Fluvial sequences of the Maas: a 10 Ma record of neotectonics and climatic change at various time-scales

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Cover: Parasequences of the Maas river: Quarry Mol (Mechelen aan de Maas, Belgium); Lacquer peel quarry Panheel (the Netherlands). Design: Diana Moeliker.

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STELLINGEN

1. Maasafzettingen verdienen door hun karakteristieke lithologische, mineralogische en geochemische eigenschappen een eigen plaats in de stratigrafie van de Nederlandse ondergrond.

2. De hydrogeologische en geomorfologische situatie in de Ardennen bepaalt mede de gevoeligheid van het Maas systeem voor het registreren van klimaatsveranderingen.

3. De geringe aandacht die in sedimentologische overzicht werken wordt besteed aan de rol van het klimaat bij het interpreteren van sedimentaire structuren is voor een deel te verklaren door het ontbreken van zowel mogelijkheden tot datering als het ontbreken van kaders voor klimaat referentie. (A. Miall, 1996: Geology of Fluvial Deposits)

4. Het onafhankelijk van elkaar aanbrengen van een numerieke rangorde in verschillende, hiërarchisch-geordende systemen komt de overzichtelijkheid niet ten goede als naderhand blijkt dat er een koppeling tussen deze systemen blijkt te bestaan.

5. Nederlandse rivieren zijn slechts plaatselijk te beschouwen als meanderende systemen. Een omschrijving als eilandrivier met een recht tot zwak sinoïde verloop is passender voor de oorspronkelijke patronen.

6. De geodetisch bepaalde gemiddelde verticale snelheden van de bodembeweging in Zuid Limburg zijn een factor 10 hoger dan de gemiddelde snelheid bepaald over de laatste 3 Ma. Deze snelheid is op haar beurt weer een factor 20 hoger dan de gemiddelde snelheid over de daaraan voorafgaande 10 miljoen jaar. Dit geologisch perspectief suggereert dat wij op het ogenblik leven in een periode met extreme verticale korst bewegingen.

7. Indien bij de structurele analyse van een gebied de morfotectonische informatie wordt geïntegreerd in de gegevens die worden verkregen vanuit de seismiek, wordt het structurele beeld aanmerkelijk gecompliceerder en bovendien meer realistisch.

8. In gebieden met een geringe seismische activiteit, kan een morfotectonische analyse informatie geven over de oriëntatie van het heersende spanningsveld.

9. Voor het uiteenrafelen van de komponenten die de, geodetisch bepaalde, recente bodemdaling van West Nederland veroorzaken, is de kennis van de opheffing van Zuid Limburg zinvol.

10. Het accepteren van een beweeglijke landsgrens in gebieden waar een rivier de grens vormt tussen Nederland en Duitsland zoals voorgesteld door de Technische Commissie die het Tweede Verdrag tussen het Koninkrijk der Nederlanden en de Bondsrepubliek Duitsland inzake grens correcties heeft voorbereid, maakt inzicht in het tektonisch-morfologisch riviergedrag essentieel voor de evaluatie van de lange termijn consequenties van een degelijk besluit. (Elsevier, 14-11-1992)

11. De Nederlandse stuwwallen zijn waarschijnlijk gevormd tijdens de overgangsperiode van glaciale naar interglaciale klimaatscondities.

12. Het gebruik van het onderzoek naar Dinoflagellaten in de Kwartaire en Tertiaire afzettingen, is in het Noordzee bekken te weinig toegepast.

13. Veel Nederlandse natuurgebieden zijn aangewezen op marginale gronden. Het is daarom te verwachten dat het inzetten van grote grazers bij het terreinbeheer, op zogenaamde natuurlijke wijze, van dergelijke gronden, de gezondheidstoestand van deze dieren ook marginaal zal worden.

14. De vaak gehoorde uitspraak "verkeerd verbonden" als tijdens een telefoongesprek niet de gewenste verbinding tot stand is gekomen zegt meer over het beperkte vermogen tot zelfkritiek van de spreker dan over het falen van de communicatie techniek.

15. Het korte-termijn denken van vele politici staat op gespannen voet met hun taak om richting te geven aan lange termijn visies.

16. In de westerse wereld is het moeilijker om kinderen tot evenwichtige persoonlijkheden op te voeden dan het promoten van een TV ster.

17. Niet alle terrassen langs de Maas zijn Maasterrassen.

Stellingen behorende bij het proefschrift van M.W. van den Berg:
Fluvial Sequences of the Maas: A 10 Ma Record of Neotectonics and Climatic Change at various Time-Scales

Wageningen, 9 oktober 1996

Contents

Summary

Around 10 million years ago, the interplay of tectonics, climate and sea level changed markedly in the southern North Sea Basin. One of the results was the start of rapid progradation of the Rhine/Meuse delta. The sediments of this basin-filling complex have been preserved in a number of depocentres in an en-echelon arrangement which form the backbone of the North Sea Basin. These depocentres were located within the west-central part of the European intra-plate rift system. The southernmost part of this rift system, the Roer valley graben (part of the lower Rhine Embayment), was the first depocentre to be filled by sediments supplied by the rising hinterland in the south. The sediment-source area expanded gradually to include the Alpine collision front, which was first tapped by the river Rhine around three million years ago. Therefore, the sediments that filled the Roer valley graben provide information for unravelling the geological history of the north-west European plate.

Using high-resolution, sequence-stratigraphic techniques, this thesis focuses on the stratigraphic, morphologic and tectonic aspects of the upper-delta and fluvial sediments, laid down by the Rhine/Maas (Meuse) fluvial system in the present-day Netherlands. The present study benefited greatly form the large database, compiled during the past decades by the Geological Survey of the Netherlands and the Winand Staring Centre-DLO Institute. Vital data sources were Zagwijn's pollen-based regional palaeo-climatic interpretations as well as climatic data derived from deep-sea cores and ice cores. In addition, correlation of deep-sea geological records to astronomical parameters, proved of great value as it enabled conversion of the relative time scales based on palaeontological data to a linear one.

In general, precise dating of fluvial deposits is difficult. As a consequence, interpretation largely depends on circumstantial evidence, such as the fractal-type hierarchical structure of the climatically controlled fluvial systems.

A total of 57 stacked units (around 5-20 m in thickness) have been identified by studying terraces and fining upward sequences in borehole records from the Roer valley graben. In both types of exposure, the bounding surfaces of these stacked sequences generally reflect long-term, basin-wide episodes of fluvial deposition and erosion. Dating and modelling support the interpretation that the sequences represent a fourth-order cyclicity in the hierarchy of environmental changes that affected fluviatile processes and caused river reactivation (at fifth-order level in Miall's classification). For this entire period of 10 million years, a strong correlation exists between the number of (buried) surfaces reflecting river reactivation and major climate cycles with a duration varying between approximately 400 to 50 thousand years. Measured in time-steps at million years scale (1-2 Ma), the dominant average duration shifts from 200 thousand years for the period from 10-2 Ma to 100 thousand years for the last 2 million years. Such climatically-controlled cyclicity is well-known from deep-sea cores. This correlation shows that climatic change is an important control on cyclicity in fluvial sequences as well. Moreover, it demonstrates that it is possible to meaningfully correlate the oceanic record with the continental fluvial record. Consequently, a continuous series of fifth-order fluvial sequences can be used to develop a high-resolution time frame, which will be of great value in the study of basin dynamics. Fluvial sequences, which reflect the two extremes of climatic cycles, are a welcome addition to palynological records, which are restricted to warm episodes.

Zooming in on the components of fifth-order fluvial sequences has enabled us to demonstrate a fractal-type hierarchy within the coupling between sedimentary units and cli-

mate cycles. Fourth-, fifth- and sixth-order climate cycles equally reflect fifth-, fourth-, and probably third-order sedimentary cycles.

One of the problems involved in direct correlation of climate changes and changes in fluvial dynamics is accounting for the effects of the vegetation cover. Regeneration of a particular vegetation cover takes a few hundred years at the most. This introduces a time-lag factor in the process-response relationship. Obviously, this time-lag factor is particularly relevant in the case of sixth-order (millennial-scale) climate changes.

Preservation of sedimentary sequences depends greatly on regional tectonics. The present study shows that subsidence of the graben started much earlier than uplift of the south flank of the Roer valley rift. Consequently, the two processes must have different control mechanisms. Subsidence is fault-controlled, caused by deep-seated extension, governed by the dynamics of the European plate. Uplift, possibly controlled by underthrusting of the Ardennes/Rhenish Shield, resulted in overall shortening of the graben owing to foreland compression. This interaction between extension and compression is reflected in a right-lateral strike-slip movement along the principal displacement zones of the graben and in the formation of drainage divides perpendicular to the length axis of the graben. The faults are extensional in character and show predominantly normal displacements. The strike-slip dynamics are reflected in the regional morphology and the changing patterns of the palaeo-river systems. During episodes in which horizontal movement along the boundary faults prevails, a river can change course and cross major faults to start flowing to other depocentres in the same rift zone. This highlights the importance of hiatuses in the sedimentary record for the interpretation of basin tectonics.

An extensive, long-term geodetic benchmark data set, compiled over 117 years, corroborates the direction of relative displacement as inferred from geomorphology. The combination of this geodetic data set and the uplift history as recorded by the series of river terraces, is crucial for the analysis of vertical movements observed in other parts of the Netherlands. In addition, it helps to establish links between regions with different subsurface characteristics.

At the resolution scale of the fourth-order climate cycles, both uplift and subsidence show variations in pace: acceleration and deceleration. Three important episodes showed up: (1) the onset of the hinterland uplift occurred approximately around ten millions years ago simultaneously with the, type one, rapid eustatic sea-level fall known as the boundary between the supercycles TB1-TB2 from Haq's cycle chart. (2) The apparently simultaneous onset of Quaternary-type climate dynamics and the transition from relatively slow to rapid vertical crustal dynamics, which occurred approximately three millions years ago. (3) Another, albeit less dramatic indication of a geodynamic effect is the temporary cessation of uplift which occurred between 1.5 and 1.1 million years ago. An event which seems to coincide with episodes of major plate-tectonic physiographic changes.

The apparent synchronism between crustal dynamics and changes in climate system brings up an intriguing topic for future research: would it be possible to link the changing frequency of orbital forcing and its effect on climatic and geodynamic changes? The events mentioned above suggest that such a link may exist. Similarly detailed data on other basins could provide answers. Including fluvial sequence stratigraphy in general stratigraphic research may greatly improve our knowledge of basin dynamics.

1. Introduction

a) Study area

The Roer valley rift system is located in the border area of Germany and the Netherlands (see Fig. 1). The Roer valley rift system is the main structural-physiographic feature in this area. This funnel-shaped tectonic depression, opening towards the north-west, is related to the West Central European rift system (Ziegler, 1994). It is an integral part of the North Sea Basin and its tectonic history is intimately associated with the Alpine foreland of which the Ardennes/Rhenish Shield area forms an important part. The rift is bounded by normal, reverse and transcurrent faults. The main landform-shaping process was brittle deformation of the upper crust in response to intra-plate stresses.

As the Roer valley rift is part of a sedimentary basin, rivers supplying sediment were the second key parameter shaping the current land surface. The surrounding highlands were drained by the rivers Rhine and Maas (Meuse) and also by a few small Belgian rivers, which no longer exist, but which drained the northward-dipping Paleogene and Neogene strata in Belgium. The Maas is only a minor river in Western Europe. It drains the uplands of eastern France and Belgium and discharges into the North Sea. Its catchment area covers approximately 33,000 km^2. The river basin faces occasional high floods and a low basic flow regime (Berger, 1992), its mean annual discharge is currently 260 m^3/sec, but peak discharge may be as high as 2800 m^3/sec. The river basin is superimposed on a number of morpho-tectonic units and can be classified as a complex basin according to Starkel's terminology (1990).The present study focuses on the Maas river, lower reaches located in the south-eastern part of the Netherlands. In particular on the effects of climatic change and neotectonics.

South of the study area, the river system is mainly erosive, as a result of a long period of uplift. From its sources in eastern France to Grevenbicht in the Dutch province of Limburg, where the Maas crosses the southern principal displacement zone of the Roer valley rift system, the Maas river successively flows across: the permeable Jurassic rocks of eastern France, the impermeable Hercynian Ardennes Massif and the permeable Cretaceous to Tertiary deposits of the South-Limburg block. The Ardennes section affects the flow regime in the study area most strongly, because of its steep relief, impermeable substratum and large tributaries. Before entering the Roer valley graben, the Meuse crosses the South Limburg block, a tectonic unit wedged between the Midi-Aachen thrust and the southern boundary fault zone of the Roer valley rift system. Between Grevenbicht and Nijmegen, the river crosses the fault blocks of the NW-SE trending Roer valley rift system. In the South-Limburg part of the river basin, uplift prevails, whereas in central and north Limburg, subsidence plays a key role. These two different settings: i.e. uplift versus subsidence give rise to very different preservation styles for the alluvial sediments deposited by the Maas river.

b) Outline of this thesis

In **Chapter 2**, the sedimentary sequence deposited in the Dutch coastal plain between 10 Ma and 3 Ma is described. During this period, prior to the main glaciations, the climate was subtropical to temperate. These sediments are known from open-cast mining in Germany as well as from a large number of deep boreholes in the Netherlands, drilled into the deepest part of the Roer valley rift system. The distance between these two areas is more than

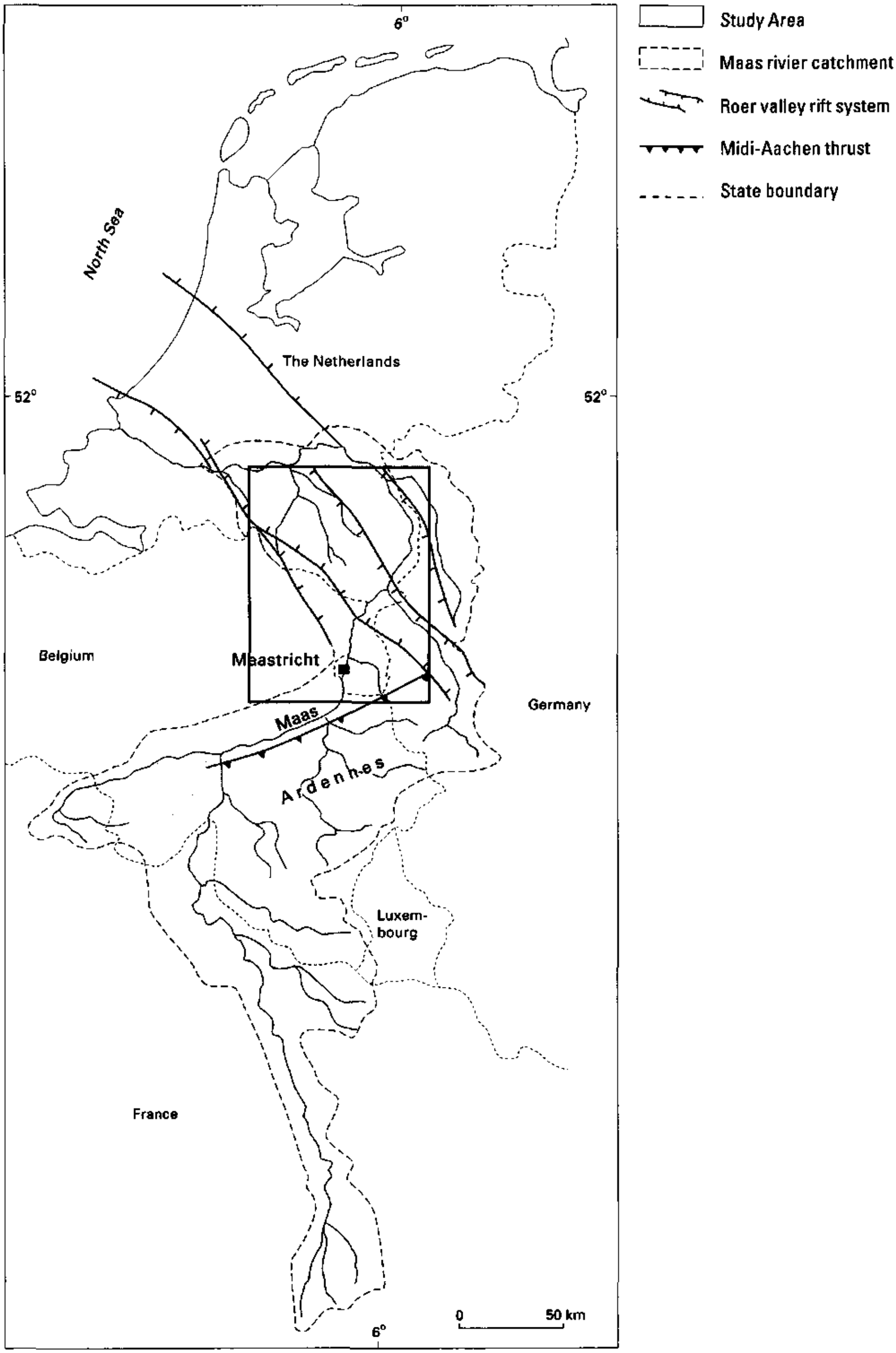


Figure 1.1 Setting of the study area

60 km, yet a comparable number (16-19) of fining-upward sequences is found in both areas. Their bounding surfaces, marked by conglomeratic deposits overlying fine-grained deposits, are considered laterally continuous. We conclude that the formation of these surfaces was controlled by external (allocyclic) factors. Correlation of the regional pollen-zone successions, last-occurrence dates of dinoflagellates and the $\delta^{18}O$ patterns from deep-sea records improves our understanding of the palaeo-climatic evolution of the northern hemisphere. The time frame resulting from these data indicates that sequence boundaries are on average 200,000 years apart (190,000 to 245,000 years). This corresponds closely with the main low-frequency $\delta^{18}O$ cycles found in North-Atlantic deep-sea cores, suggesting a common control mechanism. Whether fluctuations in sea level or in river discharge are the main controls of the formation of these bounding surfaces cannot be established. Both mechanisms are related to waxing and waning of the continental ice sheets. Ice volume directly affects the sea level. Climatic and river-discharge patterns are indirectly related to the ice volume, via the position of the polar front and its associated depression tracks.

In **Chapter 3**, a part of the sequence, deposited during the last 4 million years is described. These deposits have been studied mainly in fluvial terraces, 31 altogether, regularly distributed over the altitude range. Fluvial terraces are sedimentary successions that are preserved under conditions of relative uplift. Between 4 and 3Ma the sequences in the deeper part of the Roer valley graben, as discussed in Chapter 2, overlap in time with the terrace sequence.

During these past 4 million years, the frequency of the ice sheets' waxing and waning increased, as did that of tectonic events. The palaeogeographic changes in the study area are no longer linked with sea-level fluctuations. This did not significantly affect terrace formation. Whether the internal architecture of the terraces is different has not been established.

Field observations indicate that the Quaternary fluvial terraces formed predominantly under cold climatic conditions. By applying various dating methods, it can be shown that during the last million years, terrace formation was determined by the dominance of the orbital eccentricity (fourth-order or hundred thousands years climate-cycles). Between four to one million years ago, the frequency of orbital eccentricity was different. Periods, during which the sediment quantity stored in the alluvial plain was sufficient for terraces to survive erosive phases, were not as common as in the following million years. The interval, during which a cycle between two bounding surfaces was deposited, was longer (160-220 ka) than in the following million years. The data described in chapters 2 and 3 show that, at fourth-order climatic-cycle level, the continental record preserved in fluvial sediments implies environmental changes that match the oceanic record.

In **Chapter 4**, the inferred ages of the terraces is tested by simulating river-terrace formation in the three-dimensional MATER modelling program. This model simulates large-scale valley formation and terrace formation under generally accepted hydrological conditions, including the combined effects of tectonics and climatic changes. Modelling results are consistent with the assumptions of chapter 3. Tectonics was found to be the main control on terrace preservation.

To improve our understanding of the processes that result in terrace formation during a fourth-order climatic cycle, we describe one of these glacial-interglacial cycles in detail in **Chapter 5**. The period covering the last 130,000 years yields sufficiently detailed information on spatial depositional variations and climatic variations. The sediment data are obtained from borehole records. Stadial/interstadial alternations (fifth-order climatic cycles) show a similar deterministic effect on fluvial dynamics as the fourth-order cycles. Apparently a nested hierarchy exists.

The same combined field-observation/model approach is used in this chapter. Regional climatic conditions are set in an overall northern hemisphere context. The approaching continental ice sheet created conditions that favoured wind-blown sand transport. During the second half of the last stadial, large quantities of these wind-blown sands accumulated in the valley of the lower Maas. The resulting sediment quantity far exceeds that accumulating under 'normal' cold conditions.

At a fifth-order time scale, the effects of the cold climate seem to overrule the effect of the tectonically created accommodation space in the Roer valley rift system. The boundaries of tectonic sub-domains determine the variation in floodplain gradient. This gradient affects the river's response to the climate as well. Particularly interesting in the context of the present study is the response of the river just downstream from a tectonically determined knickpoint in the river gradient. During the last stadial, downstream the Maas had built up a low-gradient alluvial fan of up to about 10 metres thick.

Chapter 6 describes exposures of the deposits of this alluvial fan to determine in detail the response of the river to fifth-order environmental changes. The sedimentary sequence deposited during the aggradation phase and subsequent incision phase could be reconstructed in detail. The geomorphological and sedimentological evolution is related to 1000 to 2000-year-long climatic cycles (sixth-order climate cycles). The lack of dating methods at this time scale severely restricts age determination of individual sequences. Circumstantial evidence strongly suggests that the river's responses reflect short-term climatic fluctuations. This observation corroborates the nested-hierarchy concept. Vegetation evidently also played a crucial role in the interaction between climate and fluvial dynamics at this time scale.

In the previous chapter a strong correlation between climatic cycles and sedimentary sequences was demonstrated. This implies that fluvial sequence stratigraphy may provide a valuable tool for improving the time resolution in the stratigraphic record. of basins with associated fluvial systems. This principle has been verified in the last two chapters.

Chapter 7 focuses on the neotectonics of the Roer valley rift system: sedimentary sequences related to fourth-order climate cycles are used to reconstruct the uplift history of the South Limburg block. At a time scale of 100 ka, accelerations and decelerations in uplift are evident. Thus, the long-term uplift history can be divided into phases. Some of these phases can be related to major geomorphological phases that are known from other parts of the European plate. Temporary changes in the direction of the moving lithosphere plates may be responsible for this relationship. Comparison of the locations of river floodplains with fault-trace patterns shows a strong correlation. This indicates that temporary changes in the directions of river channels in relation to fault-trace patterns may provide information on the dynamics of the upper crust. From this point of view, we analyzed the sedimentary fill of the Roer valley graben. Horizontal displacement along the strike of the principal faults was found to have played a much more important role than previously thought.

In **Chapter 8**, the morphotectonics of the Roer valley rift system, as expressed by lineaments in the present-day land surface, is compared with the vertical movements deduced from the 117-years-long record of 2922 geodetic benchmarks. This shows significant differential movement, in the order of cm/cy, of wedge-shaped blocks at a 10-km scale. This movement corroborates the right-lateral sense of relative movement indicated by the lineaments. This observation is consistent with the conclusion of the previous chapter.

2. Alluvial sequence architecture and climatic change; The Late Tertiary Rhine - Maas (Meuse) alluvial/delta plain in the southern North Sea-basin

Abstract - *A 6.5 Ma long record of Late Tertiary unlithified alluvial/deltaic plain sediments preserved within the Roer Valley Graben contains 31 buried fluvial erosion surfaces. These surfaces are inferred to occur basin wide. A restricted number of these horizons is well dated by biostratigraphic and radiometric data. Pollen-dated sequences supply additional age control. The pollen ages have been sharpened by correlation of the pollen-based regional climate stratigraphy with the astronomically tuned climate patterns shown by the oxygen-isotope deep sea record.*

To arrive at an approximation of the frequency band in which the erosive surfaces occur between 9.7 Ma and 3.15 Ma, we used the climate record in conjunction with the astronomically calibrated palaeomagnetic time scale of this period. We found a frequency of around 200 Ka with a range between 190 and 245 ka. This frequency matches other records like sea-level fluctuations and temporal cycles of Late Neogene oxygen isotope compositions.

This suggests that the fluvial sedimentary record can be correlated to some extent with the oceanic record.

Introduction

In this paper we focus on fluvial deposits in a coastal lowland setting, prior to the first Quaternary glaciation. We will analyze the significance of supposedly laterally continuous lithological sequences bounded by fifth and sixth order erosive surfaces.

These sequences are contained by the Late Tertiary Rhine/Maas sedimentary record and are recognised in drillhole data and exposures.

The Roer valley rift system (Fig. 2.1) provides the setting to examine this theme. The selected study area contains the rift main depocenters and is constrained by tectonic structures. It covers an area with a down-dip length of about 65 km and a width of 25 km. Preliminary subsidence analyses suggest that subsidence in substantial parts of this depocentre was continuous at least from about 10 Ma BP till about 3 Ma BP (Geluk et al., 1994). A long, but fragmented, pollen record (Zagwijn, 1960; Zagwijn and Hager, 1987) provides a valuable tool in the correlation of drillhole records between different tectonic sub-units.

A dating frame (for associated marine sediments) comes from: (1) a few Sr-isotope dates; (2) last appearance dates of dinoflagellates; (3) (for the fluvial sediments) correlation of trend changes in the regional climate evolution, as expressed by pollen, with similar changes in the global climate as expressed by the high resolution deep-sea record. The latter is well dated by the recalibrated magneto-stratigraphy (Hilgen & Langereis, 1993; Shackleton et al., 1993; Krijgsman et al., 1994).

It will be shown that the Late Tertiary fluvial dynamics experienced important changes with a frequency of the order of 200 Ka.

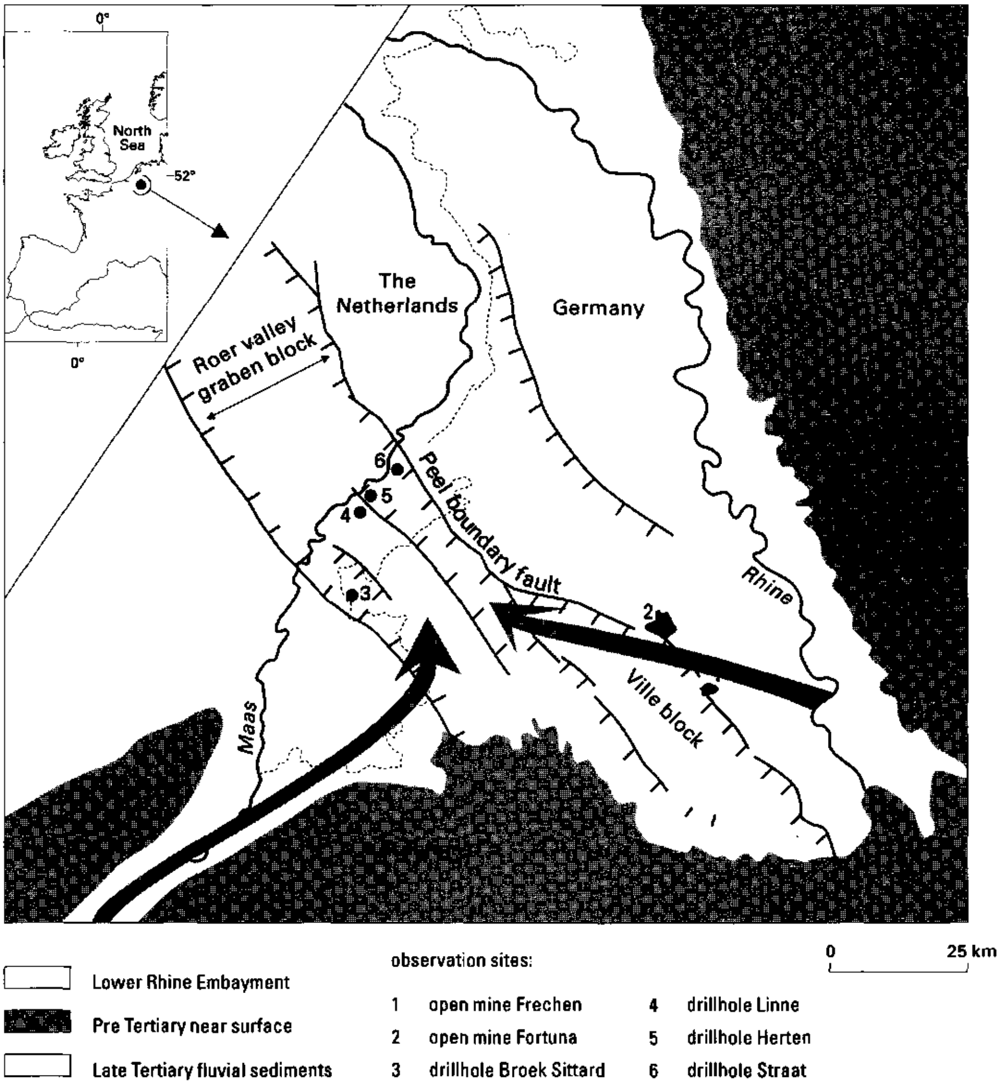


Figure 2.1 The setting of the observation sites with respect to distribution of late Tertiary fluvial sediments within the Roer valley rift system. The sedimemts NE of the Peel boundary fault are very discontinuous and probably supplied from an other (local) source.

The Lower Rhine Embayment

Basin characteristics and stratigraphy

The Lower Rhine Embayment belongs geographically (Fig. 2.1) to parts of Germany and the Netherlands. It is a funnel-shaped tectonic depression, opening towards the northwest, forming part of the west central European rift system. It is positioned between the adjacent uplifting areas (Ardennes and Rhenish Shield) and the southern North Sea Basin. The depression is composed of several north-eastward tilted blocks among others, the Roer valley graben block and the Ville block. The upper few hundred meters of the subsurface of the Roer valley block is mainly known from a dense net of drillholes, whereas very large open lignite mines have been excavated in the Ville block.

Fluvial sediments appear in the southeast of the embayment since the Middle Miocene. They pass into full-marine sediments in the northwest. Fluvial deposits have been supplied by an axial 'pre-Rhine' fluvial system together with the pre-Maas (or east Maas) as a lateral tributary.These rivers and their successors drained an area extending southwards of the Rhenish and Ardennes Shield (Boenigk, 1981), although their catchment areas were smaller than today's (Quitzow, 1975).

Studied episode (Fig. 2.2)

Due to the interplay of sea-level fall and the onset of tectonic uplift of the hinterland (see chapter 3) rivers were able to build a delta plain in the study area. Before this, mainly peat accumulated with very insignificant clastic supply. This mayor change in the basin filling complex is correlated here with the rapid (type 1) sea-level fall just after the Serravallian/Tortonian boundary (about 10 Ma). This correlation is supported by data from dinoflagellates (see below). This drop in sea-level corresponds with the boundary between super cycles TB2 and TB3 of the Haq et al., (1988) sea-level record. From this 'event' onwards, through the Pliocene, the coastline gradually shifted some 100 km north-westward (Zagwijn, 1975; Zagwijn and Hager, 1987). The overall trends in the 2nd-order sea-level position and in the position of coastline between 10 Ma and 3 Ma are opposite (Fig. 2.2). This suggests that the effect of sediment supply exceeded both subsidence and long-term sea level behaviour. This favours tectonic uplift as a key element in the driving forces beyond the delta progradation.

"Continuity" of the studied records is in fact a matter of scale. River sedimentation never reaches the same continuity as for example sedimentation in the deep-sea, but the best possible continuity is potentially reached in alluvial plains in a lowland setting. The sedimentary environment had persistently an (upper-)lowland character, under ongoing subsidence, so continuity may be assumed. The upper time boundary of the studied episode is taken at around 3 Ma. Around this time the rift system was subject to a major tectonic event. This event is expressed regionally in two ways by: (1) the onset of a 1 Million years long hiatus within parts of the graben. This is evidenced by both mineralogy of the fluvial sediments and the pollen content. Time equivalent fluvial deposits have been identified north of the Peel boundary fault (Zagwijn, 1961), indicating that at the onset of this period fluvial sedimentation was diverted within the embayment across the principal displacement zone of the Peel boundary fault line (Fig. 2.1). (2) a strong acceleration in uplift of the southern part of this rift system (chapter 7; Van den Berg, 1994).

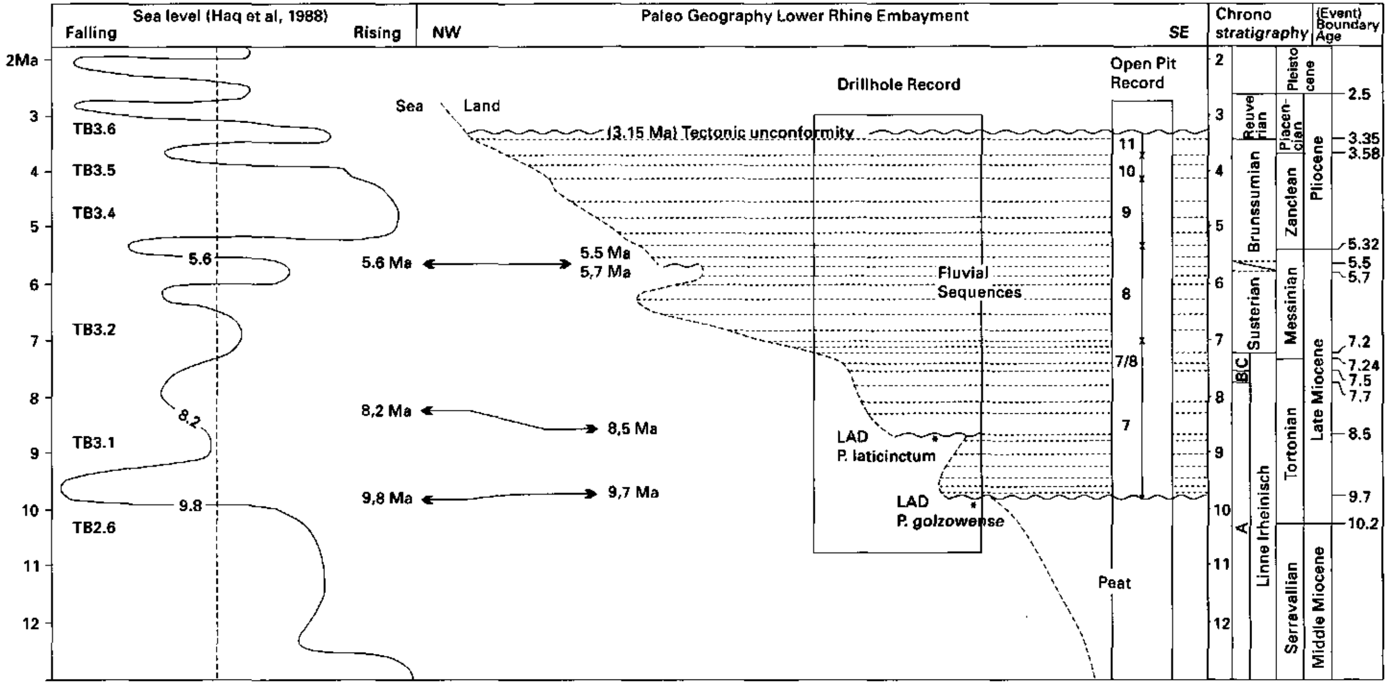


Figure 2.2 Schematic of the stratigraphic development of the alluvial- and coastal plain in the Lower Rhine Embayment in relation to third order eustatic sealevel fluctuations (Haq et al. (1988). Numbers in the open pit record refer to the main lithostratigraphic units. Stage/"event" boundary ages are discussed in text. The alluvial plain-building sequences (to be discussed below) are of fourth order.

Sedimentary units

The sediments have been described at two hierarchical levels:

1) Large-scale (tens of km, lateral), "uniform" lithological successions were distinguished by Schneider and Thiele, 1965; Hager, 1977; Boenigk, et al., 1979.
2) Detailed sedimentological information from Tagebau (= Tgb.) Frechen and Tgb. Fortuna-Garsdorf was presented by Boersma et al.(1981) and Boenigk (1981). These open pits are each a few km wide and are about 10 km^3 apart in palaeo-downstream direction, (Figs. 2.1, 2.2 and 2.3).

Ad 1: In their lithostratigraphic subdivision of unlithified sediments in the Lower Rhine Embayment, Schneider and Thiele (1965) introduced both formal names (Fig. 2.3) and numbers (01 through 19, with subdivisions, Fig. 2.2). Relevant to the present study are, in stratigraphical order: the Inden Formation (= 7): a coastal lowland facies consisting of predominantly sand, clay and peat; the Transitional Series (= 7/8): sand and fine gravels; the Main Gravels (= 8): coarse gravelly sands; the Rotton and Reuver Series (= 9 through 11): Units 9 and 11 are predominantly clay- rich deposits whereas Unit 10 consists of gravelly sands. Unit 9 is further subdivided into 9A - 9B - 9C. These labels stand for the Lower (= 9A) and Upper Rotton (= 9C), both clay-rich units separated by the generally coarse Unit 9 B.

There appears to be a fair correspondence in time between some of these litho-facies variations and 3rd-order sea-level variations (cf. Fig. 2.2). The meandering lowland facies of the Inden Formation (7) may be correlated with the TB 3.1- 3.2 highstands. The next episode between the end of TB 3.2 and the onset of TB 3.4 highstands, is a period dominated by sea-level lowstands.

We tentatively attribute the shallower facies of Unit 8 (see below) to a relative upstream position of the study area related to this sea-level lowstand episode. Evidence from channelfill structures indicative of a tidal-fluvial environment (Quarry Caumans-Scheepers near Schinveld; Ruegg, 1978) found in beds belonging to the clay-rich Unit 11 suggests that the latter unit (as well as Unit 9 with a similar facies complex) match respectively the sea-level highstands in cycles TB 3.6 and TB 3.4/TB 3.5 of the Haq et al. (1988) chart. The tripartite character of Unit 9 (fine-coarse-fine) possibly correlates with the partition TB 3.4/TB 3.5 of its marine counterpart.

Such correlations at third-order level are not controlled by sea level alone as a climate influence is present as well (see below).

Facies (Fig. 2.3B)

Ad 2: Units 7 and 7/8 are interpreted to be have been formed by meandering rivers in a lowland area (with bankfull depth between 10 and 20 m. (pers. comm. R. Boersma, 1994)). An upward change in fluvial style within Unit 8 is due to a gradual shift towards less channelized flow in shallower systems; the latter show great similarity to the (Pleistocene) braided systems in the region (Boersma et al. 1981).

The sedimentological studies by Boersma et al.,(1981) reveal the gradual change between the facies end-members 7 and 8.

Around the transition from Unit 8 to Unit 9 the braided character of Unit 8 abruptly changes into a stacked meandering succession with, several-metres-high, lateral accretion surfaces and associated oxbow fills. Complete sequences in quarry Hendrik (Brunssum, The Netherlands), suggest a palaeo-riverdepth of around 13 m.

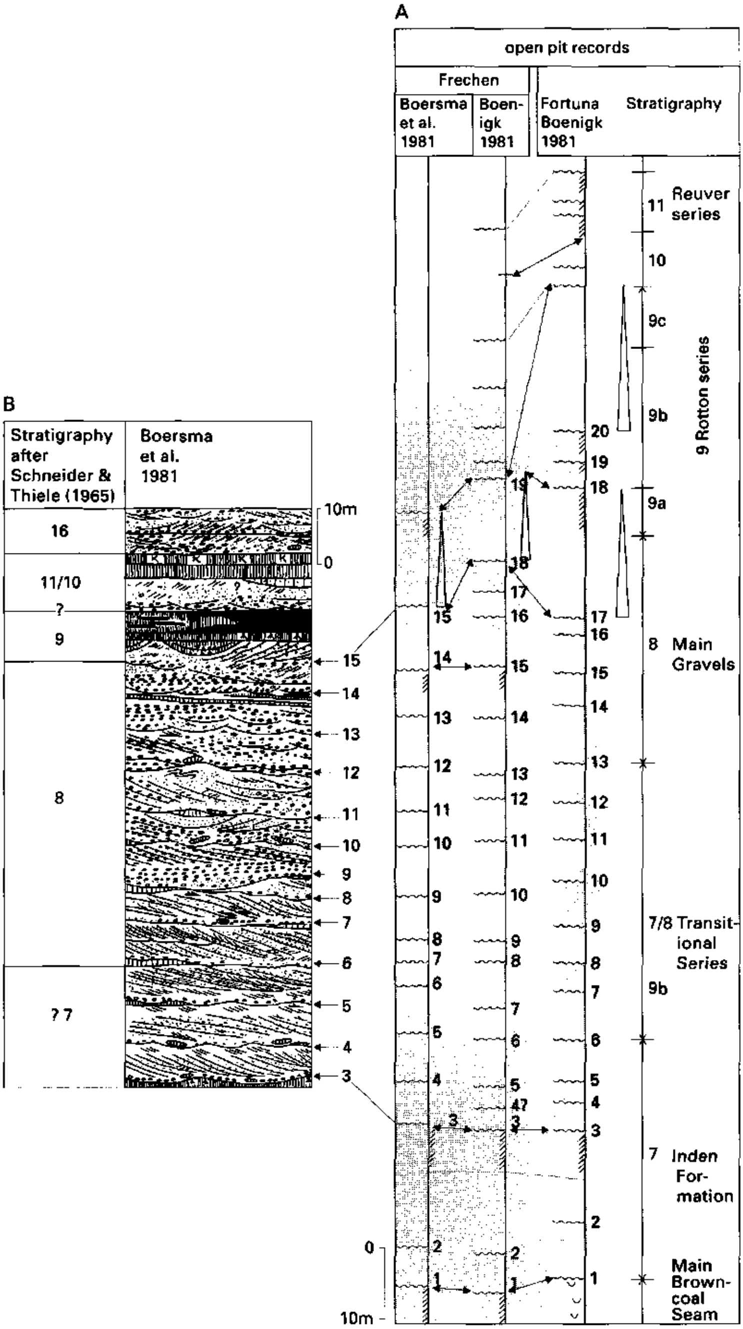


Figure 2.3A Composite of sedimentological observations made on outcrop sections in the open pit mines Frechen and Fortuna by Boenigk, 1981 and Boersma et al., 1981. To facilitate the comparison only major bounding surfaces (fifth order) as well as their correlations are indicated.

Figure 2.3B Facies drawing of a section in the Frechen open pit by Boersma et al., 1981. Numbers pointing at major bounding surfaces correspond to the section of Boersma et al, in the composite of figure 2.3A.

Bounding surfaces

Bounding surfaces are common to bed-load-dominated river sediments; they mark reactivation surfaces. Fifth- and sixth order surfaces, as described by Miall (1988) and Wizewich (1993), are most prominent within the studied sediments. Wood remains, clay balls, boulders directly underlain by better sorted and finer sediments are their lithological expression. These most prominent internal surfaces can be followed in exposures, over more than a kilometre laterally. They bound channel sheets composed of truncated channel-fill sequences which are either or not fining-upward (cf. Fig. 9 in: Boersma et al., 1981). Such features have an undulatory, but overall sub-horizontal appearance. The stacking also strongly reminds the repetitiveness exhibited by the Pleistocene counterparts of these fluvial deposits in the region. In addition they remind of features which can be recognised in drillhole data. In the study of unlithified and relatively shallow-subsurface sediments one has to combine drillhole data and exposure data. Precise correlations inevitable remain debatable by his method, but at present there is no other way available to improve the lateral resolution. This drawback may be compensated by relative high resolution dating possibilities of the subject units.
Boersma et al., (1981) and Boenigk, (1981) have shown, by studying exposures, that the Inden Formation (Unit 7), the Transitional Series (Unit 7/8) and the Main Gravels (unit 8) are together built by a stacked series of 15-18 sequences (Fig. 2.3A). These sequences are very similar in thickness in the two open pit exposures. This is in contrast to the younger Unit 9, which is only about 7 m thick in the Tgb. Frechen, whereas it reaches 55 m in Tgb. Fortuna (Fig. 2.3A) This difference between the Miocene and the Pliocene part of the sediment record might have been caused by accommodation differences due to a regional revival of tectonics during the lower Pliocene.

Inter-drillhole correlation

In the Dutch part of the Roer valley graben only along the southern margin the discussed sediments, are very limitedly exposed. To compare the data from the open pit mines with data from sections near (0 - 40 km) to the palaeo-shoreline, drillhole information has to be included.

We identified reactivation surfaces in high quality drillhole data of unlithified alluvial sediments, generally by levels with strong contrast in the dominant grainsize. To this we include drillhole intervals described as mixed gravel and clay (Fig. 2.4). As the expression is not equally well in every drillhole, gamma-logs may be helpful to find the most reliable boundary-depth.

When we compare drillholes that are a few kms apart, the depth of levels with strongly contrasting lithologies varies in the range of a few metres. This depth variation is generally significantly less than the thickness of the wedged units. So when such levels are correlated, an undulatory surface shows up with an amplitude similar to those of major fluvial-erosional surfaces, seen in exposures in this region. This type of drillhole correlation suggests that the erosional surfaces can be followed over an extensive distance. This lateral continuity suggests at least a sub-regional significance of the buried surfaces.
The roughly equal vertical spacing between the boundaries is also striking. Both characteristics of these surfaces stress their close relationship with regular system changes, fundamental to the genesis of major bounding surfaces.

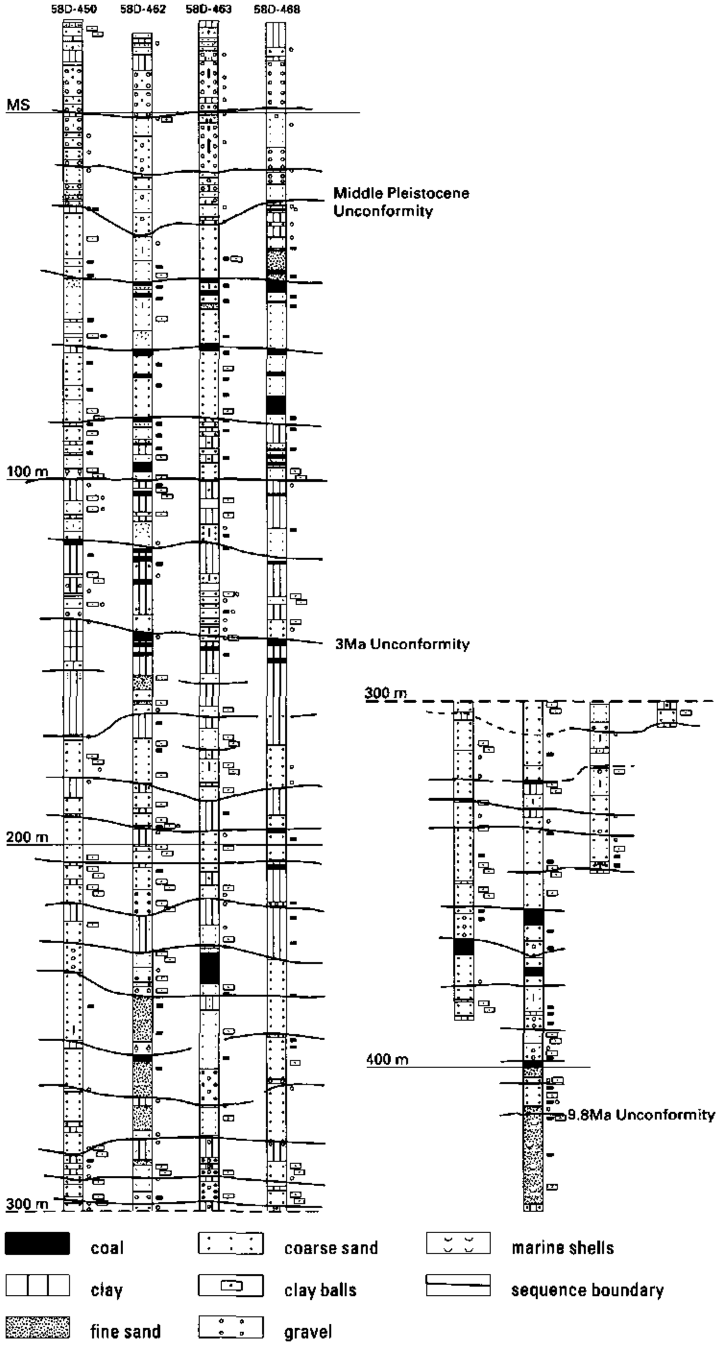


Figure 2.4 Drillholes reveal abrupt breaches in the lithological succession. These are found at regularly vertical spacing. In drillholes spaced a few hundreds of metres apart, we meet them at comparable levels and interpret them as sequence boundaries; here indicated by wavy horizontal connecting lines. Example from wells around Herten (Drillholes 58D-450/462/463/468). For location see figure 2.1.

The delineated sedimentary units are of about 8 to 20 m thick (Fig. 2.4). Often these units show an internal "fining-upward" trend, but irregularities in the grain-size trends give them a multi-storey appearance. We assume that the recognised units between two bounding surfaces, represent parts of channel-fill sequences and the separating boundaries consequently represent (fourth?), fifth and sixth order elements within the hierarchy of bounding surfaces.

The controlling factors behind the genesis of this element of fluvial architecture may be produced in various ways (channel migration or allocyclic river reactivation). To work on this problem one has to work fluvial-system wide.

Lateral continuity of bounding surfaces

The lateral continuity of the above described surfaces has been examined by the construction of a network of long cross sections through the Roer valley graben (Fig. 2.5). These sections cover the area between the graben bounding principal displacement zones. They have been chosen both parallel and perpendicular to the paleo-flow direction. The area is tectonically divided into three, north-eastward tilted, units: a central high bounded by two lows; maximum subsidence is found in the north-eastern low. It contains the sedimentary record of the rivers Maas and Rhine. Sometimes as separated systems, sometimes as a combined river.

The drillhole correlation is controlled by biostratigraphic datum lines. The regional biostratigraphy is based on internal reports of the Palaeobotany Department of the Geological Survey of The Netherlands. Between two biostratigraphically defined reference horizons (first order correlation lines in figure 2.5) we found that the number of sequences recognised in the various drillholes was very much alike alongstream. The observation that fifth - to sixth order erosional surfaces are on average sub-horizontal at micro-scale (exposures), at meso-scale (drillhole clusters) at the scale of a tectonic units (10-20 km) suggests a lateral continuity and thus a "layer-cake" sequential architecture.

During certain intervals of time, there is no lateral continuity across the graben dividing faults. This is attributed to periodic relative uplift of the central high and well explained by the stress-controlled regional tectonics in interplay with tilting (Van den Berg, 1994). The Roer valley graben is characterised by tilted tectonic units. Therefore the area of maximum subsidence (the north-eastern low) potentially contains the most reliable record of the maximum number of layers deposited in the considered time interval. Figures 2.5 and 2.6B show the 31 fluvial sequences preserved in the deepest part of the graben between the 10 Ma and 3 Ma unconformities. The drillhole section in figure 2.5 exhibits migration of the channel belt in response to the overall tectonic tilt of the Roer Valley block. Thick and vertically "continuous" series of Pliocene overbank clays and coals in the southwestern low (60D-1025) form a lateral floodbasin equivalence of a number of fluvial sequences, composed of channel deposits, in the north-eastern low. The pollen-stratigraphy of these overbank series shows that they represent long periods of time i.c. the supposed "complete" Lower Pliocene (Zagwijn, 1960; de Jong, 1982). Recognition of sequence boundaries within the thick overbank clays is very tenuous. Johnson and Fitzsimmons (1994) suggest that in this type of facies laterally persistent coals represent 'maximum flooding zones' and therefore (para)sequence boundaries. Our record occasionally suggested the same conclusion for some persistent coal layers, but we cannot generally confirm this due to the scarcity of information.

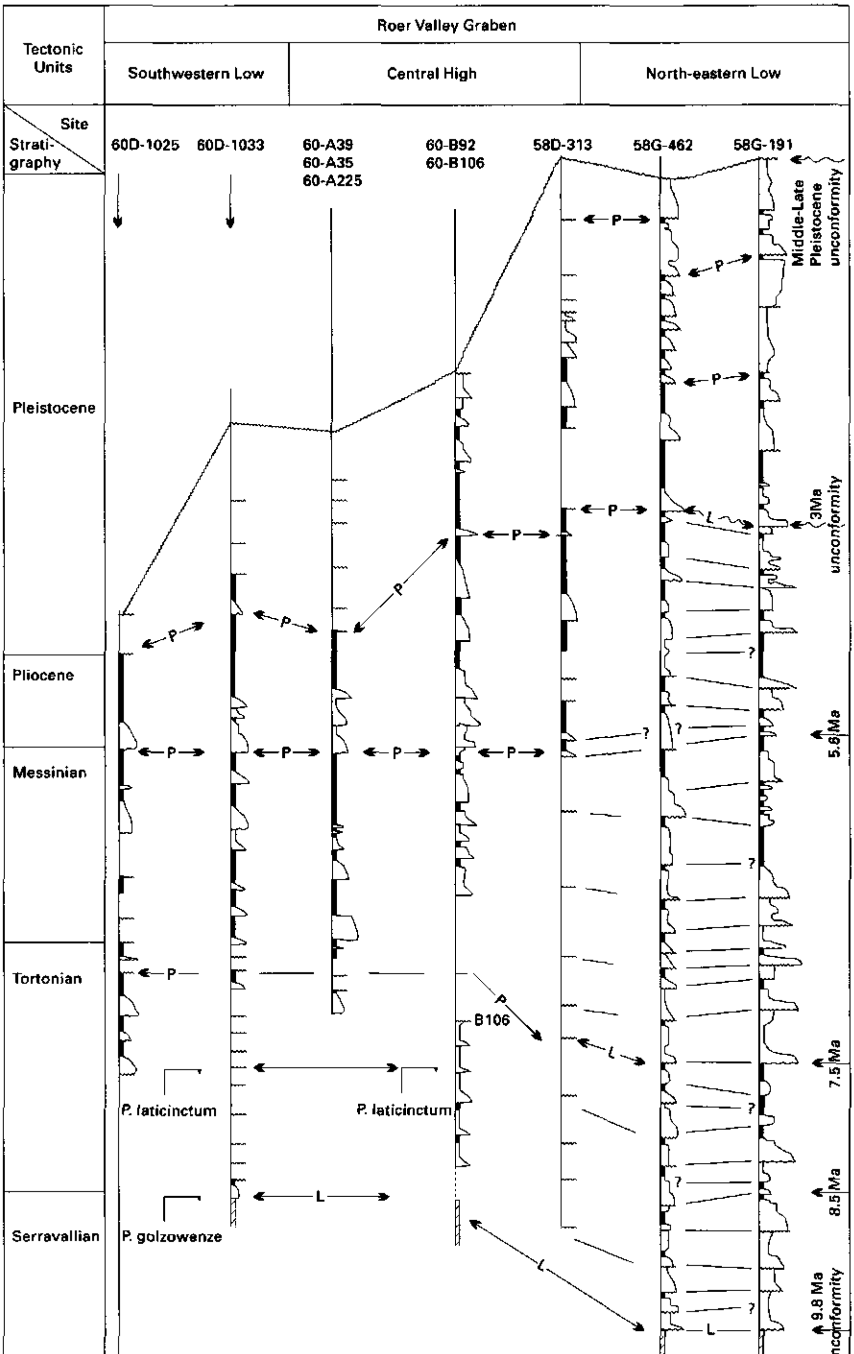


Figure 2.5 Review of the occurence of fluvial sequence boundaries in different domains of the Roer valley graben. The sections are restored with respect to the first sequence boundary after the Susterian/Brunssumian pollen-zones-boundary (5.6 Ma). First-order correlation lines are indicated by arrows. Sequences younger than the Middle-late Pleistocene unconformity have been omitted.

Regional correlation and significance of sequence boundaries

We have discussed above the occurrence of bounding surfaces and intercalated fluvial sequences, from two sources of information (exposure studies and drillhole sections). Boersma et al. (1981) and Boenigk (1981) showed that the Inden Formation, the Transitional series and the Main gravels together contain 15 - 18 channel-fill sequences in their descriptions, (cf. Fig. 2.3A).
The drillhole based cross-sections in the Dutch part of the Roer valley graben, an area 65 km downstream, revealed for the stratigraphic interval 19 sequences in the south-western low and 21 sequences in the north-eastern low (cf. Fig. 2.4). These numbers concern the interval between the 10 Ma unconformity and the 5.6 Ma time boundary. This interval presumably covers slightly more time (the equivalence of 2- 3 sequences) than the discussed exposure section.
This close correspondence in sequence-number between various sites in the palaeo coastal plain may suggest that the geological evolution within the Lower Rhine Embayment was very uniform basin wide. Not only at formation level (Zagwijn and Hager, 1987) but also at the lower level. These erosional based channel-fill sequences may be considered as parasequences sensu Van Wagoner et al. (1988); Mitchum et al. (1977); Mitchum and Van Wagoner, (1991). These authors describe coarsening-upward sequences and include the presence of marine flooding surfaces and lateral continuity in their definition. Marine flooding surfaces are not present in our case or have not been recognized. We attribute this difference to the more up-dip position of the considered facies.

