New micromorphological knowledge of the last Pleistocene glacial cycle in the loess profile at Praha-Sedlec

Nové mikromorfologické poznatky posledního pleistocenního glaciálního cyklu ve sprašovém profilu Praha-Sedlec (Czech summary)

(2 text-figs., 2 plates)

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The present contribution gives the micromorphological characteristics of the upper-Pleistocene profile at Praha-Sedlec with pedocomplexes PK II and PK III (climatic cycle B). In the subcycle B1 there are preserved remains of soil with an argillic horizon. In the upward direction, the upper subhorizons of this argillic soil are not preserved. The superposed colluvial material is the substrate for further development of the chernozem soil. The most conspicuous feature of this interglacial-glacial sequence is a thin layer of aeolian dust (the marker) contaminated by pedorelicts. Following the deposition of the marker, the surface on a large scale was stripped of vegetation cover and immediately overlapped by pellet sands-sediments of torrential rains as well as by re-sedimentation of loess and sedimentation of early glacial loess.
In the subcycle B2, a duplex humic horizon was identified. Overlapping of this soil by the new marker, repeated re-sedimentation of the subjacent soil took place. A thick loess of the last glacial was deposited over the pellet sands.

Introduction

In the frame of the International Loess Programme carried out under the supervision of G. J. Kukla of Columbia University, U.S.A., an international team of geologists and geophysicists has been formed which, in this country, works on the project: "Paleoclimatic record of oscillations and changes of the environment".

The present paper deals with the last glacial cycle B represented by pedocomplexes PK III and PK II, PK I.

The results of paleomagnetic measurements as well as a further study of snail shells using the aminoacid racemization technique oriented towards stratigraphic segmentation of the studied profile will later be published. The loess sequences in Europe and China closely parallel the oxygen isotope stratigraphy of the deep-sea sediments. Stratigraphy of the last glacial cycle in the well-developed loess sequences in former Czechoslovakia and in the deep-sea sediments is illustrated in Fig. 1 (Kukla 1975, 1987).

This paper gives the micromorphology of only the Sedlec profile type with pedocomplexes PK II and PK III. The micromorphological study is particularly suitable for making distinctions between features resulting from the sedimentation and those caused by soil formation processes. This fact enables us to specify the paleopedologic development of loess series.

History of the investigation

The remains of the Sedlec brickyards indicate the earlier intensive exploitation of loesses. In these loam pits, there still exist only two loess profiles with pedocomplexes (PK I, PK II + III, PK IV) and (PK II and PK III). The PK I in the second profile was previously exposed and it is not presently accessible. Loesses telescopically copy the surface of 5 terrace benches of the Vltava river, running NW-SE (Kukla 1961a, 1975, Ložek 1973, 1985). According to Záruba (1942), the fluvial deposits correspond to the bottom and middle terrace benches (IIIc to IIa), as far as their elevation is concerned.

In the past, there have been proved four complete cycles with fossil pedocomplexes having 3 layers of artifacts in the brickyards. These cycles were studied from the point of view of lithology, stratigraphy and paleontology by Záruba-Pfeffermann (1943), Prošek (1947), Prošek - Ložek (1952, 1954, 1957), Žebera (1949), Ložek (1955, 1956a, b, 1960, 1964, 1973, 1985) and Kukla (1961c). Significant findings at this locality were reported by Kukla (1961a, b, 1969), Kukla - Ložek (1961), Kukla - Ložek - Záruba (1961).

Petrographic and mineralogical study dealing with one of the loess sections along the north-south direction was done by Cimbálniková (1961). Judging from the description of individual layers and the footwall terrace (IIIa) itself, the autor is likely to have investigated the loess profile with PK IV.

Works of Smolíková (1964, 1967, 1990) also dealt with the typological relevance and determination of the sequence of individual phases of polygenetic development of pedocomplexes (PK II, PK II to III and PK III) of the Sedlec profile.

The interest in exposures of the Sedlec loess sections was great and a great number of older studies mentioned in the paper of Smolíková (1967), as well as the works devoted to paleobotanical research (e.g. Nečesaný 1953), and other authors argue for this fact.

In the subcycle B2 with the pedocomplex II, Kukla - Ložek - Záruba (1961) distinguished 3 humic soils in autochthonous or in para-autochthonous position in the Sedlec profile. The upper soil is covered with a soil sediment - the pellet sands - which was deposited by the hillwash after formation of the steppe soil. A conspicuously developed A horizon of the basal soil penetrates a wedge-shaped form into the footwall B horizon of the brown forest carbonate free soil. According to these authors, it is impossible to find out at the Sedlec locality whether we deal with two humic horizons of one fossil soil, or whether there are preserved two soils lying over each other. The pellet sands are separated from the underlying humic autochthonous horizon by a small loess layer which Kukla (1961b) calls "the marker" (Pl. I/1).

In the Sedlec profile by Ložek (1964), the PK II includes thick pellet sands, a typical marker and a duplex chernozem. Similarly, also Kukla (1969) includes solifluction horizon, humic pellet sands - soil sediment, marker and mottled chernozem with B horizon in this pedocomplex.

The sediments of the subcycle B1, occurring between PK II and PK III, consist of layered pellet sands

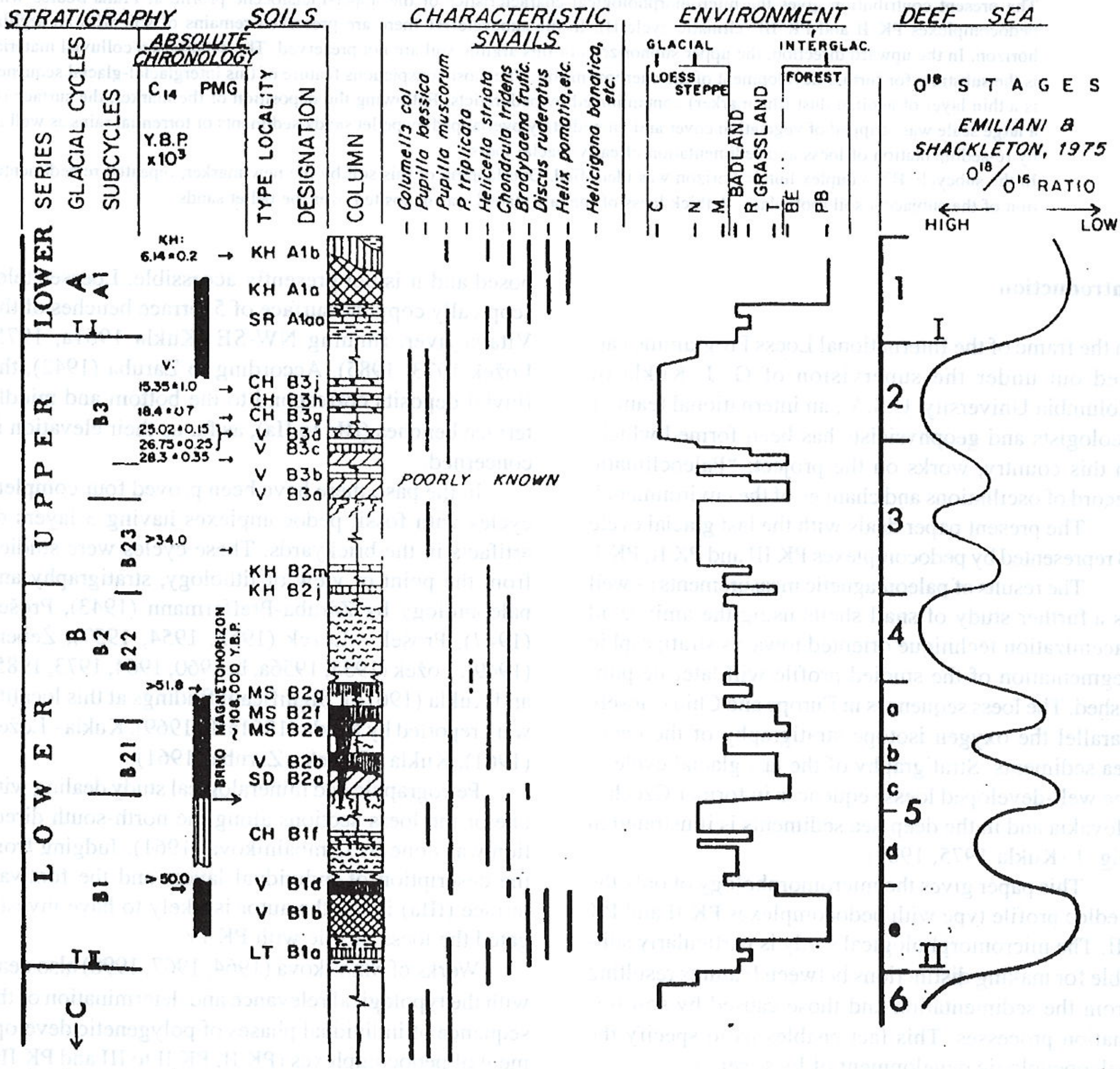

Fig. 1. Stratigraphy of the last glacial cycle in well developed loess sequence in Czechoslovakia by G. J. Kukla (1987). PMG column shows the normal polarity (black), interpreted as reversed, (blank). KH-LT abbreviations of the type localities. Lithologic column shows parabraunerde (crosshatched), braunerde (hatched), biogenic humus enriched soils (vertical hatching), pseudogleys (crosses), hillwash sediments (dashed or waves), loess (blank). The warm ^{18}O stage 5, which was subdivided into five substages, includes the last interglacial (5e) and early glacial (5a-5d).

