A continent-wide framework for local and regional stratigraphies

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Chapter 3

Contemporary Middle Pleistocene terrestrial stratigraphy of Northwest and Central Europe; a complex of local stratigraphies and palaeoclimatic stages

3.1 Climatostratigraphical subdivision of the European Pleistocene terrestrial succession

3.1.1 Historical development

Since Pleistocene stratigraphical successions in terrestrial environments are largely governed by climatic fluctuations, as indicated by lithology, structural features, fossils, soils and geomorphology, inferred climate has been used in Europe for over a century[1] as the most suitable basis for the distinction and subdivision of the Pleistocene strata and time.

The first widely-used Pleistocene stratigraphical scheme in Europe was the fourfold glaciation paradigm as initially developed by Penck & Brückner (1901-1909) for the northern Alps and mainly based on morphostratigraphical criteria. Glaciofluvial outwash terraces in the Bavarian type area were related to moraines and glacial deposits of the Alpine piedmont glaciers and used as units representing the Würm, Riss, Mindel and Günz glacial stages. Morphostratigraphical criteria were also applied to the area of northern Europe subject to ice-sheet glaciation from Fennoscandia (e.g. Penck 1879, Keilhack 1896, 1926; Woldstedt 1929, 1954). The type units of the glacial stages here, named from young to old Weichsel, Saale and Elster respectively, were originally recognised from the end-moraine belts crossing the Northwest and Central European lowlands. Another classification is related to the British ice-sheet expansions where also three major glacial stages (Devensian, Wolstonian[2] and Anglian) are well represented (King 1955, West 1963).
Unlike the Alpine region, the distribution and superposition of glacial sequences and landforms in many areas in northern Europe could be stratigraphically related to marine, limnic, fluvial and organic interglacial deposits. The latter generally contain biostratigraphical information as well as biological evidence of warmer climate conditions. Following the work of Jessen and Milthers (1928) in Denmark, palynology, together with palaezoology, became an important stratigraphical tool to define interglacials s.s. (and interstadials) as principal bio- and climatostratigraphical units and to correlate these units over wide areas in northern Europe.

The main difficulties in the climatostratigraphical interpretation of the Pleistocene sequences were the relative chronology in general and the lack of objective correlation means between the glacial sequences in northern Europe and those in the Alps. Newly developed concepts, dating methods and increased data availability in the 1950s and 1960s gradually made clear that the Alpine glacial scheme[3] could not be adopted continent-wide and became a 'straitjacket'. In the absence of such an overall framework, a complex of local stratigraphies evolved in Europe. Based on the local litho- and biostratigraphical frameworks to which genetic and causal aspects have been built in during the interpretative phase, each country or state developed its own subdivision and nomenclature of the Pleistocene Series/Epoch into palaeoclimatic stages. By counting down the units from the top, each scheme involved an arbitrary subdivision into interpreted glacial stages, defined mainly from lithological and structural evidence, and intermediate interglacial stages, generally identified from biotic palaeoclimatic indicators. Although criteria for the identification and definition of the climatostratigraphical units and their boundaries differed from country to country, the approach was to use them as a basis for interregional correlation, as advocated by Van der Vlerk (1953) among others. The most comprehensive local series of cold and temperate stages, based on superposition and in particular palaeobotanical data, are from the Netherlands, as part of the southern North Sea sedimentary basin (Zagwijn 1975).

The inherent deficiency of the composite local schemes established in the formerly glaciated areas became more and more apparent when in the 1970s evidence from extraglacial areas became available. Loess/palaeosol sequences overlying river terrace deposits from Central Europe (Červený Kopec: Kukla 1970, 1975) and long pollen records of lake sediments (Tenaghi Philippon in Greece: Wijmstra 1969, Van der Wiel and Wijmstra 1976) were hardly compatible in terms of the numbers of glacials and interglacials. Also evidence from Poland (e.g. Rozycki 1978) and the Russian Plain (e.g. Velichko 1984, 1990) did not fit easily into the classical models.
Moreover, in the light of the virtually continuous record of the Quaternary climatic history from the deep-ocean sediments (first published by Shackleton & Opdyke 1973), which demonstrates at least 11 major global cycles of glaciation in the last million years, it was shown that the frequency of glacial and intervening interglacial periods was dramatically underestimated. The vast amount of information from ocean and ice-core records, the enormous advances in geochronological techniques and the re-assessment of traditional concepts in most disciplines over the last three decades brought about continual appraisals and re-evaluations of the local and regional stratigraphical schemes and terminology. An overview of regional schemes of subdivision based on interpreted climate is shown in *Figure 3.1*.

Nevertheless, terrestrial subdivision remained constrained by stratigraphical relationships and low resolution chrono-markers. To tackle the chrono- and climatostratigraphical problems onshore, further refinement was then sought in comparison and matching of the local and regional Pleistocene evidence with the ocean and ice-core chronostratigraphies. Since the inferred palaeoclimatic stages have to fit somehow with parts of the marine relative chronological sequence, many stratigraphers in the last three decades have actually proposed and compiled MIS correlation schemes (a.o. Kukla 1975 and 1977, Bowen 1978, Sibrava *et al.* 1986, De Jong 1988, Ehlers, Gibbard & Rose 1991, Ehlers 1997, Vandenberghe 2000).
Kukla (1969, 1970, 1975) was the first who convincingly matched loess/palaeosol cycles from Slovakia and Austria (*Fig. 3.2*) with the glacial cycles of the marine isotope record for which later the long loess records from China became available (Kukla 1987, Kukla and An 1989). So far eight completed loess accumulation cycles are recognised within the Brunhes normal Chron. These

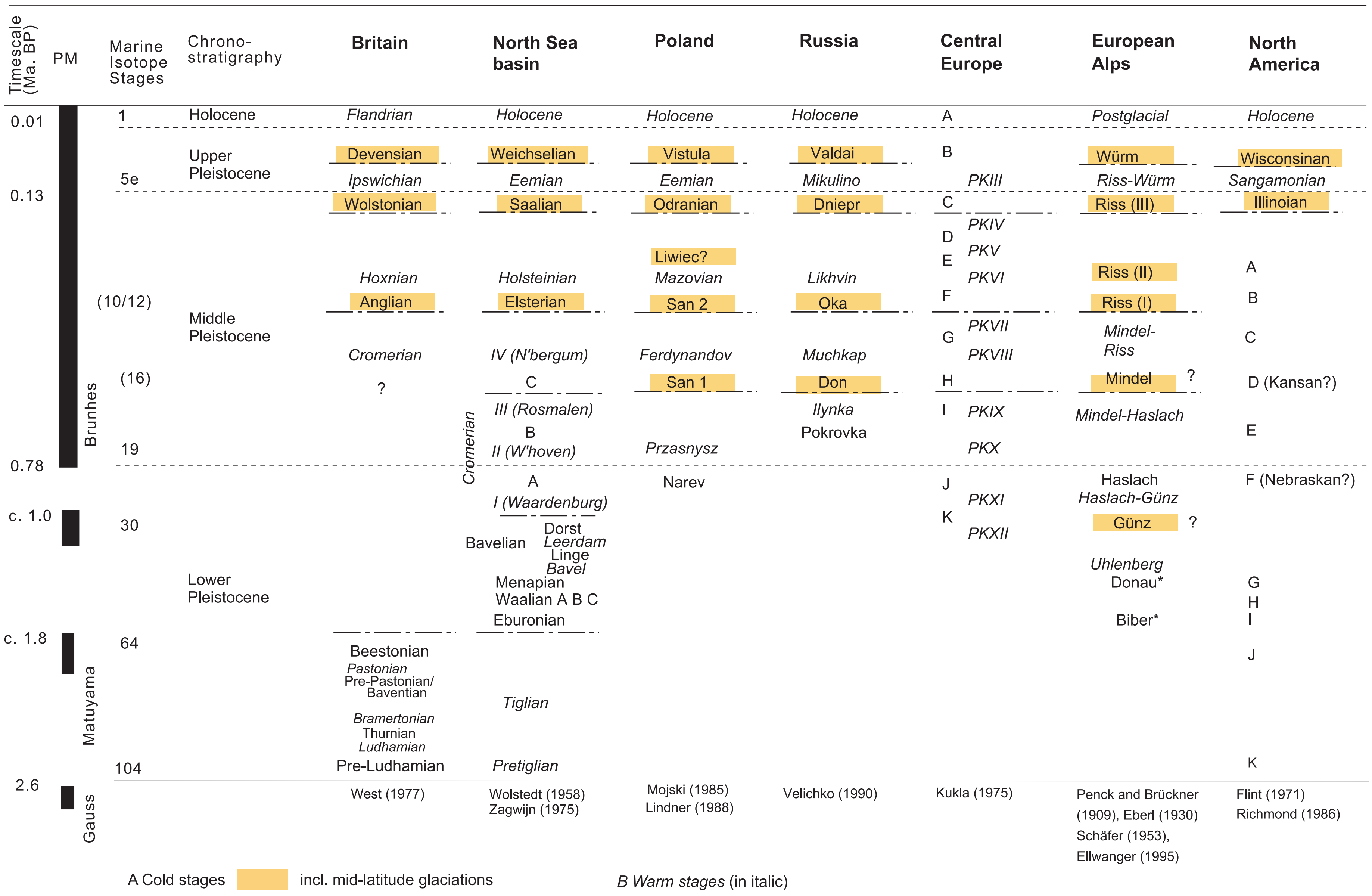


Figure 3.1 Overview of the terrestrial Quaternary (climato-)stratigraphical schemes and terminology for Europe and North America in relation to the global chronology.

cycles coincide with the 4th order glacial cycles of about 100 ka duration in the oceanic record. Up to now only the last two glacial cycles can be accurately correlated with the northern and central European mid-latitude glaciations: the Fennoscandian Weichselian / British Devensian / Alpine Würmian ice-sheet expansions and the Fennoscandian Saalian ice-sheet expansion. On a local scale, matching of pollen records from lake-core sequences with the ocean record showed similar climate-related trends for the Late Pleistocene, such as La Grande Pile and Les Echets in France (de Beaulieu and Reille 1987), Bispingen-Luhe in Germany (Field *et al.* 1994) and for the Middle Pleistocene, e.g. Tenaghi Philippon in Greece (Mommersteeg *et al*. 1995) and Lac du Bouchet/Praclaux in southern France (de Beaulieu and Reille 1995, Tzedakis *et al*. 1997, de Beaulieu *et al*. 2001).

Global matching was mostly achieved from a specific disciplinary or regional point of view and by 'counting down or up' within conventional frameworks. With the exception of Kukla's loess cycle concept, scholars did not work in a systematic way by defining unambiguous regional unit boundaries nor applied a set of large-scale correlation criteria before matching with the MIS. It is true that the use of climatic terms for the main building blocks of the glacial models met the aim of large-scale interpretation, i.e. 'the spatial reconstruction of past climate and landscapes at large (4th order) time scales'. It does, however, not satisfy for local, often temperature-related inferences from the intermediate interglacial sequences. The character, distribution and preservation of the latter are also controlled by regional (bio)geographical, geological and geotectonic variability, reflecting various short-time cyclic events.

3.1.2 Climatostratigraphical subdivision in perspective

Unfortunately, the European type localities and stratigraphical systems for at least the Early and Middle Pleistocene do not appear to be easily comparable nor synchronous (Turner 1975, Bowen 1978). In general, there is little dispute about the relative position of the dominant glacial and periglacial aeolian sedimentary units within the formerly glaciated or the non-glaciated type areas. Interregional chronological correlation of these major cold climate-driven sequences, however, is hampered by often inconsistently interpreted climatic signatures and time durations from the sequences themselves. It may also be interpreted from the scattered local, intermediate non-glacial successions, in particular from lake deposits and soil complexes comprising interbedded organic-rich horizons. Although the palaeobotanical and faunal evidence from these intermediate non-glacial deposits has particularly provided much climatic and environmental detail within the local stratigraphies, the spatial and temporal resolution of biostratigraphy and pedostratigraphy is generally limited. The fossil contents of the widely-spaced and predominantly incomplete sedimentary records show geographically-related anomalies. Moreover, most fossil groups lack substantial evolutionary change (with the exception of voles). And despite the migrations over long distances, there are similarities in species assemblages during subsequent climate stages which pose bio-correlative problems. Thus, for much of the Middle Pleistocene there are too many uncertainties for correlations to rely on palynology, pedology or other disciplines alone (*cf.* Turner 1996).

The main reasons for the unsatisfactory way in which the climatostratigraphical subdivision of the European Pleistocene has been documented, have been:

- Local (mis)interpretation of the interglacial, interstadial and glacial signatures from the sedimentary and fossil records within the geographically widely-spaced successions,
- Interregional miscorrelation of these,
- Variously and broadly defined unit boundaries.

The interpreted climatostratigraphical units are a major source of stratigraphical confusion on the continent. They likewise furnish difficulties to achieve an overall picture of the past climate. Attempts to correlate the climate-based units from one region to another have led to many discrepancies. The loess/palaeosol sequences in the extraglacial areas show more climatic cycles than the glacial sequences. Moreover, there is the problem of drawing boundaries of climate change. The interpreted climatostratigraphical units principally refer to local temperature and moisture conditions during relatively short periods of deposition in different glacial or non-glacial environments. Many of them only indirectly indicate climatically-induced events of global significance. Although climatostratigraphical units are intended to refer to climatic events as a cause for deposition, problems arise when, for example, every superimposed till in a glacial sequence is interpreted as a product of a discrete glaciation or when every organic stratum should represent an interglacial stage.

Thus, the synthetic character of the local and regional climatostratigraphical units make them inadequate for interregional correlation. Too many aspects of climatic history on the continents remain constrained by available local-scale multidisciplinary evidence and ages (Turner 1996). Regional long-term controls such as endogenic tectonics[4] may also be of importance in affecting the depositional systems and combine in different ways with shorter term exogenic climate influence in different depositional systems.

