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The History of the Great Northwest European Rivers During the Past Three Million Years [and Discussion]

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The history of the great northwest European rivers during the past three million years

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This paper is based on a review of the histories of the Rivers Elbe, Saale, Weser, Rhine, Meuse, Scheldt, Thames, Somme and Seine. Two further rivers no longer in existence, the Baltic and Channel rivers, are also included. The histories of these rivers illustrate how the interplay of tectonics and climate have influenced the northwest European drainage system through the late Cainozoic.

The foundations of the modern drainage system were laid in the Miocene when earth movements associated with Alpine orogenesis and the opening of the North Atlantic were at their height. In general, these early rivers occupied shallow valleys and transported only chemically resistant minerals and lithologies.

The Pleistocene was marked by the appearance of cold climates. These climates resulted in fluvial dissection of the landscape, which stripped first regolith, then fresh material derived by periglacial processes. This material accumulated in the river valleys as gravel and sand deposits, which make up the overwhelming bulk of Pleistocene fluvial sediments. The rivers generally adopted braided courses during cold stages. The deeply incised modern valley system has developed largely as a result of rapid climatic changes over the past 2.4 Ma or so.

Throughout this period the river system has undergone repeated adjustments in response to continental glaciation. These responses are discussed. Particular attention is paid to the impact of the Anglian–Elsterian glaciation that blocked the southern North Sea to produce a vast ice-dammed lake, the overspill from which initiated the Dover Straits.

By contrast, interglacial sedimentation comprises predominantly fine, often fossiliferous sediments with rivers normally adopting single-thread channels, while estuarine sediments were deposited in areas invaded by high eustatic sea levels. The impact of sea-level change on the length of rivers and their courses is considered.

Introduction

Northwest Europe can be regarded as an unified geographical region which has undergone a broadly consistent geological and climatic history over the past few million years. Today the region lies in the temperate climatic and vegetational zone but during the recent past it has been subjected to climatic change that has given rise to long periods of periglacial and even glacial conditions. In coastal areas, sea-level change, largely driven by glacial eustasy, has caused intermittent regressions and transgressions; in and adjacent to glaciated regions, isostatic and forebulge effects have also caused major land- and sea-level variations. All these variations are superimposed on longer-term trends in climate and tectonic evolution of the continent.

The drainage system of northwest Europe has evolved in response to the development of the continent during the late Cainozoic. The position of the region at the margin of the Eurasian continent has been crucial in determining the events that have influenced its geological evolution. By their nature, rivers owe their existence to, and are strongly influenced by, tectonic activity and climatic variability. The interplay of these two major factors is responsible for determining the form of the modern river system, which should be considered as the product of continual remodelling.

Because river valleys are major zones of terrestrial deposition, an understanding of their detailed geological record will provide a calibration of the geological evolution of the North Atlantic region as a whole throughout the past three million years. In addition, the intercalation of marine or estuarine sediments in the river sequences allows correlation with the offshore and potentially with the deep-sea facies. As will be shown below, there is considerable evidence of palaeoclimatic significance in river-sediment sequences; from this evidence it is clear that the northwest European rivers have functioned as an integrated system.

For the purpose of this paper the great northwest European rivers are taken to include the drainage systems of the Elbe, Weser, Rhine, Meuse, Scheldt, Thames, Somme and Seine. In addition, two further rivers which are no longer in existence, the Baltic River and the Channel River, will also be considered. These rivers have been selected because their individual histories illustrate the impact and interplay of the major influencing factors. They also provide an important transect across the continent from north to south, i.e. from the zone of maximum impact of glaciation, particularly during the Middle and Upper Pleistocene, to a zone which was apparently never glaciated.

In order to appreciate the evolution of the northwest European drainage system it is vital to review the history of each basin. To provide a uniform stratigraphical timetable for sequences discussed, The Netherlands' Neogene stratigraphical terminology has been adopted (Zagwijn & Doppert 1978; Zagwijn 1985). In this scheme the Plio-Pleistocene boundary is placed at the base of the Praetiglian Stage (*ca.* 2.4 Ma BP). This widely applicable boundary is thought to be 'natural' in northwest Europe because it marks the first arrival of true cold climates. It has also been used extensively throughout the region in stratigraphical schemes. Recent redefinition of the Plio-Pleistocene boundary at a younger level (cf. Aguirre & Pasini 1985) is not easily applied to already established successions and therefore, for the sake of simplicity, has not been used here.

Throughout the paper, discussions are restricted to those concerning fluvial deposits and events. Although deposition by other agencies is mentioned in passing where relevant, it is not considered in detail. This particularly concerns loess, thick accumulations of which, together with associated palaeosols, are found in many of the river valleys discussed. Loess is not fluvial in origin and therefore, although its importance to the understanding of climatic evolution is appreciated (Catt, this symposium), it has only been discussed when required for stratigraphical purposes.

Tectonic background

Throughout the Cainozoic era, northwest Europe has been affected by two major tectonic influences: fragmentation of the Eurasian–North American plate, and Alpine orogenesis. The interplay of these two factors has resulted in great complexity of both structural and depositional patterns. In essence these two processes work in opposing ways, i.e. the break-up of the northern hemisphere plate produces tensional or extensional features whereas Alpine mountain building, arising from continent-to-continent collision between the Eurasian and

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African plates, produces compressional features (cf. Ziegler 1978). The forces have led to activity along pre-existing tectonic lines that have been prevalent for long periods and in many instances have been inherited from previous phases (figure 1).

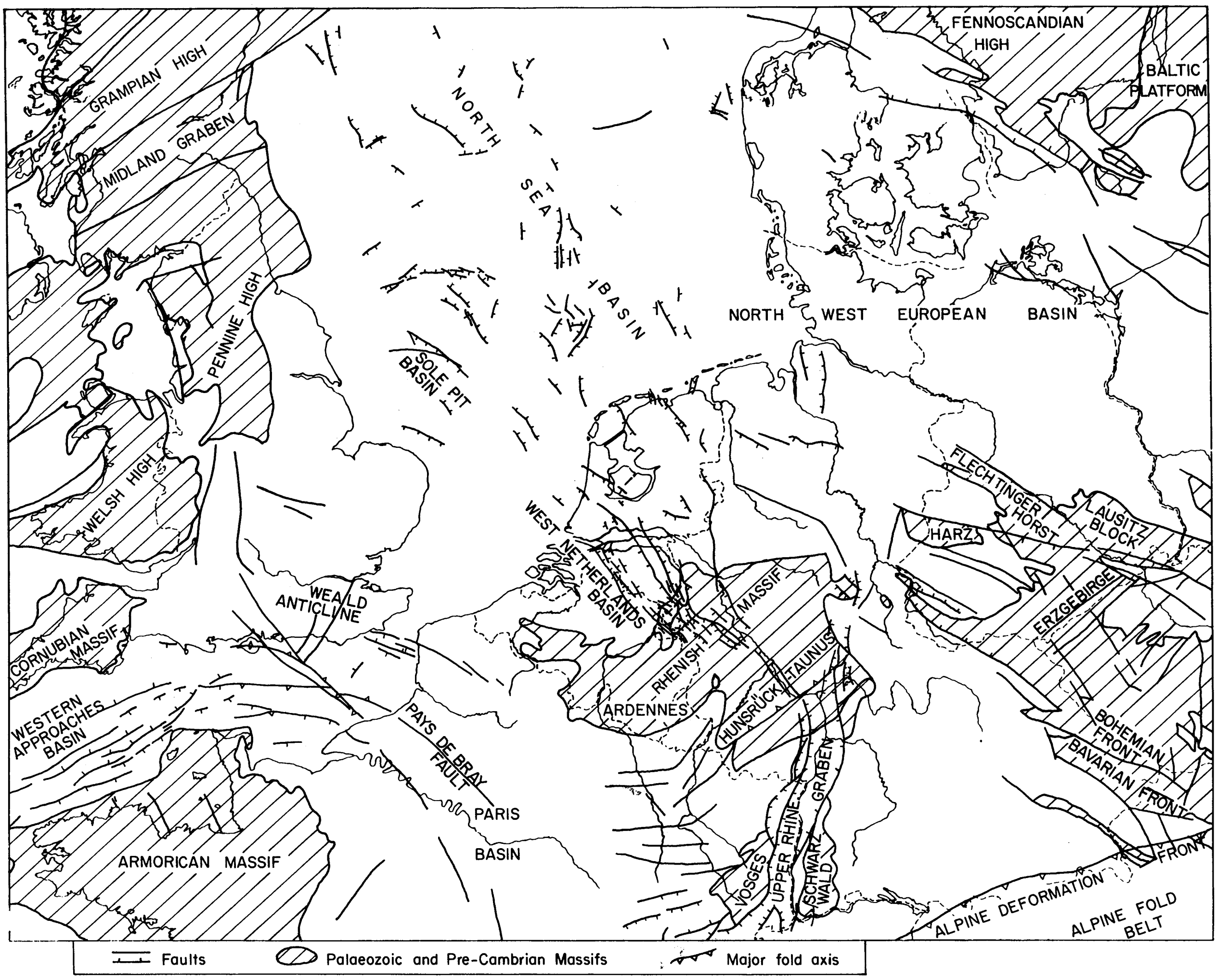

FIGURE 1. Tectonic map of the sub-Tertiary surface, modified from Ziegler (1978) with additions from Caston (1977), Zagwijn & Doppert (1978), Smith & Curry (1975), Quitzow (1974) and Ziegler & Louwerens (1979).

After the early Tertiary fragmentation of the Eurasian–North American crustal plate, the rift system of the North Sea Basin became inactive. This rift system was formed in the Mesozoic as a secondary rifting line in response to extensional stresses. Ziegler & Louwerens (1979) state that after the continental separation the crustal relaxation resulted in continued uniform basinal subsidence, which followed from a pre-existing pattern of differential subsidence initiated in the Jurassic (Ziegler 1978; Ziegler & Louwerens 1979).

The Channel may have had a similar origin, according to Smith & Curry (1975), in that ocean crust may possibly have been formed here, but that events affecting the northwest European continent may have deflected this zone southeastwards beneath the land mass. Later widespread uplift in post-Eocene times reached a peak in the Miocene resulting in inversion of the Hampshire and Channel Basins and the Bristol Channel and Western Approaches Trough, and updoming of the Pays de Bray Anticline in the Paris Basin (Ziegler 1978). This uplift was related to late phases of the Alpine orogeny. Local minor activity has continued into the Pleistocene (Bourdier & Lautridou 1974*a*).

By far the largest Cainozoic basin is the Northwest European Basin, which extends from Poland to the northern North Sea. This basin had become stabilized by the Miocene and subsequently subsided in an irregular manner (Ziegler 1978). This, together with the North Sea Basin, received huge volumes of sediment during the Cainozoic. The latter basin is estimated to contain up to 3500 m of sediment for the whole period (Ziegler 1978) compared with a thickness of over 1000 m from the Quaternary alone. This implies a tenfold increase in sedimentation during the Quaternary (Caston 1977). Its existence and infill would have been encouraged by the uplift of the Fennoscandian Shield that took place during the Pleistocene (Ziegler & Louwerens 1979).

In contrast to this basin, the foredeep basins of the Alps were much smaller and they subsided strongly during the Tertiary. Their infill was greatly deformed by late Alpine orogenic phases. Further, by late Tertiary times the basins had become uplifted and their sediments had been eroded as a result of post-orogenic isostasy of the mountain belt (Ziegler 1978).

According to the same author, the Rhine and Rhône Grabens form part of a late Cainozoic collapse system that began to subside during the Eocene and has continued intermittently since then. This subsidence is contemporaneous with the orogenesis such that, during formation of the Jura Mountains in the Pliocene, thrust sheets overrode the southern end of the Rhine Graben and the eastern margin of the Rhône Graben. The Rhine Graben extends north–northwest into the Lower Rhine Embayment as the graben and horst structure of the southern North Sea Basin (Zagwijn & Doppert 1978). There is, however, no evidence of reactivation of the North Sea rift system during the Late Tertiary. The Rhine Graben also continues northward into the North German Plain (Ziegler 1978). Moreover, rifting of the Rhine Graben is also accompanied by alkaline volcanism that continued until very recently.

Uplift of the Massif Central and the Bohemian Massif also occurred during the Neogene. In both cases this was accompanied by alkaline volcanism that began in the Oligocene and has continued up to the present. Dissection of the Alpine foreland rocks resulted from this uplift, together with that of the Vosges–Black Forest rift dome. This uplift also isolated the Paris Basin.

These important events have had profound effects on the fluvial system, which has undergone considerable adjustment and modification to adapt to the changes. As can be seen from a comparison of the tectonic and late Pliocene palaeogeography maps (figures 1 and 2) there is a close correspondence of major drainage lines and distribution of structural elements. For example, the Baltic River is aligned along the long axis of the Northwest European Basin, the central German rivers drain northwards towards this basin, the rivers Thames, Meuse and Rhine drain towards the North Sea Basin and the Seine drains towards the Channel and follows a major fault zone. In addition, the Rhine occupied the Central Graben region of the southern North Sea, Lower Rhine Embayment and the Upper Rhine Graben. Subsequent disruption of this pattern by later glaciation, sea-level change and river migration and capture has occurred, but nevertheless the tectonic effects remain an important controlling element.

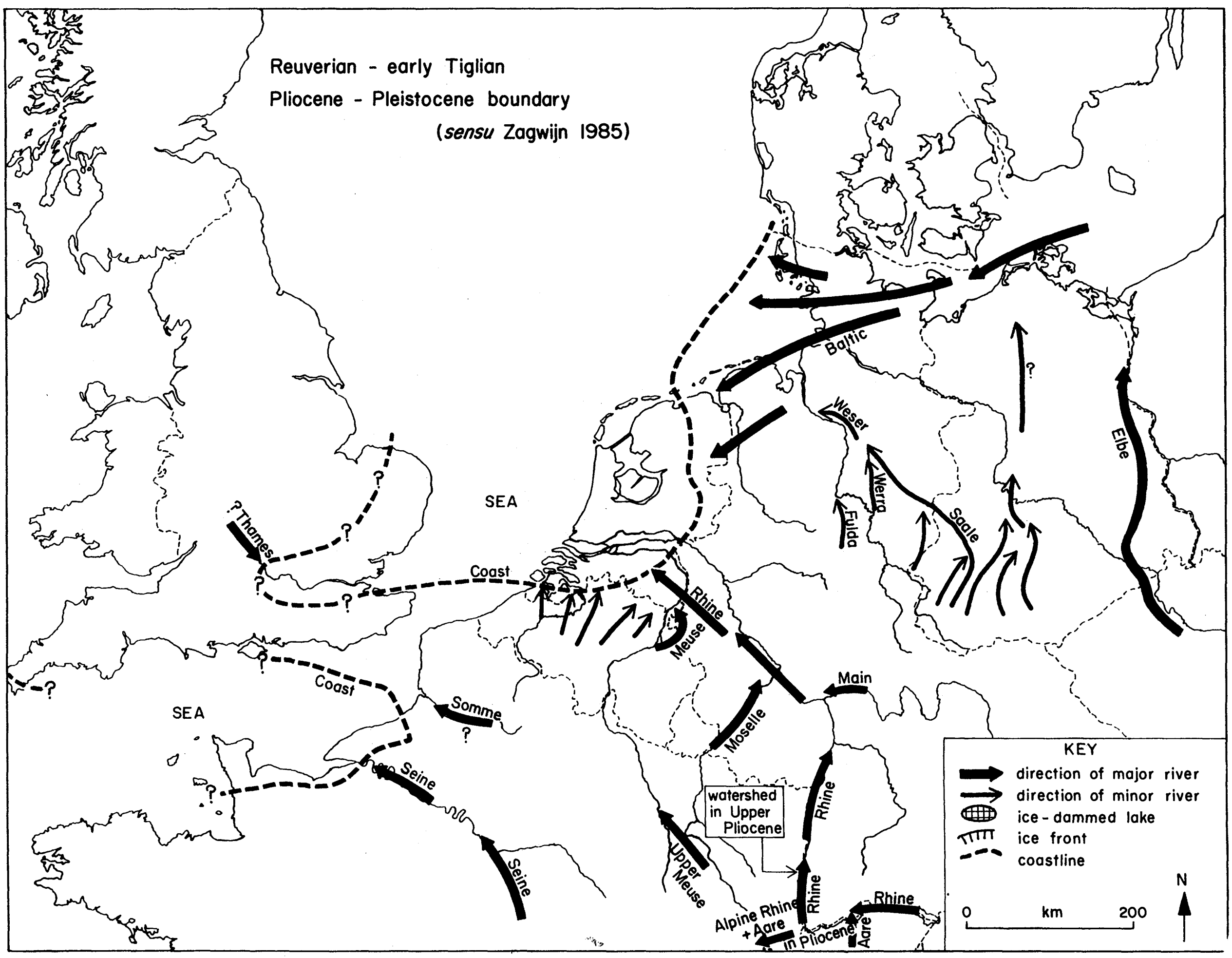

FIGURE 2. Schematic palaeogeographical reconstruction of the major drainage lines during the Reuverian (late Pliocene) to early Tiglian (early Pleistocene) Stages (*sensu* Zagwijn (1985)) based on sources quoted in the text.

THE HISTORY OF SELECTED NORTHWEST EUROPEAN RIVERS

This section comprises a series of reviews of the major river systems selected to illustrate evolution of the northwest European drainage system over the past 3–4 million years. The rivers are described in order, beginning in the north. For these reviews palaeogeographical summaries are provided in figures 2–6 for those periods for which it was practical to do so. In these reconstructions, periods of low eustatic sea level have been selected, to show the rivers at their maximum extent. During high sea-level events (i.e. interglacials), problems of location of the regional coastlines are acute, particularly for the older stages. Reconstructions for these periods have therefore been omitted.

(*a*) *Baltic River system*

Fluvial deposition became predominant in the Northwest European Basin in the Miocene (Quitzow 1953; Zagwijn & Doppert 1978) after infill of the North Sea Basin by marine sediments in the early Palaeogene. According to Bijlsma (1981) rivers from the Fennoscandian

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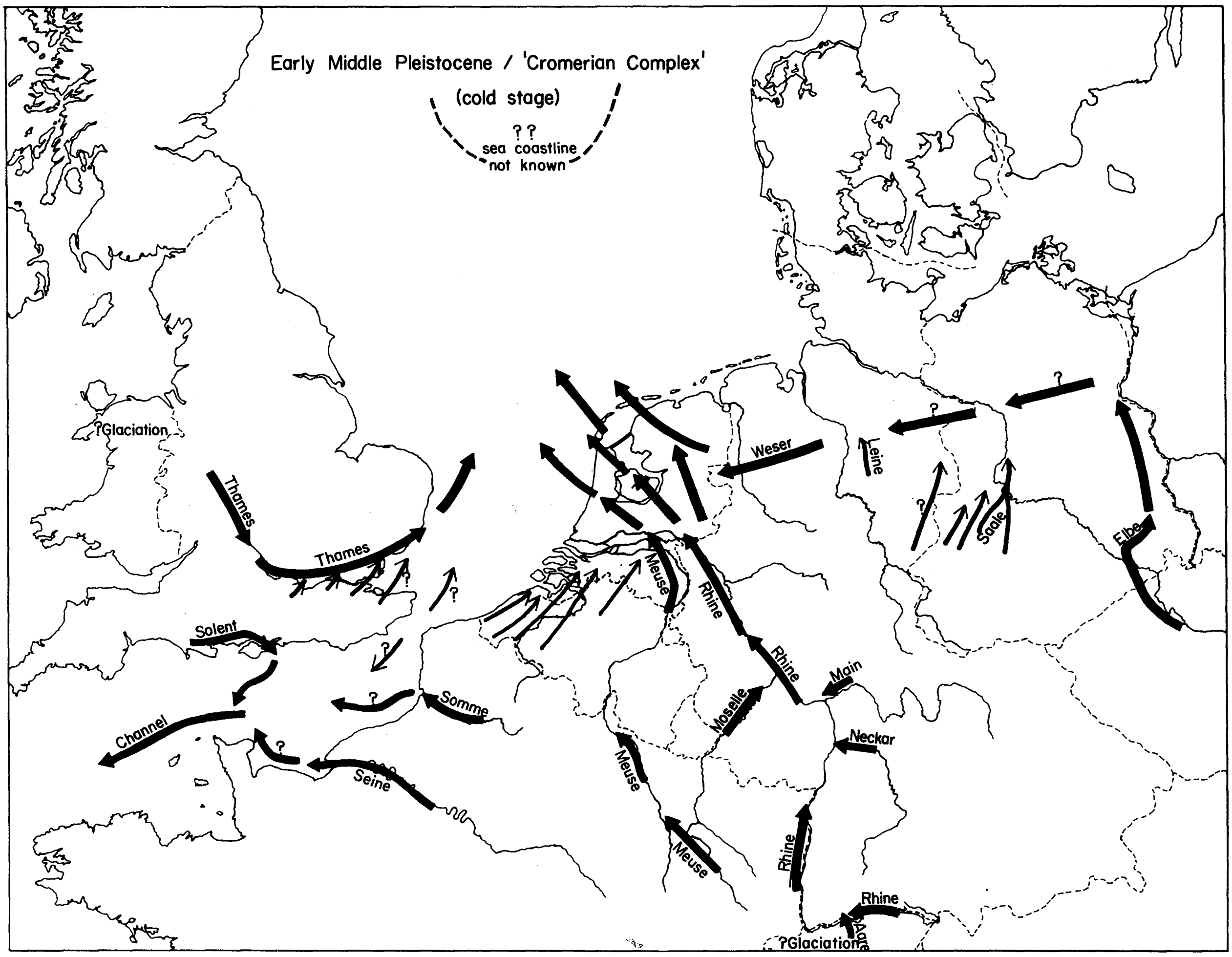

FIGURE 3. Schematic palaeogeographical reconstruction of the major drainage lines during the early Middle Pleistocene 'Cromerian Complex' at low sea-level stand. The courses are based on sources quoted in the text. (For key to symbols, see figure 2.)

