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Soils of the Plio-Pleistocene: Do They Distinguish Types of Interglacial? [and Discussion]

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Soils of the Plio-Pleistocene: do they distinguish types of interglacial?

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Soils form on land surfaces by the actions of physical, chemical and biological processes on the lithosphere, and are influenced by climate, parent material, relief, organisms and duration of formation. Remnants of Plio-Pleistocene soils may be buried beneath younger deposits or persist on present land surfaces. Their potential for rigorously differentiating interglacials by climatic characteristics is limited by problems of:

(i) precise dating of the beginning and end of soil-forming periods;
(ii) distinguishing characteristics attributable to climatic factors from those related to parent material, relief, etc;
(iii) calculating mathematical relations between measurable soil features and climatic variables;
(iv) diagenetic changes in buried soils;
(v) recognition and dating of relict features in unburied soils;
(vi) loss of many soils by erosion.

Some of these problems may be overcome if sequences of buried soils in periglacial loess deposits are used to compare the climates of successive interglacials in Europe and Asia. With the use of the length of interglacials derived from the oceanic record, the interglacials of the past million years are ranked according to approximate rate of soil development in loess. Two provisional equations relating soil development to time and climate are used; a linear relation probably overestimates the effect of time, and a logarithmic one seems to underestimate it. I tentatively suggest that oceanic oxygen-isotope stage 5e was warmer and wetter than the Holocene, stages 7 and 9 were cooler and drier than 5e, and 13–23 were generally warmer and wetter than 1–11.

1. Introduction

Soils are formed on land surfaces by processes dependent on the proximity of the uppermost layers of the earth's crust to the atmosphere and biosphere. Soil-forming processes, such as incorporation of humus, oxidative weathering of rock-forming minerals, leaching of soluble weathering products, and the downwashing (illuviation) of fine soil particles, affect various thicknesses of the crust, usually from a few centimetres to a few metres, although locally tens or even hundreds of metres. Generally, upper layers are modified more strongly and in different ways from deeper ones, thus producing a sequence of 'horizons' roughly parallel to the land surface but often disconformable with rock structures such as inclined bedding.

Simple climatic factors, such as mean annual air temperature, mean annual precipitation, and seasonal variations in temperature and rainfall, influence both inorganic soil processes (e.g. weathering and leaching) and the type and amounts of plant and animal life associated with the land surface, and also the rates and ways in which these organic materials decay after death. As the glacial–interglacial cycles of the past three million years are thought to have been reflected essentially in the same simple climatic factors, we might reasonably expect the sequence of soil horizons (or 'soil profile') at any place to preserve a history of climatic

38-2

fluctuations similar to that recorded, for example, in deep oceanic sediments. However, this is rarely true, because land surfaces are hardly ever stable for very long; they are subject to episodes of erosion by streams, glaciers, the sea, the wind, or downslope mass movement, and are also buried periodically beneath new deposits of various types. As a result, the history of soil development over periods longer than the past 10^4 years is rarely complete; the most recent episodes are quite widely recorded, but earlier evidence is either lost by erosion or preserved partly and often in diagenetically modified form as buried soils.

In addition to the direct effects of climate, soils are also influenced by four other factors or groups of factors (Jenny 1941). These are the time over which they have been forming, the physical and chemical properties of the geological materials in which they have formed, the effects of relief on water flow and profile drainage, and the activities of organisms, including plants growing in the soil, animals living in and on it, and man. The many aspects of these five soil-forming factors, and also the numerous ways in which they interact, has led to the intense lateral variation observed in the present soil cover of the earth's surface. In past periods this variation would have been equally intense, but the patterns of variation in any region were probably different from the present pattern, because it is likely that one or more of the soil-forming factors was different. On the worldwide scale climate is probably the most important soil-forming factor, but the other four are just as important in determining regional, national and local patterns of soil variation (Harris 1968).

The oceanic isotopic record provides little evidence for climatic differences between successive warm stages over the past two million years or so (Shackleton & Opdyke 1976), but this could be because of the temperature-buffering effect of the oceans and the limited precision of the isotopic determinations themselves. If we accept that the main climatic fluctuations of the Quaternary were controlled by the Milankovitch orbital cycles of 100, 41 and 23 ka duration (Hays *et al.* 1976), we might expect interglacials of different lengths and climatic intensities, because superimposition of the three harmonic cycles would have produced non-harmonic cycles of variable wavelength and amplitude. Such climatic variation is indeed suggested by the foraminiferal assemblages in north Atlantic cores, which indicate that subtropical water penetrated further north in some warm stages than in others (Ruddiman & McIntyre 1976), and by differences between local interglacial assemblages of the more climatically sensitive fossil groups, such as the Coleoptera (Coope 1977), although it is less well expressed in the pollen records from long sequences such as the Macedonian peat bogs (Van der Hammen *et al.* 1971). The extent to which it can be discerned in the palaeopedological record has yet to be examined in full.

To obtain reliable palaeoclimatic evidence from soils (whether buried or remaining at the surface), it is necessary first to disentangle climatic effects from those of time, parent material, relief and organisms, and second to date the features attributable to climate. Before discussing the pedological evidence for palaeoclimatic differences between interglacials, it is necessary to examine critically the ways in which these two distinct problems are approached.

2. Climofunctions

If we accept the general hypothesis (Jenny 1941, 1961) that soil (S) or a specific soil property (s) is a product of the five factors: climate (cl), organisms (o), relief (r), parent material (p) and time (t), it follows that we can measure the effects of any one factor only if the remainder are constant or if any variations in them can be shown to have negligible effects. A climofunction is a specific mathematical solution of the general relation

$$S \text{ or } s = f(cl)_{o,r,p,t}$$

and may be expressed either graphically or as a quantitative function. However, there are many difficulties in solving such functions, including (i) many of the factors are not completely independent variables (e.g. climatic factors are influenced by elevation, slope aspect and other relief features), (ii) some factors cannot readily be quantified, and (iii) the constancy of some factors is difficult to establish.