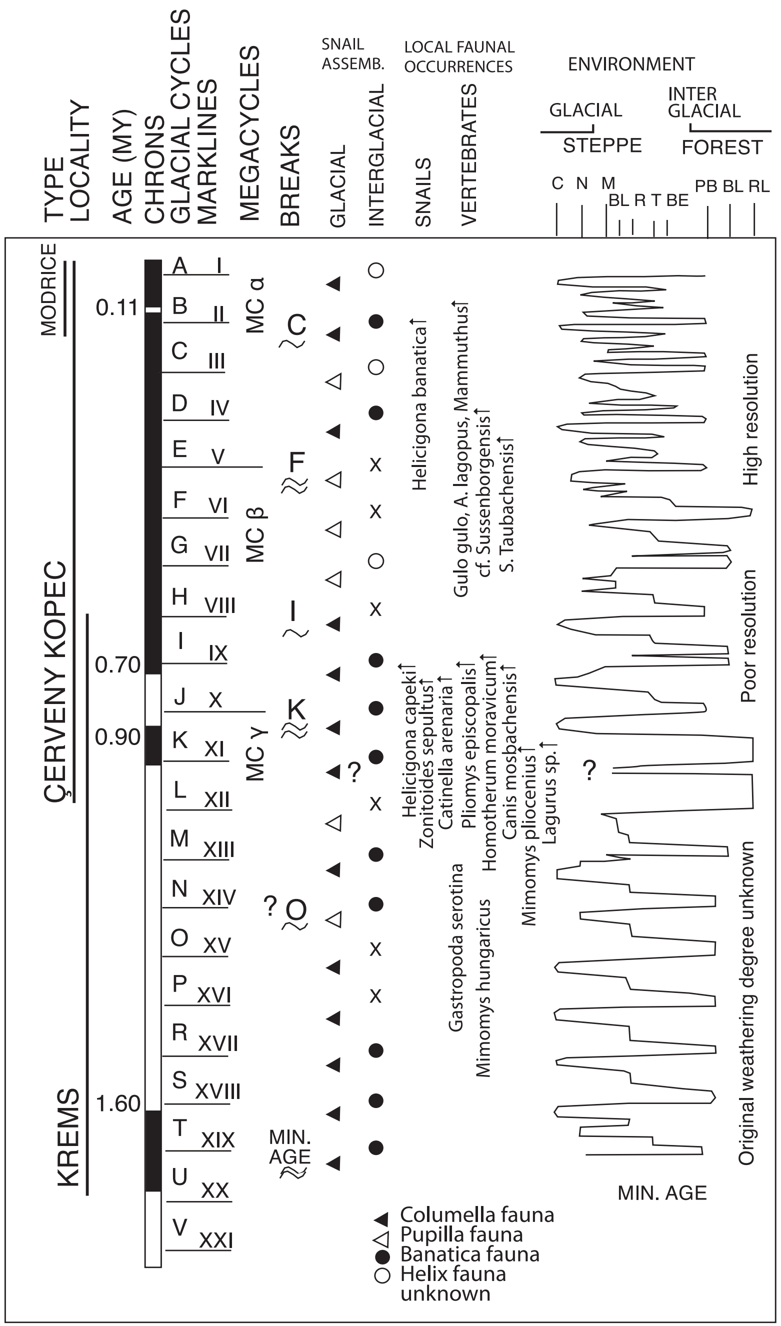
3.1.3 Persistent terminology

Climate-based units are still the principal units of conventional Pleistocene stratigraphy. Traditional terms like 'glacials'/'glacial stages' (as well as their subdivisions into 'stadials'/ 'stades' and 'interstadials'/'interstades') and 'interglacials'/'interglacial stages' have been used worldwide and remain very persistent[5]. However, these terms are actually only suitable within formerly glaciated areas. There is no clarity in the criteria by which they should be identified and defined elsewhere. Moreover, even within the glaciated regions, boundaries were identified and defined on the basis of different evidence and criteria. Furthermore, the nature of several deposits hampers unequivocal climatic interpretations to be made. Nonetheless, climatostratigraphical units were accorded formal status for a while, e.g. the geologic-climate units in the American Code (1961), but this was regarded unfeasible in the end.

Glacials/glacial stages exclusively refer to the glaciations of mid-latitude Europe as indicated by glacial deposits and landforms, for example the Weichselian, Saalian and Elsterian glaciations.

Interglacials/interglacial stages were initially used to identify erosional time units between the Alpine glacial stages, i.e. events not represented by deposits. Palaeobotanical evidence from Northwest European lake, mire and coastal marine records initiated the definition of interglacials, and interstadials, as forested periods. Following the proposal by Jessen and Milthers (1928), interglacials were defined as particular types of non-glacial conditions, as indicated by vegetational changes. Later, they also became equated with marine transgressions, periods of soil formation and other features related to relatively warm climate conditions.

In order to avoid confusion and ambiguity over nomenclature and definitions of palaeoclimatic units, the use of above-mentioned



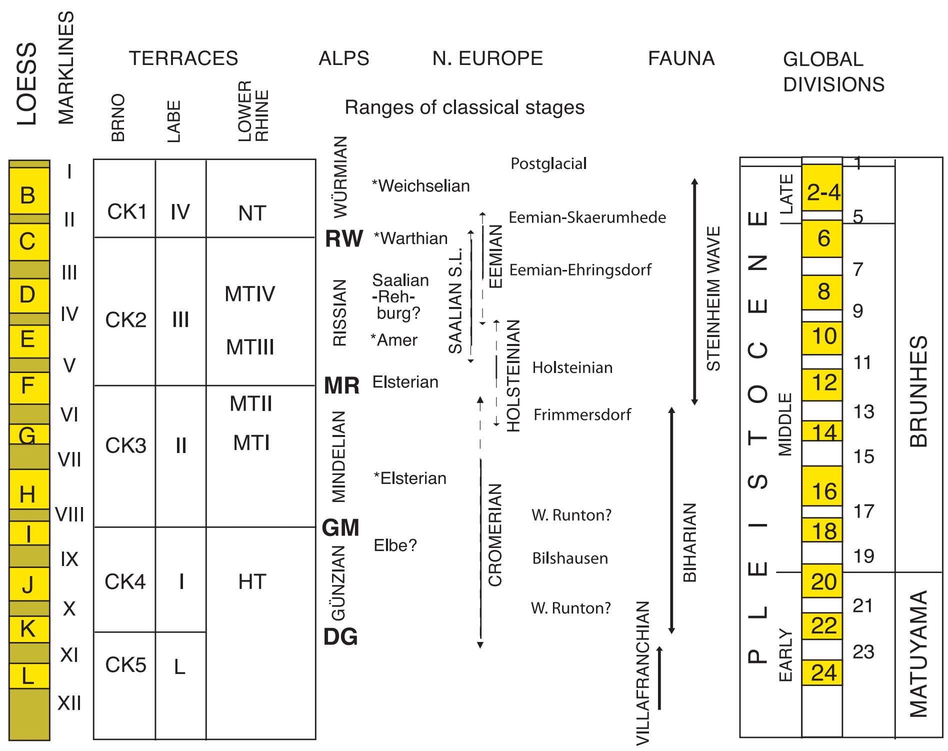


Figure 3.2 Environmental changes around Brno (Slovakia) and Krems (Austria) as reconstructed from the loess record (from Kukla 1975). Symbols for forest environment: PB: parabraunerde, BL: braunlehm, RL, cross-hatched: warm, savannah type environment favouring development of exceptionally red polygenetic soils. Symbols for snail assemblages explained in the legend. In the local fauna column, crosses mark single faunal occurrences, full dots first faunal occurrences in the loess sections. Breaks are levels of deep erosional incisions. Combined with original correlation scheme of Kukla (1977) showing his correlation of the marine isotope stages (MIS) with loess cycles, terraces and the type unis of the classical European Pleistocene subdivisions. Warm climate units stippled, intermediate dotted. Normal polarity black, reversed blank. Estimated ages of terminations and marklines from Table III (Kukla 1977). Stratigraphical positions of classical north European units at type localities are marked with a star. Stratigraphical range of most miscorrelations of the north European (climatostratigraphical) glacial and interglacial stages are shown by bars and arrows. Megacycles after Kukla (1975).

classical terms as overall climatic periods of cold (peri-)glacial conditions versus warm intervening non-glacial conditions is discouraged here. They will be as much as possible used and referred to in this thesis in their original sense and validity for the formerly glaciated regions.

A more meaningful, and widely applicable, basis for climatic subdivision on the continent is to distinguish between relatively cold stages and warm stages (*cf.* Suggate and West 1969, West 1988). These broad units of climatic change are based on local climate type interpretations[6] and compared to the present-day climate zonation of the mid-latitudes[7].
Warm or warm-temperate stages are periods characterised by forest vegetation, high sea-level stands and soil formation of substantial length comparable to the present day. In this respect they may be used as a synonym for the original 'interglacial stages' as determined from vegetational changes (Jessen & Milthers 1928). And the Eemian and Holsteinian warm stages refer to the marine transgressions in the North Sea, as well as the deciduous forest vegetations in the lake records on the continent.
Cold stages comprise all (negative) anomalies to the present-day climate zonation. They are therefore generally complex in nature and represent time periods of climatic deterioration with permafrost occurrence, tundra and steppe vegetation type, lowered sea-levels and one or more periods of ice-sheet expansion. Thus 'glaciations' or 'glacial stages' are included in the cold stages and comprise the periods of ice-sheet expansion during 'stadials'. Additionally there may be interruptions by short forested periods ('interstadials' or boreal substages) within the cold stage.
Climate type anomalies from palaeotemperature estimates can be applied to both glacial and extraglacial areas and can be recognised from deposits and structures in different depositional environments, from biota (pollen assemblages, insects) as well as from palaeosols and landforms (e.g. push moraines). Furthermore, estimates of precipitation or moisture conditions (humid/dry) can be added. Cold and warm climatic stages or periods are thus quite flexible, geographically dependent regional units, signifying several geological events, deposits or features. Both general terms will be used in discussing and reviewing the Middle Pleistocene climatic sequence in this thesis. Their informal status implies that their initial letter should not be capitalised, unless the term stage is used to refer to formal chronostratigraphical units at the (sub-)stage level, e.g. the Saalian Stage. *Figure 3.3* shows the regional subdivision of climatic stages valid for the Northwest European lowlands in relation to the chronostratigraphy.

3.1.4 Climatostratigraphy and chronostratigraphy

Climatostratigraphical units were thought to offer foundations for chronostratigraphical subdivision of the Pleistocene sequences and for interregional correlation. However, in view of the complex and heterogeneous nature of its succession, together with its brief age, the matching of locally and regionally known stratigraphical units appeared to be a correlative obstacle to chronostratigraphical subdivision. Supposed time correspondences between the climatic stages were based on the correlation of lithological characteristics, palaeontological and palynological data and other evidence like morphological and pedological features. The inherently defective and 'floating'[8] local and regional chronostratigraphical models, however, comprise few geochronological control points and many

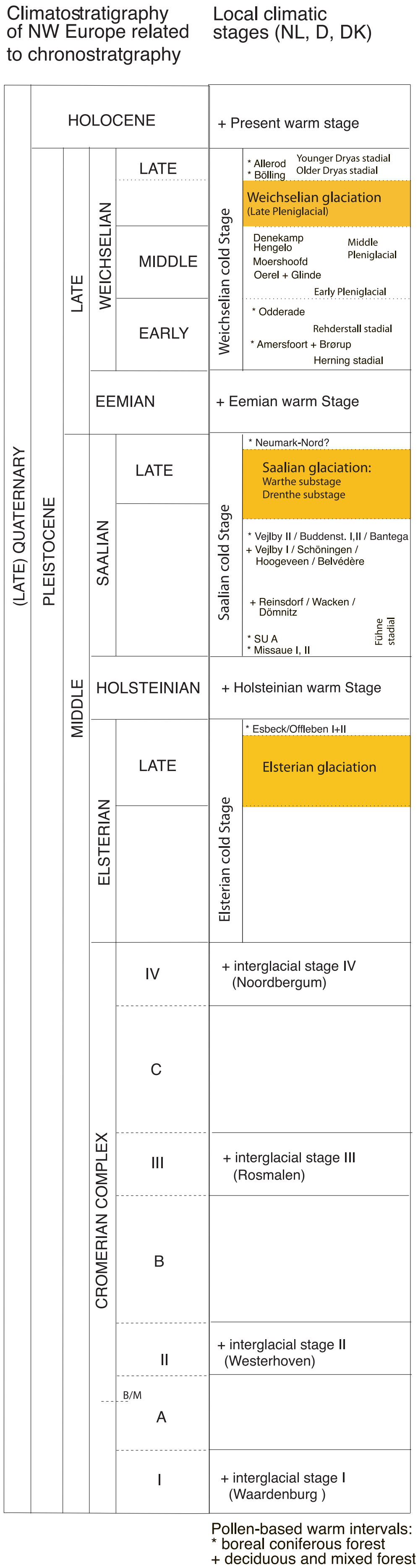


Figure 3.3 Subdivision of the Middle Pleistocene in Northwest Europe based on interpreted cold and warm climatic stages (compiled from different sources).

diachronous hiatal breaks as missing links of unknown duration. Notwithstanding the practical merits of classifying local sequences into climate episode units, they should not be adopted as chronostratigraphical stages (and hence not geochronological ages) applicable on a continental scale. Climate-based units, like the conventional stratigraphical units from which they are established, are equally time-transgressive, geographically and temporally restricted fragments. Consequently, they do not have an adequate chronostratigraphical definition, that is based on unit- or boundary stratotypes within a continuous sequence and with time-parallel boundaries. A stage or substage rank implies time correlation which is neither true for cold nor for warm stages in northern Europe. Moreover, the use of various criteria for their boundaries is inconsistent and implies the existence of gaps (and in some cases, overlaps) which is generally not shown in the palaeoclimatic tables and curves. Major erosional and subaerial unconformities filling the gaps between phases of sedimentation may span tens or hundreds of thousands years. They therefore form a substantial, but virtual, part of the chronostratigraphy in the different European type regions (Kukla 1975, Bowen 1978). A better appreciation of their relevance is emphasised and substantiated in the stratigraphical procedures followed in *section 2.5*.

The basic European glacial models may be regarded as outdated. Although intrinsicely different in nature, they are only rough structures when compared to the interglacial-glacial cycles in the oceanic record. Kukla already concluded in 1977 'that it is urgently recommended to abandon the classical terminology in all interregional correlations and to base the chronostratigraphical subdivision of the Pleistocene on the (^{18}O-record of deep-sea sediments' that showed eight, instead of four, glacial cycles during the Brunhes normal Chron. Because all terrestrial sequences contain actual and potential hiatuses, Bowen (1978) also proposed that the deep-sea cores should be used as a standard. While the temptations of direct land-sea correlations are large, the replacement of locally established terrestrial scales has never been achieved in a formal or systematical way. There are many principal objections and practical limitations involved, as noted by Gibbard and West (2000). They recommend the separation and retention of regional chronostratigraphies for each sequence-type, and that these should be correlated using event-based stratigraphy where possible. Thus, in the absence of a valid European framework, subsidiary classifications are required that better represent the terrestrial Pleistocene record and that potentially offer opportunities to correlate with the marine isotope stratigraphy.

