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The Palaeogeography of Mid- and East Europe During the Last Cold Stage, with West European Comparisons

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The palaeogeography of mid- and east Europe during the last cold stage, with west European comparisons

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In this review of work on mid- and east European palaeogeography during the last cold stage the author has used a stratigraphic division based on deep-sea sediments in the Atlantic. Difficulties apparent in palaeogeographical reconstructions are discussed. In such reconstructions the sequences of permafrost structures, loesses, fossil soils, slope sheets, fluviatile sediments, vegetation, glacial deposits and land forms are analysed. The consideration of events in both their stratigraphic order (time scale) and their areal distribution in Europe has enabled the author to give a far more accurate indication of temperature and moisture changes in the different parts of Europe. Contrasts revealed by a section through east and west Europe reflect differences in the evolution of permafrost, in the succession of vegetation and soils, and in the history of glaciation. Their source was the different degree of climatic continentality which increased throughout the last cold stage. Continentality was characterized by the contemporaneous decrease in temperature, increase in temperature amplitudes and decrease in precipitation values. The presence of permafrost in Europe was responsible for the different development of biota. This makes for difficulty in reconstructing palaeogeographic conditions on the basis of actualism.

The duration of the last cold stage and the bases of stratigraphy

The stratigraphic divisions by Woldstedt (1958), Gross (1964) and others based on loess sequences with fossil soils, and the curves of van der Hammen, Maarleveld, Vogel & Zagwijn (1967) based on sequences of both alluvia and lake deposits which have been studied palynologically did not draw the lower boundary of the last cold stage (Vistulian, Weichselian) beyond 70000 a B.P. The evidence used was based on radiocarbon dates made mainly in Gröningen. These gave an age of 50000–65000 a B.P. to the Early Glacial stadials. However, dates of sediments older than 50000 years are open to doubt.

At the same time, further evidence was presented by the dating of deep-sea sediments made by Emiliani and others. These revealed the occurrence of a warm phase some 90000–100000 years ago. Moreover, the $\delta\,^{18}O/^{16}O$ determination of age of an ice core from Greenland indicates that the cool Pleniglacial period started as early as 73000 a B.P. (Dansgaard, Johnsen, Clausen & Langway 1970).

Further studies of deep-sea sediments undertaken by Emiliani (1968), McIntyre & Ruddiman (1972), Sancetta, Imbrie & Kipp (1973) and others, based on measurements of the $^{18}O/^{16}O$ ratios in skeletons of planktonic foraminifera, on palaeotemperature reconstructions of water bodies on the basis of foraminifera species, and on terraces on the isle of Barbados (Matthews 1972) indicate that the warmest period occurred some 125000–115000 years ago. This was followed by two separate warm waves thought to be interstadial (Understanding climatic changes 1975). The simplified curves of Kukla & Kukla (1972) and others show a succession of solar radiation cycles lasting about 20000 years each. The lower unit of the last cold stage includes at least two great waves of climatic cooling and of warming with relatively smaller amplitudes. The clearly colder second unit may be subdivided into two very cold phases separated by a longer (30000 a) cool temperate period which shows numerous smaller oscillations. According to the Dutch nomenclature, the above units correspond to the 'Early Weichselian' and to the Pleniglacial with the intervening Interpleniglacial. Wijmstra's (1975) attempt to correlate vegetational changes gives clear indication of their similar rhythm in the different zones, especially in Macedonia. Additional evidence is given by the distinct change in the magnetic declination at the top of the Eemian soil at Modřice in Moravia (Bucha, Horaček, Koči & Kukla 1969) dated at 114000 a ago. The palaeotemperature curve from Greenland remains to be interpreted. The small fluctuations of $\delta^{18}O$ values beyond 78000 a B.P. and the short term and abrupt depressions at *ca.* 89500 and 109000 a B.P. may have their origin in the retarded response of the North Atlantic water-body to temperature changes.

On the base of the palaeotemperature curve derived from undisturbed deep-sea sediments (for comparison, see figure 8) the views on the age and velocity of palaeogeographical changes of the continents require thorough revision. These changes were very rapid. Both glaciation (J. Lundquist 1974; Andrews *et al.* 1975) and deglaciation (Flint 1971) took place over a period of less than 10000 years each, whereas the duration of the Early Glacial has to be extended in length. The present author subdivides the last cold stage into two main units, which include the Early Glacial and the Pleniglacial.

The reliability of information on the palaeogeography of mid- and east Europe during the last cold stage

As we go further back into the Pleistocene the available information decreases. During the last cold stage, the most essential part in obscuring older features and destroying sediments was played by the latest cold period, including the glaciation 20000–15000 years ago. The use of radiocarbon dating is limited where deposits are older than 50000 a B.P., so that stratigraphic identification and correlation of the incomplete series of deposits is frequently made impossible. More often, the correlation of events is based on the comparison of sedimentary sequences with type horizons. The following difficulties are apparent in the reconstructions of climatically controlled phenomena: underestimation of the extent of breaks in the depositional sequence, inadequate knowledge of the varied local habitat conditions and topoclimates, and the correlation of data over wide areas in the reconstructions of local events. Non-sequences are most frequently observed in sedimentary sequences that have been exposed either to erosion or to denudation. In the valleys the shift of stream channels that has been recorded from Holocene valley spreads (Klimek & Starkel 1974) is responsible for the lateral contact of rock series of different ages. Because of slope degradation more or less complete series of deposits tend to be preserved only at the foot of the slope. They frequently contain double horizons due to the redeposition of material derived from the upper slope sectors. Woodland periods are reflected in the nature of polygenetic soils whose subsequent alteration was related to the succession of plant communities.

Because other information (excluding pollen spectra) provided by deposits refers only to the site examined and to its neighbourhood, generalizations from results should be made with great caution. As in Alaska and Siberia at present, the former ecosystems and geosystems showed much local variation related to thermal and moisture conditions, and to the thickness of the active layer above permafrost.

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The radiocarbon-dated interstadial type horizons found along the North Sea coast are used in correlating horizons in other regions including Poland. However, such correlations are rather uncertain. In palaeoclimatic reconstructions, for instance, in Poland and in Hungary, the fossil floras are related either to the distant type profiles or to the ecological requirements of vegetation as observed at present, although the thermal parameters of air were different in a permafrost environment, with warmer summers and severe winters prevailing in the east.

In Poland and neighbouring countries, the palaeogeographic changes of the Vistulian stage are documented by sequences of the following phenomena: permafrost structures, loesses, fossil soils, slope sheets, fluviatile deposits, vegetation, glacial deposits and land forms.

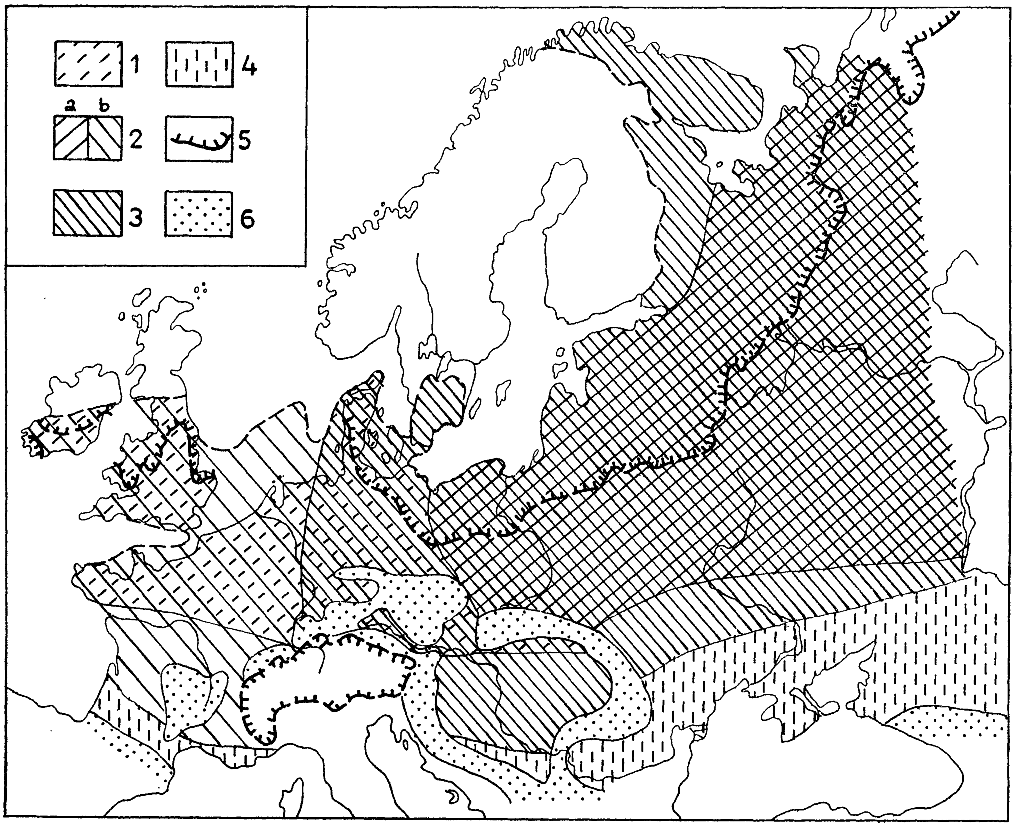

FIGURE 1. Extent of permafrost during the Vistulian (after Velichko 1973). Key: 1, Atlantic type (older Pleniglacial); 2, transitional type: (*a*) older Pleniglacial, (*b*) younger Pleniglacial; 3, Siberian type (younger Pleniglacial); 4, areas of seasonal permafrost (younger Pleniglacial); 5, extent of inland ice; 6, mountains with frost weathering features.

