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CLIMATIC FLUCTUATIONS REFLECTED IN SLOPE AND FLUVIAL SYSTEMS OF POLISH CARPATHIANS AND THEIR FORELAND DURING UPPER QUATERNARY

Abstract. On continents, besides lake and bogs, we observe non-continuous sedimentation where particular layers or whole members represent time intervals of various lengths separated by breaks of different origin. Both, fluvial and slope sediments or forms carry climatic signal. By analyzing sources of sediments, factors of transfer, duration of deposition we reconstruct climatic changes (mainly temperature, precipitations and circulation of water). We order the collected records after age and tracing longitudinal profiles of slopes and river valleys. The slopes and valley floors inform us about spatial differentiation of extreme climatic-hydrological events and on their reflection in degradation or aggradation. All this information sums up the records collected in various projects or commissions like IGCP-158 and GLOCOPH (S t a r k e l ed. 1982–1996; S t a r k e l et al. 2007), Climatostratigraphy of the Holocene of Polish territory (S t a r k e l et al. 2013) as well INTIMATE (S t a r k e l et al. 2015; G ę b i c a et al. 2015). It may be concluded that fluctuations in temperature combined with expansion of permafrost were the leading factors in transformation of landscape of analyzed area of Southern Poland during last cold stage, which is in contrary to the Holocene, when variations in humidity especially in frequency of extreme events played a leading role. The continental records very well express the role of transitional phases.

Key words: climatic fluctuations, slope system, fluvial system, upper Quaternary, Polish Carpathians,

STAND OF THE ART

Climatic changes reflected in continuous deposition on the continents can be traced only in depressions protected against rapid input or output of flowing water and mineral matter from outside. Frequently, in a vertical profile we observe non-continuous sedimentation where particular layers or their complexes or even whole members represent time intervals of various lengths: starting from products of single downpours, through multi-annual layers formed by colluvia or eolian activity, to soil profiles even developed during millennia. Such sequences of sediments include many gaps caused either by breaks in deposition or by erosional activity. Also gaps represent time intervals of various duration.

When trying to compare such different time intervals, carrying climatic signal, with continuous sequences of marine, glacial or lake deposits, not only we date these continental deposits but we also order them in spatial systems, namely in longitudinal profiles of slopes and river valleys where they reflect transfer of energy and matter. We should identify sources of removed material, types of transfer (flowing water, gravitation, wind etc.), relief of forms as well as duration of deposition. All these parameters are helpful in reconstruction of climatic changes, especially temperature, precipitation and circulation of water.

The way of ordering will be exemplified by slope and river valley longitudinal profiles mainly located in the Polish Carpathians and their forelands.

SLOPE COVERS

In the longitudinal slope profiles in their upper segments we find residual eluvia or only hard bedrock forming sometime cryoplanation terraces. Moving downslope we observe deposits removed by congelifluction (over permafrost) or connected with deep infiltration (landsliding), overland or subsurface run-off (piping) etc. On their top there are frequently loess blankets, the indicators of climatic aridity.

In the Polish Flysch Carpathians the slope sediments from the last cold stage are to 10–20 m thick at several localities. Among them, there were identified post-Brörup or Odderade lower Pleniglacial beds (75–58 ka BP), diversified Interpleniglacial (58–28 ka BP), upper Pleniglacial (28–14 ka) and Lateglacial members (Figs. 1–5). In vicinity of Wadowice town at elevation below 300 m a.s.l.,