Strictly speaking this comparable number of sequences does not prove that we are dealing with extensive, tabular shaped members within the formations, but the convergence of

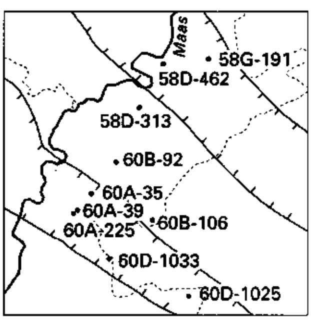
Legend to Figure 2.5

0
10 m
20 m
sequence boundary

ø = f m c g
f = fines
m = medium
c = coarse
g = gravel

insufficient data for ø scale

last occurrence
←P→ pollen-based correlation
←L→ lithology-based correlation
marine sediments



observations strongly does suggest it; (unpublished data from 3 D - seismic studies of Carboniferous fluvial deposits from the Southern North Sea indeed showed such a lateral continuity, W. Kouwe, pers. comm., 1995).

Long-distance consistency in the number of sequences consequently bears in it that they form the main building elements in a layer-cake-architecture of fluvial deposits (Fig. 2.2). A most distinctive aspect of such a geometry is that a relative limited number of major bounding surfaces has been preserved from such a long period of time. (a few tens of major bounding surfaces vs. millions of years). Lateral channel migration in combination with avulsion and subsequent preservation of channel deposits can explain the palaeo-valley wide appearance of formation boundaries (= 6th order bounding surface). It does not explain why, at the same palaeo-valley wide scale, at a regular distance above it new bounding surfaces, like the one's discussed, will be formed. These are of 5th order.

Based on the following arguments and characteristics we interpret the 5th order bounding surfaces to be of river-system scale and the result of allocyclic (external) controls:

1. Gravel analyses on Tortonian sediments taken from both tectonic lows showed that rivers Maas and Rhine were occasionally geographically separated (Burger, 1987; pers. comm. 1994). Both rivers still produced an equal number of sequences in these areas.

2. The enclosed sequences are found over a substantially wide area. It measures about 65 km down-dip with a width of 25 km.

3. Sequences occurred in a single channel system (Unit 7) or a multi-channel system (Unit 8) as well as in the gradual change between the two. Despite the differences in facies, the dimensions and their regular distribution over the accommodated record stay comparable.

These points underline the role of external parameters on the river dynamics leading to the formation of new bounding surfaces. These allocyclic controls may include tectonics, climate and/or base-level changes. In order to analyze these factors in further detail we need a good time resolution in conjunction with independent records of these parameters. Tools to obtain the best possible resolution will be discussed below.

Age - and biostratigraphic control on sequences

Absolute time

$^{87}Sr/^{86}Sr$ chronostratigraphic work (Beets, 1992) on marine sediments, offshore of the river mouth in the Lower Rhine Embayment, provided radiometric time control on local last-appearance-dates of dinoflagellates species. The same species were found in our drillhole records (Fig. 2.5).

Indirect absolute time control was obtained from correlations (to be discussed below) of the climatic evolution derived from deep-sea records (with astronomically tuned palaeomagnetic controls) with the regional pollen-based climatic patterns. To unify the timescales of various deep-sea records to an astronomical chronology we used the original palaeomagnetic interpretations of the various records but adjusted the reversal boundaries to the modified version of the recently developed geomagnetic polarity time scale (Cande and Kent,1992; Krijgsman et al.,1994). In between the reversal boundaries, we adjusted subsequently the records by linear interpolation. This assumes linear rates for the sedimentation for these stretches. This is not always correct, but it caused no substantial problems.

Biostratigraphy

Dinoflagellates. Dinoflagellates can be used for first order correlation with the standard biostratigraphic zonation. The last appearance datum (LAD) of *Palaeocystodinium golzowenze* marks the top of the marine sediments underlying the fluviatile series (Fig. 2.2 and 2.5). This is observed in drillholes Broeksittard (drillhole code: 60D-1033) and Koningsbosch (60B - 106) (Herngreen (1987; Zevenboom, 1994). From drillhole Cuyck (46A-147), positioned more seaward from the study area, Beets (1992) provided Sr-isotope dates from fossiliferous sands directly overlying a lithological discontinuity that coincides with the LAD of *P. golzowenze*. His data suggest an age of about 9.7 Ma for this unconformity. According to the standard biostratigraphic zonation this LAD marks the boundary between the Middle and Late Miocene which is dated presently at 10.2 Ma; previous work gave an age of 9.8 Ma for this boundary (Haq et al., 1988). The relatively small differences between these ages provides a fair constraint for the onset of the fluvial sedimentation.

Six sequences higher up in the record, in a fluviatile-lagoonal environment, the last appearance of *Pentadinium laticinctum* has been recorded (Fig. 2.2). This species is regarded as a reliable marker as it is little temperature dependent (pers. comm. D. Zevenboom, 1994). Its extinction is found to occur within the nannofossil zone NN11 (Cn9a) (Head, 1989). Consequently the confining dates for the extinction are between about 8.4 and 7.2 Ma. (Haq et al., 1988). A more precise Sr-isotope age of 8.5 Ma (Beets, 1992) has been reported for the extinction of *P. laticinctum* in drillhole Heumensoord (46A-260).

Arboreal pollen. For biostratigraphic control in the fluvial environment we have to rely on pollen assemblages preserved in the fine-grained sediments. Abundancy changes in arboreal species groups may reflect a wide range of environmental changes, ranging from site-specific factors to major global climatic changes. The first depend primarily on local factors whereas the latter are controlled by long-term changes in supra-regional circulation patterns in the atmosphere and Atlantic Ocean (variations in heat-release in response to the formation of North Atlantic deep water, intensity of the Gulf Stream and ocean surface water temperatures). The reconstruction of the regional palaeo-climatic evolution is primarily based on the changes in abundance of evergreen tree species (Zagwijn, 1960; Zagwijn and Hager, 1987). The characteristics of the inferred overall pattern of floral changes is consistent with the Mediterranean pollen record (Suc and Zagwijn, 1983). This consistency shows that the main zoning at least can be used for palaeo-climatic correlations at the scale of the northern hemisphere. Such correlations with deep-sea proxies of climate change may provide insight into the timeslices covered by the pollen zones. Particularly the recent progress made with the astronomically tuning of the palaeomagnetic time-scale (Shackleton, 1990; Hilgen, 1991) is of great importance to this.

Parallel trends between the pollen record and the deep-sea record

Within the Lower Rhine Embayment, Zagwijn (1960, 1967, 1990), Zagwijn and Hager (1987) and De Jong (1982) have identified a series of climatically related floral changes over the last 16 Ma (Fig. 2.6A).

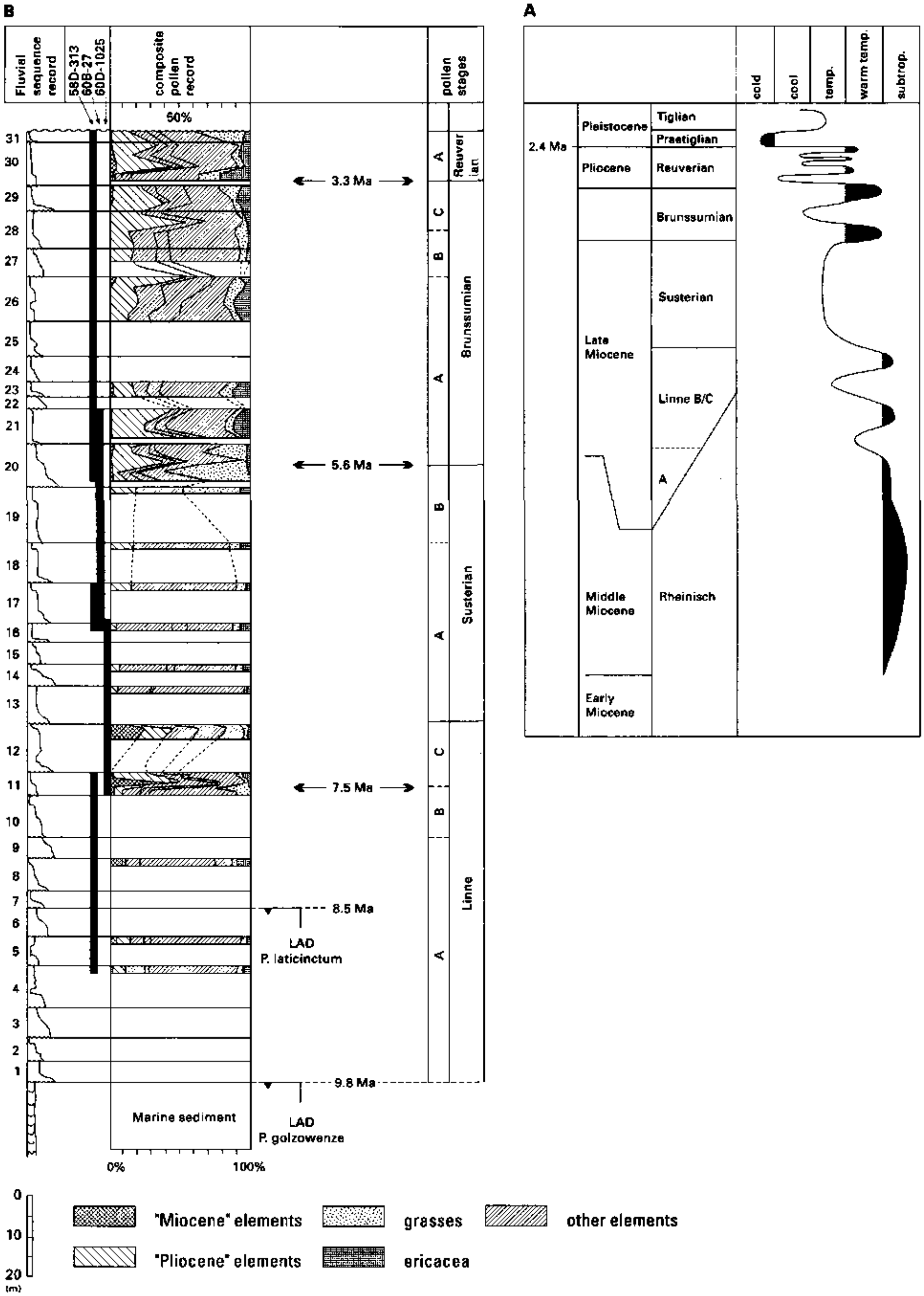


Figure 2.6A Neogene climatic development, interpreted from pollen assemblages in the Lower Rhine Embayment. (After Zagwijn and Hager, 1987)

Figure 2.6B Total number of fluvial sequences preserved in the study area and its composite pollen record. Pollen compiled with the data from drillholes 58D-313, 60D-1025 and 60B- 27 (De Jong, 1982). Pollen control is not available for the sequences: 1,2,3,6,7,9,10,15,24. Age model is discussed in the text.

Pattern similarity in inferred temperature evolution, between the land (= pollen) and the deep-sea records (= $\delta^{18}O$ and some foraminiferal species) is significant at a time-level of million years. This similarity indicates a common general climate control. It suggests that correlation of the two types of records allows us to calibrate the pollen-zone boundaries by making use of the good age-control of the deep-sea records. This procedure ignores possible time-lagging effects caused by differences in registration of climate changes by the various record-types.

The Miocene (sequences 1-20 in figure 2.6B)

Both the pollen (Fig.2.6A) and the deep-sea records (Ruddiman et al, (1987) show that after a long warm period during the Middle Miocene, the Late Miocene experienced cooling events. The distinctive pollen assemblages are known as the Rheinisch, Linne (divided up into -A,-B,-C) and the Susterian. Following the Susterian, a shift to warmer and more oceanic conditions is registered by the Brunssumian (= about Lower Pliocene). This shift loosely correlates with the transition from the Miocene to the Pliocene.

The upper part of the Miocene climatic record is covered only by a few North Atlantic deep sea records. The $\delta^{18}O$ records of DSDP-holes 552-A and 609 cover the polarity epochs 5, 6 and 7 p.p. (Fig. 2.8). Besides variations in $\delta^{18}O$ values, the coiling shifts in the planktonic foraminifer *Neogloboquadrina* (Fig. 2.7) are of great interest to our correlations also. This species exhibits, in the course of the Miocene, shifts in coiling-direction preference. The right coiling species has a preference for warmer ocean waters than its sinistral form (Hooper and Weaver, 1987; Ruddiman and McIntyre, 1981). The coiling-direction ratio trend changes from dominant right (= warm) in chron 7, to left (= cool) in chron 6 and back to dominant right (= warm) in the latest Miocene in the lower chron 4 . A very similar temperature trend is expressed by the Linne -(= warm), Susterian - (= cool) and Brunssumian (= warm) pollen zones. We come back to this subject below.

The Linne (sequences 1-12 in figure 2.6B)

The Rheinisch and its upper equivalent the Linne, represents a long period of Middle Miocene warmth. This period terminated with at least two oscillations (Fig. 2.6A). The cooler parts of these oscillations are still warm-temperate conditions. These oscillations are named the Linne-B and -C sub-zone. 'Miocene' floral elements (indicative for subtropical climatic conditions) reach a last maximum of about 40% -60% of the pollen sum at the end of the Linne B (Fig. 2.6B: depositional sequence 11). We will use this "warm spike" in our correlations.

The run down from the Middle Miocene warmth towards the Upper Miocene cooler conditions is expressed in the oxygen-isotope record of DSDP-hole 552A, by two cycles (around the boundary between chron 6 and chron 7, (Fig. 2.8). The lower of the two cycles reaches extreme low values in $\delta^{18}O$ in its optimum. This strongly resembles the above mentioned pollen patterns with the spike of "Miocene-elements" at the transition from Linne - B/ C. We tentatively correlate the two expressions of the short revival of warm conditions. This correlation may provide us with two absolute-age-approximations:

(1) A significant shift towards long-lasting cooling of surface waters in the Atlantic Ocean is well documented at DSDP-sites 552 A and 611 C and falls around 7.7 Ma. We take this age as the Linne A / B boundary.

(2) The extreme low $\delta^{18}O$ values in benthic foraminifers, which signal we correlate with the warm spike at the Linne B - Linne C boundary, occur at 7.5 Ma in chron 7 (Keigwin et al.,1987).

If indeed the present resolution of the continental and marine climatic signals, around 7 Ma, match at a one to one basis, we suggest that the Linne/Susterian boundary in the pollen zonation may equal the classic Tortonian/Messinian boundary. The latter has been re-dated recently at 7.24 Ma (Krijgsman et al., 1994).

Major climatic changes occur with a frequency of around 300 Ka in this interval (Fig. 2.8). This figure is important to the interpretation of frequencies derived from the fluvial-sedimentary record.

The Susterian (sequences 13-19 in figure 2.6B)

A general trend, during the Late Miocene, towards cooler conditions culminated on the continent in a time-interval dominated by Pine trees. This interval is called the Susterian. Pollen diagrams suggest the vegetation succession to have been rather monotonous during this interval.The succession in the coiling ratio of *N. pachyderma*, a measure for the ocean-temperature conditions, shows a similar monotonous and cool expression, for the early Messinian (Fig. 2.7) (after Hooper and Weaver, 1987).

The presence of a cooling trend in the world climates has been widely reported from both the Pacific (Keigwin and Shackleton, 1980) as well as the Atlantic Ocean (Ruddiman et al., 1987).

The pine forests on the continent as well as *N. pachyderma* in the ocean apparently respond with a similar resolution to this cool-temperate environment. This is in contrast with the $\delta^{18}O$ values of carbonate in the deep-sea and the fluvial sediments on the continent, that show more variation within this period. Both clearly express a cyclic pattern. Differences with respect to the previous period are: higher peaks in $\delta^{18}O$ values and the significantly coarser fluvial sediments (transition from Unit 7/8 to Unit 8 (= Main Gravels)).

The end of the monotonous succession in the coiling ratio of *N. pachyderma* in the deep-sea is recorded in the period between the normals N1 and N2 within chron 5 (Fig. 2.7). Hereafter a cyclic alternation of warmer and cooler intervals started with a periodicity of roughly 125 ka finally culminating in a long warm period (Hooper and Weaver, 1987). A climate evolution that bears great similarity with the succession of the cool/temperate Susterian towards the warm Brunssumian.

The Mio/Pliocene boundary

The accepted age of the Mio-Pliocene boundary is 5.3 Ma (Berggren et al., 1985). This datum is well established at the type-section by the astronomically tuned age of 5.32 (Hilgen, 1991; Hilgen and Langereis, 1993) and in the southern North Sea by a Sr-isotope calibrated age of 5.4 Ma on foraminifers (Beets, 1992). A gradual, rather than a sudden, change from Late Miocene to Early Pliocene climate conditions is registered in the Atlantic ocean (Fig. 2.8); both by the planktonic/benthic foraminiferal ratio at DSDP site 609 (50°N) (Hooper and Weaver, 1987) and by sea-level fluctuations along the east coast of the United States (Krantz, 1991). The same holds for the pollen records in our study area (Fig. 2.7). This hampers pinpointing the correlative boundary between the pollen record and other records.

A solution to this correlation problem may be found in a marked cooling event that has been registered in the oceans within this transition period (Fig. 2.7). In the Atlantic ocean (site 609, 50°N), a southward flux of polar water masses have been registered by a temporary influx of *Neogloboquadrina atlantica*. This event falls between 5.6 and 5.8 Ma. This time slice is in agreement with the shift towards lower values in $\delta^{18}O$ in the Pacific ocean, (DSDP-site 846) (Shackleton et al., 1993). The latitudinal position of the Atlantic record is comparable with the position of our study area (51°N). So we may expect also some sort of expression of this global cooling event in the pollen records. A short and strong increase in the abundance of grasses at the expense of tree species is registered in several pollen diagrams around the Susterian-Brunssumian boundary (Fig. 2.6B, sequence 20). It is suggested here that such an increase of grasses (a light-dependent plant) may be a response to a temporary opening of the canopy in the lowland forest, triggered by the cooling of the ocean-surface-water at this latitude. Occasionally, *Chaenopodaceae* and *Betula* also peak in a slightly younger part of the *Gramineae* maximum (pollen diagrams Susteren and Koningbosch in: Zagwijn, 1961). Both genera are indicative for disturbance and re-colonisation of open patches.

The Pliocene

The Brunssumian

The Pliocene is divided into the Brunssumian and the Reuverian on the basis of the pollen assemblages (Figs. 2.2 and 2.6). The lower part, the Brunssumian, represents a temperate to warm and moist climate with two climatic optima (Brunssum A and C) (Zagwijn, 1960, 1990) separated by a cooler interval: the Brunssumian B.

The boundary between the Brunssumian and the subsequent cooler Reuverian is marked by a short but dramatic reduction in Tertiary floral elements (Fig. 2.6B; drillhole Susteren in: Zagwijn, 1960). Suc and Zagwijn (1983) and Zagwijn (1986) proposed that the Brunssumian/Reuverian couplet correlates with the Zanclean/Plaisancian division in the Mediterranean. This boundary almost synchronizes the Gilbert/Gauss palaeomagnetic boundary. The latter age is astronomically tuned and established at 3.58 Ma (Langereis and Hilgen, 1991).

However a fundamental change in the environment in the North Atlantic domain occurred later. A number of indicators point to major changes following the Mid-Pliocene cooling event near 3.3 Ma (within the Mammoth subchron of the Gauss) (see for a summary: Ruddiman et al., 1987). This event marks a major change to cooler conditions of the climate system (Keigwin, 1987; Bertoldi et al. 1989).

We regard it more consistent with the general line in the pollen zonation to take this Mid-Pliocene cooling event as the time boundary between the Brunssumian and the Reuverian, rather than the Zanclean/ Plaisancian boundary. From this follows that the Brunssumian lasted about 2.3 Ma (5.6 Ma - 3.3 Ma). Correlation of the pollenzones A, B and C within the Brunssumian with the $\delta^{18}O$ pattern from the deep-sea remains problematic due to the lack of additional markers. The variability expressed by the pollen (Fig. 2.6b) is not yet fully understood.

It might be a signal close to that of the sedimentary cycles, a resolution in the order of several 100 ka (see below).

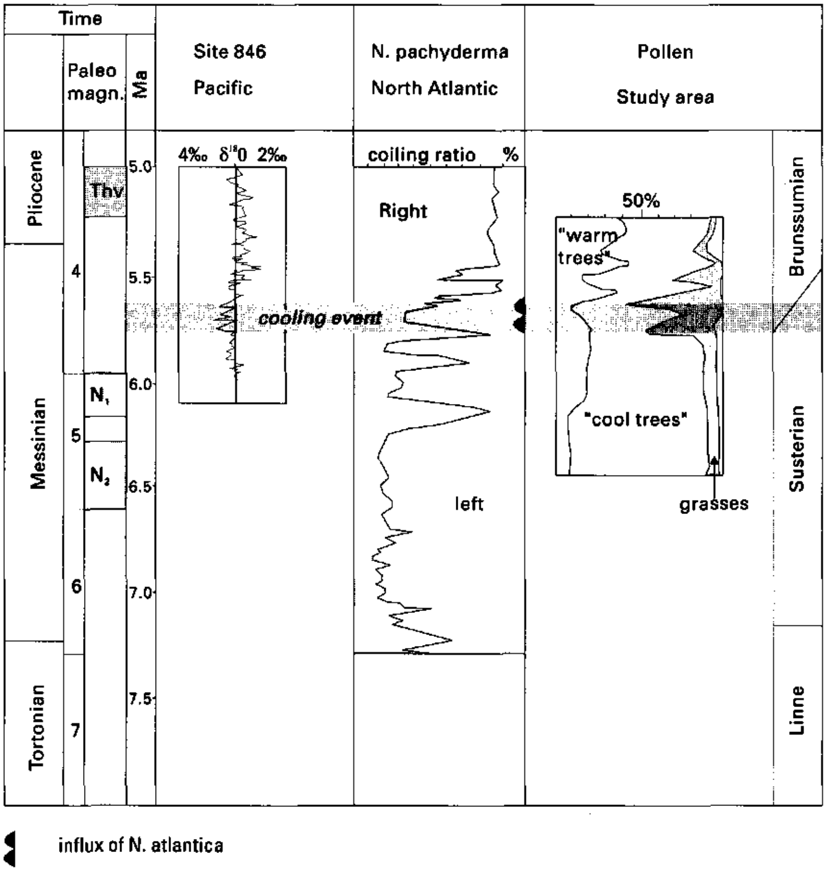


Figure 2.7 Pacific and Atlantic ocean records show around 5.7 Ma a cooling event by respectively increasing $\delta^{18}O$ values or fluxes of occurrence of a *Neogloboquadrina atlantica*. These phenomena are correlated with the sudden increase in grasses recorded in pollen diagrams. Foraminifera record from DSDP site 609 (Hooper and Weaver, 1987); $\delta^{18}O$ record of site 846 (Shackleton et al., 1993). Pollen record after De Jong, 1982)

Correlation of the subdivision of the Brunssumian and deep-sea is retarded by: (1) the incompleteness of the pollen record due to the fact that predominantly late-stage fine-grained tops of the fluvial sedimentary cycles provide abundant pollen spectra. The pollen signals related to the deposition of the lower parts of the cycles are definitely under-represented (Fig. 2.6). (2) From some of the fluvial sequences there are presumably no pollen data at all. Due to the lack of markers its not yet possible to correlate in detail the flood-basin deposits (which form the type section of the Brunssumian) with the sequences. (3) Identification of Brunssumian zones is often difficult when only a few pollen spectra are available (pers. comm. W.H. Zagwijn).

The Reuverian

The Reuverian is characterized by a rather variable pollen content and an overall reduction in 'warm' elements (Zagwijn, 1960). The Reuverian terminates with the onset of a series of major glacial phases: deep-sea stages 100 through 96, lasting from 2.5 to 2.4 Ma (Shackleton et al., 1990). They form together the Pré-Tiglian (Zagwijn, 1960). This climate interval has been regarded as the onset of the northern hemisphere glaciations and therefore has been taken as the beginning of the Quaternary (Zagwijn, 1960; 1985). Increasing evidence from deep-sea records (Loubere, 1988; Shackleton, 1993) suggests that the development of Pleistocene-type climate conditions could be regarded as a part of progressive climate deterioration. A similar conclusions follows from the Maas river terraces (see chapter 3).

The correlation of the Late Pliocene (= Reuverian) with the deep-sea signal seems to be more straightforward than that of the Early Pliocene part (= Brunssumian). The Reuverian is subdivided into A, B and C. The Reuver-A period comprises one major climatic cycle to the next major cooling. We suggest a correlation between the Reuver-A and the first 200 ka cycle after the Mid-Pliocene cooling event (from 3.3 Ma -3.15 Ma). This climate pattern is in agreement with oxygen isotope evidence where the next major cooling is found at 3.15 Ma (Shackleton et al., 1993).

The pollen show that the Reuver-B is composed of two climatic cycles with relative warm optima, the cooling event in the middle is not so extreme as the earlier ones. In the deep-sea records a next major cooling is found at 2.9 Ma (Keigwin, 1987; Shackleton and Hall, 1984; Backman and Pestieux, 1987). This suggests that the Reuver-B comprises 200 ka, composed of two 100 ka cycles.

The last Reuver interval, the Reuver-C, forms the run-down towards the glacial conditions around 2.5 Ma, another 400 ka cycle. The gradual increase of glacial maxima from deep-sea stages G(auss)-22 to stage 100 (Shackleton, 1993), as identified in deep sea-record of ODP site 846 suggests that the onset of the run down started already between 3.1 and 3.0 Ma. Such a long episode for the Reuver-C is not conventional because originally it was regarded to represent a very short period of time (Zagwijn, 1961). The frequent occurrence of cooling events as evidenced by the covariance between benthic and planktonic foraminifers (Loubere, 1988) during this period is in agreement with a long cooling episode.

The above proposed correlation suggests that the low-frequency variability (around 200-400 ka) of both records correlates well. All Reuverian sediments show a normal polarity (Van Montfrans, 1971; Boenigk et al., 1979). Their observation does not provide any additional constraints. It is outside the scope of this paper to go into further detail on this point.

Frequencies involved

We use the above established climate-based correlations to tie the pollen stratigraphy of the fluvial sedimentary record into a time frame. This resulted in a spacing in time-controlled sequence boundaries varying from 1 - 2.5 Ma. These levels are down-section used to fit the whole bounding-surfaces-record in the time frame (Fig 2.8).

To examine the likelihood, if a significant proportion of sequences had been removed without leaving any obvious trace, we compared our record with the eustatic sea-level cycle chart of Haq et al. (1988) (Fig. 2.2). As far as our time control goes we found a fair match with two 3^e order sea-level cycles. Cycle TB3.1 represents a sea-level cycle with an extreme lowstand and TB3.4/5 represents a cycle with an extreme highstand. If sequence-erosion

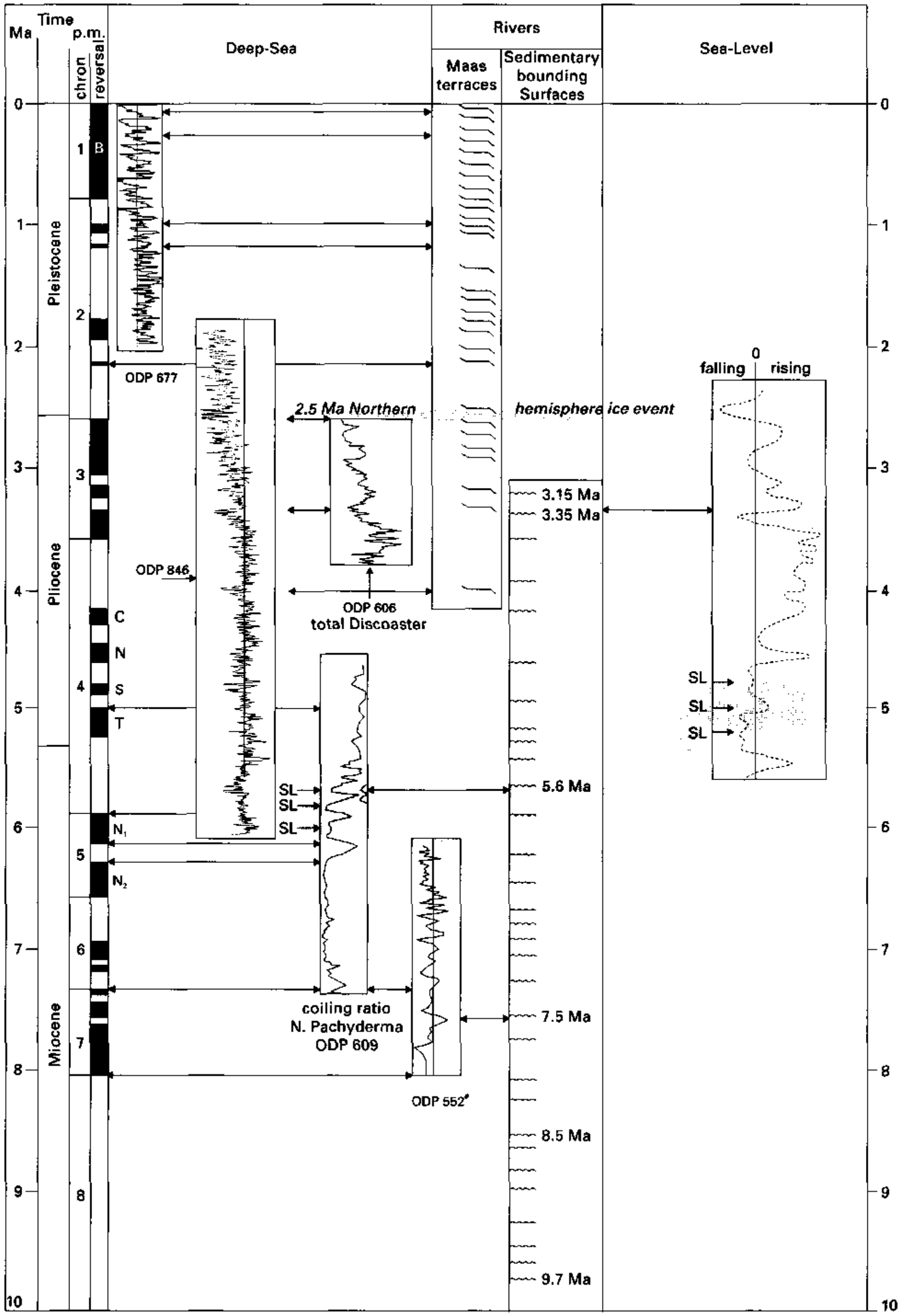


Figure 2.8 Comparison of the ocean record with the land record for the Late Neogene. Convention for interpretation of deep-sea records as climate-proxies, horizontal axis: cool to the left, warm to the right. SL = sea level events. For the age model of the terraces see chapter 3. Sources are mentioned in the text.

time period	number of sequences	average duration
3.15 -5.6 Ma	10	245 ka
5.6 -7.5 Ma	10	190 ka
7.5 -8.5 Ma	5	200 ka
8.5-9.7 Ma	6	200 ka

Table 2.1 Calculation of the sedimentary-cycle duration

played a significant role, erosion and bypassing of sediments and thus a final lower frequency, is expected during the "lowstand" cycle. The reverse is the case, that is the frequency (see below) is lower (245 ka) during the "highstand" cycle than during the "lowstand" cycle (200 ka). This suggests no major hiatus at third-order level (Ma scale). The distribution of the sequences within the time-constrained levels shows a pattern with a fair correspondences to the cyclic patterns in the deep-sea derived climate proxies (Fig. 2.8). This observation too points at internal consistency and completeness of the distribution of sequences at fourth-order level.

A first approximation of the cycle duration involved, follows from a simple division of the number of sequences over the time intervals (see Table 2.1). We took the maximum number of depositional sequences between the correlative levels to calculate the frequency. This means that for the Tortonian we recognised 11 depositional sequences within the southernmost tectonic domain (Fig. 2.5) against 10 within the northern domain. For the younger record we used the section from figure 2.6B transferred into a 'master boring' because it is positioned in the most subsided part of the graben (north-eastern part).

The outcome of these calculations points at an average cycle duration centred around the 200 ka. The uniformity in the average durations may suggest that the number of sequences is not seriously affected by erosion or by-passing. These processes may well have affected the internal architecture of a sequence (the fines/coarse-sands ratio). The longer average duration found for the period from 3.15-5.6 Ma might point to a few missing sequences. On the other hand, this is an era with long periods of general global warmth (Shackleton, 1993) and dominance of sea-level highstands (Fig. 2.2). Both factors cause stability in the fluvial system, rather than the formation of reactivation phases, so we believe that a longer duration of cycles might be expected for this period.

North of the Peel boundary fault line (Fig. 2.1) a deeper part of the rift (Venlo graben) contains a few fluvial-channel sequences of Upper Miocene age and overbank fines of Lower Pliocene age. This also might point to a period of by-passing of sediments along the study area. This is not likely as the mineralogy of the sediments indicates a local provenance rather than a Rhine or Maas origin (pers. comm. A.W. Burger, 1995). We envisage a permanent presence of both Rhine and Maas within the, tectonically constrained, study area leading to a "continuous" fluvial-sedimentation record for the period discussed.

Figure 2.8 shows that there is a wide range (between 100Ka and 400Ka) in the duration of the fluvial cycles. There appears a covariance in cycle duration both with the overlapping part of the terrace record and with deep-sea records. This suggests that causes of low frequency shifts in $\delta^{18}O$ concentrations are important in finding correlations between the land record and the ocean record.

Conclusion

Climate change knows several derivatives relevant to river dynamics: (1) sea-level change, (2) changes in vegetation type and (3) variations in precipitation.

Ad 1: Palaeo-geographic studies on the Lower Rhine Embayment (Zagwijn, 1975, Zagwijn and Hager, 1987) show that the lowland setting of the subject fluvial deposits and the lower end of stream profiles are particular affected by sea level changes (Suter et al., 1987; Schumm, 1993). Therefore in figure 2.8 a sea-level curve is included. This curve (by Krantz, 1991) is derived from high-resolution deep ocean $\delta^{18}O$ records and sedimentary records of the U.S. Atlantic coastal plain. Comparisons between paleo sea-level variations and the time-corresponding part of our record of fluvial bounding surfaces suggests a fair match. Detailed correlation is handicapped by the different age models used. The sea-level age-model has not been corrected for the new palaeomagnetic timescale as is done for the deep-sea records in figure 2.8.

Ad 2: Throughout the discussed period the drainage basins stayed forested under strongly fluctuating climatic conditions (pers. comm. W.H. Zagwijn). This argument is based on pollen records. These circumstances make it very unlikely that changes in vegetation cover acted as an upstream control on sediment delivery. So this factor may not have triggered the formation of bounding surfaces.

Ad 3: Discharge fluctuations related to periods with high magnitude floods can change river-system dynamics even under conditions of a continuous vegetation cover. The Pleistocene Colorado river in Texas (Blum, 1994) forms an example of a system that drained a continuous vegetated (grasslands) basin and produced comparable valley-fill sequences. In that case discharge fluctuations are related to latitudinal shifts of depression tracks in response to waxing and waning northern hemisphere ice sheets.

Both factor 1 and 3 point at the likely role of ice sheets.

The outcome of an average of 200 ka (with a range of 100-400 ka) as recurrence interval in formation of fluvial bounding surfaces, in conjunction with the correspondence with $\delta^{18}O$ patterns, strongly suggests a genesis controlled by a modulation of Milancovitch-type frequencies. This converging evidence requires an important role of the Late Tertiary Northern hemisphere ice sheets.

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3. A flight of 31 river-terraces of the Lower Maas, South-Eastern Netherlands: A continental record spanning 4 MA

Abstract - *The Maas river drainage basin has produced a sequence of 31 Plio-Pleistocene alluvial terraces, in the course of 4 Ma. This long continental record is formed in response to the interplay of tectonic uplift and climate change. Sediment fluxes match cold-stage marine isotope chronology in the 100 and 200 ka frequency band. Interglacial conditions are marked by an hiatus due to river incision and sediment-bypassing. The correlation is evidenced by data from palaeomagnetism, pollen, palaeosols and Tl- dating.*
The record highlights the role of the combined effect of the shielding effectivity of the vegetation cover and discharge variations (long-term presence or absence of passways of depression tracts). The record shows that northern-hemisphere fluvial systems of non-glaciated drainage basins can be a proxy of climatic change at the 100 ka timescale. Sequences of excessive sediment fluxes can be employed therefore as a dating tool in stratigraphic records in the realm of alternating interglacial/periglacial conditions.

Introduction

Rivers are continuously present in areas with humid climatic conditions. Their sediments and landforms, if preserved, are therefor long continental records. They reflect the combined effect of all sorts of drainage-basin processes, including climatic conditions, lasting on geological timescales. The resolution of this record only can be regarded as a running average of the processes involved. To appreciate the value of such long records, we need to know the sensitivity of river systems to environmental changes within known time-intervals. This condition can be met by comparing the sedimentary record with high resolution records of environmental change.

Upstream of the basin hinge line where one finds flights of river terraces, we may already exclude effects of cyclic glacio-eustatic sea level changes. Terraces are formed here under conditions of relative uplift (due to tectonic or thoroughgoing eustatic fall of sea level). They express alternating long-term conditions of accumulation or incision due to fluctuating hydrological regimes. In this paper we will focus on a terraced area in NW Europe at 51°N, shaped by the predominantly rain-fed river Maas (Meuse). The section is enclosed up-dip by the Ardennes Massif and down-dip by the southern principal displacement zone of the Roer valley rift system.

The geology and geomorphology of the area has been extensively mapped into great detail (Van Straaten, 1946; Zonneveld, 1949; Breuren, 1945; Felder and Bosch, 1984,1988,1995; Ruyters et al.,1995; Van den Berg, 1989). This large database provides a detailed overview on the distribution, lithology, sedimentology and morphology of the sediments. The terrace flight contains 31 levels, vertically separated 4-9 m apart (included a Weichselian Late Glacial level and the Holocene floodplain). It represents a considerable part of the Late Neogene and the Quaternary as evidenced by buried correlative sediments in the adjacent graben (see below). Age data on the alluvium of the terraces (superposition, palaeomagnetic, pollen, TL- and palaeo-sol data) are sparse but well distributed over a significant part of the flight. This suggests (in conjunction with the high resolution oxygen-isotope records

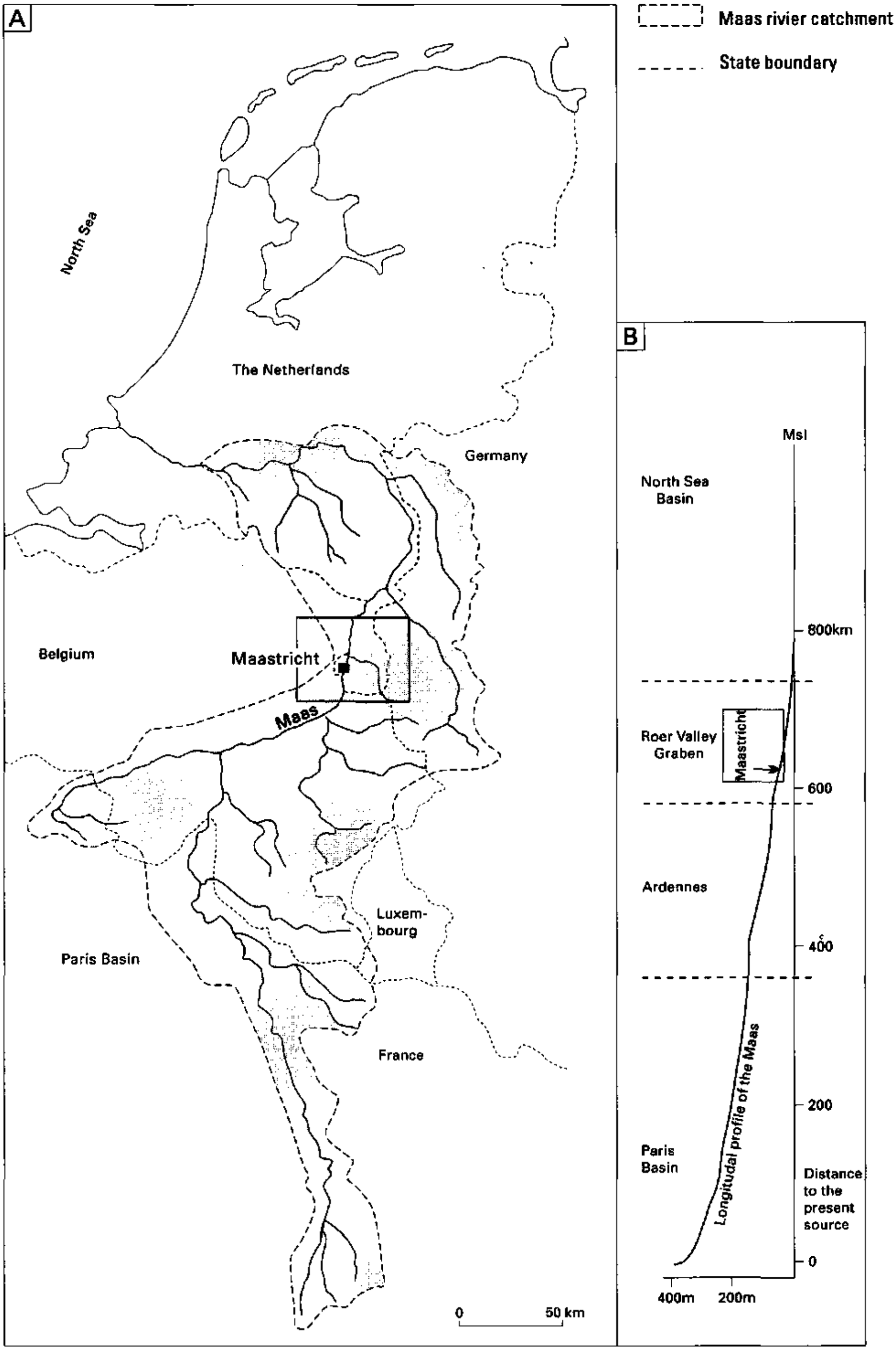


Figure 3.1 Setting of the study area within the catchment of the Maas river.

of this era) that since the second half of the Pleistocene the major cold intervals, occurring in a frequency band of 100-200 ka years, are represented by fluvial sediment bodies. This climate control on the rate of sediment production indicates that major changes in sediment production e.g. terrace formation in uplands and lateral continue sequence boundaries within their correlative sediment sequences, may be an important dating tool in basin stratigraphy. The utility of terraces to tectonic studies is obvious. Well dated series of steps can be used to deduce the regional long-term history of tectonic deformation. Such records show that, on timescales of hundred thousands of years, tectonic rates follow smooth low amplitude functions. Under the assumption that crustal dynamics change gradually, we found fluvial sequence stratigraphy in conjunction with uplift functions, to be an useful tool in the approximation of the age of non-dated terrace levels. Under these conditions we arrive at the conclusion that from the second half of the Pliocene onwards basically the same process-response relation may be applied to the interpretation of fluvial terraces.

Regional setting and stratigraphic significance

River Maas sediments have been supplied from a 33,000 km^2 wide rainfed catchment, covering the Ardennes and the eastern marge of the Paris basin. Before 250 ka the Vosges were also connected to the upper reaches of the Maas system, they were later captured by the Moselle River. (Zonneveld, 1949; Krook, 1993) (Fig. 3.1). In the study area we find remnants of the Maas river system as tracts of gravels and sands shaped into cut and fill terraces. They overly Cretaceous bedrock and Tertiary sands and are mantled by a loess sheet. The downstream gradient of the river sediments amounts about 0.75 m/km. The thickness of the sediments varies from a few meters to over 25 m. These units can be traced upstream into a narrow zone crosscutting the Ardennes (Macar,1954; Juvigné et Renard, 1992). Downstream, within the Roer valley rift system (= southeastern most tip of the North Sea basin), the main part of the flight looses its expression as terraced surfaces. The sediments here partly interfinger with Rhine sediments. Post stage 15 Maas sediments are spread side by side on the tilted Peel high within this rift (Figs. 3.2 and 3.3). This downdip morphological change is important for the interpretation of tectonics being the main driving force for the uplift of the region with the terrace record and not a sealevel fall. Otherwise more close to the shoreline a similar terraced landscape would be expected.

By its position just upstream of the basin hinge line, the study area takes a key position within the study of the history of sediment supply to the southern North Sea basin.

Positive evidence of an early existence of the lower Maas in the south-eastern Netherlands is provided by the presence of *Classopollis* pollen. These pollen have been reworked from Jurassic rocks in the NE margin of the Paris Basin (Zagwijn, 1961) They have been recognised in deposits that are ranked as the Linne pollen-stage (de Jong, 1982). The earliest occurrences can be dated about 8.5 Ma (see figure 2.6B). The regional infill of the Lower Rhine Embayment (Zagwijn and Hager, 1987; chapter 2) suggests that the system may be at least several million years older.

Tectonic controls

The main tectonic factor controlling the distribution of the terraces is the interplay of the compressional dynamics of the Ardennes massif and the extensional dynamics of the Roer valley rift system. We distinguish three zones (Fig. 3.3):

1. A zone running SW-NE parallel to the frontal thrust of the Ardennes. This zone contains in the Netherlands a palaeo-valley called the "East Maas". Towards the SW, downstream the town of Liège (Belgium), today's Maas river still occupies this zone. This major drainage pathway can be followed northeastwards into Germany (Boenigk, 1978). The northern margin of this palaeo-valley is formed in response to Ardennes-foreland compression by Neogene reactivation of the Hercynian Waubach anticline (Fig. 7.2).
2. The East Maas valley became abandoned in the course of the Early Pleistocene (Kuyl, 1980). Younger sediments were dispersed in a fan shape northwest of the Waubach anticline, together forming the West Maas area. The distribution of the terrace remnants here, shows that the Maas river slipped off the north flank of the Waubach anticline and progressively migrated northwestwards through out the Pleistocene. This resulted in the

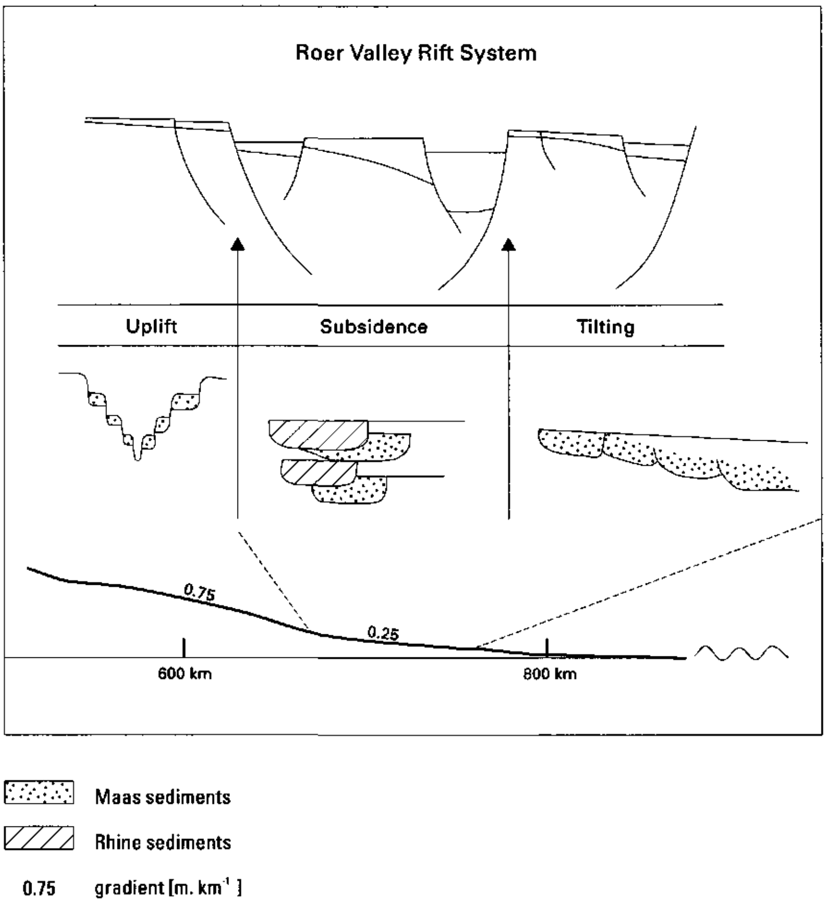


Figure 3.2 The Roer valley rift system showing various modes in tectonic style that are reflected in the morpho-sedimentological expression of the lower Maas.

preservation of terrace remnants primarily on the right bank of the valley. Exceptional is the Pietersberg-2 phase. During this phase the river migrated far to the west, probably in response to tilting of the south-western blocks of the rift.

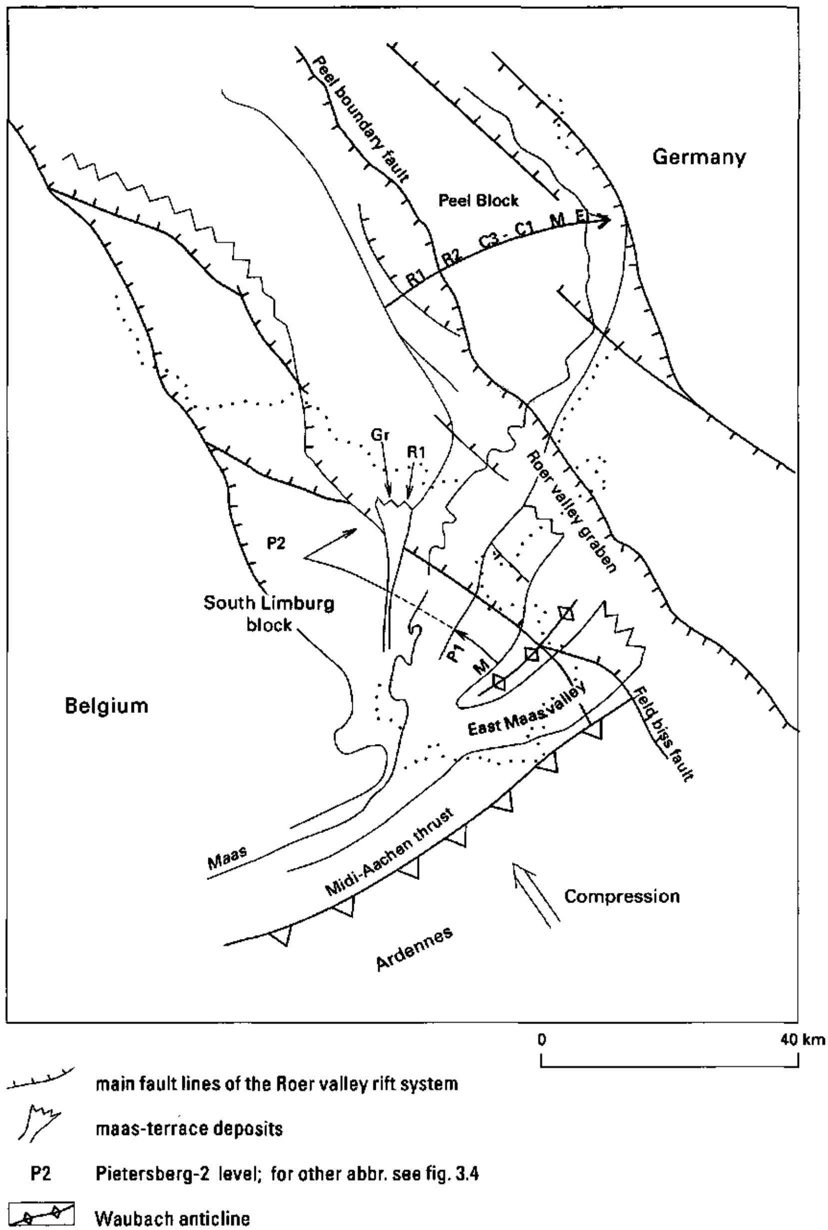


Figure 3.3 Main morpho-tectonic elements of the Lower Maas "fan", as mentioned in the text. The reactivated Hercynian Waubach anticline separates the older East Maas valley from the Pleistocene West Maas valley. Arrows indicate migration routes of the channel axis.

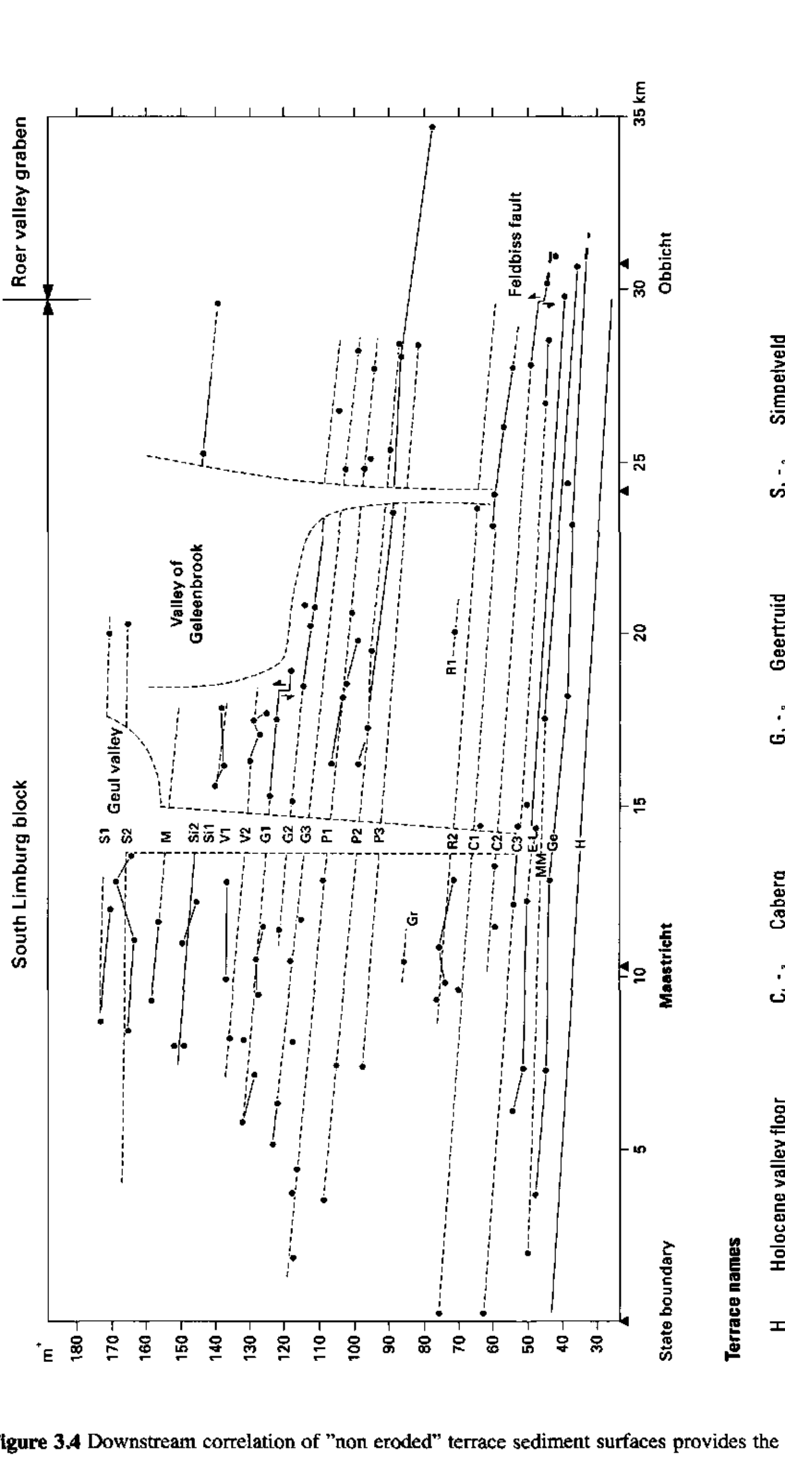


Figure 3.4 Downstream correlation of "non eroded" terrace sediment surfaces provides the reconstructed gradients of the West Maas terraces and the lower two East Maas terraces (levels S1 and S2).

3. During the Brunhes Chron (0.78 Ma - present) a "gorge"-like valley developed within the West Maas area. This morphology characterizes another phase in the tectonic history. The "fan-apex" migrated downstream towards the Feldbiss fault, the southern principal displacement zone of the adjacent graben. From two successive terraces stages (Gravenvoeren and Rothem-1) within this phase only very small patches have been preserved in the course of the incision (levels Gr. (= Gravenvoeren) and R-1 (= Rothem-1) in Figs. 3.3, 3.4). Such a restricted preservation may be indicative for the existence of tectonic sub-phases within this period. This episode was synchronously with the onset of a major reorganisation of the Maas/ Rhine system within the rift system. This will be discussed in chapter 7 (Figs. 7.7 and 7.8).

At local scale the drainage network reflects a joint system with a characteristic pattern that fits with the above mentioned deformation directions (Fig. 8.7)

Geomorphological analysis

Morphology

The area studied is roughly triangular shaped with an apex near the Dutch-Belgium border where the river debouches from the Ardennes. It measures 30 km long and 80 km wide. There is a general tilt from ± 250 m in the SE to ± 40 m in the NW concordant with the general dip. The overall flow direction of the Maas River is NNE.

Due to periglacial erosion only very few individual terrace levels still form a connected tract. The majority of the terraces cap interfluves and plateaus and flat-topped remnants become progressively smaller with increasing age.

The original terrace morphology is even more camouflaged by a cover of periglacial loess and associated colluvium.

Mapping

We used an extensive database of soil-, geomorphological- geological and topographical maps to identify the outlines of individual sediment spreads; in particular the combination of an elevation map with altitudinal information at 1:10.000 scale, and a regular net of drillholes (10 holes per km^2) reaching into the underlying fluvial sediments. Besides the shallow drillholes a less dense grid reaches the base of the fluvial sediments.

