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Stratigraphy and chronology of the periphery of the Scandinavian ice sheet at the foot of the Ukrainian Carpathians

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ABSTRACT

The article presents the results of research on the formation conditions and the age of glaciogenic deposits left by the Scandinavian ice sheet in the Eastern Carpathian Foreland, in its part drained by the Dniester River and its tributaries (Black Sea basin). These deposits (till, erratics, glaciolacustrine and glaciofluvial sediments), preserved only in some places and occurring in the range of altitudes defined by terrace VI (Early Pleistocene) and terrace V (Middle Pleistocene), are the southernmost traces of the ice sheet in Europe. They are underlain by fluvial deposits and overlain by loess deposits, and locally occur as intraloess layers. The following research methods were used: geomorphological mapping, sedimentological, lithological, micromorphological, palaeopedological, and palaeontological analyses, as well as luminescence dating. The obtained results allowed a reconstruction to be made of the main stages of morphogenesis and the style of the ice-sheet advance at its maximum extent, defined as the Sambor Phase. The area was glaciated during MIS 12 (Elsterian II = Sanian 2 = Okanian), as indicated by the occurrence of a complete Middle/Upper Pleistocene loess–palaeosol sequence. The oldest palaeosol (S4), i.e. Sokal (=Mazovian = Zavadivka) soil that developed directly on the glaciogenic deposits or loess L5 is correlated with the Holsteinian (=Likhvinian) Interglacial (MIS 11).

1. Introduction

There is currently great interest in Pleistocene ice sheets among Quaternary researchers, more specifically, the number of glaciations, their glacial age and stratigraphic position (cf. Mojski, 1993). Hence, many discrepancies exist in the stratigraphic subdivisions of the Quaternary in different parts of Europe, which complicates any reliable correlation of glaciations (coolings) and interglacials (warmings), including their stratigraphic position (Marine Isotope Stage number) and regional naming (Ehlers et al., 2011).

Such discrepancies are also unresolved in Central and Eastern Europe (e.g. Lindner et al., 2002a, 2002b, 2004, 2007). The case study presented in this paper examines the stratigraphy of the Middle Pleistocene glacial-interglacial complex in this area; this stratigraphy of Quaternary history remains unresolved, especially concerning the problem of the number of ice-sheet advances.

So far, the four Marine Isotope Stages of this complex (MIS 12, 10, 8 and 6) were believed to represent four ice-sheet advances into this area. However, geological fieldwork carried by numerous researchers from Poland, Belarus and Ukraine, among others, uncovered only two glacial tills (Lindner et al., 2007), which presumably correspond to two ice sheets: the Elsterian and the Saalian (cf. Matoshko, 2011). Among them, in the Middle Pleistocene glacial-interglacial cycles in this part of Europe, a special place is taken by the problem of the stratigraphic position of the Elsterian Glaciation, which was the first extensive Fennoscandian ice sheet extension into the central part of the continent (cf. Mojski, 1993). Recent investigations conducted in Central Germany have evidenced that this southern-most ice advance of Fennoscandian

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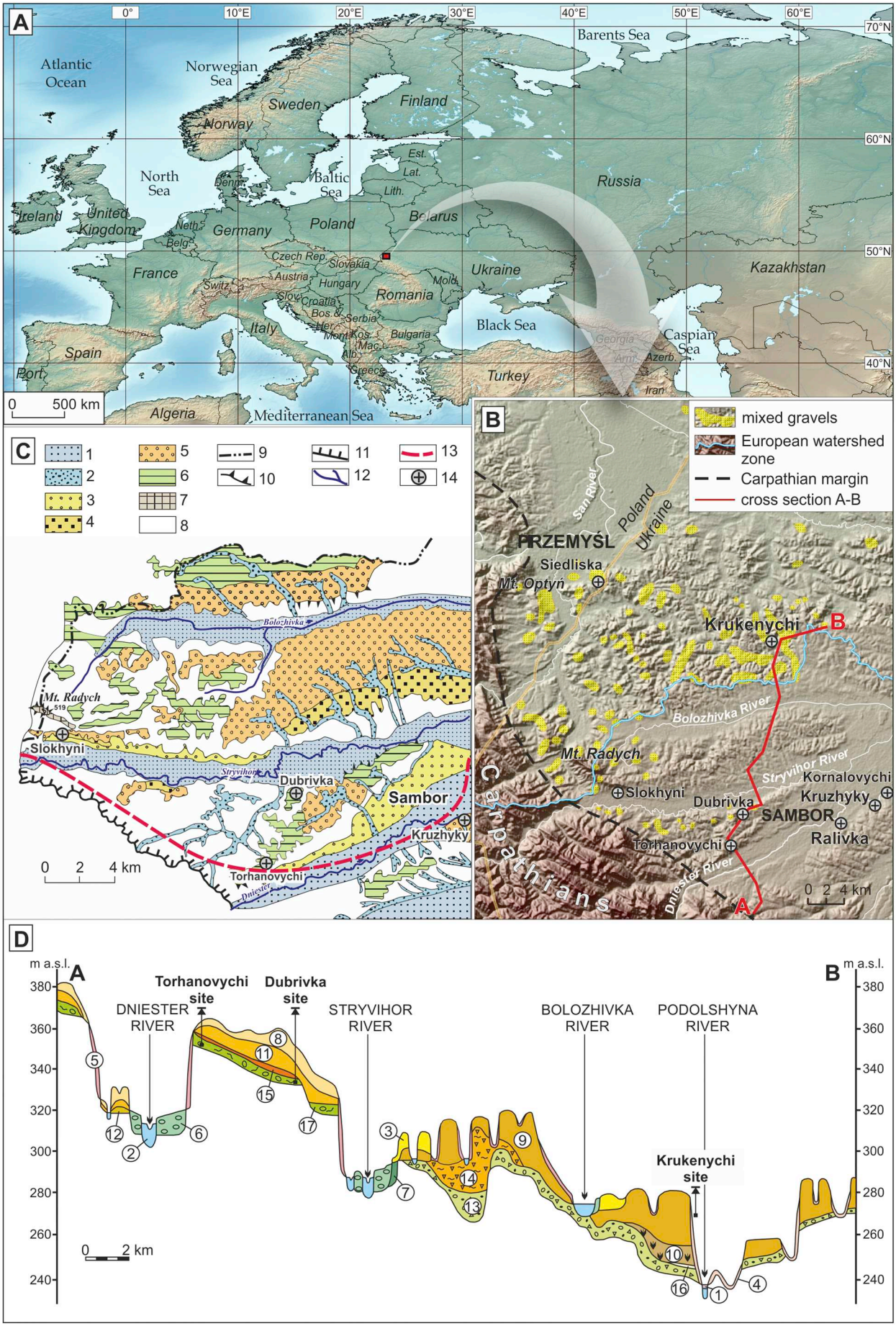

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Fig. 1. Area of study: A — location on the map of Europe (source of raster data: www.naturalearthdata.com); B — distribution of mixed gravels and sites mentioned in the text presented on relief model of Eastern Carpathian Foreland; C — main elements of relief after Łanczont et al. (2015, modified), and Scandinavian glacial limit after Demediuk and Demediuk (1995, modified); Holocene: 1 — flood plains, 2 — bottoms of smaller river valleys; Pleistocene terraces: 3 — II and III (unseparated), 4 — IV, 5 — V; 6 — Loyova level; 7 — Krasnoye level; 8 — slopes; 9 — European watershed (San-Dniester); 10 — higher erosion scarps; 11 — Carpathian margin; 12 — rivers; 13 — extent of the ice-sheet; 14 — sites; D — Quaternary deposits along submeridional morphological cross section A–B (after Gerasimov et al. (2005) modified after Hołub (2016) and supplemented): Holocene: 1 — biogenic sediments (peats), 2 — fluvial sands, gravels with pebbles; Pleistocene/Holocene: 3 — aeolian sands and sandy silts, 4 — colluvial sediments (silty loams); 5 — colluvial sediments (sandy silts) deposited on gravels and older weathered rocks; Late Pleistocene: 6 — fluvial sands, silts and gravels (I terrace), 7 — fluvial gravels, sands and silts (II terrace), 8 — aeolian-colluvial and weathered sandy silts and silts; Middle/Late Pleistocene: 9 — aeolian-colluvial and weathered sandy silts and clayey silts; Middle Pleistocene: 10 — biogenic sediments (peats, tills with organic material); 11 — Aeolian-colluvial sandy silts and loamy silts; 12 — fluvial gravels and sands; 13 — fluvial sands with clayey layers, 14 — sandy and clayey silts with gravels and pebbles, 15 — aeolian/colluvial sandy and clayey silts; glaciolacustrine silts; till, 16 — aeolian-colluvial sandy and clayey silts; Early Pleistocene: 17 — pebbles with sands (Loyova level).

ice sheets corresponds with MIS 12. This is indicated by the age of the sediments (approximately 400 ka) directly overlying the Elsterian till (Lauer and Weiss, 2018).

The western part of the Eastern Carpathian Foreland (Ukraine), located south of the European watershed, i.e. in the Subcarpathian part of the upper Dniester River catchment (Fig. 1A, B), may be the key to identifying the maximum extent and the timing of the Elsterian (= Sanian = Okanian) ice sheet in eastern-central Europe. This ice sheet, formerly referred to as the Carpathian or the Cracovian (cf. Mojski, 2006), reached the northern part of the Carpathian arc, passed by the sigmoidal bend of the Carpathians near Przemyśl, and entered the upper Dniester River catchment (Teisseyre, 1938; Herenchuk et al., 1972; Demediuk and Demediuk, 1995). The presence and extent of the ice sheet in this area are indicated by the occurrence of thin basal till (Demediuk and Demediuk, 1988) that occur only locally. Large erratics, that is till residues, occur on interfluve uplands, i.e. in situ, and on the slopes and in bottoms of river valleys — Stryvihor and Dniester as result of short-distance redeposition (Zierhoffer, 1932; Łanczont et al., 2004; Yatsyshyn et al., 2011c). The approximate extent of the ice sheet is also indicated by the presence of so-called mixed gravels (Fig. 1B), composed of Carpathian and Scandinavian material occurring in the form of thin beds in interfluve areas (Łanczont, 1997a; Łanczont et al., 2011). These represent a record of the concentrated flows of glacial meltwater streams, as well as the extraglacial waters of the Carpathian rivers (Friedberg, 1905; Przepiorski, 1938; Błachowski, 1938; Teisseyre, 1938; Demediuk and Demediuk, 1995; Łanczont and Racinowski, 1994; Bogucki and Łanczont, 2011). Other glaciogenic deposits occurring in the Dniester–Stryvihor interfluve are derived from terminoglacial lakes (Bogucki et al., 2009b). The only forms of glacial relief that have been preserved in the area of the upper Dniester River catchment are ice-marginal valleys, the most interesting of which is the Bolozhivka River valley (Romer, 1906; Yatsyshyn et al., 2011a). Unique relics of the post-glacial lakeland landscape are visible in the present-day relief of the European watershed zone (Rożycki, 1968; Bogucki (Boguckyj) et al., 1999; Łanczont et al., 2003b).

In general, the deposits and forms of glacial origin occurring in the Eastern Carpathian Foreland are covered by a continuous, 10–15 m thick loess mantle (Teisseyre, 1932/1933, 1938; Bogucki (Boguckyj) et al., 1999; Łanczont et al., 2004), referred to by Ukrainian researchers as loess-like loam (Herenchuk et al., 1972). Loess covers occur on the Pleistocene staircase terraces of the Dniester River and its tributaries (Fig. 1C, D), and the age of these terraces can be indirectly deduced from the loess stratigraphy.

Due to the extent of the extensive loess mantle, the maximum extent of the ice sheet is difficult to define, and so its boundary has been repeatedly redrawn as new information has been obtained. It is now believed that the ice sheet (Fig. 1C) advanced along the Carpathian margin and reached its maximum extent along the Chyriv–Ralivka–Kornalovychi line (Bogucki (Boguckyj) et al., 2007; Demediuk and Demediuk, 1988). This has been supported by new discoveries of till (Kruzhyky) and erratics (Torhanovychi).

In addition, the precise time that the glacial phenomena occurred in this part of the Eastern Carpathian Foreland remains unclear. Two different views are presented in literature. In the first approach, based on opinions first published in older papers (i.e. Teisseyre, 1938; Klimaszewski, 1936; Przepiorski, 1938) and later developed since the early 1970s by Ukrainian and Polish researchers (i.e. Demediuk and Demediuk, 1995; Herenchuk et al., 1972; Bogucki (Boguckyj) et al., 1999; Lindner et al., 2002a, 2002b), the area of the Eastern Carpathian Foreland was glaciated only once during the Elsterian II = Sanian 2 = Okanian Glaciation (correlated with MIS 12; *vide* Table 1). The Dniester lobe reached its maximum extent during the so-called Sambor Phase. Traces of a recessive phase, called the Krukenychi Phase, have been found about 20 km NW of the maximum extent, along the line of the present-day European watershed (Przepiorski, 1938; Bogucki (Boguckyj) et al., 1999, 2004). Glaciogenic series are thicker, and more common and complex, in this zone. The cover of sands and mixed gravels is divided by a thin discontinuous layer of basal or flow till, and locally also by glaciolacustrine deposits forming lenticular lithosomes, 2–10 m thick, occurring locally below the till (Demediuk and Demediuk, 1995; Łanczont and Racinowski, 1994). Recessive glaciofluvial deposits, occurring above the till, have a limited extent (Demediuk and Demediuk, 1995; Herenchuk et al., 1972).