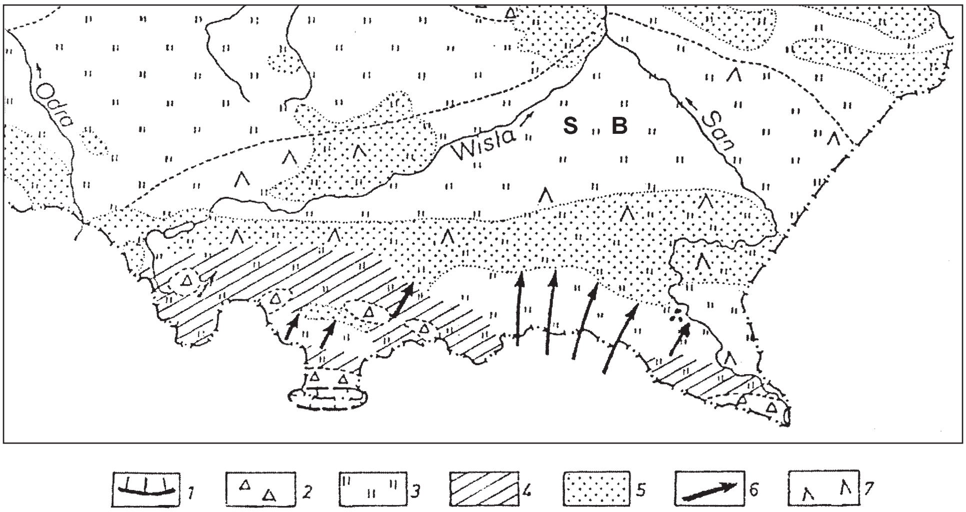

Fig. 1. Palaeogeography of the Carpathians and its foreland during Pleniglacial (Starkel 1988). 1 – glaciated area, 2 – debris covers and active cryoplanation, 3 – tundra-steppe vegetation, 4 – intensive solifluction (areas with higher precipitations), 5 – areas of loess deposition, 6 – directions of intensive deflation, 7 – areas of probable preservation of trees, SB – Sandomierz Basin

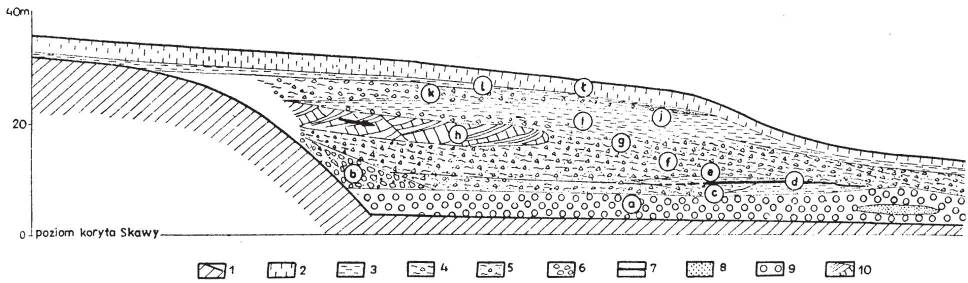

Fig. 2. Fluvial and slope deposits from last cold stage at Wadowice (after L. Starkel, in: Sobolewska et al. 1964). 1 – flysch rocks, 2 – loess (t), 3 – deluvial loam (e, i, j, l), 4 – solifluction clay with debris (f, g, k), 5 – solifluction sandy loam (c), 6 – talus debris (b), 7 – peat (d) from Brörup, 8 – fluvial sand, 9 – fluvial gravel (a), 10 – rocky landside from interpleniglacial (h)

above the Brörup peat (Odderade peat after J. Mojski 2005) a member of older Pleniglacial starts with deluvia followed by solifluction layers with coarse debris (Sobolewska et al. 1964). The Interpleniglacial with at least two warmer oscillations is represented in that locality mainly by deluvia with only rocky landslide block about 20 m long (Fig. 2) suggesting a total melting of permafrost. In Dunajec river valley the solifluction beds were dated by TL method at transition from older Pleniglacial to Interpleniglacial 60–51 ka BP (Butrym, Zuchiewicz 1991). Similar thick deluvial units in Zabrodzie separate two solifluction beds with coarse debris from Pleniglacial phases. A long phase of deep infiltration is marked by decalcification of sandstone debris (Dziewański, Starkel 1967). In Dobra (470 m a.s.l.) above fluvial gravels alternated with solifluction layer with forest-tundra vegetation dated at 32,550 ± 450 ^{14}C BP, there is about 9 meters continual deposition of colluvial loams with sandstone boulders. In these colluvial loams also *Pinus* pollen with tundra vegetation occurs (Klimaszewski 1971; Środoń 1968, Fig. 3). This indicates a very active congelifluction at the decline of Interpleniglacial, similar to that observed at present time in eastern Siberia (Baulin et al. 1984). But in the whole profile the lower part with lenses of ^{14}C dated peat do not differ from the upper ones. The whole sequence in my opinion represents the deposition during decline of Interpleniglacial characterized by thick active layers over permafrost.