The upper and lower bounding surface of a terrace sediment unit are quite different in nature. The base is strongly undulating with height differences up to several meters. The tops of the alluvium vary in general only one meter or so. The tops of the sediments are therefore regarded a less variable and much better constrained marker of a particular terrace fragment than the base of the sediments.

Drillhole-profiling showed that the loess-surfaces in general conforms the surface of the underlying river sediments. This loess blanket could not be used to date the underlying alluvial terraces as has been done in other areas (e.g. Antoine, 1994; Porter et al., 1992).

The elevation map, issued by the Topographical Survey of the Netherlands, is based on third-order levelling data points arranged in a regular grid of 100 m x 100 m. The altitude information is given with 0.1 m accuracy. To find the outlines of the individual terrace remnants, we contoured the loess-surface with 0.5 m intervals. For altitudinal information of the fluvial sediments we relied on drillhole data only. Drillhole information also helped to discriminate buried river terrace scarps from other terrain irregularities. We scrutinised all interfluves on sub-horizontal parts which may reflect buried fluvial terraces. By this we found a regular vertical spacing for the West Maas valley levels with steps of about eight metre. The separation of the East Maas terraces levels is more variable (between 4 and 9 m).

The top of the alluvium varies in general less than a metre in central parts of the flat-lying terrains of a particular level. Palaeo-channels within the tops of the terraces are under-represented, probably because many of them developed in a later stage into parts of the secondary drainage system, others may be degradated by being filled with late-stage Hochflutlehm or with products of surface denudation due to periglacial levelling and areal down-cutting of the sediment-tops. Based on the lithological composition of bars of the last glacial river we assume that only the sand-rich upper metre of these bars is prone to periglacial denudation. The underlying gravelly sands most likely prohibited deeper substantial areal surface erosion while a pavement soon will develop as a lag-concentrate of the, winnowed, gravel-rich fluvial deposits. We regard the inaccuracy in the determination of the average height of a particular terrace fragment, introduced by this method to be acceptable in the light of the 8 m steps between the various levels.

Number of terrace steps

To be able to arrive at the best possible correlation of our terrace record, with its limited dating possibilities on sparse terrace fragments, with other records it is important to tie up these fragments into levels. To this we constructed long profiles of altitudes of river terrace-tops. This allowed (1): a consistent downstream correlation between fragments and (2): provided us with the reconstruction of the original number of steps in the total terrace-flight.

ad. (1): We selected on the topographical map flat terrain patches. In case of small patches (less than 2 km^2), the highest top of the underlying fluvial sediments was chosen from all available drillhole data. Some buried surfaces can be followed over several kilometres. In such cases we took several height values in downstream direction. These data were plotted in a graph showing altitude versus distance (Fig. 3.4). These lines show the average terrace-gradient. The shortest distance to the apex of the fan was taken as a measure for their position on the horizontal axis in the diagram.

It strikes in figure 3.4 that the gradients of relative long continuous levels of the drawn lines are about parallel to the one of the Holocene Maas (H in figure 3.4). We use this parallel phenomena to attribute certain isolated remnants to a level. This is indicated by the dashed lines. This procedure worked well for the West Maas River terraces as well as for the lower two levels of the East Maas valley, because these remnants are distributed over a sufficient long distance.

A number of complicating factors did not allow to follow this procedure for the remaining higher levels in the East Maas valley. (1st) The remaining remnants of the East Maas River terrace flight are too sparse along this palaeovalley or their tops were so heavily eroded that no flat surface was left over. (2nd) This poor preservation was even more complicated by the possibility of large scale karst solution of the chalk underlying the fluvial deposits of the older levels higher up on the interfluves. (3rd) Only fifteen kilometres of this palaeovalley is undisturbed by crosscutting faulting. Long distance correlation was therefor uncertain due to the presence of many crosscutting fault lines which coincide with valleys. In the correlation of fragments across such valleys it is hardly possible to distinguish vertical displacement of a level from originally separated levels. We restricted our survey to the non-faulted transect.

ad. (2): For the West Maas valley we found 21 steps. For the East Maas valley we found 10 levels. This latter number is composed by the correlation of three sections along nearby interfluves (Fig. 3.5). 31 levels in the whole flight exceeds earlier work by Breuren (1945) who suggested 13 levels in this area. Systematically mapping allowed Felder and Bosch (1988) to extend this number to 20. The difference is not only explained by the knowledge increase over the years, but also by the method used. As discussed above, our number is based on the terrace tops whereas the other studies relied on the base of the sediments.
To name the levels we followed as much as possible the names introduced by these authors and extended an original name with a suffix 1, 2, or 3 when a level was further subdivided. When all mapped terrace fragments from the sides of the 30 km long section of the lower West Maas valley are assigned to a level, a very regularly spaced flight shows up. This may suggest that they developed as the result of regularly occurring large-scale environmental changes.

Sediment characteristics.

Many temporary small-scale quarries in the various levels existed over the years. At present only a few large-scale exposures can be studied. From older descriptions and current exposures, gravels and coarse sands dominate. Clay-beds are rare. This resembles the Scott-type system such as defined by Miall (1978). From two levels more sandy facies are known. (1) The Winterslag Sands, occupying the western margin of the Campine Plateau (= Pietersberg 2 level) in Belgium (Gullentops, 1974). (2) The Geertruid 2 terrace contains a gravel rich facies on top of a sand rich facies. Each about 12 m thick and with cold-climate indicators. We assume that the currently accessible exposures are representative for all of the gravel-rich levels on the base of the parallelism of the sediment-surface slopes and the knowledge gained from the numerous drillholes that recovered full terrace sequences. This makes it less probable that important contrasts in sedimentation are hidden within the non-exposed sediment series.

Stacked units form the most conspicuous element in the internal architecture of the terrace sediments. These units are composed of several truncated braided-channel fills. Lag concentrates, characterised by an abundance of boulders mark the unit boundaries (Fig. 3.6). These lags are directly overlain by a poorly sorted clast-rich package. The latter grades into a stacked series of laterally accreted bars. Towards the top of each unit, the distal parts of these bars frequently merge into sand wedges or rarely clay bands. The finer top series is again overlain by a boulder-rich bed. The stacking indicates net aggradation. The boulder-rich beds are wavy, but overall horizontal; they show a floodplain-wide persistency. This characteristic will be discussed in chapter 5.

The dominance of gravel lithofacies throughout the lower Maas valley indicates that they were deposited by a coarse bedload-dominated river with a braided morphology. This morphology follows from: (1) the absence of fine-grained lateral accretion pointbar sediments, as would be found in single-channel systems. (2) the sheet-like geometry of the

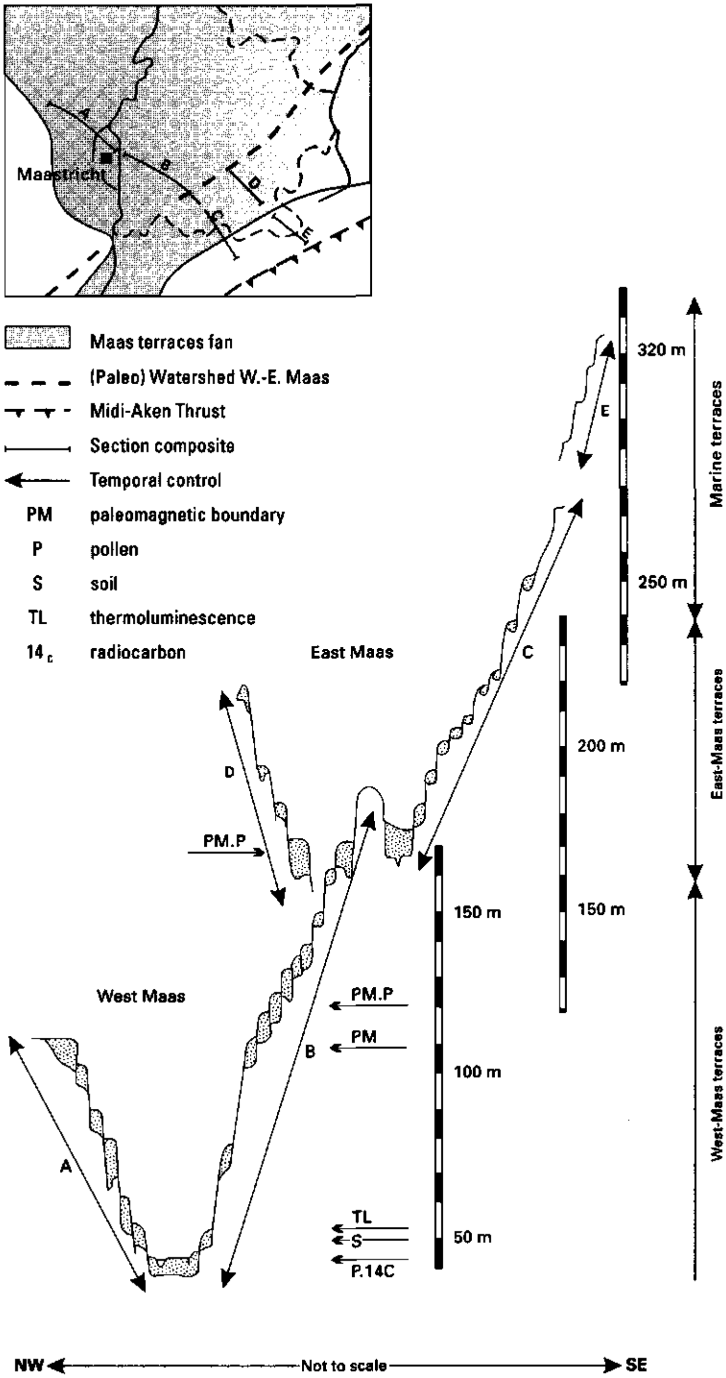


Figure 3.5 Schematic composite cross section through the upper part of the Maas terraces fan (some levels have not been indicated due to the choice of the sections).

gravel/sand units, suggesting a shallow and wide channel plain. In favourable exposures low-angle dipping accretionary surfaces have been observed. Such an internal bar-architecture fits well with those of braid-bars.

A number of associated features points towards cold-climate conditions during the aggradation; e.g. angular blocks of loose sands, obviously transported in a frozen state; syn-sedimentary ice-wedge casts and involutions (Krook, 1963; Vandenberghe, 1985; 1993) (such ice-wedge casts have been also observed at the contact between the underlying bedrock and the fluvial gravels in the quarry Obgrimbie (Pietersberg-2 level); metre-size boulders transported by ice-floes; and rarely, remains of cold-climate fauna are preserved (Van Kolfschoten, 1985). These features have been recognised individually or in combination, within the units of the flight from the Kosberg-levels downwards. Indications for sedimentation of river terraces in a permafrost environment are also reported from other NW European rivers like the Rhine in Germany (Brunnacker, et al., 1982) the river Thames in England, (Bryant, 1983); the rivers Allier (Veldkamp, 1991) and Somme (Antoine, 1994) in France, the Ohre river in Czechoslovakia (Tyracek, 1983).

Within the Lower Maas terraces, major but step-by-step shifts in the sediment composition occur both in the gravel composition (Van Straaten, 1946) as well as in the heavy-mineral and in the bulk-geochemical sand composition (Zonneveld, 1949; Krook, 1993; Riezebos, 1971; Riezebos et al., 1978; Moura and Kroonenberg,1990; Slomp,1990). An important and very characteristic change in the gravel composition occurs around the Kosberg levels in the East Maas valley; older terraces (the Waubach group) are quartz-rich, and dominated by (ultra-)stable minerals. Younger levels are progressively enriched in fresh rock components indicative for strong periglacial weathering in the drainage basin. The properties of the levels of the Kosberg-Crapoel group take an intermediate position between those of Miocene sediments and those of Pleistocene sediments. Their mineralogy

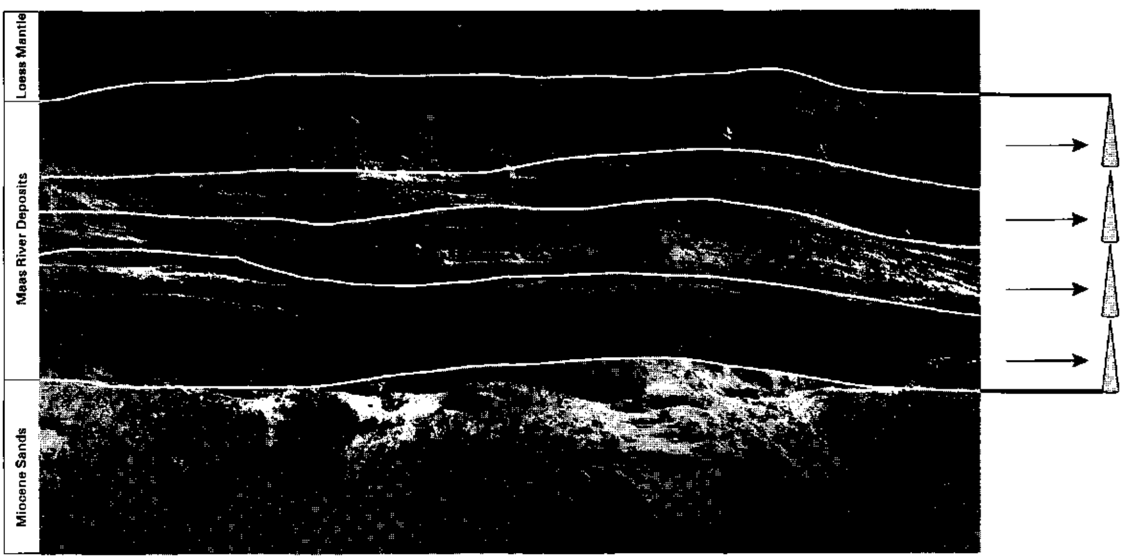


Figure 3.6 Part of the Geertruid -2 terrace sequence showing the braided stream alluvium (Scott-type) with typical planar erosion surfaces delineating (four) reactivation phases. The height of the gravelly sands approximately amounts 12 m. The alluvium is overlain by a loess mantle and underlain by Miocene marine sands.

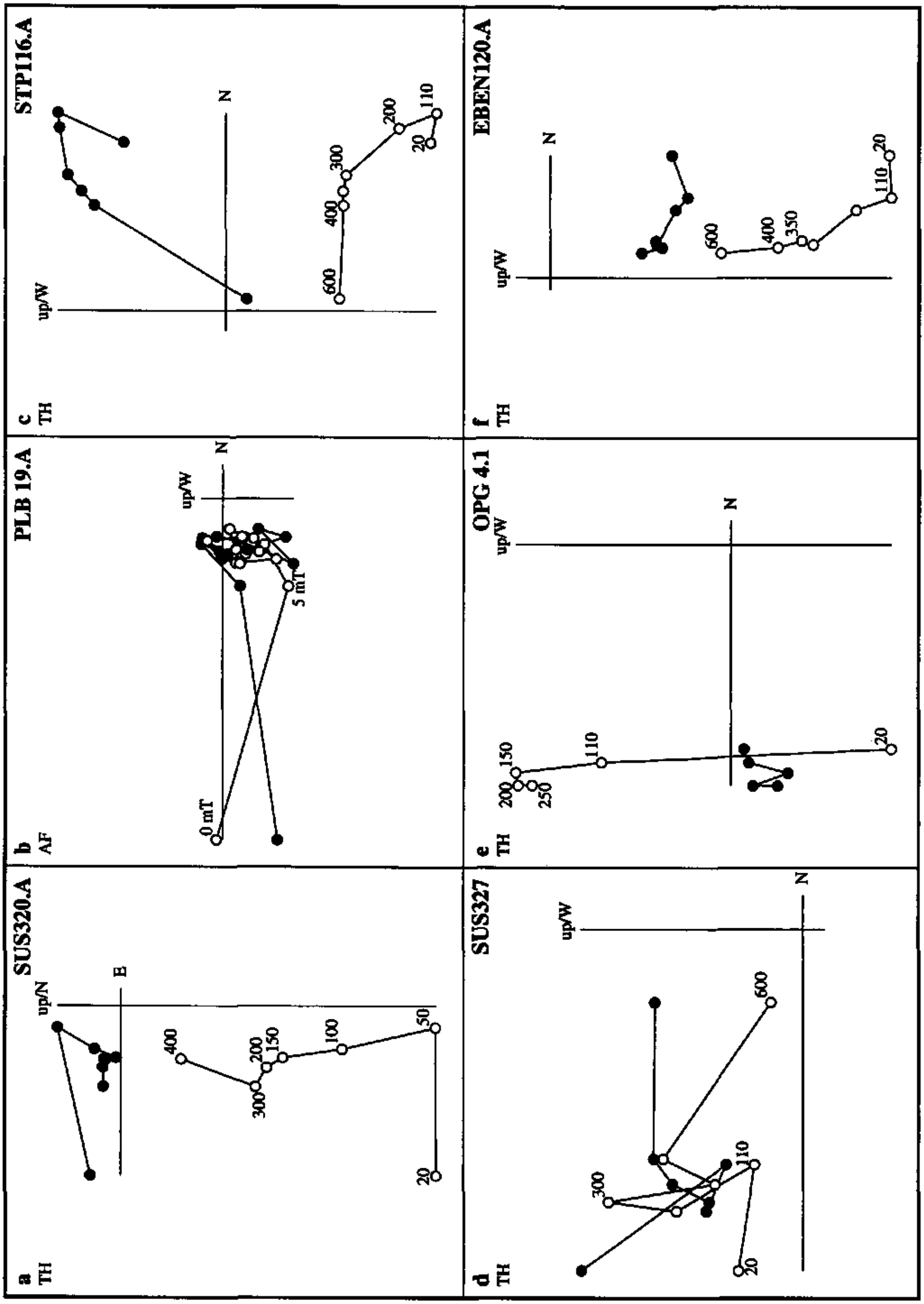


Figure 3.7 Examples of thermal (th) and alternating field (Af) demagnetisation diagrams of various terrace levels (SUS = Susterzeel quarry, Geertruid 2 level; PLB = Platte Bossen quarry, Simpelveld 2 level; STP = ENCI quarry in Pietersberg 3 level; OPG = Obgrimbie quarry, Pietersberg 2 level; EBEN = Eben Emael quarry in Pietersberg 2 level).

is closely related to the Tertiary sediments (cf. Zonneveld, 1949), their average grainsize is coarser and their quartz/rest-group ratio is more alike the Pleistocene terraces.

Terrace chronology

West Maas

The terraces are considered from their sediment characteristics to have been built during cool or cold climatic conditions. In particularly the last net aggradation period is well constrained in time to the last glacial by the combined evidence of pollen, sedimentology, morphology and ^{14}C dates; net aggradation occurred between the onset of deep sea stage 5-d through the end of stage 2 (a.i. between about 110 ka and 16 ka) (see chapter 5). So the stage 4 and 3 river phases are included into one prominent morpho-stratigraphic unit: the Mechelen aan de Maas (MM) terrace.

The present interglacial conditions show an incised and shrunken floodplain with respect to the floodplain formed during the previous period of net aggradation. The Geistingen terrace (level Ge in Fig. 3.4) forms one of a set of erosion terraces related to this incision (see chapter 6). It is also evident from the rate of lateral channel migration of the Holocene system, that significant widening of the floodplain in this stretch of the system is not likely to occur during interglacials. This geomorphological setting clearly shows that the preservation potential of interglacial fluvial sediments in this part of the system is very limited, this a.o. evidenced by the absence of positive identified fluvial deposits of Eemian age (the equivalent of deep sea stage 5^e). This interglacial has been only recognised as a palaeo-sol capping the next higher terrace (Eisden - Lanklaar terrace). The following higher terrace (Caberg-3 level) is also capped by interglacial deposits dating from the Belvédère interglacial (Kolfschoten et al., 1993; Vandenberghe,1993). Thermoluminescence dating of associated burned artefacts revealed an age of 250 +/- 20 ka (Huxtable, 1993). This age fits perfectly well with deep sea stage 7^c. The morpho-stratigraphic record of the last three terraces shows that for the long-term fluvial-landscape evolution of the last 300 ka we may apply the classic assumption that aggradation matches cold-climatic conditions and erosion and by-passing occurs during interglacials. Subtle refinements are possible in the sequence of events leading to the end forms (Vandenberghe, 1993) but they do not alter the straight forward response of the fluvial system to the climatic evolution at the longer time scale.

This conclusion forms the base for a further assessment of the chronology of our long record. To support the age assignments of the older levels in the flight, we made use of the combined evidence provided by superposition, pollen, palaeomagnetic determinations, in combination with arguments derived from lithological composition and palaeo-sols. These data were applied in age-altitude diagrams. We took for a particular terrace the end of the aggradation period (= the cold/warm shift) as a measure for its position on the time axis in the diagrams. The astronomical calibrated version of ODP 677 record (Shackleton et al 1990) provides a linear timescale.

Pollen

Over the last 50 years several researchers have made attempts to extract pollen from exposed clay layers. These clays rarely contained pollen (pers. comm. Prof. A. Brouwer;).

Even the fauna-rich interglacial deposits at the top of the Caberg-3 terrace (Belvédère interglacial) did not preserve any pollen.
Two exceptions to this rule form palaeo-channels of the Early Pleistocene levels of Geertruid-2 terrace and the Simpelveld-1 terrace.

Geertruid-2 terrace:
In a brick-yard quarry near Susterseel (at present just across the border with Germany), at the downthrow side of the Feldbiss fault, Maas sediments are underlying Rhine sediments (= Weert heavy-mineral zone). Between these sediments is an 8 m thick succession of clay and loess. This unit contains at least four palaeosols as indicated by horizons with clay-illuviation (Bruins, 1981). The lowermost palaeo-sol is a pollen bearing, humic-clay band of about 0.60-0.80 m. The rather characteristic pollenassemblage shows the presence of the Bavel 3-5 pollenzones (Zagwijn and de Jong, 1983). These zones form the upper part of the Bavel interglacial. This interglacial presumably forms the continental equivalent of deep sea stage 31. This correlation is corroborated by a polarity sequence (Van Montfrans, 1971; Bruins, 1981; this paper).

Simpelveld-1 terrace:
Gravels of the Simpelveld-1 terrace are overlain by about seven metres of a loam (in its lower part sandy-loam). Between the sandy-loam and the loam at about 172 m* (4 m below surface), an 1.50 m thick horizon with pollen-bearing organic-rich clays is intercalated. Its pollen assemblage represents a cool-temperate forest phase of Tiglien age (De Jong and Zagwijn, 1963; Kuyl, 1980).The lithological succession suggests that the organic layer forms the top of a fining-upwards sequence starting with the underlying gravels. Sedimentary structures as well as the grain size of the upper loam point at a reworking of aeolian loss.

Palaeomagnetics

Magnetostratigraphy is based on the ability of the natural remanent magnetisation (NRM) of rocks to record the polarity of the geomagnetic field at the moment these rocks are formed. The flips from one polarity to the other, i.e. the reversals of the geomagnetic field in time are quite well known (cf. Shackleton et al, 1990; Hilgen 1991), consequently a geomagnetic polarity linear-time scale can be established. For magnetostratigraphic dating the polarity pattern of a sequence of rock strata is compared with the geomagnetic polarity time scale. It is therefore required that the sequence of rock strata is deposited with a continuous rate of sedimentation, like deep sea sediments. Fitting of the obtained polarity pattern in fluviatile sediments with the standard geomagnetic polarity time scale is practically unrealisable because (1) the sequences cover a too small a part of the geological time, (2) in fluviatile deposits hiatuses in sedimentation are, of course inevitable. (3) Only the lithologies with a high clay content are suitable for palaeomagnetic purposes while these occur only rarely, generally in very small lenses intercalated in the coarse grained, sandy deposits. However, the magnetic polarity of the NRM in the clay lenses can give confirmations of the age estimates determined by the correlations of the terraces with the cold stages in the ODP 677 $\delta^{18}O$ record (Shackleton et al., 1990) as depicted below in figure 3.8a. For this purpose we have used two sets of data: firstly we have re-interpreted the palaeomagnetic data by Bruins (1981), who studied several Maas terraces mentioned in his paper and secondly: an additional set of data collected after sampling and measuring in April 1991. All together 10 levels out of the 31 have been accessible.

Sampling was done by pushing cylinders into the - preferably fresh - clay. After orientation the cylinders were cut out of the outcrop. The remanence of the samples was measured with a cryogenic magnetometer. Bruins (1981) performed the palaeomagnetic measurements of the remanence during progressive thermal (TH) or alternating field (AF) demagnetisations. Before the AF demagnetisations Bruins sometimes heated the samples to a temperature between 110° and 170° C in order to get rid of the high water content present in the samples. Because the TH demagnetisations generally give better results than AF demagnetisations, we only performed TH demagnetisations, starting with 100°C and followed by steps of 50° C.

Results

About half of the palaeomagnetic directions from the 10 terrace levels show intermediate directions (cf. Figs. 3.7a and 3.7b). Intermediate directions of the geomagnetic field occur very rare in time and therefore intermediate palaeomagnetic directions from samples taken randomly in rocks should also be observed exceptionally. Hence, it is statistically very unlikely that all intermediate directions represent the geomagnetic field. The intermediate directions which are close to N or R (cf. Figures 3.7c, 3.7d) were accepted as N or R, respectively. Other intermediate magnetic directions were considered unreliable. N results are either secondary or primary. Discriminating between the two is difficult for two reasons: (1) in geological young rocks the primary N direction is very close to the present day field direction; (2) in sediments more commonly used for magnetostratigraphic studies, secondary directions are easily detected, because they are generally bound to hematite and goethite. Since hematite and goethite in fluviatile sediments may also be detrital, these minerals can carry secondary as well as primary directions. These considerations make that the observed N polarities should be correlated to the geomagnetic time scale with great caution. The comparison between the expected polarities and the observed polarities is presented in stratigraphical order in Table 3.1. With exception of the level where only R polarities are expected, Pietersberg 1 and 2, the observed N polarities are not in conflict with the correlation. R directions are regarded as most reliable. Only two levels display R polarities (Table 3.1): Pietersberg 2 and Geertruid 2.

According to the correlation all demagnetizations from the Pietersberg 2 level should be R. The Obgrimby section has one sample which has a tendency to R direction (Fig. 3.7e), while the samples from the Eben-Emaël section have N directions (cf. Fig. 3.7f).

The Geertruid 2 level has been sampled at various locations (Table 3.1). The top has been sampled in the Süsterseel quarry displaying a R - N - R sequence.

The N part coincides with a layer of which the abundant pollen point at the presence of the Bavel interglacial (as discussed above). The observed N polarity within the Bavel interglacial is in line with previous results by Van Montfrans (1977) and Kasse (1988) in Rhine sediments. We interpret the R to N sequence as lower Jaramillo boundary followed by a hiatus.

The latter is expressed by a major change in lithology. The subsequent löss layer bears the upper R polarity.

The basal, sand-rich, part of Geertruid 2 showed mainly R directions and a few N directions. May be these N directions could represent the Cobb Mountain event.

location	Maastricht section	expected chronozone (polarity)	observed polarity
Amby (B)	Rothem 2	Brunhes (N)	N
ENCI (B)	Pietersberg 3	Brunhes (N) / Matuyama (R)	N
Eben Emaël (B); Obgrimbie (A)	Pietersberg 2	Matuyama (R)	R and N
Curf (A); Kruisberg (B)	Pietersberg 1	Matuyama (R)	N
Spaubeek (A); Martens (A); Nagelbeek (A; B); Susterzeel (B)	Geertruid 2	Matuyama (R)/Cobb Mountain (N)	R and N
Rode put (A; B)	Simpelveld 2	Matuyama (R)/Reunion 1 (N)	N
Reijmerstok (B); Platte bossen (B)	Simpelveld 1	Matuyama (R)/Reunion 2 (N)	N
Scheiberger bos (B); Crapoel (B); Imstenrade (B)	Crapoel	Gauss (N)	N
Banholt (B)	Kosberg 3	Gauss (N)	N
Landsrade (B)	Kosberg 2	Gauss (N)	N

Table 3.1: Locations of sampled Maas terraces; Maastricht section, expected paleomagnetic chronozone and polarity based on the ODP 677 $\delta^{18}O$ (Shackleton et al., 1991) correlation, and the polarity observed after demagnetization. (A) denotes results from this study; (B): results from Bruins (1981).

In summary: The limited value of magnetostratigraphy in this type of sediments is caused by the difficulties in discriminating primary from secondary magnetizations. Based on the polarity analysis only, the results of most of the terraces with a N polarity remain tentative. The observed reversals in the Pietersberg-2 and Geertruid-2 form the most important result. While we expected a R polarity from the morpho-stratigraphic position of these levels within the whole flight. This expectation was based on the process-response relationship between the occurrence of terraces and cold stages in an era which climatic evolution is forced by the 100 ka periodicity (Ruddiman, 1989). These observations in conjunction with pollen data, do confirm this anticipation and provide a further basis for the direct correlation of terraces and cold stages extended now over the last 1 Ma.

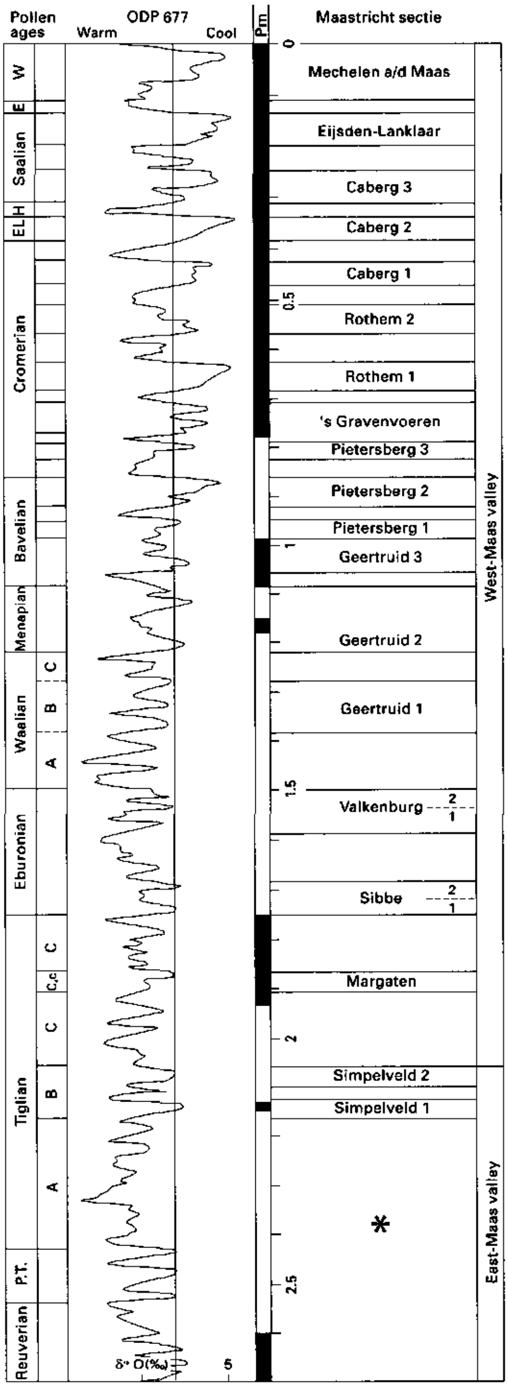
Fluvial sequence stratigraphy as a tool to infer terrace-age

A strong correlation between terrace formation and climatic change during the Brunhes magneto chron, as we showed evident from the West Maas terrace record, is also shown by Antoine (1994) for the river Somme and by Porter et al., (1992) for a non-glaciated basin in China. Our observations on the West Maas sections showed that such a correlation may be extended back in time through 1 Ma.

For the older part of the Matuyama chron we mapped within the West Maas valley another six terraces, regularly distributed over the altitudinal range. Their chrono-stratigraphic position is not directly obvious; somewhere between the Bavelian- and the Tiglian pollenzones. Their morpho-sedimentary properties do not differ from their younger equivalents. This suggests that again these terraces represent prominent cold episodes.
The climatic evolution of this period is primarily forced by the 40 ka signal (Ruddiman, 1989). This forcing has led to many more cold-warm oscillations than the six as suggested by the terrace flight. The ODP 677 deep sea-record shows that there is a marked difference between the various oscillations (Fig. 3.8). The deep cold spells alternate with less pronounced ones. This latter pattern shows great similarity with a pollen based climatic evolution as has been developed by Zagwijn (1985), despite his different age model. This coherency suggests that the continental record is controlled not so much by the 40 ka signal, but by cold episodes with a recurrency of 100, 200 and 400 ka. This may explain why there are only six terrace levels preserved in the record. Only during those deepest cold spells enough sediment is stored in the Maas valley to arrive at conditions for terrace preservation.

To examine the consequences of such an age model for the undated terraces, we plotted an age/altitude graph and applied the ODP 677 timescale (Table 3.2) (for East Maas terraces the ODP 846 timescale was applied, see below). The obtained graph fits perfectly well with the younger record (Fig. 3.9).
The fitted graph shows decelerations and accelerations. As argued above, tectonics are envisaged to be the driving force for the capability of the river to incise to a level deeper than the previous terrace base. The variations in tectonic dynamics, derived from this correlation, fit the European context (see chapter 7). These observations suggest to confirm our assumption about the different significance of the cold episodes during the early Matuyama. These circumstantial evidences indicate that for the episode before 1 Ma, under different forcing conditions, only the extreme cold spells have been registered in the fluvial realm. They support the basic principles behind the stratigraphic significance of fluvial terrace-sequences: terrace sediments = coldest intervals and terrace-scarps = incision phases = warmest intervals. Interstadial conditions did not become expressed in the morphology. The scarps form the erosional unconformities between the various terrace building episodes. These morphostratigraphic units merge downstream into stacked series of the same cold-climate sediment fluxes separated by interglacial deposits (Ruegg,1994).

For the much warmer Late Miocene to Early Pliocene it has been shown in chapter 2 that erosional unconformities (= sequence boundaries) between the various Maas/Rhine stacked sediment packages within the Roer valley graben (adjacent to our study area) match prominent $\delta^{18}O$ variations in various deep-sea cores. In that record it is assumed that the relative coarse-grained lower parts of the sequences represent the cooler episodes and the finer



abbreviations

W	Weichselian	EL	Elsterian
E	Eemian	P.T.	Pretiglian
H	Holsteinian	*	age - model older terraces: see text

Figure 3.8 Stratigraphic chart showing the correlation between the north-western European pollen stages (Zagwijn, 1985), the ODP 677 $\delta^{18}O$ ocean record (Shackleton et al., 1991) and the Pleistocene terraces of the Maas.

Inferred age (Ma)	Altitude (m)	Terrace name
0.003	38	Holocene floodplain
0.014	49	Mechelen a/d Maas
0.13	52	Eisden Lanklaar
0.245	57	Caberg-3
0.33	63	Caberg-2
0.42	70	Caberg-1
0.51	76	Rothem-2
0.62	81.5	Rothem-1
0.715	86	Gravenvoeren
0.78	97.5	Pietersberg-3
0.87	104.5	Pietersberg-2
0.955	112.5	Pietersberg-1
1.03	119.5	Geertruid-3
1.09	124	Geertruid-2
1.28	130	Geertruid-1
1.5	136	Valkenburg-2
1.57	139	Valkenburg-1
1.69	146	Sibbe-2
1.74	150	Sibbe-1
1.87	159	Margraten
2.06	167	Simpelveld-2
2.14	175	Simpelveld-1
2.44	184	Noorbeek
2.6	190	Crapoel
2.69	198	Kosberg-3
2.81	202	Kosberg-2
2.94	206	Kosberg-1
3.14	215	Waubach-3
3.3	220	Waubach-2
3.98	235	Waubach-1
13.8	250	youngest marine cliff

Table 3.2 Inferred age/altitude relation of the Maas terrace flight taken at the latitude of Maastricht.

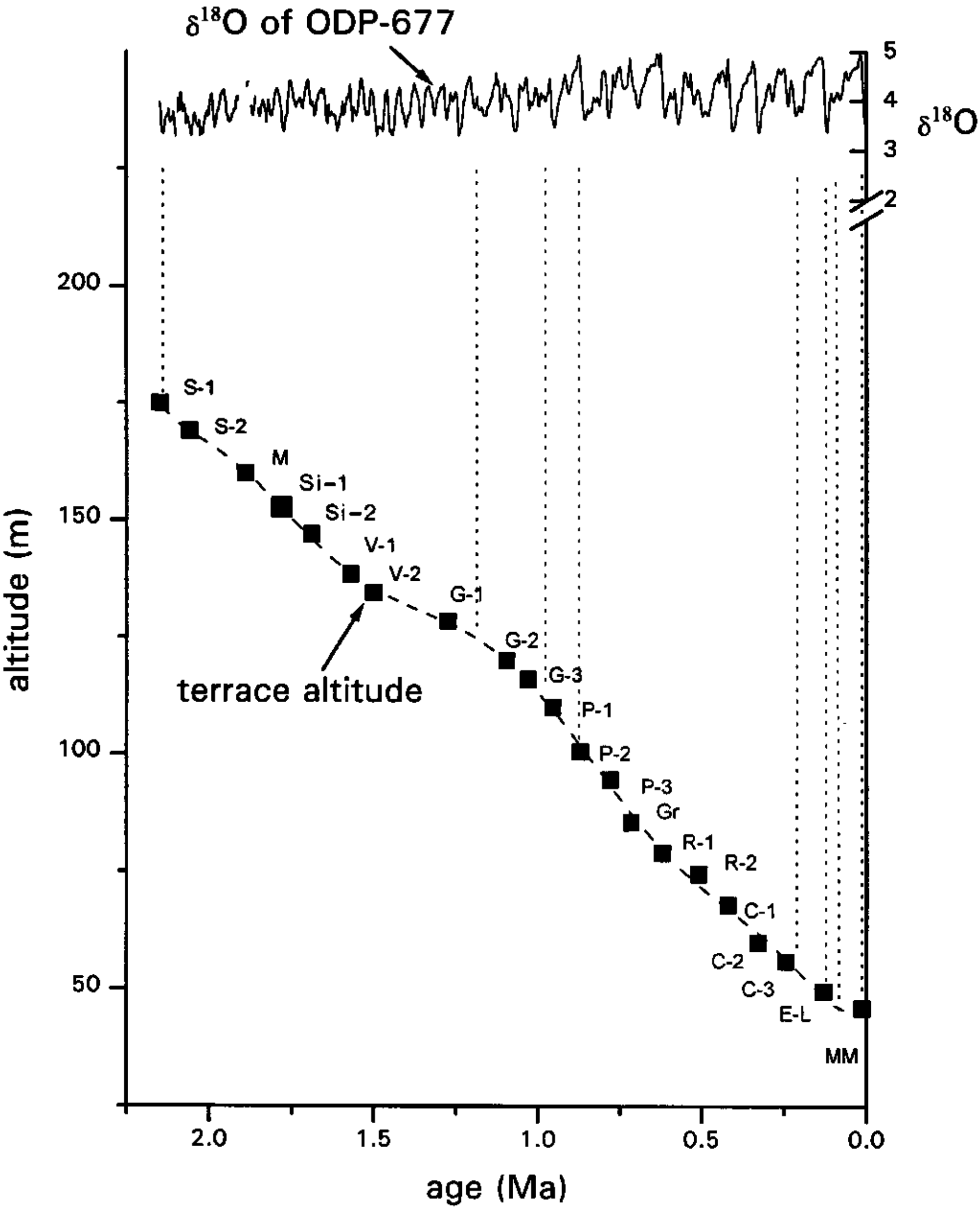


Figure 3.9 Altitude-age plot of the youngest 21 terrace levels considering sequence stratigraphic arguments and age determinations mentioned in text.

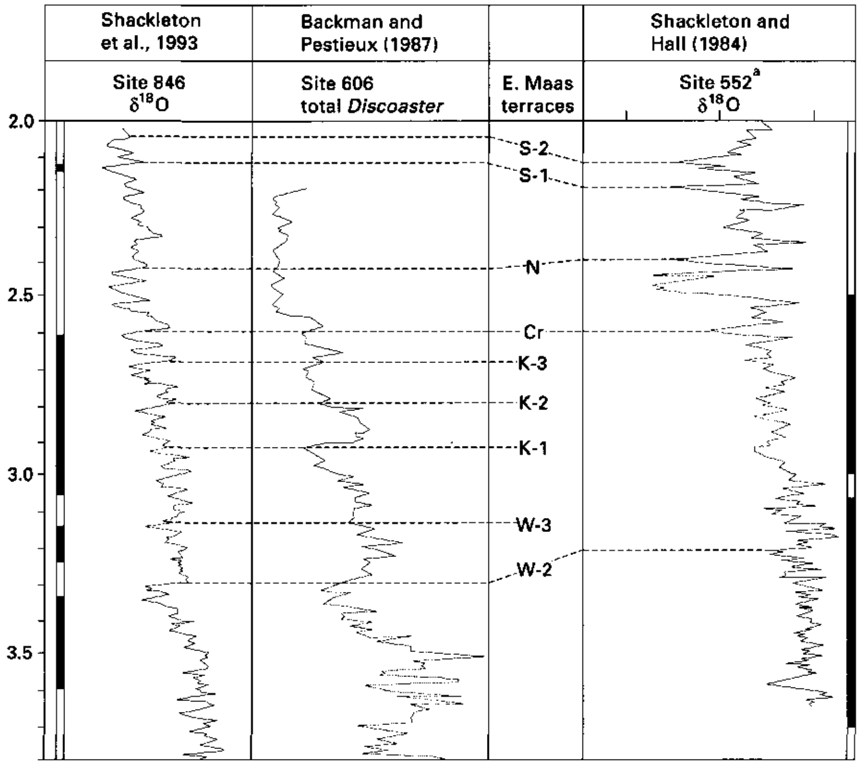


Figure 3.10 Late Pliocene to Earliest Pleistocene climate trends expressed in deep sea-records. East-Maas terraces are plotted at marked cooling events.

top series represent the relative warm intervals within the overall "warm" geological period. The sequence boundaries therefore may be regarded as the lateral equivalent of the scarps in the terraced region. Fine grained sediments have been bypassed in the terraced area during periods of incision. Although the controlling parameters and causes might not be the same as those of the Pleistocene terraces, the ultimate conclusion is that fluvial sedimentary units either expressed as buried stacked sedimentary sequences or as exposed terrace sequences can be used as a correlation tool to other (better constrained in time) climatic proxy records.

The East Maas valley terraces

Morphology

We distinguish 10 terrain steps on the East-Maas valley flanks underlain by fluvial deposits. This number is based primarily on the morphology of the waterdivide between the brooks Geul and Gulp (section C and D in Fig. 3.5).
From old to young, our series comprises the following levels: Waubach-1 (= W-1) through 3; Kosberg-1 through 3; Crapoel; Noorbeek and Simpelveld 1 and 2. The vertical distance

between the various levels varies between 15 m and 4 m (Table 3.2). The altitude differences between East-Maas terraces appears to occur with steps of around 4-5 m, or their multiples. This suggests little variability in the ultimate combination of factors controlling terrace formation.

Age model

Up to now, age models for the terraces older than the Simpelveld levels are very loose due to lack of time constraint. The widest likely age-range is between the 13.8 Ma Middle Miocene sealevel highstand and the 2.5 Ma northern hemisphere ice event.

The Neurath Sands deposits evidence the 13.8 Ma transgression into the Lower Rhine Embayment (Zagwijn and Hager, 1987; Herngreen, 1987). We correlate the coastline of this youngest transgression, that intruded far enough southward, with a low fossil cliff at 250 m altitude in the East Maas valley (e.g. section C in Fig. 3.5). This cliff is marked by a terrain step mantled with well rounded beach pebbles. The correlation is based on a palaeogeographic argument (the strike and position of the coastline) and stratigraphic argument (transition of marine facies to a fluvial facies).

After the Waubach terraces, the Kosberg terrace deposits are the first to contain large boulders transported by ice-floes. This feature is traditionally correlated with the Prae-Tiglian (P-T in Fig. 3.8; 2.5 Ma), the first date that pollen indicate ice-age conditions in the Netherlands. This correlation is problematic in our line of reasoning about the relation between terraces and cold stages. From figure 3.8 follows that there is not enough time-space available for the Noorbeek- and Crapoel terraces. Sediment characteristics of the subject terraces (high quartz content; heavy minerals marked by a stable association; palaeosols with the characteristics of a red-yellow podsolic soil) all might point at a Pliocene age (Kuyl, 1980).

Below we introduce uplift rate as a new element in the range of circumstantial evidences for an age-approximation of the East-Maas terraces. It is evident that uplift rates for the period between 13.8 Ma and 2.5 Ma were nowhere near those after 2.5 Ma, if simply measured by dividing total altitude over time.

ODP site 846 (leg 138) is the first, astronomical calibrated, high resolution record covering the Pliocene (Shackleton et al., 1993). This provides the timeframe for the cooling events that went prior to the 2.5 Ma event. Among many other deep-sea records, the general pattern of the pollen record (Zagwijn, 1960) is in fair agreement, as far as long term trends concern, to that of ODP 846 despite some significant differences. This trend (see figure 3.10) shows a general warm part of the Pliocene (the Brunssumian p.p) between 4.5 Ma and 3.4 Ma, followed by a less warm part between 3.4 Ma and 2.5 Ma (the Reuverian). The first half of the Reuverian (3.4 Ma- 2.95 Ma) clearly falls apart into 2 cycles separated by a cooling at 3.15 Ma. The oldest of the two corresponds to the Reuverian A; the younger, showing a bipartion like the pollen do, corresponds to the Reuverian B. The run down to the 2.5 glacial event (= Reuverian C in the pollen zoning) shows in various ocean records several cycles, together spanning 400 ka. Some records show a gradual build up in intensity like $\delta^{18}O$ in ODP 846 with occasionally deep cooling events. Similar to this is the picture by abundancy variations in the sum of six nannofossil species of genus *Discoaster* in Site 606 (Backman and Pestiaux, 1987) (Fig. 3.10).

Other records suggest a sudden intensification like the pollen and DSDP Site 552A (Shackleton and Hall, 1984). The latter record shows a similar frequency in climatic cycles

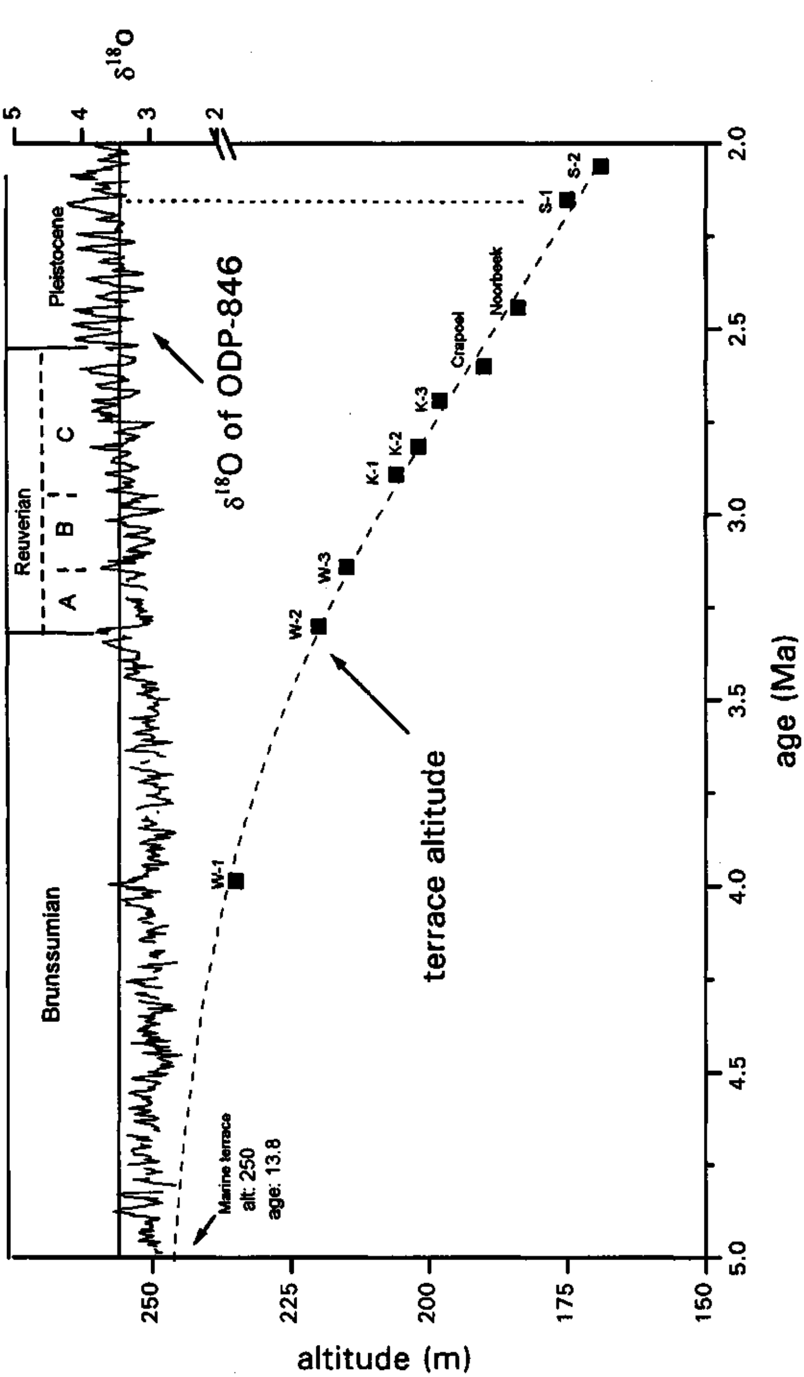


Figure 3.11 Altitude-age plot for the East Maas terraces. As a measure for the age assessment we took the age of a cool-warm shift in ODP 846 record (Shackleton et al.,1993) under the considerations mentioned in the text. The fitted function includes the youngest marine terrace. This graph is considered as a proxy of the regional uplift.

as the others, only the amplitude differs. The strong covariance in the ocean records for the period between 2.5 Ma and 2.9 Ma strongly suggests that climatic variability before and after 2.5 Ma is basically alike. This observation may provide more time-space for the cold climate Kosberg terraces.

The general valley-side morphology does not indicate an important lapse in geomorphic processes but rather a continuing succession in the evolution of the landscape providing the uplift rate decreases back in time.

All these circumstantial evidences together favour a simple solution by assuming that the East-Maas terraces represent Late Pliocene climatic events rather than similar events from the Late Miocene or the Early Pliocene.

We took as working hypotheses: (1) Periods with clear evidence of surface-water temperature fluctuations are either evidenced by ice-rafted sediments in the North Atlantic Ocean or by events with strong fluctuations in the *Discoaster* spp. record (Backman and Pestiaux, 1987). (2) The normal polarity found in the sediments of the Crapoel level does not reflect a present day field direction. (3) We accept a correlation of the Simpelveld-1 level with deep sea stages 82/81 on the basis of combined circumstantial evidences of palaeomagnetism, pollen and morphostratigraphy.

In Figure 3.11 we worked back in time from the altitude/time position of the Simpelveld-1 level and plotted subsequently the height of the various older East-Maas terrace-top levels against cool to warm shifts at the end of palaeoclimatic deterioration events with clear evidence of global cooling.

The sigmoidal graph fitted through the various terrace-time/altitude positions is interpreted to image an approximation of the Late Miocene through Pliocene regional tectonic uplift. This graph elegantly fits with the one obtained by the Pleistocene terraces-age correlations (Fig. 3.12). To this we considered the episode till the formation of Valkenburg-2 terrace (V-2) because after that the stressfield conditions changed apparently. The latter will be discussed in chapter 7.

The interpolated graph also suggests that a change in the regional uplift, after a long period of quietness, sets in after about 10 Ma. These inferred tectonic dynamics, despite the coarse interpolation, are in good agreement with the tectonic evolution and lithological infill of the adjacent Lower Rhine Embayment. The sequence of continental infill in this depocenter shows very little siliciclastic input combined with insignificant fault activity during the Early - and Middle Miocene (Gliese and Hager, 1978). During this period of over ten million years, siliciclastics have been replaced by wide spread peat growth (Main Seam or Hauptflozgruppe). As late as 10 Ma hinterland uplift is expressed by significant subsequent siliciclastic sedimentation. This event is well constrained in time by dinoflagellate assemblages (Herngreen, 1987) in conjunction with strontium-isotopes dating (Beets, 1992).

This independent dataset on the sedimentation record is in agreement with the uplift consequences of our age approximation of the Pliocene terraces.

Alternative Pliocene-age/altitude interpretations for the latter will inevitably lead to a more fluctuating uplift pattern. But the general trend stays.

A superimposed pattern is for the time being less satisfactory because it suggests for this episode more than can be sustained by other controls. The general shape of the obtained curve and its step-wise pattern between 1.5- 1.1 Ma will be discussed in the light of other records in chapter 7.

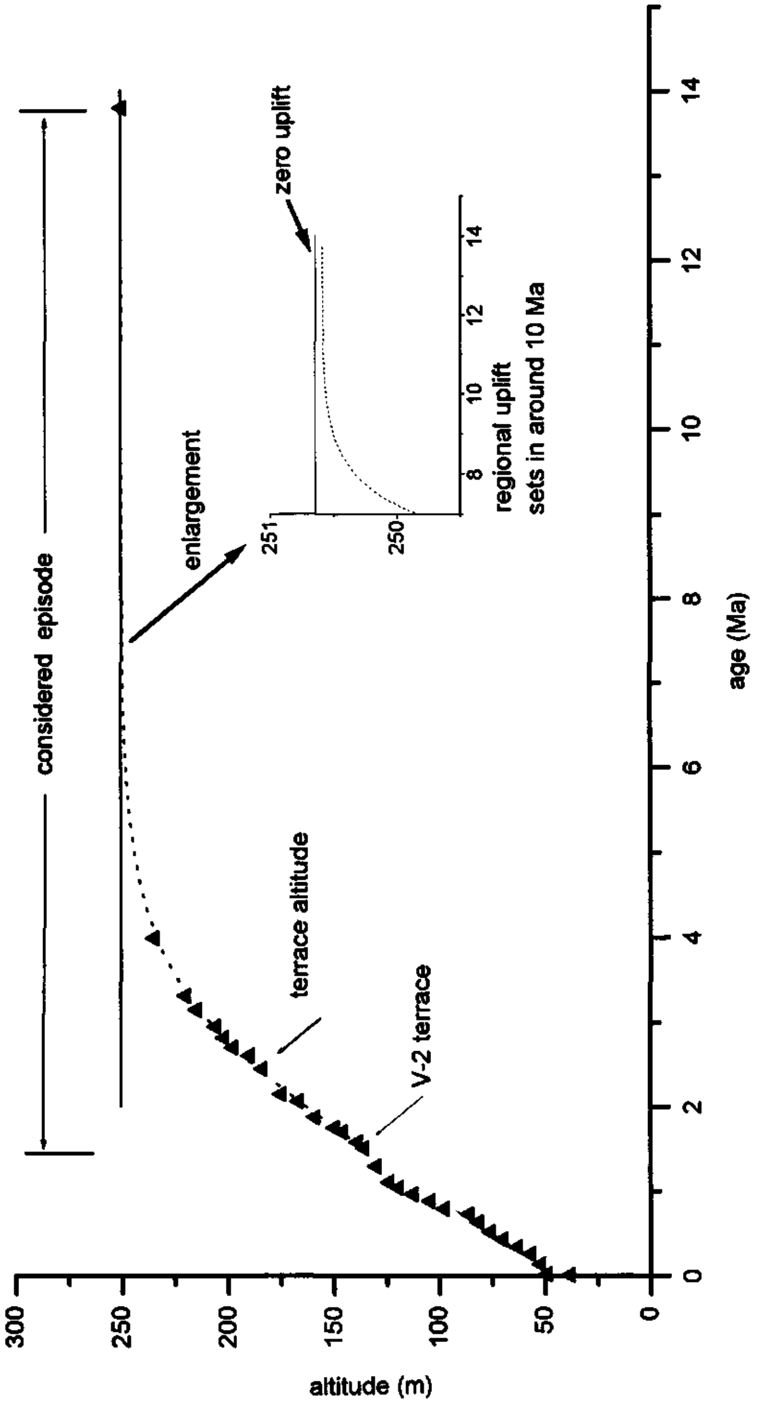


Figure 3.12 The combined East and West Maas terrace record plotted in the discussed age/altitude relationship. The fitted function through the terraces older than V 2 is considered as a proxy for the regional uplift. It shows that uplift sets in around 10 Ma and reaches a maximum rate of change around 3 Ma.

Conclusions

The multi-disciplinary investigation of morpho-sedimentological evolution of the lower Maas valley provides an unique opportunity to clarify the long-term river dynamics, expressed as alluvial terraces, in terms of climatic change and tectonic uplift.
Based on sparse but favourable distributed evidence on age control in conjunction with a fluvial sequence stratigraphic approach, the valley showed to be an unparalleled storehouse of a 4 Ma long continental record.
Progressive uplift made this discontinuous record continuous in its registration of repeated climatic shifts. Within our model each cool and cold episode (controlled by the eccentricity frequency of the orbital system) is registered in the river system by storing sediments, whereas warm temperate episodes are marked by terrain steps. This model shows by the presence of river terraces that the middle part of the Pliocene was characterized by considerable climatic variability.
From the morpho-sedimentological point of view the change from Pliocene to Pleistocene conditions is a very gradual one.