3.1.5 Chronostratigraphical boundaries of the Middle Pleistocene subseries

Internal dating of the Middle Pleistocene succession in the Northwest and Central European type regions is primarily relative and based on superposition and correlation of preserved depositional (lithostratigraphical) units and their biostratigraphy, using palynological and various palaeozoological zonations. Geochronometric and geomagnetic dating methods, developed since the 1950s, have to some extent proved valuable supplementary means on the chronostratigraphical position of deposits (see also *section 3.4*). The resolution of these methods, however, decreases with time. The radiocarbon method, established by Libby (1955), provided a sound basis for dating the last 40,000 to 50,000 years. Dating techniques such as K/Ar, Ar/Ar, TL and OSL, U-series and ESR[9] yield ages up to 300-400 ka, or even more, for suitable sediments and fossils, but are not very reliable yet and still in development.

Consequently, the discontinuous and genetically diverse Middle, and likewise Early, Pleistocene terrestrial subseries have a low resolution classification[10]. In fact only the lower and upper boundaries can be accurately defined:

The lower boundary of the Middle Pleistocene is proposed at the first sedimentary units where palaeomagnetic dating of the sediments show normal (Brunhes) geomagnetic polarity (Richmond 1996). A lower dating limit of about 780.000 years[11] ago then can be set as a maximum age which corresponds to MIS 19.

Based on different criteria the upper boundary of the Middle Pleistocene on land is defined at the beginning of the last interglacial/glacial cycle. This in practice appears to be a diffuse non-synchronous boundary. The transition of glacial and subaerial periglacial sequences, related to the penultimate completed glacial cycle (C), to the last non-glacial (Eemian) sedimentary cycle of marine, lacustrine and fluvial origin, or to soil formation (starting with decalcification) or to forest vegetation, is represented by different starting points in the time interval between the MIS 6 global ice-volume maximum and MIS 5e global ice-volume minimum, i.e the deglaciation. The Middle/Late Pleistocene boundary is set at the transition of MIS 6 and 5e for which the midpoint at 128 ka ('termination II') has been chosen arbitrarily as stage boundary (Broecker and Van Donk 1970, Gibbard 2002). Recently the Amsterdam-Terminal borehole (Van Leeuwen *et al.* 2000) has been proposed as the Eemian boundary stratotype for Northwest Europe (Gibbard 2003).

3.2 Material building blocks of the Northwest and Central European Pleistocene stratigraphy

The shallow subsurface of Northwest and Central Europe is one of the best geologically investigated areas worldwide. Material evidence of unlithified Pleistocene deposits from numerous field research localities, such as open-air sections and boreholes, have been described and subdivided into local, regional and national classification systems. The factual units structuring the local stratigraphies are of a lithostratigraphical, biostratigraphical and morphostratigraphical type in which (litho)genetic aspects play an important role. They do represent many different environments having repeatedly coexisted in Pleistocene time. This section reviews the building blocks of the local and regional stratigraphies of this part of Europe and the relationship between the stratigraphical sequences at one locality to those at another.

The basic sedimentary components building and contributing in different ways to the local and regional stratigraphies are:

- Sediments generated in glacial depositional environments,
- Sediments generated in subaerial periglacial depositional environments,
- Marine coastal and shallow sea sediments,
- Fluvial and deltaic sediments produced by the large river systems,
- Sediments deposited in lakes, mires and bogs.

These categories represent the dominant depositional systems which form the main building blocks from which the regional Quaternary stratigraphies of Northwest and Central Europe are constructed. Since most formations in the different European stratigraphical systems include lithogenetic criteria, they generally correspond to one of the five categories. With the exception of the lacustrine deposits, the sediments of the other categories have dispersals that can be mapped over large areas. They largely corre-

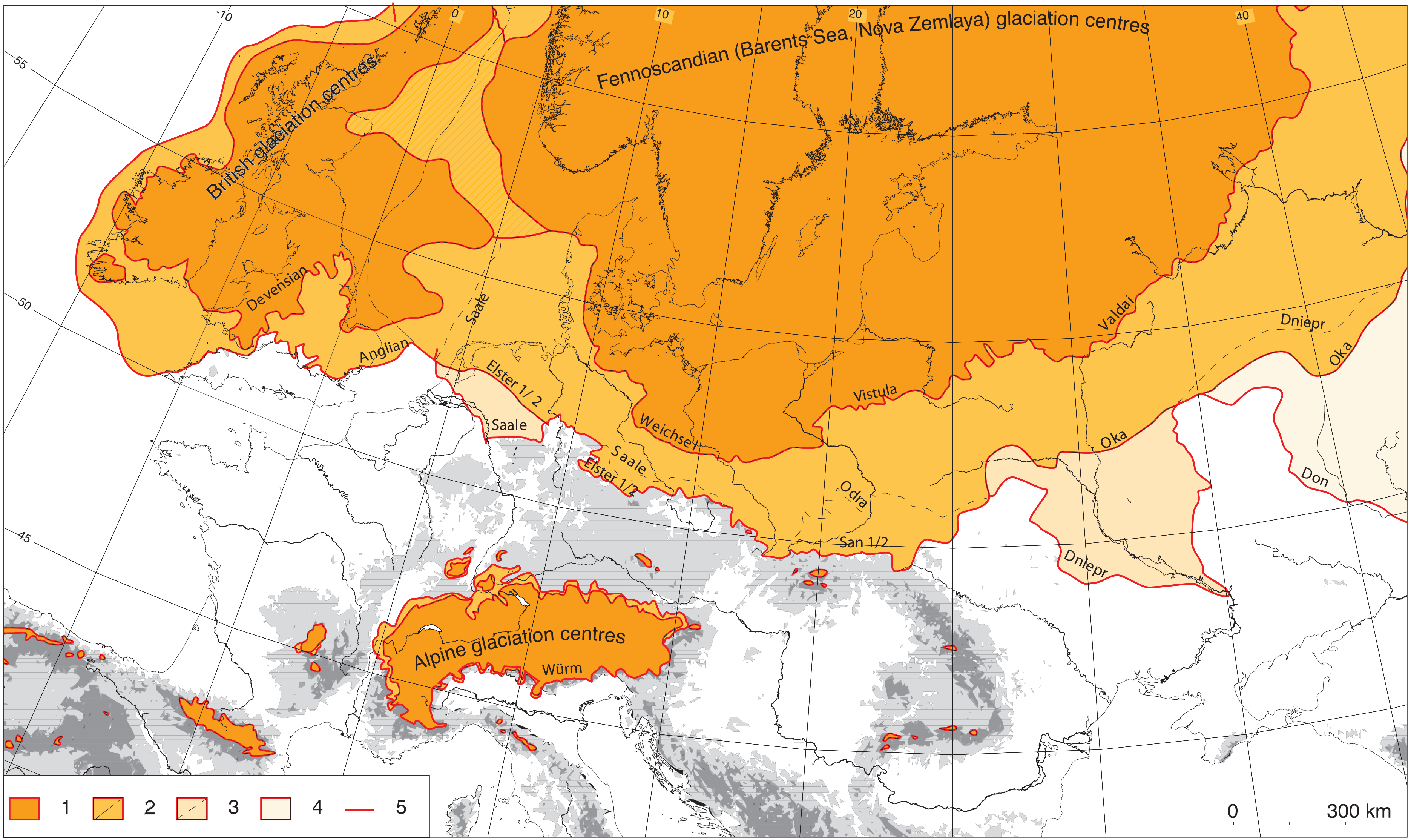


Figure 3.4 Distribution of Pleistocene glacial sequences in Northwest and Central Europe (from 'International Quaternary Map of Europe', UN/BGR 1965-1995). 1. Extent of Weichselian, Devensian and Würmian glacial sequences, 2 Extent of Elsterian, Anglian and Mindelian glaciation limits, 3. Extent of Saalian and Rissian (I) glaciation limits, 4, ice-free area during the Fennoscandian glacial cycle B (Weichselian, Devensian), 5. maximum limits of the Pleistocene glaciations.

spond to the legend units of the 1:2.5 million scale 'International Quaternary Map of Europe' (UN/BGR 1965-1995) from which the distribution maps in *Figures 3.4* up to *3.8* have been compiled.
In the next sections the litho- and biofacies characteristics and stratigraphical significance of the categories will be discussed in relation to depositional processes, climatic change and regional tectonic effects that controlled their formation. Emphasis is put on glacial and periglacial sedimentary sequences since their geometries largely structure the local stratigraphies in the glaciated areas respectively the extraglacial areas in Europe. From both areas compilation schemes are produced, arranged along W-E transects, which are presented and further discussed in *section 4.2*.

3.2.1 Sediments generated in glacial depositional environments

Glacial sequence includes till (glacial diamicton), glaciofluvial sand and gravel, glaciolacustrine and glaciomarine clay, silt and ice-rafted detritus. Tills do not occur beyond areas covered by ice-sheets and glaciers. Glaciofluvial and glaciolacustrine sediments preserved on land areas do not extend far beyond the maximum extend from which they were derived (Boulton 1990). Glaciomarine sediments are laid down beneath and in front of ice-sheets which entered the sea. Also non-glacial sediments which have been glaciotectonically deformed and/or dislocated, and are commonly incorporated in push moraines, may be considered part of the glacial depositional system.
The extent of glacial sediments and landforms in Northwest and Central Europe is shown in *Figure 3.4*. The subdivision of glacial stages in Northwest Europe was formerly based on morphostratigraphical criteria ('Endmoränenstratigraphie' - e.g. Keilhack 1926, Woldstedt 1929 and 1954) but is now based on till stratigraphy. Nonetheless, the classical subdivision of glacial stages (Elsterian, Saalian and Weichselian) is still regarded as valid, although till stratigraphical studies have revealed several phases of glacier advance and retreat within each stage. In eastern Europe (Poland and Russia) this three-fold subdivision may be extended by an older glacial stage (Donian). The Don tills extend far south into the Don basin (Velichko and Faustova 1986).
Because of the strongly erosional effects of ice-sheets, preservation conditions of glacial (and non-glacial) deposits predating the latest glaciation are limited and cause for major unconformities. Most extensive depositional glacial sequences are found in former ice marginal positions where they intervene with fluvial, aeolian and slope sequences of the periglacial zone.
The major sedimentary elements in glacial environments are:

[a] Tills

Till units[12] play an essential role in structuring stratigraphical subdivisions in northern Europe. They were deposited by wide-spread glacial events which were well-integrated on a continent-wide scale. Till units can therefore expect to be correlatable as part of a wide-spread sedimentary product whose properties also vary systematically on a continent-wide scale. Appropriate sedimentological analyses can therefore yield reasonable correlations which permit lateral connectivity to be established between otherwise disconnected exposures. They permit workers to reduce the high degree of uncertainty in stratigraphical reconstructions in an otherwise poorly represented time/space domain such as that shown in *Figure 4.2*. Boulton *et al.* (1997) have argued that tills are generally deposited in a relatively narrow zone close to an ice-sheet margin and that in more proximal zones erosion will dominate. As a consequence, except near to the limit of glaciation, much of a glacial phase will be taken up by erosion and only the last phase of glaciation will be represented by till at any one site. Genetic distinctions between basal or lodgement tills, ablation tills and flow tills are rather irrelevant then. The till unit produced during a simple glacial cycle may thus be highly diachronous. Its deposition may be complete near to the maximum of glaciation thousands of years before deposition begins in areas of final decay. Nonetheless, it still plays a vital role in defining the stratigraphical level within which a glacial phase may lie. Direct dating of glacial tills has, as yet, proved illusive, in spite of published thermoluminescence (TL)-ages on Polish tills (Rzechowski 1986).

There are contrasts in approach to till stratigraphy in different parts of Europe. In the Netherlands all sediments deposited during a single glacial stage are combined into a single formation. In Denmark (Houmark-Nielsen 1987), Great Britain (Rose 1989) and Poland (e.g. Mojski 1985, Rzechowski 1986) the glacial stages are defined on the basis of individual till units which comprise formations with well-marked upper, lower and lateral boundaries and defined with reference to a type-locality. In Germany, most till units are described with reference to their supposed chronostratigraphical position (Ehlers *et al.* 1984, Ehlers 1990).
Correlations between tills are based on superposition and lithological criteria likely to reflect large-scale integration of sedimentary processes, such as large-scale patterns of mineralogy, granulometry, clast lithology and sedimentary and tectonic fabric (e.g. kineto-stratigraphy in Denmark; Berthelsen 1978). Sedimentary structure alone is often a poor guide, as it may merely reflects local depositional processes.

Vertical lithological differentiation in till beds, even across sharp discontinuities, cannot be used as unequivocal evidence of deglaciation separating two glacial events. The existence of extraglacial (or non-glacial) sedimentation at some point is required.

By tracing the indicator erratics or matrix composition of tills to their source, it has proved possible to show that each glaciation which extended across the Northwest European plain underwent a systematic change in flow direction through the glacial cycle. An early-glacial northerly source progressively gives way to a north-easterly then easterly source, presumably reflecting the progressive migration of the ice-sheet's flow divide in an easterly direction (Ehlers 1983). Most approaches link the percentage of erratic pebbles within the till with the provenance areas in Scandinavia. The method was first developed by Hesemann (1930, 1934) and has been widely used both as an indicator of flow directions and a correlation tool in Germany (Lüttig 1958, Meyer 1983) and the Netherlands (Zandstra 1974, 1987). In the Saalian till cover of the Netherlands for example, several till facies can be distinguished (Zandstra 1987) representing changes in the source areas of the erratics and the ice-flow direction (Rappol 1983, 1987) during one ice advance. Changes in the ice-flow directions have also been reported from fabric measurements in Saalian tills from eastern Germany (Eissmann & Müller 1979; Böse 1990) and Denmark (Sjörring 1983, Houmark-Nielsen 1987).