As an example, consider the climofunction of the effect of mean annual rainfall on the depth to which calcium carbonate has been removed by acid dissolution. This would be determined by measuring the depth to carbonate in several profiles, each situated in areas where the rainfall has been consistently different but other climatic factors have been the same, and each developed for the same period of time, in parent materials of similar composition, and occurring in similar geomorphological situations, where the vegetation history and effects of animals (including man) also have all been the same. Some of these prerequisites may be met by simple field observation (e.g. similarity of geomorphological situation) or specialist studies (e.g. palynological studies of nearby lake deposits or the profiles themselves to establish similarity of vegetation history), and others (e.g. similarity of non-human animal influences) can perhaps be assumed. But other preconditions are very difficult to establish. In particular, many soil-forming processes, including decalcification, are so slow that their effects are measurable only over long periods (10^3–10^5 years), and during this time the climatic factor of interest (annual rainfall) is likely to have varied considerably at each site, certainly more than is indicated by recent meteorological records. This may not matter if all the profiles are from the same region, as they would probably have experienced similar climatic variations, and the present rainfall differences between sites could well have persisted while the various profiles developed. But if the selected profile sites are from different regions, it is less likely that their present rainfall differences have persisted for long; consequently they are unlikely to yield a reliable climofunction.

Another problem in evaluating climofunctions is establishing similarity of time over which the soil profiles have formed. The age of a soil cannot be measured directly by any known method, and must be estimated from two main lines of evidence.

(a) *Stratigraphic evidence*

A soil must obviously have begun to form some time after deposition of its parent material, but this may not help very much in dating if (as is very common) the soil has formed on an erosion surface truncating the parent material. It is then more useful to know the age of the erosion surface, but this can be determined only by tracing the surface laterally as far as possible and applying the rule that it is younger than the youngest deposit it truncates. With a soil on the present land surface the time of soil development extends from its initiation to the present day, although there could have been an intervening period when pedogenesis temporarily ceased because the soil was buried and subsequently exhumed. In a buried soil the period of pedogenesis was terminated by burial, and this is indicated by the age of the oldest deposit overlying the soil.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

(*b*) *Internal evidence*

Features developed within the profile can indicate how long it took to form, providing we know the rates at which the processes responsible for those features occurred. Such rates, known as chronofunctions, are derived in a similar fashion to climofunctions, that is they are mathematical solutions of the general relation

$$S \text{ or } s = f(t)_{cl,\,o,\,r,\,p}.$$

They are determined from soil chronosequences (Vreeken 1975), which are sets of profiles whose individual members differ in age but have similar climatic histories, similar parent materials, similar geomorphological situations, and similar histories of vegetation, animal and human influences. Establishing soil age from chronosequences is obviously subject to many of the same constraints as our original objective, that of determining an accurate climofunction. Many chronofunctions are not straight-line relations; instead, the soil property changes rapidly in the early stages of soil development, then later slows down and eventually reaches a 'steady state' with little or no further change over long periods.

Because of the difficulties involved in calculating how long individual profiles have taken to form and in establishing similarities of other soil-forming factors, we have as yet few reliable and generally valid climofunctions (Yaalon 1975). Many relate to limited regions, often in the tropics, and are based on assumptions or imperfect assessments of other (non-climatic) soil-forming factors. Another problem is that most are for soil properties which are easily modified and therefore unlikely to survive unaltered during burial or exposure to climatic change (% organic C, %N, % base saturation, pH, cation exchange capacity). However, Bockheim (1980) showed that useful qualitative information concerning climatic effects on soil properties can be obtained by comparing chronofunctions for the same property in two or more chronosequences from areas of markedly different climate. Also, we may be able to infer climatic effects by comparing topofunctions derived from toposequences (soils differentiated only by geomorphic situation) in regions of different macroclimate (Yaalon 1975).

3. Dating of soil features

The dating of soil profiles and of individual features within them (e.g. accumulations of illuvial clay) involves two distinct problems: (i) calculating how long the profile or feature took to form (as discussed above in § 2), and (ii) deciding when that period of formation was. The stratigraphic evidence outlined above provides some information for both (i) and (ii) although it is often rather imprecise. In contrast, internal evidence on the extent of soil development provides information only on the length of the soil-forming period, and this can be related to years BP for the beginning of soil development only if the soil is still at the surface, or if the date of burial of a buried soil is known.

Several methods are available for approximate dating of buried interglacial soils, but they usually provide some date during the period of soil development rather than the dates when soil development started and burial occurred. These methods include thermoluminescence studies (Wintle *et al.* 1984), palaeomagnetic measurements (Heller & Liu Tungsheng 1984), amino acid analysis (Limmer & Wilson 1980), U–Th disequilibrium relations of carbonate concretions (Ku *et al.* 1979) and palaeontological studies.

Where sequences of buried soils are preserved in thick Plio-Pleistocene loess accumulations, as in eastern Europe (Kukla 1977), Soviet Central Asia (Dodonov 1979) and China (Heller & Liu Tungsheng 1982), fairly precise dates for older Pleistocene and even Pliocene soils have been inferred by matching the sequences to the dated oceanic isotope curves, assuming that loess was deposited in cold stages and that the soils formed in warm stages. As Pye (1984) pointed out, the correlation between these long loess–soil sequences in Europe and Asia is far from perfect, and this does cast doubt upon some of the inferred dates. The assumptions that the Plio-Pleistocene history of these areas was simply alternating loess deposition and pedogenesis, and that the climatic cycles recognized in deep sea sediments had similar effects throughout the world, are probably too simple to provide a completely reliable means of dating these older buried soils. But when combined with palaeontological and palaeomagnetic evidence, they have given a much better indication of the age of older Plio-Pleistocene soils than any other method.