Curve of reconstructed environment shows cold (C), normal (N) and mild (M) loess steppe; badlands characterized by hillwash deposits; sparse or ephemeral grasslands favoring rendzinas (R); rich or persistent grasslands favoring chernozems (T); coniferous or short/lived mixed forests favoring braunerde (BE); and deciduous, long lasting forests favoring parabraunerde (PB). Cold and normal loess steppe in studied area is characteristic of full glacial (GLAC), parabraunerde for full interglacial (INTGL)

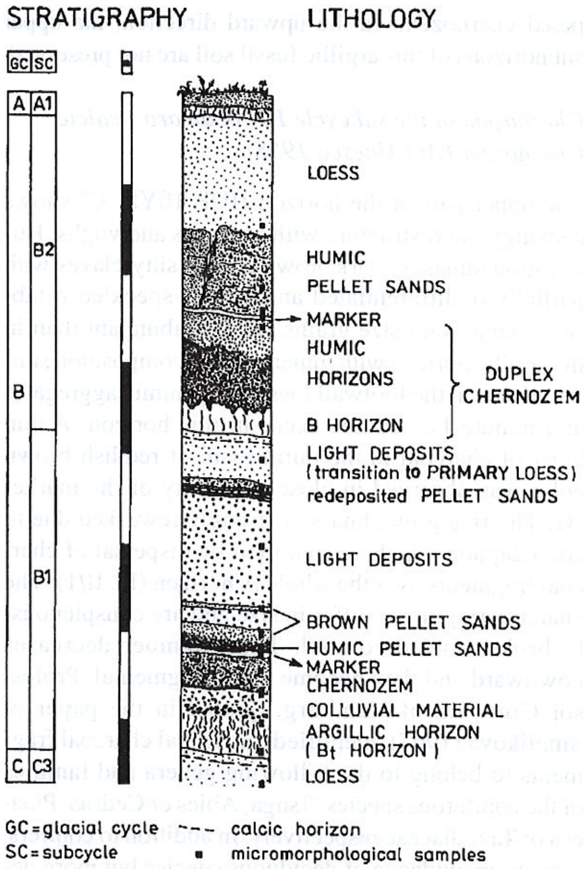

Fig. 2. Loess profile at Praha-Sedlec

with a conspicuous portion of humic and pararabraunerde materials and of superposed light loess. Kukla - Ložek - Záruba (1961) consider the formation of the basal humic material to have taken place immediately after the deposition of the pedocomplex III, when the "autochthonous rendzina" was already formed and whose development was terminated by deposition of hillwash loams.

The pedocomlex III of the subcycle B1 is composed of two independent soils, i. e. the soil of humic character and the parabraunerde (illimerized soil). The latter is underlain by a carbonate horizon (Kukla - Ložek - Záruba 1964).

Micromorphological observations of Smolíková (1967) indicate that the PK II is, for the given profile, formed by two independent soils of which the upper soil typologically corresponds to chernozem and the bottom one to degraded chernozem up to brown forest soil.

The less developed soil between the PK II and the PK III can be marked as "arctic" pararendzina. The basal complex of the cycle B is also formed by two individually developed soils, namely by an atypical chernozem and illimerized soil (parabraunerde) with the so-called "Braunlehm Teilplasma" which was floculated for its major part (Smolíková 1967).

Methods of investigation

In 1992, oriented and loose samples from selected horizons of the loess profile were collected for micromorphological studies and grain - size analysis. The samples were impregnated with polyester resins under vacuum using the method of Čurlík (1971). The samples ware described in accord with the terminology used by Bullock et al. (1985). Moreover, a grain-size distribution of individual layers of the section (Table 1) was done. The layers are described from the top of the cycle C (Fig. 2).

New micromorphological findings

Loess of the subcycle C3; sample 1; HUE 2.5Y 6/4

The loess microstructure is almost massive with transitions to platy and preserved layering. The original channel microstructure is obscured by biochannels which are filled with micritic to microsparitic calcite. Some calcite coating can be observed on transversal sections of the voids. Vertical fractures are free of carbonates. Groundmass is fine-grained, composed of carbonate, clayey-silty material showing crystallitic b-fabric with dispersed micritic carbonates. The coarse fraction consists of unweathered minerals and rock fragments: quartz, K-feldspar, plagioclase, biotite, muscovite, amphibole, pyroxene, re-deposited glauconite,

Table 1
Grain-size distribution of some layers of the upper Pleistocene loess profile at Praha-Sedlec near Prague

Particle size (mm %)	<0.01	≥0.01<0.05	≥0.05<0.125	≥0.125<1	≥1
spl. 01	15.12	52.30	29.08	2.50	1.00
spl. 02	27.93	24.87	21.50	23.30	2.40
spl. 03	29.08	37.12	23.80	9.21	0.79
spl. 04	54.53	38.88	6.34	0.24	0.01
spl. 05	37.03	44.67	15.10	2.80	0.40
spl. 06	44.15	36.29	17.06	1.80	0.70
spl. 07	42.47	35.91	17.07	4.05	0.50
spl. 08	35.18	43.83	17.69	3.20	0.10
spl. 09	31.48	49.47	15.45	3.10	0.50
spl. 10	36.26	42.06	18.78	2.50	0.40
spl. 11	39.65	40.71	17.19	2.10	0.35
spl. 12	40.21	37.32	19.27	2.55	0.65
spl. 13	50.07	44.23	4.60	0.90	0.20
spl. 14	46.40	36.90	6.01	10.40	0.29
spl. 15	41.17	35.04	16.40	6.05	1.35
spl. 16	33.12	49.88	13.40	2.75	0.85

garnet, fibrous sillimanite often with quartz- sillimanite gneiss, chert, quartzite, siltstone and fragments of brown iron nodules (nucleic-nodules) with heterogenous small nuclei.

Enrichment of voids with descending carbonate rich fluids led to the formation of calcareous tubes. Calcareous enrichment (re-precipitation) is connected with processes of degradation due to decarbonatization resulting from heavy precipitation, which is supported by an occurrence of tiny Fe-Mn intercalations in the voids and in the matrix. The overlying carbonate (calcic) horizon provides an evidence of carbonate leaching in the upper horizons.

Argillic horizon of the subcycle B1

This soil was studied in two horizons: the basal horizon HUE 10YR 6/6 and the "upper" horizon HUE 7.5YR 5/6 (sample 2) showing a different grain-size distribution ("upper horizon" is likely to represent the central part of the soil section whose upper horizons were denuded).

"The upper" horizon exhibits a subangular blocky microstructure with smooth planes, vughs and small channels and consists of the finest, yellow-brown silty-clay up to orange clayey material. Domains of the translocated clay (the iron-clay fine material forms a stipple-speckled b-fabric and a granostriated b-fabric) can be observed.

The microstructure of the basal part preserves the structure of the parent material. In this part of the soil horizon, there are relict iron-clay coatings of cryptoanisotropic character. Cryptoanisotropy is due to a higher content of iron. Fragments of oriented clay (original coatings or infillings) in the matrix argue for possible colluviation of the material. Locally, there occur tiny brown Fe nodules and Mn coatings. In the voids, there can be also seen laminated Mn infillings.

"The upper" horizon of the original soil horizon is characteristic of more abundant coarser sandy material (see Table 1; sample 2) of various roundness. Coarse grains are represented by fragments of quartz, feldspar, metamorphic rocks, claystones and siltstones. Feldspars, biotite and other clastic grains show signs of various degree of weathering. The base of the soil profile is epigenetically re-carbonatized. Oxidation conditions under which the soil originated are supported by the presence of tiny Fe nodules.

The colluviated horizon HUE 10YR 5/4 above the B_t horizon shows signs of a brown soil which was epigenetically re-carbonatized. Mn segregations indicate oxidation conditions. As follows from the above mentioned information, there exists a lithological horizon with a different grain-size composition with markedly abundant rock fragments and a conspicuous clay translocation.

Judging from the above mentioned facts, a lithological discontinuity is believed to occur under the superposed chernozem. In the upward direction, the upper subhorizons of this argillic fossil soil are not preserved.

Chernozem of the subcycle B1; (Calcaro - calcic Chernozem FAO-Unesco 1988)

The upper part of the horizon HUE 10YR 3/2 shows a spongy microstructure with channels and vughs. Humic groundmass of dark brown color is silty-clayey with partially undifferentiated and stipple-speckled b-fabric. Coarse sand-size grains are less abundant than in the argillic horizon with mineralogical composition similar of that in the footwall loess. Tiny humic aggregates and minute Fe-nodules occur in this horizon. A thin layer of charcoal in the burnt loam of reddish brown color was observed in close proximity of the marker B1. The fine groundmass is strongly reworked due to zoo-edaphon which is supported by dispersal of charcoal fragments over the whole A horizon (Pl. II/1). The charcoal fragments differ in size and are conspicuously broken and deformed. Their number decreases downward and they become more fragmental. Professor Gottwald of Hamburg, quoted in the paper of Smolíková (1967), identified individual charcoal fragments to belong to the following genera and families of the coniferous species: Tsuga, Abies or Cedrus: Pinacea or Taxodiaceae respectively. In addition to conifers, there is an evidence of deciduous species but more detailed specification is missing. These species are rather exotic and atypical for a steppe cover.