Shield and the Baltic Platform in the north and the Variscan Massifs to the south (see Elbe and Weser systems, below) brought detritus into the Northwest European and East German–Polish Basins (cf. Ziegler 1978) from early in the Miocene. This sedimentary input was encouraged by uplift of areas marginal to the North Sea and East German–Polish Basins.

The deposits comprise white to grey–white quartz-rich sands intercalated with clay, gravel and brown coals (Crommelin 1954; Friis 1974). The gravel is predominantly formed of rounded quartz with subsidiary quartzites and silicified sediments, the latter mainly limestones derived from the eastern Fennoscandian Shield and southern Baltic Platform (Zandstra 1971). The river was therefore aligned through the area now occupied by the Baltic Sea and is therefore termed the Baltic River (Bijlsma 1981).

The first sediments of the Baltic River system occur in the Early to Middle Miocene of the G.D.R. and Poland, and are associated with brown coal deposition. According to Quitzow (1953) they can be subdivided into three sedimentary cycles, which may reflect periodic uplift of the source region. In the southwestern part of the North Sea Basin, fluvial and intercalated

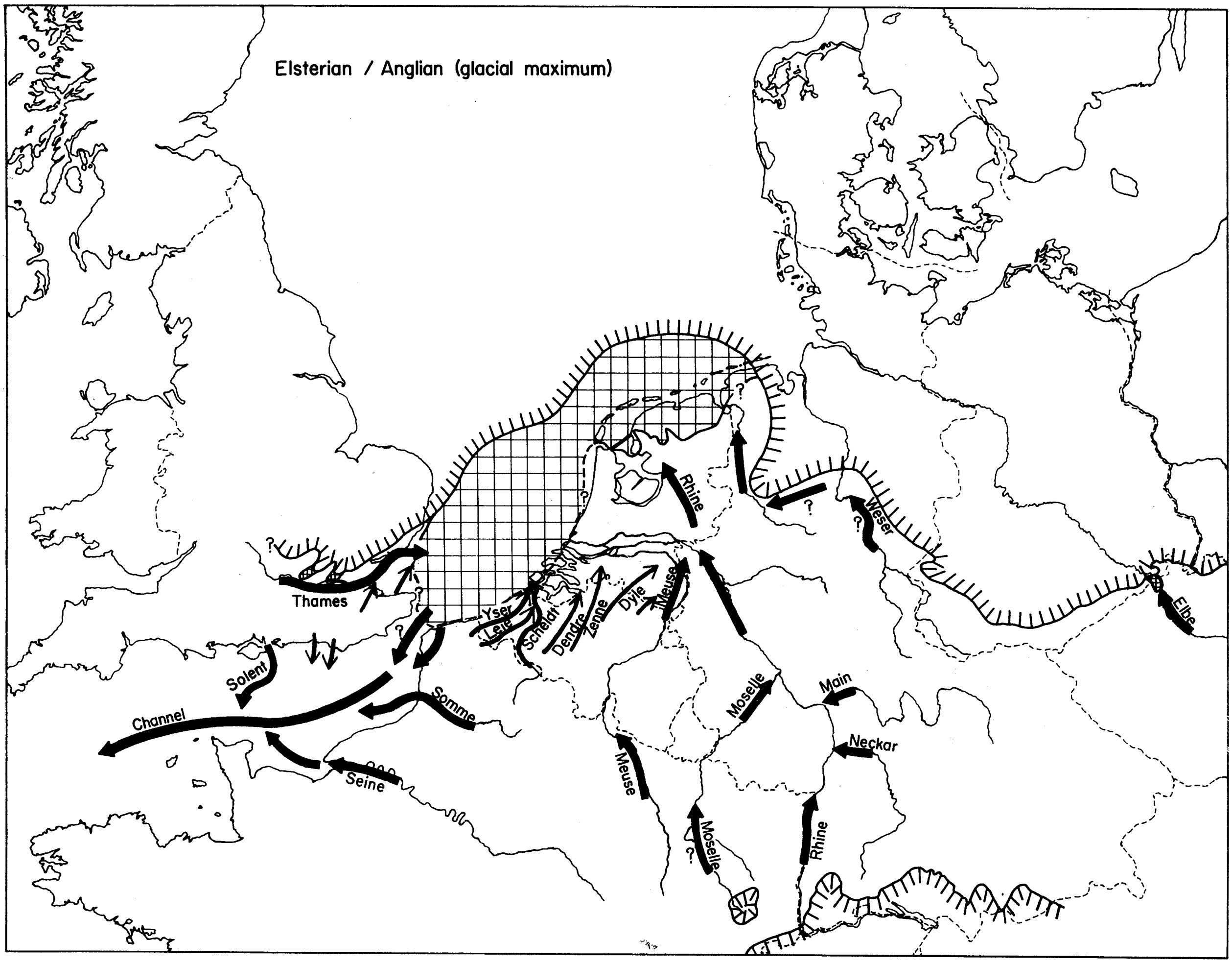

FIGURE 4. Schematic palaeogeographical reconstruction of the Elsterian–Anglian Stage at the glacial maximum. The ice-front positions and courses are based on sources quoted in the text, with additions from Kaiser (1960), Cepek (in Erd 1970) and Woldstedt (1967). (For key to symbols see figure 2.)

coastal sediments were laid down during the Miocene. In the F.R.G. and Denmark they are termed the Brown Coal Sands (Rasmussen 1961; Hinsch & Ortlam 1974).

Pliocene deposits of the Baltic River system are known from a narrow zone in the F.R.G. and The Netherlands. In the F.R.G. these Kaolin Sands have been described by Hinsch (1974). Their distribution is much affected by halokinetic movements: the greatest thicknesses are found between salt diapirs. The Kaolin Sands are exposed on the Isle of Sylt in Schleswig-Holstein, where the coarse to fine sands are of coastal and fluvial origin. They have been referred to the Brunssumian Stage by Weyl *et al.* (1955) and Averdieck (1971). Recent reinvestigation of these sands by Ehlers (1987) has shown that they contain evidence of presumed seasonal frost. Further south in Germany, brown coal layers interbedded in the sands have been found, for example at Oldenswort (Menke 1975) and Hamburg (Hallik, in Koch 1954).

The full areal extent of these deposits in northern Germany is not known. However, the course of the river seems to have been determined by subsidence in the North Sea Basin.

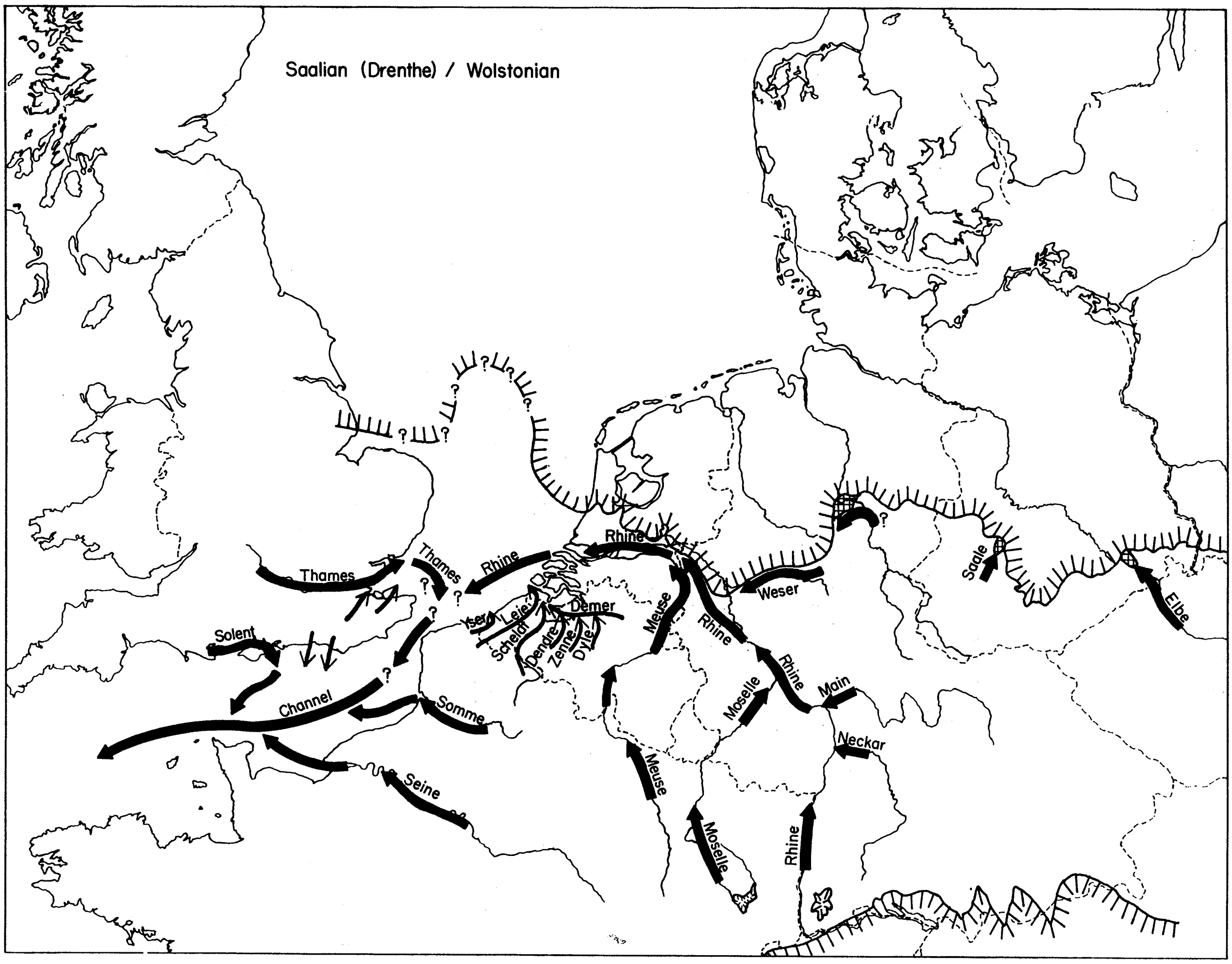

FIGURE 5. Schematic palaeogeographical reconstruction of the major drainage lines during the Saalian–Wolstonian Stage at the Drenthe substage glacial maximum. The ice-front positions and courses are based on sources quoted in the text, with additions from Kaiser (1960), Oele & Schüttenhelm (1979), Cepek (in Erd 1970) and Woldstedt (1967). (For key to symbols, see figure 2.)

In the northeastern Netherlands, deposits, predominantly fine sands but containing the characteristic pebble assemblage, are included in the Scheemda Formation of Doppert *et al.* (1975). This ranges from Brunssumian to Reuverian or possibly Early Pleistocene.

Baltic River system sediments are only known in the southern parts of the North Sea Basin in the Pleistocene, from northern Germany near Bremen to the Dutch–German border (Duphorn *et al.* 1973). They contain materials such as the characteristic lydite of the central German Uplands, as well as some large boulders that may have been ice-rafted from the eastern Baltic (Gripp 1964). In The Netherlands these coarse sands and fine gravels of the Harderwijk Formation (Doppert *et al.* 1975) range from early Tiglian to Waalian age. The earliest part of these braided river sediments is restricted to the northeast and eastern part of the Netherlands; however, by late Tiglian times and subsequently they had expanded both southwest and westwards to cover a large part of the country (Zagwijn & Doppert 1978). They also extend offshore to form a vast clastic sediment wedge (Zagwijn 1985; Balson & Cameron 1985). There are no younger sediments of this Baltic River. However, in the Enschede

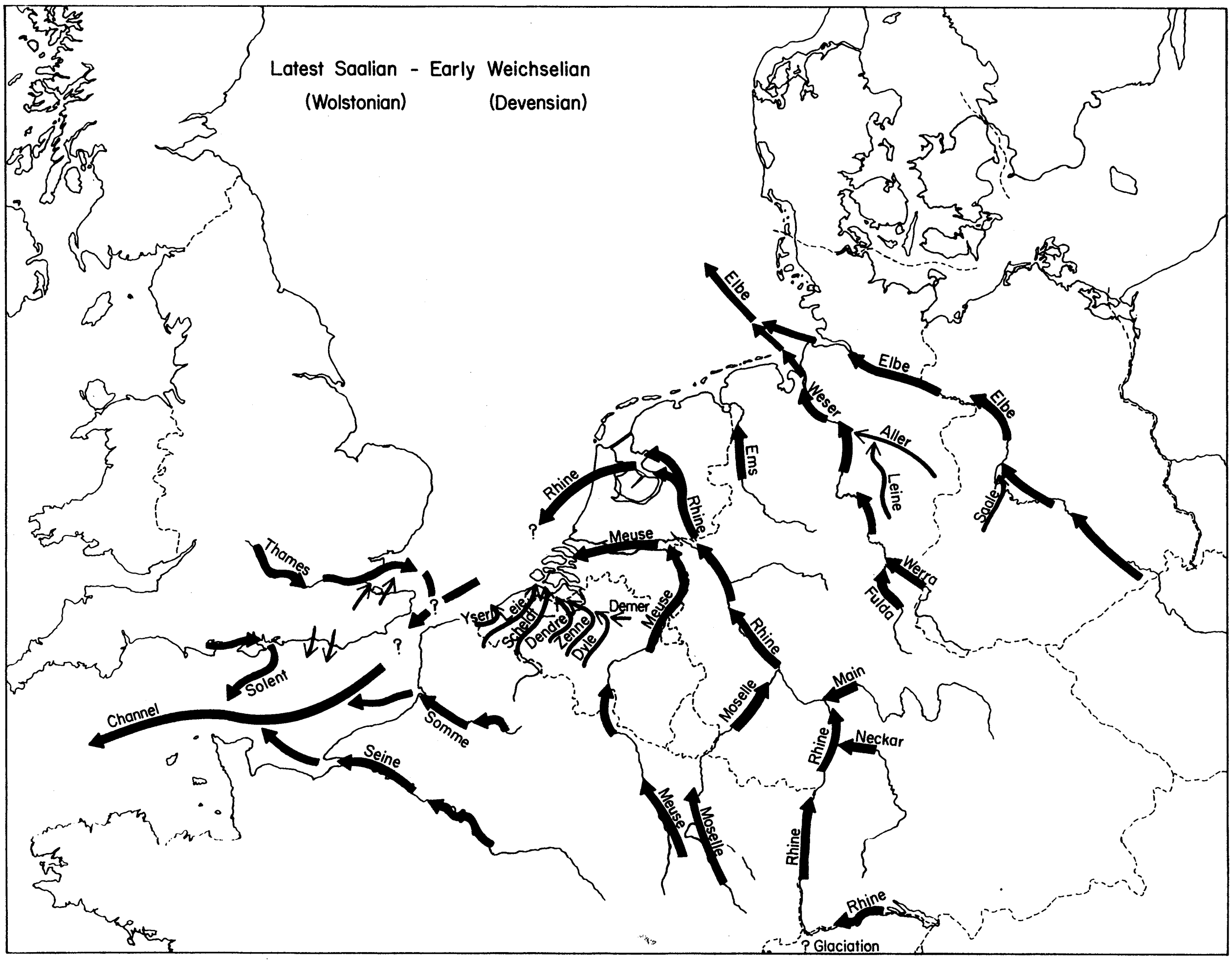

FIGURE 6. Schematic palaeogeographical reconstruction of the major drainage lines during the interval late Saalian–Wolstonian to early Weichselian–Devensian Stages at low sea-level stand. The courses are based on sources quoted in the text. (For key to symbols, see figure 2.)

Formation, which is largely of Menapian age (Doppert *et al.* 1975), crystalline rocks of Fennoscandian origin occur as large boulders or cobbles in the so-called 'Hattem Layers' of Lüttig & Maarleveld (1961) and Zandstra (1971, 1983). However, here they are intermixed with large volumes of gravel derived from central Germany via the Weser, etc. (i.e. north German rivers) and, near their southern margin, with Rhine and Meuse sediments. These 'Hattem Layers' indicate a change in depositional style and have therefore been interpreted by Bijlsma (1981) to result from contemporary glaciation of much of Fennoscandia in the Menapian. The meltwater streams from the ice sheet are thought to have discharged through the Baltic River and been confluent with the periglacial streams from central Germany. There is, however, no direct contemporary evidence of the Baltic River itself, nor is there evidence for its existence subsequently. Its destruction is attributed to glacial erosion, which resulted in formation of a proto-Baltic Basin.

After disappearance of the Baltic River, the north German rivers continued to flow into the eastern Netherlands until the Middle Pleistocene (cf. Maarleveld 1954) (see Weser (§*c*), figure 3).

(*b*) *Elbe and Saale Systems*

(*i*) *Elbe system*

The present River Elbe and its tributaries drain large areas of Czechoslovakia and the German Democratic Republic, with only a very short stretch of the river in the F.R.G. Its history is marked by repeated glacial diversion throughout the Middle and Upper Pleistocene. Indeed, its upper and lower courses are of strikingly contrasting antiquity, as has been demonstrated by Lüttig & Meyer (1974) and Eissmann (1975).

After regression of the sea, development of the northeast German–Polish Basin, and possible uplift in southern central Europe (cf. Ziegler 1978), northward-aligned drainage tended to develop. By the late Pliocene the Elbe had already established a valley through the northern Erzgebirge and northern Bohemia. This so-called Seftenberg Elbe was aligned from Dresden to Okrilla–Cunnersdorf–Hoyerswerda (Geneiser 1955, 1957). It is presumed that from here it flowed northwards to the Oder River, close to the Baltic, where it was confluent with the Baltic River (see §(*a*) above).

The actual first appearance of the Elbe has been disputed; for example Cepek (1967) favours the Miocene. Nevertheless, the course was certainly well established by the earliest Pleistocene; the gravels contain evidence of cold climates (Lüttig & Meyer 1974).

By the beginning of the Middle Pleistocene a second or Bautzen course (Geneiser 1955, 1957, 1962) was established. This course ran from Dresden via Klotsche–Okrilla–Cunnersdorf and thence east to Bautzen and the Oder. It has been suggested by Šibrava (1972, 1986), among others, that this course continued to exist until the early Elsterian, because two terraces (E terraces) in Bohemia are associated with Elsterian glacial sediments. However, Eissmann (1975) suggests that the river was aligned through Meissen to Riesa by this time. The Nestemice terrace of Bohemia was formed before the Elster I advance; the later Bohatice terrace was formed during both Elster I and II advances. Evidence for subdivision of the Elsterian glaciation has been found consistently in the southern G.D.R. (Präger 1970). A massive ice-dammed lake was formed in the Elbe and other valleys in front of the ice margin south of Dresden (Eissmann 1975). In this area the Elbe took up a course through Schmiedeberg, late in the Elsterian. However, this course must pre-date the glacial retreat, because Elsterian till rests on gravels of this phase. After the retreat of the ice sheet the river apparently established a course towards Berlin, where typical southern gravel has been found, in the Rathenow–Zossen–Wietstock area, filling valleys excavated into Elsterian glacial sediments. Mixing of reworked glacial and Elbe sediments has given rise to a characteristic pebble assemblage. Intercalated in these gravels are the so-called '*Paludina*' beds of Holsteinian age (Genieser 1962). According to the same author, the upper Elbe course through Meissen was not established before the Holsteinian, but this conflicts with the views of Eissmann (1975).

Advance of continental ice into eastern Germany during the Drenthe Stadial of the Saalian Stage caused the northerly course to be overridden as far south as Meissen, where extensive ice-dammed lakes were formed in the Dresden area (Eissmann 1975). The glacial lakes are recorded by sands and glaciolacustrine clays overlying fluvial gravel and sand of the so-called 'Main Terrace'. This 'Main Terrace' suffered later dissection so that upstream in Czechoslovakia (Šibrava 1972) it forms an important morphostratigraphical marker.

Renewed glaciation of the region during the Warthe Stadial resulted in profound changes in the upper Elbe drainage. The river was forced to flow towards the northwest by ice reaching

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the Fläming line. This diversion caused the Elbe to become linked for the first time with the Saale and Mulde rivers, which lost their lower courses (Woldstedt 1956). The continuation of the Elbe to the north took place late in the Warthe Stadial after ice recession, when the river cut through the Warthe Moraines at Magdeburg. It flowed through a series of glacial channels and topographic lows and established the present Elbe course through Hamburg into the North Sea Basin (Ehlers 1978; Meyer 1983).

By the earliest Weichselian the Bohemian–Saxonian drainage was operating as an integrated system. In the Late Weichselian, when ice advanced into eastern Schleswig-Holstein, meltwater from the ice sheet greatly enlarged the Elbe Valley in the Hamburg region (Meyer 1983). Moreover, its northwest extension can be recognized to over 150 km south of Helgoland on the North Sea floor (Figge 1980, 1983), and demonstrates that the river was graded to low eustatic sea level. The Elbe was also confluent with the Weser in the offshore area.