At another locality, in Krościenko (> 400 m a.s.l), there are exposed 13 meters of solifluction colluvia with 5 horizons with coarse debris alternated with 4 clayey layers (Klimaszewski et al. 1939) which probably represent the same period with cyclic oscillations. Degradation of permafrost has been dated by OSL at 47.2 ± 3.5 ka BP in colluvia filling upper course of the Jamne valley in the Gorce range (700–1100 m a.s.l.) (unpublished).

In the wide transversal gate between Western and Eastern Carpathians wind played a leading role during the Pleniglacial (Gerlach at al. 1972; Gerlach 1990; Starkel 1988; Fig. 1). At the northern margin of the intramontane Jasło-Sanok Depression the lower, 5 m thick, part of the Carpathian silty-sandy

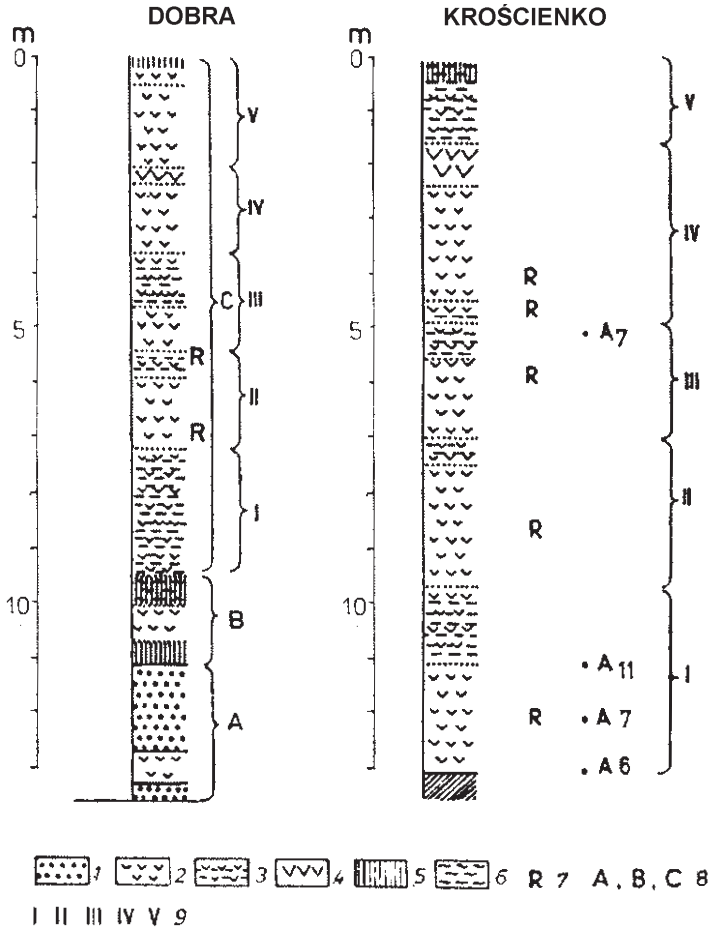

Fig. 3. Interpleniglacial slope deposits at Dobra (Klimaszewski 1971; Środoń 1968) and Krościenko (Klimaszewski et al. 1939) in Polish flysch Carpathians. 1. fluvial gravels, 2. solifluction with debris, 3. solifluction dominant clay,4. coarse debris, 5. silt and clay, 6. organic remains (horizon dated 32550 ± 450 BP), 7. rare organic remains, 8. sediment members, 9. main sediment units

loess was dated by TL at >35–30 ka BP in Humniska (Gerlach et al. 1991, 1993; Fig. 4). The local source of material is documented by composition of heavy minerals identical with flysch bedrock. It means that during the Interpleniglacial some phases of strong eolian activity had taken place. To the south, the zone of deflational depressions surrounded by higher plains with ventifacts extends (Gerlach 1990).

