the top part they are enriched with iron oxides (up to 3–6%), and black spots of manganese oxides also occur. The upper part is characterized by the predominance of very resistant minerals, i.e. zircon (>60%), and rutile (20%), in the heavy mineral assemblage.

The Wartanian age of the alluvial series is indicated by the TL dates (164–168 ka BP) obtained for the fluvial deposits, and primarily because they form a continuous sequence with the overlying Eemian biogenic and mineral-organic deposits that accumulated in the meander cut-off, or with the fossil forest soil.

EEMIAN MINERAL-ORGANIC DEPOSITS (OIS 5E)

The shape of the meander cut-off is difficult to fully reconstruct, but its length can be estimated at about 200 m, i.e. similar to the sizes of the Holocene cut-offs (meander wavelength

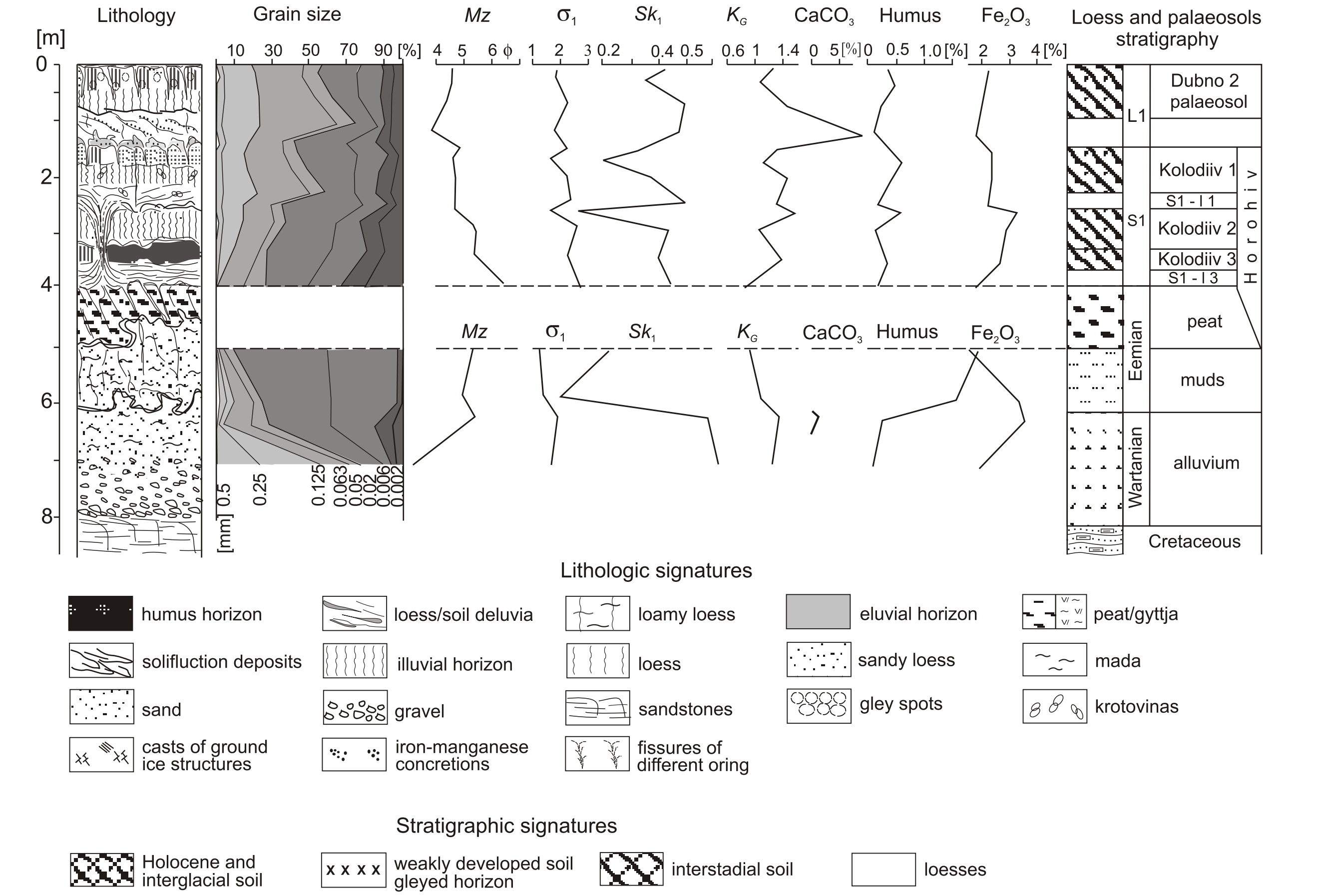

Fig. 9. Kolodiiv 5 profile

Diagram of grain-size distribution: Mz — mean grain diameter, σ_1 — sorting index, Sk_1 — skewness index, K_G — kurtosis index, $CaCO_3$ — carbonate content, Humus — humus content, Fe_2O_3 — content of free iron oxides

150–250 m) in the bottom of the Sivka River valley. The maximum depth of the Eemian ox-bow lake was 2–2.5 m; its erosional bottom is sandy or gravely. The interglacial sequence is represented by both mineral and biogenic deposits. These last consist of gyttja and peat, both very rich in plant macrofossils such as wood fragments (up to 30 cm long), cones, nuts, and seeds.

Profile 5 is typical of the littoral part of the ox-bow lake (Figs. 7, 9 and 10). The bottom 0.1 m thick layer of sandy muds is disturbed by load-cast structures and involutions (amplitude up to 0.3 m), probably of sedimentary origin, which penetrate into the overlaying middle layer that is 1.3 m thick. These layers are separated by thick and disturbed iron hardpan. The greenish sandy loam occurring above them gradually changes into dark grey silty mud. This deposit is rich in organic carbon compounds (up to 1%), and rust-coloured spots of iron oxides. The Eemian sequence is topped with peat 0.9–1.2 m thick (Fig. 11), dark brown to black, light, macroporous, layered, with secondary calcite crystals. Pebbles and gravel layers occurring within this unit indicate that river water periodically flowed into the ox-bow lake. Profile 3 (Fig. 12) represents a deeper part of the ox-bow lake, in which dark grey gyttja with a silty-clayey mineral component accumulated; the gyttja is horizontally layered in places, with gley spots, and in its bottom part is enriched with carbonate concretions. The fossil molluscs in gyttja indicate that this water body developed in two stages and in relatively warm climatic conditions (Alexandrowicz and Dmytruk, 2007). The mineral-peat unit in profile 5, and the gyttja in profile 3 are cut by a system of narrow fissures, oblique and almost vertical, filled with very fine orange sand. This indicates that before burial these deposits underwent intensive weathering and desiccation, and mineral material got into open fissures probably under periglacial conditions of the Earliest Vistulian. Traces of probably seasonal, deep frost penetration, i.e. structures of segregated ground ice (2–4 cm in diameter) occur in the gyttja, which contains hygroscopic clay, and in places it is brecciated as a result of frost weathering.

Casts of faunal structures up to 2 cm in diameter, filled with grey sand, found in the gyttja, reflect younger events and should be related to organisms living in the Early Vistulian soils.

The peripheral zone of the ox-bow lake is recorded in the deposits of profile 3A (Fig. 7). The compact, bluish-grey, strongly gleyed loam about 1 m thick is separated by a sharp boundary from the underlying Wartanian alluvium. This loam was probably deposited in the last part of the Eemian.

The lithologically variable Eemian deposits, which fill the fossil cut-off, can be arranged according to age on the basis of palaeobotanical data (Kalinovych, 2001, 2002). The pollen dia-

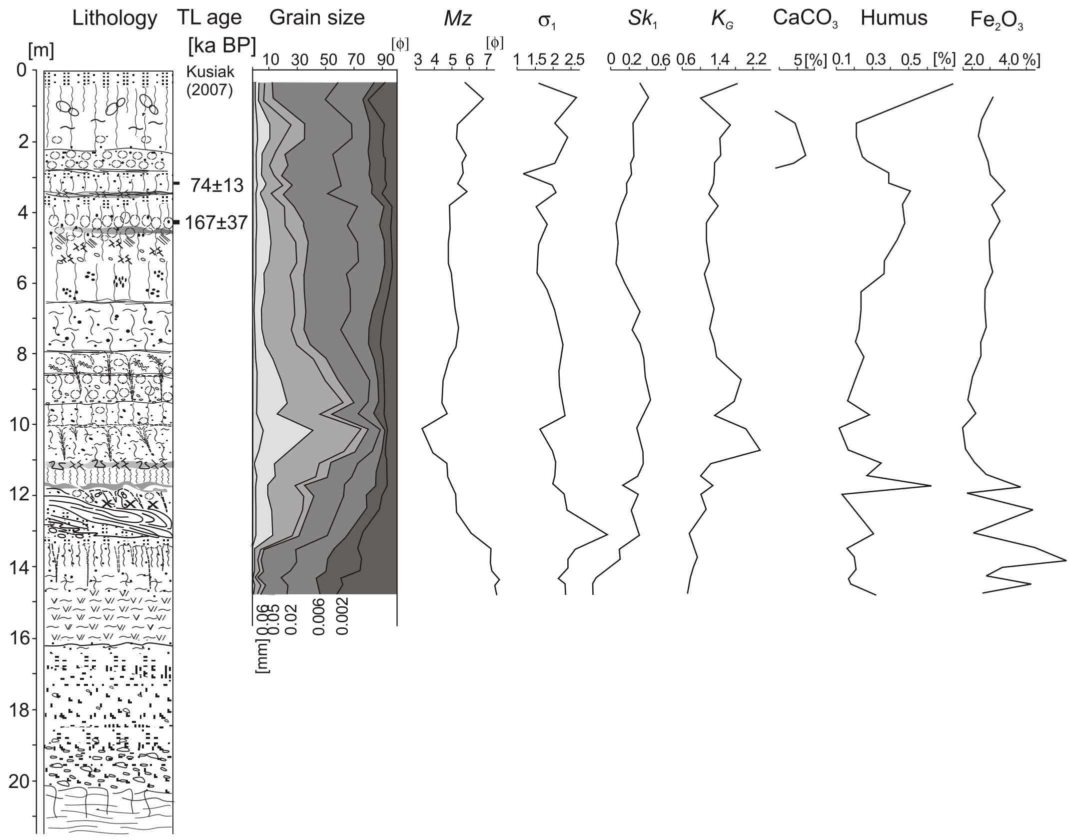

Fig. 10. Kolodiiv 3 profile

For other explanations see Figure 9

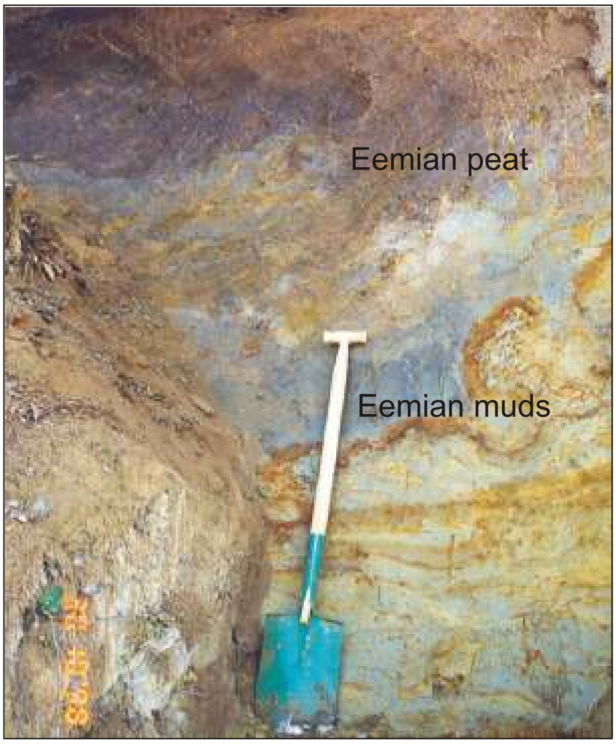

Fig. 11. The Eemian muds and peat in the Kolodiiv 5 profile

gram of gyttja from the Kolodiiv 3 profile is divided into two local pollen assemblage zones. The older one, a *Tilia–Corylus* L PAZ, is characterized by high pollen values of *Tilia* and *Corylus*, and considerable frequencies of *Carpinus*. Pollen grains of *Quercus*, *Ulmus*, *Acer*, and *Fraxinus* occur sporadically, while pollen of *Pinus* and *Picea* is absent. A considerable increase in *Alnus* values is found in the upper sample. The zone represents the interval, in which lime-hornbeam dry-ground forests predominated, with an admixture of oak, elm, and maple-tree. Hazel was abundant at forest margins or formed separate thickets. Following Mamakowa (1989), this zone is similar to *Corylus–Quercus–Tilia* R PAZ, and represents the optimum phase of the Eemian Interglacial. The younger local pollen assemblage zone (*Carpinus–Corylus*) is characterized by the continuing high pollen values of *Corylus*, a decrease in *Tilia* frequencies, and a simultaneous rise in the percentages of *Carpinus*. Pollen grains of *Quercus*, *Ulmus*, *Acer*, and *Fraxinus* occur sporadically. Values of *Alnus* are moderately high but variable. Therefore, one can conclude that forest communities were similar to these of the preceding zone but with a higher contribution of hornbeam. This is a counterpart of the *Carpinus–Corylus–Alnus* R PAZ of Mamakowa (1989).