The sequence of permafrost structures

Permafrost surviving over many years was the primary feature of the cold period in Europe (figures 1 and 2). Thus the structures associated with permafrost are the best indicators of former changes of climate (Dylik 1969*a*). In mid- and east Europe permafrost built up slowly. Its earliest traces are materials enveloping the ice cores of fossil pingos (Dylik 1967). Below the horizon of the first Early Glacial interstadial ice-wedge casts also occur, even in Belgium (Paepe & Zagwijn 1972). Two Early Glacial ice-wedge horizons are reported from the Małopolska Upland (Jersak 1973) and from Hessen (Rohdenburg 1967). In west Europe, predominant plastic deformations indicate seasonal freezing (Velichko & Berdnikov 1973).

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The very beginning of the older Pleniglacial is marked by a phase of severe climate accompanied by permafrost of Atlantic type, indicated by ice-wedge casts, 2–4 m deep. Temperatures of the ground surface are believed to have dropped to −3 or −4 °C. The thickness of permafrost must have exceeded 100 m, following Velichko's calculations (1973). The mean annual temperature ranged from −6 to −8 °C (this is the annual mean required for the growth of ice wedges (Péwé 1966)). According to Rohdenburg, several generations of ice wedges indicate a discontinuous permafrost with recurrent phases of formation. However, typical 'Siberian' forests of the Interpleniglacial and shallow gley soils give evidence rather of seasonal changes in thickness of the active layer. During the Interpleniglacial permafrost disappeared in the Hungarian Plain.

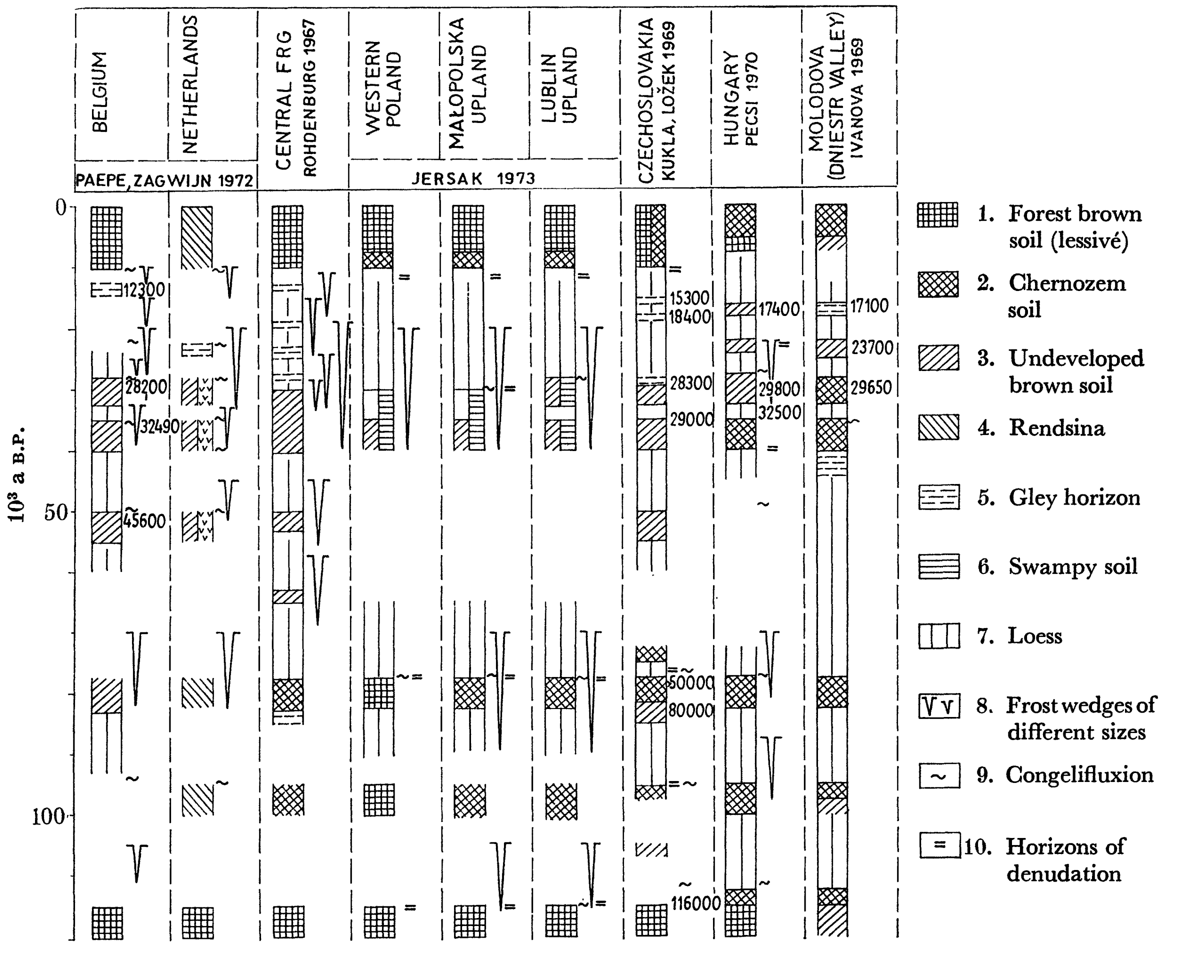

FIGURE 2. Age of fossil soils and permafrost phenomena in different European loess areas during the last cold stage.

The younger Pleniglacial is characterized by a repeated cooling of climate. Permafrost penetrated to greater depths then since the middle horizon of the youngest loess contains ice-wedge casts, 4–6 m deep, and polygons to 20 m in diameter (Rohdenburg 1967; Jersak 1973). Pseudomorphoses which developed from sand wedges devoid of ice give evidence of themarked desiccation of climate (for comparison, see Goździk 1973). The average ground temperature dropped to −6 °C. Permafrost was 200 m thick and extended as far south as latitude 45° N (figure 1). The severity of climate is also indicated by permafrost expanding in areas left freshly exposed by ice retreat (Maruszczak 1960). For example, ice-wedge casts may be found in deposits that have been laid down between two glacial oscillations in Wales (Saunders 1973).

Sand wedges filled with aeolian sand occur even within the boundary as the Pomeranian phase (Kozarski 1974) and point to the marked continentality of climate in east Europe. Kozarski suggests that during the deglaciation the mean annual temperature must have been by 14–17 °C lower than now. The degradation of permafrost was relatively rapid towards the end of the Last Glacial because in central Poland the last phase during which permafrost became reactivated corresponds to pre-Bølling times (Chmielewski 1970). During the Allerød the blocks of ice deeply buried in the subglacial channels were already melting. Degradation of the deeper and drier permafrost proceeded faster under continental climatic conditions than in west Europe. It might have been much slower there since pingo-like forms were still active during the Younger Dryas, for example, in Belgium and the Netherlands (Pissart 1963; Ploeger & Groenman van Waateringe 1964).