Some buried soils may represent single major climatic episodes, such as interglacials, but many could have formed over longer periods because it is possible for parts of some land surfaces to remain stable through major climatic changes without suffering erosion or burial beneath fresh sediment. In soils on the present land surface, features inherited from earlier periods, when soil-forming conditions were different, are termed relict features, and the soils, relict soils. Buried soils showing a similarly complex history are usually termed polycyclic or composite soils. The most useful technique for identifying polycyclic soils and dating pedological features relative to one another is micromorphology. Thin sections show, for example, the relations between successive episodes of clay illuviation, iron and manganese mobilization, soil faunal activities and disruption by frost action, so that a history of climatic changes during soil development can be reconstructed (Kemp 1985*a*). Where two or more soils, perhaps partly truncated, are separated by thin sediment layers showing pedogenetic alteration, the resulting profile is termed a compound soil or pedocomplex.

4. Effects of climate on soil properties

Although there are few rigorously evaluated soil–climate relations (climofunctions), much is known in qualitative and sometimes semi-quantitative terms about the influence of rainfall and temperature on some soil characteristics. However, problems arise in the application of these generalized relations to the reconstruction of past climates because of the modifications that many soil features undergo on burial (diagenesis) or when soil-forming conditions change. Some features are more likely to persist than others, but much depends on the conditions to which they are subjected. For example, the humus in a buried soil is likely to be oxidized if it is above the groundwater table and the overlying deposits are permeable, but it may be preserved indefinitely in anaerobic conditions below the groundwater table.

The following climate-related soil features are especially relevant to the various types of interglacial soil found in loess sequences, which will be discussed in §§ 5 and 6.

Organic matter content

The amount of total organic matter in soils is determined by the balance between input and decomposition rates: thus soils of hot and cold deserts have little or no organic matter, because there are few plants and input is very small; tropical forest soils usually have fairly small amounts because, although input is rapid, so also is decomposition; the most organic soils are those of humid temperate regions, where high rainfall and weak evapotranspiration encourage lush vegetation but the low temperature prevents rapid decomposition. The distribution of organic matter with depth can suggest general climatic conditions because it is often affected by vegetation type: under forest the organic matter decreases rapidly with depth below a thin humic horizon, but under grassland organic matter is fairly evenly distributed over a greater thickness of soil, as in chernozems.

Clay content and type

In areas where there is little or no mineral weathering because of low rainfall or low temperatures, the clay (equivalent particle diameter less than 2 μm) content of soils derived from most parent materials is not increased, even over long periods of soil development. In contrast, the clay content of soils in warmer, wetter regions is often greatly increased by weathering, but it is very difficult to infer climatic conditions from a simple measure of clay content. This is because many soils are derived from clay-rich parent materials, and in subsurface horizons the clay content may be increased by physical illuviation from overlying horizons, although changes of clay content with depth may also be inherited from an inhomogeneous, stratified parent material. Clay produced by weathering must therefore be distinguished in each horizon from clay inherited from the parent material, a problem that usually involves careful quantitative granulometric, mineralogical, geochemical and micromorphological comparisons between horizons.

Clay in subsurface horizons which has been illuviated from above can usually be recognized in thin section because it forms birefringent coats (argillans) on sand grains or the walls of channels. The process of clay illuviation, typical of parabraunerde, is favoured by a soil pH of 4.5–6.5 (or higher if associated with much exchangeable sodium), small amounts of cementing and flocculating agents (carbonates, humus, sesquioxides, exchangeable Al, Mg, Ca), a system of fissures and other voids such as those formed by dissolution of limestone clasts, and a seasonal rainfall distribution (McKeague 1983). It is associated with woodland rather than open vegetation, but is usually a slow process, so large amounts of redeposited clay imply a long period (several thousand years or more) of soil development in cool or temperate, humid conditions. In mid- and high- latitude regions this implies that argillans formed in interglacials rather than in other Quaternary periods.

In clay-rich soils, birefringent aggregates of oriented clay within the soil matrix are also formed by seasonal shrink–swell processes. Lehm soil microfabrics are characterized by clayey matrices strongly reorganized in this way and by physical re-incorporation of clay from illuvial accumulations. The distinction between these two types of oriented clay in the matrix is often difficult. However, in soils formed in loess or other sediments originally containing little clay, a lehm-type fabric implies considerable clay enrichment by prolonged interglacial weathering and illuviation.

The main climatic factor influencing clay-mineral transformations is the excess of rainfall over evapotranspiration, which determines the extent of leaching. Increased leaching results mainly in the loss of bases and silica, with consequent accumulation of alumina and iron oxides. The main effect of increasing temperature is to increase the rate at which a particular mineral type is formed, and thus its abundance in profiles of a particular age. In the subpercolative soils of semi-arid regions, where evapotranspiration exceeds rainfall for much of the year so that there is little or no leaching of silica or bases, the clay fractions are progressively enriched in 2:1 minerals (Brown 1984), such as smectite. In the percolative soils of more humid regions without a marked dry season, 1:1 minerals, such as kaolinite and halloysite, are formed in preference to 2:1 minerals. Further leaching of silica, especially in the humid tropics, results in an excess of residual alumina, which forms gibbsite. Impeded drainage, even in hot humid regions, causes retention of silica and bases in the profile, and thus favours formation of smectite.

Soil colour

Soil colours are either inherited from the parent material, or result from soil-forming processes, such as incorporation of humus and oxidation or reduction of inherited or newly formed iron compounds. Any red or brown colours in subsoil horizons not derived from similarly coloured parent materials usually indicate oxidation of iron compounds in a fairly warm climate. Brown colours (Munsell hue 10YR) often result from crystallization of goethite from the amorphous iron released by biodegradation of iron–organic complexes (Schwertmann *et al.* 1974) and are typical of brown earths formed in cool humid regions with little seasonal variation of climate. Redder colours (hues of 7.5YR to 10R) often correlate with increasing abundance of haematite (Kemp 1985*b*; Barron & Torrent 1986), formation of which requires neutral conditions, little organic matter, and a strongly seasonal climate with hot, dry summers (Guillet & Souchier 1982). They occur in some parabraunerde, the haematite often being illuviated with silicate clays to form red argillans in subsoil horizons, and also in soils with lehm-type fabrics (rotlehm).