The pollen diagram of peat from the Kolodiiv 5 profile is divided into two local pollen assemblage zones: *Carpinus–Alnus–Picea* and *Pinus–Picea–Betula*. The older one is represented by two lower samples, and is characterized by a predominance of *Alnus* pollen, considerable values of

Fig. 12. The Kolodiiv 3 profile: Eemian gyttja and loess cover

Carpinus and *Picea*, and low percentages of *Pinus*, *Abies* and herbs. Pollen grains of *Quercus* are infrequent, and these of *Tilia* occur sporadically. This pollen spectrum represents the last part of the Eemian Interglacial when hornbeam was replaced by boreal elements (alder, spruce, fir) in forest communities. Following Mamakowa (1989) this is a counterpart of the *Picea–Abies–Alnus* R PAZ. The younger local pollen zone is characterized by a predominance of *Pinus* pollen, continuing high values of *Picea*, and a decrease in *Alnus* frequencies. The pollen values of sedge rise considerably. Such a spectrum indicates an expansion of boreal forests, a higher proportion of sedge meadows, and corresponds to the *Pinus* R PAZ of Mamakowa (1989). Therefore, gyttja and peat deposits from the Kolodiiv profiles represent different periods of sedimentation. Gyttja is related to a relatively short period in the middle part of the interglacial optimum. Peat developed in the late interglacial phase, when climatic conditions deteriorated; only the lowest part of this deposit represents the end of the interglacial climatic optimum (Kalinovych, 2002).

EEMIAN-EARLY VISTULIAN SET OF PALAEOSOLS (OIS 5)

Following Martinson's model (Martinson *et al*., 1987), an age of 129.8 ka was assigned to the OIS 5/6 boundary, and an age of 73.9 ka to the MIS 4/5 boundary (Kukla *et al*., 1997). Different time intervals have been assigned the to Early Vistulian (OIS 5d–a): 115–73 (Liedtke, 1993), 115–75 ka BP (Haesaerts *et al*., 1999), 116(114)–73.9 ka BP (Kukla *et al*., 1997) and, 116–75 ka BP (Mojski, 1999). Within the Early Vistulian, a succession of alternately short cold and longer temperate phases has been documented (Paepe and Vanhoorne, 1967; Behre, 1989; Emontspohl, 1995; Bińka and Grzybowski, 2001; Günster *et al*., 2001; Guiter *et al*., 2003; Müller *et al*., 2003; and others). These phases are clearly recorded in long terrestrial pollen sequences in central and western Europe. Three interstadials have been distinguished, the Amersfoort, Brörup, and Odderade, but the Amersfoort is sometimes considered to be a warm oscillation at the beginning of the Brörup. According to west European data, the time interval of the Amersfoort together with the Brörup (OIS 5c) is 105–94 ka BP, and that of the Odderade (OIS 5a) is 82–75 ka BP (see Mojski, 1999), and after Haesaerts *et al*. (1999) these are 108–95 ka BP and 85–75 ka BP, respectively.

Referring to the name Horohiv, which is regularly used in the stratigraphic scheme of the loesses in the Western Ukraine (Boguckyj, 1986, 1987), we define the entire Eemian–early glacial soil succession as the Horohiv *sl*. (S1) stratigraphic unit, which represents OIS 5e–a. The Eemian part of the succession forms the Horohiv *ss*. (SS1) unit, which is attributable to soil formation during the Eemian period (OIS 5e). The early last glacial (OIS 5 d–a) deposits at Kolodiiv are subdivided into six palaeosol and loess units, each representing significant climatic oscillations. Equivalents of warmings are three interstadial soils (Kolodiiv 1, Kolodiiv 2, and Kolodiiv 3, i.e. S1-s1, S1-s2, S1-s3), which are correlated with the main interstadials of the Early Weichselian Glacial (Amersfoort, Brörup and Odderade).

The Horohiv *sl*. soil succession (S1) at Kolodiiv shows spatial differentiation, which indirectly reflects the microrelief of the contemporary terrace surface. This succession is complex and atypical. It contains two morphologically and typologically differentiated sets of palaeosols of different ages. The Eemian interval, corresponding to OIS 5e, is represented either by a single well-developed forest soil or two distinct complete soil units, also of forest type. Interstadial soils lie in direct succession on the Eemian palaeosol or are separated by thin mineral sediments, equivalents of cool episodes. However, these three soils are not present together across the exposure. In its southern part we find the entire set of three soils, while in the northern part some layers are absent, probably due to denudation processes as seen in the occurrence of intersoil solifluction layers.

Eemian palaeosol. Profiles with a single interglacial soil (SS1) occur in the northern part of the Kolodiiv site (Fig. 7; 1A and B profiles). This is an automorphic lessivé soil (to 2 m thick), which is characterized by a very well-developed Bt horizon (the so-called argillic horizon). The Bt horizon is enriched in clay (up to 18%), and cemented with iron oxides (up to 2.75%). The Eet horizon is moderately thick (0.2 m), bleached, impoverished in clay minerals, unconsolidated, and contains many black ferromanganese nodules (Boguckyj and Vološyn, 2006). A distinct boundary separates the Eet horizon from the

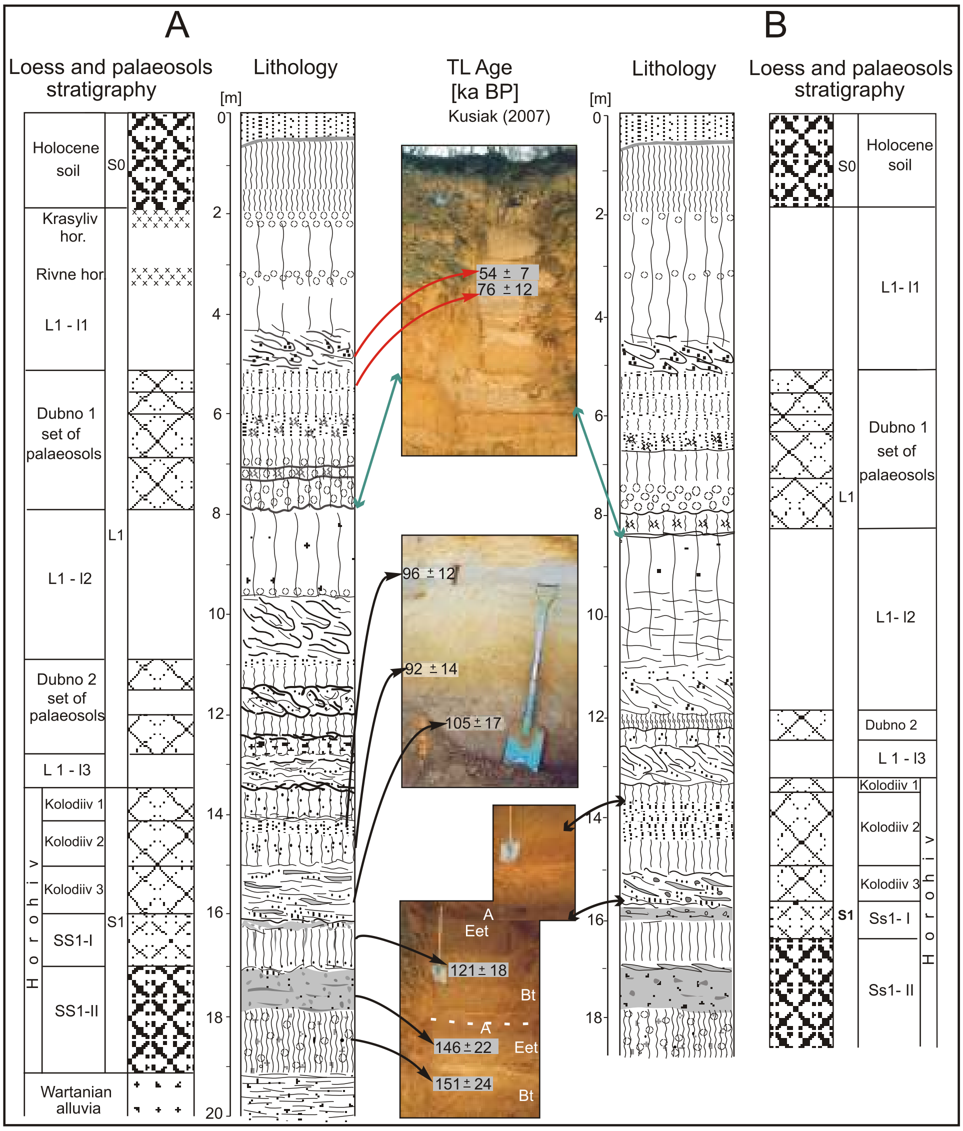

Fig. 13. Kolodiiv 2 profile: state of the exposure in 1999 (A) and 2003 (B)

A — humus horizon, Eet — eluvial horizon, Bt — illuvial horizon; for other explanations see Figure 9

superimposed humus horizon. We relate it to interstadial (Early Vistulian) pedogenesis occurring in open grassland.

Profiles 2 and 2 A (Figs. 13 and 14) contain a succession of two similar soils, both of interglacial type, and their total thickness is 4 m. The upper soil (SS1-I) is directly superimposed on the lower one (SS1-II), without a distinct boundary between them. They were probably associated with a local buried land depression. Each one contains a well-developed Bt horizon, diagnostic of lessivé soil, enriched in iron compounds (up to 2.5%), gleyed, with aggregations of iron-manganese concretions. This specific set of soils contains typical eluvial horizons with large iron-manganese concretions. These horizons are very well developed (in the lower soil about 1 m thick), with a platy structure, a low content of free iron oxides (1.6–1.7%), and with charcoal. The humus horizon of the lower soil is thin, with 0.3% of humus, and also with charcoal. It is macroscopically less visible because it has been transformed by subsequent pedogenesis. The interglacial set of soils described indicates that, within depressions on the Eemian palaeosurface on the Sivka River, the lower soil was buried and was thus isolated from the influence of living organisms and associated pedogenesis. Then — still under interglacial conditions — the upper soil developed on the freshly accumulated deposits.