The accumulation of loess

Loess is a soil of the cold continental climate (Gierasimov 1962). In mid-Europe its source was the seasonally flooded plains and areas exposed to frost disintegration (Malicki 1950; Maruszczak 1968; Jersak 1973). The distance of transport of the loess dust with increased size-grades to the west was short. In mid- and east Europe loess is found in the hypsometric belt of 100–500 m above sea level (Velichko & Khalcheva 1973). Its lower limit decreases to the west and south. The distribution of loess coincides with the boundary of areas occupied by steppe communities in which the capillary upward movement of water and a dense grass cover favoured deposition and growth of the dust layer (Jersak 1973). Both the distribution and thickness of loesses varied throughout the last cold stage, reflecting the continentality of climate (figure 2). According to Velichko & Khalcheva (1973), the estimated annual rates of loess deposition were 0.07 mm for the older Pleniglacial, and 0.4 mm for the younger Pleniglacial. In southern Poland the Early Glacial loesses are thin or not apparent. The older Pleniglacial loess attains 1 m in the west and 5–7 m to the east. It is clearly gleyed, suggesting wetter conditions. The younger Pleniglacial loess is a typical subaerial deposit, locally 10–15 m thick. Associated with it may be three or four horizons of frost cracks, congelifluxion and gley (Kukla 1969; Jahn 1970; Jersak 1973) which indicate oscillations of climate and its increasing continentality. At that time, the loess belt expanded markedly from the periglacial zone to the border of basins in Bulgaria, Serbia and Italy. The general indication is of a vegetation rather similar in character over wide areas because there were no continuous forest barriers (Velichko & Khalcheva 1973). In Podolia loess deposition continued at least into times postdating a phase of climatic warming some 17000 a ago (Ivanova 1969).

Fossil soils

The sequences of fossil soils are best known from loess profiles in which the particular soil horizons point to pauses in sedimentation (figure 2). Usually there are two different soil complexes recognized, namely the interglacial–Early Glacial complex and the periglacial complex. The older one is called the Mezinski soil complex (Velichko & Morozova 1973) or the Nietulisko type (Jersak 1965). In Czechoslovakia it includes the fossil horizons PK III and PK II (Kukla 1969). The truly interglacial series comprises three horizons which overlie one another: chernozem, brown-earths-lessivé and again chernozem clearly reflecting the succession of vegetation (Jersak 1973). The PK III (uppermost part dated at 114000 a B.P.) is separated either by a hiatus, thin diluvia or loess with frost structures from the overlying soils (PK II in Czechoslovakia) of chernozem type (Mojski 1969 and others). Their two or three horizons indicate three or four oscillations of a dry temperate climate. The brown horizon with mollusc faunas *Brachybaeba triticum* and *Chondrula tridens* (Ložek 1969) is related to the development of forest communities during the Brørup Interstadial. Dating them is beyond the range of the ^{14}C method (at Vestonice >55000 a B.P., Kukla 1969).

In the lower part of the Pleniglacial loess horizons of initial chernozems or rendsinas can be seen (Kukla 1969). On the River Dniestr (Ivanova 1969) and on the Rhine (Paas 1969) datings of gley horizons suggest an age >45000–50000 a B.P. The Interpleniglacial period is clearly differentiated. It includes soils dated at 35000–25000 a B.P. which belong to the so-called Paudorf complex (Fink 1969) or Briansk (Velichko & Morozova 1973) or Komorniki (Jersak 1965) and PK I (Kukla 1969). This complex consists of one to three horizons which formed in a permafrost environment and are expressed as pseudogley or arctic brown soils showing poorly developed humus accumulation processes. In the Pannonian Basin under waning permafrost conditions this complex was represented by initial chernozems, or even brown earths (Kukla 1969). Their main horizon is dated at 30000–25000 a B.P. (at Molodova on Dniestr 29650 ± 1320 B.P. (Ivanova 1969); at Mende 29800 ± 600 B.P. (Pécsi & Hahn 1969), at Vestonice 28300 ± 300 B.P. (Kukla 1969)). These correspond in the west to the Denekamp interstadial and to the Zelzate soil 28200 ± 270 a B.P. (Paepe & Zagwijn 1972). The older clearly brown horizon may correspond to the Hengelo Interstadial. The frequently cited dates of the order of 25000 a B.P. (Velichko & Morozova 1973; Mojski 1969) suggesting a younger age of the former Paudorf interstadial are questionable and mark rather the beginning of intense loess deposition which stopped the accumulation of humus in the soil.

Lying on top of this soil complex are initial pseudogley soils, with 3 or even 4 horizons reported, for example, from the loesses at Zwierzyniec by Kraków (Sawicki 1952) and at Molodova on Dniestr. Radiocarbon determinations on the latter indicate an age of 23000–23700 a B.P. and 17100 ± 180 a B.P. (Ivanova 1969). The above horizons are related to periods of climatic warming and increased humidity. In Hungary the chernozem-like humic horizons indicate a forest–steppe community (dates of 18600 ± 150 and 17400 ± 440 (Pécsi 1970)). The sandy Allerød soils (Manikowska 1966) are also of poorly developed brown earth type showing a distinct humic horizon.

The frequently observed distinct boundaries between the buried soils and loess horizons may suggest rather sudden changes of climate in mid- and east Europe (Kukla 1972).

Slope deposits

The slope deposit sequence traced in the loess profiles is complex. At the base of each subaerial loess a solifluxion or deluvial horizon is commonly developed. In the loess-free areas the scheme is simpler. There is evidence of two distinct complexes of slope sheets comprising weathering debris and solifluxion material (for example, in the Carpathians, figure 3) associated with permafrost conditions. This fact is recorded even by devotees to the periglacial denudation cycle that comprises the ascending phase, the climax, and the descending phase (Dylik 1969*a*). Apart from a few sites (e.g. Józefów) distinct evidence for important phases of slope degradation during the Early Glacial is lacking.

The lower Pleniglacial series rests on Early Glacial peats (Sobolewska, Starkel & Środoń

1964) and attains in places a greater thickness than the upper and younger one. This may suggest that a relatively long period of cool climate (Klatka 1962; Starkel 1964) has been available for its formation. Thick layers of scree found in caves, for instance, in the environment of Kraków (Nietoperzowa Cave – Różycki 1967) and in Dobrogea (Adam Cave – Samson 1969) support this interpretation. On the border of the Warta (= Warthe) end moraines, a cryopediment, together with the correlative deposits in depressions predates the Interpleniglacial (Rotnicki 1966, 1974).

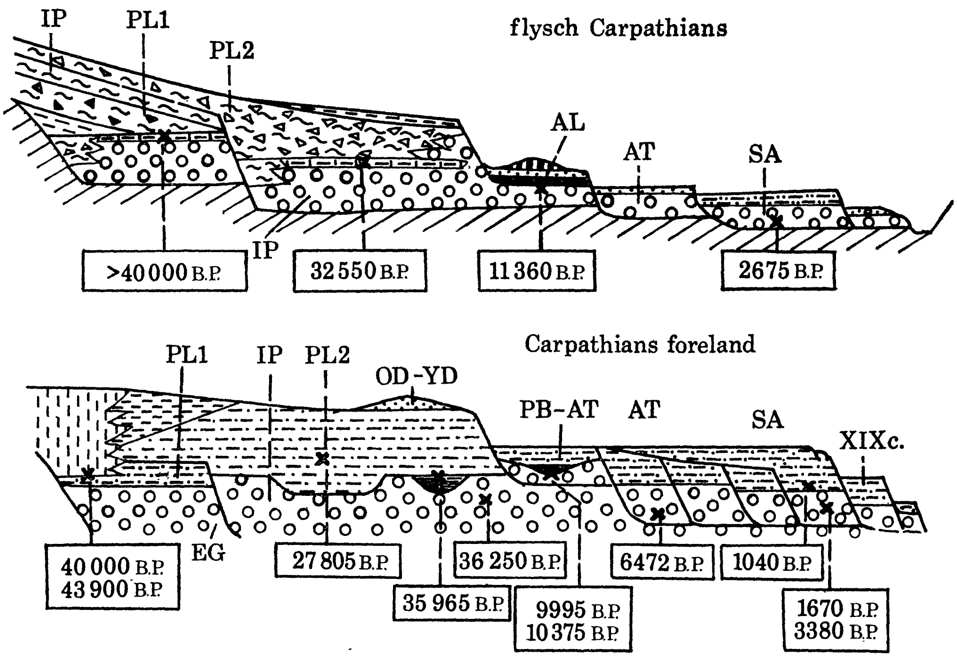

Figure 3. Terrace sequence of Vistulian and Holocene age in the Polish Carpathians and their foreland.

The Interpleniglacial phase is well characterized by a weakening of denudational processes. For example, the long section through a hill slope at Zabrodzie in the Polish flysch Carpathians shows changes from solifluxion processes to diluvial processes, and restriction of solifluxion to the steep slope sectors (Dziewański & Starkel 1967).