Some voids are filled with humic-carbonate coatings (micritic form) and aggregates of lenticular gypsum. The content of microsparitic calcite in voids and in the matrix increases towards the central part of the horizon HUE 10YR 3/3 (sample 3). Mn intercalations and coatings in the matrix and in voids in the central part and at the base of the soil horizon give an evidence of oxidizing conditions during the soil formation.

Marker of the subcycle B1; sample 4; HUE 10YR 4/3

Interruption of the pedogenesis occurred during sedimentation of the marker (6-8 cm in size) on the chernozem soil (Pl. II/2). The microstructure is mostly massive, but occassionally also channel-like or granular. The groundmass is very fine (Table 1), silty-clayey and enriched with humic material. Minute amount of coarse fraction contains quartz, feldspars, micas, fragments of lydite and sillimanite gneisses, fragments of charcoal and small Fe nodules.

Traces of the parallel layering are due to tiny humic aggregates alternating with silty-clayey loess material without humic admixture. The layering becomes more distinct towards the marker´s surface. The channels and the neighbouring matrix are impregnated by black Mn oxides. The voids may contain microsparitic rhombohedrons of calcite.

Micromorphological features indicate that this horizon is of polygenetic character. The marker is generally finer-grained than regular loess (see Table 1) with nearly no difference in petrographic composition. The layering is likely to reflect the seasonal alternation of dust storms with the hillwashed slope material of re-deposited steppe soils.

Humic pellet sands of the subcycle B1; sample 5; HUE 10YR 3/3
Brown pellet sands of the subcycle B1; sample 6; HUE 10YR 5/4

The humic pellet sands are separated from the lower marker B1 by a sharp boundary. They transit into brown pellet sands through an inverse grading. Immediately above the marker, there is a dark humic layer 2.5 cm in size, with a fine humic clayey-silty groundmass exhibiting a massive microstructure with very fine intra-aggregate voids. Relict charcoal and Mn intercalations occur in the matrix. The mixture of humic-clayey micro-aggregates and clay fragments with signs of clay migration argue for the existence of colluviation processes. This small layer was thought to be the existence of so-called "autochthonous rendzina" (see above). However, micromorphological observations indicate that the thin layer represents very fine humic pellet sands exposed to subsequent weak pedogenesis.

The size of individual aggregates and their abundance increase toward the surface, being less than 1 mm large. The abundance of clayey fragments also increases. The coarse fraction is qualitatively identical with fractions in the footwall soils. It is enriched in re-sedimented fragments of charcoal and brown Fe nodules. The channel voids and fissures are filled with epigenetic carbonates.

The pellet sands exhibiting granular to crumb microstructure have often intergranular space filled with fine material. Relict clay coatings were found to occur in fine silty-clayey groundmass of brown colour. The brown pellet sands are also pedogenetically influenced. They show strong re-carbonatization, matrix affected by epigenetic carbonates and Mn intercalations of dendritic forms.

V. Ložek provides examples of interglacial fauna occurring in the pellet sands.

Light deposits [re-sedimented loess C3] of the subcycle B1; sample 7; HUE 10YR 6/4

The light yellow- brown loess deposits as much as 200 cm thick are intercalated with brown to dark brown humic layers of various thickness. They have a channel microstructure. The grain-size composition is different from that of the typical loess (Table 1). Subangular to round microaggregates close to each other, separated by packing voids occur within one aggregate. Single microaggregates consist of fine silty-clayey, carbonate, humic and clayey matrix with the signs of clay translocation, and with Fe nodules. Domains of micritic to microsparitic calcite crystals were observed in the fine mass. Fabric of the fine material has crystallitic and speckled b-fabric. Microsparitic and also acicular calcite fill the chanels. Coarse fraction is identical with that of the footwall pellet sands. The re-calcification appears to be strong. Mn intercalations are also occasionally present.

A small layer 6 cm in size occurs close to the base of these light sediments. It consists of brown pellet sands HUE 10YR 5/6 with relict clay coatings and an admixture of coarse fraction of shales. Across the slope, there are spread greyish brown pellet sands HUE 10YR 5/4 with fine clayey-silty carbonate groundmass showing preserved fragments of mollusca shells (Pl. II/3), large charcoal relicts and brown Fe nodules. They occur about 130 cm above the pellet sands. These 5 cm thick pellet sands acquire towards the top wall a crumb structure with numerous vughs and channels with a chaotic distribution of aggregates 1-1.5 mm in size. The aggregates have humic and clayey matrix with marked orange laminated coatings.Cemented aggregates of both matrices are also present.

Some aggregates are broken due to transport. The pellet sands are believed to be re-deposited and later exposed to strong re-carbonatization.

The upper 8 cm thick pellet sands of the sample 8, HUE 10YR 4/3 are rather rich in humic aggregates exhibiting strong re-carbonatization. The matrix contains gypsum in addition to carbonates in voids. An organic shell of Cretaceous age filled with Fe pigment was found in the fine matrix. These dark pellet sands are conspicuously subdivided into numerous thin horizons down the slope, and they were believed in previous works (see above) to be the "arctic" pararendzina (Pl. I/2).

The grain-size distribution of the superposed loess deposits to the primary loess of the sample 9, HUE 10YR 6/4 indicates a silty admixture (Table 1). The microstructure and fabric of the groundmass are the same as those of the above described deposits. In the coarse fraction, there occur shales with Fe pigment, marls, Fe nodules and fragments of micritic carbonates. Individual grains are partly weathered.

This sediment is strongly impregnated by secondary carbonates and also by lenticular gypsum. The voids and the matrix are either copletely filled with acicular (so-called lublinite) to micritic carbonates which sometimes form micritic nodules, coatings, hypocoatings and often even microsparitic forms of calcite. Leaching processes indicate an existence of superposed carbonate horizon.

Degraded chernozem of the subcycle B2; (Luvi- haplic Chernozem FAO-Unesco 1988)

The humic horizon is more than 100 cm thick. It can be divided into several subhorizons. The basal subhorizon HUE 7.5YR 3/3 (sample 11) shows subangular blocky microstructure with transition to a channel microstructure. Subangular aggregates are separated by packing voids. The fine groundmass is clayey, clayey-silty with a stipple- speckled b-fabric. This brown-black groundmass preserves the features of clay transloca-tion having orange to orange-brown coatings which copy the surface of microaggregates. The clayey groundmass is differentiated. Clayey domains exhibit a parallel striated b-fabric. The clayey material of the voids and the matrix are microlaminated and mostly cracked. Some voids are filled with colluvial material. The coarse fraction is worn. Besides typical Fe nodules, there occur black disjointed Mn nodules. The horizon with darker color of the sample 12, HUE 10YR 3/2 contains greater portion of dispersed humus showing spongy and markedly granular microstructures. The fine humic silty-clayey groundmass has an undifferentiaded and partly stipple-speckled b-fabric. The fragments of clay coatings are also present in the matrix. Fragments of charcoal, Fe nodules and spherical excrements of mesoedaphon are rare in the groundmass. The upper horizon (HUE 10YR 4/3) was by faunaturbation partly transfered in the superposed marker B2 which ceased to have a sharp contact boundary. This horizon has spongy microstructure. The fine humic mass has a considerable admixture of silt and does not contain fragments of charcoal. B-fabric is undifferentiated and has no carbonates.

The brown forest soil HUE 10YR 5/4 (sample 10) exhibiting channel microstructure and brown clayey-silty groundmass belongs to the B2 cycle. This fine groundmass has a mosaic-speckled b-fabric and grano-poro-striated b-fabrics. The channels and vughs have mostly smooth walls; only larger voids are coated with dendritic Mn segregation which frequently also occur in the neighbouring matrix. Larger voids are filled with acicular calcite. The coarse fraction is identical with the underlying light material. The soil is characteristic of a weak re-carbonatization.

The B horizon belongs to the overlying degraded chernozem.

Marker of the subcycle B2; sample 13; HUE 10YR 5/4

It is represented by a very fine material (see Table 1) with nearly massive microstructure having abundant horizontal pores. The fine silty-clayey groundmass with admixture of humic material has undifferentiated b-fabric. The coarse fraction is less abundant and consists of rimous quartz, muscovite, dark biotite, feldspars, amphibole, pyroxene, siltstone, shales, Fe nodules and tiny particles of charcoal.

The marker contains humic microaggregates with nearly no carbonates. The carbonates occur in the upper pellet sands and in the lower soil in the form of infillings of offset fractures 1 mm in size.

Humic pellet sands of the subcycle B2; HUE 10YR 3/3

The bottom of the sample 14 as well as the central part of the sample 15 exhibit fine crumb to subangular blocky microstructure. Table 1 indicates the influx of the coarse fraction which is qualitatively identical with the underlying chernozem. Some microaggregates represent weathered pedorelicts with clayey matrix. Single aggregates are angular, subangular to rounded indicating a chaotic deposition and exhibiting various grain size (Pl. II/4). The humic silty-clayey groundmass with various intensity of coloring contains charcoal and relict coatings or infillings of the clay translocation from the underlying chernozem. The admixture of sand grains between the aggregates as well as the amount of relict coatings increase upward. Fe nodules and re-deposited fragments of Mn pedofeatures are present in the humic matrix. The voids are filled with sparitic calcite.