The heavy mineral analysis results illustrate the complex structure of the interglacial part of the Horohiv soils (Fig. 14). Very resistant minerals constitute 70–80% of transparent heavy mineral assemblage. Together with garnets they are the dominant minerals, while the content of epidotes and pyroxenes is low. The heavy mineral resistance indices are high. Two subordinate boundaries occur within the succession. The first boundary separates a lower palaeosol of interglacial rank form an upper one, and is expressed by the changing ratio of zircon to rutile. The second boundary occurs between the Eemian and early glacial parts of the succession. It is recorded as a rise in

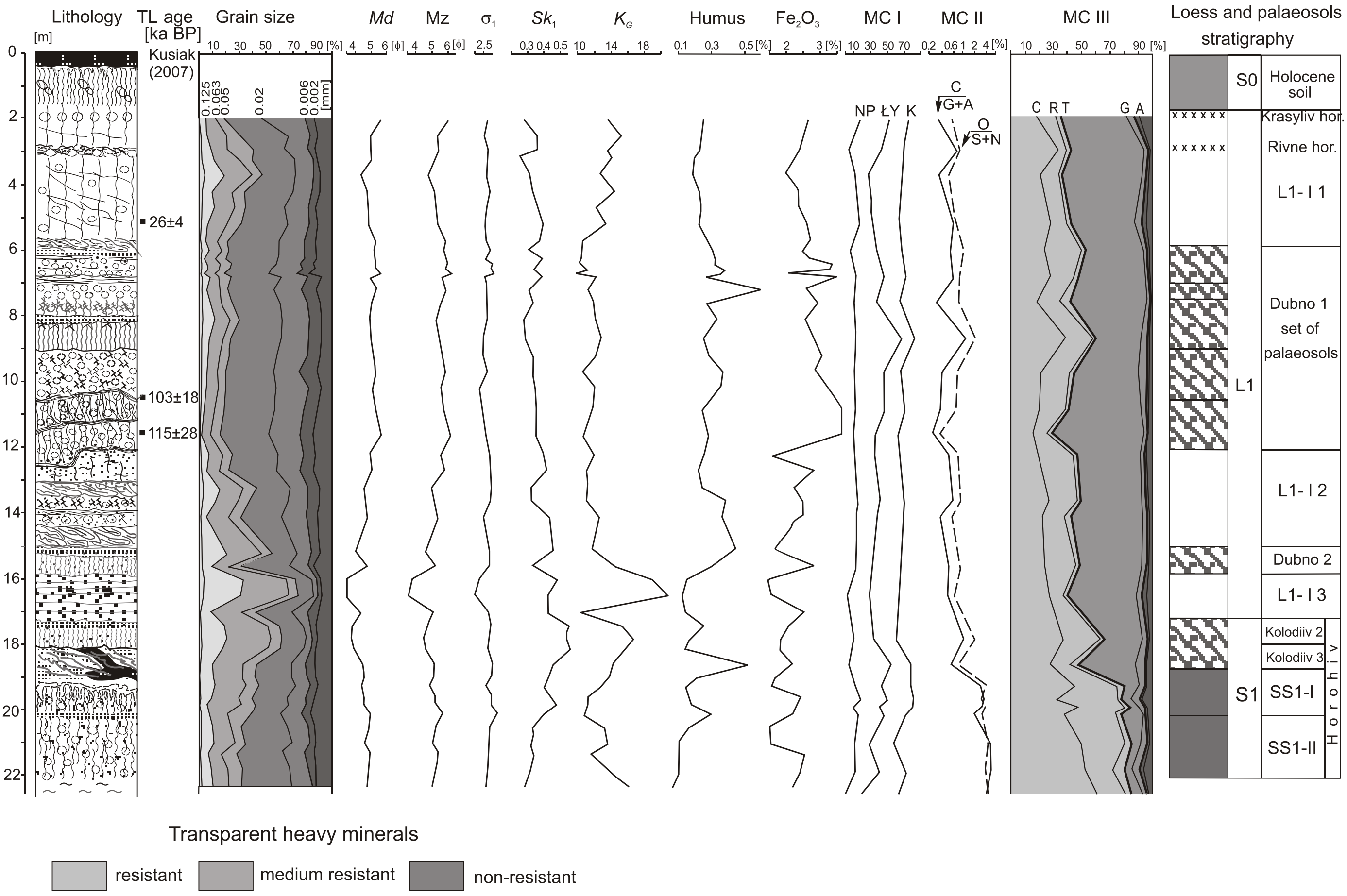

Fig. 14. Kolodiiv 2 A profile

Diagrams of heavy mineral composition: MC I — content of opaque minerals (NP), micas (ŁY) and concretions (K); MC II — indices of transparent minerals; MC III — transparent mineral composition; letter symbols of transparent minerals: C — zircon, R — rutile, T — tourmaline, G — garnets, A — amphiboles, O — resistant minerals, S — medium resistant minerals, N — non-resistant minerals; for other explanations see Figure 9

the content of garnet, and a simultaneous decrease in the percentages of zircon and rutile at the base of the Early Vistulian deposits. These changes correspond with lithological features indicating stratigraphic subdivision of the Horohiv succession (Racinowski, 2007).

Palaeomagnetic investigations of the Kolodiiv 2 profile showed an episode of reversed polarity, related to the Blake event, in the Bt horizon of the SS1-I soil. That is why this palaeosol was correlated with OIS 5e1 and why older palaeosol (SS1-II) was correlated with OIS 5e3 (Nawrocki *et al*., 2007).

The Eemian in north-central Europe was characterized by a relatively stable climate (Litt *et al*., 1996) but new investigations of some profiles within biogenic deposits provide evidence of climate change, with slightly cooler conditions in the middle and late part of the interglacial (Guiot *et al*., 1993; Guiter *et al*., 2003; Klotz *et al*., 2003). Such climatic fluctuations (OIS 5e4 and 5e2) were recorded in GRIP (Greenland Ice-Core Project Members) Eemian ice (Johnsen *et al*., 1995), and in deep-sea deposits (Martinson *et al*., 1987; Johnsen *et al*., 1992; Keigwin *et al*., 1994). Evidence for noticeable climate instability during the Eemian, but without extreme cooling, were provided by Klotz *et al*. (2003) based on European pollen data. Winter cooling within the *Carpinus* Zone (E5) was reported by Field *et al*. (1994) and Guiter *et al*. (2003) in the second part of the interglacial. Guiter *et al*. (2003) indicated a possible opening of the forests before the establishment of boreal coniferous forest that marked the end of the interglacial in central-west Europe. The inter-Eemian spell is dated at about 122 ka (Kukla *et al*., 1997). Disturbances of sedimentation were recorded in some pollen profiles of the Eemian Interglacial in Poland during the pollen phase E5 *Carpinus vel Carpinus–Corylus–Alnus* R PAZ (Mamakowa, 1989; Tobolski, 1991; Janczyk-Kopikowa, 1991; Niklewski and Krupiński, 1992; Kuszell, 1997; Granoszewski, 2003 and others). A significant indication of such processes was found in the Eemian palaeosol from the Tarnawce profile (situated at the Carpathian margin, about 155 km NW of Kolodiiv). The Eemian vegetational succession is very well represented in the Tarnawce pollen diagram but the phase of hornbeam forest development is not recorded, probably because of denudation of the deposits (Komar and Łanczont, 2002). Therefore, conditions favouring activation of erosional processes and supply of clastic material from slopes to sedimentation basins may have periodically occurred also at Kolodiiv in the later part of the

Eemian. These conditions may have been related to increasing continentality of climate, causing lower humidity, shortening of the growing season, and more rapid erosion. Such processes usually resulted in a partial removal of soil (as seen in the Tarnawce profile), and subsequent continuation of pedogenesis without visible changes in the soil profile. However, local extreme events at Kolodiiv (provoked by climatic disturbances) interrupted the pedosedimentary evolution because the soil was buried, and a new but similar soil developed subsequently.

In the west European loess sequences, a counterpart of the Horohiv *ss*. palaeosol is the level of intensive pedogenesis named the Rocourt soil in Belgium (Paepe and Vanhoorne, 1967; Antoine *et al*., 2001; Guiter *et al*., 2003), and PKIII in Czech Republic (Kukla and Ložek, 1961; Frenzel, 1964). These soils with a textural B-horizon resemble a present-day Luvisol (Haesaerts *et al*., 1999). They developed under conditions favourable to leaching.

In the Central Ukrainian loess profiles, Eemian soil is represented by a brown forest soil in the lower part of Pryluki horizon (Veklič, 1968; Gozhik *et al*., 1995). According to some authors (Gerasimenko, 2000; Rousseau *et al*., 2001), a grey forest soil, named the Kaydaki horizon (kd), represents OIS 5e, and the stadial-interstadial (loess-palaeosol) sequence named the Pryluki horizon (pl) is referred to OIS 5 a-c.

Early Glacial palaeosol set. In the southern part of the Kolodiiv site, where Eemian biogenic sedimentation took place, deterioration of climatic conditions between the Eemian and the last glacial, i.e. the Herning Stadial, equivalent to OIS 5d (Liedtke, 1993), is marked by a layer of carbonate-free loess of boggy or alluvial facies (S1-l3). Boggy loess occurs as greenish-grey, gleyed, silty-clayey loam up to 1 m thick (profiles 3, 3A, 4), and loess of alluvial facies is bluish-grey, sandy loam less than 0.5 m thick, with thin laminae of sand and rare gravel layers (Fig. 9, profile 5).

The Kolodiiv 3 unit (S1-s3), which developed over the above-described deposit, is a two-horizon soil over 1 m thick, composed of Ah (or Ahg) and B (or B-like layer of strong weathered loess) horizons. A dark grey humus horizon is moderately thick (0.2–0.5 m), contains up to 0.35% of humus, charcoals, and has been reworked by pedofauna. A pole colour visible in places in the bottom part probably resulted from initial leaching. This horizon was disturbed by intensive solifluction, and processes associated with seasonal (?) frost changed its structure. A dense network of fissures filled with grey, gleyed material runs downwards from the Ah horizon, and penetrates deep into the B horizon. These horizons are separated by a sharp denudational boundary. The B (?) horizon is 1 m thick, loamy-silty, bluish-grey, strongly gleyed, enriched with iron oxides, with orange and rusty-coloured streaks and spots, iron-manganese small concretions, and concentric structures of Liesegang ring type. According to Konecka-Betley, the micromorphological structure of this horizon, examined in profile 5, is not clear because features of the Bt horizon are absent (Łącka *et al*., 2007).

In profiles 1A, 2 and 2A (Figs. 7, 13 and 14) the Kolodiiv 3 unit is a humus horizon (chernozem-like) that varies in thickness, directly overlying the Eemian forest soil. It is moderately thin (0.3 m) and undisturbed in profile 1A, while in profiles 2 and 2A its thickness reaches over 1 m, it is stratified, gleyed, and disturbed by slope processes. A considerable admixture of sand indicates a marked activity of these processes, which were accompanied by wind action (the content of silt fraction in this horizon is over 20%). The humus content is moderately high (>0.5%). This horizon has some features of pedolith, i.e. soil developed on slowly accumulating/aggrading deposits; the sedimentation rate was so slow that soil formation, and especially humus accumulation, was not interrupted. A high activity of pedofauna, typical of grassland ecosystems, occurred in this soil as shown by many faunal tube casts, accentuated by the pale colour of the infill material. Many iron-manganese concretions and charcoals are also easily visible against the dark grey colour of the horizon.

Such charcoals, though redeposited, together with scattered flint artefacts belonging to the East Micoquian Culture, occur in the solifluction horizon overlying the Kolodiiv 3 soil in profile 3. The solifluction layer is characterized by variability in thickness from several to 60 cm, and a complex internal structure with interlayers and small lobes of sandy-loamy loess, sand, and material coming from the eroded Kolodiiv 3 soil. The archaeological find has all the features of the remains of the a Mousterian cultural layer, and shows distinct stratigraphic and typological similarity to the cultural layer II at the Yezupil site. This indicates Palaeolithic human occupation of the central part of the Eastern Carpathian Foreland at the beginning of the Vistulian stage (Cyrek and Sytnyk, 2002; Łanczont and Madeyska, 2005; Sytnyk *et al*., 2007).

The middle soil of the Early Vistulian, i.e. the Kolodiiv 2 unit (S1-s2), directly overlies the older soil or is separated from it by the solifluction layer mentioned above (e.g. in profiles 3A, 3, 4/5). The Kolodiiv 2 soil is a local stratigraphic marker at site owing to distinct changes in the soil substrate of the Ah horizon caused by fire (many clods of rusty-coloured, burnt loam; Fig. 15A). This soil resembles the older one as regards morphology and depth of profile but it is characterized by a higher content of humus (up to 1.2%), stronger spotty gleying, a slightly more distinct eluvial horizon, and a stronger transformation by pedofauna and by mammals (mole tunnels). An interesting epigenetic structure is developed in this profile at the boundary between the loess and the Kolodiiv 2 soil (Fig. 15B). This is a wedge-shaped fissure, over 1 m deep and, up to 0.3 m wide at the top, filled with organic (soil) and mineral (loess) material. The filling material is generally massive in the upper part of the fissure, and subvertically foliated in its lower part. The neighbouring layers are bent upwards in the contact zone with the wedge. We have determined that these fissures form polygons up to 6 m in diameter along the Kolodiiv outcrop walls. These dimensions are consistent with seasonal ground frost (Jahn, 1970).