In temperate regions, such as the northern U.S.A. and northwestern Europe, reddened soils occur mainly on pre-Eemian deposits or land surfaces, whereas younger soils, formed since the Eemian, are usually brown. This observation suggests that in these regions reddening occurred in the Eemian and some earlier interglacials, but rarely in the Holocene. Either the Holocene was too short or its climate in the northern U.S.A. and northwestern Europe did not fully favour haematite formation. Reddening in undoubted Holocene soils (those formed on late Weichselian or younger sediments) is known from Israel (Dan *et al.* 1968), Morocco (Sabelberg 1977), southern France (Bresson 1974), southern Spain (Torrent *et al.* 1980), the alpine foreland in southern Germany (Schwertmann *et al.* 1982), and a few isolated sites in western Britain (Clayden 1977), but many of these soils are sandy or gravelly, and probably have a warmer pedoclimate than finer, more water-retentive soils. In south Germany, soils on silty or clayey Würm moraines adjacent to the reddened soils on glaciofluvial gravels have yellowish-brown subsurface horizons, and the redness of the coarser soils increases westwards with a decrease in mean annual precipitation from 1200 to 500 mm and an increase in mean annual temperature from 7 to 11 °C. This suggests (i) that there are at least two constraints on reddening, climate and parent material, both of which affect the pedoclimate; and (ii) that during the Holocene the summers in most of northwest Europe were just too cool and wet for reddening to occur in most water-retentive soils.

In the American system of soil classification (Soil Survey Staff 1975), unburied interglacial soils with reddening as a presumed relict feature are separated as various 'Pale-' great soil groups, and in the system used in England and Wales (Avery 1980) they are in 'paleo-argillic' subgroups. Many paleo-argillic soils also have a lehm microfabric. Pale-great groups and paleo-argillic subgroups both include soils with a wide range of particle-size distribution, so in Eemian and some other interglacials in these countries there was less parent material constraint on reddening than during the Holocene. This observation suggests that the summer climate during those interglacials was warmer and drier than it was in the Holocene. The pedological contrast may have been accentuated by soil development over longer periods in the interglacials than in the Holocene, but it is unlikely that time alone accounts for the much more widespread occurrence of reddened soils in interglacials.

Carbonate content

The amounts of calcium carbonate in soils depend on the original composition of the soil parent material, any continuing additions of carbonate from aeolian deposition or rainfall, and the balance between leaching losses and reprecipitation from the soil solution. Reprecipitation of carbonate occurs in the still calcareous subsoil beneath decalcified horizons, or nearer to the surface in soils subject to a persistent soil-moisture deficit in a prolonged dry season. However, secondary carbonate deposited close to the surface in a dry season may be redissolved and leached downwards in the wet season if the rainfall exceeds the field capacity of the soil for any significant period. In seasonally dry climates the depth below the soil surface at which carbonate first appears therefore reaches an equilibrium determined by the ratio between the mean number of days per year when soil-moisture deficit exceeds a certain value and the mean days when the soil received rainfall in excess of that required to maintain field capacity. As these periods depend on the type and density of the vegetation cover and on soil particle-size distribution, structure and porosity, as well as on the seasonal distribution of rainfall and air temperature, calculation of palaeoclimatic variables from the distribution of calcium carbonate in a soil profile is fairly complex. However, it is clear that the presence of secondary carbonate close to the soil surface results from low annual rainfall; the carbonate concretions (loess 'dolls') often distributed throughout beds of loess probably originated in this way, showing that even as it was being deposited loess was often subject to arid-region pedogenesis.

5. Important sequences of Quaternary soils in loess

Although precise dating and palaeoclimatic interpretation of soils formed in past Quaternary periods presents considerable problems, we can avoid the complications of different parent materials and geographical variations of climate within each interglacial by comparing buried interglacial soils formed within limited areas in which only loess accumulated for much of the Quaternary. In some periglacial regions, loess was deposited throughout most of each cold stage, because extensive glacial grinding or frost shattering of rock was necessary to produce large quantities of silt. However, silt production decreased in the intervening warm stages, and soils developed in areas where the loess surface remained stable. Although much loess must have been eroded from these regions, sequences of numerous loess-soil cycles were preserved in suitable sediment traps, such as subsiding basins or the sides of valleys above the level of later fluvial activity. Terrestrial gastropods and sporadic pollen preserved in these sequences show that the loess accumulated in cold, dry conditions, and that the soils usually represent periods of warm, humid climate with interglacial forest development (Kukla 1977; Lazarenko *et al.* 1981; Liu Tungsheng *et al.* 1982). Another advantage of using soils buried within loess sequences for palaeoclimatic distinction between interglacials is that deposition of loess usually results in a level land surface, so slope, the main component of the relief factor (r) in pedogenesis, is virtually uniform in space and time.

The main sites at which Quaternary loess successions with multiple buried soils have been

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

studied are in central and eastern Europe, Germany, southern Ukraine, northern China and Soviet Central Asia. Palaeomagnetic and other datings of the soils and intervening loess layers in these areas often allow fairly clear correlations to be drawn with the worldwide sequence of oceanic oxygen-isotope stages, but only for about the past million years. Earlier parts of the terrestrial successions are often less complete, and earlier oceanic stages are less clearly defined.