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4. Three-dimensional modelling of Quaternary fluvial dynamics in a climo-tectonic dependent system. A case study of the Maas record (Maastricht, The Netherlands)

Abstract - *A three-dimensional model (MATER) simulating Quaternary terrace development along the river Maas at Maastricht was made. Simulation results indicate that Maas terraces are most probably the result of both climatic and tectonic changes in time. Terrace formation is mainly climatically determined but their long-term preservation is closely related to the prevailing tectonic regime. Quaternary uplift of the Maastricht area has known several phases with different 'constant' uplift rates. These differences in uplift rate are a main cause of the general valley morphology near Maastricht. Whereas the deep-ocean climatic record is clearly dominated by the tilt cycle between 2.4 and 0.9 Ma, the continental record in the form of the cold-stage floodplain aggradation-dominated periods appears to register primarily modulations of the eccentricity cycle over this episode. During the last 0.9 Ma, when the general uplift rate was of the order of 0.1 m/ka, a 'complete' climate-related terrace sequence developed. Every 100 ka climatic cycle is registered in this part of the terrace sequence. Both model and field evidence suggest that Maas terrace sediments are mainly records of the 'cold' periods. As such they are valuable continental counterparts of palynological records registering 'warm' episodes. Tectonism determines long term terrace preservation and the value of a specific fluvial system as record of past climatic change.*

Introduction

Long continental and marine records are of great value in understanding global changes. Understanding the correlation between the continental and marine records is crucial for gaining a comprehensive view of global changes. One of the physical links between the continent and the ocean is formed by the fluvial system. As such it can help us to make a more detailed analysis of basin margins filled with continental deposits. Intra-plate stress fields relate the tectonic history of the centre of the basin with its margins (Cloetingh, et al. 1985). Basin margins that experience uplift become dissected by the fluvial systems that are the feeder systems to these basins. So combined geomorphological-sedimentological and stratigraphic studies of the fluvial system in the realm of the rising basin flank may provide us with a detailed insight into the temporal evolution of the basin dynamics.
In the previous chapter the Maas river terrace flight has been discussed with respect to its morphological, sedimentological and age-control aspects. From a morphological point of view terraces are quite simple features but from a sedimentological point of view they are very complex. Each terrace unit is usually made up of several stacked, often incomplete sedimentary cycles, representing several alternating depositional and erosional stages. Generally three major driving forces behind terrace formation are used: changes in climate, tectonism and base level. Climate change and tectonic uplift are inferred to have been key factors in the formation and preservation of the West Maas terraces, although rivers may make terraces without there necessarily being an external change (Schumm, 1977). Base level changes have only importance in fluvial reaches downstream a terrace intersection. They loose their significance updip of the basin hinge line.

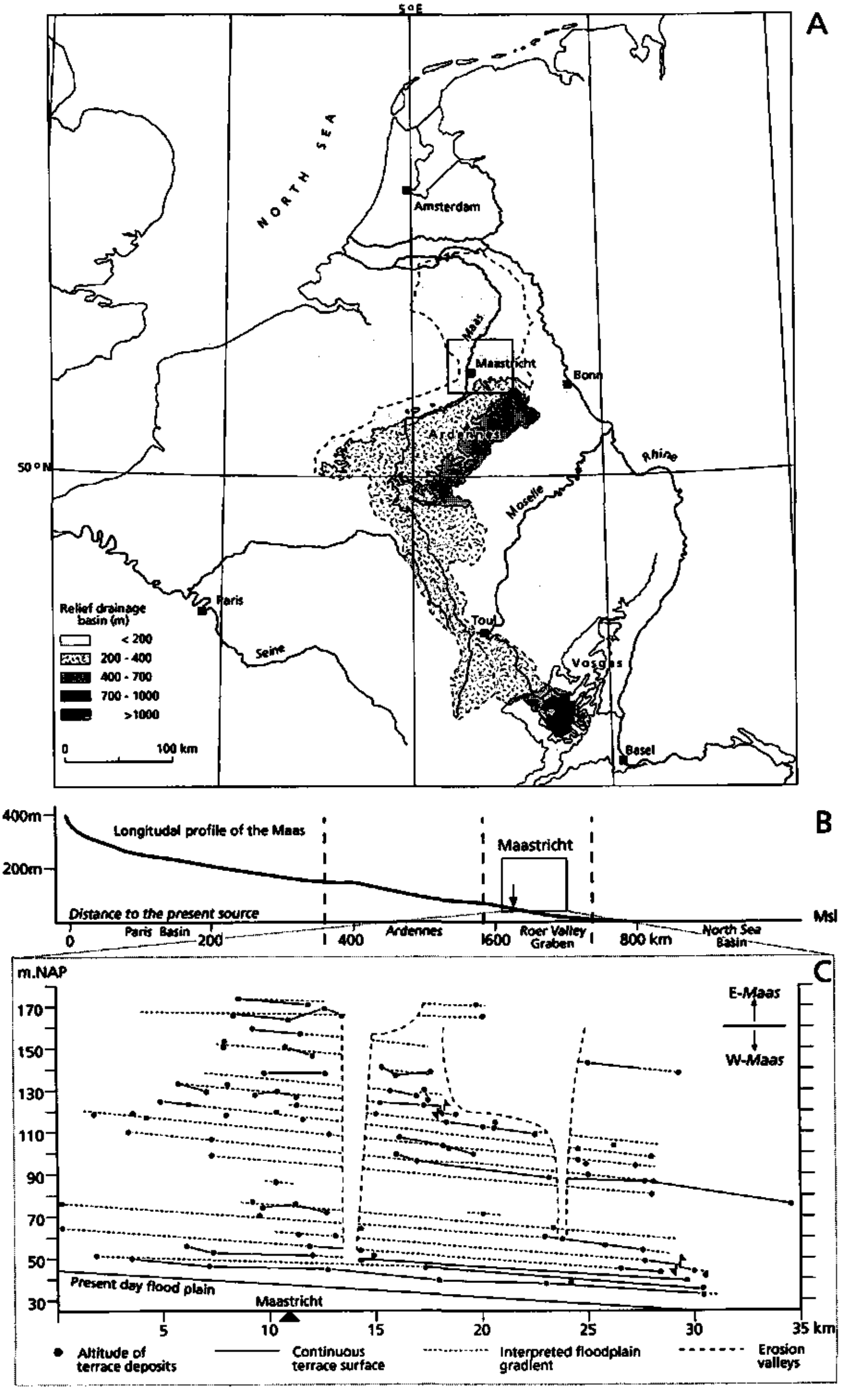


Figure 4.1 Setting of the study area within the catchment of the river Maas (a); Present day longitudinal profile of the Maas, the major tectonic domains are indicated (b); Reconstruction of the Pleistocene longitudinal profiles of the West-Maas.

In this chapter we will evaluate assumptions made in the previous chapter by modelling the West Maas terrace flight. Models that do describe fluvial system dynamics for longer time spans are rare because most knowledge of fluvial systems is based on short-term, well controlled experiments (Schumm, 1977; Gregory, 1983; Dawson and Gardiner, 1987).
A conceptual, long-term (2000 ka) macro-scale (100 km^2) fluvial system model, a so-called coarse-scale model (Thornes, 1987), was developed by Veldkamp and Vermeulen (1989). This model simulates the development of fluvial terraces as the result of strongly simplified external and internal changes in the simulated fluvial system. An adapted and extended version of this model was applied to the Allier terrace sequence (Veldkamp, 1992). However, the major limitation in the Allier drainage basin was the limitation of the terrace record.

The adapted MATER model (MAas TERraces) is a PASCAL program which runs on a VAX 8600. It simulates climate and tectonic dependent terrace formation three dimensionally. MATER simulations will give insight into possible effects of climatic and tectonic changes and internal fluvial dynamics on general alluvial stratigraphy and valley morphology of the Maas valley near Maastricht.

The Maas terrace flight

The river Maas is a rain fed river draining an area of approximately 33000 km^2 in Western Europe with a mean annual discharge of 260 m^3/s and peak discharges up to 2800 m^3/s at Maastricht. The river basin is superimposed on various morpho-tectonic units and can be classified as a complicated basin as mentioned by Starkel, 1990. (Fig. 4.1). The study area is situated in the south-eastern most part of the Netherlands near Maastricht. Here the river has left the Ardennes and has spread its courses over the southern flank of the Roer valley Graben divided over two successive valleys, the East-Maas and the West-Maas (Fig. 4.2).

Maas terraces

East and West-Maas valleys together contain about 31 different terrace levels with steps upto about ten meters. Quartz content of terrace gravels decreases with altitude in favour of fresh rock fragments (Van Straaten, 1946). Simultaneously the heavy mineral composition changes from an extremely stable association towards a more unstable spectrum with a significant proportion of garnet, epidote and hornblende (Zonneveld, 1949) indicating a renewed incision of the uplands. Terrace gravels are generally coarse and contain abundant poorly rounded gravel together with large drift blocks (boulders larger than 0.5 m in diameter) and blocks of non-cohesive sediments, apparently transported in frozen state, indicating cold climate fluvial transport (Krook, 1961; Brunnacker, et al.,1982).

The West-Maas valley contains about 20 terraces (Figs. 4.1 and 4.2) above the present day floodplain. The reconstructed longitudinal profiles are parallel to each other and to the Holocene floodplain. They represent top levels of cold-stage aggradation in the course of the valley formation. This parallelism indicates that terraces developed out of reach of sea-level fluctuation. Moreover floodplain aggradation took place during sea-level low stands. First-order geological age control on the West-Maas terrace sequence comes from climatic (Gullentops, 1954), biostratigraphical (Zagwijn, 1974; Zagwijn and de Jong, 1983) and

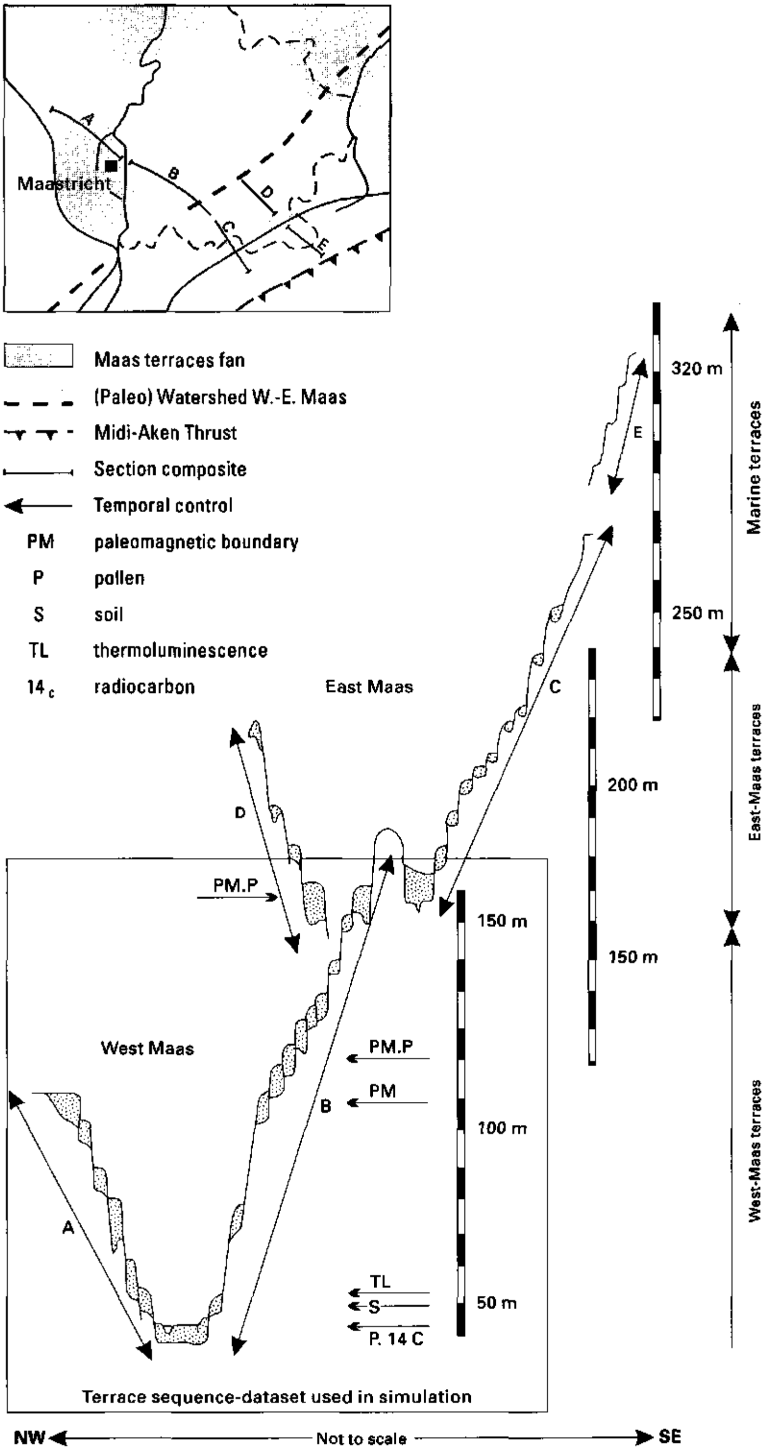


Figure 4.2 Schematic cross-section of the Maastricht terrace sequence

palaeomagnetic (Bruins, 1981) indications of the sediments, combined with C^{14} (De Jong, 1984) and thermoluminescence datings (Huxtable and Aitken, 1985) (Fig. 4.2). This age control also shows that at least the last three glacial stages are represented in the Maas terrace record as an individual terrace. This frequency suggests an individual terrace level for approximately each 100 ka. Nine terrace levels higher up the flight, the Geertruid 3 level is found. This older level contains in its top-sediments Interglacial pollen (Zagwijn and de Jong, 1983) and the top of the Jaramillo palaeomagnetic event (Bruins, 1981) and was therefore dated around 1 Ma BP. Now we have left 8 undated terrace levels to divide over 750 ka. Assuming a terrace level for each 100 ka a match of the undated terraces with known cold stages with the ODP 677 record (Shackleton et al., 1990) is possible. This correlation allows us to assign an approximate age to the individual terrace bodies between Geertruid 3 and Caberg 3 levels. Of the nine terraces above the Geertruid 3 level only the Geertruid 2 and Simpelveld 1 levels contain pollen (Zagwijn 1974) and/or palaeomagnetic boundaries (Bruins, 1981). These data allowed us to match these older West Maas terraces with the ODP 677 record (Fig. 4.3). The match of the ODP 677 record with the pollen record (Zagwijn, 1985) relies partly on a correlation of Maas and Rhine deposits. This exercise revealed an interesting new positioning of some of the pollen stratigraphic stages within the Cromerian complex.

Changes in the Maas catchment area

At least since the Pliocene, the Ardennes and Vosges massifs have belonged to the source areas of the Maas system. There is ample evidence in the Vosges that glaciers were active during cold stages of the Pleistocene epoch (Autran and Peterlongo, 1980). Since the deposition of the Caberg 3 sediments (about 250 ka BP, Huxtable and Aitken 1985) the Maas lost its upper reaches in the Vosges to the Rhine system (Moselle) as evidenced by changes in heavy mineral assemblage (Zonneveld, 1949; Bustament de Santa Cruz, 1976;). Another, much smaller part of the headwaters was lost to the Aire-Aisne-Seine system (Fig. 4.1). These captures did not only cut off more than 10% of the catchment (Bosch, 1992), but the Vosges with their relatively high altitudes also formed a significant source area in generating rain and melt water.

Tectonic control

In the study area, the successive Maas floodplains show a strong tectonic control. The mapped Maas terrace sequences (Van den Berg 1989; Felder and Bosch,1989) showed a strong relation in their spatial distributions to the position of folds and overthrusts known from the top of the Carboniferous. This strain pattern, initially generated by the Hercynian deformation phase (Waterschoot van der Gracht, 1938), is known in great detail from geophysical records and coal exploration and mapping (Patijn, 1963; Drozdzewski et al. 1985). Known structural antiforms of Hercynian age are reflected in the limits of palaeo-lateral migration of the river. These lineaments are parallel to the northern Ardennes thrust (the Midi-Aachen thrust). We interpreted this topographical expression to reflect incipient crustal buckling, a strain pattern caused by the neotectonic thrust-induced foreland stress (Van den Berg, 1990) (Fig. 4.4). During its evolution the river valley shifted away from the thrust, crossing several thrust-parallel blocks. This tectonic history is relevant to modelling because realistic simulations will require an unequal uplift for different valley slopes. In the

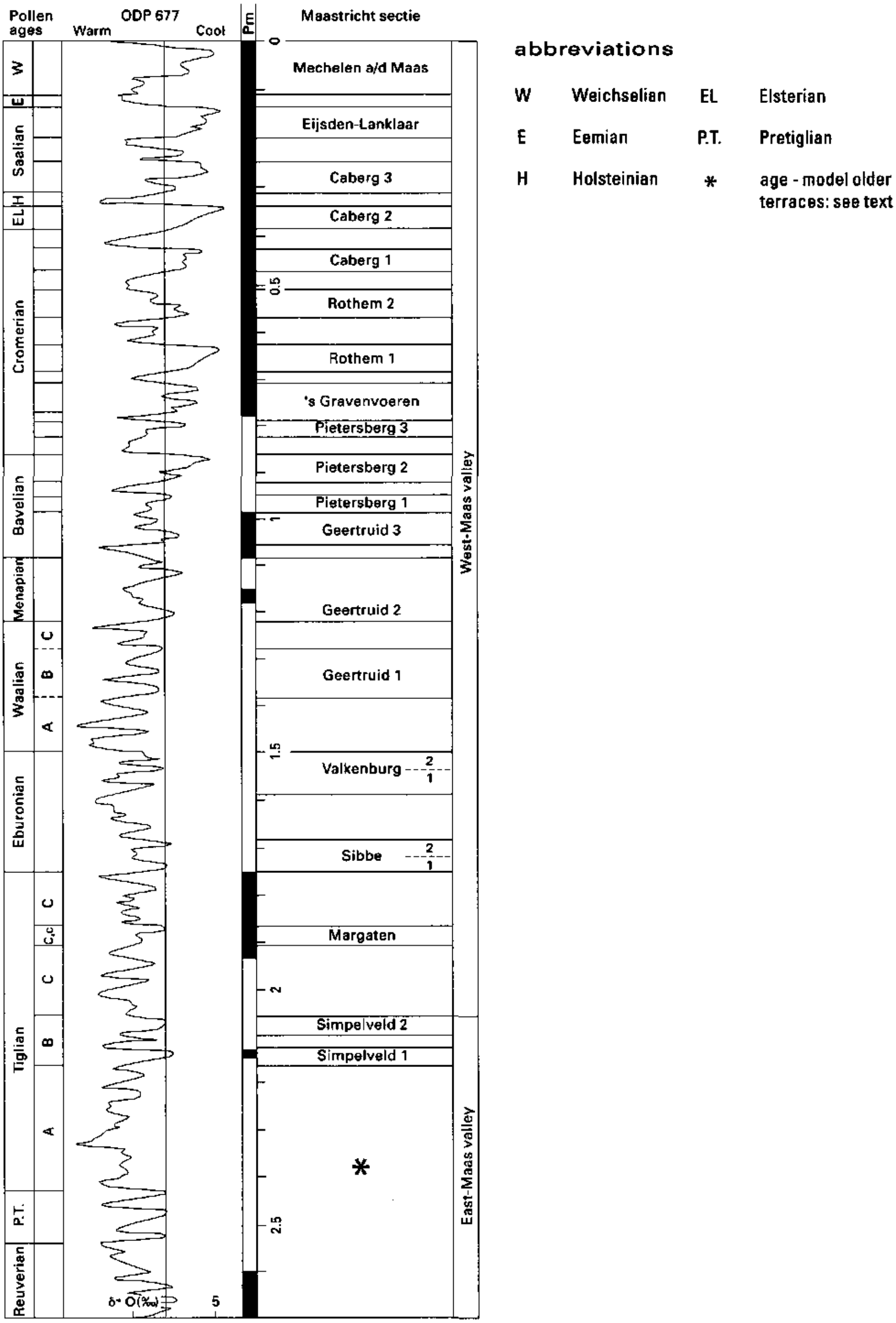


Figure 4.3 Correlation between the north-western European pollen stages (Zagwijn, 1985), the ODP 677 $\delta^{18}O$ ocean record (Shackleton et al., 1991) and the Pleistocene terraces of the Maas.

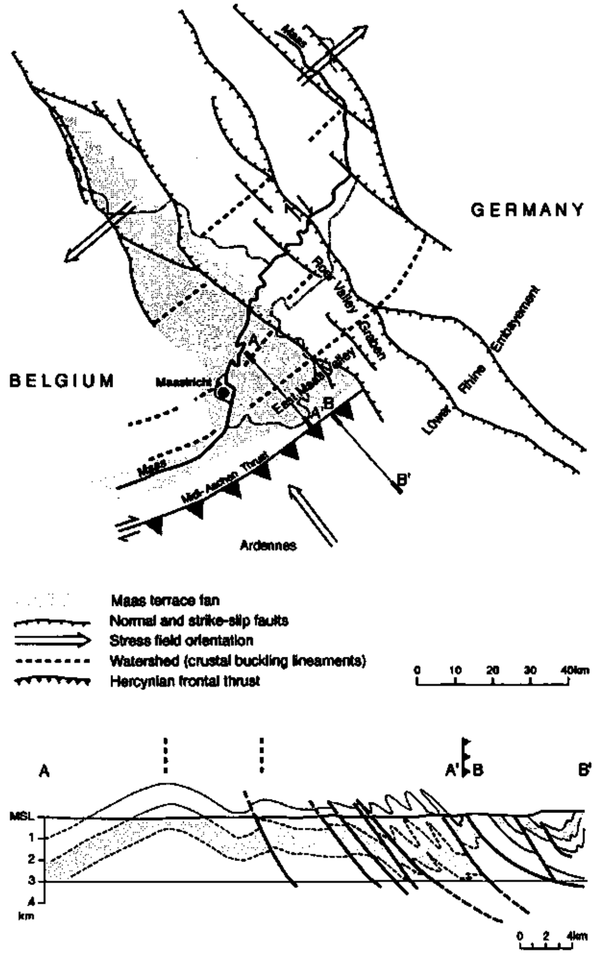


Figure 4.4 Structural setting of the Maas terrace fan on the southern flank of the Roer Valley Graben. Section B-B' compiled and simplified after Meissner et al.(1983).

study area, the Maas valley near Maastricht, the western and eastern valley slopes of the West Maas belong to different tectonic units, the eastern valley slope being the most uplifted one.

To gain insight into the rate of uplift of the study area, we reconstructed an uplift curve as a time-altitude diagram, using the altitude of the various terrace surfaces as marker points relative to the present floodplain. With respect to the time-axis this marker is placed at the end of a glacial period. The resulting uplift curve for the Pleistocene shows a pattern of fluctuating uplift rates over the Pleistocene, this pattern will be discussed in another paper. For the purpose of this paper we only wanted to know the effect of variations in the rate of uplift within a realistic domain. The values of uplift rates learned from this curve fall within the range of 0.02 and 0.7 m/ka. For the Pleistocene as a whole we arrived at an average value of 0.06 m/ka. Although these values differ a factor ten, they still fall within a

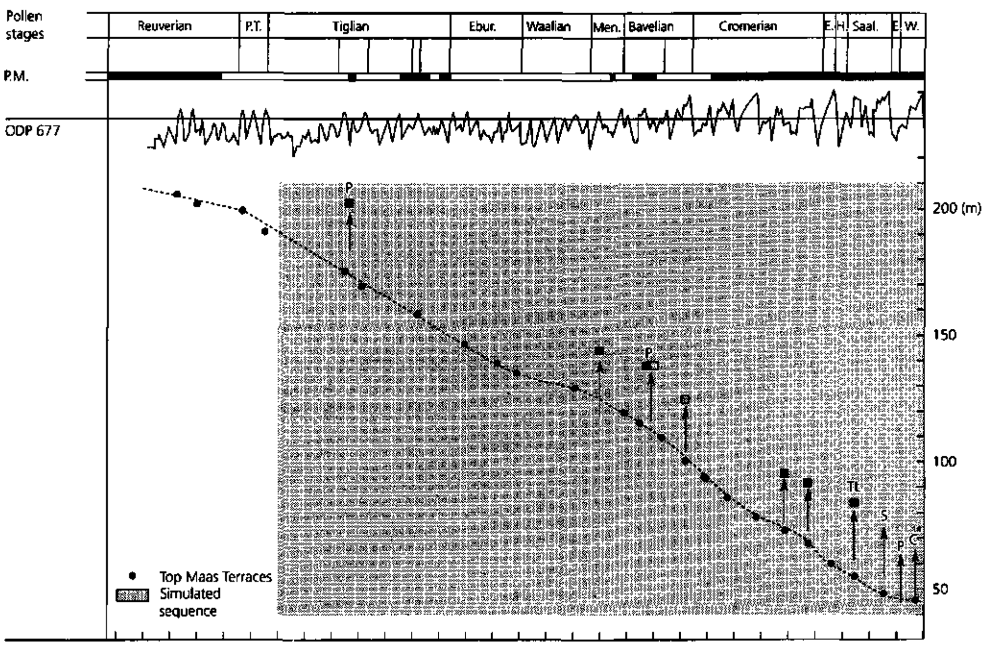


Figure 4.5 Reconstruction of the regional uplift, based on the altitude/age position of the Maas terraces.

range found elsewhere in North-Western Europe (e.g. Zuchiewitcz, 1991). Primary levellings in the Maastricht area provided values of the order of 0.8 $\pm$ 0.3 mm/a averaged over the last five decades (Groenewoud et al, 1991).
Because the reconstructed record of crustal uplift, as shown in Fig. 4.5 in detail still contains a number of uncertainties, we generalized model uplift rates by averaging over intervals between time-constrained terrace levels. In the modelling procedure the following uplift rates were used:
2150 ka to 780 ka BP an uplift rate of 0.05 m/ka
780 ka to 250 ka BP an uplift rate of 0.105 m/ka
250 ka to 10 ka BP an uplift rate of 0.02 m/ka

The exercise discussed in this paper will be restricted to the development of the West Maas because the time control on the sediments of this section is much better than that on the East Maas section.

Model characteristics

General model construction, organization and operation have already been described in previous papers (Veldkamp and Vermeulen, 1989; Veldkamp, 1992). Only some generalities and adaptions and extensions for the Maas system will be described here. The Model was organized with entities and attributes. An entity is an independent unit in the object system. The entities in this model are LANDSCAPE and RIVER. An attribute is a property, mark or characteristic of an entity, such as discharge for RIVER and valley_depth for LANDSCAPE. The attributes have estimated values within, for the Maas system, realistic domains (Table 4.1).

Table 4.1

A.	Time	:1 < Time < 2400 (ka) (Time steps of 2 ka)
B	Entities	
1. Entity: RIVER		
Attributes		Domain
Discharge		:$1\text{x}10^{12}$ < Discharge < $1.5\text{x}10^{13}$ (m^3/ka)
Input_load		:$-0.6\text{x}10^{9}$ <Input_load<$1.4\text{x}10^{8}$ (m^3/ka)
Form		:meandering or braided
Width		:0.2 <Flood_plain_width< 19.6 (km)
Maxload		:$7.05\text{x}10^{-5}$ < Maxload < $4.5\text{x}10^{-3}$ (m^3s^{-1})
Erosion		:$-2.1\text{x}10^{-3}$ < Erosion < $4.5\text{x}10^{-3}$ (m^3s^{-1})
2. Entity: LANDSCAPE		
Attributes		Domain
Quplift		:0 < Quplift < 0.5 (m/ka)
Quplift_unequal	:0 < Quplift_unequal < 0.05 (m/ka)	
Relief x,y,z		:150 < x < 15000 (m)

	150 < y < 15000 (m)
	1 < z < 500 (m)
Valley_depth	:20 < Valley_depth < 500 (m)
Stratigraphy	:0 < stratigraphy < 1000 (ka)
sed_composition	:1 < sediment_composition <= 4 (-)
Valley_width	:0.4 < Valley_width < 19.60 (km)

The interaction between the two entities are the erosion and sedimentation processes acting in time steps of 2 ka. These processes are large-scale analogies of the real processes. They are artificially constructed processes reacting to changes in the 2 ka discharge/sediment load equilibrium and they are both defined as a function of climate. When the sediment input load exceeds the sediment transport capacity, which is a function of discharge, the difference is deposited (NOT EROSION = TRUE), and in case the transport capacity exceeds the input load the difference is eroded in the simulated system (EROSION = TRUE). The LANDSCAPE is changed by simple straightforward processes that add grid cells to or remove them from the LANDSCAPE in a way similar to deposition and erosion processes. Before these processes start acting, the boundaries and conditions controlling these processes are determined. For instance, erosion in a meandering river can only occur within the calculated floodplain width which is related to the 2 ka discharge. The changed LANDSCAPE is stored in a geographical information system (GIS). The model simulates the existence of a fluvial system within a valley and not the river dynamics itself.

Which processes act in the LANDSCAPE is determined by decision rules extracted from general literature on hydrology (Schumm, 1977). These decision rules are:
IF EROSION & UPLIFT & MEANDER THEN INCISION
IF EROSION & UPLIFT & BRAIDED THEN INCISION & BANK_EROSION
IF EROSION & NOT_UPLIFT & MEANDER THEN INCISION & BANK_EROSION
IF EROSION & NOT_UPLIFT & BRAIDED THEN BANK_EROSION
IF NOT_EROSION THEN DEPOSITION (disregarding all other conditions)

UPLIFT and NOT_UPLIFT are system states indicating the impact of tectonic uplift. UPLIFT is true (NOT_UPLIFT = false) when the fluvial erosion is not able to compensate for the LANDSCAPE uplift. MEANDER is true when the Maas is meandering, whereas the BRAIDED state is true when the Maas is braiding. Erosion processes are headward migrating along the longitudinal profile, whereas sedimentation migrates in a downstream direction. Irregular headward erosion during erosion of a meandering RIVER is caused by changes in the effective floodplain width during the simulation. The resulting irregular bank erosion has an effect similar to that of a meandering river in a real system.

Model Input

Climate

As shown by various environmental records on the two hemispheres, the Quaternary climatic changes are global and synchronous. The astronomical parameters included in Milankovic's theory satisfactorily describe the waxing and waning of continental ice-caps as evidenced from tuning procedures carried out on both continental (Kukla, 1987; Kukla et al., 1988) and long deep-sea records (e.g. Imbrie et al., 1984). Continental ice-sheets and related sea-ice fields in the Northern Hemisphere strongly influenced the ocean-continent heat transfer and the jet stream related depression tracks (COH members, 1988; Broecker and Denton, 1990) thus effecting North-Western European climate.

The relationship between climatic behaviour and fluvial dynamics depends on the rate of climatic change and the effect of such changes on discharge distribution and sediment load (Lowe and Walker, 1984).

During glacial episodes the southward expanding high pressure areas forced the depression tracks and the vegetation belts towards lower latitudes. The shift of the vegetation belts led to a vegetation type (tundra) in the Maas catchment which was particularly sensitive to high frequency changes in the climatic system and caused an increase in sediment supply to the Maas. The southward shift of the depression tracks led to more continental and drier conditions (Guiot et al., 1989; Ruddiman et al., 1989) causing relatively low mean discharges in the rivers in these periods.

Interglacials yielded an opposite picture, an increase in the mean discharge and a complementary decrease in sediment supply to the Maas.

As model simulations require a climatic input with a constant reliability during the simulated time span a simplified Milankovic curve was used as basic climatic input. The mean discharge per 2 ka is simulated as the sum of three sine (SIN) functions with the periodicities of the precession (23 ka), obliquity (41 ka) and eccentricity (96 ka). As the sediment load seems to have a similar behaviour complementary to the 2 ka discharge, it was simulated as the sum of the cosine (COS) functions of the same three periodicities.

The resultant harmonic functions are not weighted by the 400 000 eccentricity modification. This signal is clearly present before 1 Ma BP; also the dominance of the obliquity in this period (Raymo et al., 1989) is not accounted for in our climate-related functions.

Although the climate functions used do not exactly match the climatic curves derived from deep-sea cores and lake sediments, they sufficiently describe climatic changes during the Quaternary for our long-term modelling purposes. This is because these functions are used in a qualitative way. The finite-state model MATER determines states using the climate-related functions.

The effects of the climatic dynamics are thus simulated as changes in discharge and sediment load in the Maas. Except the difference in behaviour of the fluvial dynamics in glacial and interglacial periods the transition from a glacial to an interglacial environment was also taken into account. As a result of glacier melting in the Maas headwaters (Vosges) an extra sediment flux of 1.5 times a normal flux is supplied in the fluvial system. This extra sediment is only supplied to the Maas when the glacial duration is longer than 10 ka. This subjective criterium which can be seen as a threshold between the impact of a glacial and Stadial, is indirectly derived from a sediment flux reconstruction in the river Allier (Veldkamp, 1991). This river within the Loire basin (Central France) showed a strong climate control over its sediment fluxes during the Late Quaternary due to glaciers in its headwaters.

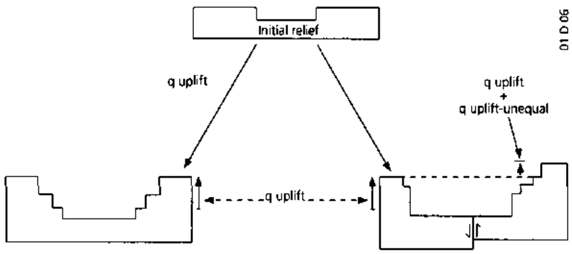


Figure 4.6 Tectonic components used in MATER simulations.

Tectonism

Two different components of tectonism have been incorporated in the MATER model (Fig. 4.6). A component of gradual uplift of the whole simulated landscape (QUPLIFT), and an uplift component describing the difference in uplift rate of the landscapes on both sides of the Maas (QUPLIFT_UNEQUAL). It was already argued that both sides of the Maas valley at Maastricht belong to two different tectonic units with slightly different movements in time. Based on the slight differences in terrace altitudes in the Maas valley we assumed a difference in uplift rate of only 0.0015 m/ka whereby the eastern valley slope experienced the highest uplift rate.

The general uplift reconstruction showed already the changing rates in the Maas basin in time. During the simulation the following constant uplift rates were used: from 2.4 to 0.78 Ma BP a rate of 0.04 m/ka, from 0.78 to 0.25 Ma BP a rate of 0.105 m/ka, and during the last 0.25 Ma a rate of 0.02 m/ka.

Initial relief

The chosen initial relief (Fig. 4.7a) has a surface area of 400 km^2 (20 km x 20 km) a maximum altitude of 180 m and a minimum altitude of 30 m. The initial relief consists of a valley with a width of 2000 m and a depth of 5 m only. Terrace stratigraphy displays only age 0, indicating that no fluvial sediments occur in the initial LANDSCAPE.

Model Output

The output of the MATER model is a data file (GIS) with LANDSCAPE altitudes and stratigraphy for each time step (2400). With these data, cross-sections, maps, three-dimensional relief graphs, or three-dimensional graphs of one cross section development in time can be drawn. It is also possible to make a time related graph showing changes in climate and the accompanying sediments and terraces. (Figs. 4.7, 4.8 and 4.9)

Model tuning

The model was tuned using the mean 2 ka discharge and 2 ka sediment load and their amplitudes. Two known system conditions were used as references during tuning: the total net

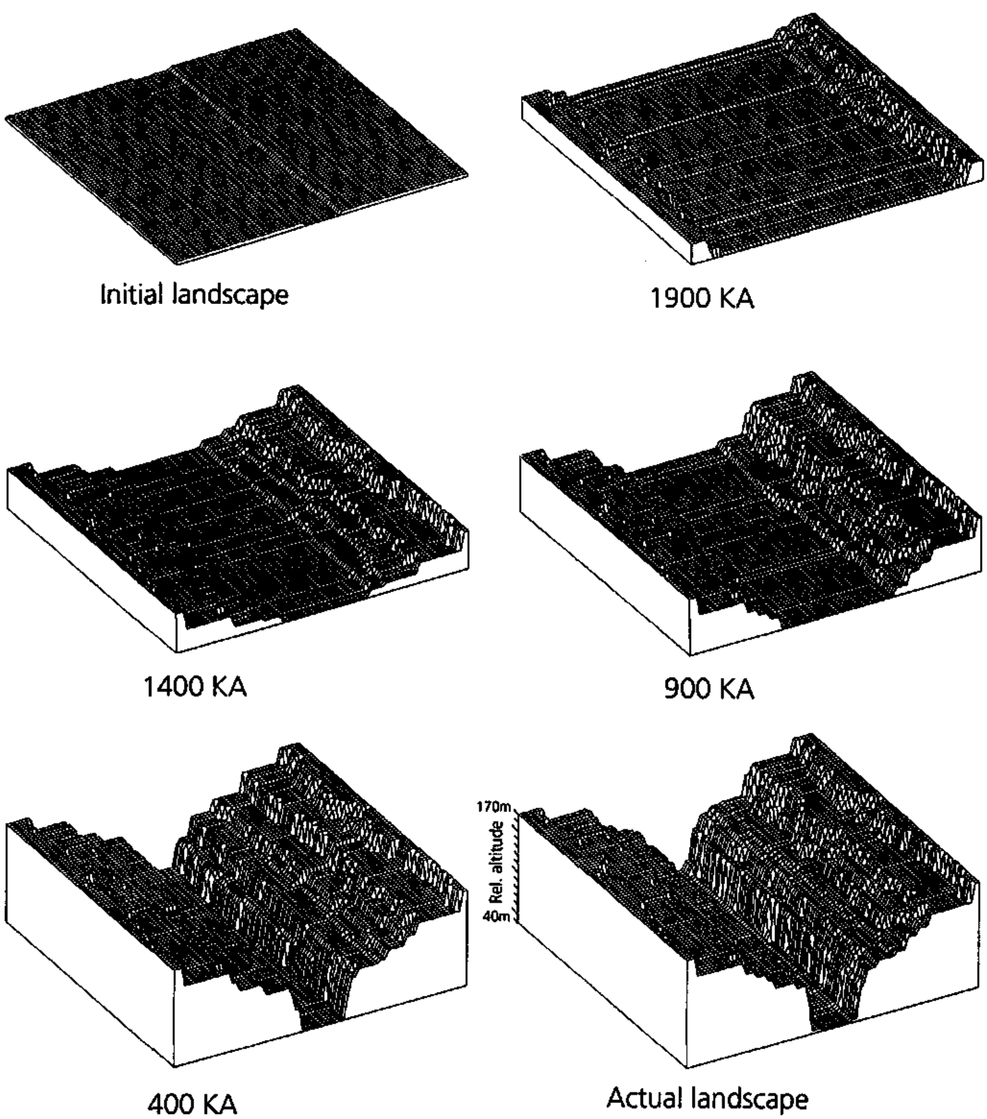


Figure 4.7 Three-dimensional valley relief development during simulation, in time steps of 0.5 Ma from 2.4 Ma BP (a) to present (f).

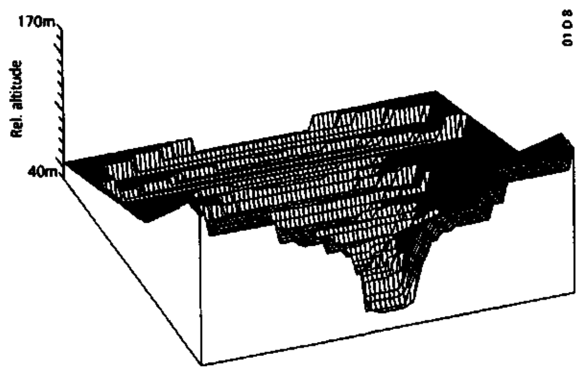


Figure 4.8 Three-dimensional relief development of one cross section in time.

volume eroded during the Quaternary, and the occurrence of alternating erosion and sedimentation. The tectonic inputs were relatively well known and not used for model tuning. Previous sensitivity testing (Veldkamp and Vermeulen, 1989) of the river terrace formation model showed the model to be sensitive to changes in its input variables: discharge, sediment load, uplift rate and unequal uplift rate for different valley sides, all in 2 ka time steps, as well as the decision rules described above. Model sensitivity could not easily be assessed as certain input variables are interrelated, such as the two uplift components and the 2 ka discharge and sediment load. The decision rules cannot be tested to standard procedures as their validity is based on the literature used. Terrace preservation is especially sensitive to landscape uplift rates from 0.05 to 0.2 m/ka.

Simulation results

The simulated LANDSCAPE development in time, under conditions described by the model input, is visualized three dimensionally for each 500 ka in Fig. 4.7a to f.

To illustrate the incision and sedimentation dynamics in more detail a cross-section development in time is shown in Fig. 4.8, showing relief changes during the 2400 ka of the simulation. This figure clearly shows the alternating incision and sedimentation of the Maas in time causing different terrace levels.

The role of climatic change is obvious. The simulated Maas system tends to form terraces at most changes from glacial to interglacial although terraces were also formed without a major change in climate. Such intrinsic threshold crossings are not very frequent and cause relatively small terraces which disappear during the simulation because climatic-induced terraces overrule such events. The general terrace formation mechanism is deposition at the end of a glacial period followed by an incision in the subsequent interglacial period. Therefore it is not surprising that, although the river has known a meandering state many times, the accompanying interglacial sediments are rarely found in the stratigraphical record.

This limited occurrence of meandering sediments is true for both the described simulation and the actual Maas terrace sequence. This illustrates that terrace sediments mainly contain a registration of the colder periods and their transition to the warmer periods. Terrace sed-

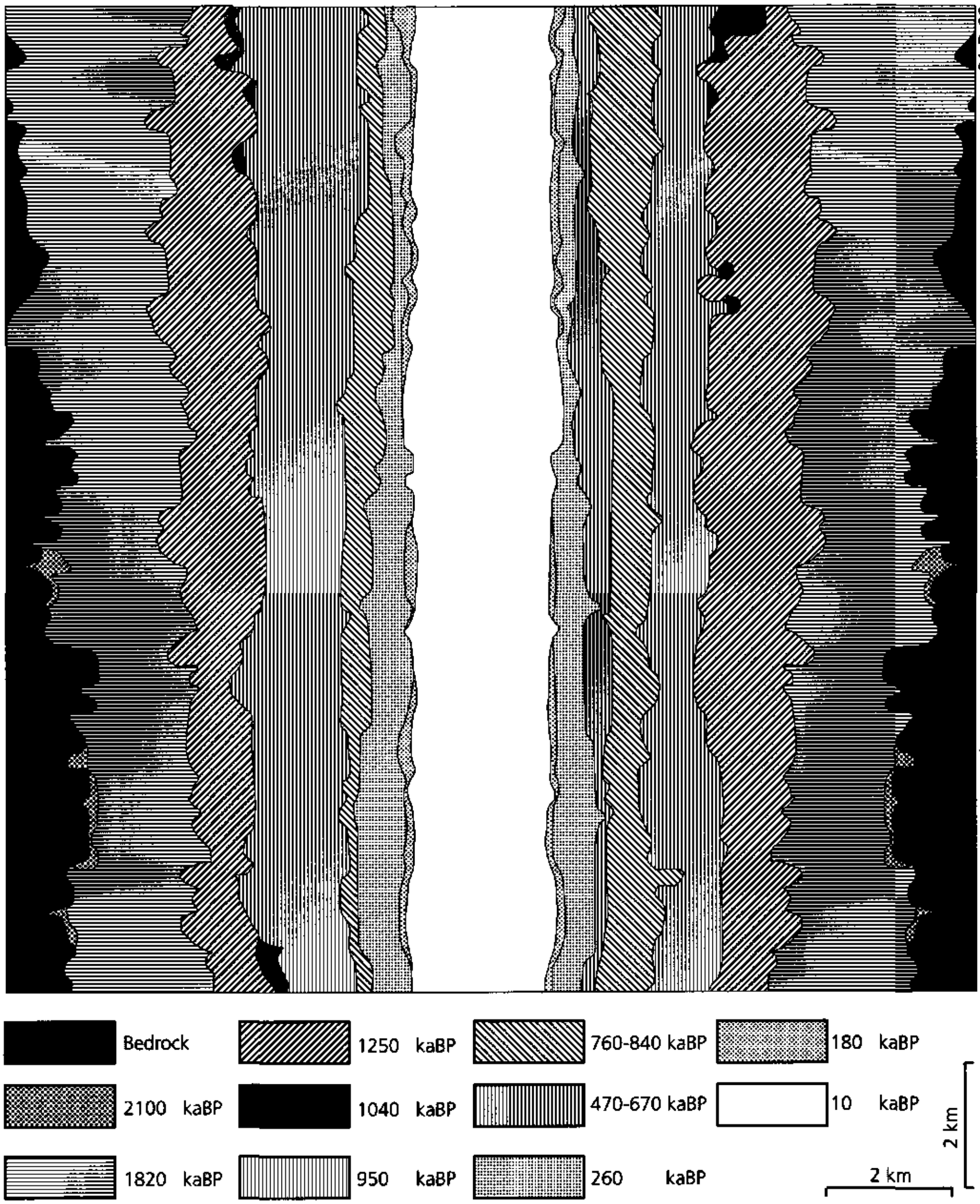


Figure 4.9 Age of surface sediments in the end relief.

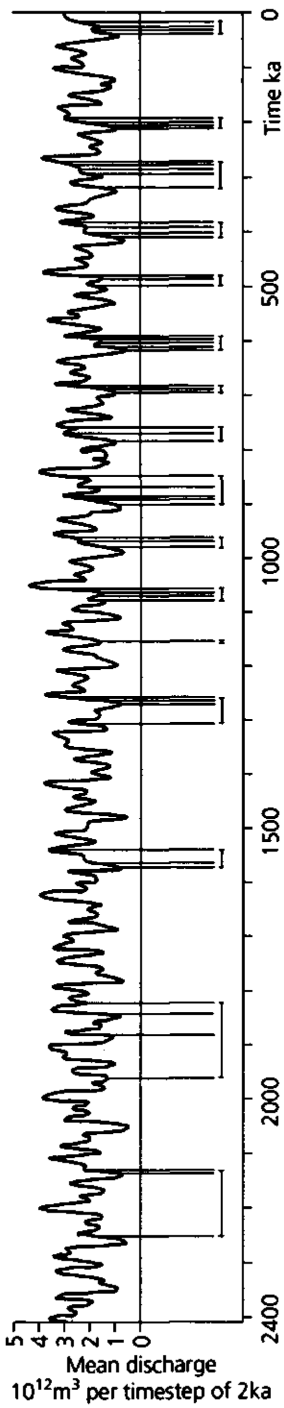


Figure 4.10 Terrace sediment ages on the Quaternary time-scale with the simulated 2 ka discharge. Horizontal bars indicate terrace bodies, vertical bars indicate sediment sequences separated by erosional phases.

iments seem therefore complementary to palynological records mainly recording the warmer climatic spells.

The general role of different uplift rates becomes obvious from the valley morphology as displayed in Fig. 4.7. The higher uplift rate between 780 and 250 ka BP caused a stronger incision shaping steeper valley slopes and smaller terraces. In Fig. 4.9 displaying the age of the surface sediments in the end relief, the general decrease in terrace area with age during the simulation is obvious. The slower uplift rate during the last 250 ka diminished the chances of a terrace to survive the erosional stages.

In Fig. 4.10 terrace sediment ages are given, on the Quaternary time scale, in relation to a simulated and strongly simplified climate curve. This curve serves as a proxy of 2 ka steps discharge variations. The lower mean discharges represent glacials and the higher interglacials. It can be seen that most sediments are deposited during glacials. Especially during the period with an uplift rate of 0.105 m/ka between 780 and 250 ka almost every glacial is registered as a terrace. This illustrates that under certain tectonic and climatic conditions a 'complete' terrace record is possible. A similar result was found for a comparable model for the Allier terraces (Massif Central, France). That model demonstrated that an apparently complete terrace sequence can have large time gaps (Veldkamp, 1991; 1992). The present modelling suggests that the older terraces represent a much longer time span and are polycyclic. This might suggest difficulties in correlations directly with the climatic history. This is in contrast to the results from chapter 3 where fitted functions suggest that such correlations reveal a smoothly developing uplift history. Below we will come back to this point.

Comparison of simulation results with the Maastricht terrace sequence

In Fig. 4.11, a simulated schematic cross-section after 2400 time steps (ka) has been plotted together with the schematic cross-section near Maastricht to compare the MATER simulation results with the actual Maas system. Both cross-sections display terraces as they occur along a few kilometre of the Maas projected on one cross section. It has to be realized that the Maastricht cross section in Fig. 4.11 is only a schematization of reality. The effects of loess cover, mass movements on slopes and the dissection of terraces by minor tributaries are not included in the schematic cross-section. Because MATER has only conceptual value the correspondence between the model and the Maastricht sequence must be considered qualitatively only.

The major points of agreement between MATER output and the Maastricht sequence are the number of terraces, general terrace sediment thickness, and the relative altitudes of the terraces on the valley slopes. There are also some clear differences between the model and reality. The difference in distribution of the older terraces and the number of younger terraces in the model and reality are the main diverging results.

The older terraces (above P-1) are more regularly distributed in the Maas valley than in the model. A possible explanation can be found in the change in lithology in which the Maas valley has incised. In reality the oldest terraces have eroded in Tertiary sediments whereas the younger terraces have incised in more resistant Cretaceous chalk. This differ-

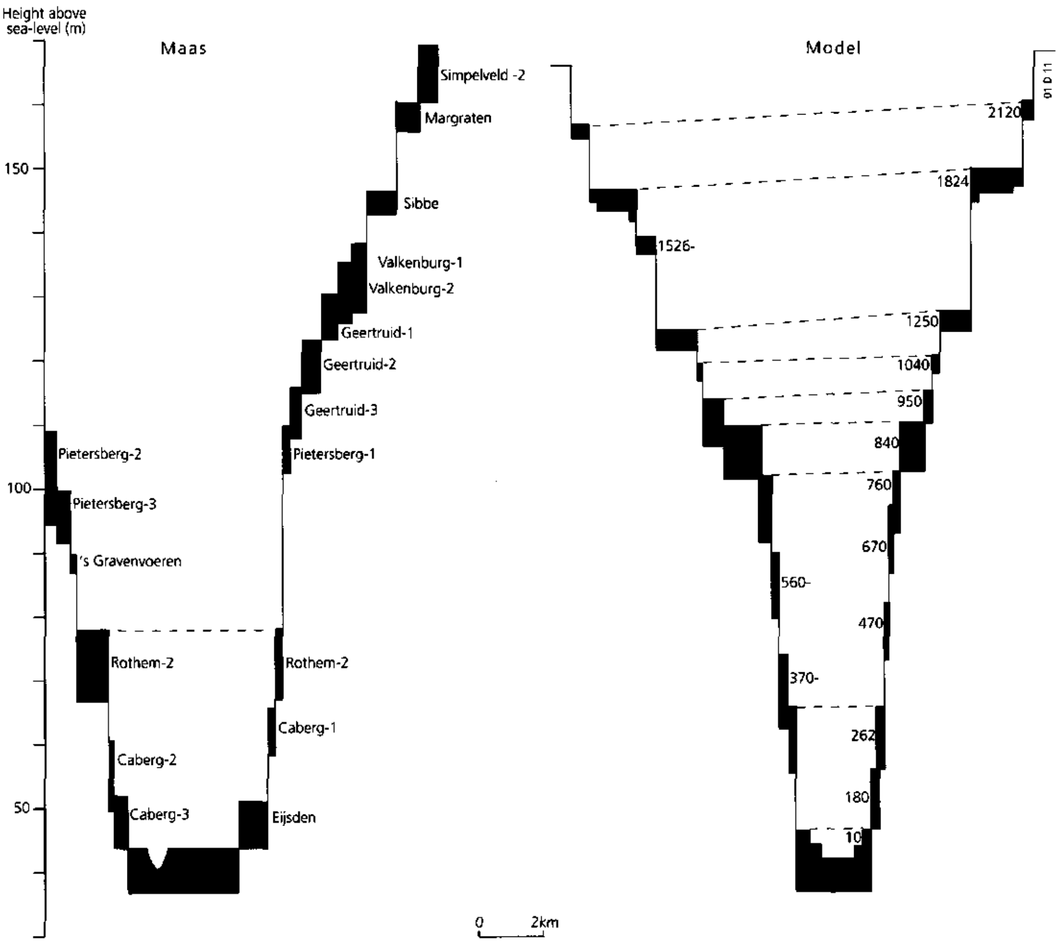


Figure 4.11 Comparison of the Maastricht terrace sequence and the model output.

ence in lithology was not incorporated in MATER. Besides lithology, climate and valley depth effects may also account for the observed differences. The difference in distribution between reality and MATER output, and the observed poly-cyclicity of the older terraces may be due to a valley depth effect. A fluvial system in a shallow valley has more lateral (erosional) dynamics than a similar system in a deeper valley. But one must realize that we also excluded an extra weight upon the harmonics of the 400 and 40 ka climatic cycles as found in deep sea record registering climatic changes older than 1 Ma.

The other difference between model and reality concerns the younger terraces. In the simulated valley only a few terrace remnants are found witnessing the last 200 ka. In reality the younger terrace records appear very complete. This difference may also be explained by an event not incorporated in the MATER model: the capturing of the Maas headwaters in the Vosges by the Rhine approximately 250 ka ago. This capturing caused a decrease in both discharge and sediment load.

Despite the discussed differences it is striking that relatively simple model inputs can already accomplish such a complicated terrace sequence. Despite the differences between the MATER output and the Maas terrace sequence the simulation shown is thought to display a possible scenario for the development of the Maastricht terrace sequence.
The simulation results clearly suggest that even an apparently regular and complete terrace sequence as found in the Maas valley at Maastricht is only partly complete. With the present data it is not possible to appreciate how much of the climatic variability is hidden within the sedimentary successions of each individual terrace. Only a part of the sequence at Maastricht from terrace level Geertruid-3 down to the Caberg-3 level (1000 to 250 ka BP) displays a 'complete' terrace chronology since they were formed in a period with stronger uplift rates than the older and younger terraces. In a 'complete' terrace sequence younger than 1 Ma the glacials of every 100 ka cycle have been registered by the occurrence of a terrace. The older terraces (> 1 Ma BP) in the simulation are polycyclic and thus might represent a longer time span than the terraces from 1 Ma to 250 ka BP.

Conclusions

Long-term simulations of the climo-tectonic dependent Maas system by the MATER model suggest that the Maas terraces at Maastricht are mainly the result of general discharge and sediment load ratio response to both climatic and tectonic factors. MATER illustrates that in the Maas system tectonism must have played a dominant role in determining the terrace preservation and the actual general valley morphology, whereas climatic dynamics seem to have strongly determined terrace formation by causing changes in water and sediment supply. The differences between model simulation results and the actual Maas system may be well explained by the strong simplifications of climatic en tectonic parameters of the simulated system. The observed differences may also indicate that variations in lithology and valley depth did also contribute to the terrace distribution in the Maas valley. A plausible explanation of the deviation between modelling output and the actual Maas system for the youngest terraces is the capturing of the Maas headwaters by the Moselle system 250 ka ago.

In general it can be concluded that a climo-tectonic dependent fluvial system can register Quaternary oscillations as terraces. Essentially, terraces are records of cold periods. The past and prevailing tectonic settings determine the sensitivity of the fluvial terrace record. It can therefore be concluded that a successful attempt to compare and correlate different fluvial systems can only be made when their tectonic setting and history is well known.

Co-author to chapter 5:

Antonie Veldkamp

5. Climate controlled alluvial sequences and para-sequences of the last 130 KA: A field and modelling study of the Lower-Maas record in the Netherlands

Abstract - *The internal architecture of a sequence of river Maas alluvium of Weichselian age shows a stack of four para-sequences. Their erosional bounding surfaces were found to occur floodplain-wide. Associated dated deposits allowed to relate these para-sequences to the Stadial-Interstadial climate couplets within the Weichselian. This match illustrates that nested hierarchical cyclic climate patterns can have their equivalents in river sediments. In order to investigate this relationship in more detail, we tested the concept within the MATER model which was calibrated already to a 2 Ma record of alluvial stratigraphy and valley morphology in the Maas valley. Simulations demonstrated that the stratigraphical impact of climate has been both direct and indirect. Direct by controlling alluvial aggradation and erosion and thus the formation of mayor bounding surfaces, indirect via extra-basin eolian sands brought into the floodplain, leading to more voluminous para-sequences.*

Introduction

In certain regions on earth climate change is of primary importance in controlling the variability of river systems on time scales covering several 100 ka cycles (Porter et al., 1992; Antoine, 1994; this thesis). In NW Europe aggradational terrace units reflect cold-stage aggradation whereas dissection matches interglacial conditions, in this way reflecting the 100 ka forcing frequency of climate change. This rule of thumb holds certainly true for the last 1 Ma.

To study the relationships of long-term fluvial dynamics with climate and tectonism Veldkamp and Vermeulen, (1989) constructed a model that operates with time steps of 1000-years and which is able to simulate terrace formation. Climate changes were modelled as the interference pattern of the three main periodicities basic to the orbital parameters. This simulation model was further elaborated and tuned to the last 2 Ma of the 4 Ma long record of the lower Maas in the southern Netherlands (MATER; Veldkamp and van den Berg, 1993). MATER simulations illustrated that the used combination of input parameters served as a fair approximation of past conditions.

Furthermore, it was demonstrated that the match between terrace stratigraphy and ocean records, in harmony with eccentricity- dominated climatic cycles, depends strongly on the tectonic setting. In rapidly subsiding areas the effects of climate are hidden in an often thick and extensive sedimentary record (Ruegg, 1994). Contrary to the rapidly subsiding areas hardly any fluvial record is preserved in rapid uplifting regions. In areas, upstream of the basin hinge line, where intermediate (uplift) tectonic velocities prevail, river terraces witness alternating aggradation- and erosion processes controlled by past climates.