The youngest soil in the succession, i.e. the Kolodiiv 1 unit (S1-s1), directly overlies the immediately older one (Kolodiiv 2) in all the places examined, apart from in profile 5 where a thin layer of sandy loess disturbed by solifluction separates the two soils. The full profile of the Kolodiiv 1 soil is preserved in the central and southern parts of the site (profiles 3, 4, 5; Figs. 7, 9 and 10). In comparison with older units, it is slightly less thick (0.8 m), with a lower content of humus, and without visible traces of leaching. Casts of faunal structures,

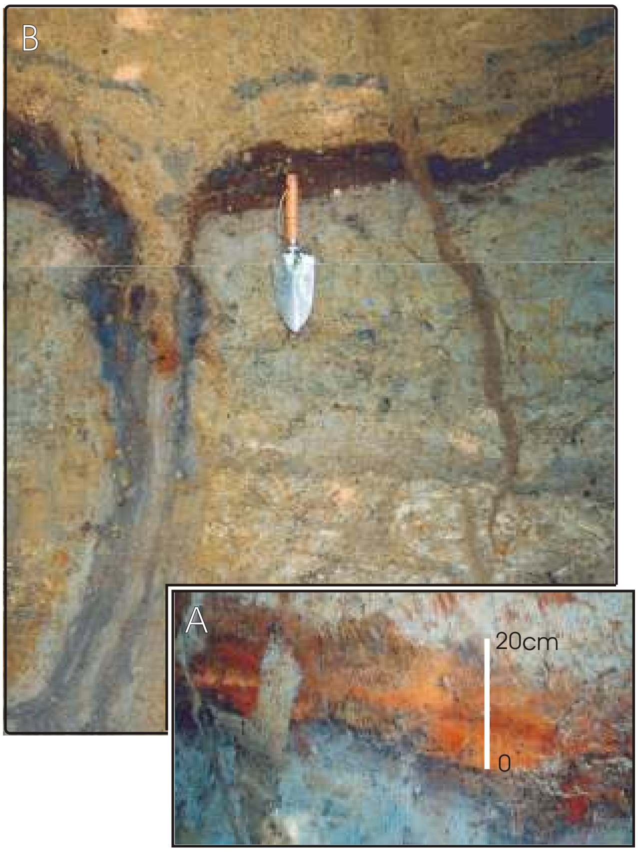

Fig. 15. The Kolodiiv 5 profile

A — the "fire" Kolodiiv 2 palaeosol; B — epigenetic fissure structure running downward from the top of this soil

also mole tunnels, are common features of all the Kolodiiv palaeosols.

To sum up, the earliest phases of pedogenesis following the Eemian Interglacial are represented by interstadial forest-steppe soils, which underwent evolution caused by change from a generally wet and warm boreal climate towards a drier continental climate. Most of these soils are gleyed, with a less or more distinct E horizon and a truncated B horizon; all these soils are carbonate-free. Traces of translocation downwards of clay and iron compounds (the content of iron oxides varies from 2.5 to 6%) are distinctly visible in the soil profiles. They are also characterized by the occurrence of numerous faunal structures. It seems that these soils may be of polygenetic origin, and can indicate different palaeoclimatic phases within each of the interstadials. We relate the development of the forest soils to the climatic optimum of the early glacial interstadials. In the last parts of the interstadials, biogenic soils developed under sparse-meadow or steppe vegetation (Łącka *et al*., 2007). They formed under conditions of the strong activity of a soil edaphone (an inference strengthened by the presence of mole tunnels), with a supply of fresh mineral material.

Some information about the early glacial interstadial flora of the Dniester valley around Kolodiiv comes from the Yezupil site. Komar (2002) gave the results of pollen analysis of a chernozem that probably corresponds to the Kolodiiv 1 horizon (OIS 5a). The upper part of chernozem horizon at Yezupil was TL dated at 78±11 ka BP, and the lower part at 85±13 ka BP (Łanczont and Madeyska, 2005). The spectra represent open pine-birch forests, with *Corylus* and *Salix*, and a small admixture of *Quercus* and *Carpinus*. *Poaceae*, *Asteraceae* (including *Artemisia*), Chenopodiaceae, Cyperaceae and *Helianthemum* are important components of the NAP (non arboreal pollen). They form open steppe communities. A mosaic of forest patches and open places with steppe vegetation characterized the landscape. Wet places were overgrown by meadow communities with sedges (Łanczont and Madeyska, 2005).

Each of three pedogenesis stages at the Kolodiiv site ended with climatic deterioration. Aeolian processes were not intensive. It may be supposed that permafrost was absent, only long-lasting, deep seasonal ground frost occurred, which caused the development of small reticulate structures of segregated ground ice. Runoff was very active during these cold oscillations. The cold phase following the formation of the Kolodiiv 3 soil was rather wet, and solifluction developed. During the next, perhaps more continental, cold phase (after the formation of the Kolodiiv 2 soil), contractional fissures (Fig. 15B), and structures of subsidence and melting of seasonal ground frost developed (Fig. 16).

The results of TL dating of these deposits at Kolodiiv are shown in Table 1. The TL age obtained for the Kolodiiv 3 soil in profiles 1A, 3, and 4 is too old. The TL age obtained for this soil in profile 2 (Fig. 13), in which it has more distinct features of a pedolith enriched in fresh aeolian material, corresponds to the age of the Amersfoort Interstadial (Bosinski, 2001; Guiter *et al*., 2003). The TL age for the Kolodiiv 2 soil in profile 2 (92–96 ka BP) is as expected, and in profiles 3 and 4 it is too old (132–121 ka BP). The youngest soil, i.e. the Kolodiiv 1 unit, was TL dated at 95±15 ka in profile 3, so this age is also too old. The deposits of cold phases are dated at 164–203 ka BP (the phase directly following the Eemian), and at 111–117 ka BP (the phase between the Amersfoort and Brörup).

Only some of the dates given here are consistent with the stratigraphic interpretation. Some of them are too old by far. However, despite the short distances between individual sites, sedimentation conditions were spatially variable. In more open places, more distant from the penecontemporaneous slope, sedimentation was dominated by aeolian processes (profile 2), while the deposits that accumulated near the slope consist also

Fig. 16. The Kolodiiv 4 profile

The "fire" Kolodiiv 2 palaeosol with load-casted structure pressed into the B horizon to a depth of 30 cm

Table 1

TL dating of palaeosol and loess samples taken from the deposits of OIS 5 (obtained by Kusiak)

Profiles	Stratigraphic units		TL age [ka BP] Kusiak (2007)
	Loess units	Palaeosol units	
3		Kolodiiv 1 palaeosol (S1-s1)	95±15 (Lub-3688)
2		Kolodiiv 2 palaeosol (S1-s2)	96±12 (Lub-4014) 92±14 (Lub-4015)
4			121±18 (Lub-3353)
3			132±19 (Lub-4172)
1A	S1-l2		111±21 (Lub-3351)
3A			117±32 (Lub-3688)
2		Kolodiiv 3 palaeosol (S1-s3)	105±17 (Lub-4016)
3			164±15 (Lub-4173)
4			146±20 (Lub-3522)
1A			148±26 (Lub-3352)
4	S1-l3		164±26 (Lub-3523)
3			183±18 (Lub-4174)
			203±24 (Lub-4175)
2		SS1-I (Bt horizon)	121±18 (Lub-4017)
2		SS1-I (Eet horizon)	146±22 (Lub-4018)
2		SS1-II (Bt horizon)	151±24 (Lub-4019)
1A		Horohiv ss. (Bt horizon)	164±30 (Lub-3353)

of slope material (Łącka *et al*., 2007), probably redeposited over a short distance as rapid sedimentation events, often as large mass movements (solifluction lobes). This phenomenon is strongly marked in the deposits accumulated in the meander cut-off. At the beginning of the early glacial, the deposition of fresh silt was accompanied by that of aeolian material derived from wind-eroded older loess, which occupied higher Pleistocene terraces, and deluvial material from strongly eroded Eemian soil. These mixed deposits formed the parent material of the Kolodiiv 3 soil so that the TL age obtained is considerably older than the depositional age.

The stratigraphical equivalent of the Kolodiiv horizons is grey forest soil (Terhorst *et al*., 2001) or greyzem (Antoine *et al*., 2001), found in the early glacial part of loess-palaeosol sequences in western Europe. This indicates a forested continental environment with cold snowy winters and warm summers (Haesaerts *et al*., 1999). At present grey forest soils are zonal soils of the temperate reach of boreal climate, which develop under deciduous forest communities alternating with meadow or meadow-steppe ones (Gerasimova *et al*., 1996).

In the Central Ukrainian loess profiles, the EarlyVistulian is represented by chernozems (Veklič, 1968) or by three subunits of the Pryluki soil succession, i.e. a meadow soil, a brown forest soil, and a humic chernozem (Gerasimenko, 2000; Rousseau *et al*., 2001; Fig. 8). Thin loess interbeds occur between these soils in the most complete profiles (Gozhik *et al*., 1995). Climatic reconstruction based on pollen data from loess profiles (Bolihovskaja, 1995; Gerasimenko, 2000; Komar, 2002) indicates that chernozems developed under alternating conditions of dry steppe climate, and warmer interstadials, in the optimal parts of which deciduous forest grew.

LOESS OF THE LOWER PLENIGLACIAL (OIS 4)

The Lower Pleniglacial corresponds to OIS 4 dated at 73–61 ka BP (Liedtke, 1993), 75–58 ka BP (Mojski, 1999), or 72–60 ka BP (Guiter *et al*., 2003). This is the oldest period of maximum cooling in the last glaciation (Guiter *et al*., 2003). It was probably the first period of four major ones of loess accumulation, which occurred in the western Europe during the last glacial, with luminescence age estimates ranging from 67 to 60 ka BP (Frechen *et al*., 2001). The central European pollen data document the main opening of the vegetation, and suggest the presence of cold steppe biomes in this interval (Vandenberghe, 1992; Liedtke, 1993; Bolihovskaja, 1995; Müller *et al*., 2003).

In all profiles of the Kolodiiv site, the boundary between the deposits representing OIS 5 and OIS 4 is sharp and erosional. We infer that the fine, yellow aeolian silt that fills fissures which cut the Kolodiiv 1 palaeosol in profile 5, may be an equivalent of the marker bed of fine aeolian silt deposited by continental-scale dust storms at the boundary between OIS 5 and OIS 4. This phenomenon is suspected as the cause of rapid environmental degradation then, associated with a sharp transition from the rich grasses and forests of the early glacial to the barren landscapes of the Pleniglacial (Kukla and Ložek, 1961; Rousseau *et al*., 1998; Antoine *et al*., 2001). In the heavy mineral composition (Fig. 14) there is a conspicuous inversion of the proportion between resistant minerals and the rest, which is marked at the boundary between the deposits of OIS 5 and OIS 4.

The Lower Pleniglacial is represented at the Kolodiiv site by a loess-like deposit (L1-l3), 0.5–3 m thick, of deluvial- and solifluction-aeolian facies. This is sandy and silty loam (Mz = 3.9 ϕ) with sand laminae. This carbonate-free deposit is characterized by low contents of humus (0.13%) and iron oxides (1.8%). In the heavy mineral assemblage, garnet (about 50%) exceeds zircon (<30%), and rutile (about 10%). Amphiboles, staurolite, biotites, and tourmaline occur in small quantities (Fig. 14). The high content of garnet suggests that the source material of the loess from the Lower Pleniglacial at Kolodiiv was formed under conditions favouring enrichment in garnet, i.e. in a fluvial environment. This indicates that periglacial alluvium played an important role as a source of fresh silt (Łanczont, 1995). The deflated loesses of older covers, spread on high terraces (V and VI) of the Dniester River near the Kolodiiv site, may have been an other source of dust.

The Lower Pleniglacial Loess has not been dated at Kolodiiv but we have its TL age (63±9 ka BP — Lub-3994) from the nearby profile at Yezupil. At Kolodiiv, this loess bed appears stratigraphically uniform. Signs of weak tundra biogenesis, large iron-manganese concretions, manganese spots, and initial Liesegang rings are scattered. Separate interphase gleyed horizons, which occur in some loess profiles in Poland (Maruszczak, 1985; Łanczont and Alexandrowicz, 1997; Dolecki and Łanczont, 1998), in Ukraine within the Uday Loess unit (ud) (Gozhik *et al*., 2001) and in west-central Europe (Guiter *et al*., 2003), are absent at Kolodiiv.

Interphase tundra horizons representing a subarctic climate within the loess of OIS 4 are named the Niedereschbacher Zone in Germany (Semmel, 1967), and the Ognon 1 and Ognon 2 horizons in France (Woillard, 1975; Van Vliet-Lanoë, 1987).

Based on the curves representing the intensity of aeolian dust accumulation in the Late Pleistocene, Hovan *et al*. (1989) and Jouzel *et al*. (1993) deduced the occurrence of oscillations during its deposition in OIS 4.