During the postglacial (Holocene) period this valley has been flooded as a result of high interglacial sea level in the estuary and fine sediments have accumulated on the late Weichselian fluvial gravels and sands.

(*ii*) *Saale River*

During the Pliocene, the German Central Uplands, Thüringia and the Weisse Elster basin were drained by northward-flowing streams. This probably resulted from uplift of the region (cf. Ziegler 1978). Before the Elsterian, rivers in the western Thüringia Wald transported crystalline rock pebbles to the Werra and Weser Rivers. Whereas rivers in the eastern Harz followed courses close to those at present, those from central Germany flowed northwards, to judge from discoveries of the typical heavy-mineral association in the Braunschweig area (Lüttig (1974, in Lüttig & Meyer (1974)). The latter, which includes topaz from Vogtland, is found in The Netherlands (Crommelin 1953, 1954; Maarleveld 1954; Bijlsma 1981) and is interpreted as evidence for inflow by rivers from the north German area before the Elsterian.

With the advance of Elsterian ice a series of ice-dammed lakes was formed in the river valleys in the Halle–Leipzig area. These may have coalesced into a single massive lake 450 km^2 in area (Eissmann 1975). The advance during the Elsterian brought ice to the Thüringia Basin and the Eastern Harz. In Elster I, glaciolacustrine clays were laid down on river gravels and till was deposited above. Subsequent ice retreat allowed laminated clays to accumulate once more, but these were overridden by the Elster II advance. The substantial moraines remaining after retreat of the ice caused many of the North German rivers to adopt new courses, some following glacial valleys; for example, the Saale, Mulde and Elster rivers formed a single river north of Zörbig (Knoth 1964; Ruske 1964).

In the Holsteinian, fine sediments were laid down in river valleys, for example, at Edderitz in the Riesdorf Saale Valley. A period of downcutting seems to have followed the Holsteinian and an Early Saalian interstadial (Dömnitz). Subsequent aggradation gave rise to a 'Main Terrace' found in each of the river valleys. For instance, this early Drenthe aggradation can be followed along the Saale river as far as the confluence with the present Elbe Valley. The lobate advance of ice in the Drenthe caused ice-dammed lakes to be formed in the river valleys. On retreat of this ice the rivers were again forced to adopt new courses; for example, the Saale changed its course to the west downstream of Merseburg (Schulz 1962; Lüttig & Meyer 1974).

In contrast to the Elbe, the Saale River was not overwhelmed by ice in the subsequent

Warthe Stadial. Instead, the river deposited gravels and sands downstream as far as Magdeberg. In the Gerwisch area an Acheulian industry is found in these sediments. The Saale joined the Elbe, as mentioned above, after diversion of the latter by the Fläming ice margin of the Warthe Stadial. Subsequent development took place under interglacial conditions in the Eemian and Holocene (Flandrian) with the accumulation of 'lower terrace' gravels during the Weichselian.

(c) *Weser system*

The River Weser is formed by confluence of two important rivers, the Fulda and Werra, and has been investigated by Lüttig (1974). Both these rivers drain the uplands underlain by Palaeozoic to Mesozoic rocks of the Rhenisches Schiefergebirge and the Thüringia Wald. However, below Porta Westfalica the Weser flows across the unconsolidated Quaternary rocks of the North German Plain. Near Verden, the broadly northwest to westerly flow direction of the Weser becomes more northwesterly where it is joined by its major tributary, the Aller.

In the Weser catchment there were two major tectonic influences during the Tertiary. Firstly, uplift of the Palaeozoic–Mesozoic rocks was associated with Saxonian tectonics. These late Cretaceous–early Tertiary movements reflect compression from the south during late Alpine orogenesis. This uplift is accompanied by contemporary downwarping of the Northwest European Basin (Ziegler 1978).

The earliest evidence of northward drainage from the Thüringia Wald is found in the Miocene lignite sands of the Kassel area, in which abundant flint and crystalline rocks occur. In the Neogene, uplift of the Lower Saxony uplands became more intense associated with faulting and basaltic volcanism. A series of gravels and sands derived from these upland areas are found but are restricted to local tectonic basins and depressions between salt domes. For example, post-basaltic flint and crystalline rock gravels and sands occur in association with clays of Reuverian (Late Pliocene) age in the Uslar Basin.

In the Pliocene, uplift continued south of a line from the North Harz to the Wiehengebirge of Lower Saxony. During this period the sea retreated to beyond the present coast. At the same time the rivers seem to have flowed towards the north and northwest, to judge from the occurrence of 'southern-type' quartz-bearing gravel spreads from areas between the present Weser and the Ems. Although these spreads are recognized, the source of the material and the actual river courses are not identified. Indeed Lüttig (1974) has postulated that the distribution suggests that the rivers may have been migrating across much of northwest Germany at this time.

Late in the Pliocene the rivers may have established more fixed courses, to judge from evidence in the upper Werra, upper Weser and Fulda valleys. Here gravels are found as elevated remnants at scattered localities. Pebbles from the present source area of the Weser are also found in a broad strip that extends generally westwards from the Minden–Nienburg area to the Netherlands and the Ems river. Pebbles of granite and porphyry from the Thüringia Wald are present. Similar material is also found at Braunschweig, deposited by the early Saale river which also flowed north of the Harz and followed the present Leine course. This westward-flowing stream was part of the Baltic River system (Maarleveld 1956; Bijlsma 1981).

Because of the structural relations in this region, i.e. uplift of the uplands and subsidence of the lowlands, older deposits progressively dip downvalley and pass below the modern floodplain level one after another in a transition zone. North of this zone the fluvial deposits

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occur in superposition with the oldest at depth. The transition zone at the upland front is not simple but is separated into structural strips marking the boundary between the two areas.

The so-called 'Upper Terrace' is much better known than the higher earlier deposits. Sediments of this complex unit are widespread in the upper Weser, Leine and other tributaries. The presence of well preserved 'Upper Terrace' remnants in the Leine–Aller system has been established by Lüttig (1960) on the basis of Thüringia Wald pebbles. At this time the Leine flowed westwards from Elze towards the Netherlands where the gravels are assigned to the Enschede Formation (Zagwijn 1985).

With the onset of continental glaciation in the Elsterian of north Germany, stratigraphical control is much improved. The lower courses of the rivers are still poorly known from this period; however, the upper courses have provided important evidence. The gravels and sands relating to this period are still included in the 'Upper Terrace' complex. In the Harz rivers, gravels and sands derived from the devegetated ground surfaces were laid down at the beginning of the Elsterian when the ice sheet entered the region and overrode river valleys. It brought about a complex series of drainage changes. A short recession was followed by a major readvance that took the ice across large areas of Lower Saxony and may have dammed the Weser at Hamelin, although this has recently been disputed by J. P. Groetzner *et al.* (unpublished). Water from the upper river courses, as well as the meltwater, must have been directed south of the Wiehengebirge into the Osnabrück area from where it may have continued into the Ems valley at this time. The contemporary ice margin extended from north of Minden west into the Emden area at Salzbergen (Meyer *et al.* 1977), and eastwards into Saxony and Silesia. East of the Weser Valley this margin marks the maximum extent of Pleistocene glaciation (figure 4 and 5). However, to the west the Saalian glaciation was the most extensive.

After glacial recession, proglacial lacustrine clays of the Lauenburg Clay Formation were laid down in basinal areas in the lower Weser area. This was accompanied by incision of the upper sections of the river valleys. With the climatic amelioration of the Holsteinian Stage, widespread interglacial sedimentation took place. These predominantly fine sediments have yielded much fossil evidence (see Lüttig (1974) for references). Reworked faunal material derived from interglacial sediment is found in the next unit of gravels and sand of the 'Middle Terrace' complex. This unit, which contains evidence for cold-climate braided river deposition, also includes Palaeolithic artefacts. The gravels and sands comprise local products of periglacial weathering. These early Drenthe Stadial deposits were overridden by continental Drenthe ice in the lower courses but remain undisturbed in the upstream areas. During the Rehburger phase the ice penetrated as far south as the central Netherlands, and to Braunschweig. The deposits of the 'Middle terrace' can be shown to be contemporary with this ice advance and ice-dammed lake sediments were laid down at Porta Westfalica (J. P. Groetzner *et al.*, unpublished). After an ice recession, downcutting occurred in the valleys; it was into the newly deepened valleys that ice flowed and deposited till. Drainage during this phase was aligned marginally to the ice and entered the Rhine in the Ruhr area (Thome 1980). This is supported by finds of exceptionally 'eastern' material in The Netherlands (Zandstra 1983). This second, Heisterberg-phase advance caused the ice to bulldoze preexisting fluvial and glacial deposits into a series of extensive marginal ridges.

The rivers, after having flowed westwards since the early Pleistocene, breached ice-pushed ridges and flowed northwards. The ice recession allowed the Weser to follow a series of

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glacially overdeepened areas to reach its present northwest course in the late Drenthe Stadial; this course may already have been exploited, before the glaciation, by the Aller River.

The subsequent interstadial and Warthe Stadial periods are not represented in the middle and upper Weser System except possibly by sands in the northern tributaries of the Aller.

By Eemian Stage times, however, the rivers had almost cut down to present levels and aggradation of fossiliferous sediments of argillaceous and/or calcareous lithologies took place (see Lüttig (1974) for references). The glacioeustatic rise in sea level caused the lower reaches of the river to be drowned by the sea. The river course was then almost as today.

Glaciation during the Late Weichselian did not reach the Weser system. Instead the area was subjected to intense periglacial conditions, which provided large volumes of coarse material by slope processes. The gravels and sands derived from these materials, the 'Lower Terrace' complex, reach substantial thicknesses and were deposited by the river in a braided mode. In some places the 'Lower Terrace' can be subdivided into lower and upper subunits; radiocarbon dates on material from these units indicate that sedimentation continued into the Late Glacial (Lüttig 1974).

In the Holocene fine sediments were once more laid down, dominated by the so-called 'flood loams' of overbank origin. Lowered humidity in the Atlantic period gave rise locally to sand-dune formation on the north and east banks of the Aller, the Leine below Hanover and the Weser below Minden. Today, however, these dunes are generally stable.

(*d*) *Rhine system*

The Rhine System is the longest and therefore the most important member of the present northwest European drainage system. At present it rises in the central Swiss Alps, where it is fed by glacial meltwater, and flows into the northern Alpine Foreland. In northern Switzerland it enters the Bodensee (Lake Constance) which marks the lower limit of the so-called Alpine Rhine. In this section the river, aligned northeast–southwest, has a very steep gradient and falls over 2000 m. Downstream of the lake, as far as Basle, the river passes through the Upper Rhine Valley. The gradient in this stretch, although much shallower than previously, is still rather steep and is broken at intervals by rapids and a large waterfall. Midway between Basle and Schaffhausen the first major tributary, the Aare, which drains much of the Alpine Bernese Oberland region, joins the Rhine. At Basel, the river turns northwards and enters the Upper Rhine Graben in which it flows to Oppenheim. No change of gradient occurs, however, even though this trough is one of active subsidence in which the river is confined for about 250 km. The Upper Rhine Graben is a major structural element that originated in the Eocene as a result of the late Alpine orogenic phases (Ziegler 1978). At Mannheim the second major tributary, the Neckar, joins the Rhine. The Neckar system drains much of the Black Forest.

Downstream of Oppenheim the Rhine passes into the Middle Rhine Valley by entering the Mainz Basin. The latter extends as far as Koblenz, an area of net uplift. At Mainz, the Main is confluent with the river; at Koblenz, it is joined by the Moselle. The former drains much of central Germany, whereas the latter brings water from the Vosges and Hunsrück regions. Throughout the Middle Rhine the river passes through the tectonically active uplifting Rhenish Massif (Rheinisches Schiefergebirge). The resistant lithologies in this region cause the river once more to steepen its gradient and occasionally to give rise to rapids (subsequently removed by man). This course has been occupied by the Rhine at least since the Pliocene (Bibus 1980). At Koblenz the river passes through the Neuwied Basin, which extends approximately to Andernach. Downstream the river crosses the northern boundary faults of the Rhenish Massif and enters the Lower Rhine Embayment, from where it has a gentle gradient into the Netherlands and the sea. However, in The Netherlands and southern North Sea Basin, subsidence has affected the Rhine downstream of Nijmegen (Quitzow 1974; Brunnacker 1975; Zagwijn & Doppert 1978). In The Netherlands the Rhine and Meuse (Maas) share a common delta and their history has been linked throughout the period considered here.

The tectonic complexity of the region through which it flows has had great impact on the type and preservation of deposits of the Rhine system. In essence this means that, in areas of net uplift like the Rhenish Massif, the vertical distance between terrace accumulations is increased relative to stable regions. By constrast, in areas of crustal depression, such as downstream of the Nijmegen–Cologne hinge zone and in the Upper Rhine Graben, thick sequences of younger sediments overlying older sediments are laid down (Brunnacker 1975). By its nature the Rhine has a long and detailed history, which is based here on Quitzow (1974), Boenigk (1978), Brunnacker (1975, 1978, 1986), Semmel (1973), De Jong (1965), Zagwijn (1974, 1979, 1985) and Zonneveld (1974).

The River Rhine seems to have formed in the middle Miocene, so that by the Pliocene it had already evolved considerably. According to Boenigk (1978) the early Pliocene Rhine deposited large quantities of coarse clastics in the Lower Rhine. The materials comprise highly stable minerals and rounded pebbles, mainly of quartz and quartzite derived from the Moselle area. These 'Hauptschotter' or Waubach Gravel sediments are included in the Kieseloolite Formation. At this time, the Rhine was only a local stream in the Rhenish Massif area.

In the Upper Rhine Graben, the southern Vosges and southern Germany, water flowed southwards to join a pre-Alp river aligned towards the southwest. This river almost certainly entered the Rhône system. Uplift of the Swiss Jura during the Middle Pliocene forced the Aare to flow to the northeast to join the Rhône tributary: the Doubs and the pre-Alp river southwest of Basle (Quitzow 1974). The Danube (Donau) was also created at this time. By the Middle–Late Pliocene the Rhine had cut back to a major watershed in the Upper Rhine Graben at Kaiserstuhl near Freiburg (figure 2). Depression of the Graben caused considerable expansion of the rivers Neckar and Main by capture of Danube tributaries in Swabia and north Bavaria (Quitzow 1974).

In the Lower Rhine the deposition of clastics decreased into the middle Pliocene; Waubach Gravel was replaced by widespread lagoonal clays and organic sediments (Red or Brunssum Clay) deposited under a warm-temperate climate. The contemporary river system seems to have adopted a meandering pattern with fine clastics restricted to narrow channels. Zagwijn (1963) has interpreted the couplet of Waubach Gravel and Brunssum Clay as a major sedimentary cycle. This he compared with stratigraphically higher levels deposited by the Rhine (such as the Late Pliocene Schinveld Sand and Reuver Clay; Early Pleistocene Tegelen Gravel and Tegelen Clay) and interpreted them as related to intermittent tectonic uplift of the central European uplands. Further coarse clastics of the Schinveld Sand and Gravels overlie the fine sediments and contain higher frequencies of less durable lithologies than the previous unit. These rocks, derived from the Rhenish Massif, mark the breakthrough of the Rhine into the Upper Rhine Graben. The gravels show evidence of deposition in an oscillating-discharge 'braided-type' system under a cool dry climate (Boenigk 1978). Deposition by the Rhine in The Netherlands was largely restricted to a delta in the Central Graben area (Zagwijn 1974).

In the Upper Reuverian the coarse clastics are replaced by widespread deposition of clay,

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sand and peat under a broadly humid, warm-temperate climate (Zagwijn 1960). According to Boenigk (1978), contemporary with this period there is a major change in the sediments in the basin, which he attributed to the expansion of the Rhine into the Alsace molasse region. This resulted in a great increase in the rivers' catchment.

In the Upper Pliocene–Lower Pleistocene, the southwest-flowing Aare filled a wide valley that developed between Basle and the present Doubs Valley with very coarse gravels termed the Sundgau Gravels. Quitzow (1974) equated these gravels with the earliest known deposits of the Upper Rhine Valley, the Older Deckenschotter, which he thinks were laid down by meltwater discharging into the Aare–Rhône system. By the time that the Younger Deckenschotter were deposited in the Upper Rhine, the Aare had broken through into the Upper Rhine Graben and established a connection with the Rhine. Exactly when this happened is not agreed; Quitzow (1974) equates it with the 'Hauptterrasse 3' (HT III; cf. Brunnacker (1975)) of the Lower Rhine on the basis of alpine rocks in Rhine gravels, whereas Zonneveld (1974) states it was not earlier than the Plio-Pleistocene boundary.

The Pliocene–Pleistocene boundary is represented in the Lower Rhine and The Netherlands by a widespread hiatus (Zagwijn & Doppert 1978), a possible consequence of local uplift or erosion. The southeast Netherlands and the Lower Rhine Embayment are marked by the development of a series of northwest–southeast aligned tectonic blocks at this time, differential movement of which gave rise to a series of basins and highs (cf. Zagwijn & Doppert 1978). The greatest thickness of Pleistocene sediments is found in the Central Graben near Eindhoven and its upstream extension, the Erft Basin; sediments also accumulated in the Venlo Graben and covered the intervening Peel Horst–Ville Block. In the earliest Pleistocene The Netherlands was beneath the sea, the Rhine discharging in the Central Graben region (Zagwijn 1974, 1979). Upstream in the Lower Rhine Embayment the Rhine deposited a series of gravels and sands which are interbedded with clays and organic deposits of limnofluvial origin. This complex of sediments forms the 'Hauptterasse 1' (HT I) unit of Brunnacker (1975, 1978, 1986). The basal Pleistocene gravels at Ville (Gravel b, ?Praetiglian) are of cold-climate origin, as are the later gravel units. This is overlain by temperate fossiliferous clays and intercalated sands referred to the Tiglian interglacial stage (Zagwijn 1960, 1985; Brunnacker 1975, 1978, 1986; Boenigk 1978). The next gravel deposits (gravel D), representing the Eburonian Stage, are the first to be deposited in thick units rather than sheets and for the first time include large ice-rafted blocks. However, according to Semmel (1973), drift blocks were already present in Praetiglian gravels in the Rhine–Main area. The river adopted a braided pattern during this and all subsequent gravel and sand depositional periods (Brunnacker 1975, 1978, 1986; Boenigk 1978). The remaining gravel and clay interbedded units of HT I are referred to the Waalian and Menapian Stages. The upper part of the HT I gravels and sands are probably the equivalent of the combined Rhine–Meuse Kedichem Formation in The Netherlands. The latter is overlain by the important Sterksel Formation, a unit of crossbedded coarse gravel-rich sands, about 50 m in thickness, which contain materials of Alpine provenance (De Jong 1965). It spans several interglacial and glacial periods (Zonneveld 1974). The uppermost part of the formation, the Weert Zone, which is equivalent to 'Hauptterrasse 2' (HT II) in the Lower Rhine (*sensu* Brunnacker 1975, 1978, 1986) dates from Glacial B of the 'Cromerian Complex' (Zagwijn 1985). The Weert Zone, a garnet-poor assemblage, is an important market horizon in The Netherlands. During the intervening period the Rhine–Meuse system, together with the North German rivers, had greatly expanded their deltas to the west and northwest into the

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southern North Sea Basin (Zagwijn 1974); these sediments are almost certainly the Yarmouth Roads Formation of the offshore area (Balson & Cameron 1985; Bowen *et al.* 1986).

In the Upper Rhine Graben, subsidence has resulted in the preservation of very thick Pleistocene successions; the thickest sequence (over 328 m) occurs near Heidelberg. The depression has also caused the tributaries to incise very deeply into the the valleys adjacent to the Graben (Quitzow 1974). In the Mainz Basin parts of the uplifted Kelsterbach Terrace may be equivalent to the Lower Pleistocene (von der Brelie, in Semmel (1973)).

During the deposition of the 'Hauptterrasse 3' (HT III) of the Lower Rhine, a general eastward shift of the river was in progress. This prevented deposition of the gravels and sands across the whole basin (Boenigk 1978). The gravels are rich in Eifel volcanic material (Brunnacker 1975). An important climatic development is also recorded in the sediments; the local appearance of widespread permafrost with ice-wedge casts and cryoturbation occurs for the first time in HT III (Brunnacker 1975; Boenigk 1978). The deposits of HT III are equated in The Netherlands with the Veghel Formation and are therefore of late 'Cromerian Complex' age. The continued eastward shift of the Rhine begun in HT III was completed by HT IV, when the river finally left the western basin because of uplift and infill by the Meuse.

Contemporary subsidiary deposition upstream continued in the Neuwied Basin, where equivalents of the Main Terraces can also be recognized. The early Middle Pleistocene sequence can be equated with the Upper Rhine Graben. The gravels and sands preserved in the Graben are intercalated with widespread fossiliferous sands of the Mosbachian, the equivalence of which is debated, but is probably of early Middle Pleistocene age. This deposit is also found in the Rhine tributaries, the Main and Neckar (Quitzow 1974). It is possible that the Mosbach deposits may also be in part equivalent to the Kelsterbach Terrace of the Mainz Basin.