Pattern analysis of long term climate variability suggests that its variation is composed of a hierarchy of nested cycles. Sedimentological analysis of Maas river terraces shows a full terrace unit (of approximate 100 ka) to be composed of a stack of units (para-sequences) bounded by erosional unconformities (Fig. 3.6), suggesting a similar nested principle for the terrace stratigraphy.

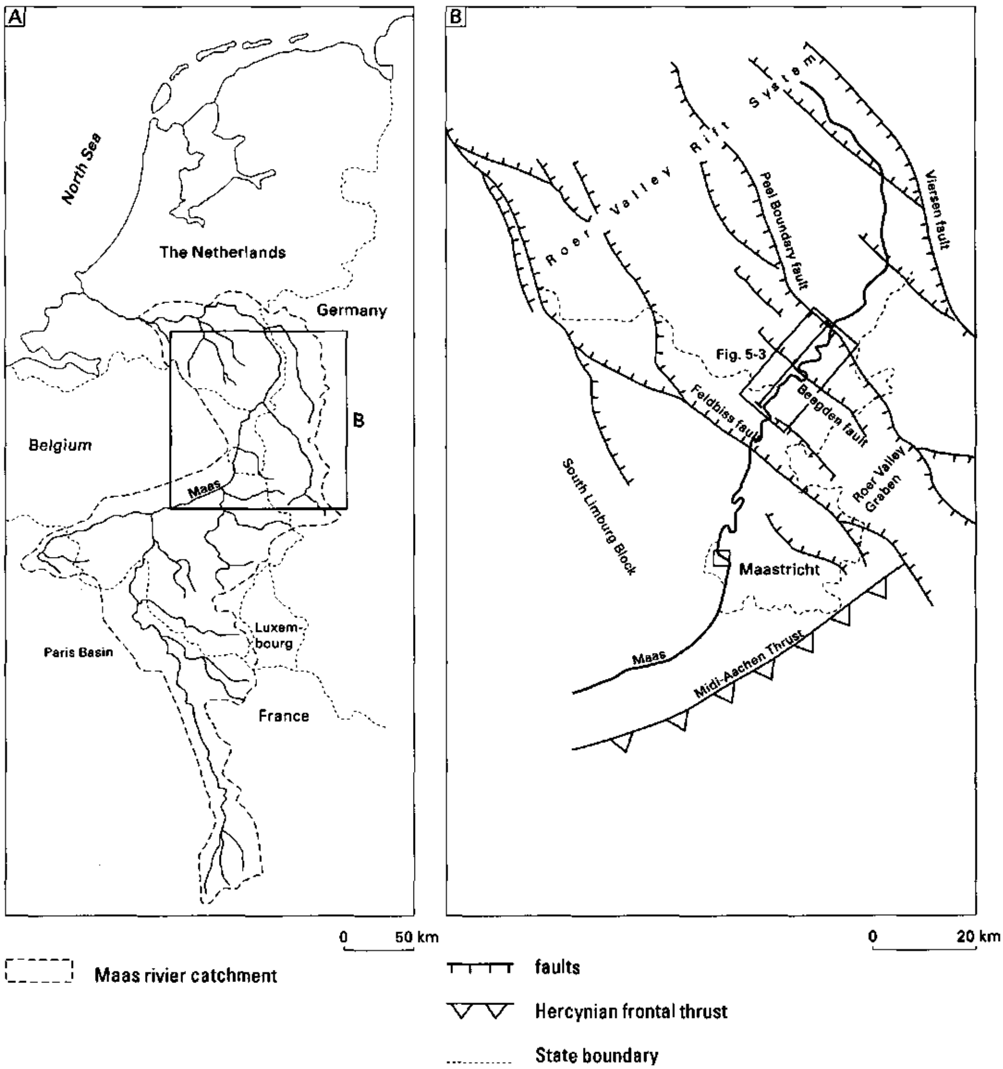


Figure 5.1 Maas river catchment area and the structural setting of the study area within the southern Netherlands.

In this paper we will investigate whether the nested climate cycles and the stacked pattern of para-sequences in fluvial terraces are linked. The research will be based on an integrated field and modelling study of the record of the last 130 ka.

Study area

In previous chapters we focused on the uplifting South Limburg block to study longterm terrace formation. We will now concentrate on the adjacent subsiding graben. Both sections have their own specific preservation conditions for the river Maas alluvium.
The sediment accumulation in the South Limburg section is generally less than equivalent

buried units in the downstream graben section. For example: the thickness of the Weichselian series increases up to two times in parts of the graben, giving better registration in the record. The average river gradient decreases from 0.75 cm/km in the uplifted section to 0.25 cm/km in the graben (Fig. 3.2). The lower gradient favours deposition of more fine-grained facies assemblages, yielding pollen records important for dating.
Pollen assemblages from the Late Saalian to Holocene, are relatively well known in northwest Europe (Zagwijn, 1961; 1975; van der Hammen et al. 1967). Moreover the climate evolution derived from these pollen records fits very well with climate records from other sources (marine and ice-core records) implying a well established climatical and chronostratigraphical framework for this episode (Behre, 1989; Guiot et al., 1989; Woillard and Mook, 1982; Vandenberghe, 1985). To study the erosional/depositional evolution of the river during one climatic cycle is therefore most promising for the period equivalent to deep sea stages 6 through 1 (cf. Fig. 5.4).
Probably the most complete river Maas sediment record for the last 130 ka is preserved in the middle section of the rift system: the Roer valley graben. Therefore data collection for this study was mainly restricted to the graben setting (Fig. 5.1).

Numerous high-quality boreholes of the graben area have been collected over the years by the Geological Survey of the Netherlands. Besides borehole information, data from quarry exposures and geomorphological inventories were available, they concern the period since the Late Pleniglacial (equivalent of the deep sea stages 2 and 1).
Deposits, positively attributed to an Eemian age (stage 5^e) are not known upstream of the basin hinge line. It is a common feature in the record of the lower Maas valley that (older than Holocene) interglacials register very poorly in the sedimentary record. This is explained by the morphology of an interglacial floodplain. Like the Holocene floodplain, such floodplains were deeply incised and strongly shrunken with respect to the previous glacial floodplains and therefor prone to erosion during the next glacial.

Fluvial sequence stratigraphy

The internal architecture of bedload dominated fluvial systems, commonly shows bounding surfaces. The better the contrast in grain size the better the recognition. Deposits of the lower Maas are relative inhomogeneous, bounding surfaces are therefore easily recognisable. Allen (1983) and later Miall (1988) developed a framework to classify bounding surfaces in fluvial sediments. Six levels were distinguished, ranging from contacts between sets (= first-order) to contacts between formations (= sixth-order level).
This hierarchy is well recognizable in exposures in various Maas terraces (Fig. 3.6). The sixth order levels (= terrace base and -top) reflect eccentricity (100 ka) related climatic changes.
Fifth order levels show up as erosive contacts (marked in Fig. 5.2) with associated conglomeratic sediments which are composed of large blocks, boulders, and rolled mud-clasts. We will focus our attention on these fifth order levels to determine their relationships with possible forcing frequencies higher than eccentricity dominated climate cycles.
Exposures in Maas terrace-sequences generally comprise a limited number (about three to four) fifth order bounding surfaces within one 100 ka cold climate sedimentary sequence. The bounding surfaces are usually undulatory in nature with an amplitude of a few metres.

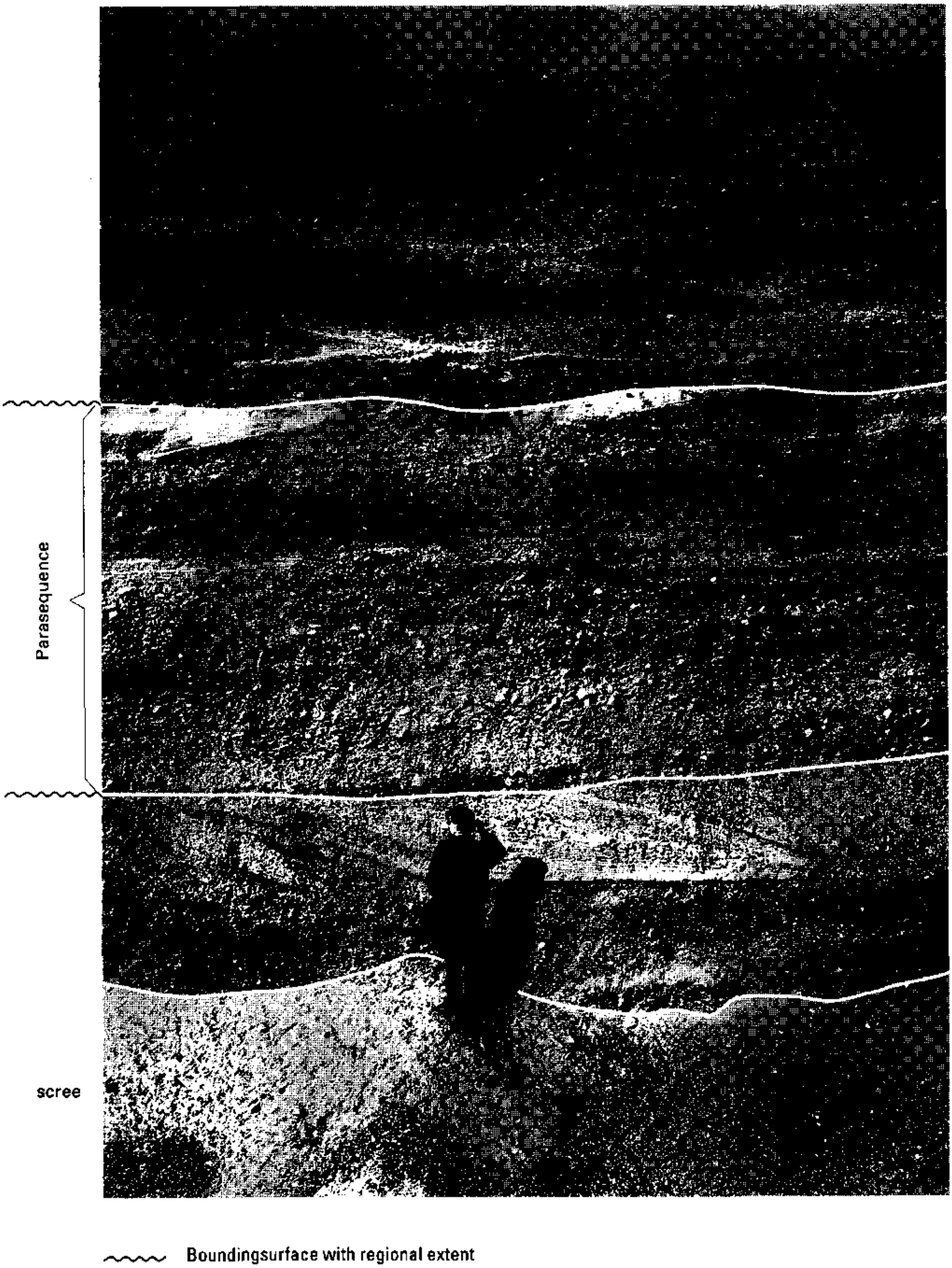


Figure 5.2 Cold-stage river Maas deposits are devided into units by bounding surfaces of regional extent. Maximal lithological contrasts mark these surfaces. They bound a parasequence (= FUS, see text). Photograph from the Pietersberg-2 terrace level as an example. Quarry Mol, Mechelen a/d Maas; Belgium.

The 3 - 6m thick units in between (= para-sequences), show an incomplete or poorly developed grainsize fining upward trend: more sand lenses and occasionally clay beds towards the top of the commonly gravel rich units (hereafter called fining upward sequence; FUS). These units have been deposited in a braided fluvial environment (see Fig. 5.2). Generally speaking they are, exposure-wide, laterally continuous but this is insufficient evidence to fully clarify the geometrical relationships between the various FUS at floodplain level.

To tackle this problem of parasequence continuity, we constructed several cross sections from available borehole data. The boreholes are spaced hundreds of metres up to a kilometre apart. These cross-sections cover several square kilometres of the Weichselian floodplain. They were chosen both perpendicular as well as parallel to the paleo-flow direction. Borehole data generally do not reveal bounding surfaces with limited lithological contrast, we thus assumed that strong lithological contrast in the cored sediments reflect fifth and sixth order bounding surfaces, as they do in exposures. Data on heavy-mineral assemblages (Zandstra, 1966) allowed to define the sixth-order levels (Fig. 5.3).
All sections indicate a lateral continuous stacking of four para-sequences: "fining upwards" units (I-IV) (Fig. 5.3). Their bounding surfaces are frequently marked by clay-rich horizons. It was of course not possible in every individual boring on the chosen sections to pin point the bounding surfaces of all four units, in such cases we used for correlation the abundance of scattered borings close to the sections.
The main exception to the longitudinal and lateral continuity is found where Late Weichselian/Holocene erosion locally removed underlying sequences.
The bounding surfaces are found at comparable depths over several kilometres long distances, thus our four FUS appear to represent, more or less tabular, floodplain-wide units. The coarse-grained sediment bodies can be interpreted to represent a period with abundant and large scale migration of channel bars: a braided environment. The subsequent deposition of clays marks a change to a more channel-bound regime with more limited discharge fluctuations.

Section description and interpretation

Geological age of the parasequences.

Humic clays found at the top of the FUS-1 (= the lowest para-sequence) showed Early Weichselian pollen spectra of the Amersfoort-Brørup complex (Zagwijn, 1961;1963) (Fig. 5.4). This Interstadial is the northwest European equivalent of oxygen isotope stage 5^c (Behre, 1989) (Fig. 5.5).
In most boreholes clay layers are absent at the top of FUS-2. Where encountered there was no independent evidence for its geological age.
The third unit has a well developed clay layer in the Roer valley graben (Fig. 5.4). Zagwijn, (1976) attributed the pollen spectra of the clay to the (Middle) Pleniglacial (stage 3). Its upper age is more precisely constrained by an associated peat layer found downstream in the Venlo graben. This peat revealed a ^{14}C age between 26 and 25 ka (GrN-16949 and GrN-16948).
The top of the fourth parasequence is part of the present day landscape. Pollen from organic-rich mud preserved in chutes in this level revealed an early Allerød age (Van den Broek

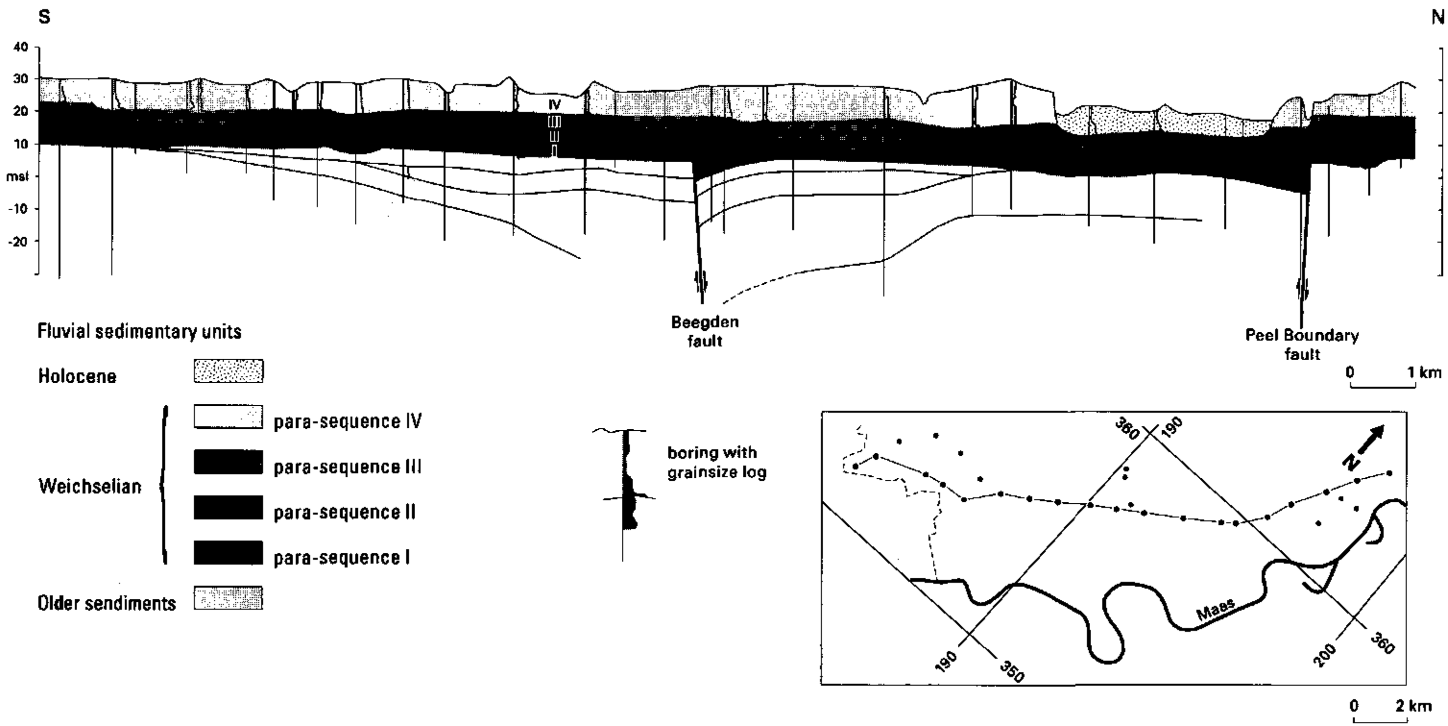


Figure 5.3 Longitudinal section based on borehole data though the Weichselian alluvial record in the Roer valley graben. Lateral continue boundary plains are marked by maximal grain size contrast (gravels on top of fines). They separate four parasequences (I through IV), together composing the full Weichselian aggradation record. For location see Figure 5.1B.

and Maarleveld, 1963; W. Hoek pers. comm., 1995). This shows that this level was abandoned during the transition of the Late Pleniglacial to the Late Glacial .

Summarizing, we found that the top and the bottom of the first and the fourth FUS are relatively well constrained in time (Fig. 5.4). The levels with fine-grained deposits clearly coincide with more temperate climatic conditions. The coarse-grained deposits could not be dated directly but they are believed to have been related to cold conditions during deposition. These cold-climate conditions during the aggradation were also witnessed by ice- and frost wedge casts in channel and overbank deposits of FUS 4. The overbank deposits of this unit consist of fluvio-eolian sediments indicating a strong influx of wind-blown sands into the floodplain during its aggradation (Chapter 6).

The presence of a discrete, and restricted number of floodplain-wide para-sequence boundaries together with associated fine-grained top-deposits, suggests that these boundaries represent periods of important facies shifts. Apparently large parts of the channel belt became temporary inactive and the abandoned channels in between bars became plugged with overbank deposited clays. Such facies shifts may have occurred regularly in the course of lateral channel shifting, but the record shows that the preservation potential of such local events was very low, unless they occurred floodplain wide and the activity zone of the main channel shrunk. Whether incision was associated with a reduction in floodplain width cannot be determined from our record; if so, it was not very deep (with respect to the depth of current floodplain). During the subsequent intermittent stages, the system became overloaded with bedload again and net aggradation continued.

We interpret the lithological sequence: fine-grained top deposits - erosive boundary - coarse grained bar deposits as a sequence of events that reflects the climate-change forcing on the overall style of alluvial sedimentation rather than avulsion controlled alluvial suites formed under a continuous tectonic subsidence. Below, we demonstrate that floodplain aggradation went faster than the average subsidence (Fig. 5.5).

Allogenic control on major bounding surfaces.

Generally bounding surfaces are not constrained in time giving way to interpretations in terms of autocyclic processes (e.g. Wizevich, 1993).

In this case, pollen and ^{14}C dating allowed us to assign the first and the fourth para-sequence to the first and the last stadial/interstadial oscillation within the Weichselian glacial period (Fig. 5.4). Taking these age models as an analogy for all the para-sequences, the second and the third para-sequence are supposed consequently to fit into the Early Pleniglacial and Middle Pleniglacial climatic couplet (Fig. 5.4). The position of parasequence II may be floating between stage 5b and stage 4. We tentatively correlate parasequence II with stage 4 because of the much harsher climate and longer duration of this stadial. Moreover the following Odderade interstadial stage 5a (Fig. 5.5) is rarely identified in fluvial deposits (pers. comm. J.de Jong). This may be the result of a low preservation potential due to the existence of narrow and incised floodplains during this temperate period.

A time/depth graph (Fig. 5.5) of this correlation gives an approximation of the net raising of the floodplain. A stepwise raising and lowering of the floodplain comes out. The graph shows that a substantial part of the vertical floodplain aggradation is lost to the incision phase at the end of the Weichselian. This will be discussed in the next chapter. The floodplain-height difference between the successive interglacial positions may be an

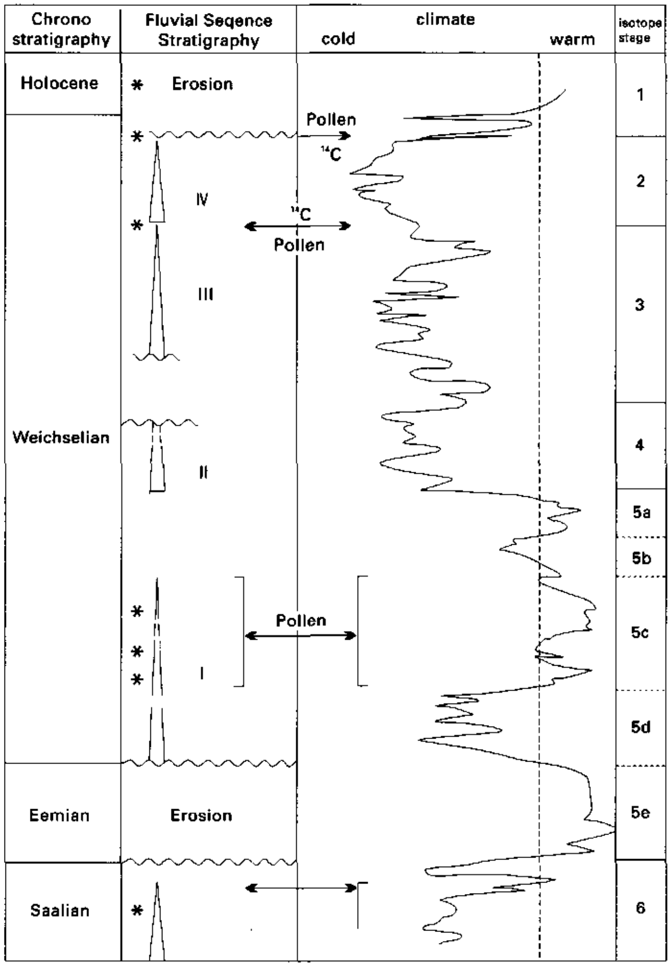


Figure 5.4 Comparison of the four alluvial parasequences (I-IV) in the river Maas alluvium with a northern hemisphere climate proxy (Pakitsoq-Greenland isotope record (Reeh et al., 1991)). The asterix mark the position of geological age controls; indicated correlation levels are discussed in the text.

approximation of the space created by subsidence (tectonic accommodation) for this site within the Roer valley graben. In other tectonic domains of the Roer valley rift system, the incision phase has brought the Holocene floodplain at or below the level of the top of parasequence 1. In both cases we suggest that sediment overloading of the system during cold stages exceeded creation of accommodation space, implying the allocyclic climate control of the alluvial dynamics. This conclusion of climate-change forcing on the fluvial dynamics is important because the major erosional (4th order) surfaces are common in channel deposits of other bedload dominated fluvial systems.

To further support the feasibility of our correlation and its interpretation we tested it with the MATER simulation model (Chapter 4; Veldkamp and Van den Berg, 1993).

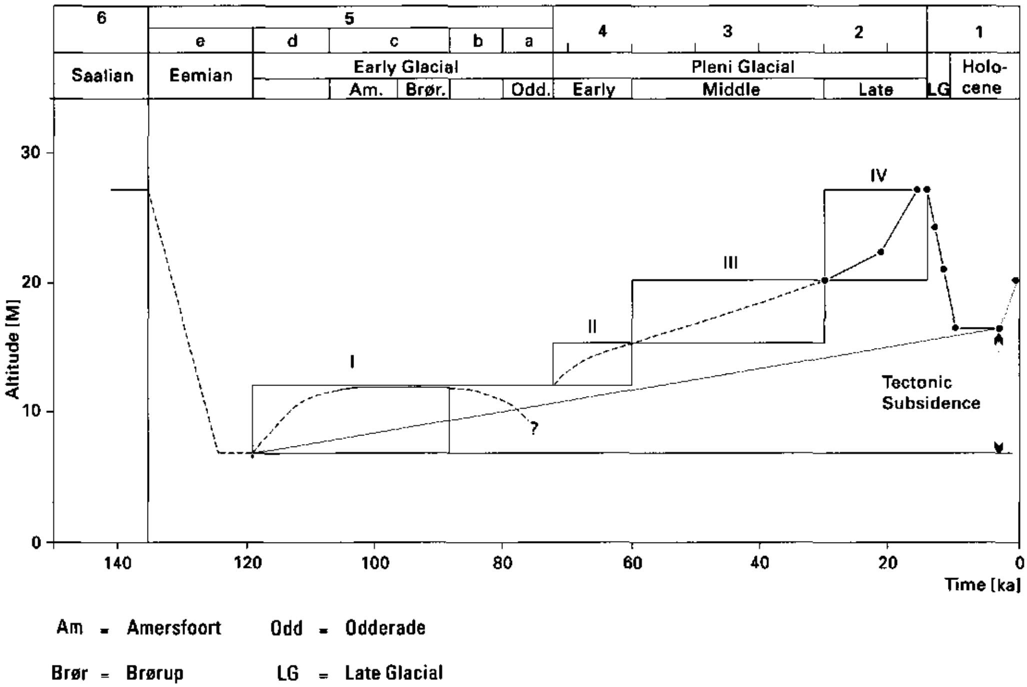


Figure 5.5 Parasequences put into a time/altitude graph. The boxes indicate the maximum available space for a unit. The drawn-dashed line indicates the (tentative) position of the floodplain surface. The altitudinal difference between the interglacial floodplain positions is interpreted as a measure for average tectonic subsidence.

Modelling of fluvial parasequences

The general principles of construction, organization and operation of MATER, a model simulating river dynamics (erosion and sedimentation) as a function of tectonism and climate have been described in previous papers (Veldkamp and Vermeulen, 1989; Veldkamp, 1991). MATER simulates the existence and activity of a fluvial system and not the internal river dynamics itself. The processes acting in the simulated fluvial system, are determined by decision rules extracted from hydrological principles (e.g. Schumm, 1977). The simulated erosion and sedimentation processes are up-scaled analogues of the real processes which respond to changes in the 2 ka discharge/sediment load equilibrium which are both defined as a function of climate. When the sediment input load exceeds the sediment transport capacity, the latter is a function of discharge, the difference is deposited, and in case the erosion and transport capacity exceeds the input load, the difference is eroded in the simulated system.

Because MATER was already calibrated for a 2 Ma Maas terrace sequence (Veldkamp & van den Berg, 1993), no new calibration is necessary. Extensive sensitivity analyses demonstrated that the model is sensitive and robust enough to simulate fluvial dynamics for the last 100 ka (Veldkamp, 1993).

Driving forces: The rate of climate change determines its effect on discharge distribution and sediment load of a fluvial system (Lowe and Walker, 1984). A relationship is assumed

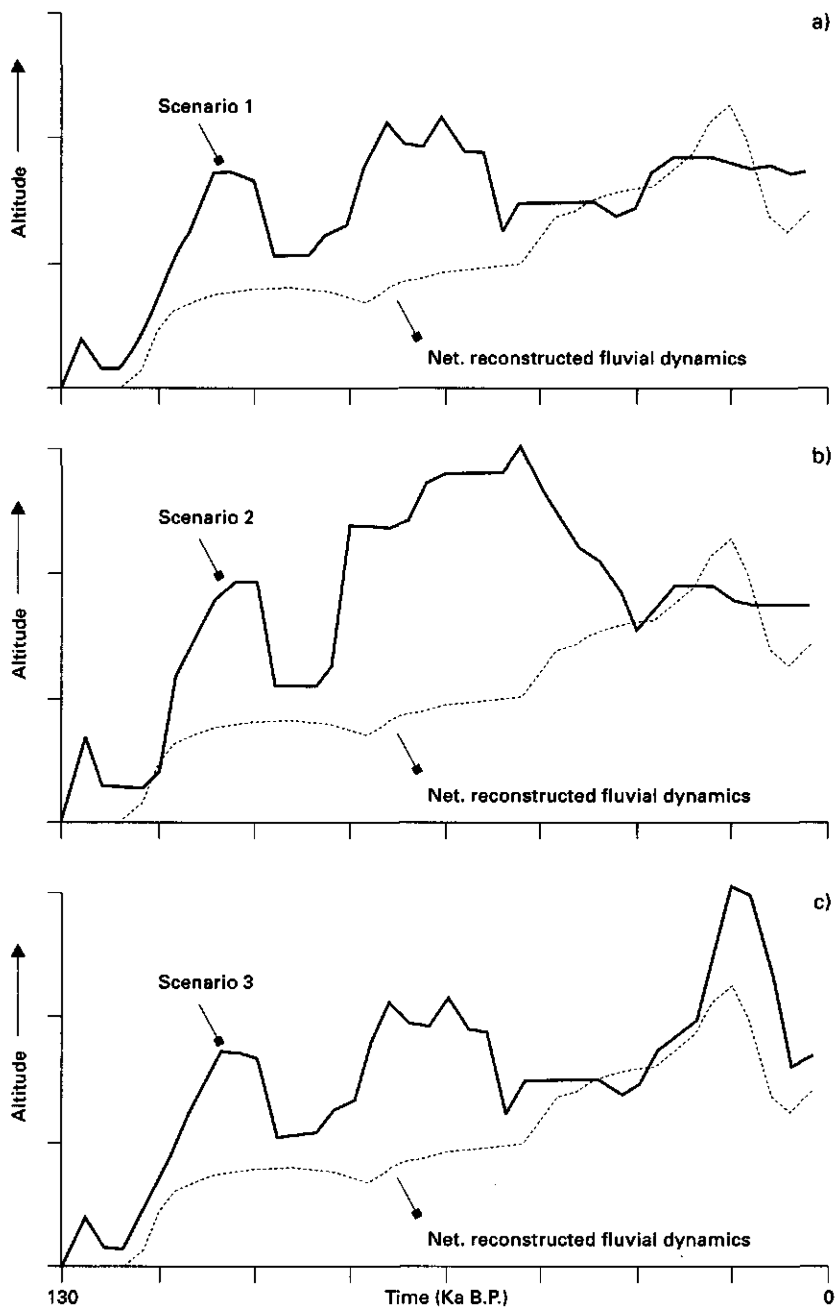


Figure 5.6 Build up of parasequences an alternation of aggradation and erosion during simulation presented as a time/altitude graph (sign convention: up = aggradation; down = erosion). The drawn-dashed line indicates the (tentative) position according to the reconstruction of Figure 5.5. The fat line shows the simulated erosion and aggradation.

between waxing and waning of continental ice-sheets, as expressed by isotope records, and mean discharge/sediment load changes. The main effect of ice-sheets to river discharge is by controlling the position of the polar front which in turn regulates the pathways of depression tracts and thus precipitation in river basins. This acts in conjunction with the temperature and precipitation controlled shifting of vegetation belts which govern the sediment supply by changing surface cover.
These climate changes appeared in earlier model research by the second author, to be an important driver of long term fluvial dynamics. The use of time steps of two thousand years holds that our climate function should have a resolution less than two thousand years but should show sufficient detail to allow the model to compute the transient response of the alluvial dynamics to the climate forcing. In this case we choose the Pakitsoq Greenland "ice core" function (Reeh et al., 1991) as a driving function. This record was believed to show just the right resolution of Northern Hemisphere ice-volume signal for the period under discussion. Other records like the GRIP core (Dansgaard et al., 1993) show too much detail in temperature change whereas deep sea records are too coarse.
Besides the difference in fluvial dynamics between glacial and interglacial periods, the transition from a glacial to an interglacial environment was also taken into account. This operates when a glacial lasts longer than 10 ka. Then, an extra sediment flux of 1.2 times a normal flux is supplied to the fluvial system during one timestep. This simulates instability in the Maas headwaters.

Simulation results

Within the discussed simulation scenarios the used tectonic setting and rate was derived from the Fig. 5.5 (The floodplain-height difference between the successive interglacial positions was used as an approximation of the subsidence rate). So instead of uplift (the tectonic setting of the 2 Ma terrace flight) a gradual subsidence was simulated. The results of three different scenarios simulating the vertical floodplain-dynamics (erosion-sedimentation) in a subsiding setting are given in Fig. 5.6.

Scenario 1: A standard simulation with equal climate model settings as used for the 2 Ma long Maas terrace sequence at Maastricht was made (Fig. 5.6*a*). This MATER simulation describes rather well the climatic and tectonic effects on the 100 ka alluvial stratigraphy and match field data for the FUS 1 and 2. The net aggradation phases of FUS 3 and 4 are represented only as a minor features during this simulation, whereas in the field the sediments from these periods form a mayor component within the whole Weichselian floodplain aggradation.
Scenario 2: To check whether this difference between model and field data finds its origin in an incomplete tuning, we doubled the sediment flux released during instable conditions due to major climatic changes in the Maas basin. This modification lead to no visible improvement in the simulation of FUS 3 and FUS 4. On the contrary the erosional discontinuity between FUS 2 and 3 did not form, and FUS 4 remains a minor feature in the total sediment record (Fig. 5.6*b*). It was consequently concluded that the overall improvement of the simulation was insufficient to accept these new conditions as more valid than those of the first simulation. Because MATER was already tuned to a Maas section it was thought to be unrealistic to change other model para-meters. Subsequently we concluded that the first MATER scenario displays the most likely simulated climate impact on fluvial dynam-

ics in the Maas system. Still the observed anomaly of the simulated and real world FUS 3 and 4 remained to be explained.
Pollen-based reconstructions of the Late Pleniglacial environment around the southern North Sea suggested this to have been a polar desert (Van der Hammen et al., 1967). Outside river plains extensive (loam-free) sand sheets were formed during this period under absence of vegetation and concordant with pre-existing regional topography (Ruegg, 1983). Since the eolian sands within the Maas valley have a northwestern and downstream provenance, their effects can not be automatically simulated with MATER. To simulate their possible effects on the fluvial dynamics we had to assume an extra eolian sand flux during the Late Pleniglacial (during the formation of FUS 4) leading to scenario 3.
Scenario 3: A simulation with standard MATER model settings incorporating the extra eolian sands fluxes during the Late Pleniglacial resulted in a section with all four FUS's. This result indicates that the FUS are most probably the result of climate controlled fluvial activity with changing contributions of eolian sediments. The youngest FUS 4 can only be simulated by the combined result of climate driven fluvial dynamics and climate controlled eolian sand inputs. In reality the FUS 3 unit was also formed in an environment with eolian sediment additions too. We did not simulate this addition because we have insufficient data on the eolian fluxes in time to make realistic simulations.
The three scenarios and the match between model output and field evidence (Fig. 5.6), without a new calibration of the MATER model, confirms the fact that MATER does reasonably well simulate the general Quaternary Lower Maas dynamics. More detailed simulations are for the time being not feasible with MATER because the time step resolution is to inaccurate to allow more detailed simulations. Furthermore, a detailed (2 ka resolution) aeolian sediment flux curve is required to allow realistic model simulations of the Lower Maas during the Weichselian.
Simulations indicated that indeed climate induced variations in river discharge and sediment load can produce a similar para-sequence stack as observed in the field. The agree between the model results and the stratigraphical correlations stresses that this sedimentary record may serve as an example of allocyclic controlled fining upwards cycles in fluvial systems.

Conclusions

We demonstrated that during the Last Glacial/Interglacial cycle, subordinate climate cycles forced by the interference pattern of the three (100 ka, 41 ka and 23 ka) Milankovitch cycles are reflected in the alluvium of the Maas river by a stacked pattern of parasequences together composing one terrace sequence. This evidence presents strong arguments for the statement that a close coupling exists between the forming of fifth-order bounding surfaces and climatic change: an allocyclic control on fluvial architecture.
The modelling study with MATER indicated that the impact of climate can be both direct and indirect.
Direct climate impact influences the aggradation and erosion processes through discharge and sediment load control, while the input of extra-basin aeolian sands represents an example of indirect climate control.
Based on the river Maas results we expect similar patterns and relationships to be found in other northwestern European fluvial systems due to their similar history and tectonic set-

ting. Especially the potential impact of eolian fluxes on regional fluvial dynamics and the resulting stratigraphical record can be expected to have played a significant role in the broad coversand and loess zones in Europe.

Co-author to chapter 6:

Jacques C.G. Schwan

6. Millennial climatic cyclicity in weichselian late pleniglacial to early holocene fluvial deposits of the river Maas in the southern Netherlands

Abstract - *In the southern Netherlands, well beyond the basin hinge-zone, the evolution of the river Maas was controlled by tectonism and climate change.*
In the study area, the Maas crosses the central graben of the Roer Valley rift system. Here, an alluvial fan was built up at a rate exceeding subsidence and this permitted stepwise incision and greater terrace differentiation than in the Maas-valley upstream of the southern peripheral Feldbiss fault. The Maas river system has responded to periodical climate change on Weichselian to Holocene, Weichselian stadial-to-interstadial and millennial time scales and the corresponding fluvial cycles find expression in the terrace morphology and deposits of the river. The millennial-period fluvial cycles are of two types. The erosional type is of Weichselian Late Glacial to Early Holocene age and consists of two successive terrace-levels and a planation surface now buried under deposits of the modern floodplain. The aggradational type dates to the Weichselian Late Pleniglacial and is represented by a sequence of fluvio-aeolian beds. Each of which is less than a metre in thickness and bounded by a periglacial surface. This deposit is interpreted as a succession of drying -up/cooling-up cycles with mean period of 1.4 ka. The cycles supposedly were driven by periodical changes in the strength and position of storm-tracks on the North Atlantic.

Introduction

Changes in river systems are related to climate changes, tectonism and sea-level fluctuations (Lowe and Walker, 1984; Gibbard, 1988). The influence of the last factor decreases in an upstream direction and beyond the basin hinge-zone its effects are fully outweighed by those of the other two determinants. At the higher latitudes of the Northern Hemisphere it was in the first place the waxing and waning of the Pleistocene ice sheets which caused an often profound modification of the fluvial environment. In mid-latitude regions beyond the reach of the glacial meltwaters, change of climate had a different effect on fluvial systems. Here, it acted more indirectly upon river behaviour by its control of variables such as precipitation, vegetation density, absence or presence of permafrost, weathering regime, discharge pattern and availability of detritus. In glacial times, an excessive supply of mainly coarse rock fragments causes aggradation by braiding rivers. During interglacial periods, on the other hand, the sediment load of the streams is both dominated by fines and generally lower in particle content. As a result, erosion will prevail in the river portion upstream of the hinge zone whereas deposition of fine-grained sediment takes place in the lower reaches of the river which are under the influence of the rising sea level (Törnqvist, 1993). The associated channel style is either straight, anastomosing or meandering. In the last case, the meanders may occur only locally or all along the course of the river. Among other things, this depends on the pattern and intensity of crustal fragmentation resulting from tectonic activity (Chapter 8; Van den Berg et al., 1994).

The territory of The Netherlands is located at the southern tip of the North-Sea basin. Largely due to this, the Quaternary evolution of the Dutch Maas-Rhine river system took

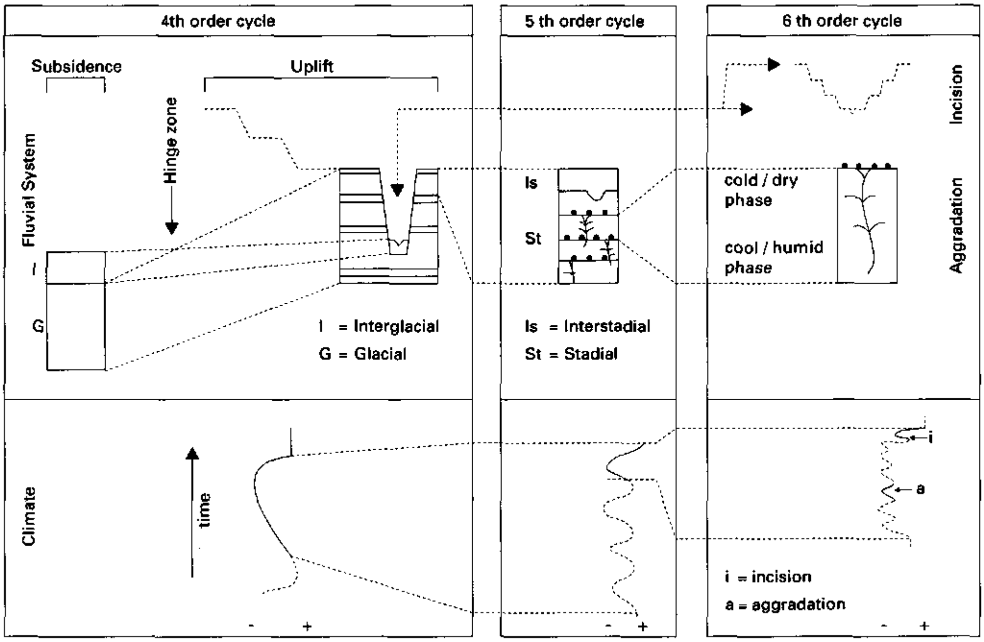


Figure 6.1 Hierarchy of climatic cycles and corresponding expression in the fluvial system.

place in two different settings (Ruegg, 1994). In the subsiding part of the fluvial basin, more or less continuous deltaic deposition left behind stacks of fining-upward sequences corresponding to climatic cycles. Upstream of the hinge zone ('terrace intersection') where uplifting prevailed, river terraces were formed. In this situation, climate change manifested itself by aggrading of coarse-grained sediment in cold times succeeded by downcutting mainly in the period of transition from cold to interglacial or interstadial conditions (Veldkamp and Van den Berg, 1993). Commonly, the end of the aggradation process is marked by a level of fine-grained channel fill on top of the sands and gravels.

For the formation of the type of terrace discussed above, two requirements must be met. These are: (i). A climate-controlled change in the mean bedload concentration of the riverwater and (ii). A relative lowering of base level. This can be brought about in two different manners viz. by tectonic upheaval or by a rate of sediment-aggradation exceeding subsidence where the landsurface is subject to downwarping. In time series representing Quaternary climatic change, small wiggles are superimposed on larger ones and these, in their turn, on still larger ones. Thus, we have to do with a hierarchy of climatic cycles with increasing period. At the highest level we find the glacial-to-interglacial or fourth-order cycles. These are followed by stadial-to-interstadial or fifth-order cycles. Next and last in the context of this paper we have sixth-order cycles with relatively short phases of contrasting climate (Bond et al., 1993). A similar hierarchy of cycles has been found in the sedimentary sequences laid down by the river Maas (Fig. 6.1). At the uplift rate prevailing in the Maas-catchment, the main river terraces are the expression of fourth-order climatic cycles (Chapter 3). Each aggradational terrace unit consists of several fining-upward sequences corresponding to fifth-order climatic cycles (Chapter 5).

This paper discusses a succession of sixth-order climatic cycles which ran their course in the lower reaches of the river Maas (Meuse) in the southern Netherlands during the time period from the Weichselian Late Pleniglacial to the Early Holocene. As will be explained below, these cycles happened to be recorded at a high level of resolution in (i). Fluvio-aeolian overbank deposits laid down during aggradation phases and (ii). A flight of low terraces formed during periods of incision.
The number designation of the cycle orders in the preceding section is the one introduced by Haq et al. (1988). The hierarchy proposed by these authors consists of six orders of cyclic sea level change with decreasing period.
Their first-, second- and third-order cycles, though lacking regular periodicities, have durations in the hundreds of millions to one million year range and are mainly related to continental drift and other tectonic processes. The lower order cycles, on the other hand have periods much less than one million years. It is generally accepted that at least the fourth- and fifth-order cycles must be attributed to astronomical forcing of the climate, i.e. to variations in the seasonal and latitudinal distribution of insolation caused by rhythmic changes in the earth orbital parameters (Plint et al. 1992). Although it is not questioned that the high frequency cycles of sixth order are a result of climatic change, it remains to be seen whether periodical astronomical perturbations on millennial time scale should be the ultimate cause of these fluctuations.

The Maas river system

From its headwaters region in eastern France to Grevenbicht in The Netherlands the Maas flows through successively the Jurassic beds of the Paris Basin, the Hercynian Ardennes Massif and the Cretaceous to Tertiary rocks of the South Limburg Block (Fig. 6.2). Between Grevenbicht and Mook the river crosses the fault blocks of the NW-SE trending Roer Valley rift system, which is the southern extension of the North Sea Basin (Ziegler, 1994). In this tectonically active zone, differential vertical movement goes on to the present day. Due to this, the transition from the rising South Limburg Block to the Roer Valley graben is marked by the following features:

1. A sharp break in river gradient from 0.75 m/km to 0.25 m/km downstream of the Beegden fault (Fig. 6.3),
2. A rapid downstream widening of the Pleistocene floodplain halfway between the Feldbiss and the Beegden faults. The outline of the fossil floodplain is demarcated by the terrace scarps which can be traced across the various fault blocks,
3. An abrupt change in lithology of the shallow subsoil from fluvial gravelly sands upstream to well sorted eolian and fluvio-eolian fine sand downstream. Preservation of sediment in the subsiding Roer Valley graben is demonstrated by the fact that the second sediment type corresponds to a younger depositional unit than the other one,
4. A pattern of inactive shallow gullies having their origin some distance downstream of the Feldbiss Fault. Probably, this feature represents the surface of a small alluvial fan which formed at the foot of the rising South Limburg Block.

Beyond the hinge line of the North Sea Basin (Fig. 6.3), the Maas enters the Holocene lowland floodplain which it shares with the lower-course distributaries of the Rhine.

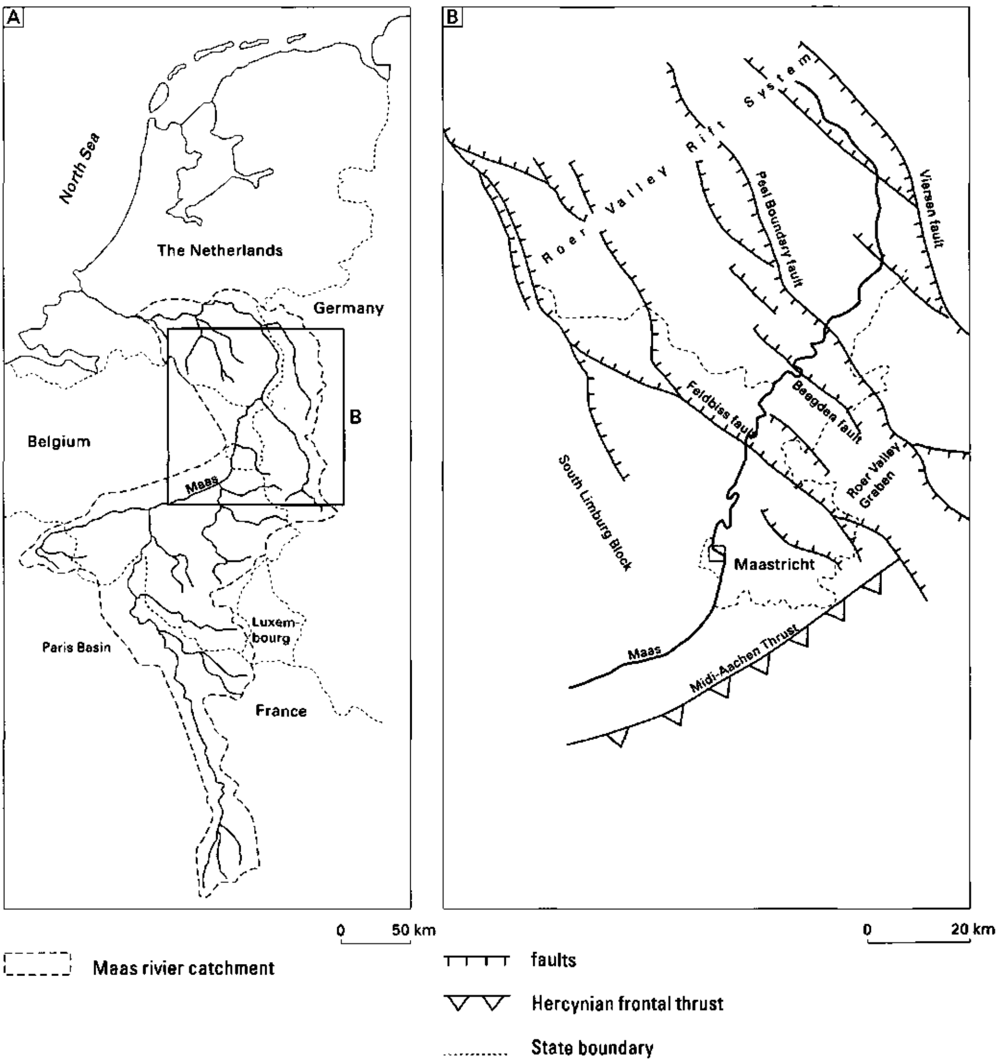


Figure 6.2 A. Catchment area of the Maas river, B. The Roer valley rift system.

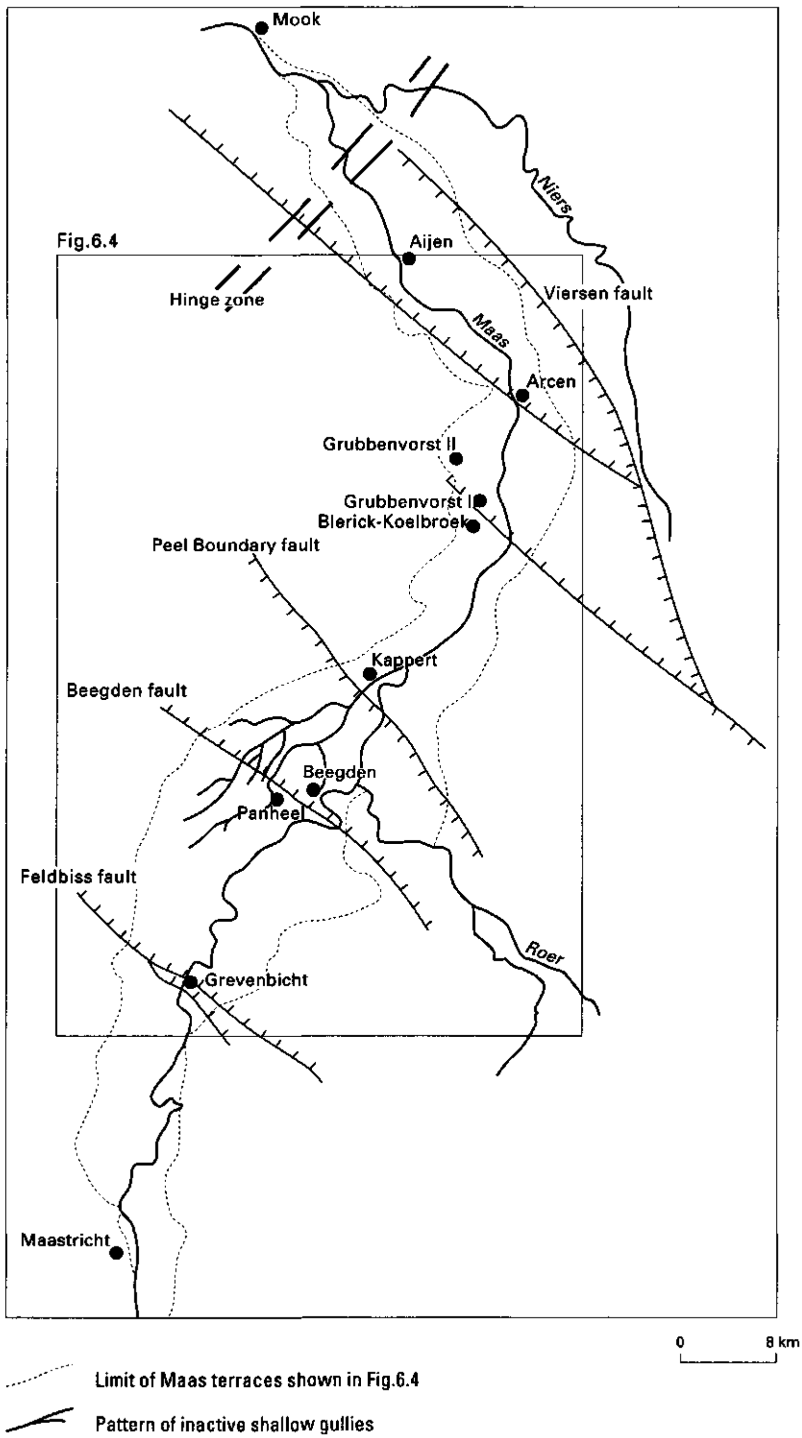
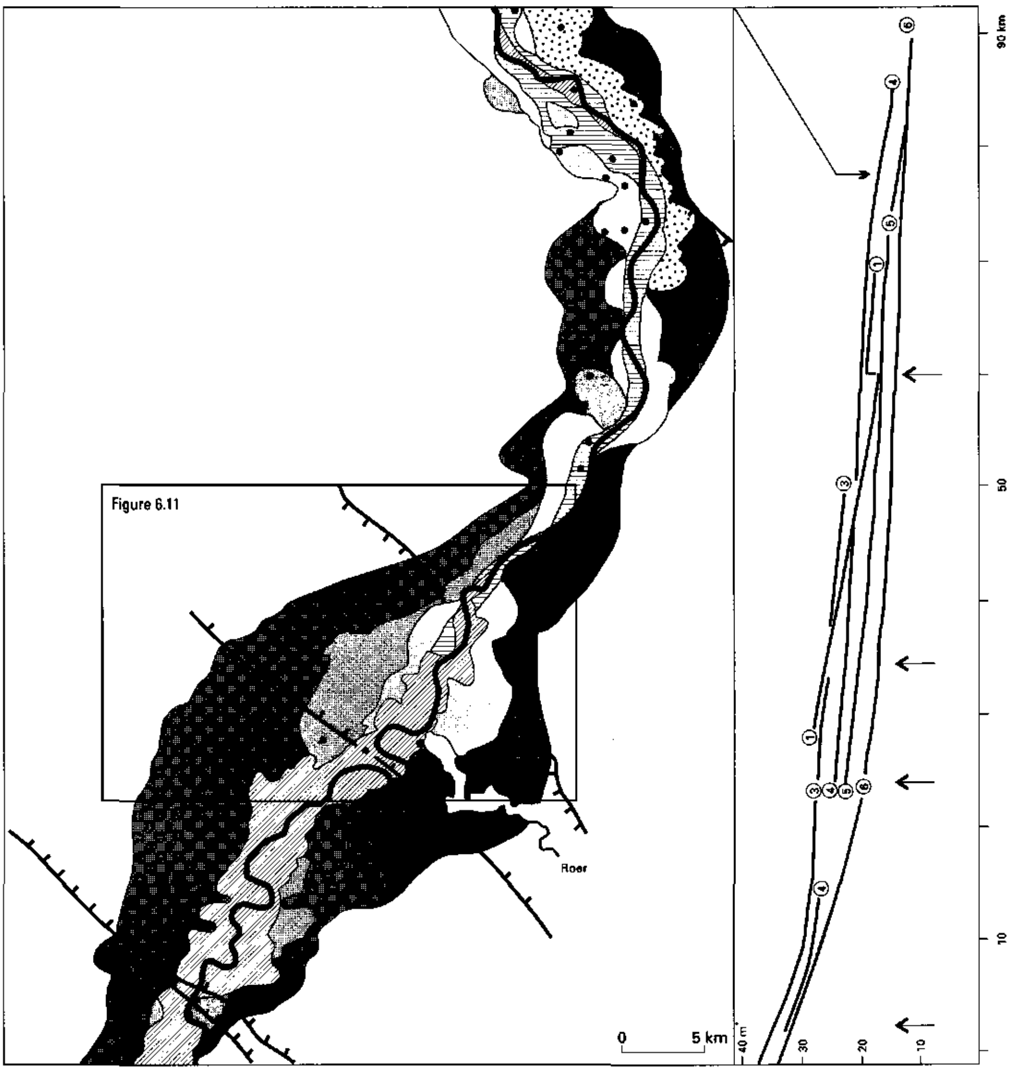


Figure 6.3 Features and locations in the southern Netherlands which are mentioned in the text.



Parabolic dunes of younger Dryas Stadial age

Sample site

Fault trace

Fault

Figure 6.4 Terraces of the Maas and their longitudinal gradients in the southern Netherlands between Grevenbicht and Aijen.