LOESS-SOIL SEQUENCE OF THE MIDDLE PLENIGLACIAL (OIS 3)

The next stratigraphic unit of the last glacial, i.e. the "Interpleniglacial" or "Middle Pleniglacial", is the equivalent of OIS 3. Different approximate time intervals for this unit have been published: 62–26 ka BP (Vandenberghe, 1992), 61–27 ka BP (Liedtke, 1993), 58–25 ka BP (Mojski, 1999), 65–25 ka BP (Arnold *et al*., 2002) and 60–29 ka BP (Guiter *et al*., 2003). This long interval was generally less cold than the preceding one (Guiter *et al*., 2003). It included climatic oscillations marked by numerous rapid switches between brief cold and longer warm times (Vandenberghe *et al*., 1998; Vandenberghe and Nugteren, 2001) known as Dansgaard-Oeschger (DO) events (interstadials), registered in the GRIP (Greenland Ice-Core Project Members) ice core (Dansgaard *et al*., 1993; Frechen *et al*., 2001). It is generally characterized by a decrease in, or cessation of, loess sedimentation and terrestrial soil-formation (Arnold *et al*., 2002).

The rhythmic stratigraphy and chronology of this interval on the Russia Plain (Zarrina *et al*., 1981) is expressed as the following stadials: Lower Graždansky (^{14}C dated at 49–40 ka BP), Middle Kašynsky (37.5–34 ka BP), and Upper (bipartite) Dunajevsky (32.5 to about 24 ka BP). In the loesses of the Dnieper region, the Middle Pleniglacial is represented by the Vytačiv set of 2–3 palaeosols (vtb_1 and vtb_2) separated by a loess bed (vtb_{2-1}), which was TL dated at 44–35 ka BP, and ^{14}C dated at 33–30 ka BP (Gozhik *et al*., 2001). In the light of palynological data (Rousseau *et al*., 2001), in the generally milder conditions of this interval, vegetation changed from an open *Pinus* boreal forest (lower soil), which included a few deciduous trees, through sparse steppe (intrasoil loess), to more productive forest-steppe mosaic (upper soil). In loess profiles of Volhynia and Podolia, the Middle Pleniglacial is represented only by the Dubno palaeosol. Pollen data from this area indicate a mosaic vegetational pattern of shrub tundra, steppe-tundra, and steppe with patches of forest-tundra, and a short expansion of boreal forest with pine and birch during the climatic optimum (Bezus'ko *et al*., 1989). The ^{14}C age of the Dubno soil, estimated at 28–29 ka BP (Bezus'ko *et al*., 1989), indicates that it probably corresponds to OIS 3.1. However, based on new results of TL dating in the Rivne profile at Volhynia (43±6 and 61±7 ka BP), this soil can be related to OIS 3.3 (Nawrocki *et al*., 2003). A similar stratigraphic correlation is also suggested by the TL ages of the Dubno soil at Yezupil: 40±5 (Lub-3992) and 61±8 ka BP (Lub-3993). It should be stressed that the older of these ages were obtained from the B horizon of this soil, i.e. they indicate the age of the substrate, which formed during the Lower Pleniglacial. The younger results represent the older part of the Interplenivistulian, when the A horizon developed.

At the Kolodiiv site, the greatest stratigraphic variability of the Middle Pleniglacial deposits is found in profiles 2A, 2, 3, and 4/5. Profile 2 was investigated twice, in 1999 and in 2003 (Fig. 13). The second time, the exposure wall was shifted about 1–2 m farther in as a result of cleaning of the face, and the exposed sequence appeared to be somewhat different from that described in 1999, both in respect of the number and the thickness of the layers distinguished. This indicates the effect of a distinct palaeosurface microrelief, simultaneously a cause and an effect of spatial variability of morphogenetic and soil-forming processes, which acted on this surface.

The sequence deposited at Kolodiiv in the Middle Pleniglacial is about 6–7 m thick, and includes two sets of palaeosols (L1-s2 and L1-s1) typical of tundra landscape, which are separated by a loess bed (L1-l2). These deposits are wholly carbonate-free. Gravels, which sporadically occur throughout the profile, were redeposited from higher parts of the slope (higher terraces). Thus, slope processes considerably influenced the formation of these deposits (Łącka *et al*., 2007).

The bottom soil of this sequence shows spatial variability conditioned by the palaeomorphology of the Kolodiiv site. Two types of soil are found; each truncated by erosion and covered with solifluction loess:

1. Wet tundra soil (gleysol) up to 1.5 m thick, developed under wet conditions with periodic supply of loess and slope (solifluction) material. This is a one-horizon soil, loamy-sandy, with nests and lenses of pure fine sand visible mainly in its middle part, and with traces of different cryogenic deformations of solifluction type (plication deformations). Traces of a well-developed small reticulate structure of ground ice occur in places at the bottom of the soil. Rusty-coloured spots, ferruginous streaks, and iron-manganese concretions are also found.

2. Authomorphic soil associated with higher situated, better drained relief elements. This is a distinctive subarctic brown cambisol, with a thickness of 0.6–0.8 m, in which the Ah horizon constitutes half the thickness. This soil is gleyed from the top, rather rich in humus (up to 0.5%) and in iron compounds (about 3%), with numerous iron-manganese and iron concretions.

In profile 2, 1999 only, this stratigraphic unit was found as two similarly developed brown soils separated by a solifluction layer.

The middle unit, i.e. an intersoil loess, with a maximum thickness reaching 3 m, contains a considerable amount (45–50%) of loess with a mean grain diameter ranging from 5.4 to 4.6 ϕ; the deposit is poorly to very poorly sorted (σ_1 = 1.6–2.2), with positive and very positive skewness (Sk_1 = 0.2–0.5), and with a very variable kurtosis index (K_G = 1.2–2.1). The contents of iron oxides (1.5–3.15%) and humus (0.15–0.37%) are variable, with most humus in the lower layers, which contain an admixture of material from the underlying soil. The loess is reworked by solifluction in the lower part of this unit, and by sheet-wash in its middle part as seen in rhythmic parallel or almost parallel bedding, irregularly undulatose, and inclined down palaeoslope. As well as synsedimentary structures, discontinuous postsedimentary ones are present that disturb stratification. Specific agglomerations of iron and iron-manganese concretions, and larger structures resembling initial Liesegang rings, which occur throughout the loess unit, may be related to tundra biogenesis.

The upper set of palaeosols is composed of two (profile 4/5) or even four to five tundra soils lying in direct succession (profiles 2A, 2, 3). These soils generally resemble the older ones in respect of typology. Their sequence is regular. The lowest is a thick gelic gleysol, which is characterized by the occurrence of thick limonitic hardpans in its bottom and middle parts. This palaeosol is overlain by a series of several (up to 4) similar, thin, two-horizon, weakly developed soils of cambisol type (Łącka *et al*., 2007). The content of humus is higher in cambisols (up to 0.6%) than in gleysols (0.25–0.4%) but free iron oxides occur in similar quantities in both soil types (3–4%). It seems that this set of soils developed in an interval comprising alternating, very short, slightly warmer and colder phases. Cold phases were characterized by weak deposition of loess alternating with erosion and redeposition of sediments, so the TL ages for the samples from this soil set in profile 3 are too old (74 and 167 ka BP). The material building these layers derived probably from nearby, from eroded older loess deposits, so it was insufficiently bleached before deposition due to a short transport distance. Thus, this lack of reliable dating results for these deposits provides indirect information about environmental conditions during their formation, and also about the climate during the younger part of OIS 3.

The heavy mineral assemblage of the loess-soil sequence from the Middle Pleniglacial resembles that from the loess representing OIS 4. It is characterized by a garnet content reaching 50%. The main minerals form the following sequence: garnets>zircon>rutile. Staurolite, tourmalines, amphiboles, and pyroxenes are accessory components. A lower intensity of weathering causes a decrease in the values of the mineral indices: O/S+N = 0.25–1.2, and C/G+A = 0.4–1.5. Within these intervals, higher values of deposit maturity indices are found in those loess layers affected by pedogenesis.

The Middle Pleniglacial loess-soil sequence at Kolodiiv reveals several climatic oscillations during this period. Its great thickness and pedostratigraphical variability are unique features in comparison with other Ukrainian loess profiles, but resemble those recorded, for instance, in northern Europe (Guiter *et al*., 2003). With reference to the stratigraphic scheme of the western Ukraine loesses (Boguckyj, 1987), we named the palaeosols of the lower set Dubno 2 (Dubno — older palaeosol(s)), and those of the upper set Dubno 1 (Dubno — younger palaeosol(s)).

The Dubno 2 palaeosol/palaeosols, which reflect the first climatic improvement during the Middle Pleniglacial, probably correspond to the Oerel and Glinde interstadials of the west European stratigraphic schemes (Behre and Lade, 1986; Liedtke, 1993; van der Hammen, 1995). According to the calibrated dating results from the Oerel site in Northern Germany (Behre and Van der Plicht, 1992), these interstadials occurred at 58–54 ka BP and 51–48 ka BP, respectively. The solifluction loess separating the two palaeosols of the Dubno 2 set probably correlates with the Ebersdorf Stadial (Liedtke, 1993), and with the second period of loess accumulation in western Europe (Frechen *et al*., 2001). If we agree with the opinion of Guiter *et al*. (2003) that the two episodes called Oerel and Glinde may respectively correlate with Ognon 1 and Ognon 2, we should recognize the Dubno 2 set as interphase soils corresponding to OIS 4. However, this seems to be a wrong conclusion because the Dubno 2 palaeosol/palaeosols reveal much more intensive pedogenesis than could have developed during the short warm interphase episode/episodes of the cold OIS 4.

The lower Dubno 2 unit is separated from the upper Dubno 1 unit by a loess bed. It indicates a cooler stadial, which may be correlated with the Lattrop Stadial (Liedtke, 1993), and so comparable to the third period of loess accumulation in western Europe (Frechen *et al*., 2001), most likely between 50 and 40 ka BP (in the light of luminescence dating).

It seems that the Dubno 1 set of palaeosols belongs to the younger part of the Interpleniglacial, with a series of warm oscillations (Van der Hammen *et al*., 1967; Zagwijn, 1974; Van der Hammen *et al*., 1995), which contains the long Moershoofd Interstadial Complex (50–43 ka BP), and two shorter but intense interstadials: Hengelo (38–37 ka BP) and Denekamp (34–29 ka BP) (Guiter *et al*., 2003). These are separated by the Hasselo and Huneborg stadials, respectively (Liedke, 1993). In the opinion of Van der Hammen *et al*. (1995), a few minor oscillations can be found within the Huneborg Stadial and the Denekamp Interstadial.

Based on the type and intensity of pedogenesis at Kolodiiv, it may be that the climate was wetter and colder during the Moershoofd Interstadial than during the succeeding warm periods. During the Hengelo and Denekamp interstadials, the climate was probably more variable, and numerous organic layers formed in the mineral deposits.

In comparison with the entire Kolodiiv site, profile 1B is atypical because the part corresponding to the Middle Pleniglacial is reduced. Above the loess correlated with OIS 4, one well-developed gley horizon occurs, which is overlain by massive loess with features typical of the Middle Pleniglacial Loess. Two TL ages were obtained for the same sample from this gley horizon: 46±6 and 50±7 ka BP.

LOESS OF THE UPPER PLENIGLACIAL (OIS 2)

The Upper Pleniglacial (equivalent of OIS 2) was the coldest period of the last glacial, representing the strongest aridity (Frechen *et al*., 2001; Guiter *et al*., 2003). In the opinion of Antoine *et al*. (2001), a thick loess deposits of this age, which occurs in the upper Rhein valley, shows three pulses of loess sedimentation at 27–25, 22–19, and 16–15 ka BP. Based on investigations of the west European loesses (Frechen *et al*., 2001), it can be deduced that the accumulation of the upper young loess probably started after the Heinrich 3 event and lasted to the Heinrich 2 event, i.e. from 25 to 20 ka BP. Also Greenland ice records indicate a major period of dust accumulation during this time (Bond *et al*., 1992, 1993; Grootes and Stuiver, 1997). The youngest loess-like sediment accumulated between 17 and 14 ka BP (Frechen *et al*., 2001). Two phases of transitory climatic improvement are recorded in western and central Europe on sequences at *ca*. 23–22 ka BP and between 16–15 ka BP (Lascaux warming) in the Upper Pleniglacial (Guiter *at al*., 2003).

Uniform, typically developed aeolian loess (L1-l1) 2–6 m thick accumulated at the Kolodiiv site during OIS 2. The heavy mineral assemblage of this loess resembles that of the older unit but contains less rutile, and higher percentages of non-resistant minerals (amphiboles, pyroxenes, and biotite). On the basis of the vertical variability of grain-size distribution of this loess in profile 2, three cycles of deposition can be distinguished, the middle cycle being characterized by the most distinct, short-term oscillations of environmental dynamics (Frankowski *et al*., 2007).

Poorly marked solifluction structures occur in the bottom layers (Figs. 13 and 14), which were formed in an open environment of tundra type, under cold and humid climatic conditions (Alexandrowicz and Dmytruk, 2007). Solifluction supplied older material to the aeolian cover being formed, so the TL age of this layer was too old. A sample taken from the solifluction loess was TL dated at 54±7 ka (Fig. 13).