Once the Lower Rhine was established east of the Ville Block, a series of terraces was formed, the so-called 'Middle Terraces' (*sensu* Brunnacker 1975, 1978, 1986). In spite of a thick loess cover, four morphologically defined gravel and sand units with intercalated fine sediments can be recognized. The earliest Middle Terrace 1 (MT I) gravels include acidic volcanic minerals from the Eifel in the Middle Rhine. The influx of volcanic minerals from the Eifel marks the change in The Netherlands from the Veghel Formation to the Urk Formation. This unit, marked by a lack of permafrost structures which Brunnacker (1975) relates to a mild climate, can also be equated with tectonic activity. Uplift of the Rhenish Massif during late deposition of the 'Hauptterrasse' caused massive aggradation in the Main and Neckar Valleys upstream, according to Brunnacker (1975). This may be a time-equivalent of the Günz period of the Alpine region (Quitzow 1974). Evidence of glaciation in northern Europe is indicated at this time ('Cromerian glacial C') in the northern Netherlands by the occurrence of Scandinavian erratics in the Weerdinge Member, basal Urk Formation (Ruegg & Zandstra 1977; Zandstra 1983).

The Lower Rhine Middle Terrace II of Brunnacker (1975) shows the return of periglacial conditions but with an intercalated temperate deposit of probable 'Cromerian IV' age (*sensu* Zagwijn 1985). The subsequent terrace MT III is underlain by coarse gravels, the so-called 'Channel Gravels', which, because they are overlain by the fossiliferous Krefeld Beds of Holsteinian age, are referred to the Elsterian Stage (Brunnacker 1975, 1978, 1986).

In the Middle Rhine Valley deepening continued into the Elsterian (Quitzow 1974), and

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similarly in the Upper Rhine. However, advance of ice in the Mindel (?Elsterian) certainly overran the Alpine and Upper Rhine. However, in The Netherlands, continental ice did not impinge on the Rhine in the onshore area. Nevertheless, ponding of the southern North Sea by ice caused the deposition of the glaciolacustrine Peelo Formation clays in the northern Netherlands (Ter Wee 1983*a*; Zagwijn 1974, 1979). The river discharged into this lake (see Channel, §*j*).

After deglaciation the North Sea transgressed into the Lower Rhine Valley in The Netherlands (Zagwijn 1974). Upstream fossiliferous sediments were laid down such as at Krefeld and Karlsruhe (Semmel 1973; Brunnacker 1975).

In the Saalian (?Riss), erosion of the Upper Rhine Valley reached a greater depth than subsequently, so that these earlier channel gravels underlie Weichselian deposits. The lower Middle Terrace can be equated with the outermost Alpine Riss endmoraines (Quitzow 1974). In the Lower Rhine the gravels of MT III, overlying the Krefeld Beds, are equated with ice advance in the Drenthe Substage (Brunnacker 1975, 1986; Zagwijn 1985) although Zagwijn equates this with MT IV. This advance caused bulldozing of Rhine sediments near Krefeld and in the Central Netherlands (Zagwijn 1974). The advance also deflected the Rhine from its northwest course in The Netherlands towards the west. However, after ice retreat the river partly readopted a northerly course by flowing through the Deventer basin towards Noordoostpolder (Van der Meene & Zagwijn 1978). This northern course, now occupied by the IJssel river, was abandoned by the Rhine in favour of a westerly course during the Middle Weichselian (Van der Meene & Zagwijn 1978).

A fourth Middle Terrace, MT IV, according to Brunnacker (1975) post-dates the Drenthe Substage. The age of this gravel unit can be defined by reference to the next lower accumulation, the Lower Terraces, which in The Netherlands are equivalent to the Kreftenheye Formation (Zagwijn 1985). These braided river gravels and sands of Rhine and Meuse origin include peats and organic material of Eemian age. The underlying gravels are therefore, by implication, of Saalian age (De Jong 1965; Zonneveld 1974; Zagwijn 1985). The Kreftenheye Formation closely follows the modern river course in The Netherlands and can be traced out beneath the southern North Sea to at least 50 km offshore (Jelgersma *et al.* 1979). According to these authors, the Rhine discharged through the Straits of Dover at this time (figure 6).

Deposition of the Kreftenheye Formation resumed in the Weichselian, a twofold subdivision of which is possible, as the upper horizons contain pumice from the Laachersee eruption (*ca.* 11 ka BP) in the Late Glacial Interstadial. Upstream two Lower Terraces can be identified by using the same criterion, according to Brunnacker (1975). Indeed equivalents of the Lower Terraces are known throughout the Rhine system and can be related to Würm endmoraines in the Alps (Quitzow 1974).

Since the last glaciation flood silts have been deposited on the underlying gravels in the Rhine Valley. In the Alpine Rhine the glacially overdeepened basin of Bodensee is being infilled by massive accumulations of deltaic deposits, which extend upstream to Chur.

(*e*) *Meuse* (*Maas*) *system*

Over the past few decades the history of the Meuse has received attention from Dutch, Belgian and French workers, most notably Zonneveld (1974), Paulissen (1973), Maarleveld (1956) and Pissart (1974).

The present River Meuse system drains the eastern part of the Paris Basin, Lorraine and the Ardennes in southeastern Belgium, from where it flows northeastwards into the Lower Rhine

Graben. After crossing a series of northwest–southeast-trending major faults, the river turns first northeast then eastwards to parallel the Rhine through the southern Netherlands to the sea. In their lower reaches the Meuse and Rhine have closely linked histories; they share the same general course and the Meuse has been a Rhine tributary for part of its Neogene existence.

According to Pissart (1974) a primitive Meuse system was already in existence in the Miocene and was represented by northwest flowing streams. However, by the late Pliocene–early Pleistocene, a series of captures in the region north of Mézières had resulted in the formation of northeastward flowing drainage. The Meuse system has lost a considerable part of its basin during the Quaternary as a result of river capture: for example, the Aire was lost to the westward-flowing Oise in northern France and the Upper Meuse was lost to the Moselle and thus to the Rhine.

The drainage-system changes must have been encouraged by differential local tectonic movements such as subsidence of The Netherlands' Basin region and uplift of the Paris Basin. These differential movements have given rise to local deviations of the terrace deposits and varying presentation of remains; for example, on the Mesozoic rocks of the Lorraine basin the Meuse deposits are very poorly preserved, whereas between Mézières and Liège they are better preserved on the more resistant Palaeozoic lithologies.

Divergence of older aggradations upstream of Namur indicates that the movement axis migrated northwards progressively during the Quaternary from Givet to Anseremme; local uplift of the Liège region may account for strongly inclined downstream aggradations. In South Limburg, where the river is no longer restricted to the valley, very wide accumulations are found. According to Zonneveld (1974) the quartz content of the aggradations decreases with age; this result indicates progressive incision of the Ardennes Massif through the Quaternary.

In Plio-Pleistocene times the river had adopted a wide, shallow valley trending in a southwest–northeast direction. Several levels of braided river sand and gravel aggradations have been distinguished. A peat bed within sediments of the Sempelveld Terrace is probably of late Tiglian age (Zagwijn 1985). The oldest deposits in the Lower Meuse, often referred to as 'Traînée Mosane' and comprising only very resistant quartzite and flint, occupy a very shallow valley, represented by gravel on the Ubachsberg Massif and in the Lower Rhine Basin (Boenigk 1978). It is referred to the late Pliocene. The oldest terraces are aligned south and east of this massif.

The next lowest aggradation in the South Limburg area, the Margraten Terrace, includes the first evidence of detritus derived from the Upper Meuse Vosges catchment into the Lower Meuse. This material indicates that by this time the drainage link from the Vosges region had been established, although Bustamente (discussion in Pissart (1974)) believes that this link was already established in Plio-Pleistocene times. Also at this time the Dutch–German border area was greatly influenced by tectonic movements along the Rhine Graben boundary faults. These led to the Meuse's adopting a more northerly course by periodic northwest migration; the resulting thick gravel aggradation forced the Rhine to flow further to the east in the Lower Rhine Valley. Organic sediments resting on gravels beneath the Valkenburg–Sibbe Terrace (Boenigk 1978) have been referred to the Bavel interglacial Substage of the Bavelian by Zagwijn (1985). Equivalent terrace aggradations of these older terraces exist upstream of Liège at least as far as Mézières according to Pissart (1974), although this conclusion is almost entirely based on altitudinal correlations, which provide a well-developed 'stairway'.

The so-called 'Middle Terraces' in southern Limburg, and their downstream equivalent, the

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Veghel Formation in the buried valley of Middle and North Limburg and northern Brabant, span the Middle Pleistocene. The 'Middle Terrace' sequence was studied by Paulissen (1973) who noted that almost all the deposits accumulated under periglacial conditions with the river adopting a braided mode. The only possible exception is the Lanaken Terrace, which may be of interglacial origin. These deposits form part of the Veghel Formation of the southern Netherlands (Zonneveld 1958, 1974). The lowest part of the Veghel Formation contains clay deposits at Rosmalen equated with 'Cromerian Complex Interglacial III' (De Ridder & Zagwijn 1962). Deposits of this formation can be traced into the central Netherlands. According to Bustamente (1974) by the time of deposition of the Caberg Terrace gravels, heavy-mineral analyses indicate that supply of Vosges material had stopped entering the Meuse. This cessation reflects capture of the Upper Meuse by the Moselle. Because the Caberg Terrace is thought to be of early Saalian age (Zagwijn 1985) this capture most probably took place during the preceding Elsterian Stage (figures 4 and 5). At Maastricht–Belvédère a fossiliferous temperate deposit overlies gravels and sands of the Caberg Terrace. Detailed investigation of the sediments indicates deposition early in the Saalian Stage (Vandenberghe *et al.* 1985; van Kolfschoten & Roebroeks 1985).

From late in the Saalian the Kreftenheye Formation sediments were laid down by broad braided streams in the southern Netherlands (De Jong 1965). These Rhine–Meuse sediments include interglacial organic sediments of Eemian age. According to Van der Meene & Zagwijn (1978) diversion of the Rhine by the Drenthe ice caused confluence with the Meuse. This was severed after the ice retreat, but a westerly Rhine course was re-established in the Middle Weichselian.

Aggradation of the youngest of the Saalian terraces in Limburg, the Eisden–Lanklaar Terrace (Paulissen 1973) was disturbed by movement on the northwest–southeast-trending Feldbiss Fault, a major line from The Netherlands (Paulissen *et al.* 1985). Minor faulting also occurred in the early Weichselian. This gives rise to local thickening of the gravels north of the fault scarp.

The gravels and sands of the Kreftenheye Formation of The Netherlands have an upstream equivalent termed the Mechelen–Grosveld Terraces (Paulissen 1973; Zonneveld 1974). According to Juvigné (in Pissart (1974)) the Mechelen Terrace contains volcanic minerals derived from the Eifel area, eruptions from which are known to be of post-Eemian age.

The youngest or lowest terraces of the Meuse have been assigned to the Late Glacial and earliest Flandrian (Holocene) by Van den Broek & Maarleveld (1963) and Paulissen (1973). Consequently, formation of the present Meuse floodplain is thought to post-date the Preboreal.

(*f*) *Scheldt system*

The Scheldt system comprises a high-density network draining a series of small, low-altitude basins. It is surrounded on three sides, north, south and east, by the Meuse catchment and on the fourth by the North Sea where the valleys of Yser, Aa and Hem, etc., are partly drowned. The most important members of the Scheldt system include the Lys (Leie), Scheldt (Escaut), Dender (Dendre), Zenne (Senne), Dijle (Dyle) and Demer rivers. They drain the northern border of the Dinant Basin, the upper table country of the Namur Basin, and the northernmost part of France. The main streams, the Yser, Dendre, Zenne and Scheldt, are generally aligned parallel to the present coast but this alignment is abruptly interrupted by the east–west-trending Scheldt–Demer valley towards the Scheldt estuary. In general the rivers are flowing over Tertiary rocks, except for the upper headwaters of the Leie and Scheldt which drain the Mesozoic rocks of north France. In these cases the sole source of coarse detrital material is the Cretaceous rocks which provide flint, although silicified sandstones (sarsen) are also known from the Tertiary.

The primary consequent fluvial system seems to have developed on a marine regression surface towards a west-southwest–east-northeast coastline north of the present Scheldt–Demer valley during the late Neogene (Tavernier & De Moor 1974). This regression appears to be associated with general lowering of the North Sea Basin. By the early Pleistocene the coastline was aligned southwest–northeast and subsidence of the Lower Rhine Embayment and Central Graben caused the development of a bay (Zagwijn & Doppert 1978). This, accompanied by slight uplift of the Brabant Massif, encouraged the streams to flow northwards (Vandenberghe *et al.* 1986) (figure 2). According to Tavernier & De Moor (1974) the deposits of the pre-Quaternary streams are intensely chemically altered and suggest that warm, moist conditions prevailed during their deposition.

The detailed pattern of early Pleistocene drainage in Belgium is not known but it is presumed that deltaic deposits were laid down during the Tiglian under a perimarine environment (cf. Paepe & Vanhoorne 1970) in the Campine area. Regression of the sea allowed expansion of the rivers northwards into the southern Netherlands. Here a 'pre-Scheldt' river has been recognized by Vanderberghe *et al.* (1986) at Galder, where it was confluent with the early Meuse (see above). This river, flowing under a cold climate, had a braided form.

The onset of rapidly changing climates seems to have initiated alternating periods of downcutting and aggradation in the rivers. Early high-level terrace accumulations seem to be restricted to interfluve areas; this observation is interpreted by Tavernier & De Moor (1974) to reflect lateral displacement of streams. This is best seen in the trunk streams Leie and Scheldt. In the northernmost part of the area, during the Eburonian and Menapian, an alternation of fluvioperiglacial and perimarine deposits is found. This alternation reflects sea level oscillations, the former being formed by rivers during cold periods and the latter being deposited during the higher sea levels of the temperate periods when the sea invaded the lower reaches of river valleys, a pattern that continues to the present day. The northward-directed drainage system seems to have persisted throughout the Early Pleistocene and much of the Middle Pleistocene as well. For example, braided river deposits of the Scheldt can be traced to Dordrecht during the 'Cromerian Complex'. By the Elsterian the drainage system, particularly the upper parts of the southern valleys, were established with fluvioperiglacial terrace deposits, mostly gravels and sands, occurring on modern valley sides (e.g. Leie and Scheldt). The drainage probably continued to flow towards Breda early in this period. Later in the Elsterian the so-called Flemish Valley, an east–west-trending depression south of the present estuary occupied by the Scheldt and Leie, may have been excavated. Its formation may have been associated with the opening of the Dover Straits, which favoured east–west aligned drainage (Vandenberghe & De Smedt 1979) (figures 4 and 5).

Deep erosion of river valleys during the Elsterian enabled penetration of their lower courses by the sea after a eustatic sea-level rise during the Holsteinian. From this point onwards the Yser seems to have drained directly into the sea at the coast, whereas the Scheldt–Demer system entered the Flemish Valley. Meanwhile the lower courses of the rivers were accumulating estuarine or perimarine deposits, and fine freshwater sediments such as peats and organic clays were accumulating upstream.

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The Saalian Stage is well represented in the Scheldt System by two marked periods of valley incision each followed by aggradation of fluvioperiglacial gravels and sands. The Flemish Valley and the upstream extension, the Scheldt–Demer Valley, were greatly enlarged by lowering of interfluves by periglacial processes. During this time the series of obsequent northern valleys draining into the Demer river were probably established. With the onset of temperate conditions in the Eemian, marine trangression again invaded the lower parts of the valleys and submerged the Flemish Valley. The latter became a large gulf, the floor of which was subjected to tidal scour. Upstream, organic sediments accumulated in the valleys; on interfluve surfaces pedogenesis was active and peat was formed in the obsequent northern valleys.

Regression of the sea in the Weichselian accompanied a return to deep incision, followed by rapid deposition throughout the drainage system, the form of which was virtually as at present (Vandenberghe & De Smedt 1979; Tavernier & De Moor 1974) (figure 6). The fluvioperiglacial gravels and sands were laid down as before with the rivers adopting a braided mode. Organic sediments are occasionally found intercalated in the sediments formed as abandoned channel or hollow fillings. The dates, according to Tavernier and De Moor (1974), group between 45600 and 21113 years BP. In the Late Weichselian, cover-sand deposition became dominant in the area. This sediment buried the gravels and in some areas the rivers became sluggish and took on a meandering form (Vandenberghe & De Smedt 1979). Contemporary organic sediments from the cover-sand period have yielded radiocarbon dates of 12856–10025 years BP.

In the Flandrian (Holocene) a return to interglacial conditions has led once more to drowning of the lower parts of valleys by the sea and accumulation of fine organic and inorganic sediments upstream, interspersed by minor periods of incision (Vandenberghe & De Smedt 1979).

(g) *Thames System*

The River Thames System is the largest drainage basin in Britain. The Middle and Lower Thames forms the axial stream of the London Basin syncline, which comprises Cretaceous chalk and overlying Tertiary clays and sands. However, at Reading the river is aligned northwards, passing through a deep chalk valley from the so-called Upper Thames region. Here the river crosses a series of gently dipping Mesozoic clay and limestone formations. Upstream of Oxford several tributaries converge to form the trunk Thames river. This part of the catchment includes the Cotswold Hills and the south Midlands.

In the late Pliocene to earliest Pleistocene, the London Basin was submerged beneath the sea. Fossiliferous sands with a lithology and fauna similar to that of the Red Crag Formation of Suffolk occur at heights of up to 180 m on both the north and south sides of the basin (Chatwin 1927; Dines & Chatwin 1930). After reinvestigation of the North Downs localities these sands were assigned by John (1980) to the Headley Formation; Moffat & Catt (1986) have shown that the sands and possibly some gravel are of marine origin. Contours on the sub-Red Crag surface indicate a relative tilting of *ca.* 180 m between the western London Basin and the Suffolk coast localities. Assuming a broadly uniform age (Pre-Ludhamian = ?Praetiglian (Zalasiewicz *et al.* 1988)), either the eastern part of the basin must have subsided (West 1972; Moffat & Catt 1986), or the western part been uplifted in the early Pleistocene. It is not known whether the Thames existed at this time.

The subsequent marine regression was followed by the development of an early Thames drainage system. This is indicated by fragmentary deposits, termed Pebble Gravel, which occur at 120–130 m OD on the south Hertfordshire plateau. The gravels are composed predominantly of flint, much of which is derived from the Tertiary rocks. However, in some areas they contain a characteristic chert derived from the Lower Greensand of the Weald. These finds indicate the existence of south-bank tributaries (Wooldridge & Linton 1955). The Pebble Gravels, part of which was termed Nettlebed Gravel by Gibbard (1985), fall in height towards the northeast. Their precise age cannot at present be determined.

After a period of incision, a marked change in gravel composition is recorded in the next lowest unit, the Stoke Row Gravel (Gibbard 1985). This is dominated by a dramatic increase in the quartz and quartzite clast frequency; this material is apparently derived from Triassic conglomerates in the west Midlands. The downstream equivalent of this unit, the Baylham Common Gravel of Allen (1984), can be recognized in southern East Anglia and forms the uppermost member of the Kesgrave Formation. This abrupt increase is probably the result of an upstream capture, but it has also been attributed to glaciation in the west Midlands and north Wales (Bowen *et al.* 1986). The first influx of quartz-rich gravel into northern East Anglia was found in sediments of pre-Pastonian age by Hey (1976, 1980), Funnell *et al.* (1979) and West (1980*a*).

Four further quartz-rich gravel units occur in the Upper and Middle Thames and can be followed both into the Upper Thames and northeast into East Anglia (see, for example, Clarke & Auton 1982), i.e. north of the present eastern course. In the latter area they are referred to (with the Stoke Row and equivalents) as the Kesgrave Sand and Gravel Formation (Rose & Allen 1977), whereas upstream they form part of the Middle Thames Gravel Formation (Gibbard 1985) in the Middle Thames and the Northern Drift Formation (Hey 1986) in the Upper Thames (see Bowen *et al.* (1986) for summary). The next youngest member, the Westland Green Gravel, can be traced throughout the area and contains evidence of deposition under a periglacial climate with the river adopting a braided mode. It contains ice-rafted blocks and smaller clasts of volcanic rocks derived from north Wales (Hey & Brenchley 1977; Hey 1980; Green *et al.* 1980). The same sedimentary structures and lithologies occur in the lower units and the latter have been interpreted as evidence of multiple glaciation in the north Welsh Snowdon and Berwyn districts (Whiteman 1983; Bowen *et al.* 1986). These deposits presumably accumulated during the later Lower and early Middle Pleistocene, because temperate fossiliferous sediments of 'Cromerian Complex' age have been found intercalated within the younger members of the Kesgrave Formation (P. L. G., unpublished observations) (figure 3). The quartz-rich gravels are also present in the offshore region east of East Anglia, where they form part of the Yarmouth Roads Formation (Balson & Cameron 1985; Bowen *et al.* 1986). At this time, therefore, the Thames was contributing to the greatly expanded deltas that occupied the southern North Sea (cf. Zagwijn 1974).