After about 25 ka BP during distinct cooling and rise of aridity in the whole Central Europe (Maruszczak 1980) slope processes probably decreased and in the Carpathian foothills the youngest loesses were deposited, mainly blown out from valley floors (Łanczont 1995). That member in Humniska was 5 m thick, but it was even thicker locally, like at the margin of Carpathians along

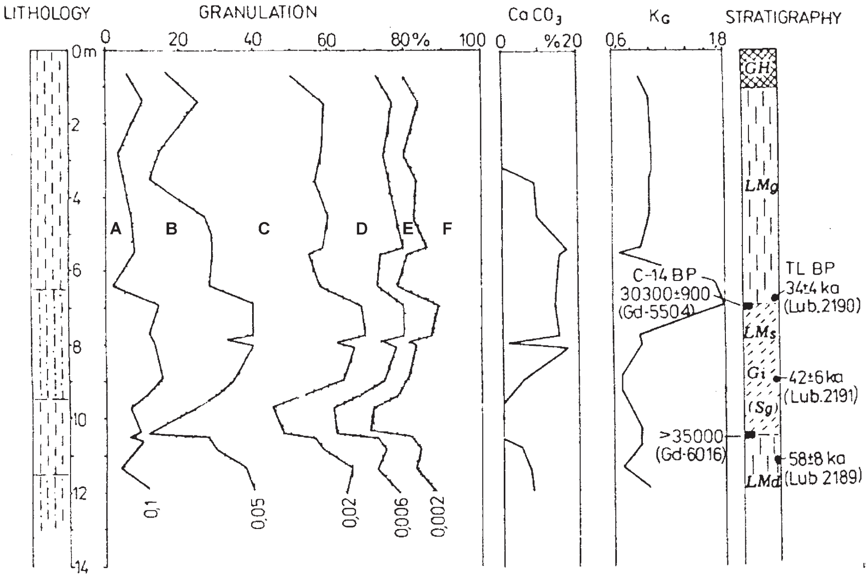

Fig. 4. Profile of eolian deposits (Carpathian loess) in Humniska at margin of the Jasło-Sanok depression (Gerlach et al. 1991) Grain size analysis, $CaCo_3$ content, C_{14} and TL datings

the uppermost Vistula (Niedziałkowska, Szczepanek 1994) and San valleys (Łanczont 1995). In upper courses of some greater mountain valleys, above the Interpleniglacial overbank alluvia with peat layers there were deposited either solifluction beds or alluvial fans of tributary valleys (Birkenmajer, Środoń 1960; Krysowska-Iwaszkiewicz, Wójcik 1990). In several terrace vertical profiles it is reflected in gradual decline in roundness of gravels (Starkel 1965). In a small valley of Lipowe I found the interfingering of thin congelifluction tongues with alluvial gravels reflecting the alteration of late spring snowmelt floods and late summer creeping (Starkel 1960a).

For the Lateglacial there are several reconstructions of temperature based on vegetation changes and periglacial structures (Madeyska 1995). The upper timberline (coinciding with July temperature about +10°C) in the Allerőd raised to about 1000 m a.s.l. in the Polish Carpathians (at present it is located 1500 m a.s.l.) and during the successive Younger Dryas dropped down to about 500 m a.s.l. (Ralska-Jasiewiczowa 1980). Comparing the ^{18}O curve from Greenland for the Allerőd and Younger Dryas with fluctuations of similar order during the Interpleniglacial we may expect that similar changes of an upper limit of trees as well as a permafrost limit repeated.

In higher elevations of the flysch Carpathians the last phase of activation of slope processes is recorded on the periglacial cryopediments-glacis in Gruszowiec in the Beskid Wyspowy Mts. (Klimaszewski 1971; Starkel 1960a; Fig. 5). At the outlets of gullies cutting the upper steep slopes rising to 1000 m a.s.l. there are superimposed on congeliftuction glacis the small torrential fans built of

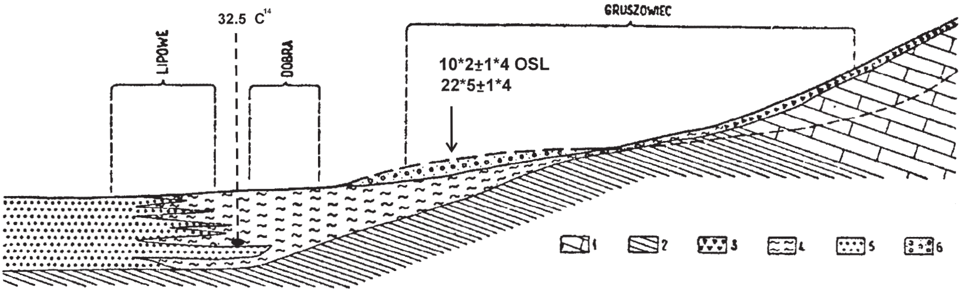

Fig. 5. Periglacial covers over slopes and in valley floors of the Beskid Wyspowy (after L. Starkel 1960b). 1 – thick-bedded Magura sandstones, 2 – Submagura sandstones and shales, 3 – debris covers, 4 – solifluction covers, 5 – alluvia, 6 – overlying lateglacial fans at the outlets of gullies dissecting slopes (dated by OSL); localites representing various parts of synthetic profile underlined