The other, younger part of loess L1-l1 contains a large amount of coarse silt (about 36%), an admixture of very fine sand (about 20%), and of the < 0.002 mm fraction (about 10%). The mean grain diameter of this loess falls in the middle of the grain-size range of typical loess, i.e. between 5 and 6 ϕ. This is carbonate loess but the content of $CaCO_3$ does not exceed 7–8%. A sample taken from the typical loess was TL dated at 26±4 ka (Fig. 14). The depositional environment of this loess is characterized by a mollusc association indicating cold and dry climate, and an open environment of arctic steppe type (Alexandrowicz and Dmytruk, 2007).

Two gleyed horizons occurring within this loess in profiles 2, 2A, and 4/5 indicate more intensive pedogenesis. Gleyed, clayey loess forms these horizons. Each of them is 0.2–0.5 m thick, often with thin sand laminae and plication deformations, which suggest phases of slightly milder or rather wetter climate. By their intraloess occurrence they resemble the Rivne and Krasyliv horizons distinguished in the Volhynian and Podolian loesses. However, the stratigraphic position of the Rivne horizon, which has been accepted until now on the basis of the investigations of the Volhynian and Podolian loesses, is questionable (Maruszczak, 1995, 1996). Based on new TL dates in the Rivne and Torčyn profiles, this soil can be correlated with OIS 3.1 (Nawrocki *et al*., 2003).

The horizon, named Rivne to conserve terminology, occurs rather frequently in the loesses of the Halyč Prydnistrov'ja region (Łanczont and Boguckyj, 2002) as weak gley palaeosols, weathered horizons or layers with solifluction structures. The vegetation during this phase may have been subarctic grassland or steppe (Komar, 2002). Rich archaeological materials of the Gravettian Culture are associated with this horizon in the Palaeolithic sites at Halyč and Mežygirce. At Halyč they occur just below the soil, and at Mežygirce within the Ah horizon of the Rivne Unit. This difference is supported by radiocarbon ages, which range between 25.2 and 19.6 ka BP at Halyč, and from 17.2 to 20.4 ka BP at Mežygirce (Cyrek and Sytnyk, 2002). The TL ages of loess near the artefacts obtained in the Lublin Laboratory are 17.7–20.1 ka BP (Łanczont and Bogucky, 2002), and those obtained in the Gdańsk Laboratory are 20.8–21.8 ka BP (Fedorowicz *et al*., 2003). This horizon probably corresponds to so-called Eben-Zone distinguished in the Upper Pleniglacial Rhein loess, and dated at about 18 ka BP (Schirmer, 2000). The Upper Pleniglacial Erbenheimer palaeosol succession, which is developed in the Central Germany loess area, is similarly TL dated at about 17.5–22 ka BP (vide Terhorst *et al*., 2001).

In the Volhyno-Podolian Upland, the Krasyliv horizon is characterized by a polygonal network of large ice-wedge casts, which reach to a depth of 6 m, and were formed in the most severe climate of the Upper Pleniglacial. However, such forms are absent from the East Carpathian Foreland. At Kolodiiv, the Krasyliv horizon occurs just beneath Holocene soil (profile 2) or is undistinguishable where transformed by the Holocene pedogenesis.

CONCLUSIONS

The unique loess records from the Kolodiiv site provide key information for the correlation of European loess stratigraphy. They represent a sequence in which the climatic changes over the last 120 ka seem to be recorded. The sequence is essentially continuous and has a soil of the last interglacial and organic deposits at its base. The record of regional environmental changes during OIS 5 and OIS 3 is especially detailed and complete in comparison with loess profiles of the Dnieper River basin.

Detailed multidisciplinary study of the loess-palaeosol sequence at Kolodiiv leads to the following conclusions:

In the early glacial, aeolian accumulation took place (under conditions of seasonal ground frost) but it was insignificant; typical loess accumulation occurred during the Pleniglacial (OIS 4–2) in three main stages (probably under conditions of discontinuous, and then continuous permafrost). At Kolodiiv, loess was deposited on the slope exposed to the west, and it was a westerly wind, which carried dust as inferred from, for example, the spatial variability of mineral composition in loesses spread over an extensive area, from Volhynia to the East Carpathian Foreland (Chlebowski *et al*., 2004). That direction of transport is also indicated by the mean grain diameter, which is progressively smaller from the west to the east in the adjacent Podolian loesses (Tokarski, 1936). In the Sivka River valley at Kolodiiv, loess accumulated on different relief elements. Aeolian silt originated mainly from local source areas. When redeposited downslope, it was enriched in material from up slope (including from older terraces). Several factors favoured loess accumulation in the vicinity of the Kolodiiv site, i.e. the windward and rather "warm" exposure of a long slope, which was steep in its upper part, and gentle near the terrace, and a nearby source of silt. Older deposits were preserved, and younger ones accumulated at the same place. This slope loess sequence was sensitive to climate oscillations.

Four types of palaeopedological taxa have been differentiated within the loess terrace cover at Kolodiiv:

— an interglacial (Eemian), soil succession which includes one or two soil-forming episodes;

— interstadial one- or two-horizon palaeosols over 1 m thick;

— interstadial thin two-horizon palaeosols;

— monogenetic incipient/embryonic palaeosols.

Palaeosols from the first and second group form the Horohiv, soil succession distinguished in the stratigraphic schemes of loesses in the western part of Ukraine, and correlated with OIS 5. Palaeosols from the third and fourth group occur as different soil types within the Dubno Unit (also recognized in the west Ukrainian loesses), which corresponds to OIS 3.

The Eemian soil succession forms a catena pattern with the organic deposits having accumulated in an ox-bow lake. These deposits represent only fragments of the Eemian succession because of hiatuses, which could have been because the ox-bow lake was small, rather shallow, but periodically water flowed in it as it intermittently joined with river channel. The soil succession occurring in the lower part of the section is unique. Two similar and undoubtedly interglacial forest soils lying in direct sequence are the first palaeopedological evidence of Eemian climatic disturbances, recorded in pollen profiles, and represented by an episode of increased soil erosion resulting in the absence of the hornbeam forest phase. The next important erosion episode occurred towards the end of the Eemian, when the top part of the upper forest soil was truncated.

There was found a succession of three soil-forming processes separated by cold episodes correlated with OIS 5d–a. In the case of the Kolodiiv there was not found the typological differentiation of soils such as the succession of meadow soil, brown soil, and chernozem in the sub-units of the Pryluki horizon of the Central Ukraine (Rousseau *et al*., 2001). At Kolodiiv, each of the warm episodes was complex, and includes an earlier forest phase of taiga type with lessivé soil, which was replaced by a younger one with steppe or meadow soil. Erosion developed between these pedogenetic stages. During cold phases, solifluction and surface wash processes were more intensive than aeolian ones, so the deposits and soils developed on them are difficult to date by TL. Permafrost was absent but deep seasonal ground frost caused the formation of segregated ground ice and the development of cryo-desiccation cracks.

The Middle Pleniglacial is represented by a dozen or so climatic oscillations of very unequal duration. The climatic heterogeneity of this interval is very well reflected in the variation in grain-size distribution through this stratigraphic unit (Frankowski *et al*., 2007). It can be classified as three stages representing different climates:

— an older stage (OIS 3.3) contained two distinct phases of stabilization separated by a phase of weak loess accumulation, and strong denudation resulting from solifluction under moderately wet conditions;

— a middle stage (OIS 3.2) characterized by more intensive loess deposition under colder and more continental climatic conditions, with stronger precipitation and spring thaw;

— a younger stage (OIS 3.1) marked by alternating episodes of cool and cold climate, more or less dry, which probably became shorter through time. Aeolian accumulation was less intensive, and slope processes more active, as is indirectly shown by TL ages of the deposits that are too old. A series of tundra gley soils or arctic brown soils (from two to five according to the geomorphological situation) lying in direct succession represents episodes with ameliorated climatic conditions. The oldest of these is the thickest, and its strong gleying indicates that it developed under the most humid conditions.

Higher rate of aeolian accumulation during the Upper Pleniglacial resulted in the formation of a moderately thick (up to 6 m) bed of uniform loess. Deposition was less intensive during the slightly wetter episode, when gley tundra developed.

Acknowledgements. Investigations of the Kolodiiv profile are partially supported by Grant 6 PO4E 031 15 from the Committee of Scientific Research and are a part of the series of monographic studies of the key profiles of loesses of the Carpathian Foreland. The authors are grateful to Dr M. Wilgat who has prepared the English version of this paper. Many thanks to M.Sc. E. Sadowska for technical preparation of figures.

REFERENCES

ALEXANDROWICZ W. P. and DMYTRUK R. (2007) — Molluscs in Eemian-Vistulian deposits of Kolodiiv site (East Carpathian Foreland) and their palaeoecological interpretation. Geol. Quart., **51** (2): 173–178.

ANTOINE P., ROUSSEAU D.-D., ZÖLLER, LANG A., MUNAUT A.-V., HATTE C. and FONTUGNE M. (2001) — High-resolution of the last interglacial-glacial cycle in the Nussloch loess-palaeosol sequences, Upper Rhine Area, Germany. Quater. Intern., **76** (77): 211–229.

ARNOLD N. S., Van ANDEL T. and VALEN V. (2002) — Extent and dynamics of Scandinavian ice sheet during Oxygen Isotope Stage 3 (65.000–25.000 yr B.P.). Quater. Res., **57**: 38–48.

BEHRE K. E. (1989) — Biostratigraphy of the last glacial period in Europe. Quater. Sc. Rev., **8**: 25–44.

BEHRE K. E. and Van Der PLICHT J. (1992) — Towards an absolute chronology for the last glacial period in Europe: radiocarbon dates from Oerel, Northern Germany. Hist. Veget. Archeobot., **1**: 111–117.

BEHRE K. E. and LADE U. (1986) — Eine Folge von Eem und 4 Weichsel-Interstadialen in Oerel/Niedersachsen und ihr Vegetationsablauf. Eiszeit. Gegen., **36**: 11–36.

BEZUS'KO L. G. and BEZUS'KO A. G. (2003) — Some problems of the palynology of the Riss-Würm deposits of the section Kolodiiv (Halyč Prydnistrov`ja region, Ukraine). XII Ukraińsko-polskie seminarium terenowe "Stratygraficzna korelacja lessów i osadów lodowcowych Ukrainy i Polski" pt. Lessy i paleolit Nadniestrza halickiego. Univ. M. Curie-Skłodowskiej, Lublin: 6–8.

BEZUS'KO L., BOGUCKYJ A. and KLIMANOV W. (1989) — Rastitel'nost' i klimat zapadnyh oblastej USSR v dubnovskom (brianskom) mežstadiale (na primere Malogo Poles'ja). In: Palaeoklimaty i oledenenija v plejstocene. Nauka: 86–91.

BIŃKA K. and GRZYBOWSKI K. (2001) — Early Vistulian deposits at Świnna Poręba, Western Outer Carpathians (Southern Poland). Stud. Quarter., **18**: 11–16.

BOGUCKYJ A. B. (1986) — Antropogenovye pokrovnye otloženyja Volyno-Podolyy. In: Antropogenovye Otloženyja Ukrainy. Nauk. Dumka: 121–132.

BOGUCKYJ A. B. (1987) — Osnovnye ljoessovye i palaeopočvennye horyzonty peryglacyalnoy ljossovo-počvennoy serii plejstocena jugo-zapada Vostočno-Evropeyskoy platformy. In: Stratigrafyja i korrelyacyja morskyh i kontinentalnyh otloženyj Ukrainy. Nauk. Dumka: 47–52.

BOGUCKYJ A. and ŁANCZONT M. (2002) — Loess stratigraphy in the Halyc Prydnistrov'ja region (in Polish with English summary). Stud. Geol. Pol., **119**: 315–327.

BOGUCKYJ A. B. and VOLOŠYN P. (2006) — Inžynierno-geologična charakterystyka vidkladiv opornoho rozrizu Kolodiiv 1. Visnyk L'viv. Un-ta, ser. Geogr., L'viv: **33**: 19–25.

BOLIHOVSKAJA N. S. (1995) — Evolucyja liossovo-počvennoj formacji Sieviernoj Evrazji. Izdatel'stvo Moskov. Univ., Moskva.

BOLIHOVSKAJA N. S. and BOLIHOVSKYJ V. F. (2001) — Ljossovyje porody Evropy. In: Ljossovyj pokrov Zemli i jego svojstva (ed. V. T. Trofimov): 15–73. Izdatel'stvo Moskov. Univ., Moskva.