Advance of the continental ice sheet in the Anglian–Elsterian caused burial of the Kesgrave Formation in all but the southernmost part of East Anglia. A characteristic palaeosol, the Valley Farm Soil, formed on the terrace surfaces; this soil was also buried by Lowestoft Till deposited by the ice sheet and forms an important marker horizon (Kemp 1985; Rose & Allen 1977; Rose *et al.* 1985, 1976). In addition, the Anglian ice sheet advanced into and overrode the contemporary valley of the Thames in Hertfordshire and also advanced into southern tributary valleys. This dammed the rivers and the resulting lakes progressively overspilled until the water reached an unglaciated valley, the Medway, via which the river could return to its earlier course in easternmost Essex (Gibbard 1977, 1979, 1985; Bridgland 1980, 1985) (figure 4). During the course of this glaciation the connection with the source of quartz-rich material seems to have been severed; the late Anglian Black Park Gravel and all subsequent units are poor in quartz in the Middle and Lower Thames (Gibbard 1985). This accords well with the Upper Thames, where the post-Northern Drift gravels are similarly poor in quartz (Briggs & Gilbertson 1973, 1980).

During the Hoxnian–Holsteinian temperate Stage, the rise of the eustatic sea level caused inundation of the Lower Thames. This transgression deposited estuarine sands, silts and clays on organic freshwater sediments at Clacton (West 1972; Bridgland 1980, 1985). Upstream the brackish water extended to Swanscombe, east of London, where fossil-bearing fine channel and overbank sediments accumulated. This interglacial period is poorly represented further up the valley, although organic sediments at Sugworth, previously correlated with the Cromerian, may be of this age (Shotton *et al.* 1980; Gibbard 1985). The river seems to have adopted a meandering channel pattern at this time.

The Wolstonian–Saalian of the Thames Valley is represented by a series of gravel and sand deposits laid down under periglacial conditions (figures 5 and 6). In the Middle and Lower Thames three major units, the Boyn Hill, Lynch Hill and Taplow Gravels, accumulated. These can be correlated in the Upper Thames with the Hanborough Terrace gravels, equivalent to both the Boyn Hill and Lynch Hill members; the Wolvercote Terrace gravels are equivalent to the Taplow member. The Wolvercote gravels contain flint, thought to have been derived from till at Moreton-in-the-Marsh (Bishop 1958; Briggs & Gilbertson 1980). Because the till has been equated with the Wolstonian glaciation of the Midlands, the gravels must either be contemporary with or post-date the glacial event. The previous aggradations, Boyn Hill, Lynch Hill and their equivalents, characteristically include considerable quantities of Late Middle Acheulian artefacts throughout the valley, but particularly in the Middle Thames region (Wymer 1968; Gibbard 1985).

At the end of the Wolstonian, after deposition of the Taplow Gravel and downcutting, accumulation of a further thin gravel unit gave way to deposition of sands, silts, clays and organic deposits in both the Thames and its tributaries during the Ipswichian–Eemian Stage. Once again sediments from this period are poorly preserved upstream of the estuary area, although some sites exist in the Upper Thames (Briggs & Gilbertson 1980; cf. Briggs *et al.* 1985). However, from Central London downstream a series of localities such as Trafalgar Square, Peckham, Aveley, Purfleet, West Thurrock and Grays, correlated with the Ipswichian, record a drowning of the valley by a rise of sea level (Hollin 1977). Water level in the estuary rose to *ca.* 10 m OD and caused considerable aggradation of silts and clays.

Fall of sea level in the Devensian–Weichselian led to incision into these estuarine sediments and the climatic deterioration resulted in a return to gravel and sand deposition by the braided river. During cold stages solution of Chalk bedrock often associated with scour has considerably modified gravel and sand accumulations leading to collapse or local thickening of sequences (Gibbard 1985). One such feature occurs in the Thames tributary Kennet Valley at Brimpton where Bryant *et al.* (1983) have described an Early Devensian profile preserved beneath younger gravels. Evidence of two interstadial and three stadial events was preserved at this site. In the Upper Thames, probable Early Devensian gravels occur beneath the Summertown–Radley Terrace (Goudie & Hart 1975; Seddon & Holyoak 1985) and these can be followed into the Middle and possibly the Lower Thames Valleys (Gibbard 1985 and unpublished). They were laid down under periglacial conditions (figure 6).

By the Middle Devensian further downcutting, followed by aggradation, led to accumulation of gravel and sands. These occur below the modern floodplain in the Upper Thames, but downstream they emerge and are found above the valley floor. In the London area organic deposits interbedded in the Kempton Park Gravel contain herb-dominated floras indicating a full glacial environment and give radiocarbon dates of 45–30 ka BP. Further downcutting and aggradation of valley-bottom Shepperton Gravel and equivalents occurred during the Late Devensian. In the Middle and Upper Thames dates of 14.5–10 ka BP have been obtained for organic deposits intercalated in the gravels. The organic deposits again contain fossils indicating the prevalence of cold climates with treeless shrub- or herb-dominated vegetation. This evidence implies a period of downcutting or non-deposition from 30–*ca.* 15 ka BP in the Thames system. During this interval, clayey silt 'brickearth' (Langley Silt Complex) sediment accumulated in the Middle Thames. This sediment, largely of local origin, contains a loess component and has been dated by thermoluminescence to a period centred on 17 ka BP (Gibbard *et al.* 1987).

In the Flandrian, organic sediment, tufa, clay, silt and sand have been deposited on the gravels and sands underlying the modern floodplain. This deposition has led to an infilling of depressions and the adoption by the rivers of single meandering courses. Since the Neolithic period, land clearance has given rise to accumulation of overbank clay and silt by flooding.

(*h*) *Somme*

The Somme is not strictly one of the 'great European rivers'. It is, however, very important historically: it has been investigated for over 150 years because of the abundance of mammalian remains and Palaeolithic artefacts recovered from its deposits.

The course of the Somme is strongly related to the geological structure of the region. It occupies the axis of a major northwest–southeast-trending syncline in which smaller anticlinal structures separate neighbouring minor tributaries; the Yères, Bresle, Anthie and Cache parallel the Somme and similarly occupy the floors of small synclinal structures. The age of this structure is thought to be Miocene, but later earth movements in the late Tertiary and possibly early Quaternary are also thought to have effected the area (Bourdier 1974*a*; Bourdier & Lautridou 1974*a*).

The present Somme Valley is cut for its entire length in Chalk. It drains the Picardy region or the northern part of the Paris Basin and flows broadly east to west to Amiens where it turns towards the northwest. The interfluve areas represent the dissected remains of a gently undulating surface underlain by a thick spread of clay-with-flints similar to those found in southern Britain (Bourdier 1974*a*; Bourdier & Lautridou 1974*a*; cf. Catt 1986).

No late Pliocene and early Quaternary deposits are found in the Somme region, but close proximity to the Seine region suggests that the two areas have evolved similarly (see below). It is therefore unclear when downcutting began, but it is attributed by Bourdier (1974*a*) to the initiation of periglacial climates and possibly to marine regression. The highest and therefore the earliest deposits are plateau gravels, which occur as thin patches near Grâce and downstream of Abbeville. They are much disturbed by later movement, cryoturbation and solutional collapse (Bourdier & Lautridou 1974*b*). They contain no evidence of dating and can therefore only be assigned to a period before deposition of the subsequent Very High Terrace deposits. Between deposition of the plateau gravels and the gravels and sands of the so-called Very High Terrace, about 25–30 m of downcutting occurred. At Montières–Grâce they are calcareous, in contrast with those occurring at lower levels (Bourdier *et al.* 1974*a*). Moreover, the gravels and overlying loess have yielded an assemblage of small vertebrates which according to Chaline (1974) demonstrate a 'Cromerian Complex' age. They also pre-date the Brunhes–Matuyama palaeomagnetic epoch boundary (Biquand 1974). A possible find of Ardennes sandstone in this deposit by Commont (1911), which would indicate a link with the Oise Basin, has never been confirmed by recent finds (Bourdier 1974*a*).

The next lowest sequence is that of the High Terrace, a complex unit which mainly comprises periglacial fluviatile gravels and sands much disturbed by cryoturbation. Throughout the valley Palaeolithic artefacts have been found in the gravels. For example, the type site of the Acheulian, St Acheul near Amiens, occurs in this terrace (Bourdier 1974*b*). In the Abbeville area the gravels are overlain by white sands of temperate origin. These sands, which may be in part of estuarine origin, have yielded the remains of large vertebrates that suggest correlation with the Cromerian *sensu stricto* of Britain (Bourdier 1974*b*) (?'Cromerian Complex Interglacial IV'). However, the Abbeville sands are most important for their rich so-called Early Acheulian ('Abbevillian') artefact assemblages.

The Middle Terrace is the next complex unit and can be subdivided into three subunits, MT I–III (Tuffreau *et al.* 1982). Throughout the Somme and its tributary, the River Avre, they again record a return to the deposition of coarse sands and gravels thought to have originated under periglacial conditions. Each subunit has yielded Acheulian artefacts. In the Avre Valley, particularly at Cagny near Amiens, sands, clays and loess overlie the gravels of MT II. The floral and faunal evidence (Bourdier *et al.* 1974*b*) indicate a climatic amelioration; a soil in the upper part of the sediments is equated with the Holsteinian Stage. This implies that the gravels beneath and possibly those of MT I are Elsterian, whereas those of the lower MT III may be of early Saalian age.

During subsequent Saalian time, deep incision of the valley was resumed; indeed the total downcutting in the Middle Pleistocene is of the order of 42 m (cf. Bourdier 1974*a*). Deposition of coarse gravel and sand of the Low Terrace (Tuffreau *et al.* 1981) again took place under a periglacial climate, according to Bourdier *et al.* (1974*c*); loess also accumulated. These gravels have again yielded Acheulian artefacts. However, overlying clays and loess contain abundant Levalloisian and Mousterian material. The gravels are covered at Longpré by fluvial sands, clays and tufa of temperate origin (Tuffreau *et al.* 1981; Sommé *et al.* 1984). These sediments are presumably of Eemian age, although the latter authors equate a palaeosol in the overlying loess with this period, and therefore the deposits would have to be pre-Eemian. At Montières, silts containing hippopotamus remains have also been recorded from beneath loess (Bourdier *et al.* 1974*c*).

The Weichselian of the Somme Valley is represented by loess covering older sediments and by downcutting to 12 m in total below the pre-existing sediments. The deposits of the Very Low Terrace are coarse gravels and sands and can actually be subdivided into two aggradations, one with its surface above the modern floodplain and a second underlying the floodplain. The latter are overlain by Postglacial peat and in turn by alluvial silts and clays. The silts and clays are attributed to inwash of fine sediment from land clearance and agriculture (Bourdier 1974*a*).

(*i*) *Seine System*

The River Seine and its major tributaries, the Marne, Oise and Aube, drain much of the north, central and eastern Paris Basin. Throughout almost the whole catchment the bedrock comprises Mesozoic and Tertiary rocks in which flint is the major pebble-forming lithology,

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although Tertiary siliceous sandstone (sarsen) is also present. The drainage courses are greatly influenced by local geological structure. For example, the Lower Seine downstream of Paris closely follows a shallow northwest–southeast-trending synclinal structure, which turns towards the southwest downstream of Rouen. This structure is also penetrated by two major faults trending in the same direction (figure 1): the Seine fault, which crosses the whole Paris Basin from Rouen to Limagne, and the Fécamp–Bolbec Fault, which runs from near Rouen westwards and continues beneath the Channel (Bourdier & Lautriodou 1974*a*). Minor tectonic activity has continued in the region into the Pleistocene.

Similarly to the Somme, the present Lower Seine is incised deeply into the gently rolling chalk country of the Upper Normandy Plateau. The surface is likewise underlain by often thick accumulations of clay-with-flints (Bourdier & Lautridou 1974*a*).

The earliest evidence of the Seine is recorded in the Lozère Sands (Kuntz & Lautridou 1974; Cavelier 1980). The deposits comprise quartz-rich sands, which are found in widespread patches through the Lower Seine, south of Paris and as far south as the Bourbonnais country south of Nevers. In the Lower Seine, the Lozère Sands were deposited in the delta when the river was apparently draining the northern Massif Central, an area now within the catchment of the Loire (figure 2). According to Macaire (1984), epeiorgenic uplift of the south and southwest part of the Paris Basin in late Pliocene–early Pleistocene caused drainage to flow to the west to form the Loire, rather than towards the Seine. At La Londe, southwest of Rouen, a small tectonic downfaulted depression contains a sequence that spans the Pliocene to Lower Pleistocene. These sands occur at the base and are overlain by the marine St Eustache Sands, which have been assigned to the Upper Brunssumian or Lower Reuverian (Clet 1982*a*). The Lozère Sands must therefore be presumed to be of early Pliocene age.

Renewed marine incursion in the Late Pliocene is represented by the St Eustache Sands, but these give way to lagoonal deposits in their upper levels and are replaced by a series of estuarine–lacustrine black and brown silts and clays of the La Londe Formation. Cooling of climate towards the top of the silt and clays is recorded by the palynology, but is not paralleled by any sedimentary change. The clays are correlated by Clet (1982*a*) with the Reuverian to Praetiglian.

The return of fluviatile conditions is signalled at La Londe by deposition of the bedded sands of the Fourmetot Formation (Kuntz *et al.* 1974, 1979). They are overlain by fluviolacustrine sand and gravel of the Roumois Formation. Intermediate between these units are ice-wedge casts, which indicate that periglacial climates prevailed during the intervening phase (possibly equivalent to the Eburonian (Kuntz *et al.* 1979)). However, these deposits are not thought necessarily to represent the Seine itself, but record cutting of local stream valleys under a cold climate. This suggests that major river incision in the region followed regression from the relatively high Tiglian sea levels (recorded at Bosq d'Aubigny in Cotentin (cf. Clet 1982*b*)) and the development of periglacial conditions.

The Lower Seine is characterized at present by its spectacular series of deeply incised meanders. On the basis of the history of fluvial aggradations it appears that these meanders were formed very early in the river's history and have subsequently been progressively modified (cf. Chancerel 1985, 1986). Although some rare pebble spreads occur above it, the first true terrace-like alluvial deposits are the strongly weathered gravels and sands of the Madrillet Formation, which were laid down after *ca.* 35 m of valley incision. The meanders cut directly into bedrock were already fully formed (Chancerel 1986). Further incision is followed by deposition of the Bardouville Formation. These first two aggradations show very little downstream gradient and may therefore indicate uplift of the coastal region (Alduc *et al.* 1979; Chancerel 1986); all subsequent terraces show a normal gradient upstream to Paris (Lautridou 1982). Two younger gravel and sand accumulations separated by a period of incision follow: the Radicatel and Berville Formations of Chancerel (1986). These two units may be equivalent respectively to the Forêt de Bord and the Rond de France Terraces of the Elbeuf meander (Lefèbvre *et al.* 1986). By their relation to younger aggradations these units are assigned broadly to the early Middle Pleistocene (Alduc *et al.* 1979) (figure 3).

Considerable problems arise for longitudinal correlation of fluvial aggradations in the meanders of the Lower Seine because of the varying style of deposition and erosion. This is attributed to the orientation of the meanders and to the possible migration of a 'knick-point' between the Elbeuf and Mantes regions (Lautridou *et al.* 1984). The result is that from Elbeuf downstream there are clearly developed aggradational units separated by periods of incision. In contrast, upstream near Mantes an almost continuous overlapping sequence of small-scale sedimentary units separated by low step-like cliffs is found (Lautridou 1982). Correlation between the regions can only be achieved by using 'marker' units (Lautridou *et al.* 1984; Lécolle 1984; Chancerel 1985, 1986).

The next youngest aggradation is the St Pierre-les-Elbeuf Terrace of the Martot Formation or Middle Terrace (Alduc *et al.* 1979; Lautridou 1982; Lautridou *et al.* 1984). This unit is equivalent to the Bois Delamare Complex as seen at Anneville, where the sediments contain evidence for at least three erosional and depositional events (Chancerel 1986). The gravels are periglacial in origin (Lefèbvre *et al.* 1986). At the St Pierre type section, fluvial gravels and sands underlie an interglacial tufa and a very thick loess sequence which includes four palaeosol horizons. Correlation of the basal soil with the Holsteinian suggests that the terrace gravel and sands are of pre-Holsteinian, i.e. Elsterian or more probably late 'Cromerian Complex', age. Upstream the 'High Terrace' (P1, XII–XV) of Lécolle (1984) can be equated with these sediments on the basis of stratigraphical relations. The deposits are greatly disturbed by post-depositional bedrock solutional collapse and periglacial cryoturbation.

Possible estuarine deposits resulting from high eustatic sea level drowning the valley during the Holsteinian Stage may occur at Cléon near Tourville (Lautridou 1982) and possibly in part of the Tankerville sequence.

The next lowest aggradation is the Oissel Formation or Pont de l'Arche Terrace (Lefèbvre *et al.* 1986); this formation is equated with the Elsterian Stage (formations VII–XI of Lécolle 1984). The subsequent Tourville Formation and the Criquebeuf Terrace (Lefèbvre *et al.* 1986) are broadly of Saalian age (Lautridou 1982; Alduc *et al.* 1979). The base of the Tourville Formation comprises periglacial, fluvial gravels and sands that have yielded an abundant fauna of cold affinities. Within these gravels a silt horizon of warm climatic character is preserved. The gravels and sands are overlain by estuarine sands of the Tankerville Member, the latter also well developed at Tankerville itself and at Cléon (Alduc *et al.* 1979; Lautridou 1982). Part of this sequence is probably of Eemian age.

Post-Eemian dowcutting was followed by two further gravel and sand aggradations that, in the modern Lower Seine, are buried by Flandrian (Holocene) alluvium and estuarine sediments (Lefèbvre *et al.* 1974). These two units, the Rouen Formation (Porcher 1977) or Formations I and II of Lécolle (1984), are assigned to the Weichselian: the higher to the

Middle Weichselian, and the basal gravels flooring the valley to the Late Weichselian, according to Lefèbvre (1974), Alduc *et al.* (1979) and Lautridou *et al.* (1984).

Investigation of the offshore Seine Bay–Fosse Cotentin area by Alduc *et al.* (1979) has shown that it is possible to follow equivalents of Weichselian, Saalian and possibly pre-Saalian formations into the Fosse Centrale, north of Cotentin. At least two Weichselian accumulations, together with probable equivalents of the Tourville, Oissel and St Pierre Formations, i.e. six terraces, are represented (Lautridou 1982) (figures 4–6).

(*j*) *Channel River system and the Straits of Dover*

Detailed investigations over the past 30 years have revealed considerable evidence of the morphology and geology of shelf seas adjacent to northwest Europe. Investigations by French and British workers have shown that the Channel floor is generally smooth with its surface gently inclined from the continental shelf margin to the Dover Straits. Near the coasts, steeper, ramp-like features are present (Smith 1985). This smooth surface is underlain over most of the channel by a thin, often mobile sediment cover beneath which bedrock, much of which is of Mesozoic age, occurs (Larsonneur *et al.* 1979; Smith & Curry 1975). Cut into this bedrock in the central and eastern Channel is a complex series of narrow valleys or channels, most of which are infilled by unconsolidated sands, gravels and clays (Dingwall 1975). The best known of these, the Hurd Deep, has been shown by Hamilton & Smith (1972) to contain a multiple infilling and to have a composite origin by combined fluvial deposition and tidal scour. Dingwall (1975) and Smith (1985) have subsequently shown that the valleys are interlinked to form a drowned drainage system formed during periods of lowered sea level. The concept of greatly lowered sea levels during cold stages is well established; however, local evidence suggests sea-level minima of 90–130 m below present levels (for summary see Hamilton & Smith 1972). Fluvial downcutting and infilling during these periods would be followed by tidal scour by marine transgression during periods of rising and falling sea levels.

In detail, the valleys form a complex anastomosing system and many are overdeepened, particularly at confluence points, a feature also found in fluvial systems on land, e.g. the Thames (Berry 1979). Smith (1985) has demonstrated that they can be linked with the Seine, Somme, Béthune, Solent, Arun and other coastal rivers. The central or Lobourg valley can also be followed to the Dover Strait where it merges into the Fosse Dangeard of Destombes *et al.* (1975), a series of scour hollows formed on the Gault–Lower Chalk contact (Smith 1985).

There is little doubt therefore that the valley system is mainly of fluvial origin and that it functioned during periods of low eustatic sea level. This is contrary to suggestions that the system may have originated by glaciation, as proposed by Destombes *et al.* (1975) and Kellaway *et al.* (1975) (see Zagwijn (1979) and Oele & Schüttenhelm (1979) for further discussion).

Evidence for the age of this river system and therefore for the present form of the Channel is rather limited by problems of access. However, the system must certainly have been at least partly in existence by the time of deposition of the earliest river sediments in the neighbouring terrestrial river valleys. On the basis of the work by Alduc *et al.* (1979) this would indicate that valley formation pre-dates deposition, possibly of the Quillebeuf and certainly of the St Pierre-des-Elbeuf Terrace gravels of the Seine, i.e. it is at least pre-Elsterian or 'Cromerian Complex'. Earlier river terrace deposits are apparently not represented and the Plio-Pleistocene sequences

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found on the neighbouring land appear to be unrelated to the Channel palaeovalley system (cf. Smith 1985). Certainly the Channel existed in the Pliocene, to judge from the occurrence of marine sediments at St Erth, Cornwall (Jenkins *et al.* 1986), at La Londe, Normandy, and at several sites in Brittany (Lautridou *et al.* 1986) (figure 2). It would seem likely, therefore, that for much of the Lower Pleistocene the area was one of net erosion and, because there is some evidence for tectonic activity in the Western Approaches Basin during the early Pleistocene (Smith & Curry 1975), this may be partly the cause. It is conceivable that this uplift may have been contemporary with that experienced in the London Basin (§ (*g*) above).