coarse gravels dated at the top by OSL method only at 10.2 ± 1.4 ka BP (Younger Dryas – Starkel 2014). Similar OSL age of 11.16 ± 0.99 ka BP was found from alluvia of an overflood terrace in a middle segment of a small Jamne valley in the Gorce Mts. The Allerőd was the time of a first phase of deep rocky landslide formation in the flysch Carpathians (Margielewski 2006). It means that permafrost had totally disappeared and new groundwater regime established.

ALLUVIA

The correlation of upper courses of valley floors (with alternating erosional–depositional phases) with dominant aggradational Subcarpathian basins shows very distinct, climatically controlled phases.

In the mountains the last cold stage is represented by two terraces. The higher one that occurs in uplifting valley segments and is built of gravels with erosional rocky benches. It was dissected at the decline of Interpleniglacial (Starkel et al. 2007) and later covered by slope deposits. In the marginal part of the Carpathian foothills the younger fill from the upper Pleniglacial has a base only 1–3 m above the present water level. In some valley segments the downcutting in bedrock is continuing (Starkel 1960b). At the foreland of the High Tatra Mts., in the Podhale intramontane basin the upper Pleniglacial advance of glaciers is reflected in formation of a younger system of glacifluvial fans and their retreat is marked in gradual formation of Lateglacial marginal moraines (Baumgart-Kotarba 1983; Baumgart-Kotarba et al. 2008).

In the Subcarpathian Basins the extensive Vistulian and Holocene alluvial fans occupy over 40% of the whole area. The Vistulian is represented by two horizons 15–20 m and 6–12 m above present channels (Starkel 1995a; Gębica 2004; Starkel et al. 2007; Gębica et al. 2015; Fig. 7). The higher one in Wisłoka valley has at base the remains of the Early Vistulian and early Pleniglacial alluvia as well as older Interpleniglacial stadial Hengelo with remains of boreal forest dated above 40 ^{14}C ka BP (Mamakowa, Starkel 1974). The younger fill in that higher terrace contains overbank sediments dated about 30–28 ^{14}C ka BP with distinct cryoturbation and tundra soils near the top. Several kilometers to the south of Brzeźnica at a foot of the Carpathian escarpment is buried even a younger fill built of sands with gravels dated by TL at 22–23 ka BP (Mamakowa et al. 1997).

Along the left bank of Vistula river these sediments, dated by OSL at 25–22 ka BP, were covered by several meter thick loess (Gębica 1995). It means that a change to a cooler continental climate and a dissection of the higher level took place 2–5 ka before the maximum extent of the last Scandinavian ice sheet (Rotnicki 1987; Maruszczak 1980).

The lower terrace, 12–8 m high, is formed by a younger fill which developed on an erosional platform (Starkel 1995a) and its base was dated at 17800 ± 1300 ^{14}C BP yet in a site located farther north in the Wisłoka valley (Gębica 2004). This lower level has a shape of a fan extending downstream and its surface with braided channels was modified by Lateglacial dune fields (Gębica et al. 2015). Near the outlet of the Wisłoka fan to the wide Vistula valley floor there were described two localities which evidenced the distinct transition from a fluvial unit with periglacial structures through fluvio-aeolian member with cryoturbations to aeolian unit on the top (Wojtanowicz 1996; Superson et al. 2010). The lowest part was dated by TL at between 29 and 16 ka BP, the middle at 14–13 ka BP and the upper one at the Lateglacial to early Holocene (Zieliński et al. 2014). It means that downcutting was onset at least from 15 ka BP and large palaeomeanders started to develop on the lowest level (see Fig. 6).