In other regions, such as northwestern Europe and North America, loess deposition was apparently much more sporadic and often confined to later Quaternary cold stages, so that only a few of the later warm stages are represented by buried soils developed in loess. Buried soils developed in other parent materials (tills, aeolian sands, river-terrace gravels, volcanic ashes, colluvial and gelifluction deposits) are also common in these regions, and some of them even form fairly continuous sequences (Rohdenburg & Sabelberg 1973). However, most are less closely related to interglacial stages than the soils in more continuous successions of periglacial loess, because deposition in the intervening episodes was not as closely controlled by cold conditions. Even glacial deposits such as tills were deposited by relatively short-lived ice advances within long cold stages; consequently a soil between the tills of two different cold stages could have formed over a much longer time interval than just the warm interglacial between the cold stages. The same problem exists with soils developed in the 'hot' loess deposits associated with hot deserts. This type of loess is not dependent on frost or glaciations; the silt it contains was probably formed mainly by salt-weathering (Goudie *et al.* 1979) and was then concentrated by short-lived fluvial transportation followed by wind action. Consequently any buried soils within it represent wetter periods, which may or may not coincide with the warm stages of the oceanic oxygen isotope record; a good example is provided by the Netivot section in Israel, which contains six similar semi-arid calcareous soils dating from various times over the past 130 ka approximately (Bruins & Yaalon 1979).

I shall examine the palaeopedological evidence for interglacial climates from five European and Asian areas of semi-continuous periglacial loess deposition and three areas in northern Europe with shorter sequences. At present it is not worth considering the evidence for interglacial climates from soils in other parent materials; Quaternary sediments apart from periglacial loess are too variable for the parent-material factor to be eliminated, relief is less closely controlled, and the soil-forming periods are often not closely related to warm stages. Interglacial soil types at each site are given numerical values 1–7 indicating increasing degree of profile development (table 1). This sequence is based on evidence from the area considered first (central and eastern Europe). Apart from the climatic variables mean annual temperature and mean annual rainfall, the main factors affecting soil type were probably time (lengths of interglacials) and organisms. During pre-Holocene warm stages, organisms were limited to vegetation and animals (i.e. man had negligible effects) and were probably strongly related to climate, so that they can be combined with climate as a single major factor in genesis of loess soils. To minimize the effects of geographic variations in this combined climate–organisms factor within each interglacial, and thus compare the combined effects of climate–organisms plus time between interglacials, average numerical soil values for each interglacial were calculated for three broad regions of loess deposition (table 3).

The effects of man during the Holocene were probably quite variable, both geographically and in terms of soil type. Deforestation and man-induced erosion could both result in some Holocene soils' having characteristics indicating cooler, drier conditions than the actual Holocene climate.

TABLE 1. TYPES OF INTERGLACIAL LOESS SOIL AT SELECTED SITES

(Numbers indicate: 1, chernozem; 2, slightly decalcified brown earth; 3, strongly decalcified brown earth; 4, parabraunerde with weak argillic horizon; 5, parabraunerde with strong argillic horizon; 6, braunlehm; 7, rotlehm.)

			central-eastern Europe								northern Europe								Asia			
Glacial cycles	European loess marklines	Oceanic warm stages	Cerveny Kopec, Czechoslovakia	Kutna Hora, Czechoslovakia	Krems, Austria	Tutrakan, Bulgaria	Costinesti, Rumania	Paks, Hungary	Dunaföldvár, Hungary	Stari Slankamen, Yugoslavia	Kährlich, Nordrhein-Westfalen	Terrace sites of the lower and middle Rhein	Bad Soden, Hessen	Buggingen, Baden-Württemberg	Heitersheim, Baden-Württemberg	Normandy, NW France	Achenheim, NE France	Eastern Poland	Lochuan, N. China	Charvak, Tashkent	Chasmanigar, Tajikistan	Ukraine
A	I	1	5	5	1	1	1	1	1	1	4	4	4	4	4	4	4	3	1	3	3	1
B	II	5e	5	5	—	3	4	4	3	3	—	4	5	4	4	5	4	4	3	4	3	4
C	III	7	5	—	—	3	—	3	3	6	—	4	—	4	4	5	4	4	3	4	3	1
D	IV	9	5	—	—	5	1	—	—	3	4	4	—	4	—	4	—	—	2	3	3	2
E	V	11	5	—	—	5	5	2	3	3	—	6	—	—	4	5	—	—	5	3	3	2
F	VI	13	7	—	—	7	7	3	—	7	4	—	—	6	—	—	—	—	6	1	—	3
G	VII	15	6	—	—	5	5	—	—	7	4	.	5	3	—	—	—	—	6	4	3	4
H	VIII	17	6	—	—	7	7	3	—	—	—	—	5	4	—	7	—	—	3	2	—	3
I	IX	19	6	—	5	7	7	3	4	—	—	—	4	—	—	7	—	—	5	3	3	1
J	X	21	3	—	5	—	—	—	—	—	—	—	5	—	—	5	—	—	3	3	3	3
K	XI	23	7	—	6	—	7	3	—	7	—	7	5	—	—	—	—	—	3	2	3	2

Central and eastern Europe

For much of the Quaternary the Carpathian basin (parts of Czechoslovakia, Austria, Hungary, Yugoslavia and Rumania drained mainly by the Danube and its tributaries) experienced strong climatic fluctuations, from cold dry periglacial continental conditions with loess deposition, to warm wet Atlantic (interglacial) conditions (Kukla 1977). Different types of soil were formed under all conditions between these extremes. Many have been dated by radiocarbon, magnetostratigraphy, thermoluminescence, and the assumption that the 'marklines' of Kukla (1969), which are boundaries between (i) thick layers of loess, with gastropods of cold, dry conditions, and (ii) overlying interglacial soils or hillwash, indicating abrupt ameliorations of climate, are equivalent to the dated 'terminations' of the oceanic oxygen-isotope curve (horizons of rapidly decreasing ^{18}O content, indicating rapid warming from a glacial to an interglacial stage). The marklines delimit glacial cycles (Kukla 1970), designated A, B, C, D, etc. backwards in time; the present incomplete (Holocene) cycle (A) began with markline I *ca.* 10 ka BP. Within each glacial cycle, submarklines form boundaries between deposits and overlying soils which are less well developed than the interglacial soils immediately above the marklines.