[b] Glaciofluvial and glaciolacustrine sediments

Subglacial fluvial and lacustrine sediments, as found for example in eskers and drumlins, are volumetrically unimportant in modern glaciers compared to their proglacial equivalents, and there is no reason to assume that this situation was different for former ice-sheets. Both glaciofluvial and glaciolacustrine sediments occur predominantly near the glacier margin, sometimes in ice-contact positions. They show strong spatial and compositional variability, from extensive coarse-grained lithofacies associations to local silt and clay beds. An example of the former are the so-called '*Vorschuttsande und -kiese*', sandur deposits overlain by tills, in northern Germany (Meyer 1983).

In many instances the deposits are associated with temporary ice-dammed lakes which formed during the advance as well as during the deglaciation. Although not amenable to direct dating, the duration of these proglacial sedimentation phases in lakes is regarded short. The features are generally of little value for wide correlation. Most occurrences are therefore left unclassified or are only used in relation to morphostratigraphy to identify glacial limits. Where deglacial ice-margins remain stable for longer periods, large glaciofluvial masses frequently give rise to hummocky, kettled topography or they are pushed into major push moraines during subsequent glacier re-advances. Indeed, many of the largest moraines are for the greater part composed of outwash sediment, sometimes associated with glaciotectonic structures reflecting ice-pushing or collapse of buried ice masses.

As the environment in which fluvioglacial and glaciolacustrine sediments form is so dynamic, they tend to represent relatively short periods of time. However, some distinctive glaciolacustrine sediments are wide-spread, such as the Peelo Formation clays in the Netherlands and their correlatives in Germany, the Lauenburg Clay. They appear to fill in the upper parts of a system of elongated basins, dissected under subglacial conditions by the Elsterian glaciation and are overlain by Holsteinian warm stage deposits. A similar sequence occurs in subglacial basins produced during the later Saalian glaciation in the same area where tills are overlain by, often varve-like, laminated clay and fine sand, followed by Eemian warm stage deposits.

[c] Glaciomarine deposits

Glaciomarine deposits also tend to be deposited in relatively narrow zones (Boulton 1990) and therefore represent short time periods when found in the geological record, although the high sedimentation rates common in glaciomarine environments can produce large thicknesses in short periods. In high latitudes, it is normal to find that, at modern sea-level, tills are overlain by glaciomarine beds, reflecting high local relative sea-levels during glaciation because of the strong lithosphere subsidence beneath and just beyond the ice-sheet (Boulton 1990). Rapid subsequent uplift produces emergence and the glaciomarine units are overlain by beach deposits. They can therefore represent very short time periods and are highly diachronous. The sequence is however a highly distinctive marker for glaciation.

In mid-latitude coastal areas however, there is a marked lack of evidence of such a glaciomarine phase above modern sea-level, with the possible exception in the Irish Sea basin during the last glacial cycle (Eyles and McCabe 1991). This may be a result of glacio-isostatic rebound or reflect low ice-sheet surface slopes, and therefore less ice-loading, at the southern margins of the North European ice-sheets resulting from flow over a deformable bed (Boulton and Jones 1979). Deglacial glaciomarine sequences are common along the mountainous west coast of Norway (Mangerud 1983, 1991). These sequences are of special interest because they show phases of ice-rafting reflecting ice margin fluctuations during the deglaciation (Baumann *et al.* 1995)

Glaciomarine (and glaciolacustrine) deposits are, however, widely found below modern sea-level in the North Sea (Cameron *et al.* 1988). The deposits which fill the Elsterian depressions in the southern part of the North Sea (Swarte Bank Formation) are a marker bed in the offshore stratigraphy. Similar deposits are found in channels originating from the last two glacial cycles. Glaciomarine deposits in the central North Sea overlying the Swarte Bank Formation indicate deposition at distance from the Saalian and Weichselian ice-sheets which entered the North Sea from Fennoscandia and Britain.

[d] Glacial landforms and glaciotectonic features

Glaciation has a fundamental impact on earth surface morphology through erosional[13] and depositional[14] processes, which create a new landscape on which subsequent sedimentary and environmental events occur. The palaeogeography of the Northwest and Central European lowlands indeed has been drastically remodelled as a consequence of repeated glacial activity.

The most striking geomorphological features developed by glacial surface processes are the moraine belts and associated basins delimiting ice-limits (*figure 3.4*). They comprise highly variable pre- and syndepositional units, often incorporating older (deformed) formations while their lower boundaries are surfaces of décollement. They also permit the reconstruction of the areal pattern of ice-sheet development, which would be impractical from till stratigraphy alone. As has been mentioned previously, subdivision of the glaciations of the north European lowland was originally based on the so-called '*Endmoränenstratigraphie*'. According to this morphostratigraphical concept all (push) moraines lying within the subsequent maximum glaciation limits, were assumed to be end-moraines or recession-moraines. Two glaciations were, for example, distinguished within the Saalian Stage: the Drenthe and Warthe Substages. Since no intermediate sediments incorporating evidence for interglacial vegetation has been found, it is assumed that they reflect, together with other end-moraine series, ice-marginal positions during different phases of the Saalian glaciation.

Thus, the end-moraines do not necessarily indicate major climatic change and they are in most cases related to short climatic oscillations at the ice-sheet margin.

3.2.2 Sediments generated in periglacial subaerial environments

With the repeated expansion of ice-sheets and periglacial areas during the Pleistocene the mid-latitudes also experienced cold-climate conditions. The most relevant and typical sediments that are produced subaerially in these cold, unglaciated areas include loess and local slope deposits resulting from mass wasting processes.

Loess is the most wide-spread product of Pleistocene periglacial action. The aeolian deposits have been formed in the unvegetated upland areas and lowland plains beyond the margins of the former ice-sheets and extend in a zone from France into China. They are evidence for cold, dry and windy climate conditions indicating the expansion of desert environments that coincided with the mid-latitude glaciation maxima. This concept of 'glacial aridity' is part of the correlation potential of primary loess units alternated by palaeosols.

Subaerial deposition under prevailing humid periglacial climate conditions comprised various kinds of locally derived slope deposits, one of which may be reworked loess. Distinction between the different types of slope deposits is not always clear, however. Sedimentary products associated with mass wasting and frozen ground constitute common elements in local stratigraphies as well as cryogenic structures. The latter may be post-depositional and will be discussed in *section 3.2.7*.

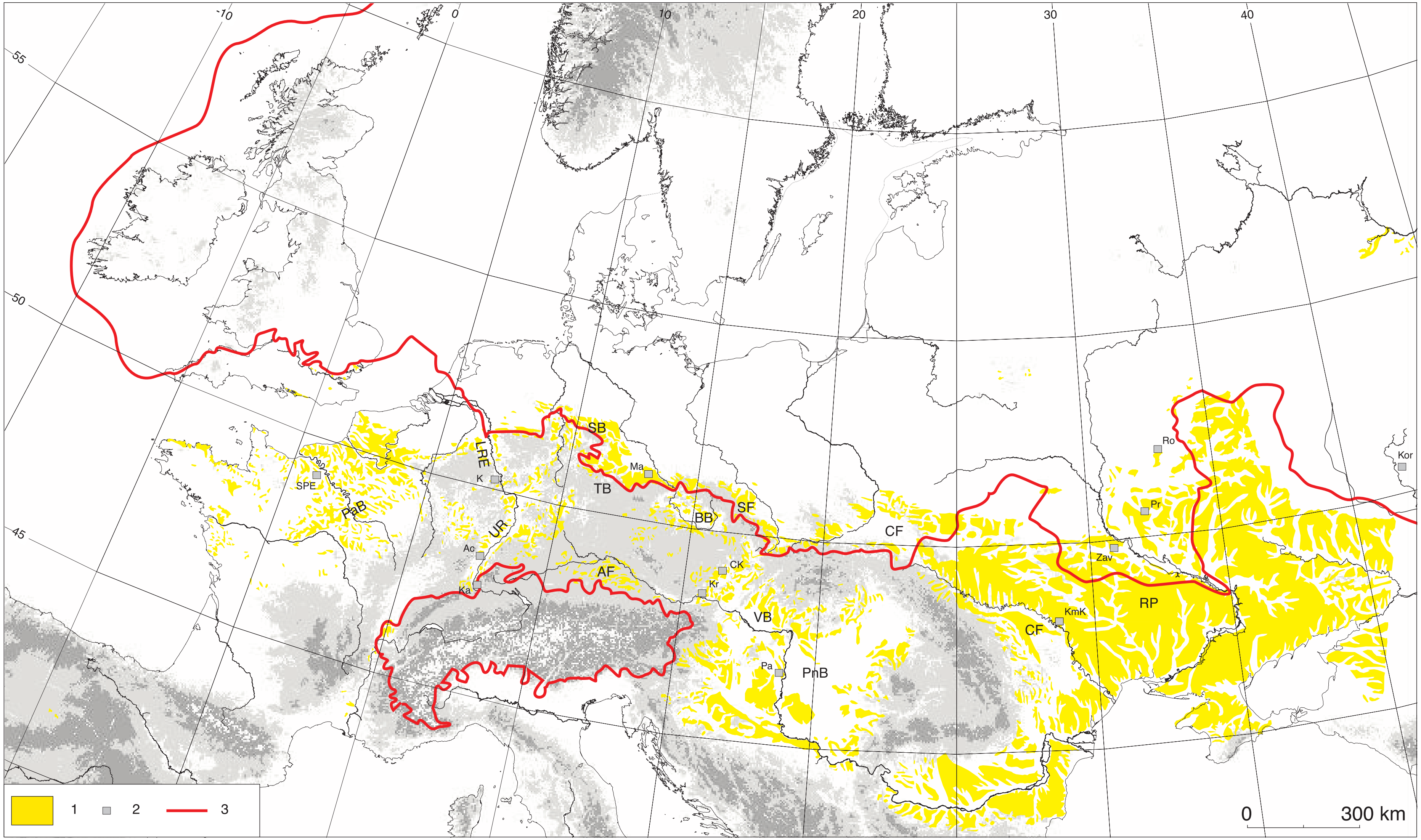


Figure 3.5 Distribution of Pleistocene periglacial subaerial loess sequences in Northwest and Central Europe (from ‘International Quaternary Map of Europe’, UN/BGR 1965-1995). 1. extent of Late Pleistocene loess sequences, 2. some key-stratigraphical localities in Northwest and Central Europe: Ac: Achenheim, CK: Cervený Kopec, K: Kärlich, Kr: Krems, Ma: Mahlis, Pa: Paks, SPE: St. Pierre les Elbeufs, 3. maximum limit of glaciation during the Pleistocene.

[a] Aeolian deposits

Aeolian periglacial deposits are primarily represented by loess and fine to medium-grained (cover) sands. Loess, consisting of wind-blown calcareous silt-sized material, covers extensive mid-latitude areas that were marginal to former Pleistocene ice-sheets (*Fig. 3.5*).

Two types of loess sequences may be distinguished in the aeolian record (Kukla and Çilek 1996): a) plateau (platform) deposits and b) valley slope deposits. Sedimentation and pedogenesis in these two types of deposits proceed in different ways. The platform sequence accumulates entirely from subaerial dust deposition (e.g. China Loess Plateau). Slope deposits are sedimentary fills of depressions usually formed at the lee-side of steep terrace faces cut in bedrock by meandering Pleistocene rivers (e.g. Červený Kopec (*Fig. 2.2*)). Next to primary loess, the latter loess sequences frequently show reworking on a local-scale commonly explained by rainwash and nivation processes as can be recognised by fine wavy laminations, lenses or horizons of sand and fine gravel or interstratified molluscs. Reworked loess deposits are generally referred to as loess derivates or have regional terms like '*brickearth*' (southern England) and '*Schwemmlöß*' (Germany: colluvial loess).

Loess sequences do not record continuous deposition. Series of loess beds are mostly interstratified with soil complexes which reflect gaps caused by non-deposition during warm and humid climate intervals of (forest) vegetation. The soils, humic or leached, can be used broadly to indicate a warm-stage character. However, because soil formation is influenced by a wide variety of independent, local factors (Catt 1988), the potential of buried soils for use as detailed palaeoclimatic markers is limited (see also *section 3.2.7*). Moreover, phases of erosion or non-deposition may result in hiatuses or polycyclic soils (pedocomplexes). Additional palaeoclimatic information from loess is yielded by microfaunal data, e.g. mollusc assemblages, that may indicate contemporaneous temperature and moisture conditions (*section 3.3.2*). Horizons between the primary loess units also may contain mammalian fauna, pollen and artefacts.

The loess stratigraphy of the Central European extraglacial zone is analogous to the till stratigraphy of the glaciated areas, although individual loess units probably represent longer proportions of each glacial phase. The stratigraphical potential of loess sequences increases when more loess/palaeosol cycles are stacked; in vertical superposition as in China and Tadjikistan or in a 'telescopic superposition' like in Central Europe (Kukla and Lozek 1961). Long stratigraphical sequences which appear to include all the principal elements of late Quaternary glacial-interglacial cycles, occur in Slovakia and Austria where they are associated with river terraces (Kukla 1970, 1977), in southern Ukraine (Veklich 1969, Veklich *et al.* 1993) and on the Russian Plain (Velichko 1990). Other less lengthy sequences are found in the Upper and Middle Rhine Valley, e.g. Achenheim (covering the last 3 climatic cycles: Heim *et al.* 1982, Rousseau & Puisségur 1990), Kärlich (Brunnacker *et al.* 1969) and Ariendorf (Brunnacker *et al.* 1975), and also in northern France along the rivers Seine (St. Pierre les Elbeuf: Lautridou 1982) and Somme (Antoine 1991, 1995). Typical loess is, however, found quite infrequent in Northwest Europe; many interruptions resulting from slopewash or some kind of gravitational flow occur.

The loess/palaeosol sequences provide a link between the deep-ocean record and the classical glacial stages on land. Kukla (1970, 1975) correlated the terminations of the marine isotope record with his 'marklines' in the loess successions of Červený Kopec (*Fig. 3.2*). These are boundaries between thick layers of loess, containing gastropods reflecting cold, dry conditions and overlying warm-stage soils[15] and hillwash, indicating abrupt ameliorations of climate. The marklines delimit glacial cycles. Within each glacial cycle, less well developed soil types are distinguished indicating climatic substages. The warm-stage soils often contain molluscs and plant remains indicating formation under typical forest vegetation. When combined with palaeontological, thermoluminescence (TL) and geomagnetic dating, the long-term loess/palaeosol sequence gives a fairly solid (age) match with the oceanic MIS.