The second approach reinterprets the above data by indicating that two separate ice sheets reached the NW part of the Ukrainian Carpathian Foreland. Firstly, this approach referred to the advance of two independent ice-sheet lobes of different age, depending on the morphology of the foreland and ice-sheet dynamics (Lindner, 2001). Secondly, Gozhik et al. (2012) and Lindner and Marks (2015) propose that the Chyriv–Ralivka–Kornalovychi line (i.e. the ice-sheet maximum extent) was reached during the Elsterian I (Sanian 1 = Glacial B within the Cromerian Complex) = Donian Glaciation (MIS 16; Table 1), and not during the Sambor Phase of the Elsterian II = Sanian 2 = Okanian Glaciation (MIS 12). They believe that the latter glaciation reached approximately the extent attributed formerly to the Krukenychi recessive phase. Two layers of glaciofluvial mixed gravels (occurring in the Krukenychi zone and attributed formerly to two phases of the Sanian 2 Glaciation) were recognized by Lindner and Marks (2015) as units formed during two separate glaciations; this proposal was based on inter alia the reinterpretation of data presented at the Polish–Ukrainian conference in 2011, during which the origin of mixed gravels was discussed (Bogucki et al., 2011a; Łanczont et al., 2011). This new idea has not been confirmed in the upper Dniester River catchment, where glaciogenic deposits older than the Sanian 2 Glaciation have not been found so far.

The present paper argues that the younger of the two ice sheets, i.e. Elsterian II = Sanian 2 = Okanian (MIS 12), reached its furthest extent towards the SE part of the study area. It proposes that the key sections (Dubrivka, Kruzhyky, Torhanovychi and Slokhyni) are located within the predominant elements of relief of the Eastern Carpathian Foreland, and that in these sections, series of preglacial (fluvial), glacial (different types), and periglacial (mainly loess) deposits are exposed. The latter typically occur above glaciogenic deposits, but can also be found beneath them in places. Glaciogenic deposits constitute the key horizon and, together with the complete loess sections, provide a sound basis for the reconstruction of the Pleistocene history of the region. By defining

Table 1
Loess stratigraphy of Ukraine and Russia and its correlation with MIS and Quaternary stratigraphy of Poland, Russia and Western Europe after Velichko et al., 2010; Zagwijn (1985), and Lisiecki and Raymo (2005).

Age [ka]	Palaeo-magnetism	MIS	Loess-soil units	Loess-soil stratigraphy: UKRAINE Western	Loess-soil stratigraphy: UKRAINE Central	Loess-soil stratigraphy: RUSSIA	Glaciations Interglaciations: POLAND	Glaciations Interglaciations: RUSSIA	Glaciations Interglaciations: WESTERN EUROPE	Stratigraphy	
14	BRUNHES	1	*S0*	*Modern soil*	*Modern soil*	*Modern soil*	*Holocene*	*Holocene*	*Holocene*	HOLOCENE	
130	BRUNHES	2 4 3	*L1*	*Dubno*	*Vytachiv* VALDAI	*Bryansk*	VISTULIAN	VALDAIAN	WEICHSELIAN	PLEISTOCENE	Late
	BRUNHES	5e 5	*S1*	Horokhiv	Pryluky	Mezin	Eemian	Mikulinian	Eemian		
	BRUNHES	6	*L2*	*Ternopil*	TYASMIN		WARTANIAN	DNIEPER: MOSCOWIAN	SAALIAN: WARTHE (SAALE III)		Middle
	BRUNHES	7	*S2*	S2-I S2-II *Korshiv*	*Kaydaky*	*Romny*	*Lublinian*	DNIEPER: Kostroma	SAALIAN: Treene		
	BRUNHES	8	*L3*		DNIEPER		ODRANIAN	DNIEPER: DNIEPER Interstadial STAGE	SAALIAN: SAALE II (DRENTE II) SAALE I (DRENTE I)		
	BRUNHES	9	*S3*	*Lutsk*	*Potygaylivka*	*Kamenka*	*Zbójnian*	*Kamenkan*	*Demnits, Wacken*		
	BRUNHES	10	*L4*		OREL		LIWIEC	PECHORAIAN	FUHNE: STAGE II Interstadial STAGE I		
420	BRUNHES	11	*S4*	S4-I S4-II *Sokal*	*Zavadivka*	*Inzhava*	*Mazovian*	*Likhvinan*	*Holstenian*		
	BRUNHES	12	*L5*		TYLIHUL		SANIAN 2	OKAIAN	ELSTERIAN II		
	BRUNHES	13 14 15	*S5*	S5-I S5-II *Solotvyn*	*Lubny*	*Vorona*	*Ferdynandowian*	*Ikorets interglacial* ? *Muschap Interglacial*	CROMERIAN COMPLEX: *Interglacial IV Voigstedt* STAGE C *Interglacial III*		
	BRUNHES	16	*L6*		SULA		SANIAN 1	DONIAN	CROMERIAN COMPLEX: *GLACIAL B (ELSTERIAN I)*		
780	BRUNHES	17 18 19	*S6* *S7*	S6-I S6-II S7 *Zahvizdya*	Martonosha	Kolkotova	Malopolanian	*Okatovo* SETUN STAGE *Krasikovian*	CROMERIAN COMPLEX: *Interglacial II*		
	MATUYAMA	20			PRYAZOVSK		NIDIAN	*Akulovian*	CROMERIAN COMPLEX: GLACIAL A		Early
	MATUYAMA	21	*S8*	S8	*Shyrokino*		*Podlasian*		CROMERIAN COMPLEX: *Interglacial I*		
	MATUYAMA	PREGLACIAL COMPLEX									

the age of the individual segments of these sequences, developing the chronology of events connected with their formation, and then reconstructing the landscapes in the ice-sheet marginal zone, it is possible to obtain key data regarding the time when the Scandinavian ice sheet reached its maximum extent. Particular attention has been given to interglacial soils.

2. Study area

The study area is located in the Subcarpathian part of the Eastern Carpathian Foreland (Fig. 1A, B). As regards tectonics, the area belongs to the Carpathian Foredeep. It is the zone of fold structures composed of Miocene molasse deposits (Petrychenko et al., 1994). The youngest deposit (of Upper Sarmatian-Tortonian age) is the thick-bedded Radych

conglomerate that forms the syncline of Mt. Radych, and of Mt. Optyń in Poland (Fig. 1B).

The structural-erosional relief manifests itself as a ridge-hilly landscape. The watershed areas, stretching in a NW-SE direction and running perpendicular to the Carpathian margin, are flat and undulated plateaux separated by vast, flat-bottomed valleys of the main rivers: the Dniester, Stryvihor, and Bolozhivka (Fig. 1B, C). The watershed areas between these rivers reach altitudes of 350–400 m a.s.l. Their valleys meet in the 10–12 km wide, flat tectonic depression of the Upper Dniester Basin at an altitude of 275–264 m a.s.l. (Hołub, 2016). Terraces of different ages are well preserved and visible in the relief of the study area. The system of the so-called upper group of preglacial terraces (Teisseyre, 1938), consisting of the fragmentary Upper Pliocene Krasnoy level (terrace VII) and Early Pleistocene Loyova level (terrace VI), occurs on the watershed plateaux (Fig. 1C). In the study area, the Krasnoy level (85–95 m above the Stryvihor River channel) has been found on the slopes of Mt. Radych (519 m a.s.l.), the highest point of the watershed between the San and Dniester rivers. The lower Loyova level is the main element of the landscape of interfluve areas in the Eastern Carpathian Foreland. Within this level, four surfaces are distinguished: they occur at different altitudes and are probably of different ages (Teisseyre, 1938; Yatsyshyn et al., 2008; Łanczont et al., 2004). The subsequent terraces (V–II) form the so-called lower group of intravalley terraces (Teisseyre, 1938). The surface of terrace V occurs at 335 m a.s.l. near the Carpathian margin, 302–305 m a.s.l. in the eastern part of the study area; in the Upper Dniester Basin, it is covered by Holocene alluvia created by neotectonic subsidence (Hofstein, 1962; Yatsyshyn and Plotnikov, 2004). As the lowest-lying glacial deposits occur on its surface, this terrace can be termed the glacial level. It is believed that the ice sheet advanced from the N and NW, and the valleys of the Bolozhivka, Stryvihor, and Dniester rivers periodically functioned as ice-marginal valleys (Romer, 1906, 1907; Bogucki (Boguckyj) et al., 1999). Terraces IV–II were formed in the valleys of the Dniester and Stryvihor rivers after the glaciation. The Holocene period of the development of the upper Dniester and Stryvihor valleys is represented by terrace I and the main floodplain (5–7 m) with 3–4 lower levels.

3. Material and methods

Quaternary deposits forming covers on the high terraces (VI and V) of the Dniester River in the Eastern Carpathian Foreland have been investigated since 1999. Besides geomorphological mapping, the following studies were carried out in the selected sites, i.e. Dubrivka, Torhanovychi, Slokhyni, and Kruzhyky (Fig. 1B, C).

Primary sedimentological features, i.e. texture and structure of deposits, the scale of the depositional units, and the contacts between them, were identified in the field. Lithofacies analysis was carried out according to the methodological principles (Table 2) proposed by Zieliński and Pisarska–Jamroży (2012). Additionally, the directions of till fabric in the glacial environment (Dubrivka site; Terpiłowski et al., 2011a) and the fluvial environment (Torhanovychi site; Yatsyshyn et al., 2011b) were measured.

Laboratory analyses were mainly carried out on samples taken in the Dubrivka key section, which is one of most complete and stratigraphically-diverse complex glacial–loess sections in the Eastern Carpathian Foreland (Łanczont et al., 2004), and this section therefore has the most complete analytical documentation. However, some of the analyses described below were also carried out on samples obtained from the other sites; these were typically taken at intervals of 0.2–0.5 m from almost all distinguished layers of the studied sections, excluding glacial till and alluvial gravelly-sandy deposits (Bogucki et al., 2010; Łanczont et al., 2004, 2015). The following laboratory analyses were carried out:

- The grain-size distribution of deposits was determined in samples representing all lithological units and palaeosols distinguished in the Dubrivka, Torhanovychi, and Kruzhyky sections, and in the younger part of the Slokhyni section (Bogucki et al., 2010; Łanczont et al., 2004, 2015). The analysis was carried out by the areometric method. Based on the measurements of grain sizes (six size-classes from fine sand to clay), the main granulometric indices were calculated according to Folk and Ward (1957);
- The micromorphology of grains in the silt fraction was analyzed under a scanning electron microscope (SEM). Thirteen samples of glaciolacustrine deposits were taken from the Dubrivka section. The microrelief of 50–100 randomly-selected grains of the 0.063–0.020 mm fraction was determined in each sample. The microstructures on the surface of the grains were described according to the classification proposed by Mahaney (2002), together with microforms distinguished by Helland and Holmes (1997), Woronko (2007), and Woronko and Hoch (2011). Additionally, the chemical composition of the grains and the crusts on their surface was determined in five samples using the EDS method (Woronko et al., 2011);
- The composition of heavy mineral assemblage in the 0.1–0.05 mm fraction of deposits from the Dubrivka section was determined (the results are published in: Łanczont et al., 2004) according to Racinowski et al. (2003);
- Humus content was determined according to Tiurin (content of organic carbon converted to percentage of humus), carbonate content according the Scheibler's volumetric method (percentage of $CaCO_3$), and the content of iron oxides (total iron content) by the colorimetric (thiocyanate) method using a UV03100 PC spectrophotometer;
- The micromorphological features of the loess–palaeosol sequences in the Dubrivka, Kruzhyky, and Slokhyni sections were described using the terminology published by Mroczek (2008, 2013);
- Pollen analysis was carried out for selected samples taken from mineral deposits in the Dubrivka section (palaeosol S4 and glaciolacustrine deposits). The samples were prepared according to Komar et al. (2009). Organogenic deposits in the Kruzhyky section were sampled for analyses of pollen and macrofossils, and the results were published by Łanczont et al. (2010);
- Samples for *Ostracoda* analysis were obtained from glaciolacustrine deposits in the Dubrivka section (negative result) and from organogenic deposits in the Kruzhyky section (Łanczont et al., 2010).
- Samples for *Molluscs* analysis were taken from the selected layers of loess deposits in the individual sites, but results have been negative.