The various soil types form the following development sequence.

cold, dry: loess steppe soils
frost gley
grassland soil with little humus — interstadial soils
grassland soil rich in humus (chernozem) (1)
slightly decalcified brown earth (2)
strongly decalcified brown earth (3)
parabraunerde with weak argillic horizon (4) — interglacial soils
parabraunerde with strong argillic horizon (5)
braunlehm (6)
warm, wet: rotlehm (7)

This has been confirmed as a weathering sequence by micromorphological and mineralogical analyses of selected Holocene and buried profiles (Bronger 1976, 1979; Bronger *et al.* 1976), and its interpretation as a climatic sequence is based on the present mean annual rainfall and temperature for nearby areas where analogous loess-derived soils have developed during the Holocene. Although this ignores other soil-forming factors, it is similar to the climatic sequence expected from the properties discussed in §4. The parabraunerde, braunlehm and rotlehm soils often contain molluscs and plant remains (Frenzel 1964; Lozek 1969; Urban 1984) indicating formation under typical interglacial forest, but the brown earths and chernozems were probably formed in areas that were too dry for forest development even in interglacials. Soil types known to have been formed during interglacials or the Holocene in what can be inferred as increasingly warm and wet conditions are designated 1–7.

Germany

Sequences of buried interglacial soils in loess are also common in central Germany, at sites such as Bad Soden (Semmel & Fromm 1976) in Hesse, Heitersheim (Bronger 1966) and

Buggingen (Bronger 1969) in Baden-Württemberg, and Kärlich (Brunnacker 1978) and other sites on terraces of the Lower Rhine in Nordrhein-Westfalen (Paas 1982, Figure 17). Palaeomagnetic dating is available for some of the sequences, such as Kärlich and Bad Soden, but the exact dating of many pre-Eemian soils is still uncertain. Also there are long gaps in some successions, and some interglacials in valley successions are represented by poorly drained soils (*Nassböden*) or alluvial soils (*Auenböden*), which cannot easily be correlated with the approximate climatic sequence of well-drained soils established in central and eastern Europe. Despite these problems, it is possible to assign many buried interglacial loess soils to oceanic warm stages and to members of the soil development sequence established in central and eastern Europe (table 1).

Most of the buried loess soils in Germany are parabraunerde, which are at least as strongly developed as nearby unburied Holocene loess soils. Some of the older interglacial soils along the Rhine valley are weak braunlehms (Bronger 1969), and in Rheinhessen rotlehms are developed in older loess with reversed magnetization (Plass *et al.* 1977).

Southern Ukraine

Veklich (1979) recognized eight loesses and eight soils (including that of the Holocene) formed in southern Ukraine within the past million years. The dates he proposed for each loess and soil do not match those of the cold and warm oceanic stages, so correlation with the dated oceanic sequence initially seems impossible, especially as fewer soil horizons have been recognized than there are known warm stages over the past million years. However, at least four of the Ukrainian soil horizons (Priluky, Kaydak, Zavadovka and Lubny) are pedocomplexes, and each probably represents two warm interglacial stages with minor deposition of loess between. This makes it easier to correlate individual soils (rather than named soil horizons) with the oceanic succession (table 2) assuming that Veklich's date for the base of the Priazovye loess (1 Ma BP) is approximately correct and that the youngest buried soil at least as strongly developed as the Holocene soil was formed in stage 5e. In at least four of the Ukrainian profiles, the youngest soil meeting this criterion (the Vitachev horizon) is a weak parabraunerde. The soils of other warm stages equate approximately with the chernozems and brown earths of central and eastern Europe. In table 1 these are assigned to oceanic warm stages according to the correlation and revised dating proposed in table 2.

Northern China

The loess plateau of northern China includes large basins with long sequences of interbedded loess and soils, forming loess *Yuans* (flat uplands), such as the *Yuan* at Lochuan. The loess layers in these basins contain herbaceous pollen, remains of dry steppe mammals, and molluscs indicating cold, dry steppe conditions, but the intervening soils contain pollen of mainly broadleaved trees and molluscs of warm, humid habitats (Liu Tungsheng *et al.* 1982). Palaeomagnetic dating of the Lochuan succession (Heller & Liu Tungsheng 1982, 1984) showed that the number of buried soils above the Brunhes–Matuyama boundary is the same as the number of warm stages in the oceanic oxygen-isotope record.

An Zhisheng *et al.* (1982) classified the buried soils in the Lochuan section into the following climatic sequence.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

TABLE 2. PROPOSED CORRELATION OF UKRAINIAN LOESS AND SOIL HORIZONS WITH OCEANIC OXYGEN-ISOTOPE STAGES

Ukrainian horizons (Veklich 1979)	dating (ka BP) (Veklich 1979)	deposits and soil types (Veklich 1979)	oceanic oxygen–isotope stages	revised dating (ka BP) based on oceanic stages
Holocene	0–10	chernozems, grey forest soils	1	0–12
Prichernomorye	10–22	loess with a brown semi-desert soil locally	2,3,4	12–71
Dofinovka	22–30	weak chernozems, brown semi-desert soils	5a	71–88
Bug	30–50	loess	5b, c, d	88–122
Vitachev	50–60	weak parabraunerde, brown earths (often reddish)	5e	122–128
Uday	60–70	loess	6	128–186
Priluky	70–100	chernozems	7	
		—	8	186–339
		chernozems, grey–brown forest soils, chestnut brown soils	9	
Tyasmin	100–115	loess	10	339–362
Kaydak	115–175	chernozems, chestnut brown soils	11	
		—	12	362–524
		grey forest soils, podzolized brown forest and grassland soils, chernozems	13	
Dnieper	175–250	loess	14	524–565
Zavadovka	250–370	chernozems, brown forest soils, weak parabraunerde	15	
		loess	16	565–689
		chernozems, brown forest and grassland soils	17	
Tiligul	370–470	loess	18	689–726
Lubny	470–650	chernozems	19	
		—	20	726–763
		brown forest and meadow soils, chernozems	21	
Sula	650–700	loess	22	763–795
Martonosha	700–920	brown grassland soils	23	795
Priazovye	920–1000	loess	24	