BOND G., HEINRICH H., BROECKER W., LABEYRIE L., MCMANUS J., ANDREWS J., HUON S., JANTSCHIK R., CLASEN S., SIMET C., TEDESCO K., KLAS M., BONANI G. and IVY S. (1992) — Evidence for massive discharges of icebergs into the North Atlantic during the last glacial period. Nature, **360**: 245–249.

BOND G., BROECKER W., JOHNSEN S., MCMANUS J., LABEYRIE L., JOUZEL J. and BONANI G. (1993) — Correlations between climate records from North Atlantic sediments and Greenland ice. Nature, **365**: 143–147.

BOSINSKI G. (2001) — El Palaeotico medio en Europa Central. Zephyrus, Salamanca, **53–54**: 79–142.

CATT J. A. (1991) — Soils as indicators of Quaternary climatic change in mid-latitude regions. Geoderma, **51**: 167–187.

CHLEBOWSKI R., LINDNER L., BARCZUK A., BOGUTSKY A., GOZHIK P., ŁANCZONT M. and WOJTANOWICZ J. (2004) — Accumulation conditions of the younger upper loesses of Sandomierz Basin, Mid-Carpathian Foreland and Podolian Upland (border between SE Poland and NW Ukraine) on the basis of the geological and mineralogical studies. Ann. Univ. M. Curie-Skłodowskiej, sect. B, **58**: 7–63.

CYREK K. and SYTNYK O. (2002) — Palaeolithic of the Halyč Prydnistrov'ja region (in Polish with English summary). Stud. Geol. Pol., **119**: 291–314.

CYS' P. N. (1962) — Geomorfologia URSR. Vyd. L'viv. Univ., L'viv.

DANSGAARD W., JOHNSEN S. J., CLAUSEN H. B., DAHL-JENSEN D., GUNDERSTRUP N., HAMMER C. U., HVIDBERG C. S., STEFFENSEN J. P., SVEINBJORNDOTTIR A. E., JOUZEL J. and BOND G. (1993) — Evidence for general instability of past climate from a 250-kyr ice-core record. Nature, **364**: 218–220.

DEMEDYUK N. S. (1966) — O stroenii pliocen-četvertičnogo pokrova Predkarpat'ja. Dokl. Soobščen. L'vov. Otdel. Geograf. Obščest. Akad. Nauk. Ukr.: 53–59.

DEMEDYUK N. S. and KRYSTOFOROVA T. F. (1975) — On the first finding of buried peat of the Mikulin period in the Forecarpathian area. Dopovidi Akad. Nauk. Ukr., ser. B, **8**: 678–682.

DING Z., YU Z., RUTTER N. W. and LIU T. (1994) — Towards an orbital time scale for Chinese loess deposits. Quarter. Sc. Rev., **13**: 39–70.

DOLECKI L. and ŁANCZONT M. (1998) — Loesses and palaeosols of the older part of the Wisła (= Würm) glaciation in Poland. Geologija, **25**: 31–38.

EMONTSPOHL A.-F. (1995) — The northwest European vegetation at the beginning of the Weichselian glacial (Brörup and Odderade interstadials) — new data for northern France. Rev. Palaeobot. Palynol., **85**: 231–242.

FENG Z.-D., CHEN F.-Y., TANG L.-Y. and KANG J.-C. (1998) — East Asian monsoon climates and Gobi Dynamics in marine isotope stages 4 and 3. Catena, **33**: 29–46.

FRECHEN M., Van VLIET-LANOË B. and Van Den HAUTE P. (2001) — The Upper Pleistocene loess record at Harmignies/Belgium — high resolution terrestrial archive of climate forcing. Palaeogeogr., Palaeoclimatol., Palaeoecol., **173**: 175–195.

FEDOROWICZ S., BOGUCKYJ A. and ŁANCZONT M. (2003) — Wyniki badań chronostratygraficznych osadów lessowych w profilu Halyč (Ukraina) na podstawie kolekcji próbek z 2002 roku. XII ukraińsko-polskie seminarium terenowe "Stratygraficzna korelacja lessów i osadów lodowcowych Ukrainy i Polski" pt. Lessy i palaeolit Nadniestrza halickiego. Univ. M. Curie-Skłodowskiej, Lublin: 11–14.

FIELD M. H., HUNTLEY B. and MÜLLER H. (1994) — Eemian climate fluctuations observed in a European pollen record. Nature, **371**: 779–783.

FRANKOWSKI Z., ŁANCZONT M. and BOGUCKYJ A. (2007) — Vistulian lito- and pedosedimentary cycles recorded in the Kolodiiv loess-palaeosol sequence (East Carpathian Foreland, Ukraine) determined by laser grain-size analysis. Geol. Quart., **51** (2): 147–160.

FRENZEL B. (1964) — Pollenanalyse von Lössen. Eiszeitalter und Gegenwart, **15**: 5–39.

GERASIMENKO N. P. (2000) — Late Pleistocene vegetational and soil evolution at the Kiev Loess plateau as recorded in the Stari Bezradychy Section, Ukraine. Stud. Quatern., **17**: 19–28.

GERASIMOVA M. I., GUBIN S. V. and SHOBA S. A. (1996) — Soils of Russia and adjacent countries: geography and micromorphology. Wageningen (Miedema), Moscow.

GOZHIK P., MATVIISHINA ZH., GERASIMENKO N., REKOVETS L. and SHELKOPLYAS V. (2001) — Quaternary stratigraphy. In: INQUA-SEQS, Subcommission on European Quaternary Stratigraphy. The Ukraine Quaternary Explored: the Middle and Upper Pleistocene of the Middle Dnieper Area and its Importance for the East-West Correlation. Excursion Guide, Kyiv: 3–62.

GOZHIK P., SHELKOPLYAS V. and KHRYSTOFOROVA T. (1995) — Development stages of loessial and glacial formations in Ukraine (stratigraphy of loesses in Ukraine). Ann. Univ. M. Curie-Skłodowskiej, sect. B., **50**: 65–74.

GRANOSZEWSKI W. (2003) — Late Pleistocene vegetation history and climatic changes at Horoszki Duże, eastern Poland: a palaeobotanical study. Acta Palaeobot. Suppl., **4**: 3–95.

GROOTES P. M. and STUIVER M. (1997) — Oxygen 18/16 variability in Greenland snow and ice with 10–3 to 105 year resolution. J. Geophys. Res., **102**: 455–470.

GUIOT J., De BEAULIEU J. L., CHEDDADI R., DAVID F., PONEL P. and REILLE M. (1993) — The climate in Western Europe during the last glacial/interglacial cycle derived from pollen and insect remains. Palaeogeogr., Palaeoclimatol., Palaeoecol., **103**: 73–93.

GUITER F., ANDRIEU-PONEL V., De BEAULIEU J. L., CHEDDADI R., CALVEZ M., PONEL P., REILLE M., KELLER T. and GOEURY C. (2003) — The last climatic cycles in Western Europe: a comparison between long continuous lacustrine sequences from France and other terrestrial records. Quater. Intern., **111**: 59–74.

GÜNSTER N., ECK P., SKOWRONEK A. and ZÖLLER L. (2001) — Late Pleistocene loess and their paleosols in the Granada Basin, Southern Spain. Quater. Intern., **76** (77): 241–245.

GURTOVAYA E. E. (1983) — Flora and Vegetation in the East of Middle Europe in Mikulino. Izvestia AN SSSR, Ser. Geograf., **4**: 78–86.

HAESAERTS P., MESTDAGH H. and BOSQUET D. (1999) — The sequence of Rremicourt (Hesbaye, Belgium): new insights on the pedo- and chronostratigraphy of the Rocourt Soil. Geol. Belg., **2** (1–2): 5–27.

HESLOP D., LANGEREIS C. G. and DEKKERS M. J. (2000) — A new astronomical timescale for the loess deposits of Northern China. Earth Planet. Sc. Lett., **184**: 125–139.

HOFŠTEJN I. D. (1964) — Neotektonika Karpat, Izdatelstvo Akad. Nauk. Ukr., Kyiv.

HOVAN S. A., REA D. K., PISIAS N. G. and SHACKLETON N. J. (1989) — A direct link between the China loess and marine $\delta^{18}0$ record. In: Milankowitch and Climate, NATO ASI series C, Math. Phys., **126** (1): 269–305.

JAHN A. (1970) — Zagadnienia strefy peryglacjalnej. PWN, Warszawa.

JANCZYK-KOPIKOWA Z. (1991) — Palynostratigraphy of the Pleistocene in Poland and the problem of the age of deposits from Besiekierz (Central Poland). Ann. Univ. M. Curie-Skłodowskiej, sect. B, **46**: 1–26.

JOHNSEN S. J., CLAUSEN H. B., DANSGAARD W., FUHRER K., GUNDESTRUP N., HAMMER C. U., IVERSEN P., JOUZEL J., STAUFFER B. and STEFFENSEN J. P. (1992) — Irregular glacial interstadials recorded in a new Greenland ice core. Nature, **359**: 311–313.

JOHNSEN S. J., CLAUSEN H. B., DANSGAARD W., GUNDESTRUP N., HAMMER C. U. and TAUBER H. (1995) — The Eem stable isotope record along the GRIP ice core and its interpretation, Quatern. Res., **43**: 117–124.

JOUZEL J., BARKOV N. I., BARNOLA J. M., BENDER M., CHAPPELLAZ J., GENTHON C., KOTLYAKOV V. M., LIPENKOV V., LORIUS C., PETIT J. R., RAYNAUD D., RAISBECK G., RITZ C., SOWERS T., STIEVENARD M., YIOU F. and YIOU P. (1993) — Extending the Vostok ice-core record of palaeoclimate to the penultimate glacial period. Nature, **364**: 407–412.

KALINOVYCH N. (2001) — Palinologična charakterystyka vykopnoj gitii rozrizu Kolodiiv (Peredkarpattja). Visnyk L'viv. Univ. Ser. Geograf., **28**: 65–70.

KALINOVYCH N. (2002) — Plant cover in the Carpathians foreland (Ukriane) during the last Interglacial epoch. In: Starsza i środkowa epoka kamienna w Karpatach, Krosno (ed. J. Gancarski): 9–15. Muzeum Podkarpackie.

KEIGWIN L. D., CURRY W. B., LEHMAN S. L. and JOHNSEN S. J. (1994) — The role of the deep ocean in North Atlantic climate change between 70 and 130 kyr ago. Nature, **371**: 323–326.

KLOTZ S., GUIOT J. and MOSBRUGGER V. (2003) — Continental European Eemian and early Würmian climate evolution: comparing signals using different quantitative reconstruction approaches base on pollen. Glob. Planet. Chang., **36**: 277–294.

KOMAR M. (2002) — Pollen analysis of the Upper Pleistocene loesses and palaeosols in the Yezupil and Halyč sites (in Polish with English summary). Stud. Geol. Pol., **119**: 245–252.

KOMAR M. and ŁANCZONT M. (2002) — Late Pleistocene loess-palaeosol and vegetation successions at Tarnawce (San River valley, Carpathian Foothills, Poland). Stud. Quatern., **19**: 27–35.

KRAVČUK J. S. (1999) — Geomorfologija Peredkarpattja. Merkator, L'viv.

KUKLA G. (1987) — Loess stratigraphy in Central China. Quarter. Sc. Rev., **6**: 191–219.

KUKLA J. and LOŽEK V. (1961) — Loess and related deposits. In: Quaternary of central and eastern Europe. Pr. Inst. Geol., **34**: 11–28.

KUKLA G., MCMANUS J., RUSSEAU D. D. and CHUINE I. (1997) — How long and how stable was the Last Interglacial? Quatern. Sc. Rev., **16**: 605–612.

KUSIAK J. (2007) — True and anomalous TL dates from Late Pleistocene loess-palaeosol deposits at the Kolodiiv site (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 167–172.

KUSZELL T. (1997) — Palynostratigraphy of Eemian Interglacial and Early Vistulian in the South Great Polish Lowland (Wielkopolska) and Lower Silesia (in Polish with English summary). Pr. Geol. Miner., **60**: 3–70. Wyd. Uniw. Wrocławskiego.

LIEDTKE H. (1993) — Phasen periglaziär-geomorphologischer Prägung während der Weichseleiszeit im Morddeutschen Tiefland. Z. Geomorph. N. F., Suppl.-Bd., **93**: 69–94.

LITT T., JUNGE F. and BÖTTGER T. (1996) — Climate during the Eemian in north-central Europe — a critical review of palaeobotanical and stable isotope data from central Germany. Veget. Hist. Archaeobot., **5**: 247–256.