The younger Pleniglacial appears to have been a time of intensification of slope processes. Colluvial deposits lie on the Interpleniglacial alluvia or soils (figure 3). In the flysch Carpathians solifluxion sheets composed of clay and scree attain locally 10–12 m. At Dobra and Krościenko several scree and clay units alternating with clay units are thought to be associated to warmer oscillations, whereas the intervening clay units may record cooler conditions and shallower summer thaw (Klimaszewski 1971). Evidence may also come from flow structures in the Interpleniglacial horizon (Starkel 1969). It is likely that the scree corresponds to wetter conditions and intense frost weathering, and the creeping clayey shales, to the drier periods. The sedimentary cycle sometimes ends with the latter deposits, indicating a desiccation of climate towards the end of the Pleniglacial (Starkel 1969). During the younger Pleniglacial formation of the younger series of cave scree (Nietoperzowa Cave and others) and of the lower step of cryopediments (Rotnicki 1974) also took place. In the wetter areas to the west solifluxion processes continued into the Younger Dryas (Rohdenburg 1967). Truncated profiles and traces of aeolian processes recorded from the immediate foreland of the last ice sheet suggest that slope processes developed especially in the wetter phase of permafrost aggradation (Dylik 1969*b*).

Fluviatile deposits

In Poland and neighbouring countries the sequences of younger Pleistocene alluvia commonly represent only short spans of time. The most reasonable explanation for this seems to be the shift of stream channels and alternation of phases of either predominant deepening or aggradation (figure 3). Alluvia of the channel facies overlain by sediments which correlate with the Early Glacial phases of climatic warming suggest that a period of erosion and gravel deposition occurred at the end of the interglacial and during the early Vistulian (Dylik 1967; Mojski 1969; Sobolewska *et al.* 1964). At the same time, aggradation is believed to have occurred in the glaciated valley heads during the maximum of glaciation (Klimaszewski 1961), whereas in the valleys of the foreland of the Ardennes a phase of revival of stream activity is attributed to the coldest periods (Paepe & Zagwijn 1972).

In central Poland sediments belonging to the Brørup Interstadial are underlain by sandy deposits assigned to the earlier stadials (Mojski 1969). In the Carpathians the late Brørup sediments immediately rest on gravels (Sobolewska *et al.* 1964; Mamakowa, Mook & Środoń 1975) which perhaps represent the earlier part of the Brørup. These are to be found deposited on the higher rock-cut benches, subsequently dissected by small channels and finally buried by younger alluvia with horizons dated at 32000–40000 a B.P. (Klimaszewski 1971; Mamakowa & Starkel 1974). This suggests that the younger gravel series corresponds to the Interpleniglacial (Paudorf) revival of stream activity. Datings of the basal alluvia in the Vistula valley at Tarnobrzeg (40000 ± 2000 a B.P. – Mycielska-Dowgiałło 1967) support this interpretation. At Kępno in the Polish Lowland dates obtained from the uppermost deposits which correlate with the older cryopediment (29450 ± 1100 a B.P. – Rotnicki & Tobolski 1969) also support a phase of erosion in the Interpleniglacial and indicate the age of valley infilling as prior to that time. The present author is of the opinion that the pre-Interpleniglacial aggradation period is also represented by the upper silty-sandy units of the 'Brørup' profiles at Podgłębokie and Konin in central Poland (Mojski & Rzechowski 1969).

During the younger Pleniglacial, outside the zone of glacially disturbed sediments, aggradation was caused by melt-water floods, as indicated by silty deposits at Brzeźnica dated at 27805 ± 330 a B.P., (Mamakowa & Starkel 1974) and by the thick series of sediments in the drainage basins of the rivers Dniepr, Upper Volga and western Dvina. These predate 25000–20000 a B.P. the maximum of glaciation (Chebotarieva & Makaricheva 1974). As the deglaciation continued the braided rivers of the extraglacial zone cut wider and deeper their channels and finally changed to meandering rivers. On the dissected aggradational plains dunes began to develop. Lying on the surface of the contemporaneous erosional plains in the Carpathian valleys are biogenic deposits considered to belong to the Bølling (Koperowa 1958; Klimaszewski 1969), and gravel and sand also overlain by biogenic deposits dated at 10300–9900 a B.P. (Ralska-Jasiewiczowa & Starkel 1975).

The history of vegetation

Since the floras of the cool stadial periods are very similar to one another the stratigraphy should be based on diagrams for the varied interstadial deposits where age determination is easier. Pollen spectra may be of two types, those typical of the Early Glacial and those representing the Interpleniglacial. The distinct sequence of Eemian interglacial plant communities with the late *Carpinus* and *Picea* phase – marked by the presence of *Buxus sempervivens* and *Ilex aquifolium* and by the absence of *Fagus* – permits the separation of the Eemian from the Early Glacial periods of climatic warming. It also shows that the climate of the Eemian was warmer and more oceanic than that of the Holocene (Środoń 1972). The Early Glacial periods of climatic warming whose age given by ^{14}C dating is more than 40000–52000 a B.P. are thought

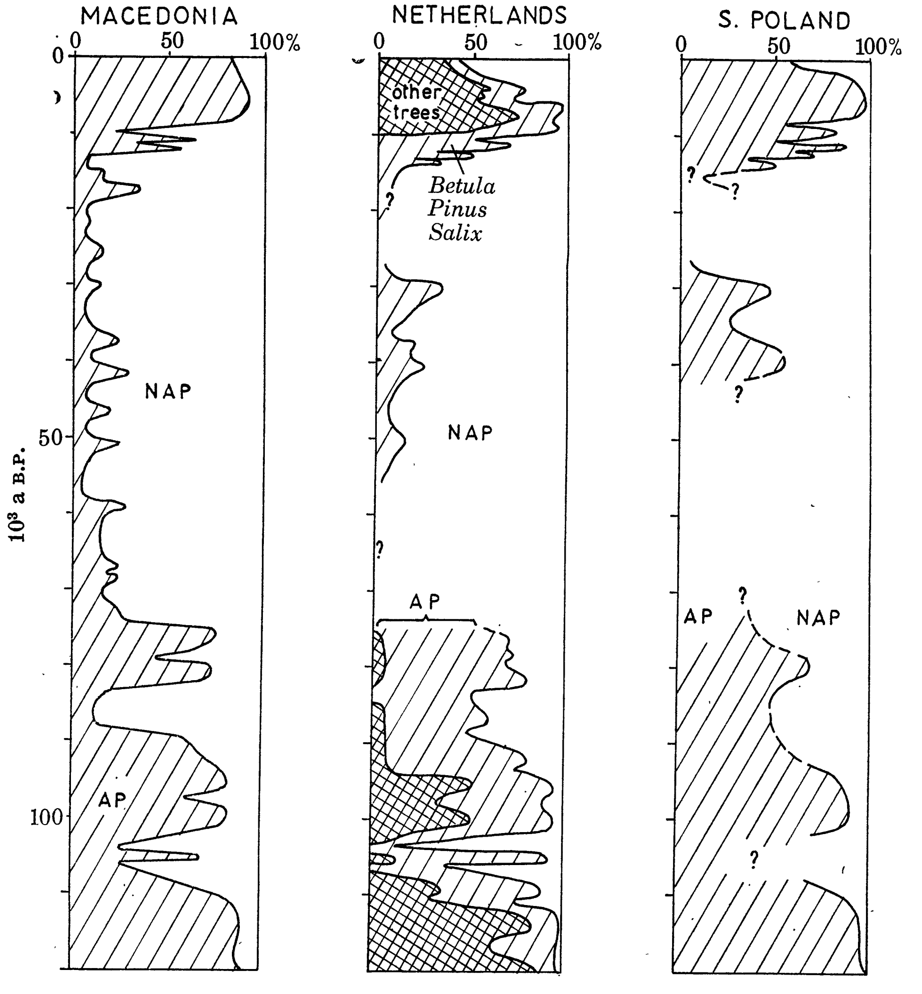

FIGURE 4. Tree and other pollen sequence during the Vistulian in Macedonia, the Netherlands (after Wijmstra 1975) and Poland (after diagrams published by Środoń 1972 and others).

to be synchronous with the Amersfoort (site Józefów, after Dylik 1968) and with the Brørup (for example, at Wadowice – Sobolewska *et al.* 1964; at Konin – Borówko-Dłużakowa 1967). These are characterized by boreal forests showing high frequencies of AP of 60–90 % (figure 4). The Brørup had an admixture of thermophilous trees (*Quercus*, *Tilia*) and of the characteristic species *Picea omoricoides* (for comparison, see Andersen 1961). The climate of the optimum was slightly cooler than that of today. In the pollen spectra for Poland the younger Odderade Interstadial (Averdieck 1967) has not been distinguished. Menke (1970) synchronized the Odderade with the Dutch Brørup and the classical Danish Brørup with the Amersfoot in the Netherlands. Pollen diagrams from Macedonia show indeed only two great interstadial phases of a warm and wet climate (Wijmstra 1969). Profiles from the Carpathians indicate that a long-term phase of climatic cooling and of solifluxion processes followed immediately a forest phase. The high frequencies of lime (> 20 %) and hornbeam (> 5 %) in the diagram from Konin may suggest the older interstadial, whereas at Wadowice and Podgłębokie (Janczyk-Kopikowa 1969) only the second interstadial of a cooler climate with spruce has been identified. In the U.S.S.R., the division into the Eemian and Early Glacial climatic oscillations remains to be clarified because of absence of well documented palaeobotanical sites.