[b] Mass wasting products

Most wide-spread in present and former periglacial areas are the subaerial deposits on and at the foot of slopes, filling depressions and stream valleys. They result from the combined effect of gravity movements (mass wasting), soil frost, rainwash and stream activity. Of all categories of mass wasting processes, the most common and effective one was solifluction: the slow downslope movement of water-saturated material (Andersson 1906). Solifluction is favoured in treeless situations and over permafrost, although the latter is not a prerequisite[16]. Solifluction phenomena on low-angle slopes like the '*Head*' deposits in Britain and Ireland or '*Fließerde*' in Germany reflect levelling of the regional morphology characteristic of areas suffering polar climates. The lithologically highly variable sediments generally are heterogeneous, unstratified and poorly sorted diamicton. The sheets and lobes in which they occur may display crude sorting into lenses or pockets of finer and coarser material as a result of differential density flow. The presence of these flow structures, indicating the degree of deformation, is one of the main criteria to distinguish them from the original sediments from which they are derived. Also loess and former soils may be incorporated in solifluction features. The latter which are termed para-autochtonous and give rise to misinterpretations in local stratigraphies.

The stratigraphical value of most solifluction deposits is limited in general, because they consist of reworked local material, including their fossils. They are, however, useful as general indicators of periglacial (cold and humid) conditions in subaerial environments, although direct interpretation is not always possible. In many cases they cover or separate other stratigraphical units or archaeological horizons and protect them from erosion.

3.2.3 Coastal marine and shallow sea sediments

Although sediments deposited in shallow seas under non-glacial conditions tend to be individually more extensive than those on land, they show many of the same problems of discontinuity in time and space. Their great advantage derives from a technique, continuous reflection seismic profiling, which permits their geometry and seismostratigraphical sequence to be established along any arbitrarily defined line, in contrast with terrestrial sequences onshore, which depend upon chance exposures or expensive boreholes.

The non-glacial marine units comprise intertidal and shallow marine sands and clays deposited during high (eustatic) sea-level stands in warm climatic intervals. Marine sequences are often incomplete as the advancing sea-level front is predominantly erosive. In most cases only the basal parts of the sequences have been preserved. Upper boundaries are time-transgressive and difficult to identify. Marine transgressions in the North Sea basin at the beginning of warm stages, when global sea-level is rising as a consequence of ice-sheet melting, have been identified for Cromerian

Figure 3.6 The distribution of basinal and terraced marine deposits in the onshore areas of Europe (from 'International Quaternary Map of Europe', UN/BGR 1965-1995). 1. areal extent of Pleistocene marine sequences, 2 maximum extent of the Eemian Sea in Fennoscandia, in the Baltic and adjoining parts of Russia (from Forsström *et al.* 1988), 3. present coastline, 4., 5., 6. important marine terrace localities.

IV, Holsteinian, Eemian and Holocene sequences. Lithostratigraphically these transgressive deposits mark the upper boundaries of the preceding Cromerian C, Elsterian, Saalian and Weichselian glacial stages.

Sediments deposited in marine environments in non-glacial conditions are now well known from the North Sea and the Baltic region from coring and seismic profiling studies. Warm-stage marine sequences are found on land in western and northern Denmark (Tornskov, Skaerumhede: Knudsen 1985, 1987, Lykke-Andersen 1987, Seidenkrantz 1996), northern Germany (Holsteinian type locality: Menke 1968, Dockenhude: Meyer *et al.* 1994), the Netherlands (Eemian type locality at Amersfoort: Zagwijn 1961, 1983 and Amsterdam-Terminal: Van Leeuwen *et al.* 2000), in northern France/Belgium (Holsteinian parastratotype at Herzeele: Sommé *et al.* 1978, Lautridou 1982) and Britain (e.g. Nar Valley: Ventris 1996). Warm-stage marine deposits along the Baltic Sea are known from Eastern Germany (Rostock: Gehl 1961) and Poland (Lower Vistula Sztum and Tychnowy marine series: Makowska 1986, Head *et al.* 2004), as well as from localities in Denmark, Lithuania, Latvia and the Kalingrad district.

Pleistocene marine deposits occur above modern sea-level where they have been uplifted by progressive isostatic uplift or where they represent warm- stage sea-levels higher than at present. In the North Sea area also factors like isostatic rebound (Lambeck 1993), hydrostatic pressure (Mörner 1980) and subsequent glacial thrusting should be taken into account when correlating and reconstructing marine sequences and their palaeogeography. Preserved Eemian and Holsteinian marine units, outcropping along the coastline of the North Sea and the Channel in England, France and Belgium form terraces up to several tens of metres. In contrast there are the marine deposits in the North Sea basin coastal areas and offshore that have been drowned or buried and occur at positions well below present-day sea-level.

3.2.4 Fluvial and deltaic sediments

Part of the history of the large river systems in Northwest and Central Europe is recorded by terrace sediment series, remains of former valley floors, along their valley sides. Terrace formation is the result of both climatic and tectonic changes in time, affecting the graded profile of the river systems. The initial development of terraces is mainly determined by climatic factors. Terrace deposits largely owe their origin to changes in discharge and sediment supply. Their long-term preservation is closely related to the prevailing tectonic regime within the different parts of drainage basins (Veltkamp and Van den Berg 1993).

The key point about terraces is that the sediment sequences and the surface developed upon them are two different things. The term 'terrace' is a morphological feature. The sediment sequences preserved below terrace surfaces are often internally complex, potentially preserving remnants of several cycles of deposition on a range of scales. The surfaces themselves may not be developed on fluvial sediments alone but on subaerial slope and aeolian sediments. And the surface may not be in its original form but may have been remodelled by periglacial and/or soil processes, post-depositionally. The age of any particular terrace surface therefore cannot be automatically assumed in either cross- or down-valley situations. It has to be interpreted from the lithostratigraphical sequence of the underlying sediments.

Many terrace sediment sequences reflect successive phases of aggradation and incision in which the older terraces lie at higher elevations due to progressive isostatic uplift and denudation in their catchments. They therefore have the advantage of showing unequivocal age sequences, although they represent very limited areas. Of particular importance in palaeoclimatic interpretation are sequences for rivers draining areas of slow continuous uplift in the extraglacial zone and which were not influenced by glacial meltwaters like the river Somme in northern France (Antoine 1990).

The response of other rivers to long-term climatic change may be a response to more complex events within the catchment (Baker 1983). Nevertheless, field evidence suggests that the greater part of the coarse-grained terrace sediments in Northwest Europe have been deposited during cold stages (Gibbard 1988). The channels cut in these terrace levels frequently contain distinctive clay and organic beds which contain evidence of warm-stage vegetation and fauna. The units of the long terrace sequence in the midstream regions of the Somme valley (Antoine 1990), for example, include fine-grained, meandering channel deposits in their upper parts.

Thickest fluvial and deltaic sequences occur in subsidence basins, such as in the Upper Rhine Graben and in the North Sea basin. Fluvial sequences here occur in vertical superpositional situations because tectonic downwarping is operating at a sufficiently high rate that the incisional phases are unable to remove pre-existing sequences. Recognition of the geometry and of erosional unconformities, as well as petrographical studies and heavy-mineral analysis here is only possible from cores.

The Middle Pleistocene alluvial plain and terrace sequences in the lower sections of rivers draining northward into the glaciated areas, like the Elbe, Weser and Rhine also reflect responses to downcutting and aggradation cycles as a result of sea-level fluctuations, glaciations and glacio-isostatic effects. They are interbedded or interfinger with marine and glacial sequences and are preserved as depositional units in both morphological terraces and in vertical superposition.

Middle to Late Pleistocene terrace stratigraphies, based on morphological, sediment-petrological, lithological (and palaeontological) criteria, have been developed for the middle and/or upper sections of the Thames (Gibbard 1985, Bridgland 1994, Bridgland and Schreve 2001), Lower Thames (Gibbard 1994, 1995), Somme (Bourdier *et al.* 1974), Maas (Veltkamp & Van den Berg 1993, Van den Berg 1996), Lower Rhine (Brunnacker *et al.* 1978; Klostermann 1992), Middle Rhine (Bibus 1980), Weser (Lüttig 1974), Elbe/Saale (Eissmann 1975), Elbe/Ilm (Mania 1989), Elbe/Vltava (Tyracek 2001) and the Danube/Srvatka (Kukla 1975, Gábris and Nádor 2006). Loess sequences and palaeosols covering the terrace surfaces provide supplementary stratigraphical means to elaborate the non-glacial succession in Central Europe.

3.2.5 Lacustrine sediments deposited in lakes, peat bogs and abandoned meander channels

Lake sediments form in small-scale basin settings which owe their origin to a variety of geological and geomorphological aspects. Their formation may be due to:

- Tectonics such as the lakes of intramontane and non-marine subsidence basins (e.g. Tenaghi Philippon and Ioannina),
- Various kinds of glacial processes like the lakes in glacial outwash or till,
- Periglacial phenomena such as remnants of pingo's and thermokarst,
- Volcanic activity (e.g. the crater lakes in the Central Massif),
- Subsidence and collapse structures resulting from salt dissolution or karst,

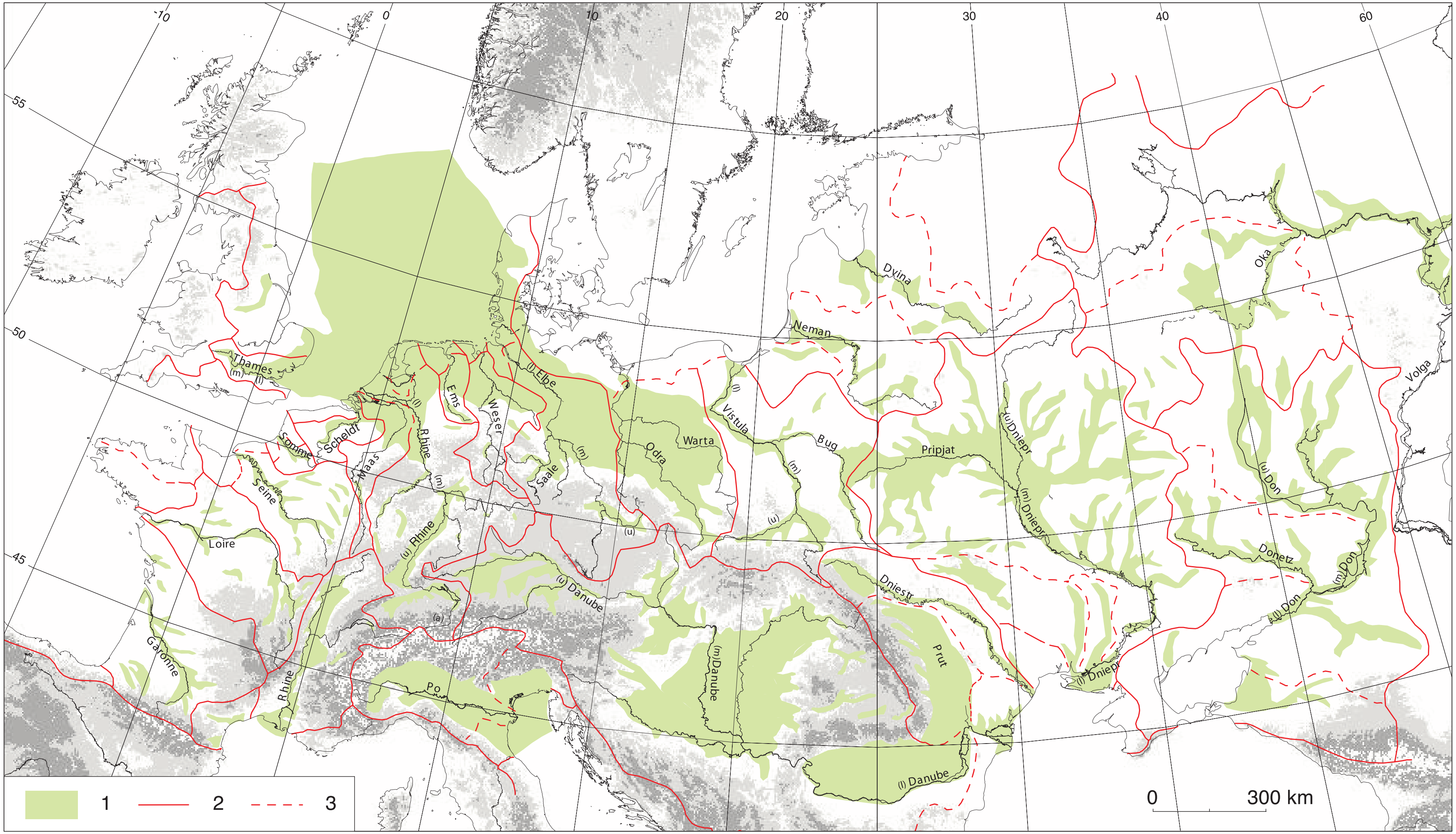


Figure 3.7 Distribution of fluvial sediments and drainage basins in Northwest and Central Europe (from 'International Quaternary Map of Europe', UN/BGR 1965-1995). 1. extent of fluvial and deltaic sequences, 2. main water divide, 3. secondary water divide.