Table 2
Lithofacies code of the examined sediments after Zieliński and Pisarska–Jamroży (2012).

Textural symbols		Structural symbols	
D	diamictons	t	trough cross-stratification
G	gravels	h	horizontal
S	sands	m	massive
T	silts	i	low-angle inclined bedding
T	muds	r	ripple cross-lamination
C	organic deposits	f	flaser lamination

3.1. Stratigraphy and chronostratigraphy findings

The key stratigraphic units in the Dubrivka and Torhanovychi sections were TL dated. A set of 40 TL dating results was obtained for loess, glacial, glaciolacustrine, and fluvial sediments (cf. Berger and Easterbrook, 1993); some of the results indicate the age of > 250 ka (Supplementary materials 2, 7 and 8). The all TL ages were determined by the late J. Kusiak in the Lublin laboratory in the years 1999–2010 according to the same methodology as described by Kusiak (Kusiak et al., 2012, 2013; Fedorowicz et al., 2013).

The reliability of the TL method remains uncertain. Some authors assert that reliable dating results can be obtained for the deposits up to an age of 250–300 ka and that deposits older than 300–400 ka cannot be correctly dated (e.g. Fedorowicz, 2006; Bluszcz, 2000; Frechen et al., 1997, 1999). Alternatively, Singhvi et al., 1982, Berger et al. (1992), Kusiak, 2002, and Kusiak et al., 2013) indicate that reliable results can be obtained for loess deposits aged up to about 800 ka, with the reservation that dating could be difficult if the equivalent dose considerably exceeds 2000 Gy and the TL signal is almost saturated. It is believed to be the source of potentially quite considerable errors when determining equivalent dose, and thus the TL age (Kusiak et al., 2013).

The large series of TL results (presented in the supplements) obtained for the two profiles in the years 1999–2010 (cf. Łanczont et al., 2004; Bogucki et al., 2010, 2011d) can be regarded as historical material, because the method used for determining the age of mineral deposits is currently not recommended (Duller, 2015; Zöller and Wagner, 2015). Some TL ages were inverted or overestimated for a particular layer; the reasons for this phenomenon have been attributed to methodological aspects or the features of mineral material. Still, despite the limitations of TL, it nonetheless allows the studied events to be ordered in time: the obtained TL ages can be allocated into several age groups, generally corresponding to the delimited pedo- and lithostratigraphic units, thus allowing a complete chronostratigraphy to be created. The Matuyama/Brunhes boundary (MBB — ca. 0.78 Ma) and the palaeomagnetic excursions found in several profiles of western Ukraine (Korolevo site, Transcarpathia, Nawrocki et al., 2016; Zahvizdja site, East Carpathian Foreland, Nawrocki et al., 2002; Skala Podils'ka site, Podillia Upland, Bogucki et al., 2009a) has been indicated on the basis of TL results.

Based on the gathered analytical data and observations made in the field, the litho- and pedostratigraphic units were distinguished. The soil horizons were described by applying palaeopedological criteria. The stratigraphy of the studied loess sequences was prepared according to the current stratigraphic scheme of Western Ukraine and the regional names of interglacial palaeosols (stratotypes) from the Volhynia-Podolia Uplands and the Eastern Carpathian Foreland (Table 1; Bogucki, 1987; Bogucki (Boguckyj) and Łanczont, 2002). The main loess and soil stratigraphic units of glacial and interglacial rank were labelled using the system of the letters L (for loess) and S (for soil), respectively, and numbered according to their increasing age (from the Holocene soil S0). This system of letter-number symbols was adapted from the Chinese loess stratigraphy (Liu et al., 1985) by Kukla (1987) and Kukla and An (1989). It is quite often used in the stratigraphy of European loesses (e.g. Antoine et al., 2009; Bogucki (Boguckyj) and Łanczont, 2002; Bogucki and Łanczont, 2018; Obrecht, 2017; Marković et al., 2006, 2015; Zöeller, 2010). The loess stratigraphy of the Western Ukraine (Table 1) was correlated with the formal Ukrainian stratigraphic scheme of the Quaternary deposits (Veklitch, 1993; Krokhmal et al., 2011) and with that of the neighbouring Russian territory (Dodonov et al., 2006). The correlation was based on combined geological, palaeopedological, and partially palaeobiological criteria. Our scheme was also correlated with terrestrial Pleistocene stratigraphy in Poland, Russia and Western Europe, and with marine isotope stages (Zagwijn, 1985; Lisiecki and Raymo, 2005; Lindner et al., 2002a, 2002b; Velichko et al., 2010; Head and Gibbard, 2015; Cohen and Gibbard, 2016).

4. Results

The deposit sequences in the individual sites were divided into main units of different origin and age. Three main lithofacies complexes are distinguished in the studied sites (except for one — Slokhyni): fluvial complex A (basal), glaciogenic complex B (middle), and loess–palaeosol complex C (upper).

4.1. Dubrivka site

This key site is located in the Stryvihor–Dniester interfluve, on the second (from top) hypsometric level, i.e. terrace VI: about 40–50 m above the Dniester River channel and 50–60 m above the Stryvihor River channel. The research was conducted within the extensive excavation of an active brickyard (Fig. 2, Supplementary material 1 — photo 1.A). Due to its continued exploitation, it was possible to gain access to new exposures in successive years, and to systematize and verify previous observations (Łanczont et al., 2004; Bogucki et al., 2009b; Bogucki et al., 2011b; Terpiłowski et al., 2011b). The compiled scheme of the Dubrivka section is presented in stratigraphic order in Fig. 2.

Complex A (*fluvial*; *preglacial*) is estimated to be 4–6 m thick. Its upper part (up to 2 m thick) consists of large-scale sets of massive gravels (Gm) and accessory medium-scale sets of massive sands (Sm) (Fig. 2, Supplementary material 1 — photo 1.1). Gravels mainly consist of various Carpathian sandstones with a noticeable admixture of chert boulders with a diameter of > 0.3 m. Gravels are flat, well-rounded, usually with a diameter of 10 cm. In the transparent heavy mineral assemblage, occurring in a silt insert, the predominant minerals are garnets (over 45%), zircon (25%), and rutile (almost 21%); the amphibole content is very low (0.4%) and pyroxenes are absent (Supplementary material 3).

Complex B (*glaciogenic*; *Sanian 2 = Okanian Glaciation*; *MIS 12*) has a complex structure and shows distinct spatial variability in the thickness and relations between glaciolacustrine silts (unit B1) and diamicton (unit B2).

Glaciolacustrine deposits consist of horizontally-laminated silts (Th) about 50–100 cm thick, and rhythmite about 1.5 m thick, comprising horizontally-laminated thin (about 5–10 cm) silts (Th) and massive thin (about 1.5–2 cm) clayey muds (Mm) (Fig. 2, Supplementary material 1 — photo 1.2) with single clasts of Scandinavian gravels (G) embedded in them. These deposits contain many involutions as well as normal and/or reversed faults, which often form complementary systems. In the eastern part of the excavation, glaciolacustrine deposits are locally interstratified by large lenses of sandy diamicton with crystalline gravels. Glaciolacustrine silts (B1) are the main deposits on the eastern wall of the excavation. These greyish-yellow silts texturally resemble loess (Supplementary material 3). The high content (50–80%) of very resistant minerals indicates the high maturity of the material (Supplementary material 5). Differences in the micromorphology of grain surface in the silt fraction, both in the nature and degree of the microrelief (Supplementary material 4), are recorded in the vertical section of glaciolacustrine silts.

Mineralogically, the silt in the lower part of glaciolacustrine deposits is characterized by the occurrence of feldspars accompanied by quartz. In addition, sodium feldspars about 20 μm in diameter also occur in the fraction. The silt grains demonstrate great diversity in microrelief and degree of crusting (Supplementary material 4A–C; Woronko et al., 2011); the former suggests that a considerable part of the grains were derived from initial cycle of rock weathering (Supplementary material 4A).

In the diamicton lenses, the grains of coarse and fine silt distinctly differ with regard to their surface. Coarse silt grains are covered with a thick crust that masks any irregularities of the surface and have no signs of mechanical destruction (Supplementary material 4D), while fine silt grains are usually sharp-edged and completely fresh or with a slight degree of crusting (Supplementary material 4E, F). This may indicate their different origin. The former may be the result of chemical weathering of older deposits, probably including soils, and the latter may indicate mechanical weathering, e.g. frost weathering (Woronko and Hoch, 2011).

The mineralogical composition of the upper part of glaciolacustrine deposits is characterized by a high content of potassium feldspars that accompany quartz, especially in the coarse silt fraction. Their origin

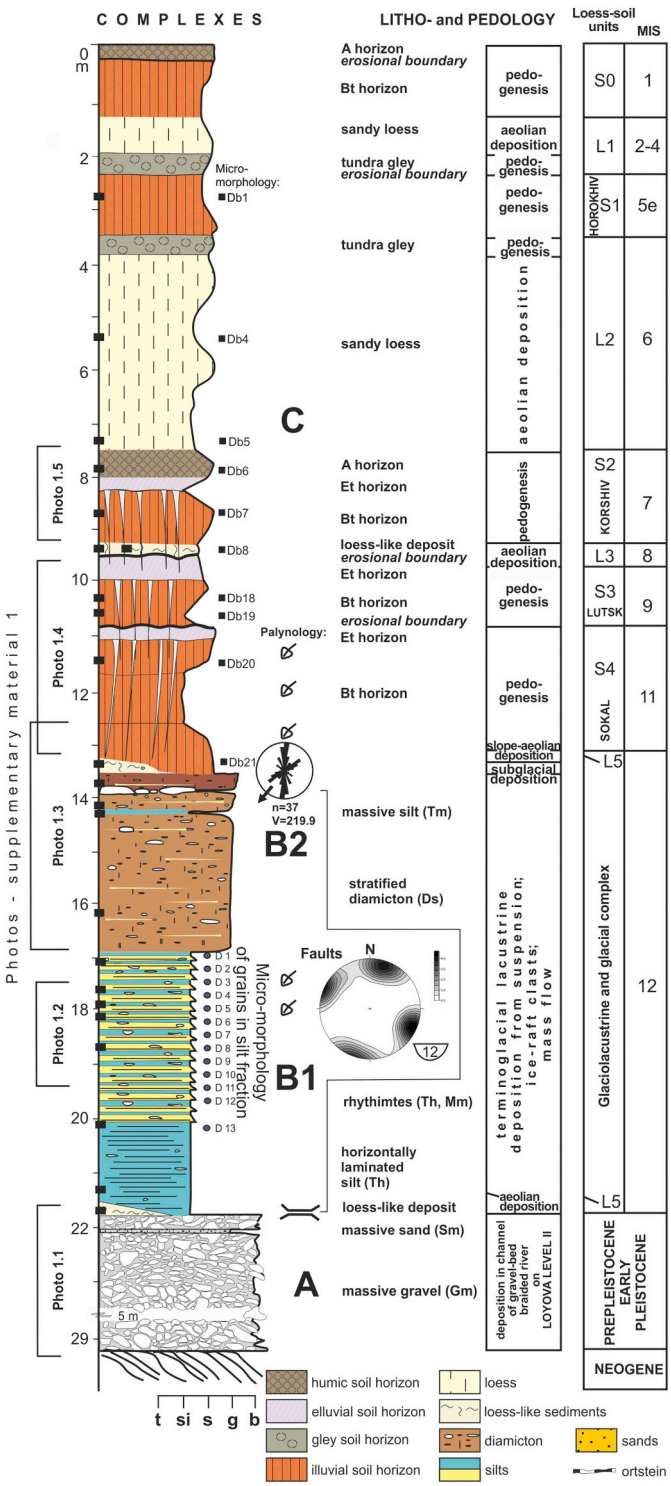


Fig. 2. Lithofacies complexes of Pleistocene deposits in the Dubrivka site. Lithofacial code is presented in Table 2.

varies as indicated by the co-occurrence of grains with a completely fresh surface (Supplementary material 4G, H), grains strongly affected by chemical weathering (Supplementary material 4I), and grains with different degrees of crusting (Supplementary material 4J). Additionally, small, very fresh and quite smooth conchoidal fractures are observed on the surface of many grains. It is also noteworthy that some grains display overgrowth quartz (Supplementary material 4K), which indicates that they originated from materials like quartzitic sandstones that slowly weathered in subaerial conditions. A considerable proportion of the grains also originated from weathered tills or other deposits where strongly mineralized solutions rich in the clay fraction circulated. Mechanical weathering, in this case most probably frost weathering, was the last process that affected the studied deposit in situ. It is indicated by the occurrence of grains with a partially fresh surface, without crust (Supplementary material 4L). In the top of the glaciolacustrine deposits, bowl-shaped depressions filled with stratified diamicton (Ds) with a large number of boulders (single ones even > 0.5 m) were found.