Loess of the subcycle B2; sample 16; HUE 10YR 6/4

The pellet sands are covered with a thin solifluction tiny layer and by loess having channel structure. The carbonate clayey-silty groundmass exhibits a crystallitic b-fabric. It consists of aggregates with an independent humic mass and also of clayey groundmass containing relict domains of oriented clay. The coarse fraction is abundant and contains quartz, K-feldspar, plagioclase, microcline, muscovite, biotite, pyroxene, amphibole, sillimanite, lydite, siltstone, Fe nodules and also relict clay textural pedofeatures. The carbonatization represented by diffuse micritic carbonate mass and calcite rhombohedrons in matrix and in voids is characteristic of this material.

Discussion

Micromorphological characteristics of the Sedlec profile indicates that the fossil soil of which probably only a "torso" with an argillic horizon has been preserved, belongs to the PK III of the subcycle B1. An accumulation of the coarse-grained material had taken place prior to the clay translocation. The horizon B_t appears to be product of a mixed deciduous forest. In the upward direction, the upper subhorizons of B_t horizon were perhaps not preserved.

The colluvial material from these upper subhorizons was probably a substrate for further soil development.

The superposed humic soil has developed under the steppe up to wooded steppe cover. The soil was partly affected by increased moisture content and particularly by strong bioturbation. A tiny layer of charcoal pieces with red-brown rim occurs in the surfacial part of the humic horizon. This tiny layer represents a remnant of a taiga fire. The soil next to it is covered by the marker (Pl. II/3). The aeolian sediment is con-

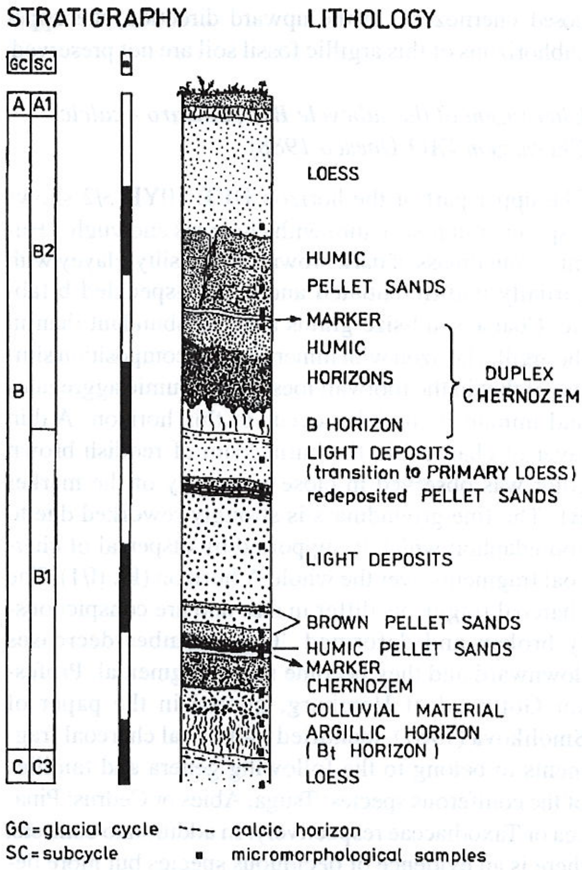

Fig. 2. Loess profile at Praha-Sedlec

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spl. 15	41.17	35.04	16.40	6.05	1.35
spl. 16	33.12	49.88	13.40	2.75	0.85

garnet, fibrous sillimanite often with quartz- sillimanite gneiss, chert, quartzite, siltstone and fragments of brown iron nodules (nucleic-nodules) with heterogenous small nuclei.

Enrichment of voids with descending carbonate rich fluids led to the formation of calcareous tubes. Calcareous enrichment (re-precipitation) is connected with processes of degradation due to decarbonatization resulting from heavy precipitation, which is supported by an occurrence of tiny Fe-Mn intercalations in the voids and in the matrix. The overlying carbonate (calcic) horizon provides an evidence of carbonate leaching in the upper horizons.

Argillic horizon of the subcycle B1

This soil was studied in two horizons: the basal horizon HUE 10YR 6/6 and the "upper" horizon HUE 7.5YR 5/6 (sample 2) showing a different grain-size distribution ("upper horizon" is likely to represent the central part of the soil section whose upper horizons were denuded).

"The upper" horizon exhibits a subangular blocky microstructure with smooth planes, vughs and small channels and consists of the finest, yellow-brown silty-clay up to orange clayey material. Domains of the translocated clay (the iron-clay fine material forms a stipple-speckled b-fabric and a granostriated b-fabric) can be observed.

The microstructure of the basal part preserves the structure of the parent material. In this part of the soil horizon, there are relict iron-clay coatings of cryptoanisotropic character. Cryptoanisotropy is due to a higher content of iron. Fragments of oriented clay (original coatings or infillings) in the matrix argue for possible colluviation of the material. Locally, there occur tiny brown Fe nodules and Mn coatings. In the voids, there can be also seen laminated Mn infillings.

"The upper" horizon of the original soil horizon is characteristic of more abundant coarser sandy material (see Table 1; sample 2) of various roundness. Coarse grains are represented by fragments of quartz, feldspar, metamorphic rocks, claystones and siltstones. Feldspars, biotite and other clastic grains show signs of various degree of weathering. The base of the soil profile is epigenetically re-carbonatized. Oxidation conditions under which the soil originated are supported by the presence of tiny Fe nodules.

The colluviated horizon HUE 10YR 5/4 above the B_t horizon shows signs of a brown soil which was epigenetically re-carbonatized. Mn segregations indicate oxidation conditions. As follows from the above mentioned information, there exists a lithological horizon with a different grain-size composition with markedly abundant rock fragments and a conspicuous clay translocation.

Judging from the above mentioned facts, a lithological discontinuity is believed to occur under the superposed chernozem. In the upward direction, the upper subhorizons of this argillic fossil soil are not preserved.

Chernozem of the subcycle B1; (Calcaro - calcic Chernozem FAO-Unesco 1988)

The upper part of the horizon HUE 10YR 3/2 shows a spongy microstructure with channels and vughs. Humic groundmass of dark brown color is silty-clayey with partially undifferentiated and stipple-speckled b-fabric. Coarse sand-size grains are less abundant than in the argillic horizon with mineralogical composition similar of that in the footwall loess. Tiny humic aggregates and minute Fe-nodules occur in this horizon. A thin layer of charcoal in the burnt loam of reddish brown color was observed in close proximity of the marker B1. The fine groundmass is strongly reworked due to zoo-edaphon which is supported by dispersal of charcoal fragments over the whole A horizon (Pl. II/1). The charcoal fragments differ in size and are conspicuously broken and deformed. Their number decreases downward and they become more fragmental. Professor Gottwald of Hamburg, quoted in the paper of Smolíková (1967), identified individual charcoal fragments to belong to the following genera and families of the coniferous species: Tsuga, Abies or Cedrus: Pinacea or Taxodiaceae respectively. In addition to conifers, there is an evidence of deciduous species but more detailed specification is missing. These species are rather exotic and atypical for a steppe cover.

Some voids are filled with humic-carbonate coatings (micritic form) and aggregates of lenticular gypsum. The content of microsparitic calcite in voids and in the matrix increases towards the central part of the horizon HUE 10YR 3/3 (sample 3). Mn intercalations and coatings in the matrix and in voids in the central part and at the base of the soil horizon give an evidence of oxidizing conditions during the soil formation.

Marker of the subcycle B1; sample 4; HUE 10YR 4/3

Interruption of the pedogenesis occurred during sedimentation of the marker (6-8 cm in size) on the chernozem soil (Pl. II/2). The microstructure is mostly massive, but occassionally also channel-like or granular. The groundmass is very fine (Table 1), silty-clayey and enriched with humic material. Minute amount of coarse fraction contains quartz, feldspars, micas, fragments of lydite and sillimanite gneisses, fragments of charcoal and small Fe nodules.

Traces of the parallel layering are due to tiny humic aggregates alternating with silty-clayey loess material without humic admixture. The layering becomes more distinct towards the marker´s surface. The channels and the neighbouring matrix are impregnated by black Mn oxides. The voids may contain microsparitic rhombohedrons of calcite.

Micromorphological features indicate that this horizon is of polygenetic character. The marker is generally finer-grained than regular loess (see Table 1) with nearly no difference in petrographic composition. The layering is likely to reflect the seasonal alternation of dust storms with the hillwashed slope material of re-deposited steppe soils.

Humic pellet sands of the subcycle B1; sample 5; HUE 10YR 3/3
Brown pellet sands of the subcycle B1; sample 6; HUE 10YR 5/4

The humic pellet sands are separated from the lower marker B1 by a sharp boundary. They transit into brown pellet sands through an inverse grading. Immediately above the marker, there is a dark humic layer 2.5 cm in size, with a fine humic clayey-silty groundmass exhibiting a massive microstructure with very fine intra-aggregate voids. Relict charcoal and Mn intercalations occur in the matrix. The mixture of humic-clayey micro-aggregates and clay fragments with signs of clay migration argue for the existence of colluviation processes. This small layer was thought to be the existence of so-called "autochthonous rendzina" (see above). However, micromorphological observations indicate that the thin layer represents very fine humic pellet sands exposed to subsequent weak pedogenesis.