During the end of the Pleniglacial, with a gradual warming and expansion of forest–tundra vegetation a decline in sediment load occurred which was reflected in continuation of incision of river channels even below the present water level. A similar deep incision was recorded even earlier in the lower course of the Morava river where tree trunks in the base gravels were dated at between 22 and 16 ^{14}C ka BP (Havliček 1991). The filling of one of such palaeochannels in

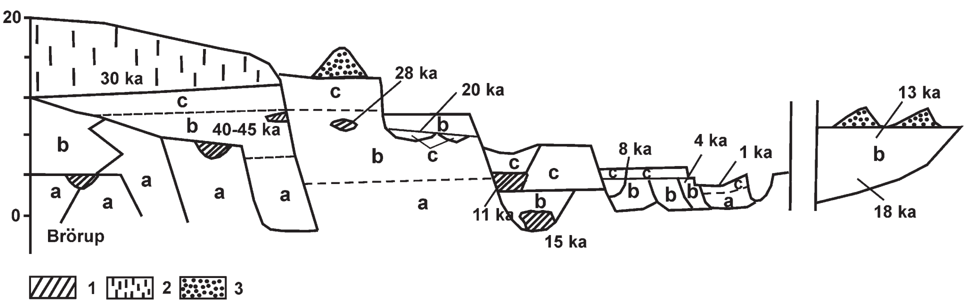

Fig. 6. Synthetic cross-section of valley floors of Carpathian river at their outlet from the Carpathians. a – channel facies of alluvia, older than interpleniglacial, b – channel facies of alluvia-younger (from Interpleniglacial – to Holocene), c – overbank and palaeochannel fill facies, 1. peat, 2. eolian loess, 3. sand dunes

the San valley started about 15 ka BP (Klimek et al. 1997) and the other in the Vistula valley near Krakow at 13260 ± 160 BP (Kalicki 1991). Such buried palaeochannels were common at the level of 6–9 m high terrace, which was extended laterally by Lateglacial large meanders. Only in the valleys with steeper gradients, rivers carrying coarser gravels and draining higher Carpathians preserved braided pattern even till the early Holocene (Sokołowski 1995; Starkel et al. 1996). Some large meanders were abandoned very early (in Bieruń dated 12500 ± 230 BP – Klimek 1995), the other were active till the transition Younger Dryas – Preboreal (Starkel 1991). Their parameters were greater then those of the Holocene (widths 3–5 times larger, meander radii 2–5 times larger) and bankfull discharges 8–20 times larger (after R. Soja – in Starkel et al. 1996). These two phases of distinct changes (braided – large meanders – small meanders) were connected with two factors: first, with expansion of forest; second, with a change of types of precipitation and floods (from snowmelt to summer rainy floods – see: Kozarski et al. 1988). Therefore, farther to east of the Dnieper river large palaeomeanders were abandoned 3–4 ka earlier due to much earlier expansion of forests over permafrost (Panin et al. 2001; Starkel et al. 2015).

During the Holocene a gradual up-building or dissection of the Lateglacial level took place during several phases of cutting and filling connected with phases of frequent floods leading also to avulsions of river channels (Starkel et al. 1996). Thanks to more than 100 radiocarbon datings and about 200 dendrochronological datings in the Vistula valley and in the valleys of its Carpathian tributaries it was possible to distinguish several phases of increased fluvial activity accompanied by formation of new fills and avulsions of channels dated at 8.5–7.8 ^{14}C ka BP (9.5–8.5 cal ka BP), 7.3–7.0 ka, 6.6–6.0 ka, 5.5–4.9 ka, 4.4–4.1 ka, 3.5–3.0 ka, 2.7–2.4 ka, V–VI c. AD and VI–XIX c. AD (Starkel 1995c; Starkel et al. 1999, 2006). During these phases it was possible to distinguish several, decades-long, clusterings of heavy rains which can be evidenced by subfossil oaks (Krąpiec 1998).

Excellent example of ca. 95 rainy events occurring between 8390 ± 105 and 7785 ± 145 ^{14}C years BP deliver the over 5 m thick alluvial fan of small creek at Podgrodzie (Niedziałkowska et al. 1977; E. Czyżowska – in Starkel et al. 1996). The records from this locality coincide in time with other records from the Carpathians, Alps and other parts of Europe documenting the first, early Holocene wet phase before the climatic optimum (Starkel 1999). Several landslides in the Flysch Carpathians and reactivation of debris in the alpine belt of Tatra Mts. (Kotarba, Baumgart-Kotarba 1997) coincide with that phase.