Legend to Figure 6.4

Topographic level	Feature		Pleistocene aggradation	Time of fluvial erosion	Beginning of floodplain abandonment or burial	Holocene aggradation
6	Modern flood plain covering terrace 6	Cut terraces		Early and Middle Holocene	Transition from Middle to Late Holocene	Late Holocene
5	Terrace 5 with braiding surface pattern			Younger Dryas Stadial	Preboreal	
4	Terrace 4 with meandering surface pattern			Allerød Interstadial	Younger Dryas Stadial	
3	Terrace 3 with meandering surface pattern	Fill terraces	Weichselian Early Glacial + Pleniglacial		Early phase of Allerød Interstadial	
2	Aeolian beds covering terrace 2				Late phase of Weichselian Late Pleniglacial	
1	Terrace 1		Saalian glaciation phase		Transition from Oxygen isotope stage 6 to 5	

The terraces of the Maas

Geomorphology

Of particular relevance to the subject of this chapter is the part of the Maas-valley located between Mook to the north and Grevenbicht to the south (Fig. 6.3). Here, the effects of tectonism and climate are well expressed in alongstream variations in channel pattern, floodplain width, stream gradient, depth of incision and terrace differentiation. A similar variation, both within and between terrace levels is also present in the details of the river morphology, i.e. in features such as terrace scarps, abandoned channels, fluvial bars and aeolian landforms.

As shown in Fig. 6.4, the terraces in the subject region are identified by topographic levels. Each topographic level corresponds to a surface of either a terrace proper or a deposit which has buried the original terrace. Accordingly, six different terraces could be distinguished. These terraces are of two types, viz. fill terraces and cut terraces (cf. Merritts et al., 1994). The fill terraces have a surface formed by fluvial aggradation and this original upper level of alluviation may be with or without a cover of aeolian sand. These terraces occupy the highest positions in the terraced valley and their planform is characterised by long and gently curved terrace scarps.

The cut terraces have a top surface resulting from downcutting and planation by the river. In the valley reach under consideration, cut terraces occur in a flight carved into previously aggraded, mainly sandy sediment. In contrast to the first category, the cut terraces are bound by successions of short and rather strongly curved terrace scarps. Horizontal tectonic motion may have codetermined the morphology of the cut terraces as will be discussed below.

Age of the terraces

In general, the age of terraces is taken at the moment of the beginning of floodplain abandonment, i.e. by the beginning of the retreat of the riversystem from the former floodplain. Ideally, this point in time should be determined on the earliest fill of fossil channels left behind on the terrace surface.

The chronostratigraphy given in Figs. 6.4, 6.8 and 6.13 is based on pollen ages and ^{14}C ages compiled from Van den Broek & Maarleveld (1963), Teunissen (1983), Westerhoff & Broertjes (1990), Vandenberghe et al. (1994) and Kasse (1995). In a number of cases a relative dating was obtained by extrapolation of the long profile of a terrace of known age to one of unknown age.

Terrace 1 dates back to the Saalian glaciation phase which corresponds to the last part of oxygen isotope stage 6. In downstream direction, this fill terrace becomes increasingly covered by fluvial deposits of terrace 2 age and these, in turn, by parabolic dunes of Younger Dryas age. *Terrace 2* is a fossil braidplain abandoned in Weichselian latest Late Pleniglacial times and subsequently covered by loess and aeolian sand. The upper fluvial layers of this fill terrace are of Late Pleniglacial age with the overlying windborne strata having ages ranging from latest Late Pleniglacial at their base to Late Glacial at their top (this terrace is not shown in Fig. 6.4 for reasons that the windborn surface relief prohibited to find sufficient reliable data on the altitude of the interface of fluvial- vs. wind born strata).

Terrace 3 is the highest terrace with meandering channel pattern. Since the surface of this fill terrace and the buried braidplain of terrace 2 are on one level, terrace 3 is interpreted as the riverward extension of terrace 2, implying neither aggradation nor downcutting during the aeolian transgression from the west. The scarp separating topographic levels 2 and 3 marks the limit of the aeolian transgression. Abandonment of terrace 3 floodplain began in a early phase of the Allerfd Interstadial.

Terrace 4 is another terrace with meandering surface pattern. Counted from above, it also represents the first cut terrace resulting from incision and subsequent lateral erosion. These processes took place during the Allerfd and were concluded by floodplain abandonment beginning in the Younger Dryas Stadial.

Terrace 5 is a cut terrace with braiding surface pattern. The terrace surface formed during the Younger Dryas with floodplain abandonment starting in the Preboreal.

Terrace 6 is a cut terrace covered by deposits of the modern floodplain. Fluvial erosion supposedly began in the earliest part of the Holocene and was followed by equilibrium flow (steady-state flow) during an extended period of time (cf. Fig. 6.13). Alluviation, largely as a result of deforestation by man, began to take over around the close of the Subboreal. The history of the buried planation surface rests on a limited number of borehole data. In core Rijckholt (De Jong, 1984), the base of a hiatus of 3.8 ka is dated at 8.610 14C yr BP (GrN-11316) corresponding to roughly 9.600 calendar (cal) yr BP. This time point is supposed to mark the incipience of an interval of fluvial erosion and subsequent steady-state flow. In core Beegden (De Jong, 1970), the clayey and organic matter containing top of a fining-upward sequence is overlain by sandy gravels. This abrupt break in fluvial regime is pollen dated at the transition from Late Subboreal to Subatlantic and may represent the beginning of Holocene aggradation. Other pollen records, however, suggest that this event has an older age and should be placed in the Subboreal Chron.

Tectonic effects

Apart from the obvious situation of the Venlo graben controlling the orientation of the Maas valley between Arcen and Mook, two other tectonic effects in the subject region are discussed now.

1. In the Maas valley upstream of the Feldbiss fault there is only one terrace level in between the surface which is age-equivalent with terrace 3 and the modern floodplain. In

contrast to this, in the stretch downstream of observation site Panheel two terrace levels are found between terrace 3 and the modern floodplain. These are terraces 4 and 5 of Fig. 6.4. This greater terrace differentiation is believed to be related to a knickpoint in the longitudinal profile of the stream. The abrupt downstream decline in gradient leads to the buildup of an alluvial fan in periods of sediment deposition. Supposedly, this took place at a high enough rate to produce net aggradation in the subsiding Roer Valley graben. As a result, the groundsurface was sufficiently raised to permit downcutting and terrace formation during periods when supply of debris to the river was at a minimum.

2. In a part of the terraces, the abandoned channels and fluvial bars have suffered a variable measure of deformation. In places, this effect is severe enough to render questionable the unequivocal identification of the channel pattern. Since the orientation of the distorted features almost everywhere coincides with lineaments of the Roer Valley rift system, the latter supposedly have acted as barriers which locally diverted the course of the ancient Maas. The lineaments result from release of horizontal crustal stress (Van den Berg, 1994) and their activity appears to account for the downstream morphological variation found within one and the same terrace level.

Climatic control

The terraces 2 to 6 of Fig. 6.4 represent a cycle of aggradation and subsequent incision. When it is supposed that the deposits of the terrace 2 span most of the Weichselian Late Pleniglacial (Fig. 6.8) and equating the beginning of this interval with that of Oxygen isotope stage 2, then this cycle may have a period of at least 22 ka (from 25 ^{14}C to 3 ^{14}C). The cycle started with a phase of aggradation in a cold and increasingly dry climate; this is suggested by the fact that the fluvial strata of terrace 2 are overlain by a fluvio-aeolian or full-aeolian sand sheet. The close of the period with a polar desert or arctic tundra climate (Zagwijn, 1991) and the transition towards postglacial conditions is marked by three consecutive phases of terracing during the Weichselian Late Glacial and earliest Holocene. Lastly, a protracted interval of fluvial erosion and subsequent steady-state flow during Early and Middle Holocene times was succeeded by a phase of man-induced aggradation. Thus, there appears to be a close match between the evolution of Weichselian Late Pleniglacial to modern climate and that of the Maas valley in the southern Netherlands. A similar trend, though referring to a much longer time scale, was found for a drainage basin in Central China by Porter et al. (1992).
The response of geomorphic processes to climate change not only manifests itself in the alternation of alluviation and terracing but also in the steady narrowing of the floodplain. It is thought that this occurred in punctuated rather than continuous fashion. Accordingly a first narrowing phase supposedly took place during the Weichselian Late Pleniglacial when the depositional regime changed from fluvial to aeolian. Second and third phases of the same type are tentatively attributed to the transition from Allerfd to Younger Dryas respectively Younger Dryas to Preboreal.

Observation data

The subsoil of the topographic levels 2 and 3 of Fig. 6.4 has been studied in the exposure sites Panheel and Kappert, respectively. Site Panheel has a length of 400 m and a mean

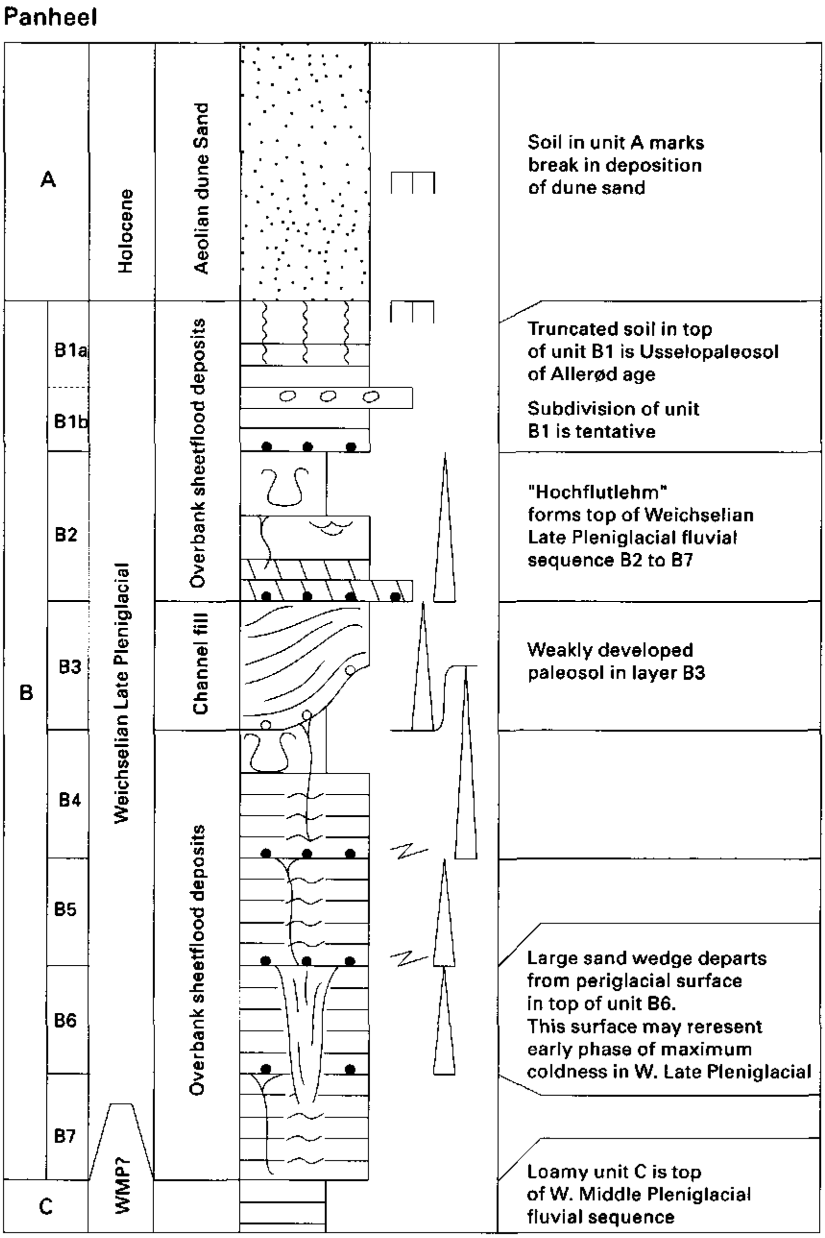


Figure 6.5 Principle features of exposure Panheel. Height of profile = 9.3 m. WMP= Weichselian Middle Pleniglacial. WL= Weichselian Late Glacial. For Legend see Figure 6.7

depth of 7.5 m. The principal characteristics of this location are given in Fig. 6.5. In this figure the terrace proper is represented by units B1 to B7 (counted upwards) and unit C. The B-group consists of seven structurally similar units which are separated from each other by sharply defined bounding surfaces. Within this group, a distinction is made between the trough-shaped unit B3 and the other B-units which have a tabular shape, in general less than a metre in thickness.

In sandpit Kappert the exposed face has a length of 10 m and a depth of 6 m. Fig. 6.6 summarizes the sedimentology of this site and shows that only its basal part (unit E) corresponds to a terrace deposit s.s. The chronostratigraphical position of the various units of the two exposures has been inferred from (i). Specific characteristics of the sediments such as palaeosols, cryogenic structures and alternating bedding in aeolian units and (ii). Tentative correlation with units of known age or chronostratigraphical position by extension of terrace-gradient lines. This procedure is based on the data of Fig. 6.8 with sites Blerick - Koelbroek and Grubbenvorst I and II serving as reference sections.

The interpretation of the two exposures will first concentrate on units B1 to B7 of site Panheel which exhibit the following characteristics:

1. Presence of periglacial surfaces which separate the units from each other. A periglacial surface may be identified as the level of departure of frost cracks, ice-wedge casts or sandwedges, or as the upper bounding plane of periglacial involutions or cryogenic distortions; otherwise as a deflation lag concentrate. Normally, several of these features occur in combination,

2. A fining-upward trend in most of the units. The degree of development of this feature is variable as can be seen when e.g. units B2 and B5 are compared. In the former stratum, the texture changes from gravelly sand at the base to sandy loam in the top whereas in the second one the upward textural change is from fine sand to loamy fine sand only,

3. A fluvio-aeolian make-up of the six tabular units (Fig. 6.9). Various types of fluvio-aeolian processes and features are discussed by Langford (1989), Langford & Chan (1989) and Schwan & Vandenberghe (1991).

In the present context, sediments of this type are characterized by (i).Predominance of a well-sorted sandy texture even to the extent that, on first glance, a pure aeolian origin may be attributed to the deposit in question, (ii). Irregular alternation of windborn beds and strata laid down by shallow current flow, (iii). Occurrence of layers of ambiguous identity which may result from either aeolian or fluvial activity, (iv). An overall trend of upward increase of aeolian influence. In the ideal case, a purely fluvial base gives way to a fluvio-aeolian zone of transition which is succeeded by a purely aeolian top. On the basis of the above features, it is suggested that each of the units B1 to B7 of exposure Panheel is a result of fluvial aggradation with gradually increasing aeolian influence followed by a period of non-deposition, subaerial exposure and formation of a periglacial surface. It is furthermore assumed that this two-phase process was caused by a climatic cycle with deposition during an interval of higher than average temperature and snowfall and periglacial activity in a subsequent phase that was colder and drier than average. Accordingly, each of the B-units represents a cooling-upward and also a drying-upward sequence.

The tabular B-units of exposure Panheel presumably formed by overbank flooding of a braiding or low-sinuosity single-channel river which was a Weichselian forerunner of the present Maas. Because of the periglacial climate, this process would have been restricted to a relatively short period in spring when melting of snow caused a peak discharge and con-

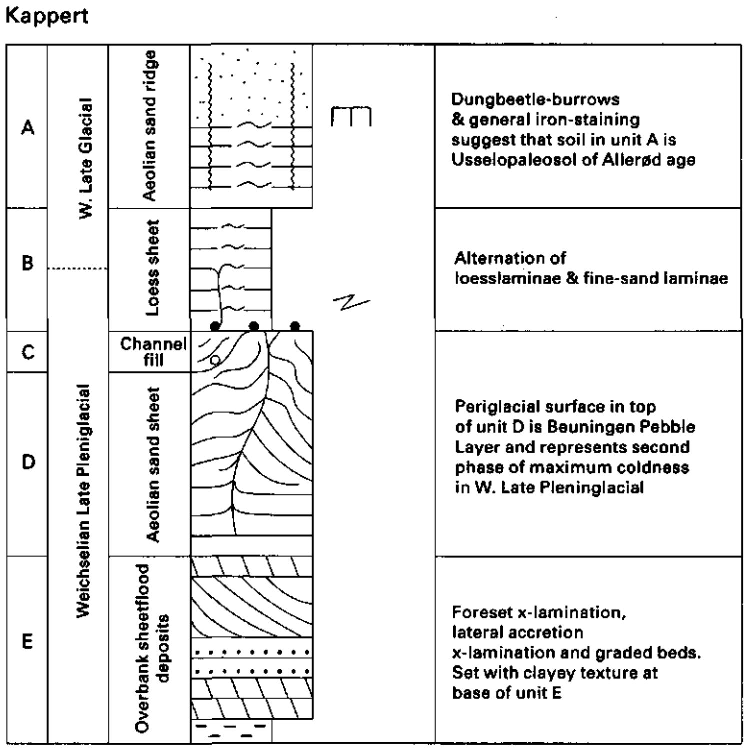


Figure 6.6 Principle features of exposure Kappert. Height of profile = 6 m. For legend see Figure 6.7

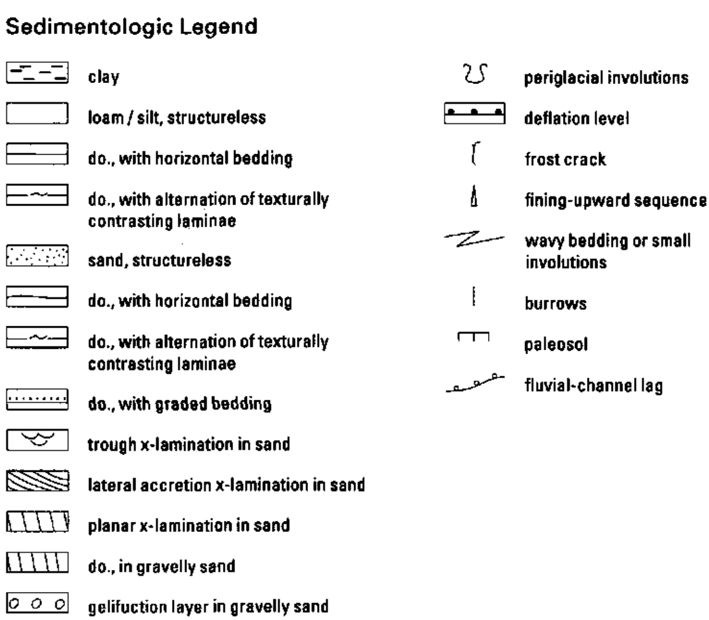


Figure 6.7 Legend to Figs. 6.5 and 6.6

comitant flooding. Thus, each of the units under consideration was built up by a succession of annual flooding events. Deposition of this type would have continued until deterioration of climate stopped or reduced to ineffectiveness the spring snowmelt floods. Here and there on the Weichselian Late Pleniglacial Maas terrace, activity of shallow, non-channelized sheetflows is suggested by features such as shown in Fig. 6.10. The trough-shaped unit B3 is part of a 10 m wide and 1.5 m deep channelfill in exposure Panheel. Its presence amidst strata of different make-up forms evidence of a single depositional event interrupting the dominant fluvial process. The feature is probably due to two consecutive random shifts of the channelzone of the river with lateral displacement first towards and subsequently away from the reference site.

Exposure site Kappert is located on the boundary between an aeolian coversand area to the west and a fluvial-terrace surface to the east (Fig. 6.11). In this section, two marker levels (the Beuningen Pebble Layer and the Usselo palaeosol) occur in fairly distinct state and the gradual upward transition from fluvio-aeolian unit E to pure aeolian unit D is well exposed. Owing to these characteristics, site Kappert is highly representative for the Weichselian Late Pleniglacial period during which fluvial aggradation was increasingly replaced by aeolian deposition.

Discussion

Allocyclic origin of overbank sheetflood units

In the preceding chapter the formation of a succession of seven fluvial beds in exposure Panheel (unit B in figure 6.5) was attributed to climatic forcing i.e. to a repetitive allogenic control; presumably at a millennial timescale as suggested by preliminary O.S.L. dating carried out at the Risö National Laboratory, Denmark (pers. comm. Nigel Poolton). Obviously, this same feature could have been brought about by within-basin autogenic processes such as avulsion and crevassing which may operate independent of tectonic activity or change of climate (Wizevich, 1993). Yet, on the following grounds it is thought that this alternative does not apply.

Aggradation by the Weichselian precursor of the Maas took place on a braidplain confined by terrace scarps. Around site Panheel the so-formed deposits consist of two belts which are approximately parallel to the modern river valley. The belt closest to the Maas roughly coincides with unit B of Fig. 6.11 and represents the channel facies of the ancient river having a texture of gravelly sand. The other zone is to a large extent covered by sheets of aeolian sand or loess and occurs in the subsoil of unit C in Fig. 6.11. This is the overbank sheetflood facies of the Weichselian Late Pleniglacial fill terrace with granular composition of mainly fine sand. The above bipartition suggests that the filling of the space between the terrace scarps proceeded simultaneously in two subenvironments viz. the channel plain and the overbank plain. Due to aggradation under periglacial climatic conditions with short-lived peak discharges during the snowmelt period, sediment-overloading of the river and choking of existing channels would occur every year in spring time. As a result, the channel plain was subject to frequent avulsion events and this affected the adjacent overbankplain in two ways. Firstly, as in a periglacial environment a river acts as a heat source, it would influence the local climate of the overbank zone. Secondly, the lateral shifts of the main channel modified the degree of sand accumulation in the subject zone.

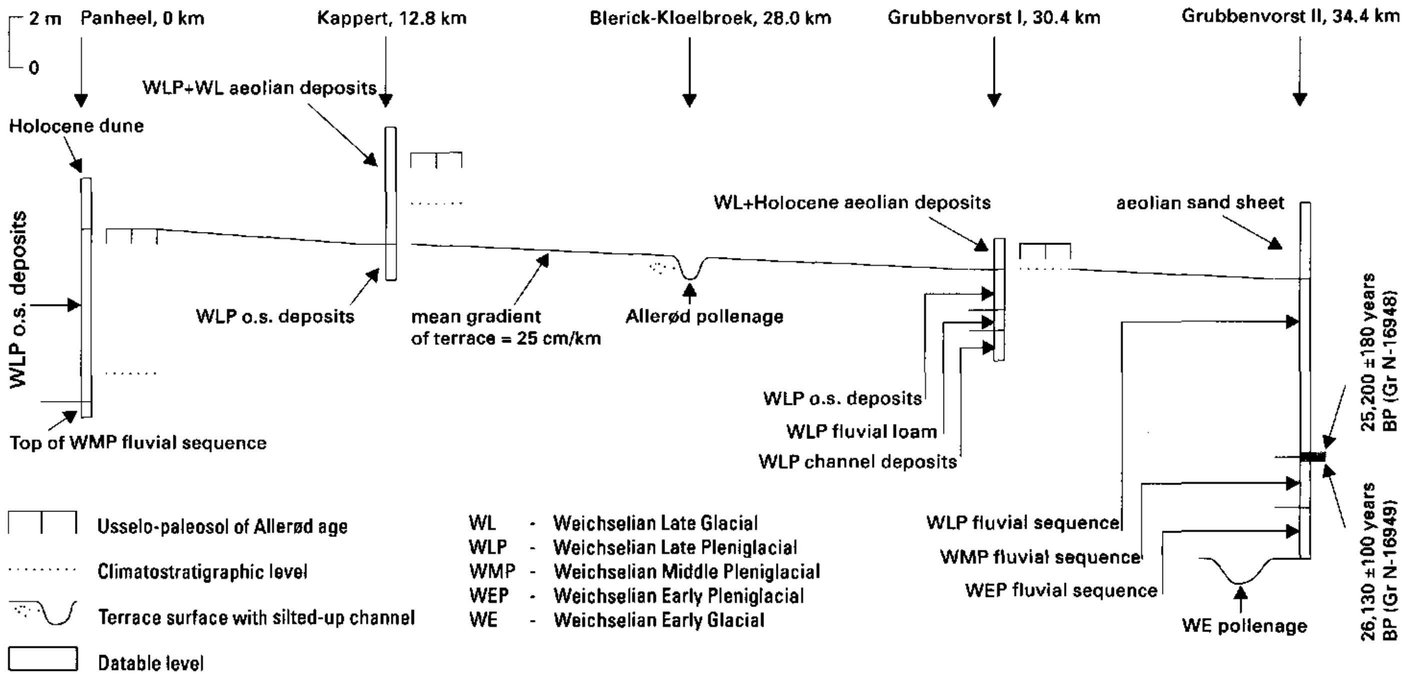


Figure 6.8 Longitudinal section of topographic level 3 of the Maas valley in the southern Netherlands. O.S.= over-bank sheet. Sources: Westerhoff & Broertjes (1990), Schwan & Vandenberghe(1991); Van den Berg & Veldkamp (submitt); P.Cleveringa, W. Hoek and miss. H.van den Bos, pers. comm.

Thus, two extremes have to be considered: (i). The main channel is close to the eastern riverbank with nondeposition and freezing prevailing in the western overbank plain, (ii). The main channel is near the western riverbank with sheetflood deposition of fine sand and thawing of the ground characterizing the western overbank sub-environment. As these two positions alternated many times in the at least ten thousand years of deposition of the fill terrace, we would expect a succession of numerous thin beds separated by periglacial surfaces. In reality, however, the terrace fill consists of no more than six fluvio-aeolian overbank units plus one channel fill unit with a mean thickness of 80 cm and this suggests that an autocyclic mechanism does not adequately account for the aggradation process.

Climatostratigraphic implications

On the basis of various information summarized in Fig. 6.8, a Weichselian Late Pleniglacial age is attributed to the strata B1 - B7 of exposure Panheel (Fig. 6.5). Data from the European continent have shown that the Weichselian Late Pleniglacial was a period of very low winter temperatures, aridity, maximal extension of the continental ice-sheet and permafrost development on a large scale (e.g. Van der Hammen, 1971; Kolstrup, 1980; Behre, 1989; Guiot et al., 1989). Because of the prevailing coldness, neither radio-carbon dating nor palaeobotanical analysis could contribute a great deal to the climatostratigraphy of the subject time span. To a certain extent, this was compensated for by levels of abiotic cryogenic structures occurring over large areas which serve as marker horizons. Increasingly, continental biological and lithological evidence has been supplemented with palaeoclimatological proxy-information contained in ocean- floor sediments and continental ice-sheets. As a result, it is now generally understood that: (i). Climatic events formerly thought to be of regional significance only, may have had a world wide impact (e.g. Porter & An Zhisheng, 1995), (ii). The climate has an inherent instability with greater variability than suggested by continental biological data alone. This is illustrated by the Greenland ice-core records which have revealed short-period climatic fluctuations of considerable amplitude in the later part of the last glaciation (Johnson et al., 1992; Dansgaard et al., 1993; Taylor et al., 1993), (iii). Whereas long-term changes of climate can be accounted for by astronomical forcing, climatic cycles of high frequency are more likely to be due to interactions of the atmosphere-ocean-cryosphere system (e.g. Berger, 1992; Hughes, 1992; Bond et al., 1993).

Basically, the Weichselian Late Pleniglacial Substage consists of two phases of maximum coldness with an intervening period of milder climate. Bond et al. (1993) identified these cold peaks in both North Atlantic marine cores and the Summit ice core and associated them with Heinrich events H1 and H2. The authors date the corresponding cold phases at 14,000 to 15,000 respectively 21,000 14C yr BP. Here, it is suggested that both spells of extreme coldness find expression in the two exposures. At site Panheel a large sand wedge departs from the top of unit B6 and this feature may represent the early period of greatest coldness in the Weichselian Late Periglacial. In similar fashion it is assumed that the Beuningen periglacial surface at Kappert relates to the second coldness peak of the Weichselian Late Pleniglacial Substage around 14,500 14C yr BP. The latter data represents a minimum age for the upper boundary of fluvial unit B at Panheel. This follows from the fact that the aeolian deposits of site Kappert directly overly the fluvial unit under consideration (see Fig. 6.8 and 6.11).

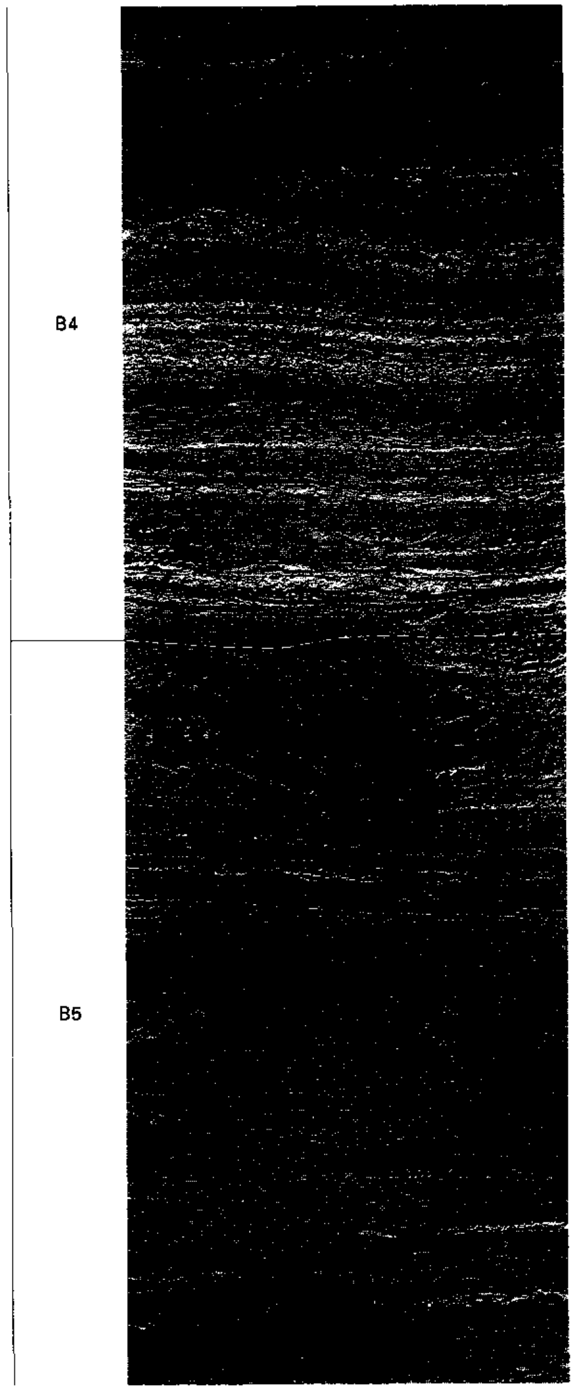
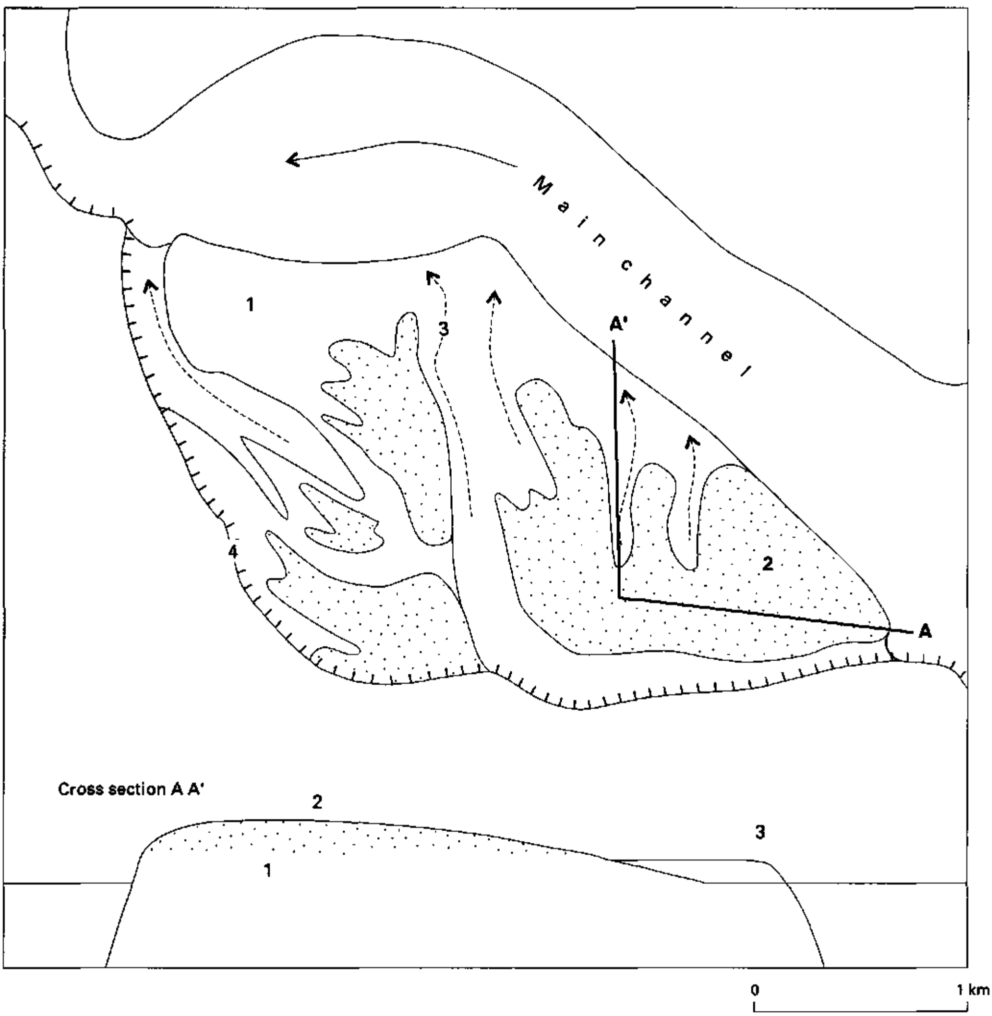


Figure 6.9 Lacquer peel of fluvio-aeolian sands in exposure site Panheel. Bounding surface between units B4 and B5 is marked by cryogenic deformations which were levelled by subsequent fluvial erosion. Length of peel = 125cm.



1 Terrace surface

2 Hand-shaped depositional lobe

3 Gullies formed during flooding event

4 Terrace scarp

Figure 6.10 Morphology of overbank sheet-flood deposits on the surface of the Maas-terraces in the southern Netherlands.

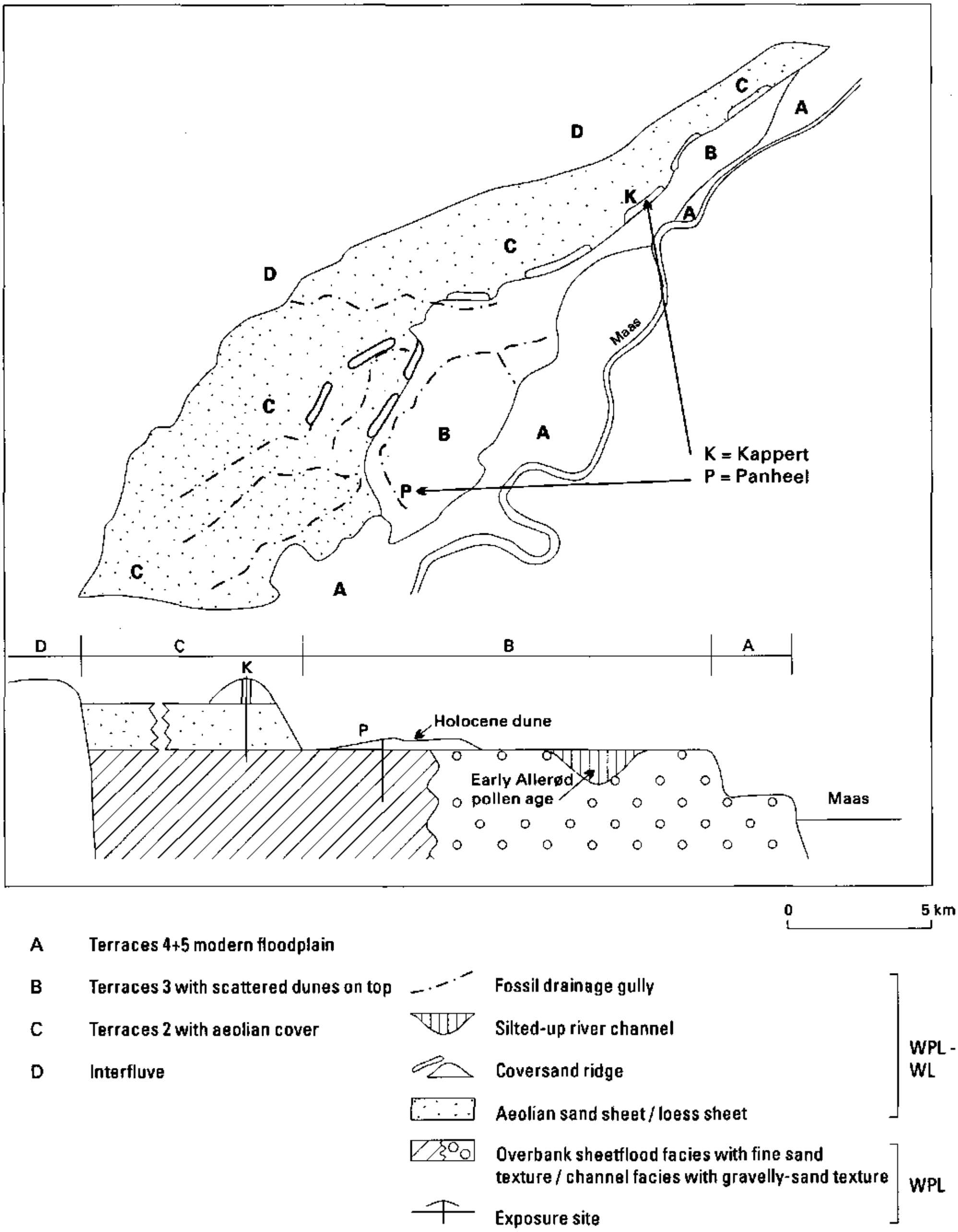


Figure 6.11 Geomorphology and stratigraphy of the exposure sites and vicinity. Cross section is not to scale and constitutes of a composite of relevant features within and outside the depicted area. WLP = Weichselian Late Pleniglacial. WL = Weichselian Late Glacial.

Period of the fluvial cycles

Working from the observation that seven or at most eight climatic cycles have been registered in unit B of exposure Panheel (see Fig. 6.5), the main period of these rhythms can be evaluated. On the basis of the data given in Fig. 6.8, the lower boundary-age of the fluvial succession may be set at 25,200 14C yr BP. Assuming that its upper boundary-age equals 14,500 14C yr BP or slightly less (see preceding section), a range of 1,440 to 1,530 years is found for the main period of the subject cycles. In Fig. 6.12, the corresponding diagrammatic curve of site Panheel is shown along with four records of climate-related parameters from various parts of the Northern Hemisphere. Though it is not possible to directly match the Panheel fluvial record with any of the other time series in Fig. 6.12, it is worth of notice that a process of short-period, recurrent climate change has been at all registered. Moreover, this occurred at a high level of resolution and in an environment which, in general, is thought to have low preservation potential.

Response of river to climate change

A discussion of the behaviour of the Maas in the southern Netherlands vis-à-vis change of climate involves consideration of three different time scales. These time scales correspond to the periods of the fourth, fifth and sixth order climatic cycles already mentioned in a previous chapter and graphically represented in Fig. 6.1. The fourth-order cycle is a glacial-to-interglacial climatic cycle which ran its course through Weichselian and Holocene times. With minor interruptions, aggradation took place during the better part of the Weichselian Age. At the onset of the Weichselian Late Glacial the cold-climate depositional process was replaced by stepwise incision which lasted to, roughly, the end of the Subboreal Chron (de Jong, 1970). From then, the river resumed its aggradational activity mainly as a result of deforestation by man. In the present context we assume a cause-and-effect relationship of climatic cycles and fluvial cycles. Accordingly, fluvial cycles are periodical changes in river behaviour which are: (i). Expressed in the lithologic, geomorphologic or biologic characteristics of the river system and (ii). Supposed to be caused by cyclic changes of climate. The subject cycle may have had a period of 116 ka when we tentatively set its beginning at 119 and its end at 3 ka BP.

The fifth-order cycles are stadial-to-interstadial climatic cycles of the Weichselian (Chapter 5). In the subsoil of the study area, the first one of these cycles is represented by an aggradational sequence with fining-upward trend, the second one by a severely truncated deposit and the third and fourth cycles again by fining-upward sequences with the clayey top of the last succession forming part of the present day landscape. Further particulars on the four cycles are given in Table 6.1. In Fig. 6.13 aggradation of the Weichselian Late Pleniglacial (=fourth) fining-upward sequence and the subsequent development of the fluvial landscape are represented in the form of a time-altitude diagram; the time points in the diagram are constrained by radio-carbon ages, pollen ages and correlations with dated ice-core records.

The sixth-order cycles are of two different types. The aggradational type is represented by the seven or possibly eight drying-upward/cooling-upward cycles of Weichselian Late Pleniglacial age which constitute the fluvial succession at site Panheel. For these rhythms a mean period of 1,430 year could be given. The erosional type occurs in the Maas- valley of the southern Netherlands as a flight of two cut terraces (terraces 4 and 5 in Fig. 6.4) and a lowest planation surface (terrace 6) which is now buried under Late Holocene sediment

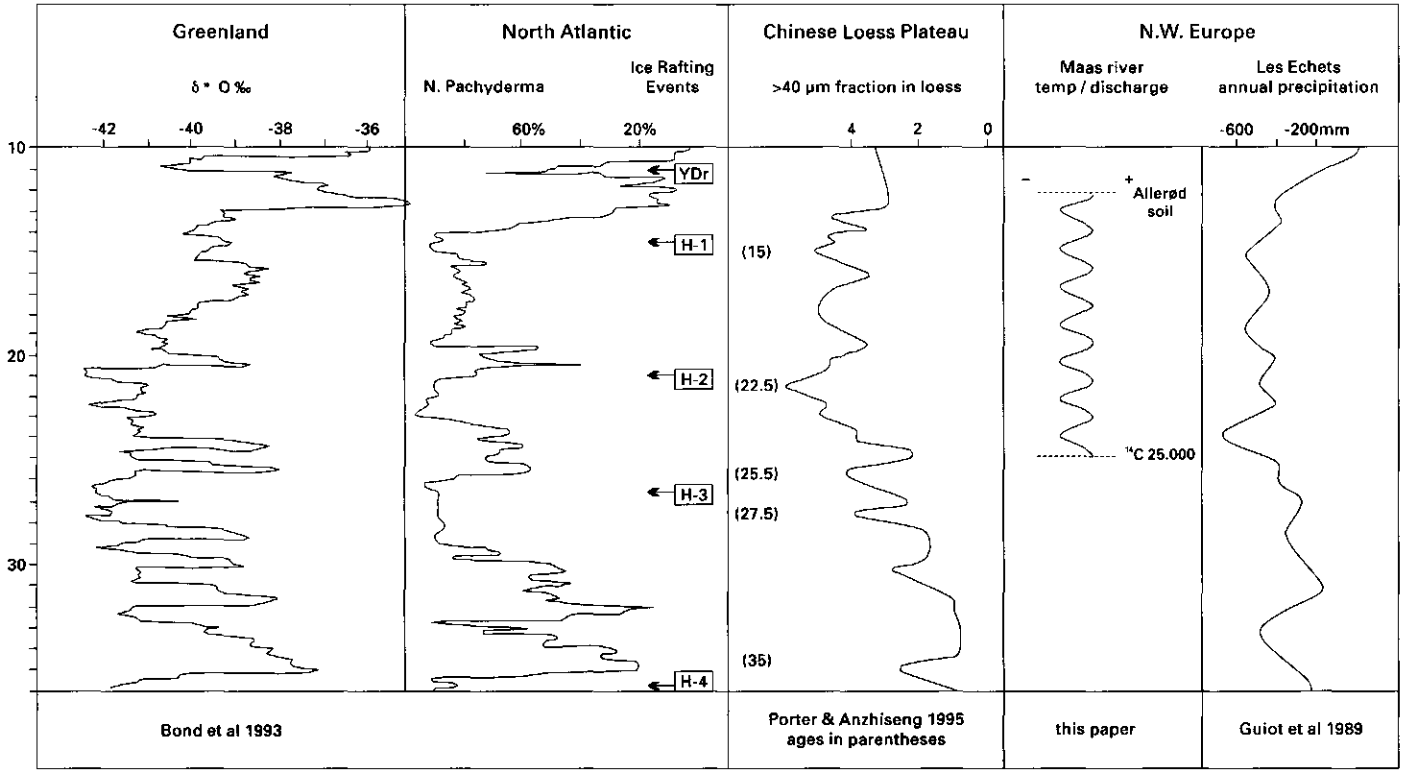


Figure 6.12 Four paleoclimatological proxy records of the Northern Hemisphere and a diagrammatic curve of site Panheel. The horizontal axis of this curve is a climate - related quality on ordinal scale.

of the present floodplain (see Fig. 6.13). The three levels formed by stepwise incision of the river into its own previously aggraded deposits and now a cycle consists of a phase of vertical erosion followed by an interval of horizontal planation. This process was paralleled by a time-punctuated narrowing of the floodplain as shown in Fig. 6.13. With incision beginning early in the Allerød Interstadial (13,900 cal yr BP) and being replaced by Late Holocene aggradation at approximately the Subboreal-to-Subatlantic transition (3,000 cal yr BP), the mean period of the sixth-order erosional cycles amounts to 3,630 years.

Causes of the fluvial cycles

It is suggested that the fourth-order fluvial cycle with period of 116 ka should be a result of astronomic forcing with dominance of the orbital-eccentricity component. Likewise, it may be assumed that the fifth-order fluvial cycles with periods ranging from 15 to 33 ka were driven by rhythmic changes in insolation caused in the first place by the interaction of the two other orbital parameters, i.e. by the combined variations of the obliquity of the Earth's axis and the precession of the equinoxes.

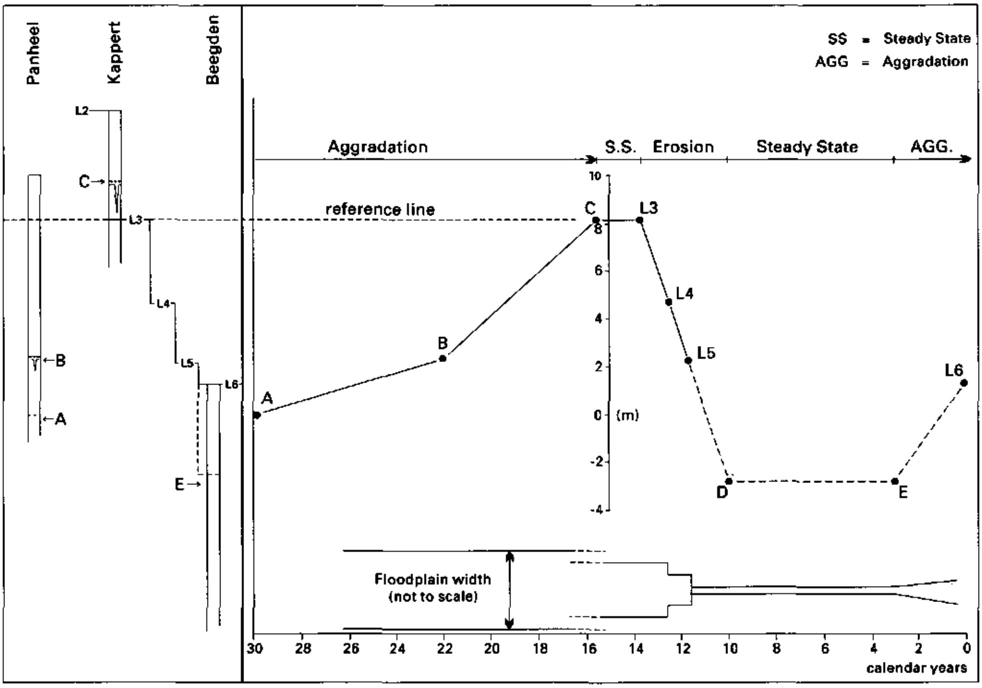
Regarding the sixth-order fluvial cycles with periods of 1.4 and 3.6 ka, it is noted that: (i). Internal dynamics of the ocean-atmosphere-cryosphere system rather than orbital perturbations should account for the high-frequency periodicities in question, (ii). Precipitation is the principal climate-parameter affecting river behaviour and (iii). Weather conditions in the drainage basin of the Maas were (and are) determined to a large extent by the proximity of the Atlantic Ocean. On the basis of these considerations it is proposed that the sixth-order fluvial cycles were driven by periodical changes in the strength and position of storm-tracks on the North Atlantic (e.g. Kapsner et al., 1995; Lowe et al., 1995). If cyclic changes in atmospheric circulation should be an acceptable explanation, a distinction must be made between the aggradational fluvial cycles of Weichselian Late Pleniglacial age and their erosional counterparts of mainly Weichselian Late Glacial age. This distinction refers to the two different palaeoclimatic settings so that the ultimate causes of the recurrent storm-track displacements may not have been the same in either case.

Conclusions

1. In the study area in the southern Netherlands, the river Maas crosses the central graben of the Roer Valley rift system. Near the Feldbiss fault which separates the rising South Limburg Block from the Roer Valley graben, the build up of an alluvial fan took place at a rate exceeding subsidence of the downthrown block. As a result, the groundsurface in the Roer Valley graben was sufficiently raised to permit stepwise incision and greater terrace differentiation than in the Maas valley upstream of the Feldbiss fault.

2. The Maas river system has responded to periodical climate changes on Weichselian to Holocene (= glacial-to-interglacial), Weichselian stadial-to-interstadial and millennial time scales. Accordingly, fourth-, fifth- and sixth-order fluvial cycles find expression in the terrace morphology and deposits of the river.

3. The sixth-order (=millennial-period) fluvial cycles are of two different types. The erosional type is of Weichselian Late Glacial to Early Holocene age and consists of two successive terrace-levels and a planation surface which is now buried under deposits of the modern floodplain. The aggradational type dates to the Weichselian Late Pleniglacial



A: aggradation since: 25 ka 14C yrs. = 30 ka cal. yrs. (Bard et al. 1993)

B: sand-wedges correlated with the episode of deepest coldness at 22 ka.

C: 15.5 ka. close of fill terrace formation; ante quem age found by correlating the overlying deflation lag-concentrate and associated m-scale frost fissures with the last deep cold spell before the onset of climatic improvement

D: erosion sequence extrapolated to the level of E. (See text for comment)

E: aggradation resumes in this section at the Subborial/Subatlanticum boundary (de Jong, 1970) (See text for comment)

L3: level abandoned at the onset of the Allerød (13.9 ka)

L4: level abandoned at the end of the Allerød (12.6 ka)

L5: level abandoned at the end of the Younger Dryas (11.6 ka)

L6: present surface of the overbank plain

Figure 6.13 Time-elevation diagram representing the evolution of the Maas valley in the southern Netherlands from the past 30 ka. Reference line corresponds to topographic level 3. Age of point D in diagram agrees with borehole section Rijckholt (De Jong, 1984) in which the base of the hiatus of 3.8 ka is dated at 8,610 14C yr BP (GrN-11316) corresponding to approximately 9,600 cal yr BP.

Substage and is represented by a succession of overbank-sheetflood units and an intercalated channel-fill unit with each of these units being bounded by a periglacial surface.

4. The aggradational type of sixth-order fluvial cycles is interpreted as a succession of seven or possibly eight drying-up/cooling-up rhythms with mean periods of 1.4 ka. The restricted number of units with mean thickness of 80 cm in the aggradational succession speaks in favour of an allocyclic origin since, in the absence of any kind of external control, a far greater number of thin units should have formed in the available depositional period.

The subject fluvial cycles supposedly were driven by periodical changes in the strength and position of storm-tracks on the North Atlantic. These short-term fluctuations must have been linked, in unknown manner, to the internal dynamics of the ocean-atmosphere-cryosphere system of the Earth.

Table 6.1

Chronology of four Weichselian fluvial cycles of the Maas in the southern Netherlands.

Fluvial cycle	Period of aggradation and subsequent incision	Tentative duration of cycle in ka
4th	WLP with stepwise incision and planation during WLG and Early + Middle Holocene	21 to 26
3rd	WMP with floodplain abandonment beginning at 26,130 14C yr BP	33
2nd	WEP	15 or less
1st	WE with floodplain abandonment during Brørup Interstadial	124 to 29

WLP= Weichselian Late Pleniglacial; WMP= Weichselian Middle Pleniglacial;
WEP= Weichselian Early Pleniglacial; WE= Weichselian Early Glacial

Geologie en Mijnbouw 73: 143-156 (1994)

7. Neotectonics of the Roer Valley rift system. Style and rate of crustal deformation inferred from syn-tectonic sedimentation

Abstract - *The Roer Valley rift system emerged since the Middle Miocene and fluvial sediments were supplied to it by the Rhine, Maas (Meuse) and local Belgian rivers. Ever since the emergence, thirty one fluvial terraces of the lower Maas river have been formed due to regional uplift. Their age-altitude record shows strong evidence for an important acceleration of the tectonic activity at the end of the Pliocene (around 3 Ma), and for high-frequency oscillations superimposed on a general continuous trend. Three relaxation periods during the Quaternary were identified, the first from 1.5 to 1.2 Ma and two short ones around 5 ka BP and after 2 ka BP, respectively. The reactivations, following these relaxation periods, appear to be of plate-tectonic importance. The observed accelerations in tectonic activity since the Late Pliocene through the Pleistocene to the present day, raise the question: are we at present living in a period of extremely high crustal dynamics? Floodplain positions of the rivers Rhine and Maas repeatedly changed in space and time. Strike-slip movements along the graben bounding faults explain this behaviour. The events point to punctuated changes in the stress field orientation, probably related to the interplay between Alpine and Ardennes-Rhenish Shield stress generators within the regional stress field.*

Introduction

Recent work on subsidence (Zijerveld et al. 1992) and stratigraphic modelling (Kooi & Cloetingh 1989; Kooi 1991; Kooi et al. 1991; Cloetingh & Kooi 1992) of the southern North Sea region provides evidence of a phase of accelerated subsidence in the Late Neogene. This phase has been recognised all around the North Atlantic (Cloetingh et al. 1990).

These observations stress the importance of the study of the time frame of the Late Cenozoic geo-dynamics (neotectonics), more so because similar observations have been made in other parts of the world (Kaldova 1988; Ruddiman et al. 1989). Whether the observed events are synchronous, however, is still uncertain.

Integrated studies of fluvial geomorphology, fluvial sedimentology and stratigraphy can improve the time resolution of regional neotectonics. This is demonstrated in the present case study of the Roer Valley rift system.

Earlier work in this tectonic domain (Van den Berg 1989) has shown a strong matching of independently mapped fluvial and structural patterns.

This observation has led to the basic assumption that crustal block movements, even if they operate at extremely low rates, due to their persistence steer the long-term position of the fluvial system. If so, the analyses of the fluvial record in space and time may reveal the ongoing crustal deformations.

This chapter focuses on the uplifting south flank and subsiding central part of the rift system.

The interpretation of their dynamics will be followed by some reflections on the plate-tectonic implications.

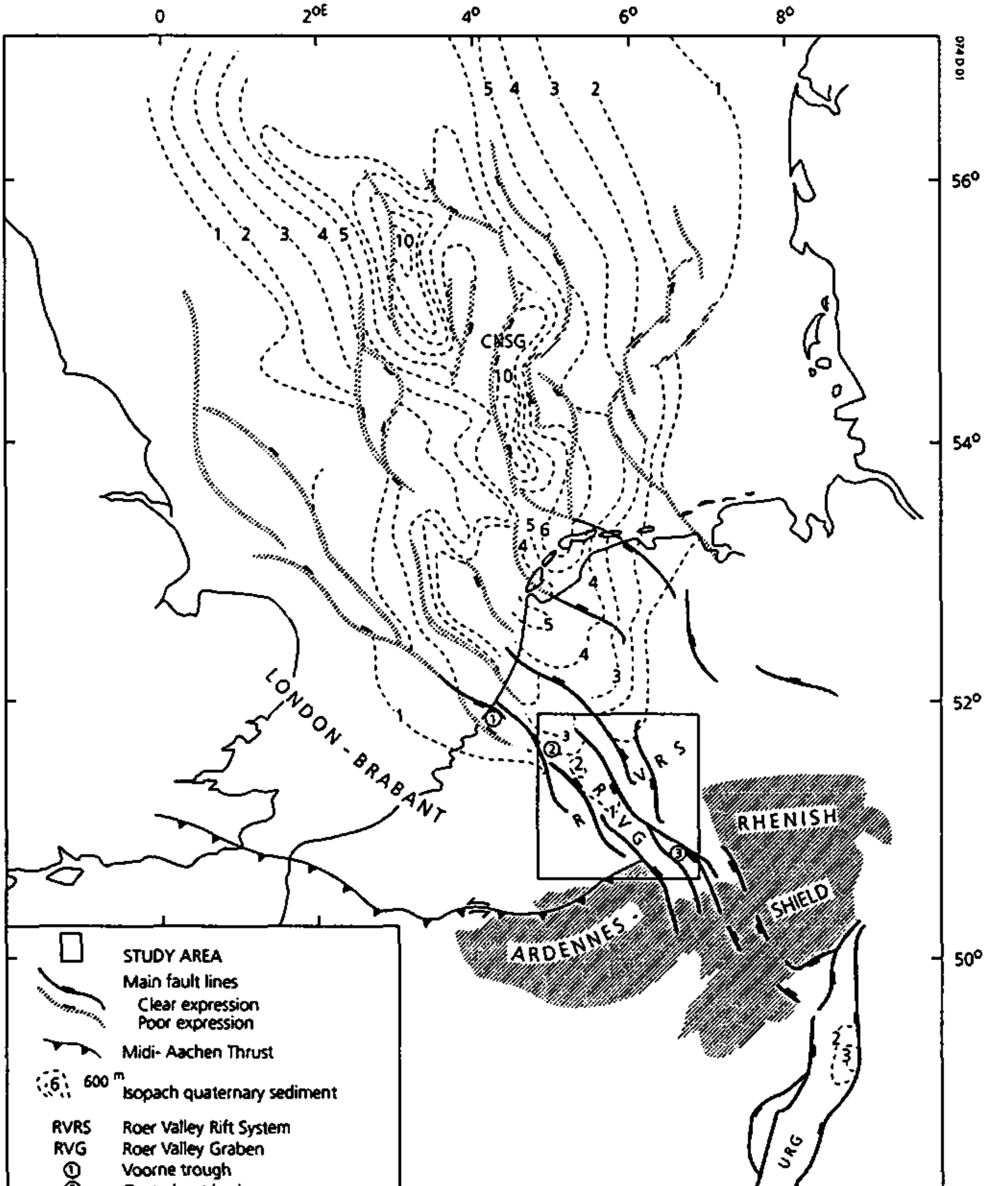


Figure 7.1 The study area (indicated by the box) within its tectonic setting. Quaternary isopach lines at intervals of 100 m (after Caston 1977; Illies 1975). (Partly after: Illies 1975 and GECO Exploration Services & Alastair Beach Associates 1989).