The size of individual aggregates and their abundance increase toward the surface, being less than 1 mm large. The abundance of clayey fragments also increases. The coarse fraction is qualitatively identical with fractions in the footwall soils. It is enriched in re-sedimented fragments of charcoal and brown Fe nodules. The channel voids and fissures are filled with epigenetic carbonates.

The pellet sands exhibiting granular to crumb microstructure have often intergranular space filled with fine material. Relict clay coatings were found to occur in fine silty-clayey groundmass of brown colour. The brown pellet sands are also pedogenetically influenced. They show strong re-carbonatization, matrix affected by epigenetic carbonates and Mn intercalations of dendritic forms.

V. Ložek provides examples of interglacial fauna occurring in the pellet sands.

Light deposits [re-sedimented loess C3] of the subcycle B1; sample 7; HUE 10YR 6/4

The light yellow- brown loess deposits as much as 200 cm thick are intercalated with brown to dark brown humic layers of various thickness. They have a channel microstructure. The grain-size composition is different from that of the typical loess (Table 1). Subangular to round microaggregates close to each other, separated by packing voids occur within one aggregate. Single micro-aggregates consist of fine silty-clayey, carbonate, humic and clayey matrix with the signs of clay translocation, and with Fe nodules. Domains of micritic to microsparitic calcite crystals were observed in the fine mass. Fabric of the fine material has crystallitic and speckled b-fabric. Microsparitic and also acicular calcite fill the chanels. Coarse fraction is identical with that of the footwall pellet sands. The re-calcification appears to be strong. Mn intercalations are also occasionally present.

A small layer 6 cm in size occurs close to the base of these light sediments. It consists of brown pellet sands HUE 10YR 5/6 with relict clay coatings and an admixture of coarse fraction of shales. Across the slope, there are spread greyish brown pellet sands HUE 10YR 5/4 with fine clayey-silty carbonate groundmass showing preserved fragments of mollusca shells (Pl. II/3), large charcoal relicts and brown Fe nodules. They occur about 130 cm above the pellet sands. These 5 cm thick pellet sands acquire towards the top wall a crumb structure with numerous vughs and channels with a chaotic distribution of aggregates 1-1.5 mm in size. The aggregates have humic and clayey matrix with marked orange laminated coatings.Cemented aggregates of both matrices are also present.

Some aggregates are broken due to transport. The pellet sands are believed to be re-deposited and later exposed to strong re-carbonatization.

The upper 8 cm thick pellet sands of the sample 8, HUE 10YR 4/3 are rather rich in humic aggregates exhibiting strong re-carbonatization. The matrix contains gypsum in addition to carbonates in voids. An organic shell of Cretaceous age filled with Fe pigment was found in the fine matrix. These dark pellet sands are conspicuously subdivided into numerous thin horizons down the slope, and they were believed in previous works (see above) to be the "arctic" pararendzina (Pl. I/2).

The grain-size distribution of the superposed loess deposits to the primary loess of the sample 9, HUE 10YR 6/4 indicates a silty admixture (Table 1). The microstructure and fabric of the groundmass are the same as those of the above described deposits. In the coarse fraction, there occur shales with Fe pigment, marls, Fe nodules and fragments of micritic carbonates. Individual grains are partly weathered.

This sediment is strongly impregnated by secondary carbonates and also by lenticular gypsum. The voids and the matrix are either copletely filled with acicular (so-called lublinite) to micritic carbonates which sometimes form micritic nodules, coatings, hypocoatings and often even microsparitic forms of calcite. Leaching processes indicate an existence of superposed carbonate horizon.

Degraded chernozem of the subcycle B2; (Luvi- haplic Chernozem FAO-Unesco 1988)

The humic horizon is more than 100 cm thick. It can be divided into several subhorizons. The basal subhorizon HUE 7.5YR 3/3 (sample 11) shows subangular blocky microstructure with transition to a channel microstructure. Subangular aggregates are separated by packing voids. The fine groundmass is clayey, clayey-silty with a stipple- speckled b-fabric. This brown-black groundmass preserves the features of clay transloca-tion having orange to orange-brown coatings which copy the surface of microaggregates. The clayey groundmass is differentiated. Clayey domains exhibit a parallel striated b-fabric. The clayey material of the voids and the matrix are microlaminated and mostly cracked. Some voids are filled with colluvial material. The coarse fraction is worn. Besides typical Fe nodules, there occur black disjointed Mn nodules. The horizon with darker color of the sample 12, HUE 10YR 3/2 contains greater portion of dispersed humus showing spongy and markedly granular microstructures. The fine humic silty-clayey groundmass has an undifferentiaded and partly stipple-speckled b-fabric. The fragments of clay coatings are also present in the matrix. Fragments of charcoal, Fe nodules and spherical excrements of mesoedaphon are rare in the groundmass. The upper horizon (HUE 10YR 4/3) was by faunaturbation partly transfered in the superposed marker B2 which ceased to have a sharp contact boundary. This horizon has spongy microstructure. The fine humic mass has a considerable admixture of silt and does not contain fragments of charcoal. B-fabric is undifferentiated and has no carbonates.

The brown forest soil HUE 10YR 5/4 (sample 10) exhibiting channel microstructure and brown clayey-silty groundmass belongs to the B2 cycle. This fine groundmass has a mosaic-speckled b-fabric and grano-poro-striated b-fabrics. The channels and vughs have mostly smooth walls; only larger voids are coated with dendritic Mn segregation which frequently also occur in the neighbouring matrix. Larger voids are filled with acicular calcite. The coarse fraction is identical with the underlying light material. The soil is characteristic of a weak re-carbonatization.

The B horizon belongs to the overlying degraded chernozem.

Marker of the subcycle B2; sample 13; HUE 10YR 5/4

It is represented by a very fine material (see Table 1) with nearly massive microstructure having abundant horizontal pores. The fine silty-clayey groundmass with admixture of humic material has undifferentiated b-fab-ric. The coarse fraction is less abundant and consists of rimous quartz, muscovite, dark biotite, feldspars, am-phibole, pyroxene, siltstone, shales, Fe nodules and tiny particles of charcoal.

The marker contains humic microaggregates with nearly no carbonates. The carbonates occur in the up-per pellet sands and in the lower soil in the form of infillings of offset fractures 1 mm in size.

Humic pellet sands of the subcycle B2; HUE 10YR 3/3

The bottom of the sample 14 as well as the central part of the sample 15 exhibit fine crumb to subangular blocky microstructure. Table 1 indicates the influx of the coarse fraction which is qualitatively identical with the underlying chernozem. Some microaggregates re-present weathered pedorelicts with clayey matrix. Sin-gle aggregates are angular, subangular to rounded indicating a chaotic deposition and exhibiting various grain size (Pl. II/4). The humic silty-clayey ground-mass with various intensity of coloring contains char-coal and relict coatings or infillings of the clay translo-cation from the underlying chernozem. The admixture of sand grains between the aggregates as well as the amount of relict coatings increase upward. Fe nodules and re-deposited fragments of Mn pedofeatures are present in the humic matrix. The voids are filled with sparitic calcite.

Loess of the subcycle B2; sample 16; HUE 10YR 6/4

The pellet sands are covered with a thin solifluction tiny layer and by loess having channel structure. The carbonate clayey-silty groundmass exhibits a crystallitic b-fabric. It consists of aggregates with an independent humic mass and also of clayey groundmass containing relict domains of oriented clay. The coarse fraction is abundant and contains quartz, K-feldspar, plagioclase, microcline, muscovite, biotite, pyroxene, amphibole, sillimanite, lydite, siltstone, Fe nodules and also relict clay textural pedofeatures. The carbonatization repre-sented by diffuse micritic carbonate mass and calcite rhombohedrons in matrix and in voids is characteristic of this material.

Discussion

Micromorphological characteristics of the Sedlec pro-file indicates that the fossil soil of which probably only a "torso" with an argillic horizon has been preserved, belongs to the PK III of the subcycle B1. An accumu-lation of the coarse-grained material had taken place prior to the clay translocation. The horizon B_t appears to be product of a mixed deciduous forest. In the up-ward direction, the upper subhorizons of B_t horizon were perhaps not preserved.

The colluvial material from these upper subhori-zons was probably a substrate for further soil development.

The superposed humic soil has developed under the steppe up to wooded steppe cover. The soil was partly affected by increased moisture content and par-ticularly by strong bioturbation. A tiny layer of charcoal pieces with red-brown rim occurs in the surfacial part of the humic horizon. This tiny layer represents a remnant of a taiga fire. The soil next to it is covered by the marker (Pl. II/3). The aeolian sediment is contaminated with the following pedorelicts: humic aggregates, charcoal and Fe nodules from the underlying soil. The layering of the marker toward the top of the section seems to indicate an increase in alternation of eolian (dust storms) and run off sedimentation. It is difficult to establish the original thickness of the marker because it was affected by bioturbation, partly transferred into the steppe soil and partly denuded.