In last millennia the effects of heavy rains are much better expressed due to progressing forest clearance and cultivation. Therefore, the deposition of thick layers of overbank loams might occur during relatively drier period of late Roman time (II–IV c. AD), of mid–Medieval period (IX–X c. AD) or late Medieval XIII–XV c. AD (Starkel 2005). In the tectonically subsiding foreland of the Silesian Beskid the overbank aggradation during the last millennium reached more than 5 meters (Niedziałkowska et al. 1977). On the contrary the ”dark ages” of V–VI c. AD which were cooler and wetter are characterized by lack of thick overbank loams but frequent deposition of oak trunks in abandoned palaeochannels and rise of lake level (Krąpiec 1998; Starkel 2005).

The fluctuations of temperature are weaker reflected in fluvial continental deposits. Good indicators of mean summer temperatures are plant communities and certain plant taxa. An expansion of *Quercetum mixtum* at ca. 10–9.5 cal ka BP reflects an amelioration of the climate while its retreat at about 5–4 cal. ka BP with simultaneous advance of *Fagus* and *Abies* point to a cooling and wetter phase (Ralska-Jasiewiczowa 1980). In the Carpathians this two successional phenomena are separated by a distinct cooler phase marked with a decline of *Picea* (Starkel 1995d). Beside that an early – mid Holocene climatic optimum is reflected in advanced deposition of calcareous tuffa. This deposition decreased later (Ložek 1991; Pazdur et al. 1988).

DISCUSSION AND CONCLUSIONS

The correlation of deposits and forms in slope profiles and longitudinal profiles of river valleys provide the information about climatic fluctuations both during a cold stage and during the Holocene. The correlation does not lead to much data on mean temperature or precipitation but tells more about the role of extreme hydrological events and their spatial differentiation. For this reason it is important to study parallel various facies of terrigenic sediments and forms, mainly related to instabilities in a hydrologic regime. As the examples of such approach may serve the last studies on Holocene climatostratigraphy for Poland, based on several hundreds dates, elaborated by a multidisciplinary team (Starkel et al. 2013). The studies of forms and sediments help to recognize in detail the character and frequency of rainfalls. During wetter phases of the Holocene frequent heavy downpours, continuous rains and rainy seasons appeared simultaneously (Starkel 1995b).

Similar procedures may be used for reconstructing cold phases of periglacial climate. The study on reconstructing climate of last glaciation in the Subcarpathian Basins, based on alluvia analyses, was prepared for INTIMATE project (Gębica et al. 2015) and included to a review paper for Central-Eastern Europe (Starkel et al. 2015). In hilly areas we reconstructed fluctuations in vertical zonality, as reflected not only in vegetation but also in expansions and decay of permafrost. We search for analogies in contemporary processes in other areas like Siberia or Mongolia (Baulin et al. 1984; Kowalkowski, Starkel 1984). By comparison with the palaeotemperature curve from Greenland we concluded about the effects of long Interpleniglacial when frequent high amplitude cyclic fluctuations of temperature (based on $^{18}0$ curve) played great role in degradation of active layer on slopes, and in deposition in upper valley segments. The material removal to the Subcarpatian Basin was more effective than during the coldest stadial when delivery was restricted by shallow active layer and eolian activity. Similar increased transfer appeared at the decline of Pleniglacial and during Lateglacial being accompanied by deepening of river channels.

In general, during the cold stage the fluctuations of temperature were leading factors in transformation of landscape in Central-Eastern Europe, which is in contrary to the leading role of fluctuations in hydrological regime during the Holocene. Also it seems that for landscape formation an important role played the sequence and duration of transitional phases during Last Glacial–Interglacial (Starkel et al. 2015). A long unstable early Vistulian (5a–5d) was a time of degradation of products of Eemian soil formation simultaneous with uplifting of southern mountain part of west – Carpathian valleys accompanied by lowering of upper forest line. This caused an incision of channels in bedrock (Binka, Grzybowski 2001) also reflected in deposition of thick alluvial fans on foreland (see Hradecky et al. 2011). Later processes were stabilized with expansion of permafrost in lower Pleniglacial (Fig. 7). After that, ca 30 ka long