In the western part of the excavation, a series of diverse deposits of diamicton type, 1.5–2 m thick, constitutes unit B2. Small (up to 5 cm) throws of silt and diamicton beds were observed along faults at the boundary between glaciolacustrine and diamictic deposits. The fault planes dip (71°–87°) towards the S. The deformations display a N–S and/or WNW-ESE vergence. The faults were found to disappear in the bottom of the diamicton unit. Beds of massive and stratified diamictons (Dm and Ds) (Fig. 2, Supplementary material 1 — photo 1.3) and discontinuous beds of massive silts (Tm) combine to form a package stretching generally along a N–S direction. The long axes of clasts dip in different directions but mostly towards the S. This package is overlain (depositional contact) by a thin sheet bed of diamicton that has features of basal till of lodgement type, with the near-bottom concentration of clasts (origin of lithofacies Dm; cf. Evans et al., 2006). This till was probably deposited at the base of the advancing ice sheet, as indicated by the following features: the sheet form of the bed, its massive structure, and distinct preferred orientation of the long axes of clasts, evidencing the ice-sheet advance from the N to S.

Complex C (*aeolian*; *Middle–Upper Pleistocene*) consists of silty (loess/loess-like), silty-sandy or silty-clayey deposits, which are carbonate-free in the whole section with a total thickness of over 12 m (Supplementary material 3).

Only the younger loess units have the features of typical loess, i.e. quite thin and gleyed loess L1 and considerably thicker loess L2, both of aeolian–slope facies; the stratified older loess series in particular, have been almost completely transformed by pedogenesis and are preserved only locally as thin layers of loess-like deposits. Altogether, four superposed palaeosols of interglacial rank S4 to S1 (S0 is Holocene soil) occur in the complex. In this sequence, the lowest soil S4 (Sokal) (Fig. 2, Supplementary material 1 — photo 1.4) has developed on the massive diamicton with boulders, or on the glaciolacustrine silts and patches (pockets/tongues) of diamicton. Long-lasting chemical weathering has resulted in the formation of crusts on the silt grains and the uniformity of surface microrelief on silt grains in the C horizon of this soil. Soil S4 forms a set together with the directly superimposed soil S3 (Lutsk) (Fig. 3M–O). The successive soils S2 (Korshiv) and S1 (Horokhiv) are separated by loess L2. Two gley horizons with numerous Liesgang rings occur in the top part of L2 and in the bottom part of L1.

All palaeosols are typologically similar. These are mature palaeopedons of luvisols (lessivé) soils, with well-developed Et and Bt horizons. Except for soil S2, the pedons are truncated, and in the case of soil S1 only the illuvial horizon has been preserved. The Et horizons have platy structure and contain numerous black manganese–iron nodules. The Bt horizons consist of 2–3 subhorizons, are enriched with the clay fraction, strongly gleyed, have a characteristic reddish-brown color, and contain streaky concentrations of numerous large black Mn—Fe nodules. They have a well-developed aggregate structure (granular and prismatic, and locally platy in the bottom part). The illuvial horizons contain numerous concentrations of clay in the form of coatings inside the fissures and channels (Supplementary material 5). Based on their occurrence, these horizons can be assumed to be Bt–argic. Additionally, soil S2 (Fig. 3, Supplementary material 1 — photo 1.5) is a pedocomplex composed of well developed luvisol and superimposed humus soil.

In all soils, numerous traces of transformation are visible in the form of dense networks of fissures filled with grey gleyed clay material. The top parts of fissures occurring in soils S4 and S3 are inclined as a result

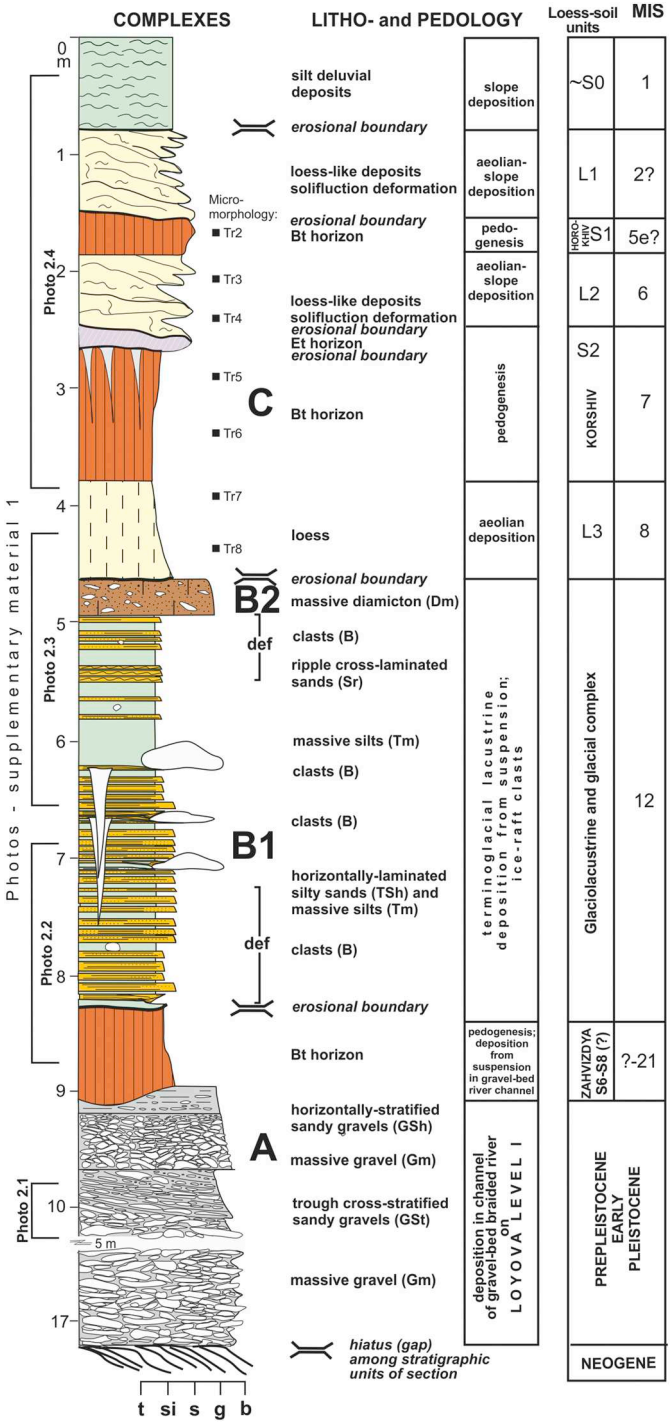


Fig. 3. Lithofacies complexes of Pleistocene deposits in the Torhanowychi 1 site. Other explanations — see the legend of Fig. 2.

of postpedogenic solifluction. Cryogenic transformations in the form of deformations of illuvial microforms are clearly visible under the microscope, especially in the oldest soils, where colloidal clay occurs as irregular pellets distributed chaotically in the groundmass. The concentrations of iron oxides are visible in the form of diffuse nodules of different sizes, coatings around fissures, and hypocoatings inside the walls. Such forms of iron compounds indicate their post-pedogenic origin (Supplementary material 5).

The composition of transparent heavy minerals is more varied in complex C than in the glaciolacustrine deposits. The content of garnets (not very resistant to chemical weathering) is the lowest in palaeosols (2–20%) and the highest in unweathered loess (35–60%) where amphiboles (up to 10%) and pyroxenes (up to 3%) also appear. The content of resistant zircon shows the opposite trend: it is lower in loess (10–20%) than in soils (25–40%).

4.2. Torhanovychi 1 site

The site is located on terrace VI of the Stryvihor–Dniester interfluve, lying at 365–375 to 381 m a.s.l.: 50–60 m above the Dniester River channel and 45–50 m above the Stryvihor River channel (Fig. 3, Supplementary material 1 — photo 2.A). Isolated hills can be seen, rising 20–30 m above this terrace and reaching 400 and 406.7 m a.s.l.; these are probably denuded fragments of the Krasnoy level. The studied Pleistocene cover in the Torhanovychi 1 site constitutes the key section (Bogucki et al., 2010, 2011c; Terpiłowski et al., 2011b). A better description of older Pleistocene deposits is obtained from the supplementary Torhanovychi 2 site, located 1 km to the west (Bogucki et al., 2011d; Yatsyshyn et al., 2011c). In both cases, observations were conducted on the walls of road gullies leading from the bottom of the Dniester river valley towards the plateau (Fig. 3A). Three main lithofacies complexes (A, B and C) were distinguished in the Torhanovychi 1 section (Fig. 3).

Complex A (*fluvial*; *preglacial*) comprises a series of deposits representing a complete fluvial cycle: from gravels to soil developed on overbank silt deposits with a thickness of 10 m. The main part of this series is a gravel layer composed exclusively of Carpathian material. The upper part of the series is a layer about 2 m thick, consisting of extensive gravel beds — mostly large or medium-scale sets of massive gravels (Gm) (Fig. 3, Supplementary material 1 — photo 2.1) and accessory medium-scale sets of sandy gravels with trough cross-stratification (GSt) and horizontal stratification (GSh) forming the rhythms: Gm → GSt and Gm → GSh. Gravels are small, usually 2–10 cm in diameter, the largest rarely reach 30 cm. The arrangement of the elongated gravels is usually chaotic or imbricate. A depression filled with silt beds is present in the top of the gravel layer; however, its original structure has been blurred by pedogenesis. The upper horizons of palaeosol have been eroded. The preserved fragment of the illuvial horizon is characterized by a higher content of the clay fraction: it is a rusty-brown color due to the occurrence of iron oxides, and a mature prismatic structure. Distinct large casts of segregated ground-ice structures indicate post-genetic frost action (Bogucki et al., 2010).

Complex B (*glaciogenic*; *Sanian 2 = Okanian Glaciation*; *MIS 12*) is < 4 m thick and consists of two parts: glaciolacustrine deposits (unit B1) and glacial till (unit B2).

The bottom part of unit B1 is a rhythmite of horizontally-laminated sandy silts (TSh) in sets up to 3 cm thick and thin (about 1.5 cm) massive silts (Tm). The upper part is a thick (about 1.5 m) package of massive grey silts (Tm) (Fig. 3, Supplementary material 1 — photo 2.2). Between these deposits, accessory fine sands with ripple-cross lamination (Sr) are present in sets with a maximum thickness of 5–10 cm (Fig. 3; Supplementary material 1 — photo 2.3). All these deposits share embedded clasts (boulders) of Scandinavian and local rocks. The entire unit shows signs of deformations in the form of fissures and involutions. A genetic connection was observed between a bipartite ice-wedge cast and involutions in the form of folds that disappear with increasing distance from the ice-wedge cast.

Unit B2 is a thin yet massive silty-sandy diamicton (Fig. 3; Supplementary material 1 — photo 2.3). Ductile and brittle deformations are observed at the contact of the diamicton with the underlying deposits of unit B1. Small isoclinal folds, involving sandy-silty deposits of the underlying layer, are clearly visible. They are also accompanied by reversed faults and systems of complementary faults, with a fault throw of up to 10 cm, which disappear in the bottom of the diamicton.

A diamicton with the same thickness occurs also above the glaciolacustrine deposits in the Torhanovychi 2 site. A large number of boulders was found in this bed.

Complex C (*aeolian–soil*; *Middle–Upper Pleistocene*) is a sequence of subaerial, completely decalcified loess-soil deposits (Fig. 3; Supplementary material 1 — photo 2.4). This section, occurring at a high altitude, is stratigraphically reduced and incomplete due to denudation processes. The sequence consists mainly of palaeosols: the thin bed of stratified loess (with a content of silt fraction > 60%) occurs only in its bottom part. Two truncated palaeosols occur on top of the loess. The lower soil, consisting of two micromorphologically well-developed Et and Bt horizons (Supplementary material 5), is clayey-silty, strongly gleyed. It probably developed in several stages, as evidenced by the erosion boundary between the Et and Bt horizons. Fissures running downward from this boundary are inclined along a fossil slope. Casts of segregated ground-ice structures indicate cryogenic deformations. The upper soil consists only of the lower part of the Bt horizon. Fragments of the destroyed upper horizons of both soils are incorporated into the silty-clayey deluvial-solifluction covers with loess inserts; these deposits are characterized by lobate structure. The described soils are probably palaeosols S2 and S1. The top part of the section consists of deluvial deposits, probably of Holocene age.

4.3. Slokhyni site

The Slokhyni site is located in the Stryvihor River valley, in the bottom part of the slopes of Mt. Radych (Fig. 1B, C). On the E and SE slopes of this massif, the Krasnoy level forms a narrow bench. The Loyova level is situated at a relative height of 60–70 m. The lower part of the slope of Mt. Radych passes into the bench of terrace IV of the Stryvihor River valley which is covered by loess (Bogucki, 2008, Bogucki et al., 2011a; Łanczont et al., 2015).

The Slokhyni research site is located in a rather extensive excavation where loam is exploited for a local brickyard (Fig. 4, Supplementary material 1 — photo 3.A). The sloping exploitation floor contains an exposed till layer covered by loess, directly over the steeply dipping layers of Miocene rocks.