The horizon with rounded aggregates - i.e. soil colluviums - of the pellet sands in PK III is separated from the marker B1 by a sharp boundary. The accumulation of the pellet sands is of inverse character because the material of A horizon was deposited prior to that of the B horizon of the soils lying on slopes. The end-member of the subcycle B1 is represented by re-sedimentation of the loess and pellet sands as well as by deposition of the primary loess.

As for the subcycle B2 with the pedocomplex II, it is very difficult to distinguish its individual phases. The humic soil with 100 to 110 cm thick horizon and the underlying B horizon belong to the degraded chernozem.

Because of the signs of the illimerization process at the base of the humic horizon and the occurrence of charcoal fragments in the middle of the horizon and the influx of the silty material into the upper part of the soil horizon in the depth of 10 to 20 cm, the author is rather in favour of its double appearance. This soil being covered with marker B2, in the PK II containing fine-grained material with pedorelicts with almost missing carbonates and layering in its upper part, indicate that the re-sedimentation of the underlying soil from neighbouring slopes may have taken place. The content of rounded humic microaggregates and fragments of clay coatings, their platy bedding and the inverse deposition of the underlying soil horizon supports the idea that the re-sedimentation resulted in formation of pellet sands. The humic pellet sands are covered by deposits of sheet erosion and by slightly sorted loess. The latter transits continually into 150 cm thick loess complex of the last glacial age. The existence of two independent humic soils distinguished earlier (see above) in the PK II of the Sedlec profile appears to by unlikely.

Bullock and Murphy (1979) consider the presence of rounded fossil aggregates in a paleo-argillic brown earth to be an evidence of certain transport of the material due to either the freeze-thaw (associated with a glaciation) or due to shifting of the active layer above the permafrost. Similarly, also Gerasimov (1971); Morozova (1972) and Fedoroff - Courty (1987) have attributed the origin of rounded aggregates to cryogenesis. Another theory on the origin of these aggregates larger than 1 mm in size occuring in sandy tills was given by Felix-Henningsen in Mücher - Morozova (1983). He attributed the formation of rounded aggregates to their transport by solifluction.

Kukla (1975) believes that the pellet sands are slope deposits of torrential rains, the so-colled storm outwashes. The pellet sands, according Kukla, are composed of entirely or partly oval pellets - aggregates of sandy fraction. These sediments originated from re-deposition of older soils transported by flash currents down the slope without being disintegrated in single aggregates. This displacement was probably due to a bare surface with scarce vegetation on slopes after a long period of dry and warm weather followed by a shorter period of intense torrential rains. The deposition of pellet sands required an existence of grassland on steep slopes. The author of this article agrees with this idea.

The processes of pedogenesis in the loess sequences are believed to have occurred either under arid conditions when the processes of evaporation (e. g. the precipitation of secondary carbonates from groundwater through capillary rise) took place or under humid conditions when the processes of leaching-translocation of substances (e.g. of carbonates and clay) prevailed. Processes in the local loesses were studied by Ambrož (1947); Čurlík (1977, 1985) et al. Brown soils (Luvisols) seem to reflect the influence of humid climate with descending accumulation of secondary carbonates in the C horizon (vertical and lateral). If a reverse water flow through capillaries together with the precipitation of carbonates occur, particularly in an arid environment, then the origin of these soils was due to a regime of evapo-transpiration, to the so-called re-carbonatization. Such phehomena can be observed in the studied profile.

Conclusion

The following conclusions can be drawn based upon a detailed investigation of the loess profile of the last glacial cycle B at Praha-Sedlec:

(a) The subcycle B2 is composed of the duplex humic horizon (100-110 cm thick) covered by marker B2. The marker consists of fine-grained material with humic pedorelicts; a resedimentation of the subjacent soil from neighbourning slopes is likely to have taken place. Humic pellet sand were covered by a thin solifluction and then by a thick loess complex of the last glacial.

(b) The subcycle B1 is built up independent soils; the lower soil - partly eroded fossil soil with argillic horizon; and the upper soil represented by steppe soil-chernozem. In the upward direction, the upper subhorizons of B_t horizon are perhaps not preserved. Consequently, a steppe soil affect by strong bioturbation was developed from a colluvial material. A superficial tiny layer of charcoal pieces represents a relict after a taiga fire. An abrupt climatic change caused a sedimentation of marker B1 after which a great accumulation of pellet sands took place.

(c) Eolian and run-off processes, solifluction and cryoturbation are characteristic of the period following the sedimentation of B1 and B2 markers. These processes were particularly significant in places where the footwall soil was not wooded which led to lateral transport of the material to new sites on slopes or in flood plains.

(d) The distribution of secondary carbonates in the loess profile argues for an ascending and descending (vertical or lateral) migration of substances.

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J. Hradilová: New Micromorphological Knowledge of the Last Pleistocene Glacial Cycle in the Loess Profile at Praha-Sedlec... (Pl. I)

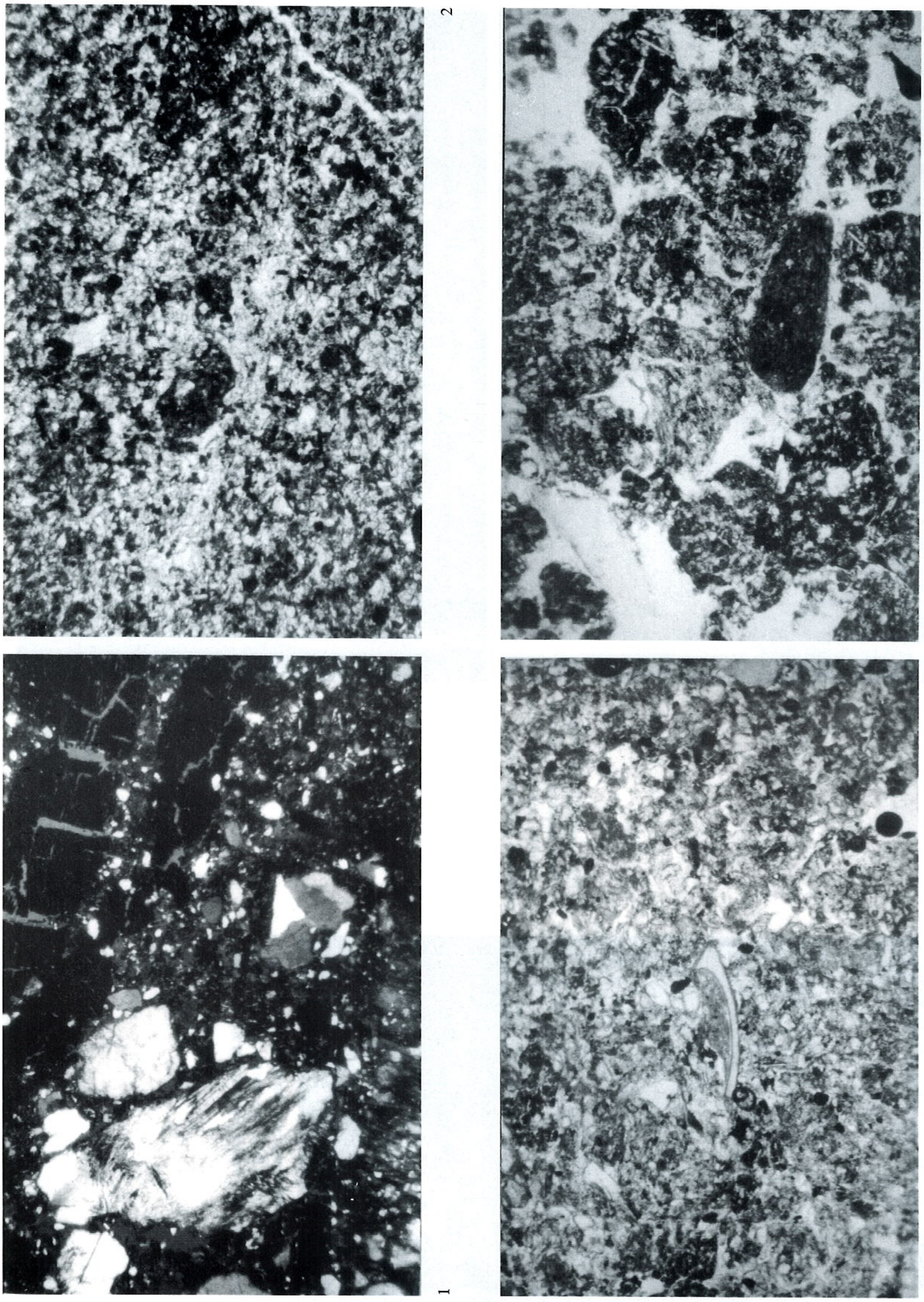
1. The interruption of pedogenesis occurred during the sedimentation of a thin light loess layer (marker B2). The marker typically occurs between the chernozem soil (here duplex chernozem) and pellet sands

2. Pellet sands in light deposits. Dark pellet sands are conspicuously subdivided into numerous thin horizons

3. A thin layer of charcoal occurs near the base of B1 marker and on the top of chernozem

Photo J. Hradilová

J. Hradilová: New Micromorphological Knowledge of the Last Pleistocene Glacial Cycle in the Loess Profile at Praha-Sedlec... (Pl. II)



2 4

1 3

For explanation see p. 329

Nové mikromorfologické poznatky posledního pleistocenního glaciálního cyklu ve sprašovém profilu Praha-Sedlec

V článku je uvedena mikromorfologická charakteristika mladopleistocenního sprašového profilu Praha-Sedlec s půdními komplexy PK II a PK III klimatického cyklu B. V subcyklu B1 se dochoval zbytek půdy s argilitovým horizontem. Na svazích paleoreliéfu se nejsvrchnější horizonty této fosilní půdy nedochovaly. Nadložní koluviální materiál je substrátem pro další vývoj černozemní půdy. Velmi výrazným znakem poslední interglaciál-glaciální sekvence je tenká vrstva eolického prachu (marker), který je kontaminován půdními relikty. Po sedimentaci markru byl povrch v převážné míře zbaven vegetačního krytu a následně překryt hlinopísky - sedimenty přívalových dešťů a také resedimentovanou spraší; dále pak sedimentací iniciální spraše.