River downcutting and deposition presumably occurred throughout the early Middle Pleistocene during periods of low sea level; contemporary periglacial fluvial deposits were laid down by the Somme, Seine and possibly some of the English South Coast rivers, e.g. the Solent River. A proto-Channel River system must have existed at the time (figure 3). However, the question of the existence of the Dover Straits is relevant here. Much debate has centred on the presence or absence of this passage and when precisely it was formed. In general it is assumed that the breach was established in the late Pleistocene (Destombes *et al.* 1975).

On the basis of Pliocene marine faunas in East Anglia, Funnell (1972) suggested that a link between the North Sea and the Atlantic through the Dover Straits was present at the time. However, Jenkins *et al.* (1986), after investigations of Pliocene foraminifera at St Erth, showed that the southern North Sea faunas were, by comparison, impoverished and indicated that no link was present. This view is supported by Zagwijn (1979).

Whether or not a Dover Strait existed in the Middle Pleistocene is also questioned, because sediments interpreted as indicating marine deposition occur in northern France at Herzeele (Sommé 1979; Paepe & Sommé 1975) and in southwest Belgium at Loo (Vanhoorne 1962; Paepe & Baeteman 1979). At least two marine transgressions have been recognized in these sediments, although part of this sequence, at least, is assigned to the Holsteinian (Vanhoorne 1962). This indeed would mean that the Straits may have been open before the Holsteinian; Zagwijn (1979) suggests possibly the Pastonian. Nevertheless, there is strong reason from indirect biostratigraphical evidence (West 1980*b*) and from subsequent events to believe that a substantial land barrier existed for much of the Pleistocene. This barrier was the extension of the Weald–Artois anticline in the Chalk (Smith 1985). Such a barrier may conceivably have been overtopped occasionally during periods of high sea level but for the prolonged cold stages it would have been dry land and drained by consequent streams analogous to those draining the neighbouring regions today (cf. Stamp 1927).

According to Zagwijn (1974, 1979), the southern North Sea Basin was greatly infilled during the Lower and early Middle Pleistocene by delta sediments, derived from the North German rivers, the Rhine and the Meuse, represented by the Yarmouth Roads and Winterton Shoal Formations offshore (Balson & Cameron 1985; Cameron *et al.* 1987). Smaller rivers such as the Thames also contributed to this huge delta complex (figure 3). Throughout this period and indeed into the Elsterian these rivers drained northwards. The advance of continental ice sheets from the north in the Elsterian–Anglian Stage would therefore be likely to have blocked the North Sea Basin, preventing water from escaping to the Atlantic, thus causing an immense ice-dammed lake to form in the remaining ice-free part of the basin (figure 4). Clearly such a lake would only have formed if the British and Scandinavian ice sheets coalesced; there is, however, much evidence to support this, such as the occurrence of Norwegian erratics (e.g. rhomb porphyry and larvikite) in East Anglia. In spite of this, until the ice sheets did form a barrier, the drastically lowered eustatic sea level and isostatic crustal depression would be expected to have caused major incision of river valleys. Just such an incision is indicated by deep valleys beneath the widespread freshwater glaciolacustrine sediments of the Peelo Formation, preserved extensively in the northern Netherlands (Ter Wee 1962, 1983*a*). Indeed, these valleys are so deep that they have been compared to those demonstrably formed beneath the ice sheet in Germany and East Anglia, even though they contain river sediments of southern derivation and were apparently never overridden by ice (Ter Wee 1983*a*; Zagwijn 1974, 1979). Nevertheless the considerable depth of the valleys (some reach to over 350 m locally (Ter Wee 1983*a*)) suggests that processes other than solely fluvial downcutting may also have been involved in their formation. Large areas of the North Sea adjacent to The Netherlands are also underlain by the Peelo Formation, according to Oele (1969, 1971); however, at least some of these deposits in the Dogger Bank area have more recently been correlated with the Saalian by Zagwijn (1979). Nevertheless, Balson & Cameron (1985) and Cameron *et al.* (1987) have identified widely distributed glaciolacustrine sediments of the Svarte Bank Formation in the British offshore sector.

Evidence from the British side of the North Sea has not previously been linked with that in The Netherlands. However, it has long been recognized that the Corton Sands, which occur intermediate between the Cromer Till and Lowestoft Till Formations in eastern East Anglia, were deposited during local ice recession (Corton Interstadial) in a substantial ice-dammed lake (West & Wilson 1968; Pointon 1978; Bridge & Hopson 1985). More recently, re-examination of Cromer Till sequences in Norfolk by the author and colleagues (P. L. G., unpublished results) has indicated that large parts of these sediments originated from the ice sheet's advancing into a large waterbody, because much of the till is of waterlain origin (cf. Kazi & Knill 1969; Gibbard 1980). Associated sand members within this formation appear to represent meltwater discharge events.

There is therefore widespread evidence for deposition in a glaciolacustrine environment during the Elsterian–Anglian Stage in the southern North Sea Region. The presence of such sediments above present sea level implies either isostatic depression of the area or elevation of lake level or else an interplay of both factors.

The water level in such a lake, formed by inwash from the rivers draining into the southern North Sea and from ice-sheet meltwater derived from much of western Europe, would undoubtedly rise until a suitable drainage outlet was reached. It seems highly likely that this outlet occurred at what is now the Straits of Dover (cf. Smith 1985). Indeed, a thick channel fill of gravels, clays and sands, deposited by a large southward-flowing river, has been identified near Wissant on the northwest French coast (Roep *et al.* 1975). However, it is probable that more than one channel functioned at this time and that the second became the Straits (figure 4).

When discussing this possibility Smith (1985) proposed that outflow from an ice-dammed southern North Sea lake would be catastrophic, comparable to the outburst from Lake Missoula in the northwest United States in the last glaciation. There is, however, a major difference between the latter and the Dover Straits examples. The drainage of Lake Missoula resulted from collapse of an ice-dam that blocked a valley, whereas a substantial bedrock Chalk ridge was present at the Dover Straits at this time. Overspill of an ice-dammed lake of the latter type would be expected to be balanced only by input, so that when a bedrock col was reached, the resulting erosional capacity of the water would be equivalent to the water entering the lake.

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Given that chalk is highly susceptible to dissolution by cold water, downcutting by the outflow stream may have been relatively rapid, but certainly not catastrophic. Such an overspill may have followed pre-existing stream valleys, joined the Channel River and discharged through this to the sea.

Progressive enlargement of the outflow col would have followed, the southward flow possibly causing rivers in the adjacent areas of the southernmost North Sea to take up courses towards the outflow as the col was lowered. Although realignment of the Scheldt System streams has been attributed to this process by Vandenberghe & De Smedt (1979) and Vandenberghe *et al.* (1985), there is as yet no unequivocal evidence for a similar adjustment of the Thames. On the other hand, Thames deposits do show a southward tendency in the offshore Thames Estuary area (D'Olier 1975; D. R. Bridgland & B. D'Olier, personal communication). In addition, no Thames deposits of post-Elsterian age have been found in the offshore region so far investigated by the British Geological Survey (T. D. J. Cameron, personal communication).

From the distribution of laminated (varved) clays of the Peelo Formation and the remarkably similar Lauenburg Clay deposits found in northwest Germany and as far north as Denmark (Ehlers 1983), as well as laminated clays filling deep glacial valleys in East Anglia, such as the Nar Valley (Ventris 1985), and offshore (Balson & Cameron 1985) it is clear that glaciolacustrine conditions prevailed in the North Sea Basin until very late in the Stage. It is not certain whether all the finds of laminated clays represent a single lake or not, particularly during the deglaciation period. Nevertheless, the widespread occurrence of these sediments strongly suggests that the North Sea lake persisted as a major feature during the Late Anglian–Elsterian period. This implies that the Dover Straits outlet may have been subject to isostatic uplift, which prevented complete drainage of the lake. Such uplift might be expected to have favoured adoption of northerly courses by the rivers. However, because the lake seems to have persisted very late, downcutting of the Dover Straits col could have continued, keeping pace with uplift, enlarging the trough and forcing those rivers near the col to adopt southerly courses.

Similar events to those in the Elsterian may have been repeated in the Saalian Stage; glaciolocustrine sediments, albeit of smaller scale, are again found associated with the Drenthe ice advance off the Dutch coast (Oele & Schüttenhelm 1979). The Drenthe advance certainly forced the Rhine and Meuse to flow westwards (De Jong 1965; Zagwijn 1974; Ter Wee 1983*b*), so that, during this period of low eustatic sea level, the Rhine, Meuse, Thames and Scheldt could have become confluent and drained through the Dover Straits into the Channel River (figure 5). Such a pattern might, however, require an ice-dam across the North Sea during this time, a possibility that seems less likely from the latest investigations (cf. Cameron *et al.* 1987). Nevertheless, both Oele & Schüttenhelm (1979) and Jelgersma *et al.* (1979) have stated that the Rhine and Meuse drained southwards after the Drenthe Substage until the Late Weichselian, on the basis of the distribution of the Kreftenheye Formation sediments. If these rivers did so, then the Thames and Scheldt must surely also have adopted this course (figure 6) despite recent studies of the bedrock topography by B. D'Olier & D. R. Bridgland (personal communication) that suggest this might not be the case for the former. Clarification of this matter must await further research on the deposits.

It seems likely, however, that the formation of the Dover Straits has been polycyclic, so that once the breach was formed by fluvial action, it was enlarged by tidal scour after interglacial

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marine transgression, then fluvially dissected once more during a subsequent period of low sea level, and so on.

The progressive increase in size of the Dover Straits is reflected in the indirect evidence of impoverishment of British interglacial palaeontological assemblages through the later Pleistocene (cf. West 1980*b*). For example, during the Holsteinian Stage the Dover Strait may have been land for much of the interglacial, only being breached during the period of maximum sea level. However, by the Eemian (Ipswichian) a substantial barrier to plant and animal migration seems to have been present.

It is interesting to note that the origin of the Dover Straits and the existence of a substantial southern North Sea ice-dammed lake suggested above were proposed originally by Stamp (1927) and regularly ever since by Zonneveld (1958), Gullentops (1974), Dingwall (1975) and most recently by Smith (1985).

It is clear therefore that the Channel River has been a major drainage line throughout much of the later Pleistocene and for at least part of its history carried the runoff of a huge area of northwest Europe. If the correlations proposed here are correct, it last functioned during the Late Weichselian and earliest Postglacial (Alduc *et al.* 1979).

Discussion and conclusions

From the foregoing, it is apparent that the response of the northwest European drainage system to the external influences of the past three million years has been complex. However, certain patterns emerge and indicate the impact of climatically and tectonically controlled variables on the system as a whole. These patterns will be discussed below.

The foundations of the modern drainage system were laid in the Miocene, when earth movements associated with both the Alpine Orogeny and the opening of the North Atlantic were at their height. Although greatly modified subsequently, the precursors of the modern rivers can be identified. The form of this system was apparently closely linked to geological structure and took a broadly simple form, although this may be an artefact of the preservation of the sedimentary sequences.

Throughout the Miocene and Pliocene the rivers were generally transporting chemically resistant minerals and lithologies, such as quartz or flint, derived from long-term weathering under moist warm-temperate climates. Relatively little mechanical weathering is recorded except when associated with tectonic uplift, which resulted in rejuvenation of rivers such as the Rhine and gave rise to deposition of gravels. Otherwise, rivers seem to have occupied shallow channels and meandered across large regions with little evidence of deep valley incision. There is no indication of the development of permafrost, although some seasonal frost activity did possibly occur in the late Pliocene.

The onset of true cold climates in the Praetiglian (Pleistocene) Stage caused a marked change in depositional style with the input of gravel and sand derived by stripping of the deep regolith inherited from the Pliocene, and the adoption of a braided form by the rivers Rhine and Meuse. This however did not have great impact across the entire region, to judge from the apparently limited change in the Seine system. The return to temperate conditions in the Tiglian caused a readoption of the meandering river form and a renewed deposition of fine sediments.

In the Praetiglian, glaciation may already have been established in the Alps; ice-rafted blocks occur in Rhine sediments. By this time the Alpine Rhine may have been captured by the Upper Rhine system, having previously been flowing to the southwest to join the Rhône. Whether or not glaciation occurred elsewhere in northwest Europe at this time is not clear.

The return to cold climates in the Eburonian–Menapian period established considerable frost weathering and therefore the loading of rivers with freshly weathered detrital material. The resulting periodic incision and later deposition under braided river régimes is well marked. The great increase in sediment deposition is recorded by the vast expansion of deltas in The Netherlands (Zagwijn 1974). Glaciation seems to have become a widespread phenomenon, not only in the Alps, but also in upland Wales and in Scandinavia at this time. The result of the latter was the destruction of the Baltic River in the Menapian, after the possible formation of a proto-Baltic Basin by glacial scour.

The first records of the prevalence of true permafrost in lowland areas comes from the Dutch–Belgian border where cryoturbation of Tiglian T4c age has been observed (J. Vandenberghe, personal communication). Moreover, at La Londe in the Seine region, ice-wedge casts occur at the base of the Roumois Formation and are thus assigned to the Eburonian Stage. Nevertheless, it would seem very unlikely that permafrost did not occur before this in the Praetiglian, to judge from the evidence for persistent cold climates.

By the later Middle Pleistocene the pattern of repeated incision and sediment accumulation was well established. However, advance of continental ice sheets into lowland northwest Europe for the first time in the Elsterian Stage had widespread effects. The area overridden by the ice was subjected to total landscape remodelling, with old river courses destroyed or buried. An entirely new landscape was formed beneath the ice by glacial and glaciofluvial erosion and deposition; the sculpturing of deep glacial valleys was to have a striking palaeogeographic impact after the ice retreat. At the ice margin, major river valleys were dammed all across the region. The Thames and its tributaries were diverted southwards, the Elbe was dammed and the North German rivers were deflected westwards. However, the most striking feature was the development of a massive ice-dammed lake in the southern North Sea, into which the Thames, Rhine, Meuse, Scheldt and possibly the Ems all discharged. Overspill of this lake almost certainly initiated the Dover Straits and greatly enlarged the Channel River system.

Subsequent glaciation in the Saalian and Weichselian Stages had comparable effects, once again causing major drainage diversion and landscape remodelling. In the Saalian, major ice-pushed ridges in the Netherlands and Germany forced the Rhine to follow a more southerly course; after deglaciation the Elbe took up its present course through Hamburg. Glaciation in the Weichselian seems to have influenced the drainage system least, as it did not reach lowland northwest Europe except in the extreme north: where meltwater discharged through the Elbe, in central and northern Britain, and in the Alps where water discharged along the Rhine. The rest of the region outside that subjected to glaciation had a periglacial climate during this period. In periglacial areas, where both direct and indirect effects of glaciation have been minimal, rivers have maintained their courses except when river capture has taken place, e.g. the Seine and Somme.

Throughout the cold stages of the Pleistocene the rivers overwhelmingly deposited gravels and sands derived either from periglacial weathering or glacial sources. The result is that valley systems contain vast thicknesses of cold-climate sediments deposited by rivers that flowed in a braided or wandering, often multi-channelled, form. These sediment accumulations are generally separated by periods of non-deposition or incision outside subsiding areas. This

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incision was previously often attributed to interglacial events. It is now, however, clear from modern process studies that incision almost certainly also occurs when river runoff is highly seasonal but when limited supplies of detritus are available, i.e. predominantly under cold climates. Highly peaked discharges characterize modern rivers in periglacial regions. This pattern is caused by storage of water as snow on the land throughout the winter and subsequent rapid melting in spring, giving rise to the nival flood event. Such events may carry much of the year's precipitation in a few days. The result is high flow velocities and therefore strong erosion and transporting power of very short duration. In glacial meltwater streams these discharge peaks may be modified by cycles of glacial melt and discharge of stored water, such as jökulhlaups. Lack of vegetation and the development of permafrost ensure a rapid return of precipitation in catchments to rivers, so that storms may also produce marked flood events in such periglacial streams.

This situation contrasts markedly with that pertaining under temperate climates, where the dense vegetation cover and the development of ground-water storage causes a cushioning of storm or spring melting effects. Vegetation cover protects the land surface by fixing and stabilizing the mineral soil, and plants also take up water that is transpired directly into the atmosphere. Storage of water and slow flow in the ground reduces flood events by controlling the volume of water entering a drainage system. This results in river discharge becoming much less peaked and causes rivers to flow all year round. In general the lack of coarse material available, except that obtained by erosion of the channel itself, and the reduced flow velocities prevent lowland rivers from greatly altering their courses. They tend therefore to adopt single-thread meandering courses with floodplains, the hollows in which become filled with organic sediments or flood silts, clays or sands.

These observations are borne out by the sequences preserved throughout the region, in which interglacial sediements tend to be fine inorganic or organic materials intercalated in gravel and sand sequences. For this reason they have a low preservation potential and in general represent less than 10% of the total fluvial sequences. The remaining 90% of sediments are of cold-climate origin.

The comparison of modern, deeply incised river valleys with the apparently shallow features predominant during the earlier Neogene strongly suggests that alternating deposition and incision, which gives rise to these deep valleys, is a direct consequence of rapid climatic change. In particular it implies that frequent downcutting and aggradation in the region, where tectonic activity can be discounted, is a result of the occurrence of cold climates and the supply of abundant fresh materials by periglacial processes.

The effects of additional climatically controlled parameters are, with one exception, poorly understood. For example, the variation in amount of precipitation and its distribution through the year will certainly influence river flow and habit. However, relatively few studies have been undertaken in sufficient detail to determine this. Exceptions include those by De Gans (1981), Rose *et al.* (1980), Vandenberghe & De Smedt (1979) and Vandenberghe *et al.* (1984).

The one exception to this general lack of knowledge is sea-level change. Glacioeustatically and isostatically controlled sea level changes are characteristic of the Pleistocene. In northwest Europe low sea levels during the cold periods caused great expansion of the drainage system onto the surrounding continental shelves. Indeed, deposition on the huge river floodplains of the North Sea and Channel areas almost certainly provided the major source of silt that formed the widespread loess deposits present in adjacent land areas.

For much of the Lower and Middle Pleistocene the southern North Sea was occupied by the

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huge delta complex of the North German Rivers, the Rhine, Thames, Meuse and Scheldt. After the early Pleistocene, marine incursions during interglacial stages remained absent until the late Middle Pleistocene Cromerian Stage. A similar feature is found in the Channel Region. Marine transgression has subsequently been particularly marked during interglacial stages, and sea levels over 100 m below present are known from glacial maxima. Truncation of the greatly expanded fluvial system by marine transgression is very significant; the Channel River, for example, was approximately 800 km in length, 3.3 times longer than the present River Thames. Moreover, the drowned fluvial deposits are subjected to tidal scour, deposition of marine sediments and remobilization.

A further effect of marine transgression has been the infilling of the pre-existing river valleys by wedge-like accumulations of estuarine sediments. These sediments comprise silts, clays, and in some cases sands interbedded with peats and detrital organic sediments (for Flandrian (Holocene) examples see Devoy (1979); Jelgersma *et al.* (1979); Huault *et al.* (1974); Porcher (1977). Interglacial examples are also widespread; in areas of subsidence they are buried by later deposits, whereas in relatively stable areas they are dissected by subsequent fluvial incision and remain as eroded remnants on valley sides.

It therefore is apparent that the impact of the various forms of climatic change on the northwest European drainage system has been very great. These changes have been superimposed on long-term climatic and geological trends resulting from plate tectonics and world geographical evolution over the past three million years or so. These trends might reasonably be expected to continue into the future.

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References

Aguirre, E. & Pasini, G. 1985 The Plio-Pleistocene boundary. *Episodes* **8**, 116–120.

Alduc, D., Auffret, J.-P., Carpentier, G., Lautridou, J.-P., Lefèbvre, D. & Porcher, M. 1979 Nouvelles données sur le Pléistocène de la basse vallée de la Seine et son prolonguement sous-marin en Manche Orientale. *Bulletin d'information des géologues du Bassin de Paris* **16**, 27–34.

Allen, P. 1984 *Field guide to the Gipping and Waveney Valleys. May, 1982.* Cambridge: Quaternary Research Association.

Averdieck, F.-R. 1971 Palynologische Untersuchungen zum Tertiär auf Sylt. *Meyniana* **21**, 1–8.

Balson, P. S. & Cameron, T. D. J. 1985 Quaternary mapping offshore East Anglia. *Mod. Geol.* **9**, 221–239.

Berry, F. G. 1979 Late Quaternary scour hollows and related features in central London. *Q. Jl Engng Geol.* **12**, 9–29.

Bibus, E. 1980 Zur Relief, Boden- und Sedimententwicklung am unteren Mittelrhein. *Frankfurter geow. Arb.* D, band 1. (296 pages.)

Bijlsma, S. 1981 Fluvial sedimentation from the Fennoscandian area into the North-West European Basin during the Late Cenozoic. In *Quaternary geology: a farewell to A. J. Wiggers* (ed. A. J. van Loon). *Geologie Mijnb.* **60**, 337–345.