Complex B (*glaciogenic*; *Sanian 2 = Okanian Glaciation*; *MIS 12*) is bipartite. The bottom part consists of sandy-silty series with a low content of erratics — a result of deposition by a subaqueous environment. The upper part is a strongly-weathered diamicton (Dm) with dispersed clasts (Fig. 4; Supplementary material 1— photo 3.1). The concentration of clasts varies: in the upper part of the slope and in the top part of the diamicton, the proportion of boulders is high; these are rocks of local, Carpathian and Scandinavian origin. Interlayers of fine sandy-silty deposits (TSh), occurring locally (Fig. 4; Supplementary material 1 — photo 3.2), are associated with the incorporation of the underlying deposits. The contact between these two series is deformational. The main deformation structure is a large-scale thrust-fold structure involving the packages of till and sandy-silty series.

Complex C (*aeolian*; *Middle–Upper Pleistocene*) consists of two loess series. The older unit C1 is a loess–palaeosol sequence, about 7–8 m thick, which overlies the till of complex B. The bottom part of this sequence consists of loess L5 (Fig. 4; Supplementary material 1 — photo 3.3) and a set of two palaeosols: S4 (Sokal) and S3 (Lutsk). These soils (Fig. 4; Supplementary material 1 — photo 3.4) are luvisols; their profiles are incomplete. The presence of a cryogenic wedge structure at the top of the S4 sequence indicates cooling between two phases of pedogenesis. The loess-like deposit overlying soil S3 corresponds to L3. A relict soil has developed on this deposit; although it appears to be of luvisol type, its stratigraphic position is difficult to determine (S2?, S1?). This is exhumed soil, cut by a dense network of gley fissures, with superimposed modern initial soil.

A common mantle is formed of the deposits corresponding to L3, the loess deposits and a set of younger deposits. This has been distinguished as unit C2, occurring within the accumulation cover of terrace IV.

The accumulation cover of terrace IV is about 11 m thick. Its bottom, alluvial part consists of channel gravels and an overbank series of sandy silts covered by the deposits of unit C2. Its lower part consists

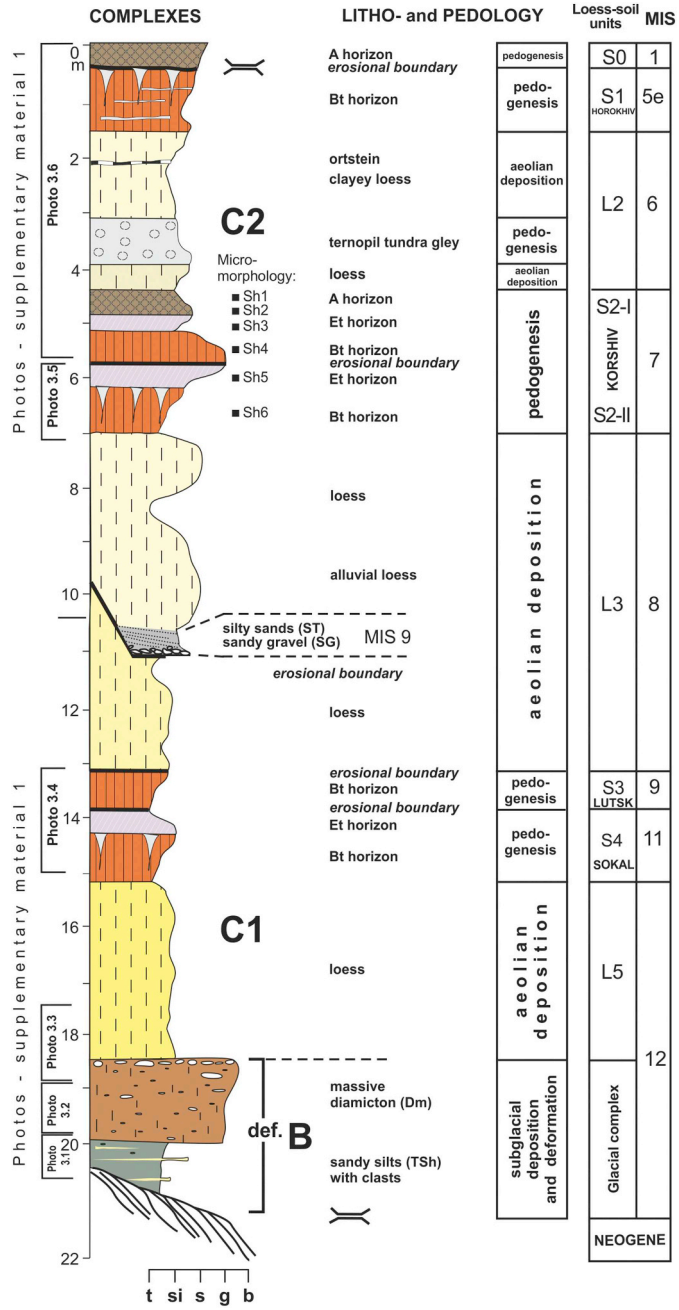


Fig. 4. Lithofacies complexes of Pleistocene deposits in the Slokhyni site. Other explanations — see legend of Fig. 2.

of loess-like sandy silts representing L3. The upper part of unit C2 is better developed and less destroyed. The S2 set of palaeosols (Korshiv) consists of two superimposed luvisols. The lower soil (S2–II) is truncated and cut by a dense network of gley fissures (Fig. 4; Supplementary material 1 — photo 3.5). The upper soil (S2–I) has a complete profile (A–Et–Bt). The Bt horizons of these soils have several micromorphological features characteristic of illuvial horizons deformed by cryogenic processes (Supplementary material 5). The eluvial horizon of the younger soil is characterized by the occurrence of features (cappings, platy microstructure) indicating that the finest fraction has been removed. The stratigraphic position of this soil is evidenced by a Middle Palaeolithic artefact (sidescraper) which has been found in its humus horizon (Bogucki, 2008, Bogucki (Boguckyj) et al., 2009; Łanczont et al., 2015). The overlying gleyed loess L2 (2 m thick) is divided by an interstadial tundra gley level with the regional name Ternopil (Table 1).

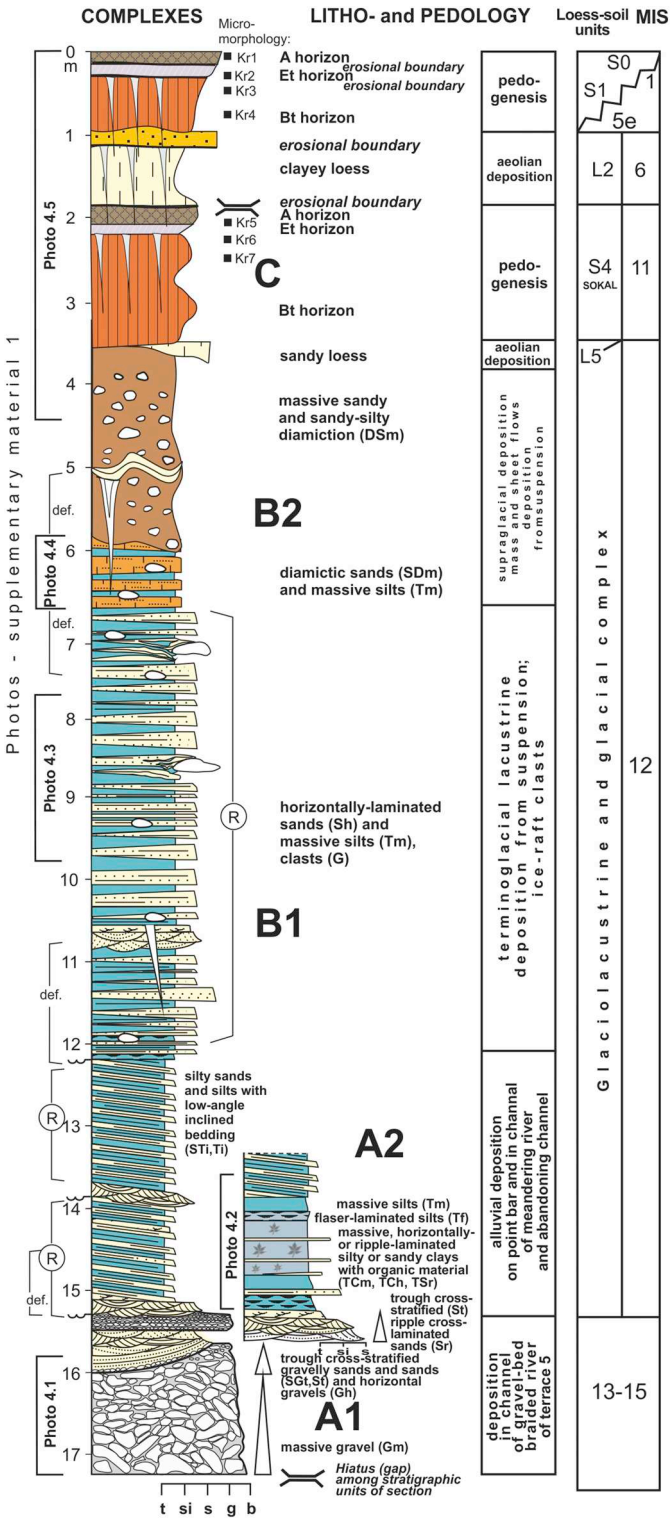


Fig. 5. Lithofacies complexes of Pleistocene deposits in the Kruzhyky site. Other explanations — see the legend of Fig. 2.

The top part of the profile consists of truncated palaeosol S1 with features of a relict soil; it is exhumed and partially transformed by modern pedogenesis (neosol S0) (Fig. 4; Supplementary material 1 — photo 3.6).

4.4. Kruzhyky site

The Kruzhyky site is located on the right bank of the Dniester River. The sequence of deposits forming terrace V is exposed along an approximately 150 m-long section in a steep, 15–17 m high bank descending directly to the river channel (Fig. 5, Supplementary material 1 — photo 4.A). This sequence consists of fluvial, glaciogenic and aeolian deposits (Łanczont et al., 2010; Zieliński et al., 2011).

Complex A (*fluvial*; *Middle Pleistocene*) is bipartite (Fig. 5).

The lower part of complex A was defined as unit A1 (fluvial; Cromerian, (?) interglacial IV, *vide* Table 1). Only the upper 1 m of this unit, i.e. the part above water level, has been described. It is a gravelly-sandy fining-up sequence composed of exclusively Carpathian gravels with a massive structure or indistinct stratification (Gm) (Fig. 5; Supplementary material 1 — photo 4.1); these pass laterally into gravelly sands and sands with trough cross-stratification (SGt, St), overlain by a medium-scale set of gravels with horizontal stratification (Gh). Based on drillings, the full thickness of the channel series is estimated at about 10 m (Yatsyshyn et al., 2011a).

The upper part of complex A is unit A2 (fluvial; anaglacial phase of the Sanian 2 = Okanian glacial). It consists of two series with a total thickness reaching 4.0 m. The bottom sandy-silty, carbonate series consists, from bottom to top, of sands with small and medium-scale trough cross-stratification passing into sands with ripple cross-lamination (St → Sr), sandy silts with ripple cross-lamination (TSr), silts with flaser and horizontal lamination (Tf, Th), and massive organic silts with numerous macrofossils (TCm) forming a lens about 9 m long and up to 1.5 m thick (Fig. 5; Supplementary material 1 — photo 4.2; more information in Supplementary material 6). Involution-type deformations including silt deposits can be seen. Alternating layers of fine sands and silts with parallel lamination and low-angle inclined stratification (STi, Ti) occur laterally (Fig. 5; Supplementary material 1 — photo 4.3). The upper series of sandy-silty deposits consists of sands with trough cross-stratification (St), sandy silts and silts with low-angle inclined stratification (STi, Ti), and accessory sandy silts with ripple cross-lamination (TSr).

Complex B (*glaciogenic*; *Sanian 2 = Okanian Glaciation*; *MIS 12*) is bipartite. The lower unit B1 (proglacial) is rhythmite, about 8 m thick. It consists of sands and sandy silts with horizontal lamination and silts with horizontal lamination or massive structure (Sh, Tm) (Fig. 5; Supplementary material 1 — photo 4.4); lithofacies of sands with small-scale trough cross-stratification occur locally (St). Two levels of deformations occur within this unit: a lower level formed of involutions and ice wedge/fissure casts and a higher level represented by folds and flexures with an SW vergence. The upper unit B2 (glacial) is up to 2 m thick. Its bottom part consists of massive diamictic sands (SDm) (Fig. 5; Supplementary material 1 — photo 4.4), and the upper part comprises massive sandy or sandy-silty diamicton (DSm) with a high content of gravel material.