V subcyklu B2 byl identifikován zdvojený humózní horizont. Po překrytí této půdy markrem došlo k následné resedimentaci podložní půdy. Na hlinopísky sedimentovala mohutná spraš posledního glaciálu.

Explanation of plate II

1. Fragment of charcoal wood species probably of coniferous species and a rock fragment (quartz, sillimanite) in A horizon, PK III. Magnification 6.3x, transmitted light, crossed nicols. *Microphoto M. Šťastný - J. Hradilová*
2. Marker B1, humic aggregates alternate with fine mass of eolian origin. Magnification 6.3x, transmitted light, nicols II. *Microphoto M. Šťastný - J. Hradilová*
3. Fragment of calcareous shell with a remnant of transported material in pellet sands disseminated in light deposits of B1. Magnification 6.3x, transmitted light, nicols II. *Microphoto J. Hradilová*
4. Humic pellet sands of B2 - typical rounded aggregates. Magnification 6.3x, transmitted light, nicols II. *Microphoto M. Šťastný - J. Hradilová*

ENVIRONMENTAL PROBLEMS - OTÁZKY ŽIVOTNÍHO PROSTŘEDÍ

Global climate change and peatland carbon balance: An overview

Globální změna klimatu a bilance uhlíku v rašeliništích (Czech summary)

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Considerable quantities of carbon have been removed from the atmosphere and stored as peat since the most recent glaciation. The conceptual framework for understanding peat accumulation and carbon balance in peatlands has long been established, yet substantial gaps exist in our understanding of processes related to carbon balance in peatlands and of the source/sink relationships for CH_4 and CO_2 between peatlands and the atmosphere. Few estimates of net primary production for peatland ecosystems exist; belowground production and its contribution to peatland carbon balance have been largely ignored. The extent to which site-specific field measurements of CH_4 and CO_2 emissions from peatlands can be generalized to broader geographic areas remains uncertain. Relatively little work has focused on the microbially mediated mechanisms of CH_4 and CO_2 production/consumption within the peat itself or on the population and community dynamics of the microorganisms responsible for carbon mineralization in peat. Such gaps in our current understanding of present-day carbon cycling in peatlands confound efforts to project how peatland ecosystem function may be altered under predicted scenarios of global climate change.

Introduction

Boreal and subarctic peatlands cover about 360 million ha of the earth's land surface and contain an estimated 455 Pg of carbon (Gorham 1991). The boreal and subarctic regions which are presently occupied by peatland ecosystems were largely covered by ice during the most recent glacial period. Thus, the quantity of carbon stored as peat is a testament to the role of peatlands as a long-term sink for photosynthetically fixed atmospheric CO_2 (Horden et al. 1992). Currently, boreal and subarctic peatlands continue to accumulate carbon at an estimated rate of 0.076 Pg/yr, acting as a net sink for atmospheric CO_2 and a net source of atmospheric CH_4 (Gorham 1991). Ongoing increases in the concentrations of atmospheric greenhouse gases (Rodhe 1990) and predictions that associated climatic warming will be most pronounced in the northern latitudes (Grotch - MacCracken 1991), have focused research on carbon cycling in peatland ecosystems.

Peat accumulates on the landscape when net primary production by the surface vegetation exceeds the sum of aerobic and anaerobic decomposition throughout the peat column. In the following paragraphs, we summarize current information regarding peat accumulation and carbon balance in northern peatland ecosystems, identifying gaps in our present understanding.

Assessing long-term peat accumulation

A common approach to quantifying long-term rates of peat accumulation is to radiocarbon date the basal peat at the deepest locality within a peatland, calculating an overall average rate of vertical peat accumulation over the lifetime of the peat deposit. Using this approach, rates of vertical peat accumulation for various regions of the world range generally from 0.2 to 0.8 mm/yr (Table 1). Notably lower rates have been obtained from sites in the unglaciated regions of Pennsylvania and West Virginia, U.S.A., which are characterized as having older, shallower, relatively highly decomposed peat deposits. It is generally acknowledged that compaction and a longer period of decomposition of the older, i.e., deeper peat dictate that vertical peat accumulation should not proceed in a linear fashion. Therefore, considerable caution must be exercised in estimating

Table 1. Rates of peat accumulation in various regions of the world, as determined by radiocarbon dating of basal peat.

Site/Region	Rate mm/yr
South Sweden and North Germany*	0.70
South and Central Finland*	0.75
Northern Europe*	0.60
Boreal USSR*	0.6-0.8
Eurasia*	0.52
Siberia*	0.2-0.4
Subarctic Canada*	0.31
Boreal Canada*	0.54
Canada Overall*	0.48
Northwest Territories**	0.17-0.76
Pennsylvania***	0.02-0.11
Big Run Bog, West Virginia****	0.017
Buckle's Bog, West Virginia*****	0.015

*Gorham (1991, and reference cited therein), **Ovenden (1990), ***Cotter (1983), ****Wieder (1985), *****Maxwell and Davis (1972) vertical peat accumulation from basal radiocarbon dates.

Using ^{210}Pb dating of shallow (<40 cm) peat cores, we have quantified rates of vertical peat accumulation at sites along a latitudinal gradient ranging from the Marcell Experimental Forest, Minnesota (U.S.A.) to Big Run Bog, West Virginia (U.S.A.; Novák 1990, Wieder et al. 1994, Novák et al. 1994, Vile et al. 1995). Contrary to what would be predicted based on the radiocarbon dating approach, there is no clear pattern of greater vertical peat accumulation in the more northern sites than in the more southern sites. Rather the greatest vertical accumulation has occurred at Big Run Bog, West Virginia (U.S.A.), which over the past 200 years has been accumulating at an overall rate of 1.8 mm/yr and over the past 50 years has been accumulating at an overall rate of 3.8 mm/yr.

Regardless of dating technique, conversion from height growth to mass (or carbon) accumulation requires information on peat bulk density measured in each section of each core. In our vertical profiles from five sites in the northeastern U.S.A., when bulk peat density was incorporated into the determination of peat accumulation, mass accumulation rates appeared to be strikingly similar, especially over the past 100 years (Wieder et al. 1994). Because of site differences in peat bulk density profiles, comparisons of vertical peat accumulation rates may not reflect site differences in actual rates of mass accumulation.

Net primary production

Net primary production is a major factor controlling peat accumulation. Considerable effort has focused on factors that affect the relative abundance of *Sphagnum* species (Gignac et al. 1991, Gignac et al. 1991), as well as on factors that affect the growth and net primary production of *Sphagnum* species (Clymo - Hayward 1982). Since temperature, moisture and light all affect *Sphagnum* growth, relationships between net primary production and climatic variables might be expected across broad geographic scales. E.g., net primary production of *Sphagnum* generally increases with decreasing latitude (Wieder - Lang 1983), but this relationship is not useful in a predictive sense.

Superimposed on broad-scale geographic patterns may be local variations in net primary production of *Sphagnum*, resulting from variable inputs of industrial pollutants. E.g., sulfur pollution (sulfate, bisulfate, and sulfur dioxide) has been suggested as causally related to the disappearance of *Sphagnum* species from areas in southern Pennines (Ferguson - Lee 1983), and high nitrogen deposition has been implicated as inhibitory to *Sphagnum* growth (Press et al. 1986). However, neither high nitrate nor high sulfate substantially affected the growth of *Sphagnum* species collected from a high acid deposition area of the eastern U.S.A (Austin - Wieder 1983). The findings of Aerts et al. (1992) support the notion that high nitrogen deposition may initially stimulate *Sphagnum* growth, but eventually cause a shift in the primary element limitation of *Sphagnum* from nitrogen to phosphorus. Further, in response to experimental acidification, the growth and production of some *Sphagnum* species was enhanced for the first two years, but subsequently returned to control levels after four years (Rochefort et al. 1990), suggesting that short-term responses of *Sphagnum* to changes in anthropogenic atmospheric deposition may not persist in the longer-term.

The central role that *Sphagnum* plays in peatland ecosystems justifies autecological studies providing information on the growth requirements of *Sphagnum*. However, from the perspective of peat accumulation, data are needed on net primary production of the entire peatland community. Particularly lacking is information on belowground production. Existing estimates of belowground production in *Sphagnum* peatlands are rather variable (see references in Wieder et al. 1989), and even less is known about the short-term dynamics of belowground production (cf. Aerts et al. 1989) or the long-term contribution of belowground net primary production to peat accumulation.