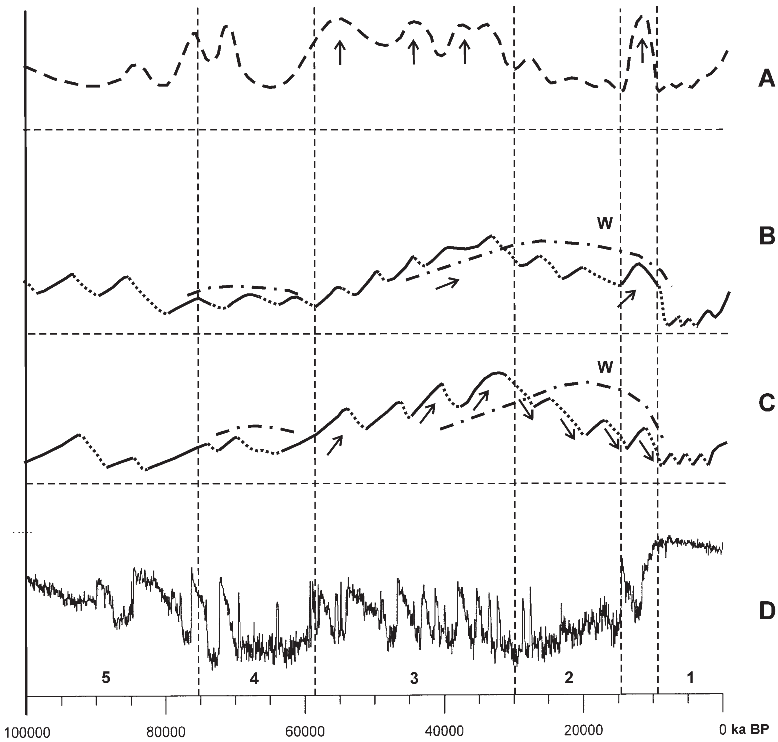

Fig. 7. Type and intensity of processes during last Vistulian in Central-Eastern Europe in relation to $^{18}0$ curve from Greenland (schematic reconstruction). A. Intensity of denudation on mountain slopes, B. Tendency to agradation and to erosion in mountain valleys, C. Tendency to agradation or erosion in submontaneous basins, D. $^{18}0$ curve from Greenland. Main stratigraphic units 1–5, w – phases of intensive eolian activity

Interpleniglacial period of great instability took place and manifested itself in degradation of slopes, alternating freezing and deep melting, finalized in removal of less resistant rocks and their deposition at the base of slopes and at a mountain foreland. The next coldest phase (Stage 2) of the upper Pleniglacial was relatively short (10–15 ka) and characterized by a protective role of deep permafrost, especially at the foreland of the Scandinavian ice sheet, exposed to strong wind activity (Dylik 1956, 1967; Goździk 1995; Starkel 1988; Zieliński et al. 2014). It was a time of gradual deepening of greater river channels, which is in contrary to previous opinions on common aggradation. This incision was alternated with formation of new fills and lateral extension of valley floors. A gradual return to forest vegetation reduced the role of wind and finally dunes were stabilized in whole Poland (Manikowska 1995). During the Lateglacial the final melting of permafrost in the mountains occurred and groundwater reservoirs were recharged, which supported piping, gullying and landsliding. In the Holocene fluctuations in precipitation and runoff regime started to play the leading role (Starkel et al. 2006). With Neolithic agrarian revolution the contrasts in sediment load increased again and the pulsation in degradation was controlled by coincidence of two factors: phases of clustering of heavy rains with floods and waves of deforestation and agriculture (Starkel 2005; Dotterweich 2008).

It seems that during the last Quaternary climatic cycle neither the J. Büdel's concept on "tote Landschaften" in Holocene (Büdel 1944) nor my concept on Holocene dense dissection of hilly landscape is fully valid (cf. Starkel 1960b). The role of permafrost was dominant in the narrow belt and was accompanied with a strong wind action (Gerlach 1990). The Holocene degradation was important on less resistant substratum, especially after deforestation supported by subsurface drainage. The phases of greatest transformation coincided with periods of transitions: retreat of vegetation cover (5d–5a) and melting of permafrost (stage 3 and transition 2-1) as well clusterings of extreme hydrometeorological events (supported by human activity).

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