Complex C (*aeolian*; *Middle–Upper Pleistocene*) is silty and sandy-silty, carbonate-free, with a total thickness of almost 4 m. Most of it has been transformed by strong soil-forming processes of different ages. The complex generally consists of two soil units. The older one is a well-developed luvisol with a complete profile that developed directly on the gravelly-sandy diamicton and locally on silt deposits (L5). This palaeosol has several features that are very typical of soil S4 (Sokal) (Fig. 5; Supplementary material 1 — photo 4.5). The A horizon shows features of redeposition. The Et and Bt horizons have an aggregate structure and are cut by a dense network of gley fissures. This soil is covered (with a large hiatus) by loess-like sandy-silty deposits L2. The overlying younger soil unit is a pedocomplex which consists of two mature soils (S1/S0): the younger, "modern" neosol S0 formed in the top of the relict exhumed palaeosol S1. The clay infillings of fissures occurring in unit L2 and the described palaeosols have features (deformations of illuvial microforms and lenticular microstructure) that unambiguously indicate post-pedogenic cryogenic transformations (Supplementary material 5).

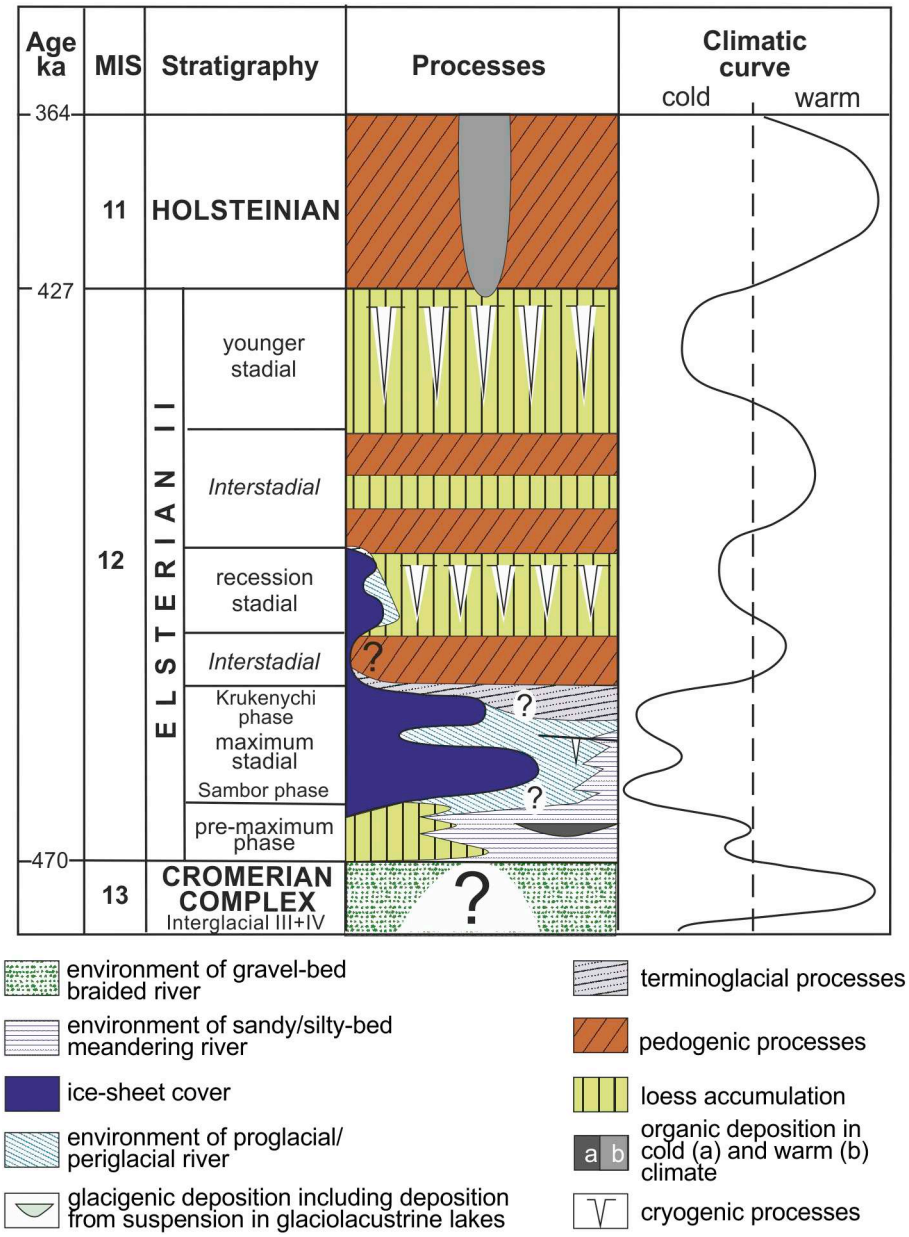


Fig. 6. Variability of depositional environments in the NW part of the Eastern Carpathian Foreland (between the San and Dniester rivers) in MIS 13–11 according to Bogucki et al. (2004) and Łanczont et al. (2010, modified).

5. Morphogenesis of the upper Dniester River valley before the ice-sheet advance (preglacial period)

A period of far-reaching consequences for the Quaternary morphogenesis of the Eastern Carpathian Foreland was the time of Elsterian II = Sanian 2 = Okanian Glaciation (MIS 12). At the time of the ice-sheet advance, the relief of the Carpathian foreland was smoother than the present-day relief, i.e. the relative heights between the watersheds and the bottoms of river valleys (the present-day terrace V) were small, estimated at 40–45 m.

The Loyova level (terrace VI) was mainly shaped by the waters of the Dniester, Stryvihor, pre-Wyrwa, and pre-San rivers, which formed a system of braided rivers in the flat plain of the Carpathian foreland (Teisseyre, 1938; Yatsyshyn et al., 2011c). Gravelly series covering individual surfaces of the Loyova level developed in quite similar depositional conditions and reached a similar thickness of 5–8 m. On the highest surface (Torhanovychi) and the second highest surface (Dubrivka), lithofacies of massive gravels can be seen; these were probably deposited as longitudinal bars or gravel sheets during flood peaks (cf. Williams and Rust, 1969; Hein and Walker, 1977). Single large boulders, even those frozen to ice floes, may have been moved during floods; in addition, the surfaces of longitudinal bars were locally washed out or gravel sheets built up during waning flood stages (cf. Zieliński, 1997; Lunt and Bridge, 2004). The analysis of gravels in the Torhanovychi area indicated that rivers flowed most probably to the NE at that time (Yatsyshyn et al., 2011c; Teisseyre, 1938). In the Torhanovychi 1 site, the deposits of a complete fluvial cycle have been preserved: the gravels are locally covered by silts settled from suspension in an abandoned channel (cf. Rust, 1978), with visible traces of a soil that had developed on them (Bogucki et al., 2010; Terpiłowski et al., 2011b). This palaeosol has been partially destroyed; the preserved B horizon has features of a mature soil, and was TL dated at about 775 ka (Supplementary materials 1 and 5). It is probably a counterpart of possibly the oldest soil unit of four superimposed paleosols with the regional name Zahvizdja pedocomplex (*vide* Table 1). The stratotype of this unit has been distinguished in the Zahvizdja site near Ivano-

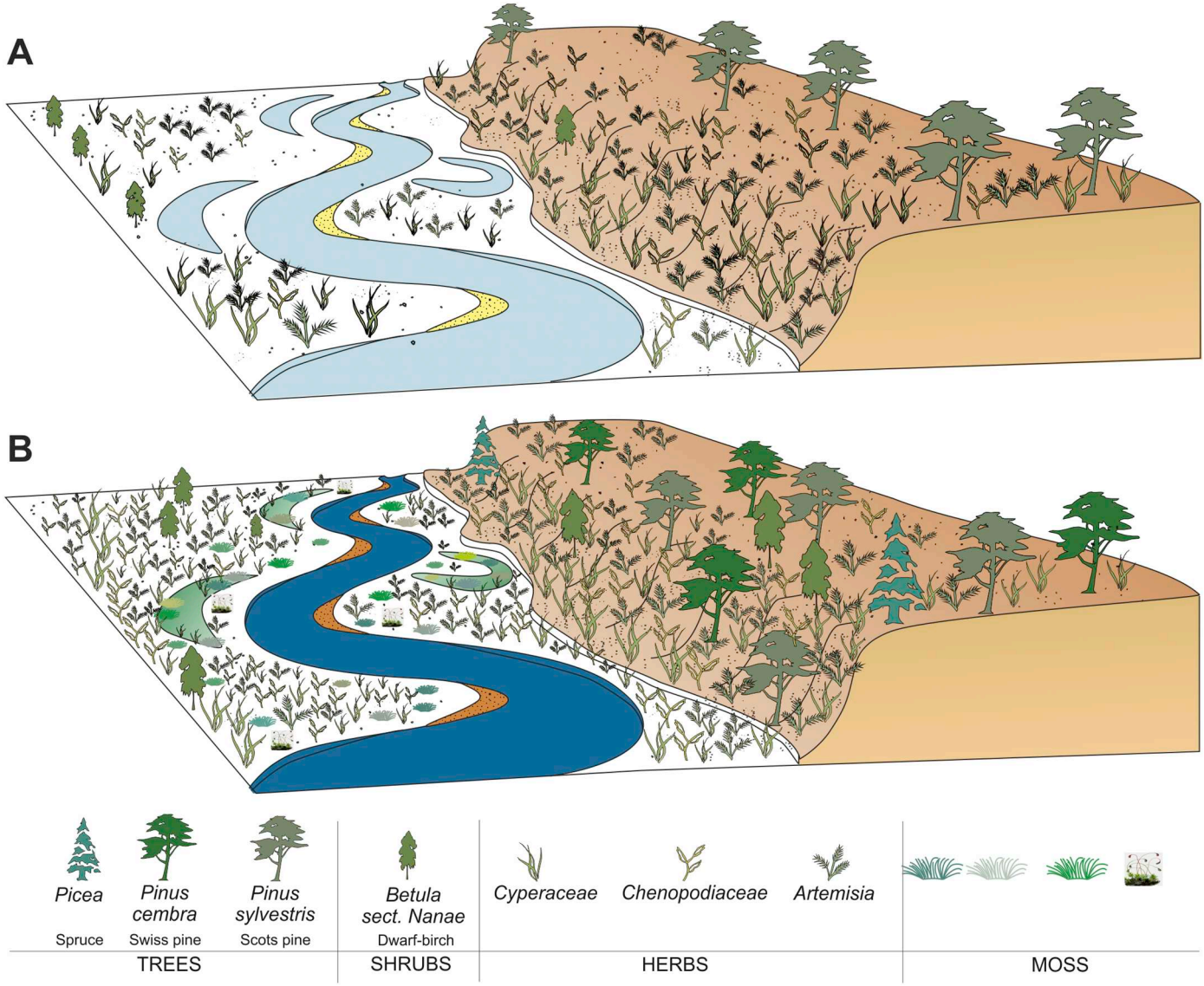


Fig. 7. Simplified reconstruction of vegetation cover in the older part (A) and younger part (B) of the pre-maximum phase of the Elsterian II Glaciation in the vicinity of the Kruzhyky site after Łanczont et al. (2010). Other explanations in the text.

Frankivsk, situated in the Bystrytsia Solotvynska–Dniester interfluve (130 km towards the SE from the study area). The Pleistocene sequence in this profile consists of fluvial deposits of terrace VI, a set of Zahvizdja palaeosols (S8–S6) in the form of four forest soils with traces of very intense postgenetic convolutions, and typical loess and palaeosol layers from L6 to S2 (Łanczont and Bogucki (Boguckyj), 2002; Łanczont et al., 2003a). The TL ages of the above-mentioned deposits range from 800 to 200 ka (Kusiak, 2002). The Matuyama/Brunhes boundary was identified within soil S7 (Nawrocki et al., 2002). In general, palaeomagnetic and TL results correspond closely with the expected age and stratigraphic interpretation of the profile based on wide criteria. The Zahvizdja set of soils corresponds to the Martonosha and the younger part of Shyrokhino units (MIS 21–17) in the loess cover of the Central Ukraine (see Table 1).

In the glaciated part of the Eastern Carpathian Foreland, a large stratigraphic hiatus exists between the described gravel series of terrace VI (Torhanovychi, Dubrivka) and the directly overlying deposits related to MIS 12.

Terrace V is the oldest intra-valley terrace in the Dniester River catchment. In the Stryvihor–Bolozhivka interfluve, the fragments of terrace V and terrace VI form the clearly visible horizontal outline of palaeovalleys with the NW-SE orientation, which indicates that the pre-San and pre-Wiar rivers were tributaries of the Dniester River (Yatsyshyn et al., 2011a).

The palaeogeographical conditions that prevailed in the period immediately preceding the ice-sheet advance were reconstructed on the basis of the features of the deposits covering terrace V in the Dniester River valley in the Kruzhyky site. This was the final period of the formation of the terrace. It consisted of two stages linked with the functioning of the Dniester River, with a different channel type, and probably occurred in the transition period between the warming of MIS 13 and the initial phase of the cooling of MIS 12 (Fig. 6).