Decomposition and carbon balance

Anaerobic conditions below the water table level, coupled with a progressive decrease in carbon quality over time, are considered to be the major determinants of peat accumulation. This view was supported by Clymo's (1984) review and mathematical models of peat growth. However, surprisingly few studies have measured either peat decomposition or aerobic versus anaerobic carbon mineralization as a function of depth in a peatland over a full annual cycle. The traditional litterbag approach continues to be used to evaluate decomposition processes in peat bogs. While the results are difficult to apply to carbon balance assessments, the litterbag approach is useful in evaluating the influence of such factors as *Sphagnum* species differences, location in hummock or hollows, and drainage intensity on *Sphagnum* decay (Lieffers 1988, Johnson - Damman 1991).

Recent studies have begun to assess decomposition in the context of carbon balance in peatlands, quantifying net CO_2 and CH_4 fluxes and identifying the factors that affect such fluxes (Crill et al. 1988, Moore 1989, Moore - Knowles 1989, Roulet - Moore 1992, Whalen - Reeburgh 1992). Such studies encounter considerable spatial and temporal variability within a particular site and results from a single study may not be generalizable to either a region or a particular type of peatland. Since peatlands have a potential to at least on occasion serve as a net sink for atmospheric CH_4 (Whalen - Reeburgh 1990, Yavitt et al. 1990), an estimation of the annual net CH_4 flux may require a rather intense sampling schedule. Even so, when the primary objective of the study is to evaluate the source/sink status for atmospheric CO_2 and CH_4 in a peatland, the data do not necessarily provide insights into carbon balance in peatland ecosystems. Moreover, even when field conditions (e. g., hydrology, temperature) are experimentally manipulated, flux measurements usually treat the peatland as a black box, essentially ignoring the microbial processes that ultimately are responsible for controlling the observed fluxes.

A more reductionist approach is to focus on quantitative assessments of the microbially mediated processes in peat that contribute to CO_2 and CH_4 production and/or consumption. Methane produced in the peat may be oxidized to CO_2, and thus would not be manifested as CH_4 flux to the atmosphere (Yavitt et al. 1988). Conversely, methane may be produced by CO_2 reduction within peat (Landsdown et al. 1992). In the absence of refixation of CO_2 by autotrophic organisms in the peat or by plants, and with export of dissolved inorganic carbon negligible, the ultimate fate of fermentation- and respiration-produced CO_2 or CH_4 is emission from the peat column to the atmosphere.

Our studies on anaerobic metabolism in two Appalachian peatlands (Wieder et al. 1990) revealed that sulfate reduction, rather than methane production, was the dominant anaerobic carbon mineralization pathway. The importance of sulfate reduction to anaerobic carbon mineralization has also been documented in the *Sphagnum*-dominated swamps of the New Jersey Pinelands (Spratt - Morgan 1990). Although West Virginia and New Jersey are in high sulfur deposition environments, the importance of sulfate reduction at these sites was not the result of high dissolved sulfate concentrations, but rather the result of rapid turnover of both the reduced inorganic S and dissolved sulfate pools. To date, the relative importance of sulfate reduction versus methane production in low S deposition boreal peatlands has not been addressed. Moreover, at both of our Appalachian peatland sites, total anaerobic CO_2 production exceeded that attributable to sulfate reduction and methane production combined. While attributable to "fermentation" reactions, the nature of these anaerobic CO_2-producing processes remains undetermined.

To assess the processes that contribute to anaerobic carbon mineralization in peat, we collected peat cores 10 cm in diameter, 25 cm deep from peatland 309 at the Experimental Lakes Area, Ontario (ELA), Bog Lake Bog at the Marcell Experimental Forest, Minnesota, and Big Run Bog, West Virginia. We measured CH_4 and CO_2 emission, as well as CH_4 consumption, from the intact peat cores incubated at 2, 12 and 17 °C. Subsequently we evaluated the temperature responses of both CH_4 production and CH_4 consumption and the response to glucose amendment of CH_4 production in flask incubations of 5 cm depth sections from each core. These studies indicated that for ELA peat, rates of CH_4 consumption exceeded rates of CH_4 production across all temperatures, leading to low CH_4 emission at all temperatures; CH_4 production was not stimulated by glucose amendment. While CH_4 production at Marcell peat was limited by organic substrate availability, CH_4 production generally exceeded CH_4 consumption especially at low temperatures, leading to high rates of CH_4 emissions that increased with temperature. Finally, at Big Run Bog, CH_4 production rates were moderate, were temperature dependent and were markedly stimulated by glucose addition. Methane consumption rates also were moderate, but were near zero at 20 °C, leading to moderate CH_4 emission that was temperature dependent (Yavitt and Wieder 1992, and unpublished data).

Results from these studies have been useful in helping to explain previously documented site differences in CH_4 source/sink relationships. Of broader implication, however, is that such site differences suggest that our three study sites may support quite different microbial communities. If further studies continue to support this suggestion, the validity of extrapolating results from single sites to broader geographic regions will be suspect.

As a novel approach toward relating present and potential future CO_2 and CH_4 fluxes from peatlands to both atmospheric chemistry and peatland carbon balance, attempts are being made to correlate carbon mineralization rates with the chemical quality of peat (Bridgham et al. 1992). Using these empirically derived regression relationships, one could predict the potential carbon mineralization of peat under a given set of temperature and moisture conditions. This approach could be useful in predicting the potential effects of global climate change on both CO_2 and CH_4 fluxes to the atmosphere and on peatland carbon balance.

The relationships between carbon quality, fermentation reactions, and carbon mineralization remain obscure. In our studies of Appalachian peatlands (Wieder et al. 1990), even at a depth of 35 cm where the peat appears to be highly decomposed, substantial anaerobic carbon mineralization was measured. Ongoing studies in which the growing vegetation in 1 m^2 plots at Big Run Bog, West Virginia (U.S.A.) was labeled with ^{14}C suggest an important role of vascular plant roots in fueling anaerobic carbon mineralization, even in deep (i. e., 20-30 cm) peat. Sampling of vegetation and peat 12 hours after labeling of the vegetation suggested substantial transport of ^{14}C to a depth of 10 cm in the peat (unpublished data). Approximately 80 % of the label that was recovered in a single plot was in the aboveground living vegetation, whereas approximately 20 % was in the dead non-green peat and associated plant roots. Of the belowground label, 93 % was recovered in the peat, with only 7 % in the roots which had been separated from the peat prior to determination of radioactivity. This distribution of label suggests that a substantial percentage of the daily C fixation by photosynthesis can be transported deeper into the peat column where leaky roots may supply high quality organic carbon to the heterotrophic microbial community. Root turnover represents another potential carbon source in deep peat. In a wet heathland, root biomass turnover rates for *Erica tetralix* and *Molinia caerulea* were 1.7 and 1.2 yr^{-1}, respectively (Aerts et al. 1989). Roots, through either leaking organic carbon or turnover, could fuel high rates of anaerobic metabolism in peat whose organic matter quality, as determined by proximate analysis, might be deemed unable to support substantial microbial activity. Data are still being collected in this study, but if this interpretation stands the test of time, vascular plant roots will merit much more attention than they have been afforded, especially in terms of understanding carbon balance in peatland ecosystems.

Summary

In this brief overview, we have tried to emphasize that substantial gaps exist in our understanding of processes related to carbon balance in peatlands and of the source/sink relationships for CH_4 and CO_2 between peatlands and the atmosphere. Net primary production obviously plays a critical role in peat accumulation, but it has been estimated for only a few sites, generally ignoring belowground production and its fate. While an increasing data set is developing regarding field measurements of CH_4 and CO_2 emissions from peatlands and the abiotic controls on emission, the extent to which site-specific studies can be generalized to broader areas remains uncertain. Relatively little work has focused on the microbially mediated mechanisms of CH_4 and CO_2 production/consumption within the peat itself. Even less is known about the microbial populations and communities that are responsible for carbon mineralization in peat. Such gaps confound efforts to project how peatland ecosystem function may be altered under predicted scenarios of global climate change.

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Globální změna klimatu a bilance uhlíku v rašeliništích

Od skončení poslední doby ledové bylo značné množství atmosférického uhlíku vázáno v biomase rašelinišť. Přestože jsou poměrně dobře známy procesy, které probíhají při akumulaci rašeliny a diagenezi pohřbeného uhlíku, existují značné mezery v našem chápání mechanismů řídících hmotovou bilanci uhlíku v substrátu rašelinišť a vztahů mezi zdroji a propady metanu a oxidu uhličitého v systému atmosféra - rašeliniště. Jen v malém počtu případů byla vyčíslena čistá primární produkce ekosystému rašelinišť, nebyla věnována dostatečná pozornost primární produkci biomasy pod povrchem rašeliniště. Není jisté, do jaké míry lze zobecnit měření emisí CH_4 a CO_2 z jednotlivých rašelinišť na širší klimatické pásy. Málo prací bylo věnováno mechanismům mikrobiální produkce a metabolismu CH_4 a CO_2 uvnitř zrajícího substrátu a rovněž dynamickým parametrům populace mikroorganismů zodpovědných za mineralizaci uhlíku v rašelině. Tyto mezery v porozumění faktorům ovlivňujícím hmotovou bilanci uhlíku ztěžují snahu předpovědět odezvu ve fungování ekosystémů během současného oteplování zemského povrchu.