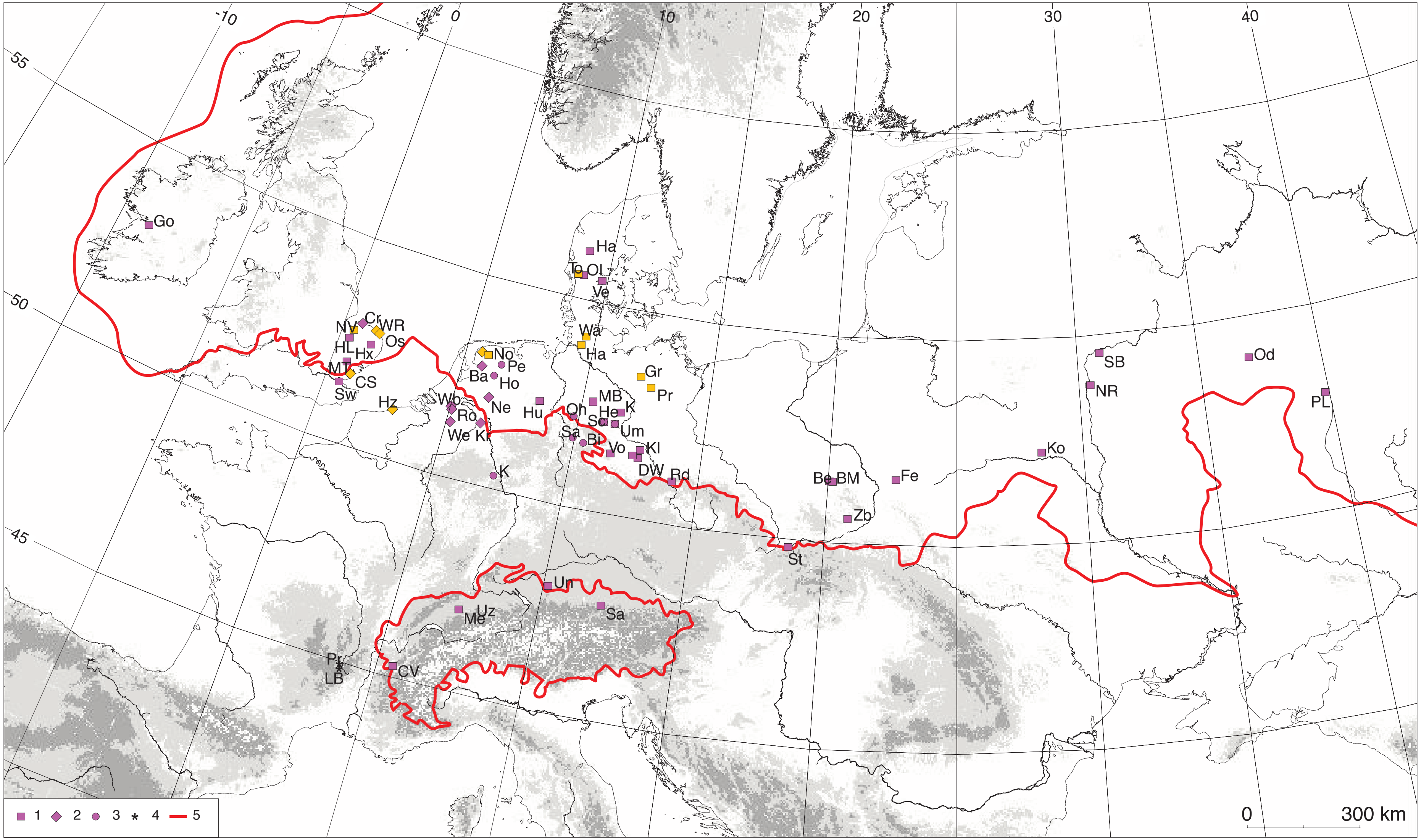


Figure 3.8 Distribution of lake and mire sequences in Northwest and Central Europe (from ‘International Quaternary Map of Europe’, UN/BGR 1965-1995). 1. glacial lake sequences, 2. fluvial lake sequences, 3. solution lake sequences, 4. maar lake sequences (LB: Lac du Bouchet, Pr: Praclaux), 5. maximum limit of glaciation during the Pleistocene.
Saalian Stage: Ba: Bantega, Bi: Bilshausen, Ho: Hoogeveen, Ka: Kärlich-Seeufer, Sc: Schöningen, Ve: Vejlby, Zb: Zbojno; Holsteinian Stage: CS: Clacton-on-Sea, Gr: Granzin, He: Hetendorf, HL: High Lodge, Hx: Hoxne, Hz: Herzeele, Kl: Klieken, Kr: Krefeld, MB: Munster-Breloh, MT: Marks Tey, NV: Nar Valley, Ne: Neede, Pr: Pritzwalk, Sw: Swanscombe, To: Tornskov, Wa: Wacken; ‘Cromerian Complex’ Stage: Cr: Cromer, Fe: Ferdinandov, Ha: Harreskov, Hu: Hunteburg, Ko: Kosteschi, NR: Nidschinski Row, No: Noordbergum, Ol: Ølgod, Os: Ostend, Od: Odintsovo, PL; Poljno Lapino, Ro: Rosmalen, Sk: Shklov, Vo: Voigtstedt, We: Westerhoven, WR: West Runton.

- Abandonment of meander channels in fluvial depositional environments.

Small shallow lake depressions in unlithified sediments are readily infilled, but even large depressions with large catchments can receive a sufficiently large clastic sediment input to have only a short life before being infilled (Mangerud 1983). Larger, deeper lakes are generally subject to strong circulation which may generate internal patterns of erosion and deposition which vary in time and space and which are subject to highly erosive turbidity flows. Thus, in such lakes, even where cores can penetrate deep enough to sample long time spans, they do not always sample continuous sequences and may only reflect major events. Selection of lake sites and the location of boreholes within a lake sequence is very important to obtain the relevant information.
Sequences particularly from lakes with a closed system and with a small catchment show basically continuous, low energy deposition. Although in the case of glacial lakes they usually do not exist for a long period of time, their preservation potential is high, being sandwiched between (or overlying) glacial deposits. Such lake sediments are important sources of high resolution data on palaeoenvironmental conditions, be it local to regional.

[a] Lake and mire sequences in formerly glaciated and periglacial areas

In the glaciated zone, the remodelling of the landscape during glacial cycles tends to ensure that no lake basin will have a continuous record of change longer than the last glacial cycle in the area. Small lakes within the glaciated zone may accumulate distinctive organic sequences with a good correlation potential before being overridden by subsequent ice or filled in by periglacial sediment. For instance, warm-stage pollen sequences attributed to the Eemian Stage, were overrun by the ice-sheet of the last glacial cycle, and correlate well with lake sequences which lie beyond the last glacial maximum but post-date earlier glaciations (*cf.* Jessen & Milthers, 1928). Some of these lakes and mires, are particularly valuable in that the initiation of sedimentation within them cannot pre-date deglaciation. They are the ideal sequences to bridge the time between two subsequent glaciations. If organic sedimentation starts early in their life, they can be a valuable guide to date for example the last deglaciation, for which ^{14}C dating is available (Mangerud 1991), or they can indicate similar 'Late Glacial/Dryas' fluctuations during earlier deglaciations, for example in the sequence of Neumark-Nord in eastern Germany.

Some well established local Middle and Late Pleistocene interglacial stratotypes and their correlatives include, from youngest to oldest:
- The Eemian lake sequences at Amersfoort (Zagwijn 1961) in the Netherlands, Hollerup in Denmark (Andersen 1965), Bobbitshole in Great Britain (West 1957), Grande Pile in France (Woillard 1979), Neumark-Nord, Gröbern and Grabschutz in Germany (Mania 1990, Litt 1994) and others,
- The Holsteinian lacustrine sequences at Pritzwalk and Münster-Breloh (Germany: Erd 1973 respectively Müller 1974), Hoxne and Marks Tey (England: Turner and West 1968, Turner 1970), and Tornskov (Denmark: Andersen 1965),
- The Cromerian lake sequences are less well established due to uncertainties in stratigraphical position, although pre-Elsterian in the glaciated areas, and similarities in pollen assemblages. In the Netherlands, Zagwijn (1975, 1996) identified four warm stages are from organic beds in fluvial deposits (I-Waardenburg, II-Westerhoven, III-Rosmalen, IV-Noordbergum; *Fig. 3.1*). Further evidence comes from the former lake sequences at Bilshausen (Germany: Müller 1965, revised in Bittmann and Müller 1996 and in *section 5.2.3*), Harreskov (Denmark: Andersen 1965) and Ferdinandov (Poland: Janzcyk-Kopikova 1975).

[b] Lake sequences from the extraglacial zone

Beyond the limit of the last glaciation, surprisingly few lakes containing a sedimentary record longer than the last glaciation have been found in Europe. The challenge is to correlate these mid- and south-European sedimentary records, extending much further back in time, with those of the former glacial lakes.
Lakes which have formed in small tectonically-controlled basins like Tenaghi Philippon and Ioannina (Greece) and in the craters of small Cenozoic volcanoes in Europe (primarily in France, Italy and Germany), the so-called maar lakes, are ideal sites for long sedimentary sequences spanning several interglacial-glacial cycles. They have very small catchments so that relatively little minerogenic material is introduced to them. However, they are ideal pollen traps. Due to the relatively low sedimentation rates, relatively short cores might represent a long period of time, whilst limited bioturbation and the dominant organic input ensures a rich source of palaeoenvironmental and palaeoclimatic information. The lake sites in eastern France at La Grande Pile and Les Echets currently have pollen sequences which extend back to about 140 ka (Guiot *et al.* 1989, 1992), although the base of the sedimentary sequence has not yet been reached. They have the potential to go back much further in time as in the maar lake sequences of the Velay region in the Central Massif, where at Lac du Bouchet (Reille and De Beaulieu 1995, Tzedakis *et al.* 1997, Reille *et al.* 2000) a lake sequence down to the base of the supposedly Holsteinian corresponding Praclaux warm Stage could be cored (*Fig. 3.9*).

3.2.6 Other sediments from local-scale subenvironments

Some other characteristic sediments in the terrestrial record, although usually local in their occurrence, will be briefly mentioned here: volcanic ash layers, secondary carbonates (travertine, spelaeothems) and cave deposits. They often comprise marker beds which may contain bio- and chronostratigraphical information of decisive stratigraphical interest.

[a] Volcanic sediments

Quaternary volcanic fields are known from several regions in Europe of which those in the Eifel region and the Central Massif have provided Middle Pleistocene chronostratigraphical units dated by tephrochronology. Of interregional stratigraphical importance is the set of tephra beds which enables dating of the Middle and Late Pleistocene loess/palaeosol and terrace sequence in the Middle Rhine Neuwied basin (e.g. Van den Bogaard and Schminke 1990). Further downstream, in the Lower Rhine Embayment and in the Netherlands, as part of the glaciated southern North Sea basin, their incorporated and fluvially transported heavy minerals are used as lithostratigraphical markers and as indirect dating and correlation tools. Unfortunately, dating and chrono-correlation are not unequivocal in these regions which will be discussed in more detail in *chapter 5*.

[b] Secondary carbonates

Carbonate sediments are formed by cementation and precipitation from springs, lakes or rivers in limestone areas. They may be of thermal or cold origin, locations of which are reviewed in Pente-

cost (1995). Travertine forms as a dense cryptocrystalline calcite occurring as subhorizontal beds in river valleys. Tufas form through precipitation on growing plants and commonly are soft and porous. Their stratigraphical value lies in the fact that they indicate relatively warm climate conditions and that their incorporated fossil remains provide often detailed local palaeoenvironmental information (e.g. Bilzingleben in eastern Germany: Mania 1973, 1993, also Hitchin in England: Kerney 1959).
Spelaeothems are subsurface carbonate precipitates found in cracks and caves. Although stratigraphically unimportant, their chemical analysis may yield local records of oxygen isotope fluctuations (*section 3.4.3*) which may be helpful in local palaeoclimate reconstructions.

[c] Cave and rock shelter deposits

Caves are the result of various processes of groundwater solution and part of karst geomorphology in limestone areas or may occur as rock shelters or abris in other bedrock. Cave deposits are highly variable and of local origin although extraneous sediments may have been brought in by water, wind (e.g. loess) and gravity action.
Caves and rock shelters are of special interest for palaeontologists and archaeologists since animals and man sought shelter in caves. Many important Palaeolithic sites are located in these environments where calcareous preservation conditions are generally good. Their chronostratigraphical position, however, generally remains indistinct because of the very low correlation potential.

3.2.7 Syn- and post-depositional structural features

Once laid down, sediments are prone to all kinds of geological and chemical processes that may alter their original properties; by collapse, slope processes/mass wasting, weathering and soil formation, compaction, diagenesis and glaciotectonics. Two features will be briefly discussed because they are indicative for climatic change or specific climatic conditions during non-depositional time intervals:

[a] Cryogenic structures

Many characteristic cold climate features may be found in any area with regular frost activity and are not indicative of any thermal limit. Cryogenic structures produced by intense frost activity are commonly found in aeolian periglacial and other cold climate sediments.
Involutions and features indicating permafrost occur at particular horizons and hence indicate contemporaneous climate. Their appearance and distribution are a valuable complementary source for palaeoclimatic interpretation (Strunk 1983; Karte 1987). Nevertheless, one should be careful with temporal interpretations because they may be post-sedimentary phenomena.
Sedimentological and morphological characteristics diagnostic of the occurrence of continuous permafrost[17] are:

- Ice- and sand-wedge casts and polygons, the most strict indicators of permafrost,
- Pingo-remnants.

Both are superficial permafrost phenomena and do not indicate to its thickness. Reconstruction of Pleistocene permafrost is so far focussed on the Weichselian cold Stage and have been undertaken in the Netherlands and Belgium (Vandenberghe 1985, 1992; Haesaerts 1984), and eastern Germany (Eissmann 1981). Evidence of permafrost of Saalian age has, so far, been demonstrated by Eissmann (1981) in eastern Germany. Based on the cryogenic deformation of lignite layers in non-glaciated areas, he indicated a minimum permafrost depth of 50 metres, and suggested a maximum of more than 100 metres.

[b] Pedogenic structures: soil complexes

Soils can be sediments in situations where sediment is pedologically modified as it is laid down - e.g. in Chinese loess sequences. But generally they are not where sedimentation does not occur during soil formation. Soil formation is polygenetic and primarily a function of climate and topography and secondarily of substrate, maturity (time) and organic activity (Catt 1988, 1995). Most soils are soil complexes and are composed of different soil types. One should therefore be cautious with soil-stratigraphy because of the lack of knowledge of the complex relationships between soil properties and climatic factors (Haesaerts and Mestdagh 2000). Moreover, the ability to estimate the lengths of soil-forming intervals is restricted. Nevertheless, buried and surface soils, in particular in aeolian sequences, hold great potential for Quaternary palaeoclimatic interpretation since they indicate warm vegetated climatic conditions. As discussed in *section 3.2.2*, palaeosols are used as marker events within the long extraglacial loess records of Eurasia, indicating periods of non-deposition during warm palaeoclimatic stages. Soil complexes found in the Central European type regions are of two main groups (Kukla 1977):

- Biogenic steppe soils (chernozems), showing accumulation of organic matter but little chemical change in mineral matrix, and
- Brown leached soils with evidence of in-situ redeposition of carbonate, iron or manganese as well as clay plasma, but reworked by pedofauna to only a minor degree.

Parabraunerde-type soils with a Bt-horizon are the product of deciduous forests. They are the most common found warm-stage palaeosols in the extraglacial areas of Northwest and Central Europe and have a high stratigraphical value.