In the older stage, this area was a gravel-bed braided river. Gravels were deposited as longitudinal bars as a result of high-energy flood flows (cf. Williams and Rust, 1969), while sandy-gravelly lithofacies represent the deposition in the interbar channels during waning flows (cf. Zieliński, 1997; Zieliński et al., 2011). In the younger stage, the Dniester became a meandering river. This is indicated by the presence of lithofacies Si, which are recorded in the lateral migration of the point-bar (cf. Ghazi and Mountney, 2009; Zieliński et al., 2016). This state of affairs rarely occurs in a cold climate; however, one such example of a river was documented in Alaska, in the transition period

between MIS 8 and MIS 7 based on fluvial deposits (Zieliński et al., 2016; Sokołowski et al., 2019). Vandenberghe (2001) proposes that the river can function in the presence of permafrost, assuming that flow energy is low; in which case, mainly highly cohesive material is deposited, and the area is characterized by a continuous cover of tundra vegetation. All these conditions were met in the Kruzhky site.

The main deposits, i.e. silts and sandy silts, were deposited in the active channel as point-bars, and in the abandoned channel. The presence of fluvial aggradation indirectly indicates a rise of the base level, and thereby a decrease of flow energy. Cold Subarctic climate conditions are evidenced by the results of the analysis of organic deposit (Supplementary material 6) occurring as a lens within the silts. Therefore, at that time, the Dniester River near the Kruzhyky site meandered in the open landscape of the tundra colonised by *Selaginella selaginoides*. The valley bottom was overgrown by moss-sedge communities with grasses, horsetail, and locally dwarf birch and single willow shrubs (Fig. 7A). The conditions favourable to the existence of Ostracoda lasted a short time but the species found in the site are characterized as common and tolerant of a wide range of water temperatures. The occurrence of only such species indicates that shallow, small but permanent water bodies, possibly of thermokarst origin, developed between sedge tufts growing on the plain. The depth of these water bodies reached only tens of centimetres, as indicated by the occurrence of *Cypridopsis vidua* (Łanczont et al., 2010). On the valley sides, tundra vegetation formed a mosaic with steppe-like vegetation (grasses, *Artemisia*, Caryophyllaceae, Asteraceae) growing in drier habitats. Single trees, such as pine and, very rarely, birch, were also found. It cannot be excluded that pine was more common in the nearby Carpathians.

The final development stage of the above-described environment in the bottom of the Dniester River valley was characterized by a slightly drier and warmer climate, perhaps of interphase rank. The vegetation cover in the Kruzhyky environs was composed of sedge, grasses, and *Artemisa*. Mosses grew in wet habitats, with valerian present in the wettest ones. The former water bodies were overgrown with carpets of *Sphagnum* and brown mosses. Trees (mainly pine and rarely birch) were still rare in the landscape but other individual trees, such as oak and spruce, appeared on the slopes (Fig. 7B).

6. Glacial history of the upper Dniester River valley — chronology and stratigraphy

In the pre-maximum phase of the glaciation, the aeolian accumulation of loess L5 (see Fig. 6) occurred probably on a small scale but in some places it survived the destructive activity of the advancing ice sheet or its waters, as exemplified by the Dubrivka site and that of Siedliska, near Przemyśl, which is located 30 km towards the NW (see Fig. 1B). Weathered Carpathian flysch rocks were the only source of material for the loess L5 deposited during the time directly before of the glacial advance in the study area (Łanczont, 1997b; Racinowski et al., 2003). Malacofauna found in this loess in the Siedliska site forms an extremely poor assemblage of very numerous specimens, represented only by *Succinea oblonga elongata* Sandberger. This fauna, typical for loess, indicates an open, quite humid habitat, and cold, subpolar climate, i.e. an environment of tundra type (Łanczont, 1997b).

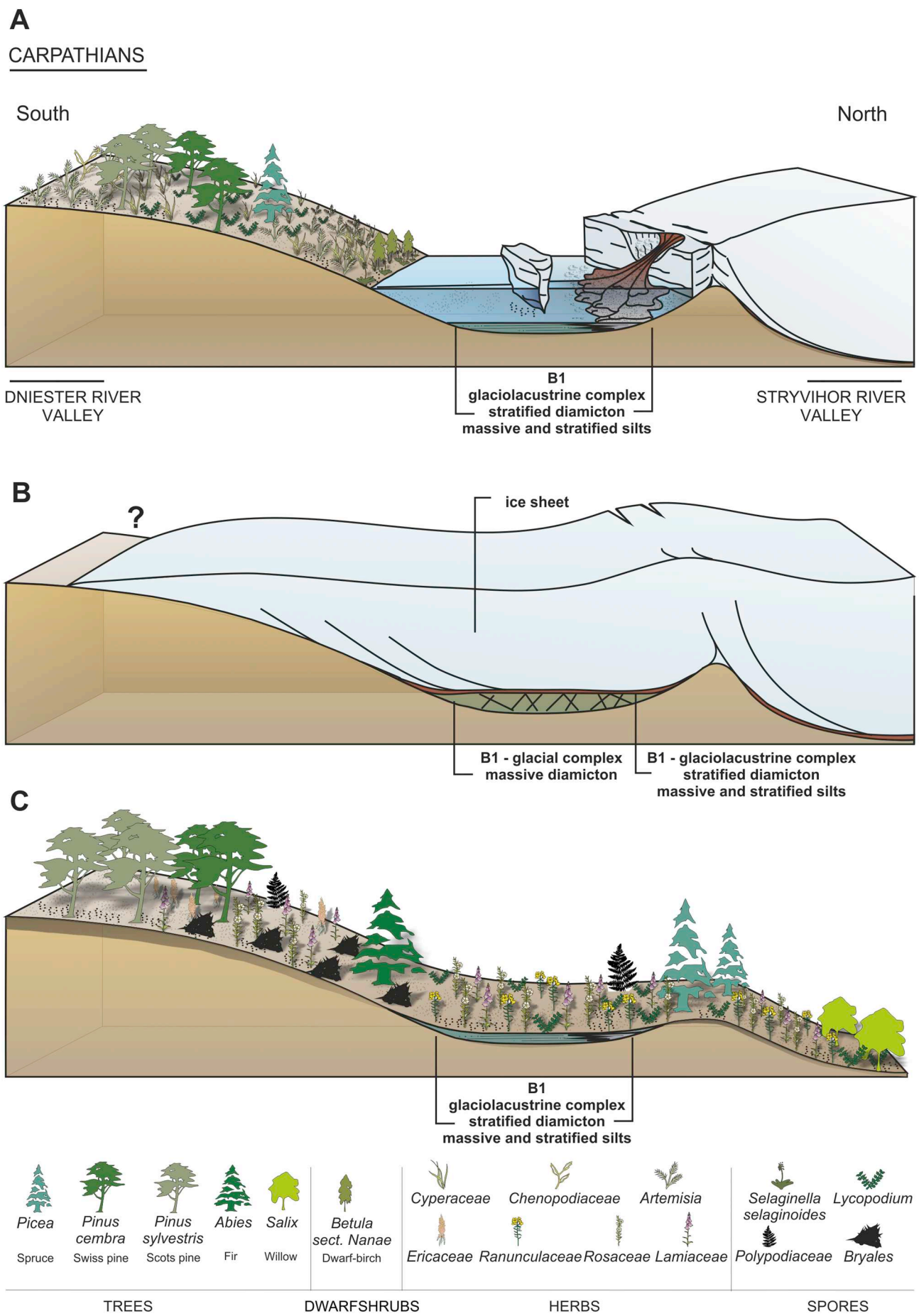
The distribution of glacial deposits in the near-Carpathian part of the Dniester River catchment allows a reconstruction of the history of the ice-sheet advance and the evolution of depositional environments in the periods when: 1) the ice sheet encroached on the study area, 2) terminoglacial lake (lakes) were formed (Fig. 8A); 3) the ice sheet reached its maximum extent (Figs. 1C, 8B).

The Sanian 2 (Elsterian II = Okanian, MIS 12) ice sheet, after passing by the bend of the Carpathian margin (Fig. 1C; cf. Łanczont et al., 1988; Łanczont and Racinowski, 1994), encroached on the Eastern Carpathian Foreland and spread to the SE as the Dniester lobe. Initially, its advance would have been fast, as indicated by the occurrence of thin deformed till directly on the bedrock on the slopes of Mt. Radych, at an altitude of 345–350 m a.s.l. (Slokhyni site). The occurrence of a single erratic found on the slope of Mt. Radych at an altitude of 410 m a.s.l. (Teisseyre, 1938), i.e. at a similar altitude as the erratics found in the Western Carpathians (Dudziak, 1961; Klimaszewski, 1936), may indicate that ice masses piled up on the slopes of Mt. Radych, the peak of which was probably a kind of nunatak. Ice masses may have piled up also near the Carpathian margin, in a narrow gorge of the Wyrwa Dobromilska River valley, along which the ice sheet advanced towards the Stryvihor–Dniester interfluve area (Demediuk and Stelmakh, 1980); the thickness of the ice sheet in this gorge was estimated at 90 m (Przepiorski, 1938) or even 120 m (Romer, 1906). From here, the ice sheet moved along the Carpathian margin, and thinned to an estimated 50 m (Herenchuk et al., 1972) or about 70 m (Romer, 1906). The present-day investigations show that, except for the case of Mt. Radych, the highest locations of glacial deposits occur at an altitude of 375–380 m a.s.l. (Yatsyshyn et al., 2011a), which may indicate that the thickness of ice masses was 40–70 m.

Glaciolacustrine deposits identified in the vicinity of Torhanovychi and Dubrivka, which are 5.5 km apart, indicate that during the Sambor Phase, the advance of the ice-sheet was slowed in the Stryvihor–Dniester interfluve (Fig. 6) by the morphological barriers created by the undulating relief of terrace VI. These conditions favoured the development of terminoglacial lake(s). The type of deposition indicates that these lakes, fed by extraglacial and proglacial waters, may have existed for a long time.

Near the Torhanovychi site, the lake was ephemeral; it was periodically drained. The glaciolacustrine series suspended in the stagnant water may have settled. The dropstones found in the silts were released from floating ice-rafts. This depositional style indicates the environmental conditions typical of a lake bottom plain (Brodzikowski, 1993). The original structure of rhythmite was deformed after periodic drainage of the lake, as evidenced by the occurrence of epigenetic bipartite ice edge cast (cf. Murton et al., 2000). Its growth caused the formation of involutions in the silt layer as a result of deposit compression (cf. Mackay, 1990; Murton et al., 2000). The occurrence of such thermal contraction structures, which are often accompanied by involutions, is considered an indicator of periglacial conditions in an area underlain by permafrost (Goździk, 1973); however, it is possible that they could have developed in ground affected by strong seasonal frost (Washburn, 1979). Of these two, the latter appears to better explain the development of thermal contraction structures near the Torhanovychi site, as indicated by the close relationship between glaciolacustrine deposition and the ice sheet. The final period of the lake's existence in this site is represented by the upper package of massive silts, whose bottom parts are characterized by deformations in their synsedimentary load structure (cf. Anketell et al., 1970).

Near the Dubrivka site, the lake developed in close proximity to the ice-sheet front. It was filled with diamictic deposits derived from subaqueous massflows, and finally covered by the advancing ice sheet. The rhythmites of glaciolacustrine complex can be considered as varves, i.e. annually-laminated deposits (De Geer, 1912; Ashley, 1975). Cyclic seasonal deposition from suspension in stagnant water is clearly recorded as thicker (summer) silt laminae and thinner (winter) clay laminae. Simultaneously with the formation of varves, dropstones were released from ice rafts floating in summer seasons. The deposition of varves in the lake was sometimes interrupted (and finally ended) by an intensive supply and deposition of material as a result of cohesive mass flows — corresponding to till of the waterlain type (cf. Piotrowski, 1994). The disappearance of this supply was followed by a slow and long-lasting deposition of silts from suspension (Brodzikowski and Van Loon, 1991). Based on the composition of transparent heavy minerals in glaciolacustrine silts (Supplementary material 3), it can be deduced that silt material originated mainly from the redeposited weathered Carpathian flysch rocks, and the weathering products of fresh glacial material were less important. Silt material was also transported to the lake by the wind, as indicated by microrelief of some silt grains (Woronko



(caption on next page)

Fig. 8. Landscape changes in the environs of the Dubrivka site during MIS 12–MIS 11: A) simplified reconstruction of vegetation cover near the terminoglacial lake in the older part of the maximum phase of the Elsterian II; B) model of development of the marginal zone of the Elsterian II ice-sheet; C) simplified reconstruction of the Holsteinian vegetation cover. Other explanations in the text.

et al., 2011). A series of TL dating results obtained for the glaciolacustrine silts seems to indicate that the age of their bottom parts is overestimated (Supplementary materials 1 and 7), which could be explained by the incorporation of older deposits washed in by water of the terminoglacial lake. Interestingly, the TL findings indicate that the silts occurring in level I of terrace VI in the Torhanovychi site are older than those of level II in terrace VI in the Dubrivka site. As the higher parts of the silts were TL dated at 470–435 ka in both sites, so they correspond to MIS 12 (cf. Preusser and Fiebig, 2009) (*vide* Supplementary materials 7 and 8).