Biquand, D. 1974 Position chronologique de la très haute nappe alluviale de Grâce (vallée de la Somme) par rapport à la limite paléomagnétique Brunhes–Matuyama. *Bull. Ass. fr. Étude Quat.* **40–41**, 157–159.

Bishop, W. W. 1958 The Pleistocene geology and geomorphology of three gaps in the Midlands Jurassic escarpment. *Phil. Trans. R. Soc. Lond.* B **241**, 255–306.

Boenigk, W. 1978 Die Flussgeschichtliche Entwicklung der Niederrheinischen Bucht im Jungtertiär und Altquartär. *Eiszeitalter Gegenw.* **28**, 1–9.

Bourdier, F. 1974*a* Essai sur le creusement de la Vallée de la Somme au Quaternaire. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale* (ed. P. Macar), pp. 241–252. Liège: Centre de Société Géologique de Belgique.

Bourdier, F. 1974*b* Le complexe Mindelien. I. La haute terrasse de la Somme. *Bull. Ass. fr. Étude Quat.* **40–41**, 165–168.

Bourdier, F. & Lautridou, J.-P. 1974*a* Les grands traits morphologiques et structuraux des régions de la Somme et de la Basse-Seine. *Bull. Ass. fr. Étude Quat.* **3–4**, 109–111.

Bourdier, F. & Lautridou, J.-P. 1974*b* Les dépôts du Quaternaire ancien. *Bull. Ass. fr. Étude Quat.* **40–41**, 129–135.

Bourdier, F., Chaline, J., Munaut, A. V. & Puissegur, J.-J. 1974*a* La très haute nappe alluviale de la Somme. *Bull. Ass. fr. Étude Quat.* **40–41**, 137–143.

Bourdier, F., Chaline, J., Munaut, A. V. & Puissegur, J.-J. 1974*b* Le Complex Mindelien. II. La moyenne terrasse de L'Avre. *Bull. Ass. fr. Étude Quat.* **40–41**, 168–180.

Bourdier, F., Munaut, A. V., Prat, F. & Puissegur, J.-J. 1974*c* Les dépôts du complexe Rissien de la Somme. *Bull. Ass. fr. Étude Quat.* **40–41**, 219–227.

Bowen, D. Q., Rose, J., McCabe, A. M. & Sutherland, D. 1986 Correlation of Quaternary glaciations in England, Ireland, Scotland and Wales. *Quat. Sci. Rev.* **5**, 299–340.

Bridge, D. McC. & Hopson, P. M. 1985 Fine gravel, heavy mineral and grain-size analyses of mid-Pleistocene glacial deposits in the Lower Waveney Valley, East Anglia. *Mod. Geol.* **9**, 129–144.

Bridgland, D. R. 1980 A reappraisal of Pleistocene stratigraphy in north Kent and east Essex and new evidence concerning the former course of the Thames and Medway. *Quat. Newsl.* **32**, 15–24.

Bridgland, D. R. 1985 The Quaternary fluvial deposits of north Kent and east Essex. Ph.D. thesis (C.N.A.A.)., City of London Polytechnic.

Briggs, D. J. & Gilbertson, D. D. 1973 The age of the Hanborough Terrace of the River Evenlode, Oxfordshire. *Proc. Geol. Ass.* **84**, 155–173.

Briggs, D. J. & Gilbertson, D. D. 1980 Quaternary processes and environments in the Upper Thames Valley. *Trans. Inst. Br. Geogr.* **5**, 53–65.

Briggs, D. J., Coope, G. R. & Gilbertson, D. D. 1985 The chronology and environmental framework of early man in the Upper Thames Valley: a new model. *Oxford*: *Br. Archaeol. Rep. Br.* Ser., no. 137. (176 pages.)

Brunnacker, K. 1975 The Mid-Pleistocene of the Rhine Basin. In *After the Australopithecines* (ed. K. W. Butzer & G. L. Isaac), pp. 189–224. The Hague: Mouton Publishers.

Brunnacker, K. 1978 Gliederung und Stratigraphie der Quartär-Terrassen am Niederrhein. *Kölner geogr. Arb.* **36**, 37–58.

Brunnacker, K. 1986 Quaternary stratigraphy in the Lower Rhine area and northern Alpine foothills. *Quat. Sci. Rev.* **5**, 373–379.

Bryant, I. D., Holyoak, D. T. & Moseley, K. A. 1983 Late Pleistocene deposits at Brimpton, Berkshire, England. *Proc. Geol. Ass.* **94**, 321–343.

Bustamente Santa-Cruz, L. 1974 Les minéraux lourds des alluvions du bassin de la Meuse. *C. r. hebd. Séanc. Acad. Sci., Paris* **278**, 561–564.

Cameron, T. D. J., Stoker, M. S. & Long, D. 1987 The history of Quaternary sedimentation in the UK sector of the North Sea Basin. *J. geol. Soc. Lond.* **144**, 43–58.

Caston, V. M. D. 1977 A new isopachyte map of the Quaternary of the North Sea. *Rep. Inst. geol. Sci.* no. 77/11, pp. 1–8.

Catt, J. A. 1986 The nature, origin and geomorphological significance of the clay-with-flints. In *The scientific study of flint and chert* (ed. G. de G. Sieveking & M. B. Hart), pp. 151–159. London: Cambridge University Press.

Cavelier, C. 1980 Miocène et Pliocène. In *Synthèse géologique du Bassin de Paris* (ed. C. Megnien & F. Megnien) *Mem. Bur. Rech. Gèol. Miner.* **101**, 416–437.

Cepek, A. G. 1967 Stand und Probleme der Quartärstratigraphie im Nordteil der DDR. *Ber. dt. Ges. geol. Wiss.* A **12**, 375–404.

Chaline, J. 1974 Les rongeurs, l'âge et l'environnement de la très haute terrasse de Grâce à Montières (Somme). *Bull. Ass. fr. Étude Quat.* **40–41**, 151–157.

Chancerel, A. 1985 *Le val de Seine D'Elbeuf à Caudebec-en-Caux. Evolution morphologique*. Thèse, 3e cycle, Université Paris 7.

Chancerel, A. 1986 Le système des nappes alluviales de la Seine en aval de Rouen. *Travaux Groupe Seine 2*, *Bull. Cent. Géomorph.* **31**, 73–82.

Chatwin, C. P. 1927 Fossils from the iron sands of Netley Heath (Surrey). *Mem. Geol. Surv. Summ. Prog. 1926*, pp. 154–157.

Clarke, M. R. & Auton, C. A. 1982 The Pleistocene depositional history of the Norfolk – Suffolk borderland. *Rep. Inst. geol. Sci.*, part 1, pp. 23–29.

Clet, M. 1982*a* The sections in the Forêt de la Londe: palynology. In *The Quaternary of Normandy. Field handbook* (ed. J.-P. Lautridou), pp. 47–49. Cambridge: Quaternary Research Association.

Clet, M. 1982*b* Bosq d'Aubigny Formation. In *The Quaternary of Normandy. Field handbook* (ed. J.-P. Lautridou), pp. 67–69. Cambridge: Quaternary Research Association.

Commont, V. 1911 La chronologie et la stratigraphie des dépôts quaternaires dans la vallée de la Somme. *Annls Soc. géol. Belg.* **39**, 156–180.
Crommelin, R. D. 1953 Over de stratigraphie en herkomst van de preglaciale afzettingen in Midden-Nederland. *Geologie Mijnb.* **15**, 305–321.
Crommelin, R. D. 1954 Über den Einfluss der nord- und mitteldeutschen Flüsse auf das ältere Pleistozän der Niederlande. *Mitt. geol. St Inst. Hamb.* **23**, 86–97.
Destombes, J.-P., Shephard-Thorn, E. R. & Redding, J. H. 1975 A buried valley system in the Strait of Dover. *Phil. Trans. R. Soc. Lond.* B **279**, 243–256.
Devoy, R. J. N. 1979 Flandrian sea level changes and vegetational history of the lower Thames Estuary. *Phil. Trans. R. Soc. Lond.* B **285**, 355–410.
De Gans, W, 1981 *The Drentsche Aa valley system.* Amsterdam: Vrije Universiteit te Amsterdam and Rodopi.
De Ridder, N. A. & Zagwijn, W. H. 1962 A mixed Rhine–Meuse deposit of Holsteinian age from the south-eastern part of the Netherlands. *Geologie Mijnb.* **41**, 125–130.
De Jong, J. 1965 Quaternary sedimentation in the Netherlands. *Geol. Soc. Am. Spec. Pap.* **84**, 95–123.
Dines, H. G. & Chatwin, C. P. 1930 Pliocene sandstone from Rothamstead (Hertfordshire). *Mem. Geol. Surv. Summ. Prog. 1929.*, pp. 1–7.
Dingwall, R. G. 1975 Sub-bottom infilled channels in an area of the eastern English Channel. *Phil. Trans. R. Soc. Lond.* A **279**, 233–241.
D'Olier, B. 1975 Some aspects of the Late Pleistocene–Holocene drainage of the River Thames in the eastern part of the London Basin. *Phil. Trans. R. Soc. Lond.* A **279**, 269–277.
Doppert, J. W. C., Ruegg, G. H. J., van Staalduinen, C. J., Zagwijn, W. H. & Zandstra, J. G. 1975 Formaties van het Kwartair en Boven-Tertiair in Nederland. In *Toelichtingen bij geologische overzichtskaarten van Nederland* (ed. W. H. Zagwijn & C. J. van Staalduinen), pp. 11–56. Haarlem: Rijks Geol. Dienst.
Duphorn, K., Grube, F., Meyer, K.-D., Streif, H. & Vinken, R. 1973 Area of the Scandinavian glaciation. 1. Pleistocene and Holocene. In *State of research on the Quaternary of the Federal Republic of Germany* (ed. E. Schönhals & R. Huckriede) (*Eiszeitalter Gegenw.* **23–24**), pp. 222–250.
Ehlers, J. 1978 Die quartäre Morphogenese der Harburger Berge und ihrer Umgebung. *Mitt. geogr. Ges. Hamb.* **68**, 1–181.
Ehlers, J. 1983 The glacial history of north-west Germany. In *Glacial deposits in north-west Europe* (ed. J. Ehlers), pp. 229–238. Rotterdam: Balkema.
Ehlers, J. 1987 Die Entstehung des Kaolinsandes auf Sylt. In *Fossilien von Sylt* (ed. U. von Hacht), Vol. 2, pp. 249–267. Hamburg: Von Hacht.
Eissmann, L. 1975 Das Quartär der Leipziger Tieflandsbucht und angrenzender Gebiete um Saale und Elbe. *Schriftenr. geol. Wiss.* **2**, 1–263.
Erd, K. 1970 Pollenanalytical classification of the Middle Pleistocene of the German Democratic Republic. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **8**, 129–145.
Figge, K. 1980 Das Elbe-Urstromtal im Bereich der Deutschen Bucht (Nordsee). *Eiszeitalter Gegenw.* **30**, 203–211.
Figge, K. 1983 Morainic deposits in the German Bight area of the North Sea. In *Glacial deposits in north-west Europe* (ed. J. Ehlers), pp. 299–304. Rotterdam: Balkema.
Friis, H. 1974 Weathered heavy-mineral associations from the young-Tertiary deposits of Jutland, Denmark. *Sediment. Geol.* **12**, 199–213.
Funnell, B. M. 1972 The history of the North Sea. *Bull. geol. Soc. Norfolk* **22**, 2–10.
Funnell, B. M., Norton, P. E. P. & West, R. G. 1979 The Crag at Bramerton, near Norwich, Norfolk. *Phil. Trans. R. Soc. Lond.* B **287**, 489–534.
Genieser, K. 1955 Ehemalige Elbeläufe in der Lausitz. *Geologie* **4**, 223–279.
Genieser, K. 1957 Ehemalige Elbeläufe im Raum zwischen Dresden, Görlitz und Berlin. *Hallesches Jb. mitteldt. Erdgesch.* **2**, 262–266.
Genieser, K. 1962 Neue Daten zur Flussgeschichte der Elbe. *Eiszeitalter Gegenw.* **13**, 141–156.
Gibbard, P. L. 1977 Pleistocene history of the Vale of St. Albans. *Phil. Trans. R. Soc. Lond.* B **280**, 445–483.
Gibbard, P. L. 1979 Middle Pleistocene drainage in the Thames Valley. *Geol. Mag.* **116**, 35–44.
Gibbard, P. L. 1980 The origin of stratified Catfish Creek Till by basal melting. *Boreas* **9**, 71–85.
Gibbard, P. L. 1985 *Pleistocene history of the Middle Thames Valley.* Cambridge University Press.
Gibbard, P. L., Wintle, A. G. & Catt., J. A. 1987 Age and origin of clayey silt 'brickearth' in West London, England. *J. Quat. Sci.* **2**, 3–9.
Goudie, A. S. & Hart, M. G. 1975 Pleistocene events and forms in the Oxford region. In *Oxford and its region* (ed. C. G. Smith & D. I. Scargill), pp. 3–13. Oxford University Press.
Green, C. P., Hey, R. W. & McGregor, D. F. M. 1980 Volcanic pebbles in Pleistocene gravels of the Thames in Buckinghamshire and Hertfordshire. *Geol. Mag.* **117**, 59–64.
Grimmel, E. 1973 Überlegungen zur Morphogenese des Norddeutschen Flachlandes, dargesteldt am Beispiel des unteren Elbtales. *Eiszeitalter Gegenw.* **23–24**, 76–88.
Gripp, K. 1964 *Erdgeschichte von Schleswig-Holstein.* Neumünster: K. Wachholtz Verlag.

Gullentops, F. 1974 The southern North Sea during the Quaternary. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale*. Liège: Cent. Société Géologique de Belgique.

Hamilton, D. & Smith, A. J. 1972 The origin and sedimentary history of the Hurd Deep, English Channel, with additional notes on other deeps in the Western English Channel. *Mem. Bur. Rech. Géol. Miner.* **79**, 59–78.

Hey, R. W. 1976 Provenance of far-travelled pebbles in the pre-Anglian Pleistocene of East Anglia. *Proc. Geol. Ass.* **87**, 69–82.

Hey, R. W. 1980 Equivalents of the Westland Green Gravels in Essex and East Anglia. *Proc. Geol. Ass.* **91**, 279–290.

Hey, R. W. 1986 A re-examination of the Northern Drift of Oxfordshire. *Proc. Geol. Ass.* **97**, 291–301.

Hey, R. W. & Brenchley, P. J. 1977 Volcanic pebbles from Pleistocene gravels in Norfolk and Essex. *Geol. Mag.* **114**, 219–225.

Hinsch, W. 1974 Das Tertiär im Untergrund von Schleswig-Holstein. *Geol. Jb.* A **24**, 34.

Hinsch, W. & Ortlam, D. 1974 Stand und Probleme der Gliederung des Tertiärs in Nordwestdeutschland. *Geol. Jb.* A **16**, 3–25.

Hollin, J. T. 1977 Thames interglacial sites, Ipswichian sea levels and Antarctic ice surges. *Boreas* **6**, 33–52.

Huault, M. F. & Lefèbvre, D. 1974 Le Postglaciaire de la Basse-Seine: sediments et chronostratigraphie. *Bull. Ass. Fr. Étude Quatern.* **40–41**, 253–256.

Jelgersma, S., Oele, E. & Wiggers, A. J. 1979 Depositional history and coastal development in the Netherlands and the adjacent North Sea since the Eemian. In *The Quaternary history of the North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) *Acta univ. ups. symp. univ. ups. a. quingent. celebr.* **2**, pp. 115–142.

Jenkins, D. G., Whittaker, J. E. & Carlton, R. 1986 On the age and correlation of the St Erth Beds, SW England, based on planktonic foraminifera. *J. Micropalaeont.* **5**, 93–105.

John, D. T. 1980 The soils and superficial deposits on the North Downs of Surrey. In *The shaping of southern England* (ed. D. K. C. Jones), pp. 101–130. London: Academic Press.

Kaiser, K. 1960 Klimazeugen des periglazialen Dauerfrostbodens in Mittel- und Westeuropa. *Eiszeitalter Gegenw.* **11**, 121–141.

Kazi, A. & Knill, J. L. 1969 The sedimentation and geotechnical properties of the Cromer Till between Happisburgh and Cromer, Norfolk. *Q. Jl Engng Geol.* **2**, 63–86.

Kellaway, G. A., Redding, J. H., Shephard-Thorn, E. R. & Destombes, J.-P. 1975 The Quaternary history of the English Channel. *Phil. Trans. R. Soc. Lond.* A **279**, 189–218.

Kemp, R. A. 1985 The Valley Farm Soil in southern East Anglia. In *Soils and Quaternary landscape evolution* (ed. J. Boardman), pp. 179–196. Chichester: John Wiley.

Knoth, W. 1964 Zur Kenntnis der pleistozänen Mittelterrassen der Saale und Mulde nördlich von Halle. *Geologie* **13**, 598–616.

Koche, E. 1954 Vom Untergrund Hamburgs. *Mitt. geol. St Inst. Hamb.* **23**, 10–17.

Kolfschoten, T. van & Roebroeks, W. 1985 The Maastricht–Belvédère project; an intermediate synthesis. *Meded. Rijks. geol. Dienst.* **39**, 119–121.

Kuntz, G. & Lautridou, J.-P. 1974 Contribution a l'étude du Pliocène et du passage Pliocène–Quaternaire dans les dépôts de la Forêt de la Londe près de Rouen. Corrélations possibles avec divers gisements de Haute Normandie. *Bull. Ass. fr. Étude Quat.* **3–4**, 117–128.

Kuntz, G., Lautridou, J.-P., Cavelier, C. & Clet, M. 1979 Le Plio-Quaternaire de Haute-Normandie. *Bull. Inf. géol. Bassin Paris* **16**, 93–126.

Larsonneur, C., Vaslet, D. & Auffret, J. P. 1979 Carte des sediments superficiel de la Manche. *Bur. Rech. Géol. miner.* 1:50000 map.

Lautridou, J.-P. (ed.) 1982 *The Quaternary of Normandy. Field handbook.* Cambridge: Quaternary Research Association.

Lautridou, J.-P., Monnier, J. L., Morzadec, M. T., Somme, J. & Tuffreau, A. 1986 The Pleistocene of northern France. *Quat. Sci. Rev.* **5**, 387–393.

Lautridou, J.-P., Lefèbvre, D., Lécolle, F., Carpentier, G., Descombes, J. C., Gaquerel, C. & Huault, M. F. 1984 Les terrasses de la Seine dans le méandre d'Elbeuf, corrélations avec celles dans la region Mantes. *Bull. Ass. Fr. Étude Quatern.* **1–2–3**, 27–32.

Lécolle, F. 1984 Phases érosives et cycles sédimentaires: les alluvions de la Seine au sud du Vexin. *Bull. Ass. fr. Étude Quat.* **1–2–3**, 33–36.

Lefèbvre, D. 1974 Le cailloutes würmien de fond de vallée, Basse Seine. *Bull. Ass. fr. Étude Quat.* **40–41**, 251–252.

Lefèbvre, D., Carpentier, E. & Evrard, H. 1986 Les terrasses de la Seine de Pont-de-l'Arche à Elbeuf. Travaux Groupe Seine 2. *Bull. Centre Géomorph.* **31**, 41–71.

Lefèbvre, D., Huault, M. F., Guyader, J., Giresse, P., Hommeril, P. & Larsonneur, C. 1974 Le prisme alluvial de l'estuaire de la Seine: synthèse sédimentologique, stratigraphique et paléogéographique. *Bull. Inf. Géol. Bassin Paris* **39**, 27–36.

Lüttig, G. 1960 Neue Ergebnisse quartärgeologischer Forschung im Raume Alfeld-Hameln-Elze. *Geol. Jb.* **77**, 337–390.

Lüttig, G. 1974 Geological history of the River Weser. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord meridionale* (ed. P. Macar), pp. 21–34. Liège: Cent. Société Géologique de Belgique.

Lüttig, G. & Maarleveld, G. C. 1961 Nordische Geschiebe in Ablagerungen prä-Holstein in den Niederlanden (Komplex von Hattem). *Geologie Mijnb.* **40**, 163–174.

Lüttig, G. and Meyer, K.-D. 1974 Geological history of the River Elbe, mainly of its lower course. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale* (ed. P. Macar), pp. 1–19. Liège: Cent. Société Géologique de Belgique.

Maarleveld, G. C. 1954 Über fluviatile Kiese in Nordwestdeutschland. *Eiszeitalter Gegenw.* **4–5**, 10–17.

Maarleveld, G. C. 1956 Grindhoudende Midden-Pleistoceen sedimenten. *Meded. geol. Sticht.* C**4** (6). (105 pages.)

Macaire, J. J. 1984 Les vallées et formations alluviales Plio-Quaternaires dans le sud et le sud-ouest du Bassin de Paris: genèse et signification dynamique. *Bull. Ass. fr. Étude Quat.* **1–2–3**, 37–40.

Menke, B. 1975 Vegetationsgeschichte und Florenstratigraphie Nordwestdeutschlands im Pliozän und Frühquartär. Mit einem Beitrag zur Biostratigraphie des Weichselfrühglazials. *Geol. Jb.* A **26**, 3–151.

Meyer, K.-D. 1983 Zur Anlage der Urstromtäler in Niedersachsen. *Z. Geomorph.* N. S. **27**, 147–160.