3.3 Biostratigraphical datasets in terrestrial sequences

Preserved fossil remains in different Pleistocene sedimentary environments provide relative, local and regional chronologies as well as information for palaeoenvironmental and -climatic interpretation. The most useful fossil groups are those that can be identified readily on a species level and those that have undergone significant evolutionary change (Bridgland and Maddy 2002). Subdivisions based on evolutionary change in flora and fauna are restricted, however, for such a short time period such as the Middle Pleistocene. In some cases it is possible to indicate a general age or constrain a time-range on the basis of extinction or of migration patterns of indicator species.

Assemblage biozones have proven to be a powerful stratigraphical tool in the recognition of Pleistocene palaeoclimatic stages in Europe (*section 3.1*). They are useful event markers when applied on the basis that the ecological response of organisms to a complex sequence of environmental and climatic changes is unlikely to generate identical assemblages at different periods. This assumption is most likely to be true for the relatively short warm stages with a rich and varied biota, but not for the long cold periods in which biotic production and biological diversity is reduced[18]. In particular, it is the rich warm stage floras and microfaunas which are assumed to be compositionally unique and which have potential to form the essential correlation horizons for the Pleistocene stratigra-

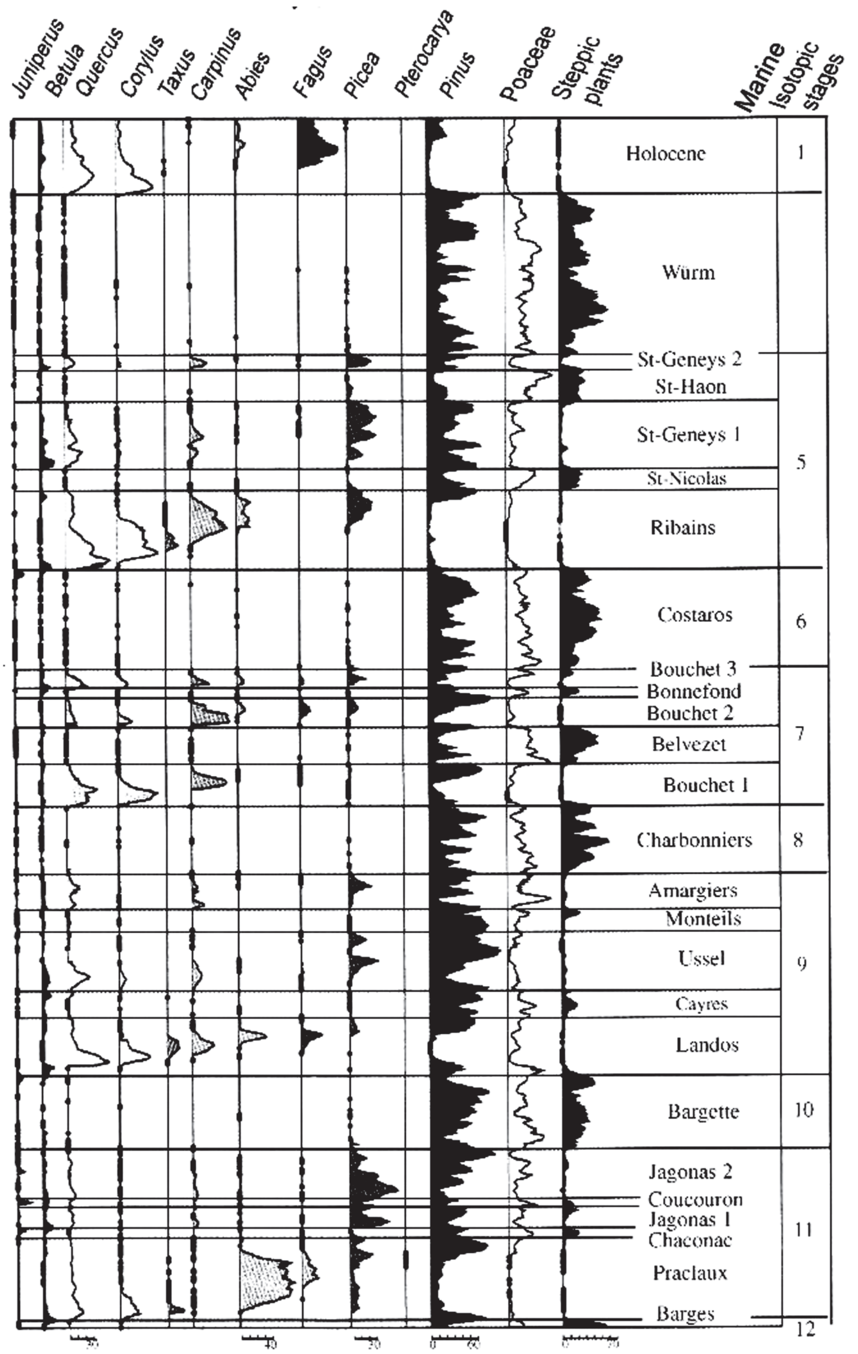


Figure 3.9 Late Middle Pleistocene pollen stages and interpreted stratigraphical relations of the maar lake sequences in the Central Massif (Reille *et al.* 2000). A tephra within the Amargiers pollen assemblage zone is dated at 287 ka (Reille and De Beaulieu 1995).

phy. One should, however, always consider its relation with the depositional environment, because of effects of reworking and inwash. Sedimentary environments and main type localities in which fossil remains are preserved have been reviewed in *section 3.2*.

A brief outline of the use and limitations of various fossil groups which play a role in distinguishing between regional stratigraphical sequences and events is given below.

3.3.1 Palaeobotanical evidence

The Northwest European continental Middle Pleistocene stage succession is largely based on a sequence of palynologically distinctive stratotypes reflecting warm-temperate climate conditions. They are generally from lacustrine environments from different origin (summarised in *section 3.2.5*) intercalating the glacial sequences. Pollen and spore assemblages are also recovered from soils and even from marine environments. Many vegetational histories only record parts of the time represented in any individual warm climatic stage. They may in later phases be influenced more and more by edaphic and other local factors which hamper correlation. Regional variations and correlation over long distances or with fragmentary data should be regarded cautiously.

Zonation schemes of vegetational development (*cf.* Turner and West 1968) enables subdivision within and between warm stage forest pollen spectra from Northwest Europe. A full climatic cycle comprises four substages of forest vegetation assemblages which, for example, have been applied to subdivide the pollen diagram for the British Hoxnian temperate Stage from Marks Tey (*Fig 3.10*):

- Substage I (Pre-temperate phase): boreal vegetation lacking

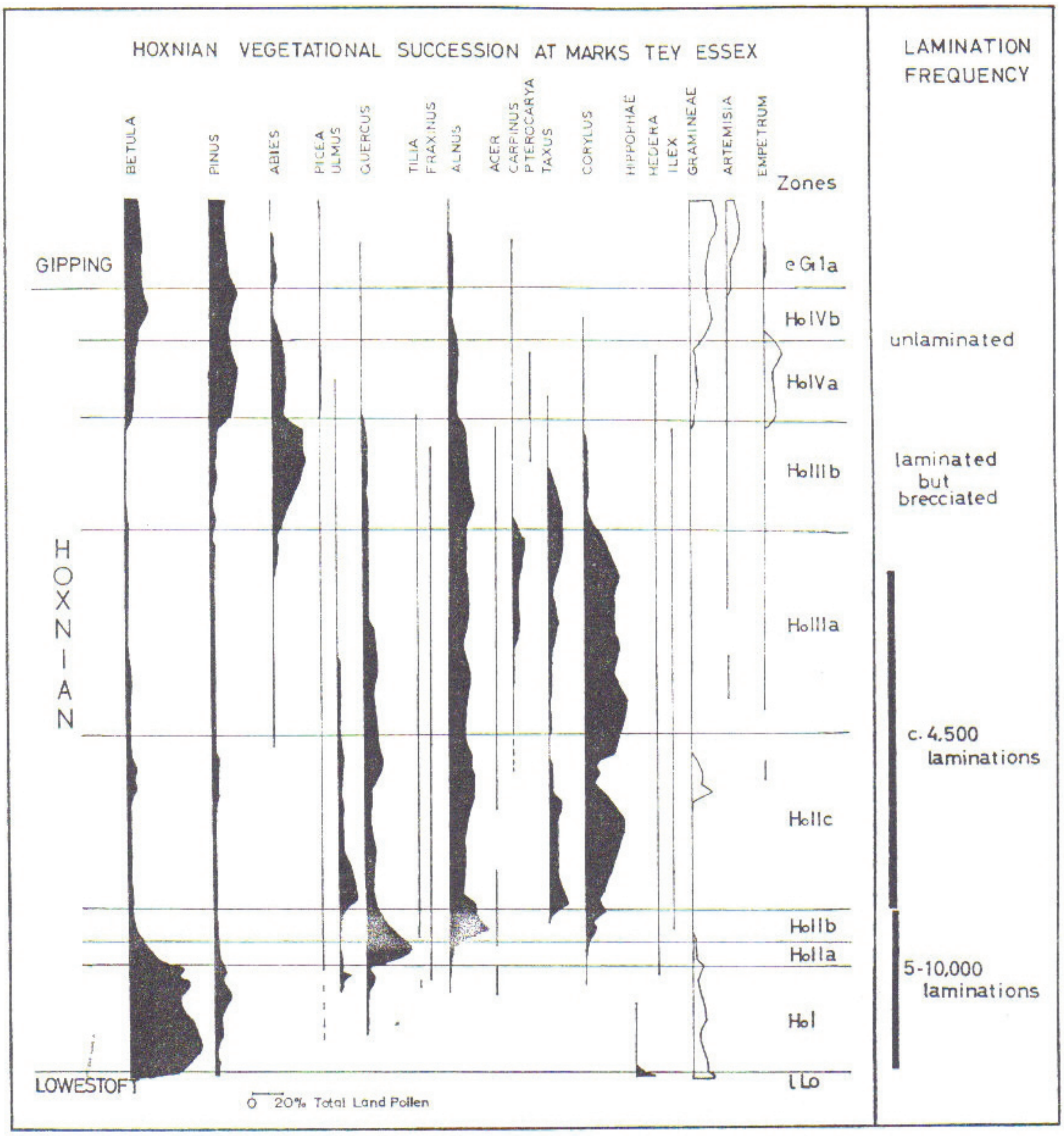


Figure 3.10 (Composite) pollen record typical for the vegetational succession of the Hoxnian temperate stage from Marks Tey (after Turner 1970).

temperate elements, but with evidence of birch and pine forest,

- Substage II (Early temperate phase): phase of deciduous forest and broadleaved trees with development of mixed oak (QM) forest type assemblages,
- Substage III (Late temperate phase): phase of deciduous forest and broadleaved trees with presence of late expanding trees like *Carpinus* and *Abies* at the expense of the mixed oak forest,
- Substage IV (Post temperate phase): in which there is a limited representation of thermophilous, broadleaved species and boreal forest is recurring.

Despite some limitations (De Jong 1988) such pollen assemblage zones provide distinctive markers of individual climatic phases and may indicate in which part of the climatic cycle a sediment was laid down. Also the behaviour of certain tree species in relation to zonal climate aspects can be studied. In general, the limited boreal-type palynofloras and the 'open tundra' vegetation assemblages (with shrubs and herbs) of stadials and interstadials inhibits their use as indicators of specific stages.

Pollen assemblages may contain specific indicators for distinguishing between temperate-climate periods. Examples of indicator species in Middle Pleistocene warm Stage pollen assemblages are:

- *Eucommia*, *Pterocarya* and *Celtis*, indicating the decrease in Tertiary assemblages and relict pollen in Middle Pleistocene deposits. *Pterocarya* for example occurs in pollen diagrams up to and including the Holsteinian warm Stage,
- The water fern *Azolla filiculoides*, indicative of a pre-Eemian age,
- *Abies*, indicative of oceanic influence in the Northwest European lowlands accompanied by high sea-level stands such as occurred during the Holsteinian and Eemian warm Stages (Zagwijn 1989, 1992).

Of major interest is how the independent, scattered and fragmentary pollen evidence from the glaciated Northwest European areas correlate with the long continuous pollen records of Tenaghi Philippon in Greece (Mommersteeg *et al.* 1995) and Lac du Bouchet/Praclaux in southern France (Reille and De Beaulieu 1995, *Fig. 3.9*). Both comprise local proxy palaeoclimate records which are a link to the marine isotope stratigraphy. Comparison, by trend matching, is now well established for the late Middle and Late Pleistocene (Tzedakis *et al.* 1997, Tzedakis *et al.* 2001). Additionally, pollen from marine cores off-Portugal have been directly correlated for MIS 5 (Sánchez-Goñi *et al.* 1999) and for MIS 11 (Desprat *et al.* 2005). On the basis of the last appearance datum (LAD) of *Pterocarya* pollen in the Praclaux warm Stage, Reille and De Beaulieu (1995) and De Beaulieu *et al.* (2001) suggest that this stage probably corresponds to the Holsteinian of northern Europe. This implies correlation with MIS 11 and the existence of two additional warm climate events within the late Middle Pleistocene Saalian Stage, preceding the Saalian glaciation sensu stricto.

3.3.2 Palaeozoological evidence

Faunal remains in the terrestrial sediments play a less important role in Middle Pleistocene biostratigraphy than does pollen. Distinctive fossil assemblages suitable for biozonation and chronostratigraphical subdivision are rare due to the fragmentary evidence, the often low frequency and the lack of distinguishing criteria. Moreover, their occurrence is often restricted to specific sedimentary environments, which makes correlations difficult. Biostratigraphical correlations on the stage of evolutionary change, extinction and migration of species, the latter factors are related to climatic change, are only possible for vertebrate and molluscan faunas and have given some subdivision in the Early and Middle Pleistocene. Their value in reconstructing climatic change in most cases is limited to the assignment to warm or cold stages. Nevertheless, many sites in Northwest Europe have yielded rich faunal remains, some of which are used as regional stratigraphical markers (summarised in *section 4.4.2* and *Fig. 4.9*) reflecting environmental conditions.

[a] Vertebrates

The preservation of mammalian bones and teeth in sedimentary environments such as non-acidic lakes margins and mires, spring surroundings, loess regions and floodplains, have allowed reconstructions of fossil vertebrate assemblages in a number of local stratigraphies. A review of presence/absence data of mammalian assemblages in Middle and Late Pleistocene warm stage fluvial terrace sequences from Britain and the German uplands is compiled by Schreve and Bridgland (2002). From their correlations they suggest there is evidence for three warm events within the late Middle Pleistocene post-dating the Elsterian/Anglian glacial sequences.