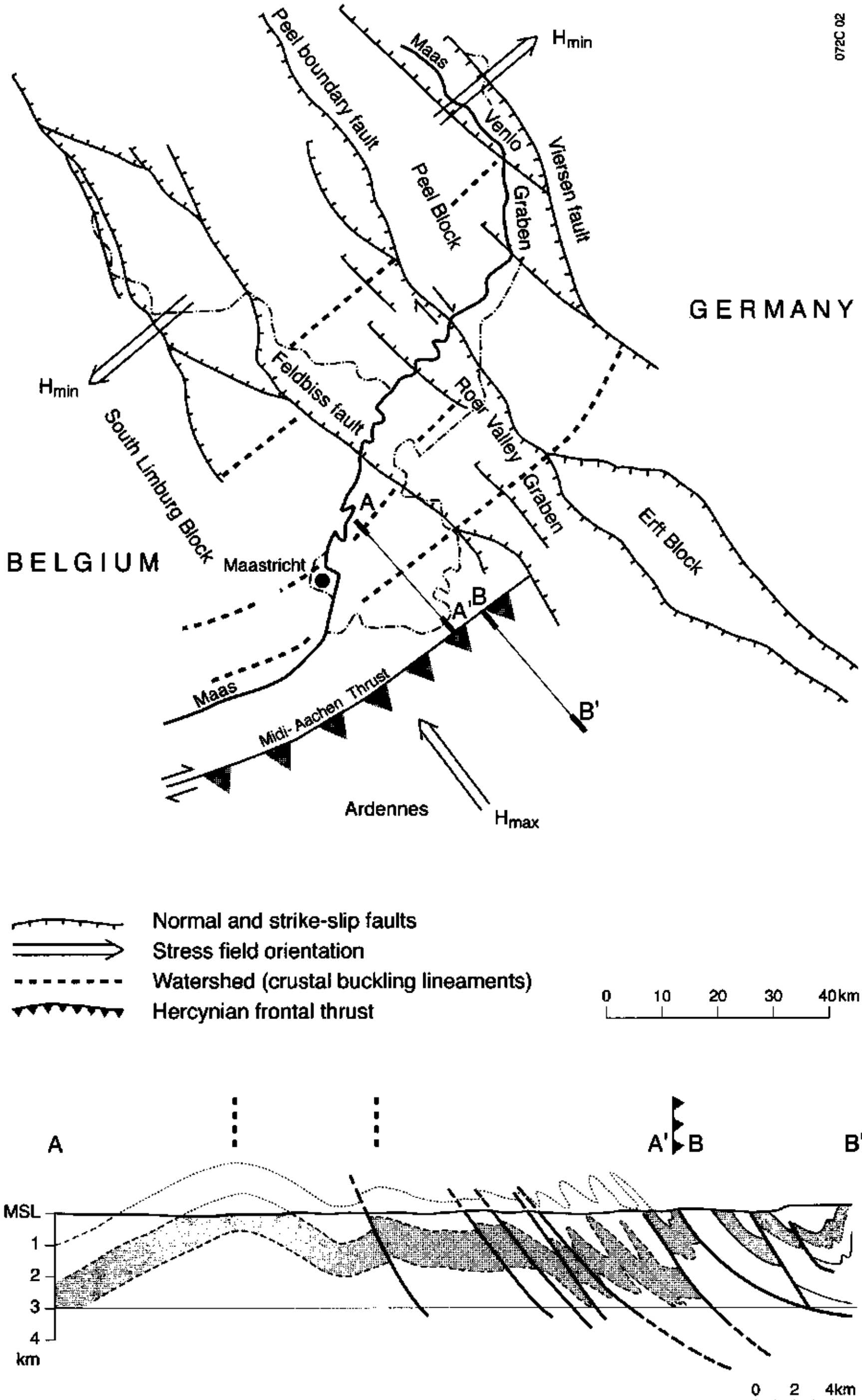


Figure 7.2 Regional (palaeo) watersheds developed parallel to the Midi-Aachen Thrust in the course of the Late Neogene. They coincide with subsurface structural highs of Hercynian age and therefore are interpreted to reflect crustal buckling caused by neotectonic foreland compression.

Structural setting

Tectonic units

The Roer Valley rift system forms the main structural-physiographic unit of the Lower Rhine Embayment. The system forms the southernmost extension of the North Sea Basin (Fig. 7.1). As such it forms part of the western and central European rift system, a chain of depocentres that connects the Mediterranean with the North Sea (Ziegler 1987). Owing to its hydrocarbon prospects, the geohistory of the North Sea Basin and the basin's structural setting are quite well known (Van Doorn & Leyzers Vis 1985; Van Hoorn 1987; Van Wijhe 1987; Demyttenaere 1988; GECO Exploration Services & Alaistair Beach Associates 1989; Remmelts & Duin 1990; Geluk 1990).

The Roer Valley rift system forms a NW-oriented fault-bounded element between the sub-parallel Upper Rhine and Central North Sea Grabens. Together these grabens form a dog-leg-like structure within the extensional setting of the NW European plate. The local North Sea depocentres are aligned with the ancient fault patterns at Triassic and Jurassic levels. Although the fault expression in the Cenozoic sediments is poor, the matching between the sediment distribution and the structural pattern suggests a strong control by the latter (Fig. 7.1).

The study area is divided into two asymmetrical segments, both having their deepest part along the northeast side:

(i) The southern segment, showing the strongest subsidence, is referred to as the Roer Valley Graben (formerly named Central Graben (Remmelts & Duin 1990; Geluk 1990) or Ru(h)r Graben (Ziegler (1992)).

(ii) The northern segment is the Peel-Venlo Block (Van Rooijen et al. 1984), the deepest part of which is referred to as the Venlo Graben. Three NW-oriented, principal displacement zones (PDZs) bound the segments: the left-stepping Sandgewand-Feldbiss-Rouw-Rijen faults in the southwest; the Peel-Roer Boundary fault forming the central element, and the Vierssen Fault in the northeast. To the south the area is flanked by the rising South Limburg Block. The northern flanking region (the Krefeld High) will not be considered.

The two main segments have been faulted. These segments display subtle to strong differences in vertical movements, both expressed in the sedimentary record as well as in their geomorphology. Analysis of their dynamics in space and time is important for a comprehensive view on the detailed evolution of the rift system dynamics, as discussed below.

State of stress

The regional crustal motions are controlled by the interplay of at least three intraplate stress fields.

(i) The subsiding North Sea Basin generates a flexural marginal bulge. The Pleistocene highs in the southern and eastern Netherlands may be an expression of this bulge which forms waterdivides: to the south with the W-E running 'Flemish Valley' in Belgium and to the east with the S-N running Ems valley. The PDZs dissect this bulge. The forming of a bulge may explain while fault zones in this region have a pronounced character in comparison with the poorer expressions more basinward (Fig. 7.1).

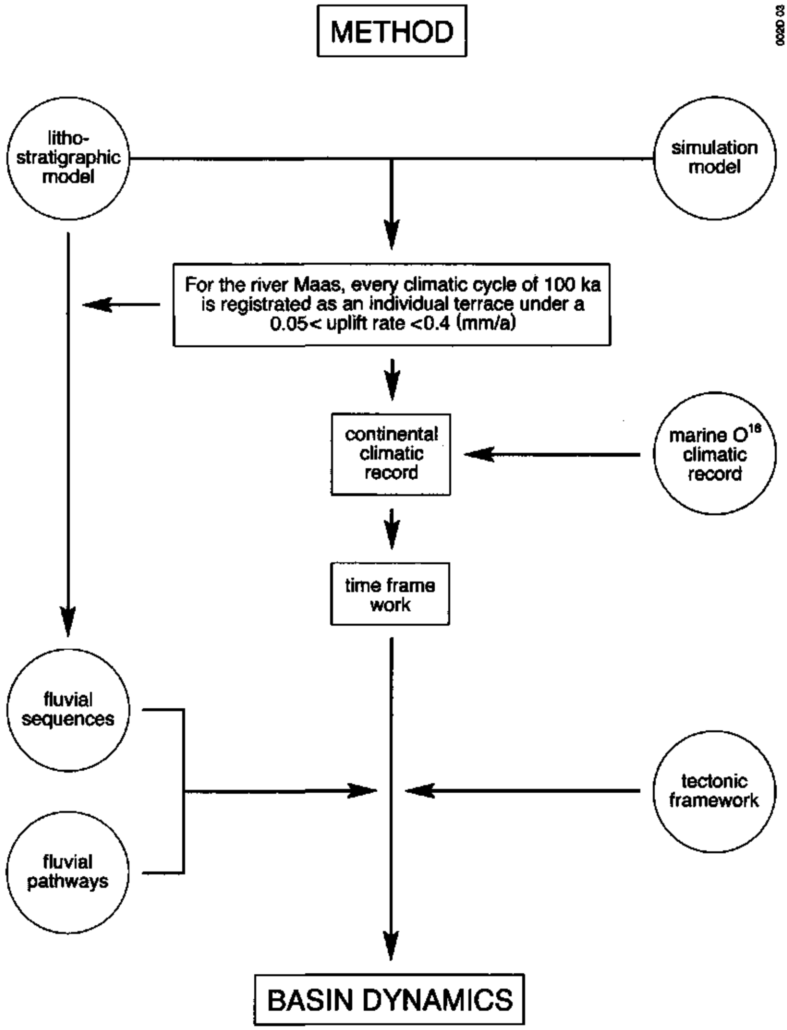


Figure 7.3 Flow chart of the method used in the present approach to unravel the Plio-Pleistocene morpho-tectonics of the Roer Valley Graben.

(ii) The horizontal principal stresses in western Europe indicate deep-seated stresses away from the Alpine collision front in a northwest direction (Klein & Barr 1986; Philip 1987; Mueller et al. 1992). This stress field (H_{max}) opens up the rift system by tensional forces (H_{min}) as is indicated by stress measurements (Ahorner et al. 1983; Illies & Greiner 1978;)

(iii) The Midi-Aachen Thrust (Meissner et al. 1983) separates the study area from the Ardennes and Rhenish Shield to the southeast. Mapping of the main (palaeo-)water divides indicates a pattern of alignments oriented parallel to this thrust (Fig. 7.2). The divides correspond with positive tectonic structures, and are therefore interpreted to reflect (reactivated) crustal buckling lineaments generated by foreland compression by the Ardennes.

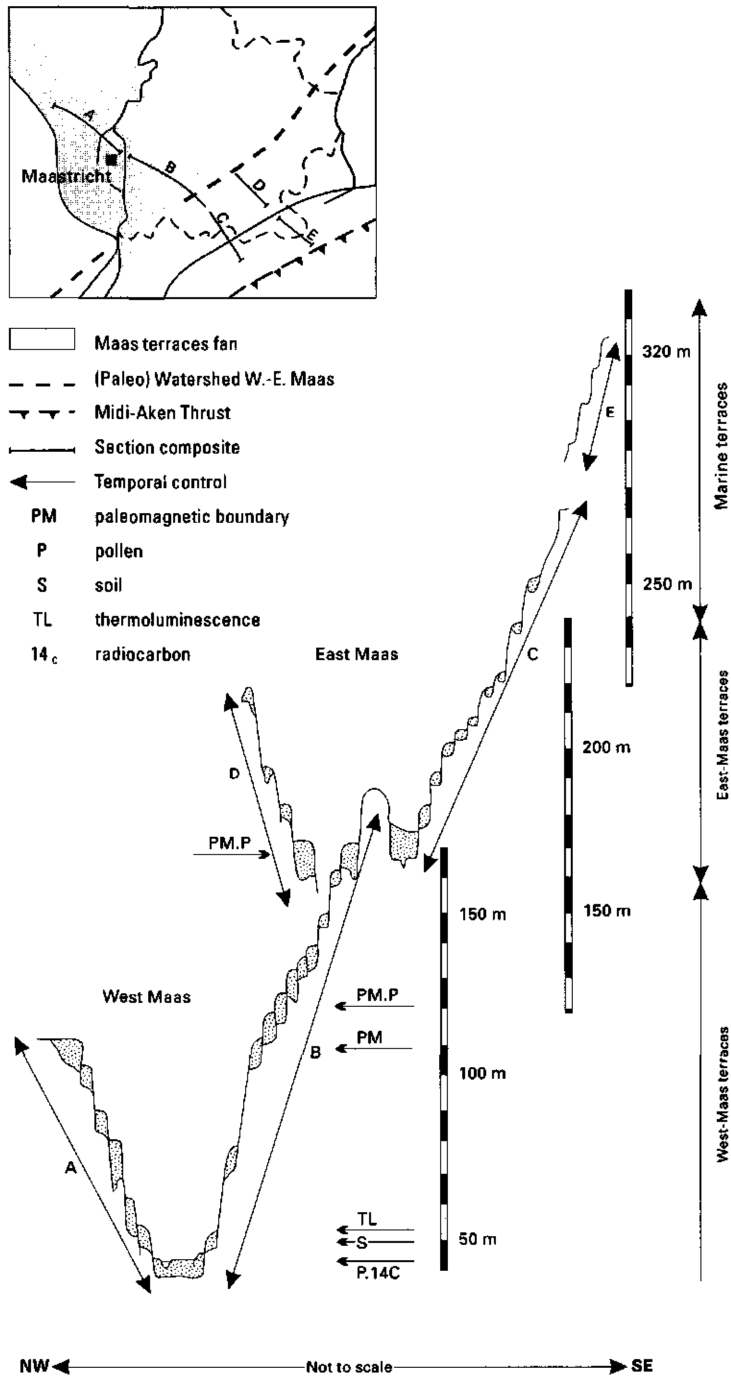


Figure 7.4 Composite section of the Maas river terrace flight in South Limburg. For an indication of ages: the transition from marine terraces to fluvial (East-Maas) terraces occurred in the Mid Serravallian at about 13 Ma; the transition from East to West- Maas terraces occurred at around 2.1 Ma.

The time frame

In order to obtain a time frame for the basin dynamics, an integrated approach was used including geomorphology, modelling and stratigraphic correlation (Fig. 7.3)

Morpho-stratigraphic studies of sediments of the river Maas in the south flank of the graben revealed a long flight of terraces (Fig. 7.4; Van den Berg 1989;).

Veldkamp & Vermeulen (1989) and Veldkamp (1992) developed a model to simulate river-terrace formation. The model operates in large-scale analogies of real processes acting in conceptual 2000-year time-steps on a macro-scale (100 km^2) setting.

The first step in Fig. 7.3 combines these two approaches (Veldkamp & Van den Berg 1993). This led to the conclusion that the interplay between the tectonics and the macro-climatic variability (ruled by the eccentricity of the earth rotation) closely determined the long-term terrace formation and preservation. Under a tectonic uplift regime ranging from 0.05 to 0.4 mm.a^{-1}, every eccentricity-related climatic cycle is represented by a terrace in a rain-fed system such as the river Maas. Terrace sediments represent the cold periods and as such they are valuable counterparts of the palynological record registering the warm episodes (erosive periods). Therefore, the combined terrace and pollen record serves as a long continental climatic record.

This allows the correlation of that record with the deep sea oxygen isotope climatic record. For this correlation we used oscillation-pattern matching, supported by palaeomagnetic data. Shackleton et al. (1990) tuned the oxygen isotope record to the astronomical timescale. Thus our correlation of the marine with the continental record transmits this absolute timescale to the latter record (Fig. 7.5). The time resolution obtained in this way reaches the 20 000 years level. This is much better than a solely biostratigraphically based level.

In subsiding parts of the Roer Valley rift, sediment packages of different source areas interfinger or are superimposed (Edelman 1933; Zonneveld 1949; Kasse 1990; Boenigk 1978). The related fluvial systems have been distinguished by means of heavy-mineral provenance studies. Their ages come from associated pollen assemblages, palaeomagnetic data and correlative terraces (Zagwijn 1960; Zagwijn et al. 1971; Zagwijn & Zonneveld 1956; Van Montfrans 1971; Zagwijn 1989). The basin dynamics in space and time has been inferred from the combination of the morpho-(litho)stratigraphic model with the time frame within the given tectonic framework. We used the terraces in uplifting parts (time-altitude position). In the subsiding part we used map-pattern correlation of the changing fluvial pathways (palaeogeography) with the known structural units.

Regional Late Miocene emergence

The study area slowly changed from a shallow marine environment to a fluvial plain since the Mid-Miocene, around 13 Ma (Quitzow 1974; Zagwijn & Hager 1987; Zagwijn 1989). This regressive coastal plain is bounded in the southeast by a palaeo-coastline . This line is either marked as a sub-areal, low fossil cliff with residual beach-pebbles (Van den Berg 1989, Felder & Bosch 1989, De Jong & Van der Waals 1971), or as a buried beach (Hager 1981; Boersma 1992). The cliff forms an important morpho-stratigraphic marker in the regional uplift history, because the onset of the regression is well correlated with Vail's cycle boundary TB 2.4/2.5 (13 Ma; Herngreen 1987).

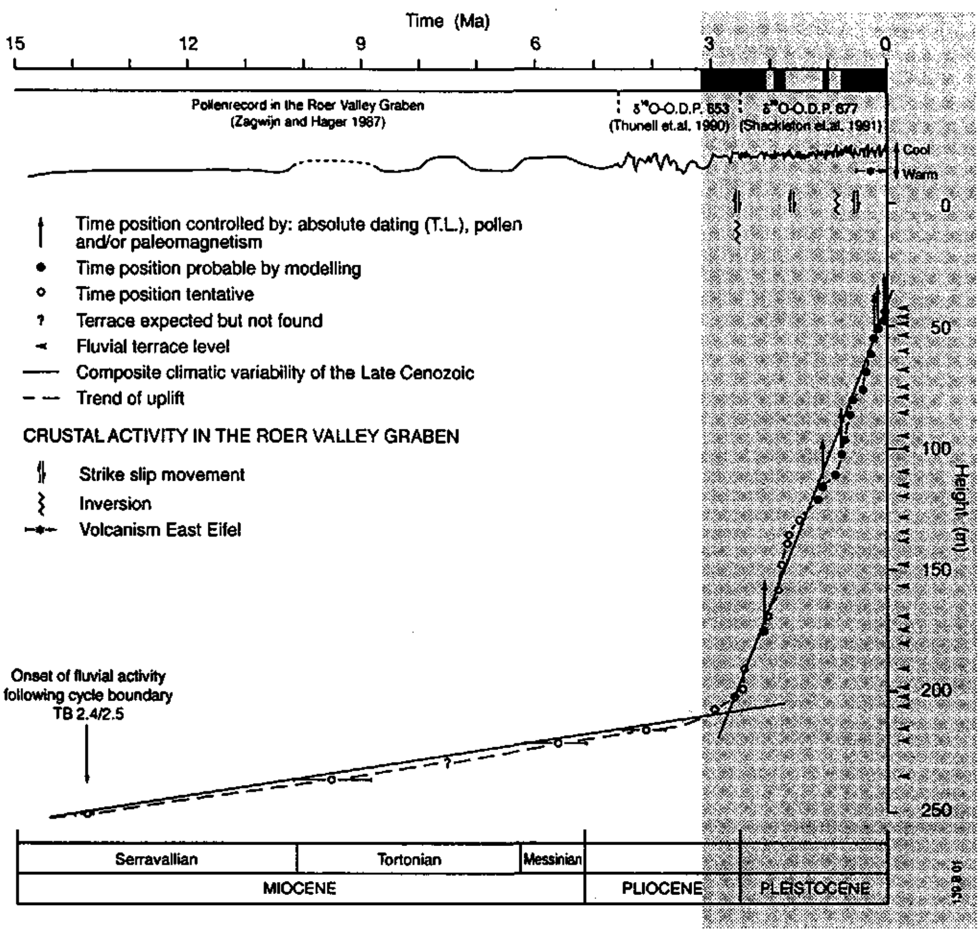


Figure 7.5 Miocene to recent uplift record at the southern shoulder of the Roer Valley Graben from dating of the terrace flight in Figure 4. The major break in the uplift witnesses the onset of the regional neotectonics at around 3 Ma.

The emerging coastal plain was fed by three feeder systems: the (Early-)Rhine, the Maas, and a number of relatively small 'Belgian' streams draining the north flank of the Brabant Massif. The uplifting south flank of the rift system together with the Ardennes-Rhenish Shield were important sediment suppliers. Net upstream erosion leaves a terrace flight. The erosion record of the Maas system is the longest and best dated of the three and can be used separately to reconstruct the uplift history of the southeast flank of the system. This record will be discussed first, followed by a discussion focusing on the sedimentation area: the subsiding grabens.

Uplift of the south flank region

The record

In the southeast part of the area, below the last Mid-Miocene marine cliff a flight of 31 fluvial terraces has been formed (Fig. 7.4). The flight is spread in a terraced 'fan'. Its formation is strongly tectonically controlled in space and time. In mapview the ,'fan' is divided by two buckling lineaments parallel to the Midi-Aachen Thrust (Fig. 7.2); the southernmost lineament separates the (palaeo) East-Maas from the (present) West-Maas, the other one separates the Main Terrace group from the Middle Terrace group. The 'fan', is bounded to the west by a series of along-strike overstepping faults (the Rauw, Hooge Mierde and Rijen faults).

The time is expressed by the height difference between the individual terrace steps. The vertical distance between the last marine cliff (age about 13 Ma) and the first Maas terrace, that witnesses cold-climate conditions of the Latest Pliocene to earliest Pleistocene (age about 2.7-2.4 Ma), amounts to only 30 m. This distance is bridged by three terrace levels of late Tertiary age. The 30 m corresponds to an average uplift rate of around 3.10^{-6} $m.a^{-1}$. On the other hand the remaining Pleistocene through Holocene record covers 160 m, giving an average uplift rate of 6.10^{-5} $m.a^{-1}$.

The major difference between these two long-term averages in uplift rates, in conjunction with the results obtained from the subsidence analysis published elsewhere (Zijerveld et al. 1992), highlights the importance of the discontinuity in the rates of the vertical motions. This break may be used to define more precisely the timing of the onset of the regional neotectonics (Fig. 7.5).

Figure 7.6 shows in more detail the regional neotectonic uplift record. The record is based on the height and age of various individual fluvial terrace surfaces. These surfaces have been reconstructed by finding best-fit plains through the tops of the respective fluvial sediments (Van den Berg 1989). In this way we were able to ignore the effects of post-depositional erosion or accumulation affecting the height of the terraces. The obtained accuracy is estimated to be in the order of about +/- 1 m.

The positions of the reconstructed terrace surfaces on the time axis were chosen at the very end of glacial periods around the transition to the next interglacial. Due to this climatic shift, floodplain aggradation is replaced by a rapid fluvial incision into the old floodplain (Van den Berg 1993). This is the response to an important change in the bedload/discharge ratio. Such a cold-warm shift lasts only a few thousand years. So within the scale proportions of Fig. 7.6, time-error bars can not be indicated.

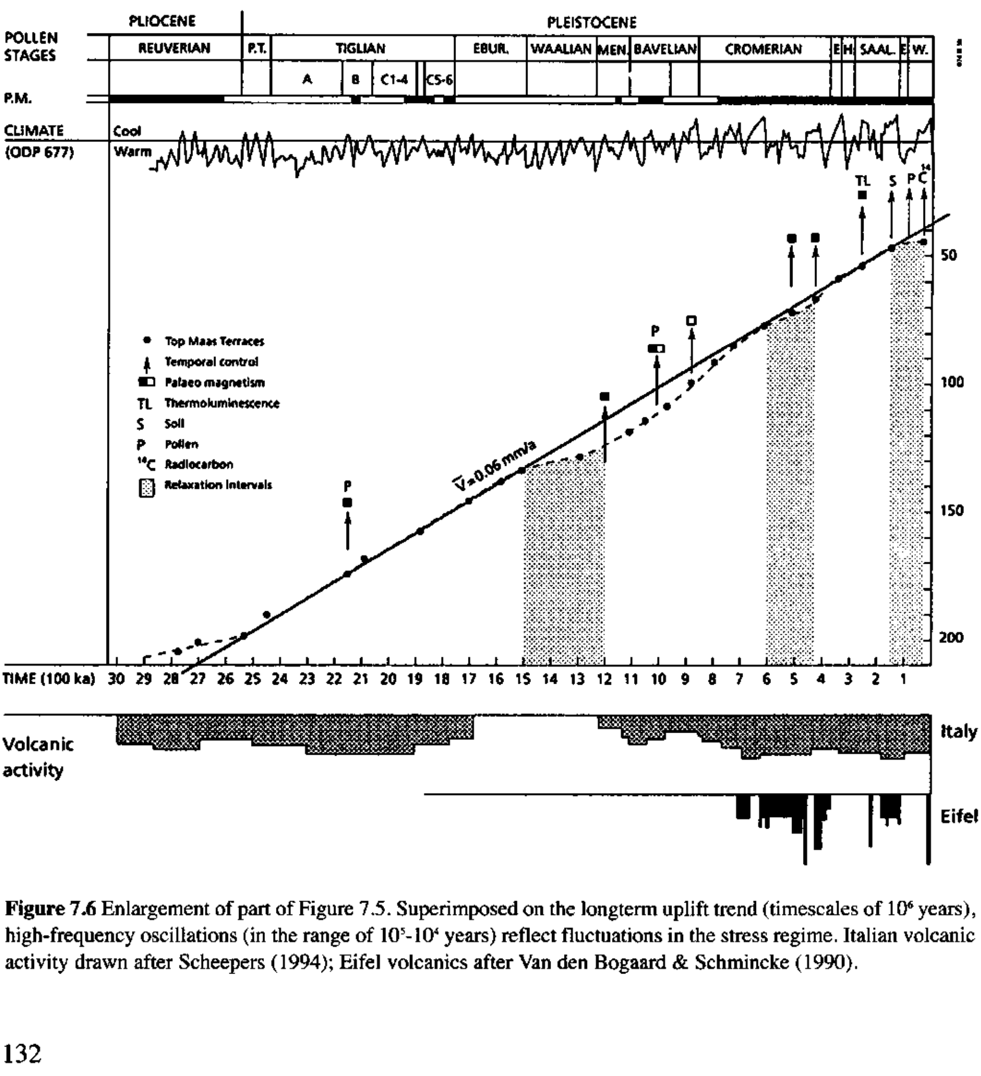


Figure 7.6 Enlargement of part of Figure 7.5. Superimposed on the longterm uplift trend (timescales of 10^6 years), high-frequency oscillations (in the range of 10^5-10^4 years) reflect fluctuations in the stress regime. Italian volcanic activity drawn after Scheepers (1994); Eifel volcanics after Van den Bogaard & Schmincke (1990).

Interpretation

The Pleistocene terrace record shows some clear departures from the average uplift rate of 0.06 $mm.a^{-1}$. These are interpreted to reflect the waxing and waning of the regional stress field. The most important reactivation occurred between 3 and 2.7 Ma, a second one around 1 Ma, and a third one just after 500 ka BP. Precision levelling indicates that the area is presently being uplifted at a rate of 8 (+/-3). 10^{-4} $m.a^{-1}$ (Lorenz et al. 1991; Groenewoud et al. 1991). This rate is in accordance with results obtained for other parts of the Rhenish Shield (Malzer et al. 1983; Ziegler 1992). It is an order of magnitude higher than the long-term average. It suggests a sub-recent strong reactivation, particularly if this value is considered in relation to the low values obtained for the last 200 ka.

The subsiding graben

The record

The Plio-Pleistocene sedimentary succession within the Roer Valley Graben is summarised as a composite section (Fig. 7.7). This section runs perpendicular to the fault systems and is consequently perpendicular or oblique to the trend of the Maas and Rhine feeder systems. The stratigraphic record shows that the river Rhine periodically flowed outside the main graben. During these periods the river was located to the east, like in the present situation. The tectonic controls on these changing palaeogeographic configurations will be exemplified for the period around the latest W-E shift of the Rhine (Fig. 7.8a-c). This level was selected because it lies at relatively shallow depth in the graben and its palaeogeography is controlled by many borings.

Rhine sediments show a characteristic mineral zoning over time (Zonneveld 1949). The youngest Rhine sediments in the graben are characterised by the 'Weert' mineral zone. The areal distribution of this mineralogy (Fig. 7.8a) shows that the system occupied a wide depositional plain with a mid-graben area of non-deposition. The river Maas formed a tributary river. In the course of the next cold stage, the Rhine sediments disappear from the graben and only Maas sediments (the Rosmalen mineral zone; Zonneveld 1964) are found in a pattern that is strongly confined by faults in the middle part of the central PDZ where there is a strong curvature on its fault trace (Fig. 7.8b).

Interpretation

-Palaeogeography:

If we assume right-lateral strike-slip motion along the central PDZ, deformation adjacent to this zone relies on its shape (Christie-Blick & Biddle 1985). The major bends in the fault trace either act as a releasing or as a restraining bend. The first opens up a local pull apart basin (confining the Maas floodplain), while the latter simultaneously forces the encompassed block (the Erft Block) to emerge. As this block is tilted towards the northeast, the river Rhine is forced into that direction.

To meet the optimal angular relationships between stress orientation and the fault line in a strike-slip setting, an additional stress component is required as normal faulting domi-

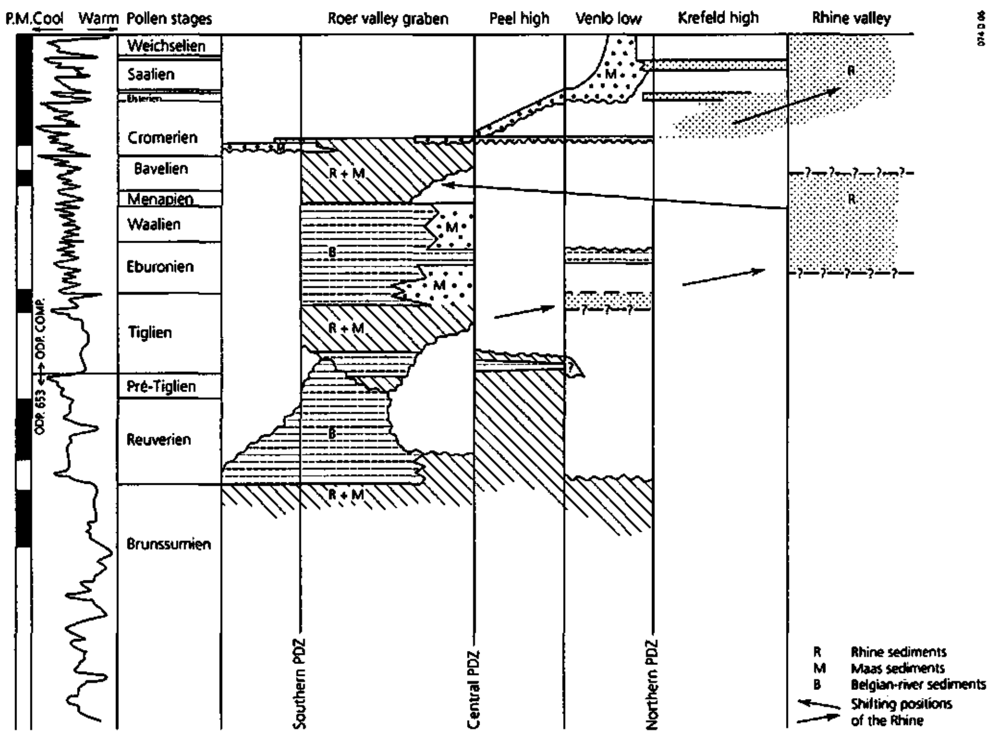


Figure 7.7 Stratigraphic chart showing the sedimentary infill of the Roer Valley rift system by the Plio-Pleistocene Rhine, Maas and Belgian rivers. Indicated by arrows is the shifting position of the Rhine (and Maas) across the principal displacements zones (PDZ). This behaviour is tectonically controlled by strike-slip motion along the central PDZ; see also Figure 7.8a-c. (Sediment record compiled partly after Zagwijn 1960; Van der Toorn 1967; Bisschops 1973; Bisschops et al. 1982; Zagwijn & De Jong 1983)

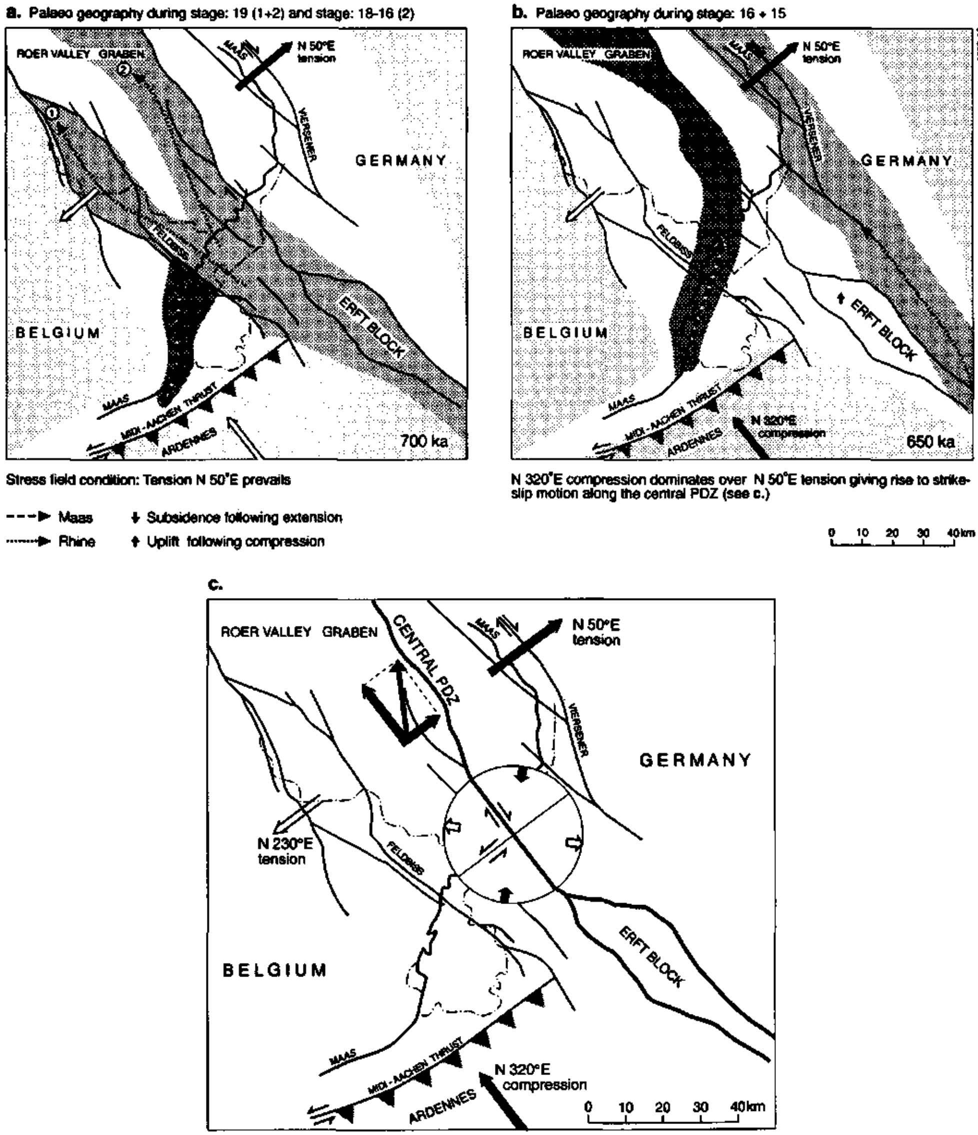


Figure 7.8a-c Palaeogeographic evolution around 700 ka BP showing the last eastward shift of the Rhine. The present-day course of the Maas is shown for orientation purposes. *a*) Over the period equivalent to deep sea stages 19 through 16, normal faulting and tension dominates in the Roer Valley Graben. *b*) In the course of stage 16, transtensional and transpressive movements along the central principal displacement zone (PDZ) cause a paleaogeographic reorganisation in and separation of the positions of the Rhine and Maas floodplains by uplift and tilting towards the northeast of the Erft Block. *c*) The change from normal faulting to strike-slip faulting along the central PDZ is interpreted to result from an increase in foreland compression by the Ardennes Massif along the reactivated Midi-Aachen Thrust.

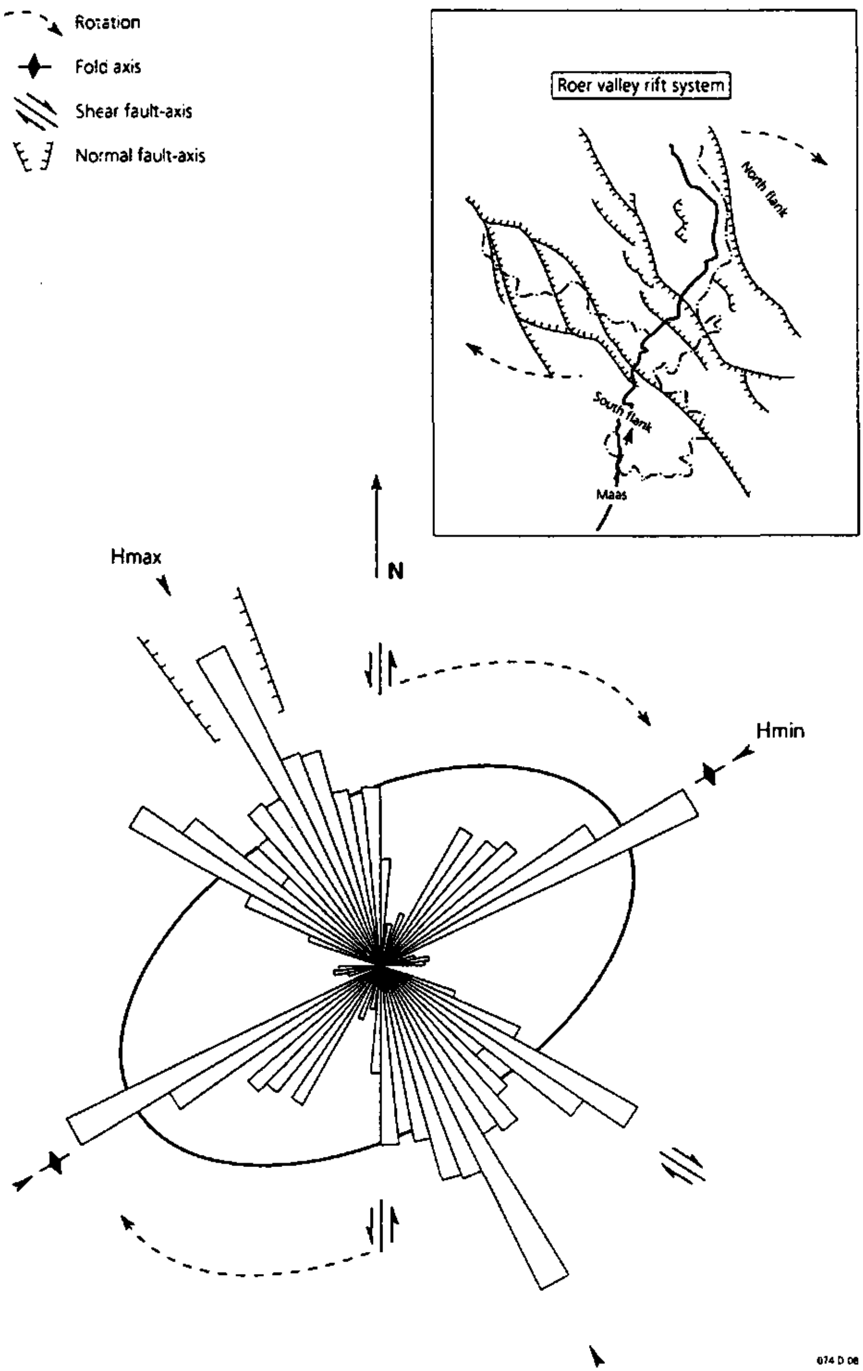


Figure 7.9 Morphologically identified lineaments in the study area formed since around 600 ka represented in a rose diagram (orientation class x cumulative-length percentage, after: Van den Berg et al. 1994). The diagram is consistent with a right-lateral rotation following the horizontal stress components H_{max} N150-160° E and H_{min} N50-65° E.

nates the regional structural style (Fig. 7.8c). It is suggested that the interplay between the regional SW-NE oriented tension and the NW-oriented compression generated by the Ardennes (see above) periodically may lead to such optimal conditions (Fig. 7.8c).

Temporarily active depocentres along the three principal displacement zones are assumed to be likewise indicative for periods when the Ardennes stress field overrules the general tension. For example, within the Voorne Trough, a depocentre arranged along the southern PDZ, about 200 m-thick deltaic deposits of Pre-Tiglian age are encountered. The Pre-Tiglian lasted only 100 000 years and represents a cold stage with low sea-levels. The unusually thick and lithologically uniform deposits suggest therefore an extremely high accommodation rate in the marginal trough during this time interval. This is consistent with the coeval strong reactivation of the uplift of the Ardennes as registered by the Maas river terrace record. Along the northern PDZ (the Viersen Fault), periodic preservation in the Venlo marginal graben also indicates periods with local extension in the releasing bend. Such simultaneously operating processes in basin extension and flank uplift are important markers in the neotectonic evolution of the stress regime.

-Stress-controlled geomorphology:
In Fig. 7.8a, the floodplain splits around a mid-graben non- deposition area within the graben. This floodplain separation is interpreted to represent a mid-graben relative high, bounded by flanking lows. In the Roermond area, the Roer Valley Graben is also morphologically characterised by a persistent central high (with terraces = uplift) and two flanking fault-bounded lows. Both from the seismic lines and from the morphology we know that such a tectonic low in itself repeats such a relief pattern. This tectonically defined morphology shows a sequence of scales: the fractal-alike pattern that emerges shows a strong similarity with the stress-responding differential flexural crustal motions at basin-wide scales in a tensional setting (Cloetingh et al. 1985). This sequence may suggest that the crustal mechanics of a divergent strike-slip setting operate at a local, a regional and a basin-wide scale. The scale determines the nature of the bounding structures: faults or flexures.
Shortly after the Rhine shifted to the east, the Maas also shifted out of the graben and slipped off the tilted Peel Block towards its present position. The present-day surface of this series of abandoned floodplains shows a well-defined deformation pattern, expressed as a low relief (1 to 2 m scale) of structural lineaments (Van den Berg et al. 1994). Figure 7.9 shows the rose-diagram (cumulative length x orientation-class) of the lineaments in the area. The orientations are consistent with the results obtained in an area east of the Viersen Fault (Plein et al. 1982).
In accordance with the palaeogeography, the geomorphology indicates the ongoing right-lateral divergent strike-slip faulting due to block rotation over the last 600 ka. This is the most recent pulse in a series.

Plate-tectonic implications and discussion

We identified a number of accelerations and decelerations (pulses) superimposed on a generally high rate tectonic phase since the last 3 Ma. There are indications from other parts of Europe that important physiographic changes tend to cluster around these pulses.

For example, near the Gauss-Matuyama magneto-boundary a thrusting pulse of the Jura mountains causes the connection of the Upper Rhine with the Swiss Alps. This has been identified in the heavy-mineral composition of the Rhine sediments (Boenigk 1982; Tebbens et al. 1995).

The Alpine geodynamics are thought to be the mechanical consequence of the Africa-Europe collision. The presence of intra-plate stresses provides a possibility to understand a coupling between the inter-plate dynamics of the two plates and the pulses observed in our study area. It is suggested here that the relaxation interval between 1.5-1.2 Ma (Fig. 7.6) may be the consequence of an observed change in direction of the relative motion of the African plate. During a part of the Early Pleistocene this direction temporarily changed from northwest towards northeast; this is extensively studied in the Tyrrhenian arc system by e.g. Van Dijk & Scheepers (1994) and summarised by Scheepers (1994). From the same area Brogan et al.(1975) report a profound change in the activity of the Tyrrhenian arc about 1.1 Ma.

Contemporaneously with this last event, uplift and consequent terrace formation begin in the central Appenines, Italy (Coltorti 1993), the Limagne, France (Allier river, Veldkamp 1991), on the north flank of the Paris Basin (Somme river, Antoine 1993) and in the Bohemian Massif (W.H. Zagwijn, pers. comm. 1993).

The relaxation phase found between 600 and 430 ka shows a strong synchronism with phase 2 of the Eastern Eifel volcanic activity (Van den Bogaard & Schmincke 1990). The other recognised phases in the Eastern Eifel volcanism are relatively short with respect to the resolution of the changes in uplift rate registered by the Maas river record, although there is synchronism between the last recorded relaxation phase and phase 5 in the volcanic activity.

Although this list of records is far from systematic and complete, they suggest that the Maas terrace flight identifies a lithospheric signal of plate-tectonic importance.

The coincidence of regional uplift of ancient structural domes and sinking basins, together with the formation of pull-apart basins due to divergent strike-slip motions in response to foreland compression by the Alps and the Rhenish Shield, may suggest that these processes are controlled by important phases in the plate reorganisation (Cloetingh & Kooi 1992).

The observed accelerations in uplift from the Tertiary to the Pleistocene, in conjunction with the acceleration of global volcanism throughout the Quaternary (Kennet & Thunell 1975) and the present-day high velocities of crustal movements with respect to the Pleistocene long-term average, raise the intriguing question: Do we live at present in a period of extreme crustal dynamics?

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8. Patterns and velocities of recent crustal movements in the Dutch part of the Roer valley rift system

Abstract - *This article presents an integration of geomorphological and geodetic data from the area of the 1992 Roermond earthquake. A dense network of lineaments is evident from major and minor terrain features, and drainage patterns also show structural control on a kilometre scale. These discontinuous terrain lineaments often of anastomosing character, match known fault patterns, and suggest that the upper crust is subdivided into many, relatively small (up to 10 km scale) wedge-shaped blocks. The lineament distribution is consistent with patterns predicted by idealized strain ellipses. It shows a right-lateral component in the motion along major faults within the Lower Rhine Embayment. The wrenching component can be related to a left-lateral motion along the Variscan Front, and a subsequent right-lateral offset of the edge of the London-Brabant Massif. The analysis of a 117-years-long data set of vertical movements at 2922 geodetic bench-marks evidences significant differential movements, and corroborates the sense of relative motion given by the lineaments.*

Introduction

In the south-eastern part of the Netherlands the NW-SE fractured zone of the Lower Rhine Embayment, an element of the central European rift system, is well expressed on the surface.

The area was the scene of the Roermond earthquake on 13 April 1992. A combined analysis of geomorphological and vertical velocity data for uplift and subsidence of this area will be presented.

The rationale for combined geomorphological and geodetic analysis in neotectonic research is that the former can give indications of the locations of lineaments and the sense of the movement. Geodetic research can give insight into the velocity of the movements, which otherwise can only be obtained as an average for long periods of geological time. This is important as recent research has shown that Quaternary velocities differ significantly from Late Neogene velocities and, that within the Quaternary significant accelerations and decelerations seem to have occurred (Van den Berg, 1994). Recently, rates of vertical movement for about fifty selected bench-marks located throughout the Netherlands were calculated from the extensive historical data base of the Dutch ordnance datum (NAP).

This resulted in a map of regional vertical movements in the Netherlands, which shows a striking similarity with geological structures in the subsoil (Lorenz et al. 1991; Groenewoud et al. 1991).

This map has now been improved using levelling data of more than 20 000 surface bench-marks. From this analysis, a detailed and precise vertical-velocity map of the study area was obtained, enabling a first check on the neotectonic movements predicted by geomorphological studies. In interpreting the results one needs to be aware of other factors causing vertical movements such as compaction, mining activities and glacial rebound.

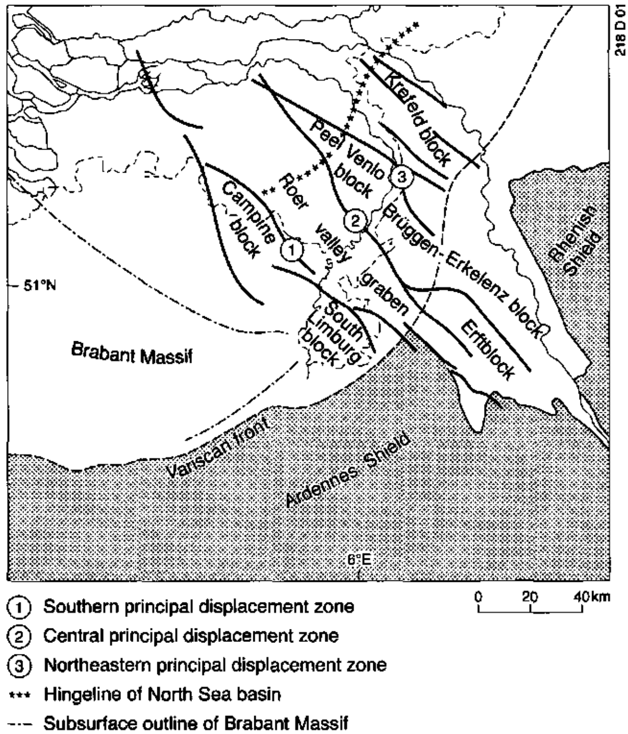

Figure 8.1 Major structural elements in the study area and adjacent parts of the Lower Rhine Embayment.

Geological setting

The study area is situated in the transition zone between the southern North Sea Basin and the foreland of the Variscan Fold Belt. It borders the north-east of the London-Brabant Massif (Fig. 8.1) and is dominated by the extensional tectonics of the Lower Rhine Embayment.

Three main landscapes have been distinguished, from south to north (Fig. 8.2):

1 the South Limburg area, characterized by strongly dissected loess-covered Pleistocene river terraces underlain by Cretaceous chalk or Tertiary sands;

2 the central area, characterized by flat Mid to Late Pleistocene (< 800 ka) braid plains covered by aeolian sand sheets. Local relief only rarely exceeds a few metres and slopes are less than 1°. Minor relief features stand out clearly in the flat terrain;

3 the northern area, comprising the Holocene flood plains of the rivers Meuse and Rhine. Lineaments are more difficult to trace here because of the young age and strong lithological contrasts; therefore, this part has not been included in the analysis.

The edge of the London-Brabant Massif strikes about N130°E. The Variscan Fold Belt forms a main structural element (thousands of km scale) in north-western Europe (Ziegler, 1990) with a regional orientation in the south-western part of the Lower Rhine Embayment of N60°E, turning to N35°E in the north-eastern part. The embayment is dissected by a few

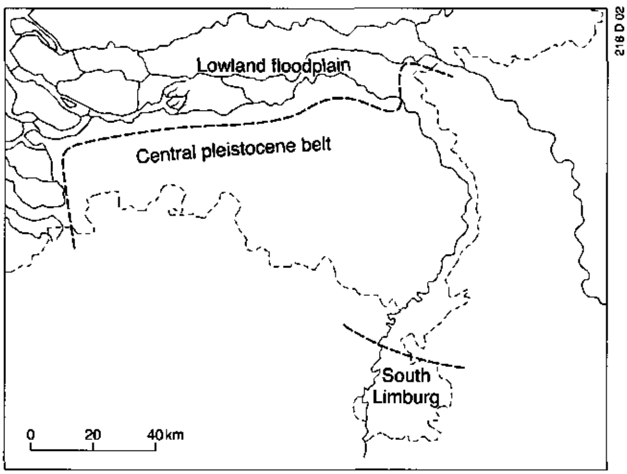

Figure 8.2 Regional geomorphology of the southeastern Netherlands.

(an order of magnitude shorter) NW-SE oriented fault zones. Their orientation with respect to the Variscan Front and the London-Brabant Massif suggests that they form an antithetic set with them (we use the terminology of Christie-Blick & Biddle, 1985). Due to the bend in the Variscan Front, some of these fault zones dissect each other a hundred kilometres away in the foreland. This pattern determines the main structural grid, encompassing a number of blocks of which the Roer Valley Graben, the Peel-Venlo Block, the Erft Block, the Krefeld Block and the Campine High are the principal elements (Fig. 8.1).

Although the fault zones form secondary structures, we will refer to them hereafter as principal displacement zones (PDZ), because they form the main grid and we want to avoid local names for different sections. We recognize three PDZs: a southern, a central and a north-eastern. The southern PDZ separates the South Limburg Block and the Campine High from the Roer Valley Graben. The latter is separated by the central PDZ from the Peel-Venlo Block, which in turn is separated from the Krefeld Block by the northern PDZ.

The main trend of the central PDZ is N150°E, whereas the trend of the north-eastern PDZ is N120°E. The Peel-Venlo Block forms a wedge between the central and the northeastern PDZ and is bordered to the northeast by the Krefeld Block. The northern PDZ forms the continuation of a series of parallel main faults in adjacent Germany (Plein et al. 1982). There these faults show well-developed flower structures on seismic reflection lines, indicative of wrenching (Christie-Blick & Biddle 1985). These main units are again subdivided into many subunits, some of which will be discussed below.

Morphostructural analysis

Due to their long term behaviour, slow crustal movements along faults show up as surface features (e.g. smoothed terrain-steps) of variable size. The study of these features provides valuable information on recent tectonics. Mapping and analysis of these surface features can be done in various ways, e.g. from satellite imagery and field documentation of geomorphic and stratigraphic relationships. A problem with satellite imagery is the discrimi-

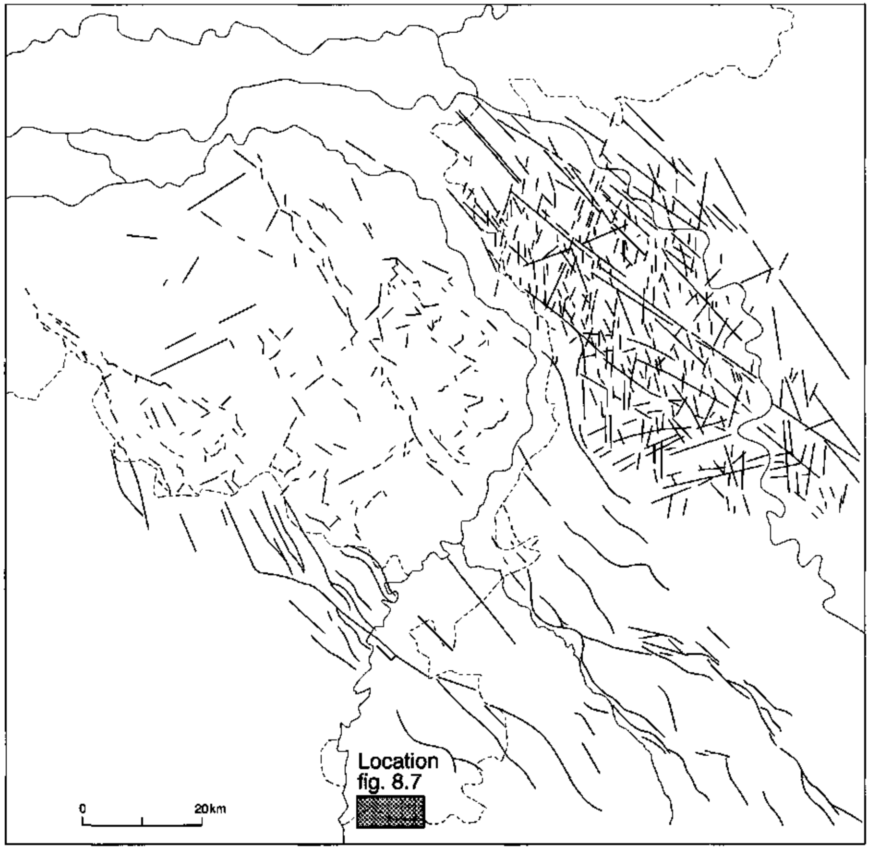

Figure 8.3 *left* Surface expressions of tectonic lineaments in the study area and its vicinity. Data in Germany from Plein et al. (1982; landsat images) and Ahorner (1962; geological study); in Belgium from Demyttenaere & Laga (1988; seismic) and Paulissen (1973; geomorphology).