Meyer, K.-D., Schmid, F. & Wolburg, J. 1977 Erläuterungen zu Blatt Salzbergen Nr 3610. Geol. Karte Niedersachsen. (111 pages). Hannover: *Niedersächsisches Landesamt für Bodenforschung.*

Moffat, A. J. & Catt, J. A. 1986 A re-examination of the evidence for a Plio-Pleistocene marine transgression on the Chiltern Hills. III. Deposits. *Earth Surf. Process. Landf.* **11**, 233–247.

Oele, E. 1969 The Quaternary geology of the Dutch part of the North Sea, north of the Frisian Isles. *Geologie Mijnb.* **48**, 467–480.

Oele, E. 1971 The Quaternary geology of the southern area of the Dutch part of the North Sea. *Geologie Mijnb.* **50**, 461–474.

Oele, E. & Schüttenhelm, R. T. E. 1979 Development of the North Sea after the Saalian glaciation. In *The Quaternary history of the North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) (*Acta univ. ups. symp. univ. ups. a. quingent. celebr.* **2**), pp. 191–215.

Paepe, R. & Sommé, J. 1975 Marine Pleistocene transgressions along the Flemish coast (Belgium and France). *North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) (*Acta univ. ups. symp. univ. ups. a. quingent. celebr.* **2**), pp. 143–146.

Paepe, R. Sommé, J. 1975 Marine Pleistocene transgressions along the Flemish coast (Belgium and France). *Quaternary glaciations in the Northern Hemisphere. Int. Geol. Correl. Prog. project* **24**, no. 2, pp. 108–116.

Paepe, R. & Vanhoorne, R. 1970 Stratigraphical position of periglacial phenomena in the Campine Clay of Belgium, based on palaeobotanical analysis and palaeomagnetic dating. *Bull. Soc. Belge geol.* **79**, 201–211.

Paulissen, E. 1973 De morfologie en de Kwartairstratigrafie van de Maasvallei in Belgisch Limburg. *Verh. K. vlaam. Akad. Wet.* **35** (127). (266 pages.)

Paulissen, E., Vandenberghe, J. & Gullentops, F. 1985 The Feldbiss Fault in the Maas valley bottom (Limburg, Belgium). *Geologie Mijnb.* **64**, 79–87.

Pissart, A. 1974 La Meuse en France et en Belgique. Formation du bassin hydrographique. Les Terraces et leurs enseignements. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale* (ed. P. Macar), pp. 105–132. Liège: Cent. Société Géologique de Belgique.

Pointon, W. K. 1978 The Pleistocene succession at Corton, Suffolk. *Bull geol. Soc. Norfolk* **30**, 55–76.

Porcher, M. 1977 Lithostratigraphie des alluvions fluviatiles holocènes de la vallée de la Seine. *Bull. Soc. géol. Norm.* **64**, 181–201.

Präger, F. 1970 Die Bedeutung stratigraphischer Untersuchungen im Quartär am Beispiel von angewandten Arbeiten im Gebiet von Nieschütz bei Meissen. *Wiss. Z. Univ. Halle*, **19** (1), 49–55.

Quitzow, H. W. 1953 Altersbeziehungen und Flözzusammenhänge in der jüngeren Braunkohlenformation nördlich der Mittelgebirge. *Geol. Jb.* **68**, 27–132.

Quitzow, H. W. 1974 Das Rheintal und seine Entstehung. Bestandsaufnahme und Versuch einer Synthese. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord Meridionale* (ed. P. Macar), pp. 53–104. Liège: Cent. Sociéte Géologique de Belgique.

Rasmussen, L. B. 1961 Mittel- und Ober-Miocän von Dänemark. *Meyniana* **10**, 59–62.

Roep, T. B., Holst, H., Vissers, R. L. M., Pagnier, H. & Postma, D. 1975 Deposits of southward flowing Pleistocene rivers in the Channel region, near Wissant, NW France. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **17**, 289–308.

Rose, J. & Allen, P. 1977 Middle Pleistocene stratigraphy in southeastern Suffolk. *J. geol. Soc. Lond.* **133**, 83–102.

Rose, J., Allen, P. & Hey, R. W. 1976 Middle Pleistocene stratigraphy in southern East Anglia. *Nature, Lond.* **263**, 492–494.

Rose, J., Boardman, J., Kemp, R. A. & Whiteman, C. A. 1985 Palaeosols and the interpretation of the British Quaternary stratigraphy. In *Geomorphology and Soils* (ed. K. S. Richards, R. R. Arnett & S. Ellis), pp. 348–375. London: George Allen & Unwin.

Rose, J., Turner, C., Coope, G. R. & Bryan, M. D. 1980 Channel changes in a lowland river catchment over the last 13000 years. In *Timescales in geomorphology* (ed. R. A. Cullingford, D. A. Davidson & J. Lewin), pp. 159–175. Chichester: John Wiley.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Ruegg, G. H. J. & Zandstra, J. G. 1977 Pliozäne und pleistozäne gestauchte Ablagerungen bei Emmerschans (Drenthe, Niederlande). *Meded. Rijks geol. Dienst* **34**, 7–9.

Ruske, R. 1964 Das Pleistozän zwischen Halle (Saale), Bernburg und Dessau. *Geologie* **13**, 570–597.

Schulz, W. 1962 Gliederung des Pleistozäns in der Umgebung von Halle (Saale). *Geologie Beih.* **36**, 1–69.

Seddon, M. & Holyoak, D. T. 1985 Evidence for sustained regional permafrost during deposition of fossiliferous Late Pleistocene sediments at Stanton Harcourt (Oxfordshire, England). *Proc. Geol. Ass.* **96**, 53–71.

Semmel, A. 1973 Area between the Scandinavian and the Alpine Glaciation. In *State of research in the Quaternary of the Federal Republic of Germany* (ed. E. Schönhals & R. Huckriede) (*Eiszeitalter Gegenw.* **23/24**), pp. 293–305.

Shotton, F. W., Goudie, A. S., Briggs, D. J. & Osmaston, H. A. 1980 Cromerian interglacial deposits at Sugworth near Oxford, England, and their relation to the Plateau Drift of the Cotswolds and the terrace sequence of the Upper and Middle Thames. *Phil. Trans. R. Soc. Lond.* B **289**, 55–86.

Šibrava, V. 1972 Zur Stellung der Tschechoslowakei im Korrelierungssystem des Pleistozäns in Europa. *Sb. geol. věd.* (A) **8**, 1–288.

Šibrava, V. 1986 Scandinavian glaciations in the Bohemian Massif and Carpathian Foredeep and their relationship to the extra-glacial areas. *Quat. Sci. Rev.* **5**, 381–386.

Smith, A. J. 1985 A catastrophic origin for the palaeovalley system of the eastern English Channel. *Mar. Geol.* **64**, 65–75.

Smith, A. J. & Curry, D. 1975 The structure and geological evolution of the English Channel *Phil. Trans. R. Soc. Lond.* A **279**, 3–20.

Sommé, J. 1979 Quaternary coastlines in northern France. In *The Quaternary history of the North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) (*Acta univ. ups. symp. univ. ups. a. Quingent. celebr.* **2**), pp. 147–158.

Sommé, J., Fagnart, J. P., Leger, M., Munaut, A. V., Puissegur, J. J. & Tuffreau, A. 1984 Terrasse fluviatiles du Pléistocène moyen en France septentrionale: signification dynamique et climatique. *Bull. Ass. fr. Étude Quat.* **1–2–3**, 52–58.

Stamp, L. D. 1927 The Thames drainage system and the age of the Strait of Dover. *Geogrl. J.* **70**, 386–390.

Tavernier, R. & De Moor, G. 1974 L'évolution du bassin de l'Escaut. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale* (ed. P. Macar), pp. 273–280. Liège: Cent. Société Géologique de Belgique.

Ter Wee, M. W. 1962 The Saalian glaciation in the Netherlands. *Meded. Geol. Sticht.* (n.s) **15**, 57–77.

Ter Wee, M. W. 1983*a* The Elsterian glaciation in the Netherlands. In *Glacial deposits in north-west Europe* (ed. J. Ehlers), pp. 413–415. Rotterdam: Balkema.

Ter Wee, M. W. 1983*b* The Saalian glaciation in the northern Netherlands. In *Glacial deposits in north-west Europe* (ed. J. Ehlers), pp. 405–412. Rotterdam: Balkema.

Thome, K. N. 1980 Der Vorstoss des nordeuropäischen Inlandeises in das Münsterland in Elster- und Saale-Eiszeit. *Westf. geogr. Stud.* **36**, 21–40.

Tuffreau, A., Munaut, A. V., Puissegur, J. J. & Sommé, J. 1981 Les basse terrasse dans les vallées du Nord de la France et de la Picardie: stratigraphie et paléolithique. *Bull. Soc. prehist. Fr.* **78**, 291–305.

Tuffreau, A., Munaut, A. V., Puissegur, J. J. & Sommé, J. 1982 Stratigraphie et environnement des industries Acheuléennes de la moyenne terrasse du Bassin de la Somme (Région d'Amiens). *Bull. Ass. fr. Étude Quat.* **2–3**, 73–82.

Vandenberghe, J. and De Smedt, P. 1979 Palaeomorphology in the eastern Scheldt Basin (Central Belgium) – the Dijle-Demer-Grote Nete confluence area. *Catena* **6**, 73–105.

Vandenberghe, J., Krook, L. and van der Valk, L. 1986 On the provenance of the Early Pleistocene fluvial system in the southern Netherlands. *Geologie Mijnb.* **65**, 3–12.

Vandenberghe, J., Mücher, H. J., Roebroeks, W. & Gemke, D. 1985 Lithostratigraphy and palaeoenvironment of the Pleistocene deposits at Maastricht-Belvédère, Southern Limburg, the Netherlands. *Meded. Rijks. Geol. Dienst.* N.S. **39**, 7–18.

Vandenberghe, J., Paris, P., Kasse, C., Gouman, M. & Beyens, L. 1984 Paleomorphological and botanical evolution of small lowland valleys – a case study of the Mark valley in northern Belgium. *Catena* **11**, 229–238.

Van den Broek, J. M. M. & Maarleveld, G. C. 1963 The late Pleistocene terrace deposits of the Meuse. *Meded. Geol. Sticht.* N.S. **16**, 13–24.

Van der Meene, E. A. & Zagwijn, W. H. 1978 Die Rheinläufe im deutsch–niederländischen Grenzgebiet seit der Saale-Kaltzeit. Überblick neuer geologischer und pollenanalytischer Untersuchungen. *Fortschr. Geol. Rheinld Westf.* **28**, 345–359.

Vanhoorne, R. 1962 Het interglaciale veen te Loo (België). *Natuurw. Tijdschr.* **44**, 58–64.

Ventris, P. 1985 Pleistocene environmental history of the Nar Valley, Norfolk. Ph.D. thesis, University of Cambridge.

West, R. G. 1972 Relative land-sea level changes in southeastern England during the Pleistocene. *Phil. Trans. R. Soc. Lond.* A **272**, 87–98.

West, R. G. 1980*a* *Pre-glacial Pleistocene of the Norfolk and Suffolk coasts.* Cambridge University Press.

West, R. G. 1980*b* Pleistocene forest history in East Anglia. *New Phytol.* **85**, 571–622.

West, R. G. & Wilson, D. G. 1968 Plant remains from the Corton Beds, Lowestoft, Suffolk. *Geol. Mag.* **105**, 116–123.

Weyl, R., Rein, U. & Teichmüller, M. 1955 Das Alter des Sylter Kaolinsandes. *Eiszeitalter Gegenw.* **6**, 5–15.
Whiteman, C. A. 1983 Great Waltham. In *Diversion of the Thames. Field guide* (ed. J. Rose), pp. 163–169. Cambridge: Quaternary Research Association.
Woldstedt, P. 1956 Die Geschichte des Flussnetzes in Norddeutschland und angrenzenden Gebieten. *Eiszeitalter Gegenw.* **7**, 5–12.
Woldstedt, P. 1967 The Quaternary of Germany. In *The Quaternary* (ed. K. Rankama), vol. 2, pp. 000–000. New York: Interscience Publications.
Wooldridge, S. W. & Linton, D. L. 1955 *Structure, surface and drainage in South-east England*. London: G. Philip.
Wymer, J. J. 1968 *Lower Palaeolithic archaeology in Britain*. London: John Baker.
Zagwijn, W. H. 1960 Aspects of the Pliocene and Early Pleistocene vegetation in the Netherlands. *Meded. geol. Sticht.*, ser. C-III-I, **5**, 78.
Zagwijn, W. H. 1963 Pleistocene stratigraphy in the Netherlands, based on changes in vegetation and climate. *Verh. K. ned. geol.-mijnb. Genoot.*, geol. ser., **20**, 173–196.
Zagwijn, W. H. 1974 Palaeogeographic evolution of the Netherlands during the Quaternary. *Geologie Mijnb.* **53**, 369–385.
Zagwijn, W. H. 1979 Early and Middle Pleistocene coastlines in the southern North Sea Basin. In *The Quaternary history of the North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) (*Acta univ. ups. symp. univ. ups. a. quingent. celebr.* **2**), pp. 31–42.
Zagwijn, W. H. 1985 An outline of the Quaternary stratigraphy of the Netherlands. *Geologie Mijnb.* **64**, 17–24.
Zagwijn, W. H. & Doppert, J. W. C. 1978 Upper Cenozoic of the southern North Sea Basin: palaeoclimatic and palaeogeographic evolution. In *Keynotes of the MEGS-II* (*Amsterdam 1978*) (ed. A. J. van Loon) (*Geologie Mijnb.* **57**), pp. 577–588.
Zalasiewicz, J. A., Mathers, S., Hughes, M. J., Gibbard, P. L., Peglar, S. M., Harland, R., Nicholson, R. A., Boulton, G. S., Cambridge, P. & Wealthall, G. P. 1988 Stratigraphy and palaeoenvironments of the Red and Norwich Crags between Aldeburgh and Sizewell, Suffolk, England. *Phil. Trans. R. Soc. Lond.* B. (In the press.)
Zandstra, J. G. 1971 Geologisch onderzoek in de stuwwal van de oostelijke Veluwe bij Hattem en Wapenveld. *Meded. Rijks. geol. Dienst.* **22**, 215–258.
Zandstra, J. G. 1983 Fine gravel, heavy mineral and grain-size analyses of Pleistocene, mainly glaciogenic deposits in the Netherlands. In *Glacial deposits in north-west Europe* (ed. J. Ehlers), pp. 361–377. Rotterdam: Balkema.
Ziegler, P. A. 1978 North-Western Europe: tectonics and basin development. In *Keynotes of the MEGS-II* (*Amsterdam 1978*) (ed. A. J. van Loon) (*Geologie Mijnb.* **57**), pp. 589–626.
Ziegler, P. A. & Louwerens, C. J. 1979 Tectonics of the North Sea. In *The Quaternary history of the North Sea* (ed. E. Oele, R. T. E. Schüttenhelm & A. J. Wiggers) (*Acta univ. ups. symp. univ. ups. a. quingent. celebr.* **2**), pp. 7–22.
Zonneveld, J. I. S. 1958 Litho-stratigraphische eenheden in het Nederlandse Pleistoceen. *Meded. geol. Sticht.* N.S. **12**, 31–64.
Zonneveld, J. I. S. 1974 The terraces of the Maas and the Rhine downstream of Maastricht. In *L'évolution Quaternaire des bassins fluviaux de la Mer du Nord méridionale* (ed. P. Macar), pp. 133–158. Liège: Cent. Société Géologique de Belgique.

Discussion

J. Rose (*Department of Geography, Birkbeck college, University of London, U.K.*) I should like to make two points concerning the topic considered by Dr Gibbard, and to ask him one question.

The evidence that Professor Smith presented for the catastrophic proglacial lake meltwater erosion which resulted in the formation of the Straits of Dover is based on the palaeochannel pattern and associated sea-bed sediments in the eastern part of the English Channel and not just on the geomorphology (Smith 1984).

The period of Quaternary time when the northwest European rivers had their greatest influence on the landscape is represented by the Sterksel Formation of The Netherlands, the Yarmouth Roads Formation of the southern North Sea and the Kesgrave Formation of East Anglia (Bowen *et al.* 1986). These bodies of clastic sediments represent the largest accumulations of Quaternary fluvial deposits within their respective areas. The vast quantities of sediment are the product of effective erosion in the headwater regions of the Alps in the case of the Rhine and of Wales in the case of the Thames, probably enhanced by the early episodes of glaciation in these upland regions. In terms of the Quaternary history of northwestern Europe these deposits were formed during the later part of the early Pleistocene (Bowen *et al.* 1986, table 2).

On all the palaeogeographical maps shown in the lecture the 'Solent River' in the region of Dorset, southeast Hampshire and the Isle of Wight is anomalous, either in terms of its relation with the coastline shown on the early reconstruction, or in terms of the river networks shown on the later reconstructions. Can Dr Gibbard explain this anomaly?

References

Bowen, D. Q., Rose, J., McCabe, A. M. & Sutherland, D. G. 1986 Correlation of Quaternary glaciations in England, Ireland, Scotland and Wales. *Quat. Sci. Rev.* **5**, 299–340.

Smith, A. G. 1984 Structural evolution of the English Channel region. *Annls Soc. geol. Nord* **103**, 253–264.

P. L. Gibbard. Mr Rose is correct in asserting that Professor Smith's interpretation of catastrophic drainage of the North Sea glacial lake was based partly on the palaeochannel pattern that superficially resembles that formed during the drainage of Lake Missoula. The palaeochannel system associated with the latter was formed in a very short, single event. This is, however, not the case for the Channel floor palaeochannels, which Smith himself concedes have a multiphase infilling (Hamilton & Smith 1972; A. G. Smith, personal communication). According to French workers the palaeochannels may in part pre-date drainage of the lake (cf. Alduc *et al.* 1979). It is possible that the palaeochannels may have been formed by superimposition of river courses following repeated marine regression.

I agree that the period before the arrival of widespread lowland continental glaciation in the Middle Pleistocene was dominated in the region by fluvial activity. The Kedichem and Sterksel Formations in the Rhine–Meuse system, and the Thames' Kesgrave Formation, mark the input of large quantities of detrital material. However, during the Early and early Middle Pleistocene, considerable quantities of sediment were deposited as the Harderwijk and Enschede Formations by the westward-flowing North German rivers. These combined sedimentary accumulations certainly indicate that contemporary periglacial erosion was very effective.

The evolution of the 'Solent River' system is fascinating and its course, at first sight, appears to be anomalous. This anomaly seems to arise from structural control. The course lies north of the Isle of Wight – Purbeck anticline, which trends west–east parallel to the present coast to beyond the east coast of the Isle of Wight. It then turns towards the southeast to cross the Channel towards the Normandy coast. I suspect that the Solent River was forced to flow eastwards by this structure, possibly as a consequence of regional uplift in the west, which may have been active in the Pleistocene (see text). The river presumably continued eastwards until it could break through into the Somme–Seine system.

D. R. Bridgland *Nature Conservancy Council, Peterborough, U.K.*). A comment was offered on behalf of myself and B. D'Olier, of the City of London Polytechnic, on the Straits of Dover.

The results of offshore reflective seismic profiling by B. D'Olier show that the Dover Strait incorporates the southern end of a submerged valley, some 50 km long, cut into the Chalk to a depth of *ca.* 55 m below sea level. This valley has no appreciable slope, but its highest point lies approximately 15 km from its southern end. At the same time, the submerged extension of the Thames valley can be traced offshore from the Essex coast towards the central North Sea, reaching 65 m below sea level by latitude 52 °N. Projection of the floor of the late Devensian

'Buried Channel' of the Thames offshore, by the route suggested by Dr Gibbard for the post-Elsterian Thames–Rhine, indicates that a depth of at least -70 m OD would be required for the river to have passed through the Dover Strait at this time. Given these facts it would seem that, unless differential subsidence or uplift is invoked, the Thames had returned to a North Sea exit during the Devensian.

P. L. Gibbard. Erosional channels are known from the Channel and Dover Straits area and are referred to by authors quoted in the text. In the past these features have been variously interpreted to result from glacial erosion or more probably from considerable tidal scour (see discussion in Zagwijn (1979)). The presence of an extensive, sub-horizontal channel lacking a sedimentary fill in this region is not unexpected and may itself also be of tidal origin. Alignment of the Rhine, Thames and other tributaries through the Dover Straits during periods of low sea level from the late Middle Pleistocene to the end of the last cold stage is based on the distribution of sediments mapped by the Dutch and British Geological Surveys. I am assured by colleagues in both organizations that there is no evidence that the Thames adopted a northerly course in the southern North Sea after the Anglian–Elsterian. Similarly, there is no evidence of a northerly directed Rhine–Meuse from sediments preserved after the Saalian Drenthe Substage. Indeed, the distribution of the Rhine–Meuse Kreftenheye Formation sediments indicates that these rivers were aligned towards the south during the Late Pleistocene.

The answer to the altitudinal problem may lie in the fact that the gradient of the Late Devensian Thames would be expected to have been reduced offshore at its confluence with the Rhine–Meuse. Such a shallow gradient would have been required for the river to pass through the Dover Straits. If the river did pass through in the Late Devensian this could be confirmed by the distribution of its deposits. It would imply that the channel to which Dr Bridgland refers in its present form postdates the Late Devensian.

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