The older Pleniglacial is represented, for example, by the uppermost silty deposits at Konin which indicate forest-tundra communities (> 52000 a B.P.), and by the basal units of the profile at Zator dated at > 40000 a B.P. (Koperowa & Środoń 1965). The presence of steppe plant pollen shows that the tundra communities have joined the steppe communities (Różycki 1967). The phases of Interpleniglacial climatic warming are also documented by the flora at Brzeźnica which has been dated at 35000–36000 a ago. The vegetation was the open forest community with *Pinus cembra* and *Larix* underlain by permafrost as indicated by injection structures (Mamakowa & Starkel 1974). Charcoals of *Larix* and *Pinus cembra* that have been found in the caves of the Kraków Upland and dated at 38160 ± 1250 a B.P. (Chmielewski 1961) appear to be synchronous with the Hengelo Interstadial. The flora tended to be dominated by woodland, as indicated by the absence of tundra faunas with *Dicrostonyx torquatus* and *Lemmus lemmus* (Kowalski 1962). The next period dated, at 32000–29000 a B.P., was relatively warm since tree clumps extended as high as 500 m.a.s.l. in the Carpathians (Środoń 1968). In the diagrams for the Interpleniglacial tree pollen appears locally at frequencies of 60–80 % (figure 4). The average temperature of the warmest month is thought to have been +12 to +13 °C, and the mean annual temperature above +2 °C. However, the presence of permafrost indicating mean annual temperatures below 0 °C has not been considered as yet. North of the South Polish Upland belt the shrub tundra was already widespread (Rotnicki & Tobolski 1969). It appears that the plant communities were similar to those of the classical profiles from the Netherlands (see van der Hammen *et al.* 1967).

The next climatic cooling in southern Poland dated at 28000 a B.P. brought about distinct vegetational changes leading to tundra communities. NAP frequencies expanded to 90 %. A similar phase of severe climate was recognized by Erd (1968) in the Democratic Republic of Germany. In east Europe high frequencies of trees (AP 20–65 %) were found to occur 30000–21000 a ago, for example, at Dunajevo on the River Lovach. The explanation may be in the still hot summers and continental conditions. After Grichuk (1973), the desiccation of climate proceeded gradually northward (see figure 7). A period of dryness started at Arapovichi – latitude 52° 30′ N – before 24200 ± 1680 a B.P., and farther north at Pokrovskaya – latitude 60° N – after 21410 ± 150 a B.P. but immediately before the advance of ice sheet. Corresponding with this dryness was the arrival of steppe elements *Ephedra distachya* and *Eurotia ceratoides*. An original map by Grichuk, based on several dated sites which document the survival of forest patches in the Russian Lowland (figure 5), shows the reconstruction of plant communities during the maximum of glaciation (20000–18000 a B.P.). The map clearly differs from the earlier maps and profiles presented by Büdel (1951), Frenzel (1960), Vigdortshik, Auslender, Znamienskaya & Dołuchanov (1972). Grichuk recognized three types of plant communities in mid- and east Europe. These were a periglacial-tundra, boreal forests (confined to enclaves in southern Europe, in the East Carpathians and the Ukrainian Upland), and periglacial-steppe. The first and last of these communities existed in a permafrost environment. A

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very narrow forest belt and the mixing of tundra and steppe are their two characteristics. The existence of refuges and the merging of tundra and steppe communities also was mentioned by Różycki (1967) and Środoń (1968). However, comparison with the present distribution of various plant communities in Yakutia and Mongolia suggests that the tundra-steppe ecosystems occurred at neighbouring sites and were closely adapted to the changing soil, thermal and moisture conditions.

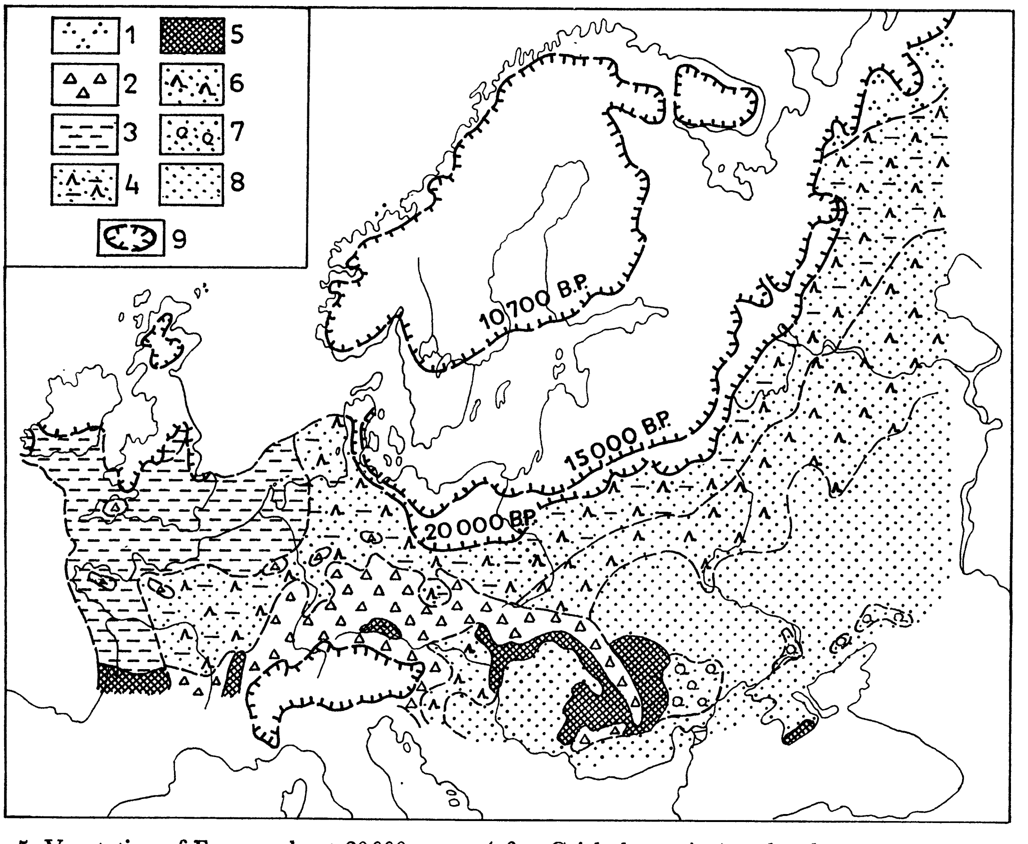

FIGURE 5. Vegetation of Europe about 20000 a B.P. (after Grichuk 1973). 1, polar desert; 2, mountain alpine vegetation; 3, tundra; 4, forest tundra; 5, birch–pine–larch open forest; 6, northern forest-steppe; 7, southern forest-steppe; 8, steppe; 9, margins of ice sheets.

Floras dating from the times of inland-ice retreat (18000–12500 a B.P.) are poorly known. South of the Carpathians there is evidence for patchy distribution of a forest-steppe (Pécsi 1970). The climate of the Mazurian (Halicki 1960) or Somino Interstadial (Chebotarieva & Makaricheva 1974) was perhaps too cool to favour growth of pine–birch forests in northern Poland. In the Soviet Union, fossil floras from the Baltic drainage basins have been dated at some 13500 a ago and indicate a severe climate. More complete profiles are available for the Oldest Dryas and the succeeding periods. For example, at Witów by Łódź (Wasylikowa 1964) tundra communities with steppe patches were replaced by birch forests in the Bølling (some 12300 a B.P.) and by pine forests in the Allerød. The initial forests thrived above permafrost (figure 6). The rapidity of both climatic warming and forest expansion during the glacial retreat is best indicated by pieces of *Pinus* and *Picea* wood contained in Allerød deposits in the environment of Helsingfors (Mölder, Valovirta & Virkkala 1957). The climatic continentality persisted in the Younger Dryas as indicated by the optimum growth of steppe elements in Poland at that time. It is not until then that the Holocene forest communities begin to develop.