LIU T. S., AN Z. S., BAOYIN Y. and JAMATO H. (1985) — The loeess-palaeosol sequence in China and climatic history. Episodes, Ottawa, **8**:21–28.

ŁANCZONT M. (1995) — Stratigraphy and palaeogeography on the Przemyśl Foothills (SE Poland). Ann. Univ. M. Curie-Skłodowskiej, Sec. B, **50**: 49–64

ŁANCZONT M. and ALEXANDROWICZ S. W. (1997) — Problem genezy i wieku terasy średniej (lessowej) doliny Sanu u wylotu z Karpat. In: Seminarium terenowe "Glacjał i peryglacjał Kotliny Sandomierskiej i Przedgórza Karpat w okolicy Przemyśla" (ed. M. Łanczont): 89–97. Univ. M. Curie-Skłodowskiej, Lublin.

ŁANCZONT M. and BOGUCKYJ A. (2002) — The examined loess sites in the Halyč Prydnistrov'ja region (in Polish with English summary). Stud. Geol. Pol., **119**: 33–181.

ŁANCZONT M. and MADEYSKA T. (2005) — Environment of the East Carpathian Foreland in the periods of Palaeolithic man's activity. Catena, **59** (3): 319–340.

ŁĄCKA B., ŁANCZONT M., MADEYSKA T. and BOGUCKYJ A. (2007) — Geochemical composition of Vistulian loess and micromorphology of interstadial palaeosols in the Kolodiiv site (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 127–146.

MAMAKOWA K. (1989) — Late Middle Polish Glaciation, Eemian and Early Vistulian vegetation at Imbramowice near Wrocław and the pollen stratigraphy of this part of the Pleistocene in Poland. Acta Palaeobot., **29** (1): 11–176.

MARTINSON D. G., PISIAS N. G., HAYS J. D., IMBRIE J., MORRE T. C. and SHACKLETON N. J. (1987) — Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 300 000 year chronostratigraphy. Quatern. Res., **27**: 1–29.

MARUSZCZAK H. (1985) — Problems of stratigraphy and paleogeography of loesses in Poland. Guide-book Intern. Symp. "Problems of the stratigraphy and paleogeography loesses". Univ. M. Curie-Skłodowskiej, Lublin: 63–80.

MARUSZCZAK H. (1994) — Chronostratigraphic correlation of loesses in southern Poland and north-western Ukraine (in Polish with English summary). Prz. Geol., **42** (9): 728–733.

MARUSZCZAK H. (1995) — Glacial cycles of loess accumulation in Poland during the last 400 ka and global rhythms of palaeogeographical events. Ann. Univ. M. Curie-Skłodowskiej, sect. B, **50**: 127–156.

MARUSZCZAK H. (1996) — Stratigraphic correlation of Polish, Ukrainian and German Loesses. Biul. Państw. Inst. Geol., **373**: 107–115.

MARUSZCZAK H. (2001) — Stratigraphic scheme of loesses and paleosols in Poland (in Polish with English summary). In: Main Sections of Loesses in Poland (ed. H. Maruszczak): 17–29. Univ. M. Curie-Skłodowskiej.

MOJSKI J. E. (1993) — Some remarks about origin of loess in Poland (in Poland with English summary). Ann. Univ. M. Curie-Skłodowskiej, sect. B, **48**: 223–228.

MOJSKI J. E. (1999) — Small stratigraphic units of Vistulian in the Peribaltic area (in Polish with English summary). Prz. Geol., **47** (3): 247–254.

MÜLLER C., PROS J., and BIBUS E. (2003) — Vegetation response to rapid climate change in Central Europe during the past 140 000 yr based on evidence from the Füramoos pollen record. Quatern. Res., **59**: 235–245.

NAWROCKI J., BOGUCKYJ A. and ŁANCZONT M. (2007) — Palaeomagnetic studies of the loess-palaeosol sequence from the Kolodiiv section (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 161–166.

NAWROCKI J., BOGUCKYJ A., ŁANCZONT M., KUSIAK J. and WERNER T. (2003) — Palaeomagnetyczne podstawy podziału i korelacji chronostratygraficznej lessów vistuliańskich Wyżyny Wołyńskiej. In: XII ukraińsko-polskie seminarium terenowe "Stratygraficzna korelacja lessów i osadów lodowcowych Ukrainy i Polski" pt. Lessy i palaeolit Nadniestrza halickiego. Univ. M. Curie-Skłodowskiej, Lublin.

NEY R., BURZEWSKI W., BACHLEDA T., GÓRECKI W., JAKÓBCZAK K. and SŁUPCZYŃSKI K. (1974) — Zarys paleogeografii i rozwoju litologiczno-facjalnego utworów zapadliska przedkarpackiego. Pr. Geol. PAN, **82**: 1–65.

NIKLEWSKI K. and KRUPIŃSKI K. M. (1992) — Sediments of the Eemian Interglacial and Vistulian in the Łomżyca Basin, NE Poland (in Polish with English summary). Stud. Geol. Pol., **49**: 43–59.

OLSZEWSKA B. (1999) — Biostratigraphy of Neogene in the Carpathian Foredeep in the light of new micropalaeontological data (in Polish with English summary). Pr. Państw. Inst. Geol., **158**: 9–28.

OSZCZYPKO N. (2001) — Evolution of the Carpathian Foredeep in Miocene (in Polish with English summary). Prz. Geol., **49** (8): 717–723.

PAEPE R. and VANHOORNE R. (1967) — The stratigraphy and paleobotany of the Late Pleistocene in Belgium. Mémoires pour servir a l'explication des cartes géologiques et mini. Pres de la Belgique, **8**.

PERYT T. M. and PERYT D. (1994) — Badenian (middle Miocene) Ratyn Limstone in Western Ukraine and Northern Moldavia: microfacies, calcareous nannoplankton and isotope geochemistry. Bull. Pol. Acad. Sc., **42** (2): 127–136.

PETRYCZENKO O. I., PANOV G. M., PERYT T. M., SREBRODOLSKYJ B. I., POBEREŽSKYJ A. V. and KOVALEVIČ V. M. (1994) — Outline of geology of the Miocene evaporite formations of the Ukrainian part of Carpathian Foredeep (in Polish with English summary). Prz. Geol., **42** (9): 734–737.

RACINOWSKI R. (2007) — Heavy minerals in the key Late Pleistocene loess-palaeososection at Kolodiiv (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 185–188.

ROUSSEAU D.-D., GERASIMENKO N., MATVIISCHINA ZH. and KUKLA G. (2001) — Late Pleistocene of the Central Ukraine. Quatern. Res., **56**: 349–356.

ROUSSEAU D.-D., KUKLA G., ZÖLLER L. and HRADILOVA J. (1998) — Early Weichselian dust storm layer at Achenheim in Alsace, France. Boreas, **27**: 200–207.

RÓŻYCKI S. Z. (1986) — Pyłowe utwory typu lessowego na świecie, ich występowanie i geneza. Stud. Geol. Pol., **85**.

RÓŻYCKI S. Z. (1976) — Conditions of the Formation of loess (in Polish with English summary). Biul. Inst. Geol., **297**: 77–92.

SCHIRMER W. (2000) — Eine Klimakurve de Oberpleistozäns aus dem rheinischen Löss. Eiszeitalter und Gegenwart, **50**: 25–49.

SEMMEL A. (1967) — Über Prä-Würm - Lösse in Hessen. Notisbal. Hess. L.-Amt Bodenf. Bd., **95**.

SEUL C. (2007) — Sedimentary environments of the Late Pleistocene terrestial deposits at Kolodiiv (East Carpathian Foreland, Ukraine) in the light of elementary and aggregate grain-size composition. Geol. Quart., **51** (2): 179–184.

ŠAKYN V. S. ed. (1976) — Geologičeskaja Karta Ukrainskih Karpat i prylegajuščyh progibov. Ukrain. Nauč. Issledov. Geologorazvjedočnyj Instytut, L'viv.

SMOLIKOWA L. (1971) — Gesetzmässigkeiten der Bodenentwicklung im Quartär. Eiszeit. Gegen., **22**: 156–177.

SYTNYK K., BEZUS'KO L., BEZUS'KO A. and BOZHKO Y. (2003) — Condition of palynological study of Riss-Würm deposits of section Kolodiiv (Ukraine, Ivano-Frankivska region). Biol. Ekol. Nauk. Zapys., **21**: 8–15.

SYTNYK O., BOGUCKYJ A. and ŁANCZONT M. (2007) — Mousterian artifacts from the unique Vistulian loess-palaeosol sequence at Kolodiiv (East Carpathian Foreland, Ukraine). Geol. Quart., **51** (2): 189–192.

TERHORST B., APPEL E. and WERNER A. (2001) — Palaeopedology and magnetic susceptibility of a loess-palaeosol sequence in southwest Germany. Quater. Intern., **76** (77): 231–240.

TEISSEYRE W.(1896) — Atlas geologiczny Galicyi, Kałusz i Halicz. Akad. Umiejętn. Kraków.

TOBOLSKI K. (1991) — Biostratygrafia i bioekologia interglacjału eemskiego i zlodowacenia Wisły regionu konińskiego. In: Przemiany środowiska geograficznego regionu Konin-Turek. Wyd. Univ. A. Mickiewicza, Poznań: 45–87.

TOKARSKI J. (1936) — Physiographie des podolischen Lösses und das Problem seiner Stratigraphie. Mémoires. Acad. Pol. Sc. Lett., ser. A, **4**.

TROFIMOV V. T., KADYROV J. V. and MINERVIN A. V. (2001) — Obszczeplanetarnyje zakonomiernosti rasprostranenija ljossovykh porod. In: Ljossovyj pokrov Zemli i jego svojstva, (ed. V. T. Trofimov): 10–15. Izdatel. Moskov. Univ., Moskva.

Van ANDEL (2002) — The climate and landscape of the middle part of the Weichselian Glaciation in Europe: the Stage 3 Project. Quater. Res., **57**: 2–8.

Van Der HAMMEN T. (1995) — The Dinkel valley revisited: Pleniglacial stratigraphy of the eastern Netherland and global climatic change. Mededel. Rijks Geol. Dien., **52**: 343–355.

Van Der HAMMEN T., MAARLEVELD G. C., VOGEL J. C. and ZAGWIJN W. H. (1967) — Stratigraphy, climatic succession and radiocarbon dating of last glacial in the Netherlands. Geol. Mijn., **46**: 79–95.

Van VLIET-LANOË B. (1987) — The role of segregation ice in the superficial formations of W Europe. The processes and their heritages. In: Abstract of the State Doctorate Thesis. Caen, Centr. De Geom. Du C.N.R. SUniv. Paris I, Sorbonne.

VANDENBERGHE J. (1992) — Geomorphology and climate of the cool oxygen isotope stage 3 in comparison with the cold stages 2 and 4 in The Netherlands. Z. Geomorph. N. F., Suppl.-Bd., **86**: 65–75.

VANDEBENGHE J., HUIJZER A. S., MUCHER H. J. and LAAN W. (1998) — Short climatic oscillations in a western European loess sequence (Kesselt, Belgium), J. Quatern. Sc., **13**: 471–485.

VANDEBERGHE J. and NUGTEREN G. (2001) — Rapid climatic changes recorded in loess succession, Global and Planetary Change, **28**: 1–9.

VEKLIČ M. F. (1968) — Stratigrafja lessovoy formatsii Ukrainy i sosiednih stran. Nauk. Dumka, Kyiv: 5–253.

VELIČKO A. A. (1990) — Loess-palaeosol formation on the Russian Plain. Quatern. Intern., **7** (8): 103–114.

WOILLARD G. (1975) — Recherches palynologiques sur le Pléistocene dans l'Est la Belgique et dans les Vosges Lorraines. Acta Geograph. Lovanien., **14**.

WU G., PAN B., GUAN Q., LIU Z. and LI J. (2002) — Loess records of climatic changes during MIS5 in the Hexi Corridor, northwest China. Quatern. Intern., **97** (98): 167–172.

ZAGWIJN W. H. (1974) — Vegetation, climate and radiocarbon datings in the late Pleistocene of the Netherlands. Part II: Middle Weichselian. Mededelingen Rijks Geologische Dienst, **25**: 101–111.

ZARRINA E., KRASNOV I. and SPIRIDONOVA E. (1981) — Detailed chronostratigraphic scale of Late Pleistocene of the North-West and Centre of Russian Plaine. In: Glacial Deposits and Glacial History in Eastern Fennoscandia, Apatity: 90–102.