cold, dry:

weak pedogenic loess (*Cathaica* assemblage)
medium pedogenic loess
pedogenic loess (*Metodontia* assemblage)
} interstadial soils

black loessial soil (1)
carbonate drab (cinnamon) soil (2)
drab (cinnamon) soil (3)
luvic drab (cinnamon) soil (5)
warm, wet:
drab (cinnamon) brown earth (6)
} interglacial soils

The black loessial soil probably corresponds approximately with the chernozem of central Europe, the drab (or cinnamon) soil with decalcified brown earth, the luvic drab soil with parabraunerde, and the drab brown earth with braunlehm. Table 1 shows the types of interglacial soil formed during oceanic warm stages 1–23 at Lochuan.

Comparisons with modern analogue soils elsewhere in China enabled Liu Tungsheng *et al.*

(1986) to propose annual average temperature and rainfall values for the Lochuan area during the various cold and warm stages over the past 900 ka. However, these palaeoclimatic inferences ignored the effect of time on development of the soils. In cold stages, when there was little or no weathering and clay illuviation, this is probably unimportant. But in the warmer and more humid interglacial climates, the length of each soil-forming episode was perhaps almost as important as temperature and rainfall in determining the type of soil produced.

Soviet Central Asia

Several long Plio-Pleistocene loess soil successions have been reported from parts of Soviet Central Asia, including southern Tajikistan (Dodonov 1979) and the Tashkent region (Lazarenko 1980; Lazarenko *et al.* 1981). In both these areas there is palaeomagnetic evidence for the age of some soil horizons; for example, the Saylyk Horizon of the Charvak section near Tashkent and Pedocomplex V at Chasmanigar in Tajikistan are both firmly related to oceanic stage 5e, and the Azadbash horizon at Charvak and Pedocomplex X at Chasmanigar are both immediately beneath the Brunhes–Matuyama boundary, so they were probably formed in oceanic stage 21. But the ages of intervening soils in these two regions are less certain, because there appear to be fewer typical interglacial soils than warm oceanic stages. This could be because of gaps in the successions, but it is more likely that some of the cooler and drier interglacials resulted in soils more like the interstadial soils of Europe, because the Holocene and most interglacial soils in Soviet Central Asia are less strongly developed than those in Europe. The most obvious point in the Charvak succession lacking an interglacial soil is between the Khandaylyk horizon (probably oceanic stage 15) and the Barrazh b horizon (probably 11); the loess between these two soils is much thicker than others at Charvak, and 1–2 'rudimentary soils' occur within it. I therefore conclude that the weakest interglacial soil at Charvak is that of oceanic stage 13.

Unfortunately the Chasmanigar section must have at least two weak interglacial soils, as there are only five typical interglacial brown earths between the Brunhes–Matuyama boundary and Pedocomplex V. It is again unlikely that there is a break in the Chasmanigar sequence, because the same number of brown earths is found at three other sites spanning this time period (Kayrubak, Lakhuti and Khonako 2). In table 1, I have provisionally allocated the soils between the Brunhes–Matuyama boundary and Pedocomplex V (all brown earths) to the interglacials which produced the strongest soils at Charvak.

Northern Europe

In Normandy the loess contains seven interglacial soils and has been divided into three formations (Lautridou *et al.* 1986). The youngest (Saint Pierre les Elbeuf) formation contains four parabraunerden (Elbeuf I–IV), the most recent of which is dated by thermoluminescence to oceanic stage 5e (Wintle *et al.* 1984). The next older (Mesnil-Esnard) formation contains two red clayey (rotlehm) soils (Mesnil-Esnard V and VI), and the oldest (Saint Prest) formation a parabraunerde (Bosc Hue VII). The sandy loess of the Saint Prest Formation has a reversed geomagnetic polarity (Biquand & Lautridou 1979), so the Bosc Hue VII soil was probably formed in oceanic stage 21 at the latest. The Mesnil-Esnard V and VI soils are tentatively correlated with the Netherlands Cromerian Interglacials and IV and II respectively (Lautridou 1977) and were therefore probably formed in oceanic stages 17 and 19. The Saint Pierre les Elbeuf formation is correlated with the Tourville formation (part of the Low Terrace

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

of the River Seine), which contains Saalian fossils. The soil Elbeuf IV is thus Holsteinian, and soils II, III and IV were probably formed in oceanic stages 7, 9 and 11 respectively (table 1).

At Achenheim in Alsace, loess equivalent to the Saint Pierre les Elbeuf formation is divided into four layers by three parabraunerde soils (Heim *et al.* 1982). From the distribution of mammal remains and Palaeolithic artefacts in this succession, it is likely that the three soils are equivalent to Elbeuf I to III of Normandy (table 1).

On the Lublin Uplands of eastern Poland, the loess overlying weathered till of the Cracovian (south Polish) Glaciation contains two buried soils of parabraunerde type (Jersak 1977). The younger (Nietulisko Soil) is correlated with the Eemian (oceanic stage 5e), and the older (Tomaszów Soil) with the Lublin interglacial. The Tomaszów Soil is developed in 'lower older loess', which is correlated with the Odra glaciation. Odranian deposits further west have been dated to oceanic stage 8 by thermoluminescence (Lindner & Grzybowski 1982), so the Tomaszów Soil probably developed during oceanic stage 7 (table 1).