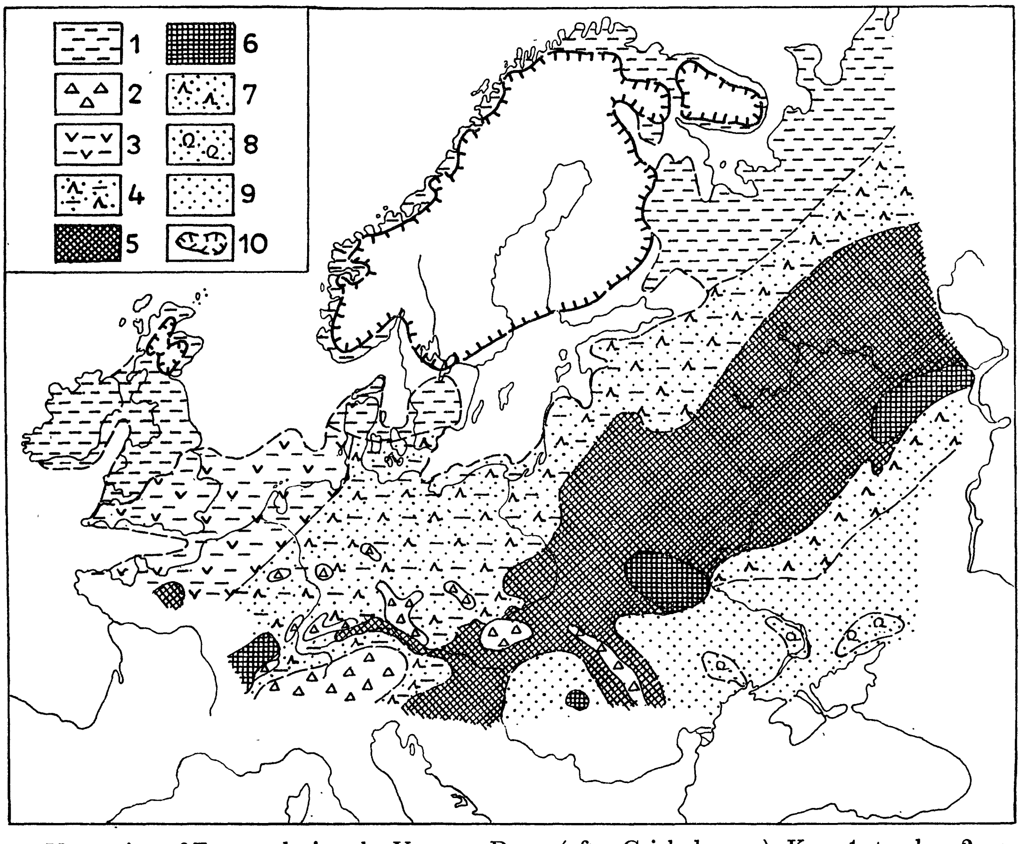

Figure 6. Vegetation of Europe during the Younger Dryas (after Grichuk 1973). Key: 1, tundra; 2, mountain alpine vegetation; 3, forest tundra with birch; 4, forest tundra (birch–pine); 5, boreal forest; 6, mixed forest; 7, northern forest-steppe; 8, southern forest-steppe; 9, steppe; 10, extent of inland ice.

The development and disappearance of the ice sheet

The size of the Scandinavian ice sheet before 20000 B.P. is partly of disputed status. The small sizes of the Early Glacial glaciers are indicated by the extent of boreal forests as far north as latitude 62° N during the Brørup (figure 7). Little is known of the ice extent during the older Pleniglacial. J. Lundquist (1971), relating the older boulder clay to this period, suggests that at that time advances brought the ice into southern Sweden. During the Interpleniglacial lowland Sweden and Finland was ice-free, since biogenic deposits have been found in the Lulea area (Fromm 1960), and marine deposits dated at 29000–26700 a B.P. occur near Göteburg (Brotzen 1961). Similar dates were obtained on the tundra deposits on the Kola Peninsula.

The time of maximum ice advance is documented by age determination on organic remains which appear under the moraines marking the maximum phase in the D.R.G. (21160 ± 800 a B.P. – Cepek 1965), and in the environments of Vologda in the U.S.S.R. (21410 ± 159 a B.P. – Vigdortshik *et al.* 1972). Without doubt, the ice advance itself is by 1000–2000 years younger than the above dates. Following Arstanov *et al.* (1971), Chebotarieva & Makaricheva (1974) record the finding of fossil floras under the moraine at Drechaluki by Vitebsk. The youngest date is some 17700 a ago. Because hazel and other pollen are present there it is difficult to accept a maximum of glaciation later than the date cited. The moraines ascribed to the oldest phases: Leszno (= Brandenburg, Bologovo) and Poznań (= Frankfurt, Jedrovo) are ice-pushed features. The deglaciation was of areal type (Roszko 1968 and others) and proceeded under extremely continental and dry conditions. The moraines of the Pomeranian (= Vepsovo) phase include frequently push moraines, indicating a different arrangement of lobes which extended from 0 to 230 km in relation to the position of the ice margin at the maximum phase (cf.

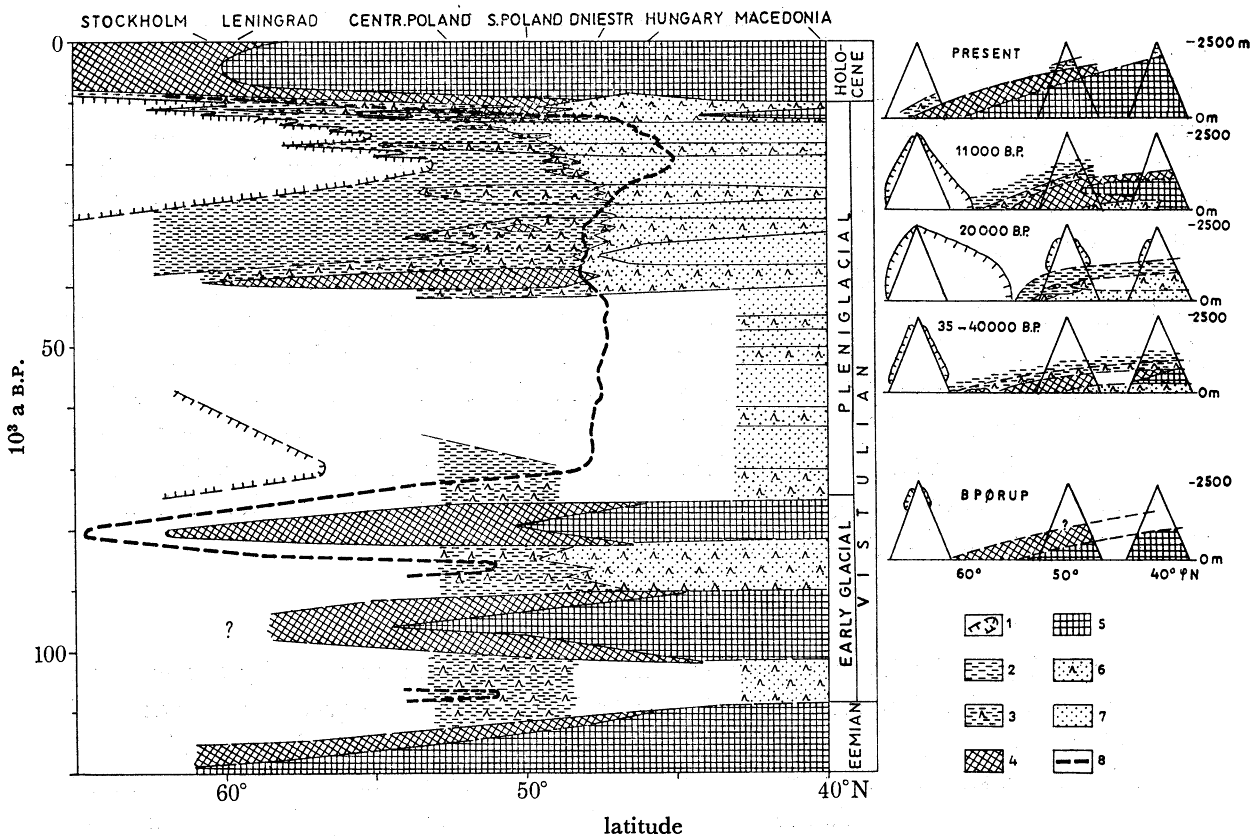

Figure 7. Changes of vegetation, permafrost and glaciation during the Vistulian in the meridional transect (Macedonia–Scandinavia) and in vertical belts. Key: 1, inland ice; 2, tundra and arctic desert; 3, forest-tundra; 4, boreal forest; 5, deciduous forest; 6, forest-steppe; 7, steppe; 8, permafrost limit.