Small patches of trees and shrubs occurred (*Pinus cembra*, *Picea*, *Betula* sect. *Nanae et Fruticosae*) in open landscape with a mosaic of separate steppe and tundra communities that prevailed near the Dubrivka site in the ice-free surroundings of the lake (more information in Supplementary material 9A). Hygrophilous *Lycopodium*, *Selaginella selaginoides* grew in the understorey. Open areas were covered by herbaceous vegetation represented by Cyperaceae, *Artemisia*, Chenopodiaceae (Fig. 8A). The climate was cold and wet, as in the Siedliska site. In general, this flora was not different from other Pleistocene glacial floras investigated in a few sites in central and eastern Europe (Van Andel and Tzedakis, 1996, 1998; Krokhmal and Komar, 2018). Small qualitative and quantitative differences in species composition resulted from local conditions.

The deposition in the lake completely stopped and glaciolacustrine environment turned into a glacial environment, as evidenced by the occurrence of basal till of lodgement type (litofacies Dm; cf. Evans et al., 2006) in the Dubrivka and Torhanovychi 1 and 2 sites. Therefore, this till serves as a record of the ice-sheet advance on the previously deposited glaciolacustrine deposits (Fig. 8B). This advance resulted in the development of discontinuous deformations: normal and/or reversed faults, often occurring in complementary systems. The vergence of faults is parallel (N–S) and/or perpendicular (WNW–ESE) to the direction of ice movement. Such bimodal directions indicate a vertical compression, i.e. glaciodynamic loading when glaciolacustrine deposits were covered by the ice sheet that moved from N to S. The TL age of the basal till in the Dubrivka site falls into the MIS 12 period (Supplementary material 8). In the Stryvihor–Dniester interfluve, this till was probably deposited in its highest locations (level I of terrace VI) under a thin layer of ice, as evidenced by its narrowness (Torhanovychi).

The further advance of the ice-sheet front during the Sambor Phase did not extend far towards the SE. Kruzhyky (Bogucki (Boguckyj) et al., 2007) and Kornalovychi (Demediuk and Demediuk, 1995) are the easternmost sites in the area of the East Carpathian Foreland. The following series of events is recorded in the Kruzhyky site. The deposits of the meandering river were initially covered by the sandy deposits of the distal part of the flat inundated fan of the Dniester River, which was blocked by the ice sheet advancing from the NE, i.e. from Kornalovychi. The proximity of the ice-sheet front is indicated by the occurrence of dropstones embedded in silts, which increases in an upwards direction. The fan deposits have almost identical lithological features in their whole section, which indicates the constant conditions of deposition. However, the deposition was sometimes interrupted. The presence of involutions and ice-wedge casts, which developed on the periodically emerging fan surface, indicates frost processes. Afterwards, the flow till was deposited — at first under water, and then on the emerging fan surface. It caused the development of strong deformations in the top part of the fan deposits. This would indicate that the ice-sheet front was then situated no more than 4–5 km from the Kruzhyky site.

During its retreat in the Krukenychi phase, the ice sheet stagnated on the northern side of the present-day European watershed. The extraglacial and proglacial waters flowed to the east along the Bolozhivka ice-marginal valley, which functioned in this role for a long time, as evidenced by the three interstadial terraces located on the sides of this valley (Yatsyshyn and Plotnikov, 2004). The transformation of the river network of the Dniester–Stryvihor system probably started at the same time as, or directly after, the retreat of the ice sheet (Romer, 1906; Bogucki (Boguckyj) et al., 1999; Yatsyshyn et al., 2008, 2011c).

During the retreat of the ice sheet, a weak, irregular accumulation of loess L5 occurred in the described area. Thin and discontinuous loess patches, with a maximum thickness of 2–3 m, accumulated on the slopes of Mt. Radych. Loess deposition was considerably more intensive in the lower part of the upper Dniester River valley (Halych Basin), where the thickness of L5 is 10 m. In the Halych site, the layers (= Tylihul loess unit in Central Ukraine) are believed to have formed under cold periglacial conditions with varying humidity during the whole MIS 12 (Łanczont and Bogucki (Boguckyj), 2002; Alexandrowicz et al., 2014).

7. The upper Dniester River valley after the ice-sheet retreat

The first interglacial after the Sanian 2 ice-sheet retreat opened a new chapter in the Quaternary history of the NW part of the Eastern Carpathian Foreland. The ice-sheet decay displays features of areal deglaciation (Bogucki (Boguckyj) et al., 1999). In the San-Dniester interfluve in the environs of Krukenychi, on the watershed plateaux with the relict landscape of post-glacial lakeland, the interglacial period is represented by organic deposits and mature luvisols that developed on the slopes of lake basins. The peat and peaty gyttja records offer an insight into the climate, environmental and vegetational history of the post-glacial palaeolake of the Krukenychy site during this interglacial. The succession of tree pollen in this site is indicative of the Mazovian (Holsteinian = Likhvin) succession. The macroflora remains reflect rich interglacial vegetation, comprising 67 taxa, and the presence of *Azolla interglacialis*, *Potamogeton pannosus*, and *Aracites interglacialis* clearly indicates the Mazovian (Holsteinian = Likhvin) age of this flora (Pidek et al., 2003; Łanczont et al., 2003b). Hence, it can be assumed that the deposits are correlated with MIS 11. In Western Europe this interval is represented by the Holsteinian Interglacial (Lindner et al., 2002a, 2002b; Krupiński, 2005; Lindner and Marks, 2012, 2015); however, the correlation of the Holsteinan interglacial with MIS 11 is sometimes questioned and it has been proposed to correlate more with MIS 9 (cf. Lauer and Weiss, 2018; Nitychoruk et al., 2006).

7.1. Long-term soil formation event during the MIS 11

Palaeosol S4 Sokal (Mazovian = Zavadivka), which is correlated with the MIS 11 (*vide* Table 1), developed directly on the deposits accumulated in the preceding glacial cycle, such as till (sites: Dubrivka and Kruzhyky), glaciolacustrine deposits (Dubrivka), or loess L5 from the final phase of the Sanian 2 Glaciation (Dubrivka, Kruzhyky, Slokhyni). The importance of the Dubrivka site should be stressed, as it is here that we can observe the spatial lithological variability of the parent material of soil S4. The highly advanced, post-depositional pedogenic transformation of these deposits reaches a depth of 1–4 m below the macroscopically-visible traces of pedogenesis in all the studied sites. There is no evidence of a hiatus between the deposition of the substratum of soil S4 and the pedogenesis phase.

Palaeosol S4 in loess profiles on the investigating area is a huge mature luvisol, with a very well-developed Bt horizon and a well-developed Et horizon. The A horizon is usually absent, being present only in the Kruzhyky site. A change of the climate conditions, correlated with the final stage of pedogenesis, is recorded as a dense polygonal

network of narrow fissures filled with grey gleyed material. Their deformation by solifluction processes is a record of the subsequent periglacial conditions.

The attempted reconstruction of interglacial vegetation, based on the pollen analysis of mineral soil material from the Bt horizon in the Dubrivka site, was not fully successful (Supplementary material 9B). The amount of pollen was small, and its composition is not typical of the succession of the climate optimum. At the interglacial time the site surroundings were covered by coniferous forest with pine, Swiss stone pine, spruce, fir, and heather in the undergrowth (Fig. 8C). Wet habitats were periodically overgrown by alder and/or willow. The forest clearings were covered by herbaceous vegetation represented by the Rosaceae, Ranunculaceae, Lamiaceae, Aristolochiaceae, Rubiaceae families as well as mosses, ferns, and clubmoss. The climate was temperate.

The fact that the soil cover recorded in a large area of the Dniester River catchment appeared to follow the same evolutionary model, and displayed a similar final image, could be attributed to uniform and constant vegetation and climate conditions (cf. Litt, 2006).

7.2. Features of younger loess-soil sequences (upper Middle–Late Pleistocene)

An important feature of the loess cover in the near-Carpathian part of the upper Dniester River catchment is the rather poor development of loess units younger than MIS 11. Their thickness relatively rarely exceeds a dozen or so metres (Łanczont et al., 2004). In this region, it is not unusual to see superimposed soils of different ages, with loess material that has been completely transformed by pedogenesis. The material for this loess originated from both local older rocks and glacial deposits (Racinowski et al., 2003); in addition, the accumulation of loess was spatially diversified, and loess layers of the same age but of significantly different thickness occur in close proximity. According to variations in the thickness of individual loess layers, that the most favourable conditions for loess accumulation occurred on the west-facing slopes of Mt. Radych (Slokhyni site). An important feature of the studied loess cover is the lack of primary loess L4. This layer was entirely transformed by the post-depositional pedogenic processes that formed soil S3. The activity of the synsedimentary redeposition process, also unfavourable to the preservation of thick loess layers, is evidenced by the predominance of the slope/solifluction facies of loess, which indicates that the periglacial climate was more humid in particular glacial periods, with the exception of MIS 6 (period when loess L2 was accumulated). In the whole Carpathian Foreland, loess L2 is characterized by a considerable thickness that unambiguously indicates a more continental climate (Bogucki (Boguckyj) and Łanczont, 2002). This loess is divided by an intra-loess interstadial Ternopil soil (Slokhyni). This gley soil, strongly deformed by solifluction, developed probably in a wetter, and maybe only slightly warmer climate. It should also be stressed that loess deposits (L1) from the last glaciation are thin and occur only in some places.

Periods of interglacial pedogenesis younger than MIS 11 resulted in the development of soils similar to soil S4. Their complete soil profiles have also been rarely preserved. This indicates that interglacial palaeosols S3, S2 and S1, respectively from MIS 9, 7 and 5, developed in the close foreland of the Carpathians under rather similar environmental conditions. These are mature luvisols, usually gleyed from the top, and cut by a dense network of fissures filled with gley material from the Et horizon. Palaeosols S2 (Korshiv, Kaydaky) and S1 (Horokhiv, Pryluki) are divided by loess cover L2 and correlated with MIS 7 and MIS 5. Lithological, palaeopedological, palynological, and archeological investigations have identified similar sequences in loess sections in the Eastern Carpathian Foreland, Volhynia-Podolia Uplands, Black Sea Lowlands and the Middle Dnieper Area, among others (e.g. Bogucki, 1987; Gozhik (Gożik) et al., 1995; Vandenberghe, 1995; Bogucki and Łanczont, 2011; Shelkoplyas and Khristophorova, 2005; Fedorowicz et al., 2013; Łanczont et al., 2014). Hence, there are no grounds for including them in the same group as the Pryluki-Kaydaki palaeosols or correlating them with MIS 5, as proposed by Buggle et al. (2009).

8. Conclusions

The studied sections of Quaternary deposits situated in the Dniester River basin in the Eastern Carpathian Foreland are key importance to the history of Pleistocene glaciations in Central Europe. The continuity of the litho- and pedostratigraphic sequence (without hiatuses) confirms that the sections are located in the marginal zone of the Elsterian II Glaciation, which is correlated with MIS 12 and Sanian 2 Glaciation according to regional glacial stratigraphy. In this area, the ice-sheet reached its maximum extent on the continental scale. The study reconstructs the main stages and style of the ice-sheet advance and main stages of loess sedimentation. Firstly, the ice sheet encroached on the region as a lobe spreading in the V terrace level during the Sambor Phase. Later, its advance slowed in the Stryvihor–Dniester interfluve, and ephemeral terminoglacial lakes formed in the depressions between the ice front and local morphological barriers. During this period, no Arctic desert was present near the ice-sheet front or adjacent terminoglacial lake; rather, the area was characterized by tundra and/or steppe-tundra vegetation with small tree groups. The ice-sheet continued to advance from N to S, which is now recorded as a thin layer of basal till overlying the glaciolacustrine deposits. Finally, during its retreat (the Krukenychi phase), the ice sheet stagnated on the northern side of the present-day European watershed.

The main palaeogeographic conclusions related to the studied loess-soil sequences are as follows: (1) only periglacial loess deposits (L5–L1), formed from local transport-weathered Carpathian flysch rocks, were used for the loess L5 of the anaglacial phase of the glaciation, while loess originating from local, older rocks and glacial deposits were used for younger deposits (over glacigenic deposits); (2) two sections (Dubrivka and Slokhyni) contain a full sequence of superimposed interglacial soils (S4–S0); (3) the similarities in the genetic type of interglacial soils that developed on loess in the close foreland of the Carpathians and the nature of their post-pedogenic transformations indicate the strong influence exerted by the local climate and vegetation conditions, which were similar in the successive interglacial periods younger than the glaciation of this area.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at https://doi.org/10.1016/j.palaeo.2019.05.024. These data include the Google maps of the most important areas described in this article.

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