Biozonation of fossil micromammal assemblages for the Middle and Late Pleistocene in Northwest Europe (Van Kolfschoten 1990, Horacek 1990) reveal temporal change in molar composition of some species. Of particular importance for the early Middle Pleistocene stratigraphy is the boundary of the rodent species *Mimomys savini* and *Arvicola terrestris cantiana*. A further marker forms the gradual change in the thickness of the enamel in the *Arvicola* lineage[19] that took place between the Elsterian and Saalian glaciations (Van Kolfschoten 1990).

[b] Molluscs

Although they supply the raw material for amino acid racemization, the Mollusca are of limited value for dating purposes. They occur in all kinds of environments, land, freshwater and marine, and are therefore powerful palaeoenvironmental and palaeoclimatic indicators.

Terrestrial gastropods in loess sequences in Central Europe (Kukla 1977) and at Achenheim, France (Rousseau & Puisségur 1990) have provided valuable information on loess and soil environments. Based on the work of Lozek (1965) several shell faunal assemblages can be distinguished, ordered in decreasing warmth requirements:

- *Banatica* fauna, indicator species *Helicigona banatica*, a warmth-loving assemblage, found in spring travertine among others,
- *Pomatia* fauna, indicator species *Helix pomatia*,
- *Tridens* fauna, indicator species *Chondrula tridens,*
- *Striata* fauna, indicator species *Helicopsis striata*,
- *Pupilla* fauna, with *Pupilla muscorum* and other *Pupilla* species, as well as *Succinea oblonga*, as indicator representatives,
- *Columella* fauna, with *Columella columella* as indicator species and a biomarker for extreme cold climates.

Freshwater molluscan fauna of river and lake deposits provide valuable biostratigraphical constraints:

- The occurrence of *Viviparus diluvianus* (= *Paludina diluviana*), characteristic of the Lower Elbe fluvial sequence, is attributed to the Holsteinian warm Stage,
- *Corbicula fluminalis,* another warmth-demanding species, is appearing in several warm-stage fluvial sequences assigned to the Holsteinian Stage and following warm Saalian Stage events, but is absent from Eemian-age deposits (Meijer and Preece 2000). Just as *Viviparus diluvianus*, the presence of *Corbicula fluminalis* is a marker in the fluvial terrace deposits in the uplands of Central Germany (the 'Mittelterrassen') intermediate of the Elsterian and Saalian glaciations.

The use of marine molluscan fauna assemblages is limited because of the lack of distinguishing criteria (Meijer and Preece 1995).

3.4 Chronostratigraphical control on terrestrial Middle Pleistocene sequences

3.4.1 Relative age markers

The age determination of distinctive lithostratigraphical units and biostratigraphical assemblages is based on relative age criteria and often the only chronological method applicable. With the exception of the last 40-50 ka, for which dendrochronology and radiocarbon dating provide reliable timing, older chronostratigraphical land-based subdivisions are mainly based on palynological criteria and superposition. Marine subdivisions, by contrast, are based on (micro-)palaeontological criteria. Layers of volcanic ash of known age and provenance are used as lithostratigraphical markers in the Late and Middle Pleistocene sequence of the Rhine valley. Interglacial marine and fluvial stratigraphical units, containing molluscan fossils from Northwest Europe, have been correlated by aminostratigraphy (Bowen and Sykes 1988). Nevertheless, this chemical method of amino-acid racemisation of molluscan shells is still of limited use for regional correlations of late Middle and Late Pleistocene transgressions. Recent developments have enhanced the reliability of the method (Sykes *et al.* 1995) and has given better results for several fluvial and estuarine sequences (Meijer and Preece 2000, Bridgland and Maddy 2002).

The duration of individual lithostratigraphical units can be determined by the counting of incremental accumulations of laminated sediments found in lakes. Of particular importance are the varve chronologies from proglacial rhythmites in Sweden (de Geer 1912) and from organic algal blooms at Marks Tey in England

(Turner 1970). The latter showed a duration of the Hoxnian temperate Stage in the order of 15-20 ka.

3.4.2 Palaeomagnetic evidence

Palaeomagnetic investigations of Pleistocene sediments provides a general, low resolution timescale requiring long fine-grained sedimentary sequences spanning hundreds of thousands of years. Of particular importance is the position of the Matuyama/Brunhes Chron boundary, the last global geomagnetic reversal, which is taken as the base of the Middle Pleistocene. It is dated at *c.* 780 ka which leaves a considerable gap with the aforementioned dating limits. Some brief excursions are noted within the Brunhes normal Chron. Langereis *et al.* (1997) in their review on short reversal excursions in the Brunhes Chron mention four significant events within the Middle Pleistocene: Calabrian Ridge (CR)0 dated at around 260 ka, CR1 dated at around 320 ka, CR2 dated at around 515 ka and CR3 dated at about 570-575 ka.
Conversely, magnetostratigraphy, in combination with magnetic susceptibility and related techniques, has yielded high-resolution records from long Chinese and Eurasian loess sequences during the last decade which show close similarity with the marine isotope record (Derbyshire (ed.) 1995).

3.4.3 Geochronometric dating

In addition to the relative age methods, whereby a sequential order to the deposits is established, different geochronometric methods such as thermoluminescence (TL, OSL, IRSL and GLSL[20]), Uranium (U)-series, Uranium/Thorium (U/Th)-dating, Electron Spin Resonance (ESR), Potassium/Argon (K/Ar)-, Argon/Argon (Ar/Ar)- and Fluorchloride/Potassium (FCl/P)-dating, are applicable as time-stratigraphical criteria (Geyh and Schleicher 1990). Although essential for the Pleistocene chronology, reliable absolute age estimations for the European continent, however, are scarce up till now. Accuracy of the dating techniques decreases with time. The chronological framework for the Holocene and Late Weichselian time has come from radiocarbon dating. The age determination beyond its limit of 40-50 ka remains troublesome because the techniques are limited, either as a consequence of their time-range or of their sensitivity to particular materials. The maximum dating limit of most of these techniques appears to be 300-400 ka or less.
The dating techniques which are used to calibrate both Middle and Late Pleistocene stratigraphies in Northwest Europe are:

[a] Luminescence dating

The luminescence methods are applicable to grains of quartz and feldspar in silty Pleistocene sediments, e.g. loess. Dates from loesses are obtained from the Rhine Valley area, northern France and Poland. Most determinations are from Weichselian loesses, and give meaningful results. Reliable dates over about 125 ka (pre-Eemian), however, have not yet been obtained, because TL age differentation of Saalian loesses is still not possible (Frechen 1991, 1993). Older (minimum) TL dates, however, are known from burnt flint at Maastricht/Belvédère in the Netherlands (270 ± 11 ka; Huxtable & Aitken 1985) and aeolian sediments and tills from Poland (Mojski 1985; Rechowski 1986). The latter TL determinations, which apparently extend into the Early Pleistocene, are questionable due to the various analysis techniques which are used. Luminescence methods such as IRSL and GLSL are used for

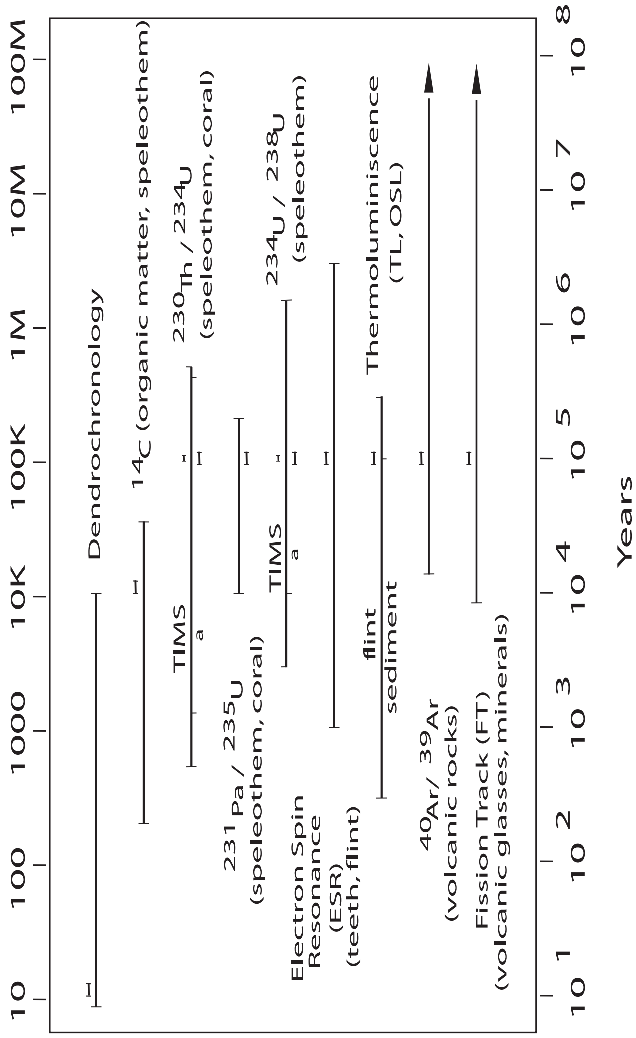


Figure 3.11 Quaternary dating methods and their time-ranges (after Stein and Linse 1993).

dating of fluvial sequences.

[b] Uranium Series dating

U-series and ESR dating can be a suitable method as far back as the Middle Pleistocene, although the dating limit appears to be about 300-400 ka. The method is most applicable to precipitated carbonate deposits and concretions. Fairly accurate dates can be obtained from cave deposits (speleothems). Presence and growth of these deposits occur during non-glacial climatic conditions, which can be derived from their oxygen isotope content. Dates are available from NW England (Gascoyne *et al.* 1983), suggesting a warm episode between 250 and 230 ka, and from Norway (Lauritzen 1991). Travertine from the Bilzingsleben terrace II has been dated by U/Th at about 320-350 ka, ESR dates range from 280-415 ka (Schwarz *et al.* 1988 in Mania 1993). Travertine from the deposits of the terraces III and IV at Verteszöllös (Hungary) have been U/Th-dated at respectively 360 ka and 200 ka (Hennig *et al.* 1983). U-series of the authigenic carbonate content in Hoxnian warm Stage deposits at Marks Tey revealed minimum dates around 400 ka (Rowe *et al.* 1999).
Moreover, estimates of the age of fossils from warm-stage marine and lacustrine deposits have been determined by U/Th- and ESR dating. Molluscan shells in Holsteinian marine deposits have been

determined from Herzeele in northern France (marine unit III: minimum age 326 ka) and from Wacken in Schleswig-Holstein. Based on U-Th dates, Sarnthein *et al.* (1986) assumes that the Wacken marine deposits are older than 350-370 ka, whereas Linke *et al.* (1985) report ESR-data from the same Holsteinian marine deposits at respectively 195 ± 25, 223 ± 25 and 218 ± 25 ka.
Dating of Late Pleistocene peat deposits from sites in Greece (Tenaghi Philippon), in the British Isles and in Germany (Schöningen), using U-series, have been reported by Heijnis and Van der Plicht (1992). Their obtained ages for the Weichselian are in good agreement with the chronostratigraphy and TL dates. Some dates probably require corrections because of problems of undefined open-system behaviour of the peat. The organic deposits of the Schöningen warm event, for example, have been dated to between 180-227 ka (Heijnis, 1992, p.132).

[c] Potassium-Argon dating

The K/Ar- and Ar/Ar-methods are applicable to volcanic material and glauconite-rich sediments that are interbedded within other depositional sequences. Several dates are available from tephras interstratified with the Middle and Lower Rhine loess and terrace sequences, dealt with in more detail in chapter 5. They are related to different phases of volcanic activity in the Eifel region, which started at about 570 ka (Frechen & Lippold 1965) or probably even earlier, before 650 ka (Van den Bogaard and Schminke 1990). The tephras provide geochronological control within key-stratigraphical sequences, such as those from Kärlich and Ariendorf (Van Kolfschoten and Turner 1996) and confine the age of fluvial sequences further downstream of the river Rhine.

[1] *Since the establishment of the modern Ice Age concept by Penck (1879) and Geikie (1895) among others (section 1.3.1).*

[2] *= Gippingian.*

[3] *Because of the authoritative status of the Alpine model one may (even) state that at least up to the 1950s a kind of reinforcement syndrom existed whereby apparent incompatibility and discrepancies (simply) were not considered or accepted because it would bring on too much confusion.*

[4] *Co-controlling accommodation space, sediment supply, grain-size and sedimentary processes.*

[5] *The Alpine glacial stages, for example, are still widely used in many mountaineous regions.*

[6] *Palaeotemperature ranges intimately related to the effect of precipitation.*

[7] *For example that cf. Köppen's classification.*

[8] *cf. Bowen 1978.*

[9] *Abbreviations are explained in section 3.4.3.*

[10] *Although they are relatively small-scale compared to subseries of pre-Quaternary systems.*

[11] *An age of 812 ka has been proposed for the Brunhes/Matuyama boundary (Langereis et al. 1997), based on rock-magnetic and geochemical properties from marine hemipelagic sediments with intercalated sapropel and tephra layers in the Ionean Sea.*

[12] *Till is a glacial diamicton and generally is only used as a genetic term.*

[13] *Downcutting of new channels and glacial basins and river diversion.*

[14] *Dumping of major sediment masses to form moraines, sandrs, kames and kettle holes.*

[15] *Typical leached soil types are 'brown earth', 'parabraunerde', 'braunlehm' and 'rotlehm' with clay-enriched B-horizons.*

[16] *Since solifluction is not necessarily confined to cold climates, the term 'gelifluction' has been proposed to describe solifluction associated with frost action (e.g. Washburn 1979).*

[17] *Requiring polar climates where mean annual temperatures are below -20 C respectively -80 C.*

[18] *Although faunal diversity was relatively high in the mammoth steppe in Alaska (Guthrie 1990).*

[19] *The enamel differentiation quotient (SDQ-value) of Arvicola molars decreases during the late Middle Pleistocene. This trend is used for correlation purposes.*

[20] *TL: Thermoluminescence, OSL: Optical Stimulated Luminescence, IRSL: InfraRed Stimulated Luminescence, GLSL: Green Light Stimulated Luminescence*