Chebotarieva, Faustova *et al* 1973). Thus the preceding interphase was sufficiently long to allow ice readvance from a distant area of at least a few dozens of kilometres. The Pomeranian phase has been dated at some 15000 a ago. The late deglaciation was very rapid. Roszko (1968) drew attention to the fast disintegration of ice leading to the formation of dead ice blocks, and to the dissection of the ice sheet by sub- and supraglacial channels which pass into wide drainage ways (= pradoliny, Urstromtäler). To the east of the River Vistula there were formed systems of extensive ice-dammed lakes. Until the recessional Gardno (= Łuzhska) phase dated at some 13000 a B.P. (Serebryanny 1969), the amount of ice retreat from the Pomeranian end moraines was sometimes 300 km in the east, so suggesting the scarcity of supply of ice and the withdrawal of the ice centre into the mountains (J. Lundquist 1974). The melting of ice blocks was retarded by the expanding permafrost. South of the Baltic Sea coast the thawing of dead ice blocks continued in the early Holocene (cf. Starkel 1966).

The climatic and palaeogeographic changes during the last cold stage

A discussion of the characteristics of both thermal and moisture conditions can be based on the subdivision of the Vistulian stage into two main units. These are the Early Glacial (115000–75000 a B.P.) and the Pleniglacial (75000–10000 a B.P.). In reconstructing past events it is useful to remember that conclusions drawn from the nature of both characteristic plants and type deposits are far from satisfactory. The succession of plants is retarded in relation to changes of thermal conditions. The thermal tree limit set by the July temperature tended to shift with the shortening of growing season and permafrost formation. The active layer supplied the forests with water under permafrost conditions. As a consequence, the Pleniglacial plant communities do not directly reflect the amount of precipitation. In reconstructions, the areal differentiation of plant communities should be considered. This could vary with the depth to which the active layer extends and with the amount of water contained in the ground. Similarly only certain periglacial structures (ice-wedge casts, sand wedges, pingos) indicate former thermal conditions and a permafrost environment (Jahn 1970; Péwé 1966 and others). In the discontinuous permafrost zone, within which Europe was lying during a great part of the last glacial, different periglacial structures may have occurred at adjacent sites. In a permafrost environment with low precipitation the water regimen influenced the soil – morphogenetic – and biological processes. Height differences brought about the formation of a cold valley bottom zone with inversions and a warmer hill slope zone above the inversional layer. It is thus safer to apply the method of comparing results obtained in different ways. This provides the bases for reconstructions shown in figures 7 and 8.

The early glacial

The beginning of the first post-Eemian climatic cooling is documented by a chernozem horizon which developed on loess above the forest soil so suggesting the gradual northwestward spread of forest-steppe (desiccation of climate) in mid-Europe. In the mid-European Lowland boreal forest was succeeded by forest-tundra. The local occurrences of both loess and frost wedges indicate the invasion of a cool continental climate. However, this period was too short to destroy completely *Tilia*, *Carpinus* and *Corylus* which survived in refuges. Consequently the mean temperature of warmest month could not have been much below 10 °C (in west Europe it was below 9 °C, after Paepe & Zagwijn). The main depressions were of the winter temperatures and precipitation values. The increased supplies of slope-derived material and accelerated floods account for the formation of gravel sheets in the Carpathian valleys (Jahn 1968; Starkel 1964; Środoń 1968). In the Brørup the re-establishment of the optimum mixed forests in southern Poland suggests that the temperature of warmest month was 2 °C lower than that of today (July +17 °C) (see figures 7 and 8). The temperature of the warmest month of the Amersfoort is believed to have been lowered by a further 2 °C (Środoń 1972). The Brørup (Odderade?) which was followed by the Pleniglacial period of solifluxion was the second and last great warm oscillation. This may be correlated with the Elevthroupolis interstadial in Macedonia (Wijmstra 1969), and with the warming up of ocean waters some 85000–75000 a B.P. (Sancetta *et al.* 1973).

The Pleniglacial

The rapid world-wide climatic deterioration some 73000 years ago brought about the continentality of climate in mid- and east Europe. The causes suggested include the shift equatorwards of the polar and arctic fronts (McIntyre & Ruddiman 1972; Wijmstra 1975) accompanied by a displacement of the longitudinal depression 'furrow' to the west so involving an increase in strength of the southerly circulation, and the disturbance of jet-streams with the resultant rare penetration of humid air masses to the east (Lamb 1971).

The continental climatic regime was widespread in mid-east Europe embracing all latitudes as indicated by the existence of steppe and forest-steppe in Macedonia and Greece (Wijmstra 1975; Bottema 1974), and by a phase of both aeolian processes (Dylik 1969*b*) and deeply frozen ground. The mean annual temperature must have been at or below −6 °C (figures 7 and 8). The persistence of forest-tundra suggests sufficiently high temperatures of the warmest month and a different seasonal distribution of precipitation than at present. A 'moist' permafrost and gleyed loess indicates rather oceanic conditions in east Europe. Because of the presence of trees it may be suggested that the amount of precipitation was at least half the present one, and winter snowfall was especially low (in Yakutia with an average annual precipitation of 250–300 mm, snowfall is only 20 %). This favoured the deep penetration of freezing into the ground.

The interstadials which constitute the complex Interpleniglacial period of climatic warming (some 50000–29000 a B.P.) were of a short duration. In the Netherlands and Belgium these were nearly treeless periods. However, a conclusion cannot be drawn from the lack of trees that the mean temperature of warmest month was below 10 °C (Paepe & Zagwijn 1972). The waterlogging of the active layer needed higher summer temperatures which however restricted the growth of trees. Forests which existed in southern Poland during the Hengelo Interstadial indicate a deeper downward thawing and summer temperatures perhaps similar to the present ones in Yakutia (mean July temperature 16–18 °C, mean annual temperature −9 °C) (Gavrilova 1973). Velichko & Morozova (1973), reconstructing the climate of the Hengelo Interstadial on the base of both soils and permafrost structures, suggest that in mid-Europe the annual precipitation amount was 200 mm, the mean temperature of coldest month was −30 °C and there were 8 months of severe cold (figure 8). The mean annual temperature must have been −1 or −2 °C. Evidence for clearly wetter conditions is missing. The greater depth of the active layer which provided sufficient moisture controlled the supply of coarse materials into the stream channels. A perennial predominance of westerly circulation might have reduced summer temperatures and restricted tree growth. On the other hand, both the pause in loess deposition and the very low frequencies of steppe elements exclude greater aridity. However, it appears that in the west the waves of climatic warming might have increased winter snowfall nourishing valley glaciers from which ice sheets developed. It has been estimated by Manley (1975) that in the last glacial the mean annual temperature in Britain must have been lowered by 1.3 °C without a decrease in precipitation to produce the ice sheets.

The younger Pleniglacial (29000–10000 a B.P.) corresponds to the time of maximum of glaciation during the last cold stage. The initial expansion of the tundra communities in mid-Europe points to a relatively wet climate. According to Geyh & Rohde (1972), a wetter phase occurred about 23500–20400 a B.P. The masses of humid air must have periodically penetrated to the east and favoured the development of ice sheets in northwest Siberia. These reached their maximum development 24000–21000 a B.P. (Zubakov 1969). The Scandinavian ice sheet reached its maximum some 20000 years ago. In east Europe this event coincides with the spread of steppe plants to the north, and with the progressive desiccation of climate (Grichuk 1973). The presence of sand wedges in Poland is clear evidence of the formation of a 'dry' permafrost now known to occur, for example, in Mongolia. It is probable that the annual precipitation amount dropped below 200 mm. In the immediate foreland of the ice sheet there extended the arctic desert zone with descending winds (Dylik 1969*b*; Maruszczak 1968). Evidence for a wetter climate to the south is lacking. It cannot therefore be concluded that the westerly depression tracks (Liljenquist 1974 and others) moved southwards. The presence of deep ice-wedge casts in west Europe (Rohdenburg 1967; Paepe & Zagwijn 1972 and others) also indicates continentality in the west.