6. Climatic comparison between interglacials by using loess soils

The interglacial soils developed from periglacial loess (table 1) show some consistent differences between interglacials. (i) The soil of stage 5e is always equally or more strongly developed than that of the Holocene (stage 1) in the same area; this agrees with the evidence of unburied soils in Europe and the U.S.A. (§ 4), but could result partly from the effects of man in the Holocene. (ii) Soils of stages 7 and 9 are equal to or weaker than that of 5e, except for the more strongly developed soils of stage 9 at Tutrakan and stage 7 at Stari Slankamen. (iii) Apart from Paks (Hungary), Ukraine and Charvak (Tashkent), the soil of stage 11 is always equal to or stronger than that of 5e. (iv) With a few exceptions, the soils of stages 13–23 are more strongly developed than those of the later interglacials (1–11).

In the absence of any other numerical data common to all the interglacial loess soils, the numbers 1–7 for the weathering sequence of interglacial soils in central and eastern Europe can be used in simple numerical calculations to compare soils in different regions and different interglacials. The mean soil values for each interglacial in the three main areas (table 3) are more variable in central–eastern Europe (2.0–6.2) than in either northern Europe (3.9–6.0, which is the upper part of this range) or Asia (2.0–4.3, the lower part of the central–eastern European range). This suggests that central–eastern Europe experienced very variable interglacial climates as well as large differences between cold and warm stages.

Because we have eliminated the parent material and relief factors in soil variation, the differences between mean soil values for interglacials can be attributed to climate–organisms and time (relative length of soil-forming periods). Before we can isolate the effect of climatic differences between interglacials, we should therefore allow for differences in the duration of pedogenesis in each interglacial, but this is difficult because many of the isotopic stage boundaries in oceanic cores are imprecisely dated. The 'terminations' at the beginning of each odd-numbered stage are reasonably well dated, but the ends of interglacials are much less clear. Also the good resolution for interglacial stages 5 and 7 shows that these were complexes of two or more warm episodes with colder periods between; the isotope curves for earlier warm stages are less well resolved and could also be complexes containing cold periods of unknown length.

Table 3. Relative development of loess soils during successive interglacials, and suggested climatic ranking of interglacials

oceanic warm stages	mean soil values: central–eastern Europe	northern Europe	Asia	mean soil ranking[a]	mean climatic ranking[b]
1	2.0	3.9	2.0	11	11
5e	3.9	4.3	3.5	8	5
7	4.0	4.2	2.8	9	9
9	3.5	4.0	2.5	10	10
11	3.8	5.0	3.3	6	8
13	6.2	5.0	3.3	1	1
15	5.8	4.0	4.3	3	6
17	5.8	5.3	2.7	4	4
19	5.3	5.5	3.0	4	2
21	4.0	5.0	3.0	6	7
23	6.0	6.0	2.5	1	2

[a] Ranking by mean, worldwide soil values.
[b] Mean, worldwide climatic ranking (from $Y = a + (B \log x)$ (Bockheim 1980)).

Bockheim (1980) found that the general equation which best explains the relationship between soil properties (Y), time (X, in years) and climate (b) over a range of climatic regions and parent-material types is the logarithmic function $Y = a + (b \log X)$, and for relations between three soil properties relevant to interglacial pedogenesis in loess (oxidation depth, percentage clay in the B horizon, and total profile depth) and mean annual temperature, the ratio $a:b$ was approximately 20:1 for each property. When the lengths of interglacials can be established more clearly from the oceanic isotopic record, this or a similar equation might be used to rank interglacials according to climatic characteristics (temperature and rainfall). At present it can be used only by taking the lengths of odd-numbered oceanic stages from the revised stage boundary dates of Imbrie *et al.* (1984). If these values are used for X, and the mean worldwide soil values (table 3) are used for Y, with $a:b = 20:1$ in Bockheim's equation, the climatic ranking of interglacials obtained is in fact little different from the ranking based on the unmodified mean worldwide soil values (table 3). This result suggests that, using Bockheim's equation, the differences in lengths of interglacials likely to emerge in future from better resolution of the oceanic oxygen-isotope record will probably make little difference to the climatic ranking of interglacials (table 3) based on the loess soils. However, Bockheim's equation may make too little allowance for time as a soil-forming factor, and should also be checked and refined. Present knowledge of the relations between climate, time and any simple soil properties likely to survive burial is scarcely adequate for the reconstruction of past climates from these soils. The tentative climatic ranking of interglacials given in table 3 is therefore probably the best that can be suggested from palaeopedological evidence until more quantitative information is available on processes of soil development in loess under different climatic régimes.

7. Conclusions

If we are unable to draw very definite conclusions at present from the long sequences of buried loess soils about climatic differences between interglacials, other soils are likely to be even less useful. They are more difficult to date, especially in terms of the exact length of the soil-forming period, and few deposits are lithologically as uniform as loess over large areas, so the effects of parent-material differences on soil properties must be considered. Also, other soils are less likely to form sequences representing all interglacials within areas small enough to have shown no geographic variation of climate during each interglacial.

Because of these additional problems with non-loessial soils, future work on palaeoclimatic interpretation of buried soils should concentrate on those in the thick loess sequences. But before these can be used effectively, it is important that reliable climofunctions for properties likely to survive burial are established in Holocene loess soils developed over known periods and in regions which have not suffered major climatic change during the Holocene. Suitable properties are depth of oxidation, humus incorporation and occurrence of secondary carbonate, thickness of Bt horizon and percentage of illuvial clay, amounts of sesquioxides and layer-silicate clays formed by weathering, and the extent of weathering of heavy minerals in the fine sand and coarse silt fractions.

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Discussion

R. Paepe (*Belgian Geological Survey, Brussels, Belgium*). What is done if palaeosols might be polycyclic or double-buried?

J. A. Catt. Polycyclic soils, as defined in the paper, were omitted, but some of the double-buried soils were attributed to successive interglacials.