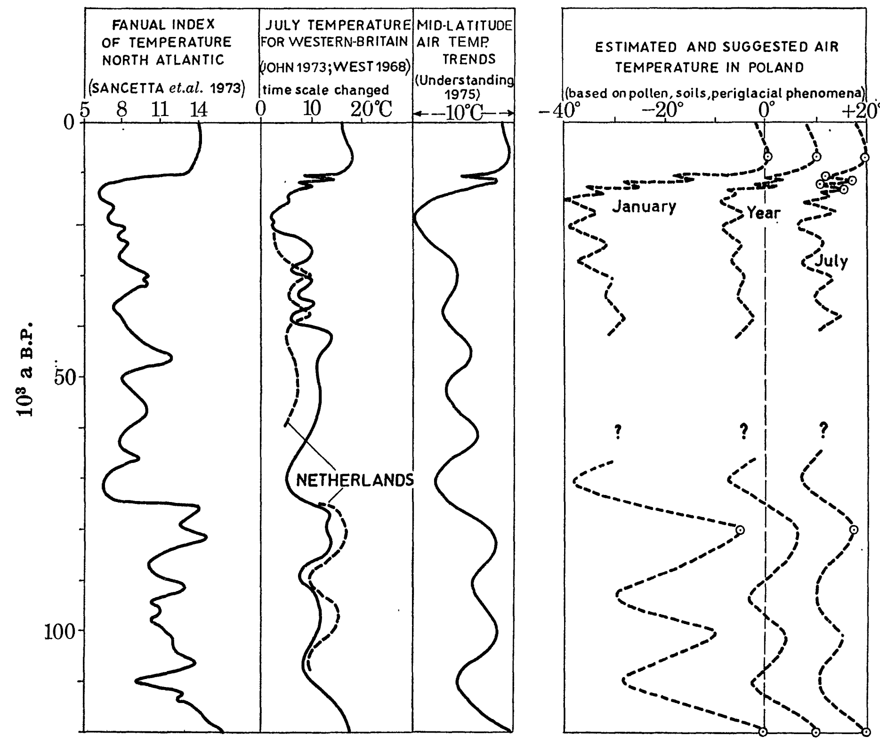

FIGURE 8. Changes of temperature during the last cold stage.

Deglaciation proceeded under extremely continental conditions with warm summers and severe winters. These were interrupted by short fluctuations of a wetter climate, with consequent ice readvances. There is evidence for the areal type of deglaciation. Concurrently permafrost expanded in areas left freshly exposed by ice retreat (also in southern Sweden – Svensson 1964). As the loess zone extended, the habitat area of tundra and forest-tundra was reduced by the steppe spread. At the end of the Pleniglacial the climate of east Siberian type has taken a turn toward the 'Mongolian' type. It seems that the ingress of humid air masses of Atlantic origin must have been short and extremely rare (it is probable that precipitation values did not exceed 200 mm). The progressive warming up of climate may have favoured the development of forest communities in the wetter phases. On the basis of herbaceous plant assemblages the mean temperature of warmest month has been established for the Bølling at +15 °C and that of the Allerød at +16 °C (Wasylikowa 1964). It appears that the mean summer temperature required for the spread of woodland in a permafrost environment must have been higher than that suggested for the present tree line. Moreover, the cause could have been the low ground temperatures and the increased instability of climate, the recurrent ingresses of humid air masses of polar-maritime origin, and of severe arctic air masses from the north which restricted plant growth, especially in spring time. The present deciduous forest zone was occupied by forest-tundra communities in the Younger Dryas and by pine forests in the Pre-Boreal (Środoń 1972). This indicates that in mid-Europe the easterly circulation system may have weakened at the beginning of a wetter climate about 8400 a B.P., with the consequent eastward spread of deciduous forest communities (Starkel 1976).

Final remarks: northwest European comparison

Analysis of the palaeogeographical phenomena recorded in deposits of the last cold stage allows the following conclusions to be made:

(1) The cold Vistulian stage includes two distinct units showing a different rhythm of climatic changes. These are the Early Glacial unit (115000–75000 a B.P.), with greater fluctuations of temperature, generally warmer and wetter, less effective in the refashioning of relief,

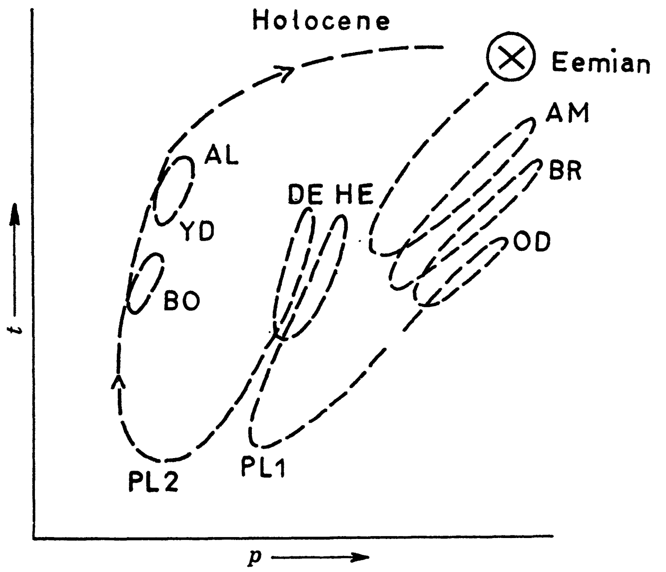

FIGURE 9. Cycle of climatic changes during the last cold stage in Europe.

and the cold Pleniglacial unit (75000–10000 a B.P.), with minor fluctuations of temperature, with a distinct continentality (great temperature amplitudes) and aridity resulting in the relief transformation and a new arrangement of plant communities (figures 7–9). Continentality manifested by changes in temperature and precipitation values has increased almost till the end of the Late-glacial. The distinct boundaries of soil horizons on loess can provide proof of sudden climatic changes of a lower order (stadials, phases).

(2) During the greater part of the last cold period plant communities differing from the present ones developed in a permafrost environment. Thus their occurrence limits the success of direct reconstructions of both palaeotemperatures and precipitation amount based on comparisons with the distribution of plant species and assemblages of today which are not underlain by permafrost. The interglacial species survived in refuge areas whose existence has already

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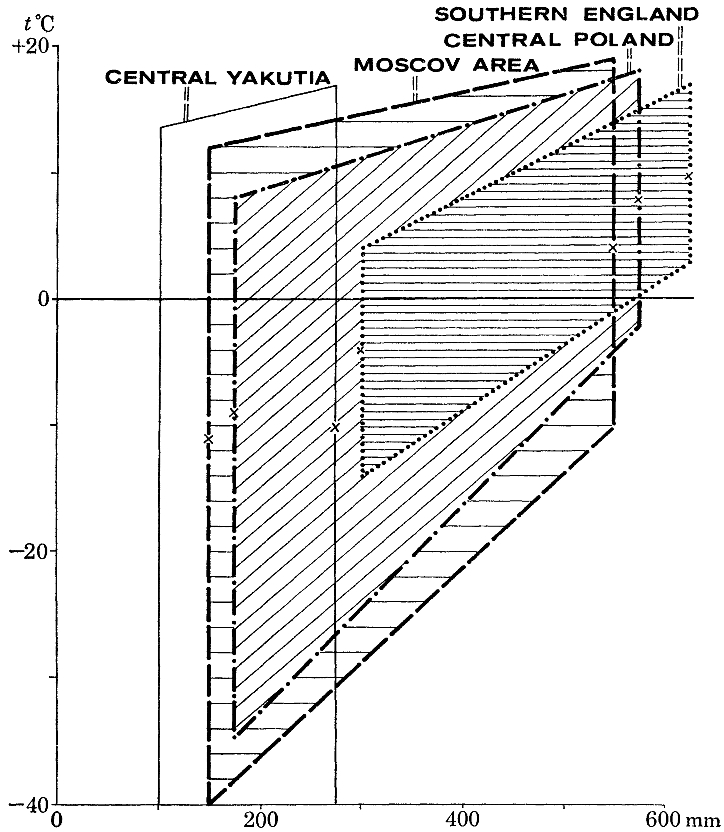

FIGURE 10. Estimated shifting toward continentality during the Last Cold Stage in different parts of Europe. Vertical lines on the right are showing temperature amplitude (I–VII) at present in the 'humid' period, on the left – during dry maximum of glaciation; ×, mean annual temperature.

been stressed early by Firbas (1949). Differences between east and west Europe were especially distinct during the Interpleniglacial. Great humidity and a thin active layer did not favour the development of forest communities then, for example, in Great Britain (West 1969; Pennington 1974).

(3) A section across east and west Europe reveals that the above areas varied by both the degrees of continentality and the annual palaeotemperature amplitudes and precipitation values. These variations are reflected in the different types of periglacial structures, in the different history of forest-tundra and forest-steppe communities in the eastern part of Europe, and in the earlier response of east Europe to the Late-glacial rise of temperature. On a diagram showing the annual temperature amplitudes in recent times and those of the Pleniglacial in Great Britain and southern Poland, in the environment of Moscow, and in central Yakutia (figure 10) not only the climatic cooling but also the increased continentality can be seen. The latter was less well expressed on the European border of the Atlantic.

(4) The formation of the ice sheet was an extreme phenomenon in the scale of the Vistulian as a whole. Ice advanced as continentality increased. This in turn was responsible for the rapid deglaciation of Europe in a slowly warming up climate. The parallel development of ice sheet and continentality may suggest both existence of a barrier, not allowing the westerly circulation system to penetrate to east Europe, and the great instability of climate in northwest Europe. A

cause may be the great fluctuations of water temperature in the Atlantic during the Pleniglacial stated off the shores of the British Isles (McIntyre 1975).

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