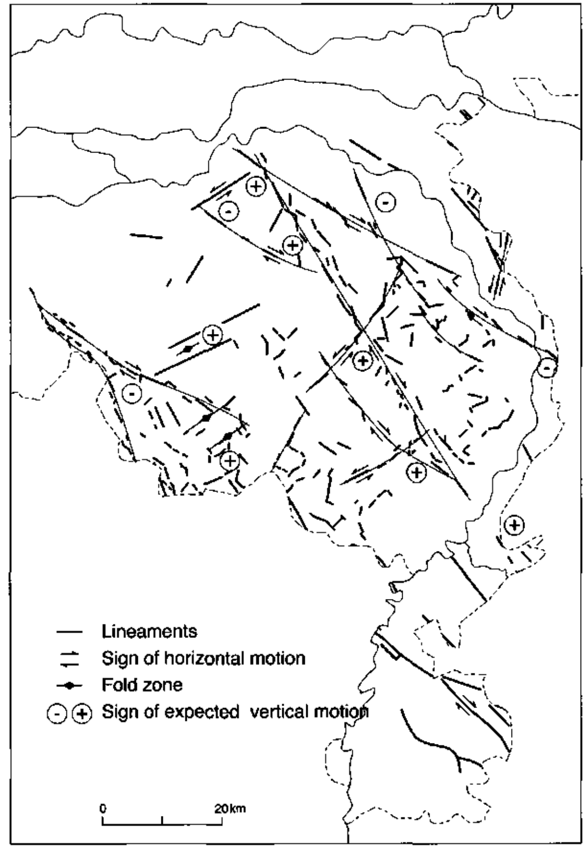

Figure 8.3 ***right*** Main structural lines inferred from surface lineament arrays in Figure 8.3*left*. The arrangement of structures indicates a right-lateral shear component within the general extensional setting.

nation between natural and artificial, man-induced, rectilinear features. However the resolution from satellite imagery is high.
The first step in the analysis was to prepare a contour map with 0.5 m intervals, drawn manually from point elevations on the 1 : 10 000 elevation map of the Topographical Survey of the Netherlands (not yet available in digitized form). The latter map is based on third-order levelling data points arranged in a more or less regular grid of 100 m x 100 m. The altitude information is given with 0.1 m accuracy. The next step was to identify discontinuities in the surface morphology. This was done by constructing long surface profiles parallel and perpendicular to the main strike of the contour lines. These profiles were drawn at regular distances of 5 km. Abrupt breaks in slope were mapped.

Using the regional geology and geomorphology we classified the identified surface discontinuities into man-made structures and natural features of aeolian, fluviatile and tectonic origin. Man-made structures are known from the topographical maps. Aeolian forms are divisible into two groups: (1) an undulating cover sand topography with local height differences less than 0.5 m, this type does not show up on the surface profiles; (2) large dune complexes (either with a strongly undulating relief, or a whale-back shape), which appear to be situated on top of a continuous surface; therefore these features could easily be excluded. The subdivision of fluviatile forms into purely fluviatile and structurally controlled forms is less straightforward. Vandenberghe (1990) has shown that local drainage systems, as developed on the Campine Block are strongly structurally controlled; but also braided or meandering channel patterns and incised valleys of the river Maas appear to line up with fault lines. So some of the riverbends form a part of the lineaments.

The main criteria, applied by several observers, to distinguish tectonic lineaments (Fig. 8.3*left*) from each other or accidental alignments were:

- a minimum length of one kilometre or alternatively arrays of shorter discontinuous lineaments in line with each other;
- rectilinear jumps in altitude or slope breaks;
- rectilinear segments in the drainage pattern.

The striking difference in lineament-density between our study area and the area studied by Plein et al. (1982) is the result of different analysis techniques.

Determining land surface deformation

The Dutch ordnance datum NAP

All heights in the Netherlands are related to the Dutch ordnance datum NAP (Normaal Amsterdams Peil). This reference level was fixed in 1684 by nine stones in sluices in Amsterdam. Due to reconstruction works, these stones have all disappeared. The last stone, which was removed in 1953, was replaced by a 23 m long foundation pile at the Dam square in Amsterdam. This pile now acts as the fundamental bench-mark of the NAP height datum.

For practical and theoretical reasons several bench-marks with very accurate heights are needed. These underground bench-marks, which serve as regional markers for the NAP, have been installed since the so-called Second Primary Levelling (1926-1940). At present there are more than 150 underground bench-marks all over the Netherlands.

For determining land surface deformation, the foundation of a bench-mark is of prime importance. In the so-called 'order' of a bench-mark an indication can be found of its foundation. For both first-order and second-order bench-marks the sites have been carefully selected, soil properties and geological conditions have been investigated, and all first-order and second-order underground bench-marks have been placed in the upper reaches of the Pleistocene sands. These sands were assumed to be stable. For third-order underground bench-marks, no geological research was carried out.

Since the stability of the underground bench-marks was of utmost importance, all of them were specially designed. This has resulted in a construction that protects the bench-mark from any vertical movements due to forces acting in the sediments above the foundation level (e.g. dragging from compacting Holocene sediments). Movements of the foundation level, however, cannot be ruled out. These movements have always been considered to be negligible. However, recent investigations have shown that significant movements do occur (Lorenz et al. 1991). This is an important observation, since the heights of the underground bench-marks have never been corrected for these movements.

Apart from the underground bench-marks, there is a dense network of 'surface bench-marks'. This network, which is much older than that of underground benchmarks, is available for practical purposes. The bench-marks predominantly consist of bolts that are placed in houses, churches, bridges, etc. The entire Dutch network consists of over 50 000 bench-marks, which means that, on average, there is one in every square kilometre.

In contrast to the underground bench-marks, surface bench-marks are 'free to move', i.e. it is accepted that, due to various mechanisms, vertical movement of the bench-marks is likely to occur. For this reason, the heights of all bench-marks are determined on a regular basis (at least once every ten years) using levelling techniques. Lithological and geological conditions must be taken into account when interpreting the movements of surface bench-marks. In the study area, the outcropping rocks include Cretaceous chalk in the extreme south, Pleistocene sands in the central part, and Holocene sands and clays in the north. Apart from the chalk, all the sediments are unlithified. Except for the Holocene sediments, all rocks are considered to be well consolidated.

Movement analysis

From repeated height determinations, an extensive data base on historical height data has been obtained. In spite of the yearly loss of about 3% of the bench-marks, long time series (up to 117 years) of changes in height are available for many of them. Various criteria have been applied for selecting the data. Of prime importance is the order of the selected levelling campaigns; first-order, second-order and third-order data are available for the movement analysis. These orders of levelling indicate the precision requirements for the levelling campaigns; a first-order levelling is the most precise. However, since the densities of the networks of second-order and third-order levellings are much higher than those of the first-order network, the redundancy of data, results in height values with fairly comparable precision for all orders of levelling (Table 8.1).

From the selected data, rates of vertical movement or "vertical velocities" can be determined quite easily. For this purpose, a regression analysis was conducted for each bench-mark record. In our study, we analyzed data from first-, second- and third-order levelling campaigns. The statistics of movement analysis are given in Table 8.2.

One has to bear in mind that the heights of the surface benchmarks have always been determined relative to the underground benchmarks. This implies, that the recently observed height changes of the underground benchmarks are not incorporated in the records of height changes of surface benchmarks. Since the abrupt lateral changes (which obviously do not depend on absolute values) in the distribution of vertical velocities are the most important indicators for neotectonic activity, the research is not seriously hampered by this lack of absolute information.

Spatial interpolation

A regional map of vertical velocities has been constructed to compare these velocities at individual points with geological and geomorphological data. In principle, various techniques of interpolation can be adapted for this purpose, each with its own theoretical background and applications.
We used the kriging technique to construct the maps shown in Fig. 8.4 . An important advantage of this technique is that it provides estimates of the accuracy simultaneously with estimates of the spatial distribution of the velocity. For detailed information we refer to Journel & Huijbregts (1978), Isaaks & Srivastava (1989), and Webster & Oliver (1990). The kriging technique was slightly modified to account for the uncertainty of the vertical velocity at the bench-mark points estimated by regression analysis (Ahmed & De Marsily 1987). In this modified kriging technique the semivariogram based on errorless measurements of velocity is used. We approximated this semivariogram by the one based on regression estimates with small error (estimated variance < 0.01).

We assumed that the mean and variance of the vertical velocity will be different for the major tectonic units.
Therefore we estimated semivariograms for these units separately and used these in stratified kriging (Stein et al. 1988). The vertical velocity at a new location was estimated by using the twenty nearest observation points in the same stratum. Figure 8.4a shows the result. The root of the kriging variance is shown on figure 8.4b. The kriging variance is strongly determined by the semivariogram, explaining the different levels of the kriging variance for the strata.
For example, the nugget and the sill of the semivariogram of the Peel Horst (Fig. 8.5a) are much smaller than for the Roer Valley Graben (Fig. 8.5b), which explains that the kriging variance is relatively small here.

Table 8.1 Precision requirements (in mm) as a function of the measuring distance l (in km) for 1st, 2nd and 3rd order levellings

	1st order	2nd order	3rd order
Section	$2.5 \sqrt{l}$	$3.0 \sqrt{l}$	$6.0 \sqrt{l}$
Traject	$2.5 \sqrt{l}$	$3.0 \sqrt{l}$	$6.0 \sqrt{l}$
Circle	$1.3 \sqrt{l}$	$1.5 \sqrt{l}$	$3.0 \sqrt{l}$

(Example: for a 49 km circle of a 1st order levelling, the precision should be better than $1.3 \times \sqrt{49} = 9.1$ mm)

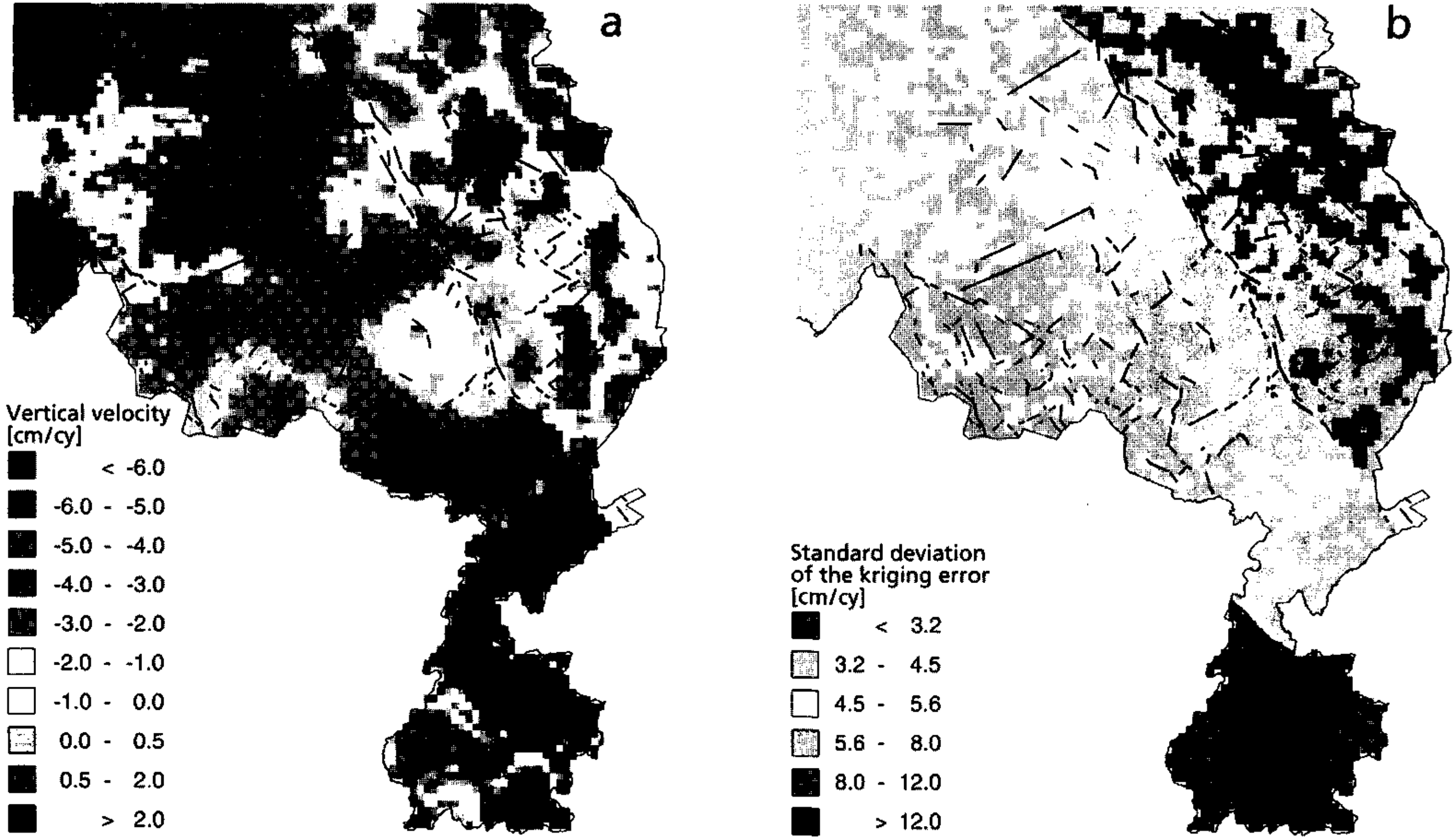

Figure 8.4a Vertical velocities estimated in cm per century (cm/cy) by kriging. Scale 1:1.000.000 (for full colour display see original paper).
Figure 8.4b Kriging standard deviation in cm per century (cm/cy) of the estimated vertical velocities. Scale 1:1000000 (for full colour display see original paper).

Table 8.2 Input and output statistics of vertical movement analysis

<u>Input statistics</u>	
number of bench-marks:	11 207
number of measurements:	34 142
selected levellings:	1st, 2nd and 3rd order
first measurement:	1875
last measurement:	1992
minimum interval:	10 years
<u>Output statistics:</u>	
vertical velocities calculated for:	2922 bench-marks
number of measurements:	17105
average vertical velocity:	-18 mm/100 yrs

Results and interpretation

Regional warping

To evaluate the pattern displayed by the lineaments, we first plotted the lineaments (cumulative length percentage x orientation-class) in a rose diagram for each major tectonic unit (Fig. 8.6). The different tectonic domains were then analyzed.

The rose diagrams show strong concentrations around NW-SE (N140-160°E, N115-130°E) and NE-SW (N55-65°E), whereas a minor peak occurs around N40°E (Fig. 8.6). The diagrams for the N.E. flanking area (Peel-Venlo Block) and the SE flanking area (Campine High) are more or less similar, whereas NE-SW orientations predominate within the Roer Valley Graben.

The lineaments are clearly discontinuous, but show grouping of the individual lines into various long, spatially separated, arrays. In many cases, these composites coincide with known fault lines and line up with structures known from Belgium (Paulissen 1973; Demytenaere & Laga, 1988) and Germany (Ahorner 1962; Plein et al. 1982; Fig. 8.3*right*). The angular relationships among the short lines as well as among the long composites are persistent at about 60°, 15° and 75°. This recalls the angles of shearing resistance in Riedel experiments (Tschalenko 1970). The central PDZ is a particularly clear example of a fault zone composed of a network of branching and rejoining elements. This pattern can also be recognized in the other composites.

All the above mentioned features together emphasize the tectonic origin of the individual lineaments. The anastomosing appearance of the PDZs (with the 15° angle of the second-order lineations) is a strong indication that horizontal movements are in progress along these fault zones. This process of shear will lead to a structurally complex pattern of local uplifting and downwarping blocks. We established a hierarchy in the lineaments to predict the local movements from the lineament pattern.

The lineament pattern breaks up into associated elements that form arrays of different lengths. This length is taken as a key to the hierarchy. The highest level is formed by the

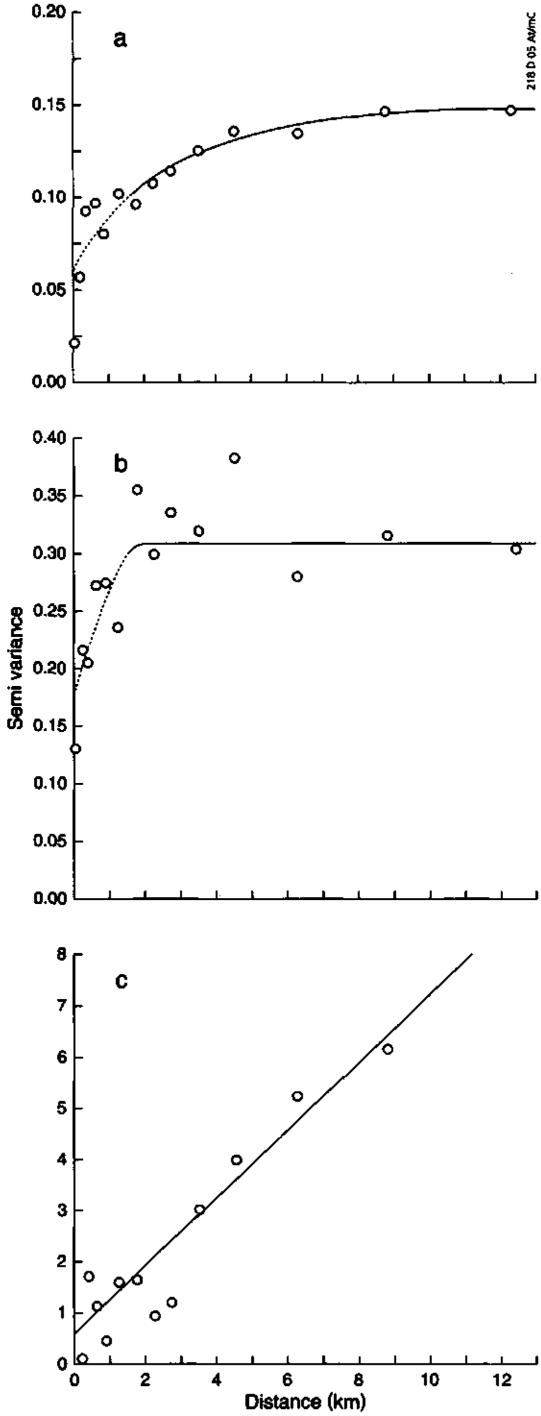

Figure 8.5 Sample semivariograms and fitted models
a: Peel-Venlo Block, b: Roer Valley Graben, c: Campine High together with South Limburg.

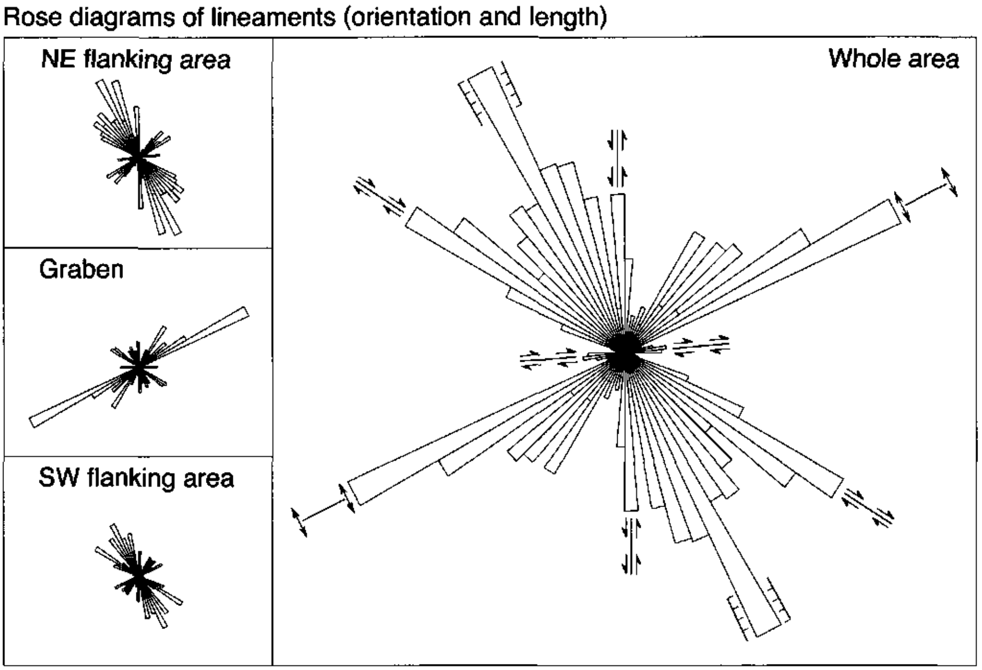

Figure 8.6 Surface lineaments presented as rose diagrams (length ¥ orientation) for the entire study area (‘whole area’) the Peel-Venlo Block (‘NE flanking area’), the Roer Valley Graben (‘Graben’), and the the Campine High (‘SW flanking area’).

longest continuous lines, i.e. the Variscan Front and the edge of the London-Brabant Massif. The second order is taken by the PDZs and the third and lower orders are taken by subsequently shorter elements. Using the mutual angular relationships between the hierarchic elements and an idealized strain ellipse (Harding et al. 1985) we determined the expected sense of movement on both sides of the shearing zones (Fig. 8.3 *right*).

The distribution of vertical velocities roughly coincides with the principal structural elements (Fig. 8.4). Topographic highs, such as Netherlands South Limburg and the Peel-Venlo Block, are rising and topographic lows, such as the graben and the riverine area, are subsiding. But several apparent inconsistencies in this general picture are also evident. The smaller-scale patchy pattern within the blocks suggests that the main domains are divided into minor domains. This is especially striking within the graben. Furthermore, a conspicuous anomaly is that the river Maas flows through an uplifting area for a considerable stretch of its course, whereas the Rhine is confined to an area of subsidence. Another striking feature is the subsiding north-eastern part of the Peel-Venlo Block. The relations of this general picture with the structural elements are discussed below.

Differential movements

-South Limburg

The South Limburg Block forms a structural high, wedged between the Variscan Front (the Midi-Aachen Thrust) and the southern PDZ of the rift. Uplifting has been going on here since the Miocene, as evidenced by the river terraces (Van den Berg, 1994). Although we

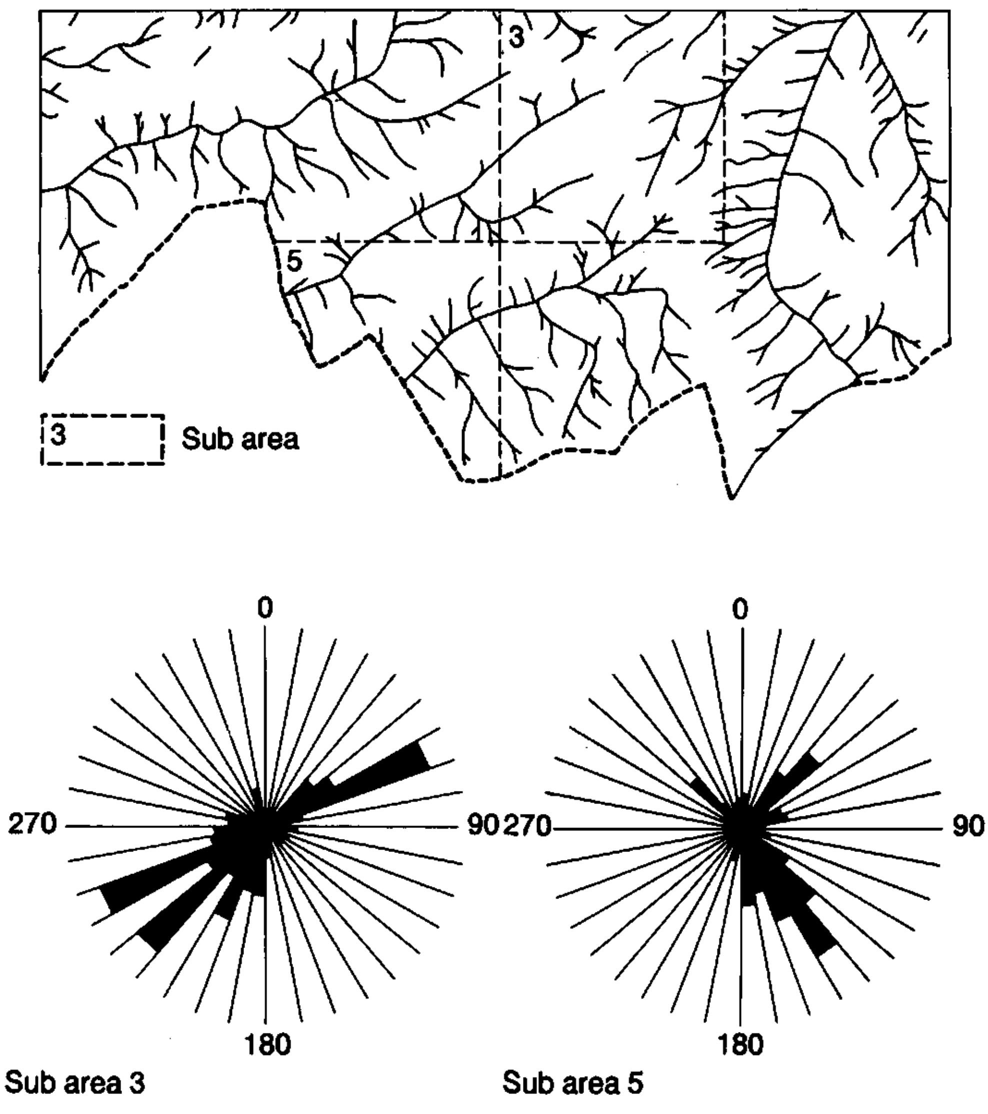

Figure 8.7 An example of the structurally controlled valley orientation of the secondary drainage pattern in South Limburg (For location and Figure 8.3). Rose diagrams (valley length x orientation) are shown for two subareas. Mapped area measures 10 km E-W.

know that the subsurface is strongly faulted (Felder & Bosch 1977), the fault system is expressed on the surface rather by the orientation of the secondary drainage pattern than by terrain steps (Fig. 8.7). For this reason the lineaments have not been indicated on the map of figure 8.3*a*

Some strong lineaments emerge from the congruency of Maas river terrace bluffs of different ages. This pattern probably reflects crustal buckling in the foreland of the Variscan Front during the Plio-Pleistocene (Fig. 8.2 in: Van den Berg 1994). Such movement is not apparent from the present-day vertical velocities.

Fault lines with a clear surface expression are almost all related to the southern PDZ. This zone itself is well expressed, both in the field by a smoothed scarp of 10-40 m height and on the velocity map by the opposite sense of vertical movements on both sides of the zone, i.e. subsidence at a rate of 0.7 mm.a-1 in the north and a general rapid uplift of 0.8 mm.a-1 in the south. The main direction of the Maas river valley north of Maastricht is controlled by the conjugate faults associated with the PDZ. South of Maastricht the main orientation changes and becomes antithetic to the Variscan Front.

For a 10 km wide area south of the southern PDZ, we would expect some relative downthrow. However, extremely high uplift velocities (> 15 mm.a-1) are recorded here. They are interpreted as artefacts. This is the old coal-mining district; the observed uplift is calculated from measurements that have been collected over a period of about 20 years, and is caused by recharge of groundwater into the abandoned mines (Pöttgens 1990).

-Campine Block

The Campine Block is one of the fractured structural highs wedged between two left-stepping fault elements of the southern PDZ (Geluk et al. 1994).

In map view, the southern PDZ splays 15°E in Belgium (Demytenaere & Laga 1988) and forms a 'lazy S' before it rejoins with faults normal to the edge of the London-Brabant Massif. A pull-apart basin is encompassed. The NW-SE oriented short lineaments within this basin reflect normal and reverse extensional faults. This extension is in accordance with the observed subsidence. The registered uplift in the south-eastern part coincides with NE-SW oriented short lineaments. Their orientation is normal to the extensional features and, within a strike-slip setting, they may consequently reflect folding associated with a horizontal motion.

-Roer Valley Graben area

The rose diagram drawn from the lineaments within the graben area is dominated by NE-SW orientations, which represent shearing structures antithetic to the PDZs. A subordinate orientation peaks at N120°E, interpreted to be formed by the associated P-shears. The latter occur preferentially along the central PDZ. Presumably this is determined by the structure of the main grid.

These third-order structures delineate blocks, which corners experience local convergence or divergence depending on the relative sense of horizontal motion. The predicted movements at the fault junctions can in many cases explain the apparent inconsistencies mentioned above.

As indicated by Geluk et al. 1994), the Roer Valley Graben can be divided into two units with different structural styles: a south-eastern part and a north-western part. The Veldhoven Fault forms the boundary between the two units. Although its eastern continua-

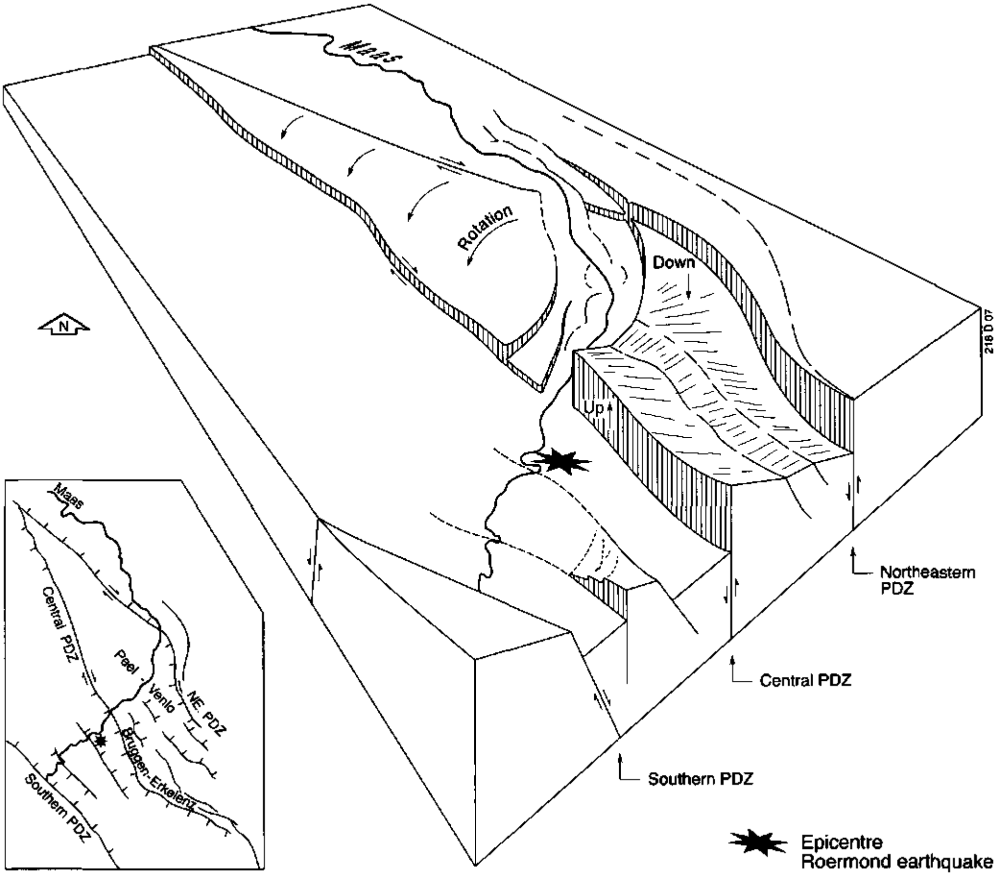

Figure 8.8 A model for the relative motion of the Peel-Venlo Block. This block and its eastern continuation, the Brüggen-Erkelenz High, is wedged between two right-lateral, principal displacement zones. Anti-clockwise rotation may generate torsion of the block. This leads to rising of the Brüggen-Erkelenz High and simultaneous sagging of the Venlo Graben, in the northeastern part of the block. (The block-diagram is schematic).

tion is not yet clear from the available seismic lines, it can be recognized on the surface as a pronounced, composite lineament with a general N40°E strike, right across the graben and continuing on to the Peel-Venlo Block. The surface elevation of both the graben and the Peel-Venlo Block is significantly higher south-east of this line than north-west of it.

In the southeastern part of the graben area, between the river Maas and the Dutch-German border, rapid subsidence can be observed (fig. 8.4a). This contradicts the geophysical and geomorphological data, from which only minor subsidence, or even uplift is expected.

The anticlinal structures in the subsurface, indicating a compressional regime (Van den Berg 1994), are believed to be responsible for this. Maybe the combined effects of increased regional groundwater extraction and of browncoal mining in the German part of the graben, overrule the expected movements. A striking feature is the narrow uplifting zone along the Maas amidst a wide subsiding area. In this area large-scale gravel extraction takes place, and possible causes the relaxation.

-Peel-Venlo Block

The Peel-Venlo Block continues south-eastward into the Brüggen-Erkelenz High. Together they form a north-eastward tilted, triangular block between the central and north-eastern PDZ. The trend of the straight central PDZ is about N150°E, whereas the north-eastern boundary has a curvilinear trace, diverging from N150°E towards N120°E. The bifurcating lineament pattern of the latter PDZ suggests a spreading between the Peel-Venlo Block and the Krefeld Block, leading to a local sagged area: the Venlo Graben (Fig. 8.8).

The stratigraphy in this Venlo Graben shows a discontinuous sedimentary record: sedimentation (= extension) alternates with non-deposition (= compression). This subregion is currently experiencing strong uplift.

The Peel-Venlo Block is wedged between two right-lateral shear zones, with the central PDZ terminating much further south than the north-eastern one (Fig. 8.8). This difference, combined with the regional NE-SW extension, causes apparently an anti-clockwise rotation of the Peel-Venlo Block, this could explain the relative uplift of the Brüggen-Erkelenz part and the consequent tilting to the north-east together with the general flow direction of the river Maas. Shortly after the Maas crosses the eastern PDZ, the main direction of the river channel turns towards N145°E. This is the orientation of associated extensional structures expected under conditions of right-lateral shear. This extension is also reflected in the observed subsidence just north-west of the river. The movements along the main faults cause stress, released by second-order shears in the blocks in between. The direction of the second-order trellis-like drainage pattern on the Peel-Venlo Block and the NE-SW oriented sets of lineaments reflect this stress. The second-order strain pattern is not evenly distributed over the wedge-shaped Peel-Venlo Block, it appears to be concentrated within the, topographically high, south-eastern half. The boundary between the two areas is formed by the long composite lineament mentioned earlier. The structural style as well as the surface altitude of the Roer Valley Graben changes across this lineament.

-Hinge line of the North Sea Basin

As already discussed, a strong lineament separates both the Roer Valley Graben and the Peel-Venlo Block into regions with different structural aspects. This lineament is not yet fully understood, but it marks an ancient structure in the subcropping Carboniferous, as already observed by Van Waterschoot van der Gracht (1918).

The difference in height of the land surface is also observed in the Maas valley on the north-eastern side of the Peel-Venlo Block. As soon as the river crosses this lineament, older river terrace levels become buried under younger ones (terrace intersection zone). Outside the study area, the lineament lines up with the zone of outcropping Tertiary deposits in the east of the Netherlands. On the basis of these features, we assume that this marks the hinge zone of the North Sea Basin. This interpretation also explains that, generally, lineaments are less clear to the north-west of this structure.

-Holocene riverine area

In the Holocene riverine area, lineaments are extremely difficult to trace at the land surface. There are at least three reasons for this:

1 The area is covered by subrecent Maas-Rhine sediments, increasing in thickness towards the west (up to 10 m). This sediment cover obliterates Pleistocene surface deformations. Moreover, Holocene sedimentation rates (Törnqvist 1993) balance the rate of present-day deformations.

2 Differential compaction of floodplain sediments complicates the interpretation differences in landsurface-altitude.
3 The area lies basin-ward of the hinge line of the North Sea Basin. Here, the fault offsets are generally less than those south of this line, hence they are less likely to be traced on the surface.

Conclusions

This study demonstrates that the resolution of structural detail becomes far greater when subsurface analyses are complemented with land surface analysis. The principal fault elements in the Lower Rhine Embayment show many characteristics of ongoing horizontal shear. They are expressed on the surface as well-defined, but discontinuous lineaments. The orientation and geometry of the composite lineaments correspond with structures predicted from the deformation of homogeneous isotropic materials.

The analysis of bench-mark records shows that the overall pattern of distribution of the vertical velocities fits with the known main structural elements. Significant deviations from the general picture can be explained by adding a right-lateral shear component to the generally accepted regional extension. Such a motion is also documented instrumentally from earthquakes showing a dip-slip or strike-slip origin (Ahorner 1975). This suggests that for other areas with a low-frequency seismicity, the combined approach of morphostructural analysis and long geodetic time series can reveal the stress field.

Our study shows that also in areas of relatively low deformation rates, tectonic processes are fundamental to the understanding of supposedly climatically controlled land surfaces and drainage systems.

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Overzicht

Doel en kader

Geologisch onderzoek komt voort uit de behoefte van de mens om de wereld waarin hij leeft te begrijpen, in haar onderdelen en vervolgens in hun procesmatige samenhang. De economisch denkende mens zoekt rechtvaardiging om aan deze behoefte, die in wezen voor een deel van culturele aard is, te kunnen werken. Het beschikbaar maken van de resultaten van zijn onderzoek aan een breed pubiek vormt een dergelijk rechtvaardiging: veranderingen in de "paleo" situatie staan model voor de veranderende "actuo" situatie.
Kartering is een goed middel om de ruimtelijke variabiliteit van het landschap en haar ondergrond in beeld te brengen. Door de complexe aard van deze variatie en de uitgestrektheid van het object is er zeer veel tijd en inspanning nodig om voldoende gegevens te verzamelen alvorens er een enigermate consistent en betrouwbaar beeld kan worden aangeboden.

Sinds het werk van de geoloog en landbouwkundige A.C.W. Staring (1808-1877), beschikt Nederland over een rijke traditie in het vervaardigen van landsdekkende inventarisaties van specialistische aard. Sinds de tijd van Staring is door de Rijks Geologische Dienst (1918-1935 en 1968-1996), haar voorgangers: Rijks opsporings dienst van Delfstoffen (1903-1918) en de Geologische Stichting 1935-1968) zeer veel informatie verzameld over de opbouw van de diepere ondergrond. De bodem en de landschappelijke expressie van de ondergrond worden bestudeerd door de Stichting voor Bodemkartering (thans Winand Staring Centrum -DLO). Daarnaast biedt de Hoogtekaart van Nederland de mogelijkheid tot een zeer gedetailleerde analyse van de landvormen van een gebied.
Overal ter wereld is bewoning geconcentreerd in de randgebieden van sedimentaire bekkens. Daarom is kennis van de opbouw van de ondergrond in deze gebieden zinvol. Nederland beschikt zeer waarschijnlijk ter wereld over de meest uitgebreide verzameling van gegevens over de opbouw van niet verharde gesteenten in de ondergrond. Dit betekent dat de beheerders van deze gegevens de mogelijkheden hebben om een hele specifieke expertise te ontwikkelen uit de verzamelde gegevens. De kracht van kaarteringsonderzoeken ligt in het integreren van gegevens op verschillende schaal niveau's. Zowel op een ruimteschaal als op een tijdschaal.

Het thema van dit proefschrift betreft het gedrag van rivieren op verschillende tijdschalen. In het bijzonder wordt ingegaan op de vraag hoe dit gedrag verandert in vergelijking met veranderingen in het klimaat en in de tektoniek.

De hier te bespreken resultaten komen direkt voort uit de geomorfologische en geologische kaartering van Zuid-Oost Nederland. Deze inventarisaties zijn een onderdeel van de landsdekkende kartering op schaal 1:50.000. Dergelijke projecten zijn niet toegesneden om nieuwe inzichten die voortkomen uit de analyse van het geinventariseerde gebied in voldoende mate uit te werken. Deze inzichten zijn echter wel uitgewerkt in een aantal (gedeeltelijk reeds gepubliceerde) artikelen die in dit proefschrift zijn samen gebracht.
Het studiegebied ligt ruwweg tussen Maastricht, Nijmegen en Tilburg. Dit gebied vormt de overgang van een erosie- naar een sedimentatiegebied van drie afvoer systemen: de Maas, de Rijn, en een (gedeeltelijk verdwenen) Belgisch systeem. Dit laatste watert thans af via de Schelde. In tektonische zin vormt het gebied (het Roerdal breuken systeem) een integraal deel van het Noordzee bekken en is de geschiedenis ervan nauw verbonden met die van het Alpiene voorland. Hierdoor past het onderzoek in een ruim kader.

De tijd-ruimte betreft de laatste 10 miljoen jaren; een periode die mondiaal zowel in klimatologische als in tektonische zin, essentiële verandering laat zien (verkoeling en een toenemend reliëf). Ruddiman en Kutzbach (1990) hebben laten zien dat de vormverandering van de aardkorst bepalend is geweest voor de richting waarin het regionale klimaat zich veranderde. Dit geeft aan dat klimaat en tektoniek niet geheel onafhankelijke parameters zijn, de mate van afhankelijkheid is tijdschaal gebonden.
Een belangrijk aspect van deze periode is dat er op allerlei tijdschalen zich cyclische veranderingen in het milieu voordoen. Deze zijn het best bekend van het klimaatsysteem maar lijken zich ook voor te doen in de tektoniek.

In het klimaat zijn zij primair het gevolg van cyclische variaties in de hoeveelheid zonlicht die de aarde ontvangt, secundair van vulkanische aktiviteit en natuurlijke zelf-regulerende mechanismen waarbij de circulatie van het oceaanwater een sleutelrol speelt. Effecten van periodische veranderingen in het spanningsveld van de aardkorst kunnen konvergeren met effecten van klimatologische periodiciteiten op het oceaan volume Cloetingh, (1988). Dit geeft aan dat beide parameters niet zonder meer kunnen worden gescheiden. Voor het scheiden van klimatologische en door de tektoniek bepaalde veranderingen op het milieu is het nodig om te beschikken over onafhankelijk ondersteunend bewijs voor elk van beide. Een probleem hierbij vormt dat voor beide zaken de resolutie in de tijd veelal verschillend is. Dit beinvloedt de scherpte van de bewijsvoering.

Sinds het laatste decennium kan een tijdraamwerk worden opgesteld dat bestaat uit een combinatie van klassieke dateringsmethoden en astronomische methoden. De nauwkeurigheid van sommige klassieke methoden is hierdoor veel beter geworden. Ook de resolutie van het raamwerk als geheel is sterk verbeterd (Shackleton et al., 1990; Hilgen, 1991; Shackleton et al, 1993). Deze astronomische tijdschaal geldt met name voor de periode van de laatste paar miljoenen jaren. Deze nieuwe tijdschaal biedt de mogelijkheden om geologische processen te kwantificeren volgens een lineaire schaal. Hierdoor is het mogelijk om verschijnselen, die het gevolg zijn van een kombinatie van zowel de tektoniek als van het klimaat, nauwkeuriger uiteen te rafelen. Omdat klimaat-bepaalde veranderingen worden gestuurd door astronomische parameters: de Milankovitch cycli.

Deze Milankovitch cycli worden bepaald door drie periodiciteiten: (1) tolbeweging van de aardas (deze duurt 21.000 jaar voor het vlak waarin dag en nacht evenlang duren). (2) De stand van de aardas (deze cyclus duurt 41.000 jaar. (3) De vormverandering van de aardbaan (deze heeft periodiciteiten van 400.000 en 100.000 jaar). Het interferentiepatroon van deze astronomische bewegingen veroorzaakt daarnaast langere cycli met een duur van 200.000, 1.300.000 en 3.500.000 jaar (Berger, 1978).
Het klimaatsysteem ondergaat ook "kort"durende variaties in de tijds-orde van decaden tot enkele duizenden jaren waarvan de oorzaken veel minder duidelijk zijn. De effecten van deze snelle cycli op geologische parameters zijn wellicht van zeer grote betekenis voor de menselijke samenleving. Het bestuderen van deze effecten is dan ook van groot belang.

De invloed van al deze verschillende klimaatveranderingen is duidelijk geregistreerd in dikke, zeer uniforme sedimentpakketten zoals deze worden gevonden op de bodem van meren, oceanen, in ijskappen en in loess pakketten. Het aantal voorbeelden hiervan stijgt ieder jaar waardoor de rol van het klimaat steeds breder wordt geaccepteerd.
Veel minder toegankelijk is de registratie van genoemde periodiciteiten in kontinentale systemen zoals rivier afzettingen. Dit komt vooral door sterk wisselende lithologische samenstelling. De eerste 5 hoofdstukken van dit proefschrift zijn aan dit onderwerp gewijd.

Variaties in temperatuur en het periodiek accumuleren van grote hoeveelheden ijs op de polen en de hooggebergten zijn het belangrijkste gevolg van de invloed van al deze astronomische cycli. Temperatuurverdeling op het noordelijk halfrond beinvloedt de patronen in de atmosferische drukverdeling. Dit beinvloedt de positie van depressiebanen en daarmee de neerslagverdeling. Omdat in de periodiciteiten van klimaatcycli een hierarchie is te onderscheiden naar de duur en de intensiteit, is het waarschijnlijk dat in klimaatafhankelijke systemen zoals oceaan volumes en riviersystemen derhalve ook een hierarchie zal zijn te vinden. De mate van direktheid van de koppeling tussen deze en het klimaat, zal bepalen hoe duidelijk deze afgeleide hierarchie zal zijn te herkennen.

De Maas is een regenrivier en haar afvoer regime wordt sterk beinvloedt door de geringe bergings capaciteit van de Ardennen. Dit maakt haar zeer geschikt om klimaatafhankelijkheid te bestuderen. Dit geldt met name voor het gedeelte bovenstrooms van de terrassenkruising (dit is de zone waar het erosiegebied overgaat in het afzettingsgebied). Benedenstrooms hiervan wordt de dynamiek gecompliceerd door veranderingen in het zeeniveau (= de ligging van de erosiebasis). Dit wordt daar vooral opgevangen door aanpassingen in de sinuositeit van de geul; effecten van veranderingen in het zeeniveau dempen uit in de bovenstroomse richting (Schumm, 1993). De vegetatie speelt een rol in de klimaatbepaalde dynamiek van het bovenstroomse deel van de rivier. Omdat het bestudeerde deel van de Maas door een gebied stroomt dat sterk door breukwerking wordt beinvloed, konden eveneens tektonische parameters (horizontale en vertikale bewegingen) worden geanalyseerd.

Volumeveranderingen van ijskappen hebben een direkt gevolg voor het volume van de oceanen. Voor variaties in oceaanvolume is door Haq et al. (1988) een rangorde opgesteld. Deze rangorde kent verschillende frequenties: 9-10 miljoen jaar (Ma), 0,1-0,2 Ma, 0,01-0,02 Ma (resp. tweede tot vijfde orde cycli). De drijvende krachten in deze tijdsbereiken zijn van tektonische en klimatologische aard.
De laatste 10 Ma beslaat één 2^e orde zeespiegel cyclus met zijn maximum in het Onder Plioceen (rond 5 Ma). Dit maximum valt samen met een periode waarin wereldwijd een warm klimaat heerst. Deze periode wordt begrensd door koelere perioden. Dit suggereert dat de tweede orde zeespiegel kurve door het landijs volume (= glacio-eustatisch) bepaald is. Uit dit proefschrift blijkt o.m. dat het begin van deze tweede orde cyclus samenvalt met het begin van een belangrijke tektonische puls. Derde en vierde orde zeespiegel bewegingen zijn eveneens van glacio-eustatische aard. Bij de derde orde bewegingen doet zich een convergentie van oorzaken voor: Lourens, (1994) laat zien dat deze periodiciteit goed correleert met die van de minima in de schommelingen van stand van de aardas (een frequentie van 1,3 Ma). Cloetingh, (1988) daarentegen verklaart deze periodiciteit tektonisch.
In het Noordzeebekken uit zich de derde orde in de positie van de kustlijn: een meer danwel minder bekkenwaardse positie van de kustlijn. Daaromheen schommelt de kustlijn nogmaals tijdens de vierde orde cycli (100.000 jaar cycli).

Binnen een bepaald traject van de rivier heeft de derde orde verandering van de ligging van de kustlijn belangrijke gevolgen voor het type van het afzettingencomplex (= facies) van de rivier. Beneden een terrassenkruising treedt daardoor, in de stapeling van de rivierafzettingen, een afwisseling op van meer proximaal (kustnabij) dan wel meer distaal gevormde afzettingen op. Uitgebreid voorkomende laagland-rivierafzettingen (klei en veen) wisselen vertikaal af met zand en grindrijke series.
Het dateren van rivier afzettingen is in de meeste gevallen (potentieel) slechts mogelijk als er voldoende met klei en veen gevulde restgeulen in voorkomen. Dit type afzetting komt

veelvuldig voor in proximale rivier trajecten maar ontbreekt meestal in distaal gevormde afzettingen. Distaal gevormde facies zijn daarom in het algemeen moeilijk dateerbaar. Maar zoals hierboven aangegeven is deze facies juist van groot belang omdat zij het meest zuivere beeld kan geven van de mate waarin het riviersysteem wordt beinvloed door veranderingen in klimaat en tektoniek. Een oplossing tot dit dateringsprobleem biedt de sequentiestratigrafie. Wij zullen laten zien dat door gebruik te maken van de sequentiestratigrafie de tijds-resolutie sterk verhoogd kan worden.

Sequentiestratigrafie poogt om complexe afzettingen te ontleden in sedimentaire eenheden die begrensd zijn door vlakken met een regionale betekenis. Van groot belang is dat deze vlakken min of meer isochroon zijn. In dat geval vertegenwoordigen zij fundamentele veranderingen in het verloop van de opvullings geschiedenis van het bekken.
De techniek wordt zowel toegepast op begraven als op ontsloten afzettingen. In de literatuur betreft het vooral afzettings complexen die worden gedomineerd door mariene afzettingen. Daarom worden sequenties veelal geïnterpreteerd in termen van zeespiegel bewegingen, maar dit is geen voorwaarde. Meestal wordt gebruikgemaakt van de frequentieanalyse van signalen die een variatie in de gammastraling weergeven, een techniek die hier niet zal worden besproken.

In het eerste deel van het proefschrift gebruiken wij vertikale lithologische afwisselingen, al dan niet gekoppeld aan geomorfologische gegevens, als basis voor de sequentiestratigrafie. Daarnaast worden gegevens uit boringen gekoppeld aan die van ontsluitingen. Sequentiestratigrafie van rivierafzettingen heeft in de literatuur tot nu toe zeer weinig aandacht gekregen. Dit kan worden verklaard door het meestal ontbreken van een combinatie van factoren: In ontsluitingen zijn de afzettingsstructuren goed te bestuderen maar ontbreken daterings mogelijkheden veelal. Daarnaast heeft vergelijkend onderzoek tussen het voorkomen van ruimtelijke variabiliteit in rivier afzettingen en het optreden van veranderingen in potentieel sturende mechanismen, weinig aandacht gekregen door het ontbreken van referentie kaders die goed in de tijd kunnen worden ingepast.

In het tweede deel van dit proefschrift wordt de neotektonische ontwikkeling van het Roerdal rift systeem in detail geanalyseerd door o.a. gebruik te maken van de rivier-sequentiestratigrafie.

In Nederland bestaat geen traditie om de relatief recente geologische geschiedenis (het Neogeen) te analyseren in termen van tektoniek (neo-tektoniek), in enkele gevallen wordt dit aspect slechts zijdelings genoemd. Waarschijnlijk omdat tektonische processen vooral geassociëerd worden met spectaculaire landvormen is dit aspect in de meeste studies onderbelicht gebleven of leefde het concept, dat aktieve tektoniek niet bestond rond het Noordzee bekken en dat de waargenomen daling slechts het na-ijl effect was van grote aktieve tektoniek uit het Mesozoicum. Gebleken is echter in andere studies dat de daling van delen van het Noordzee bekken veel te sterk is om zo te verklaren. Randgebieden langs het Noordzee bekken vertonen een sterke opheffing. Hoewel het verklarende mechanisme steeds complexer lijkt te worden, is duidelijk dat tektoniek wel degelijk een aktieve rol speelt. De tijdspanne die in dit proefschrift wordt beschreven blijkt zelfs een periode te zijn met extreem grote aktiviteit. In dit geval mogen wij dus stellen dat de "aktuo" situatie model kan staan voor het conceptuele denken over de snelheden in de "paleo" situatie.

Samenvatting

Het belangrijkste doel van dit onderzoek was om inzicht te krijgen in hoeverre veranderingen, in het verloop van de tijd, in rivierdynamiek en rivier-sedimentatiepatronen gerelateerd kunnen worden aan de dynamiek van klimaat en tektoniek. De afzettingen van Maas, Rijn en de Belgische rivieren in Zuid Oost Nederland (waar zowel tektonische daling als -opheffing aktief zijn) bieden hiertoe goede mogelijkheden omdat hier een 10 miljoen jaar lange registratie van rivieraktiviteit bewaard is gebleven. De sedimenten zijn voor een deel bewaard gebleven in een stapeling van begraven sequenties van rivierafzettingen en gedeeltelijk als rivierterrassen.
Essentieel voor de voortgang van het onderzoek was dat in dit gebied gedurende de laatste tientallen jaren door anderen zeer veel waardevolle gegevens verzameld waren; veel daarvan ligt opgeslagen in rapportvorm of als boorbeschrijvingen in de database van de Rijks Geologische Dienst; in de tekst zal er vele malen naar worden gerefereerd. Niet in de laatste plaats is het gebied uitermate geschikt omdat er reeds een klimaatstratigrafie in de vorm van pollenzonering was opgesteld. Bovendien zijn de laatste jaren zeer relevante klimaatgegevens beschikbaar gekomen uit het diepzee- en ijskernonderzoek waardoor een belangrijk referentiekader beschikbaar kwam. Dit kader geldt zowel voor de evaluatie van de pollen-klimaatstratigrafie als voor het tijdskader.
In het algemeen geldt dat rivierafzettingen zich uitermate slecht lenen voor het verkrijgen van (absolute) dateringen. Hierdoor speelt indirekte bewijsvoering (circumstantial evidence) een belangrijke rol, daarnaast ontleent de bewijsvoering ook een deel van haar kracht aan het feit dat de conclusies op allerlei tijdschalen in dezelfde richting wijzen.

Hoofdstuk 1 van dit proefschrift beschrijft het deel van de serie rivierafzettingen dat is afgezet tussen 9.7 Ma en 3.1 Ma. Dit zijn sedimenten die zijn afgezet in een kustvlakte, in een periode voorafgaand aan de eerste Kwartaire glaciaties in een subtropisch tot gematigd klimaat. De sedimenten zijn bekend uit de bruinkoolgroeves in Duitsland en in Nederland uit een groot aantal boringen uit de diepste delen van de Roerdal slenk. Deze twee lokaties bevinden zich ongeveer 60 km uit elkaar in stroomafwaartse richting. Zij blijken een overeenkomstig aantal grof-fijn sequenties te bevatten. De grensvlakken tussen deze sequenties worden verondersteld in horizontale zin zeer uitgebreid te zijn. Beargumenteerd wordt waarom deze grensvlakken beschouwd kunnen worden als extern gestuurde (allocyclische) verschijnselen. De beschikbare ouderdoms controle geeft aan dat de sequentie grenzen in dit tijdsbereik optreden met een gemiddelde cyclusduur rond de 200.000 jaar (variërend tussen de 190.000 en 245.000 jaar). Voorzover gegevens beschikbaar zijn, blijkt uit patroonvergelijking dat het aantal sequentiegrenzen zeer nauw overeen te komen met het aantal wisselingen tussen extremen in de $\delta^{18}O$ concentraties in diepzeekernen. Dit suggereert een gemeenschappelijk factor, dan wel sturend mechanisme. Het is nog niet mogelijk om vast te stellen of variaties in het zeeniveau het riviergedrag beinvloedt dan wel dat uitbreiding van de poolkappen de $\delta^{18}O$ concentratie beinvloedt en eveneens via de positie van het polaire front veranderingen in depressiebanen teweegbrengen die op hun beurt neerslagvariaties veroorzaken.

In **Hoofdstuk 2** wordt het jongere deel van de serie, dat is afgezet gedurende de laatste vier miljoen jaar, beschreven. Dit deel van de sedimenten sluit mooi aan bij de reeks uit het vorige hoofdstuk al is de expressie nu in de vorm van een ander type sedimentaire cyclus nl.

die van rivierterrassen. Wij vonden een, wereldwijd bezien, extreem lange reeks terrassen die bestaat uit 31 niveaus.

In deze periode wordt het klimaat steeds sterker gedomineerd door het optreden van ijstijden; zowel de tektoniek als de frequentie waarmee extreme klimaatvariaties optreden vertonen een versnelling. Door het uitbouwen van de kustlijn verschuift de relatieve positie van het studie gebied meer stroomopwaarts. Het gebied wordt in de loop van de bestudeerde periode aantoonbaar onafhankelijk van zeespiegel veranderingen zonder dat dit een aanwijsbaar effect heeft op de terrasvorming.

Veldwaarnemingen laten zien dat de sedimenten, waaruit de terrassen zijn opgebouwd, afgezet zijn onder koude omstandigheden. Door gebruik te maken van een combinatie van verschillende dateringstechnieken laten wij zien dat het optreden van terrasvorming voor de laatste miljoen jaren verklaard kan worden door het optreden van klimaatswisselingen met een frequentie van ongeveer 100.000 jaar (een vierde orde klimaatcyclus).
Voor het gedeelte van de afzettingen ouder dan 1 miljoen jaar blijkt dat de verdeling over de tijd van het optreden van condities die tot terrasopbouw leiden varieerd sterk. Dit wordt veroorzaakt doordat de periode van de bovengenoemde 100.000 jaar cyclus verschoven is naar 400.000; bovendien is er een signaal met een periode van 40.000 jaar sterk aanwezig. Het interferentie beeld is daardoor veranderd.

Opheffing van het gebied is verantwoordelijk voor de terras vorming. Onder de aanname dat de processen niet wezenlijk anders waren, en er dus een zelfde oorzakelijk verband bestond tussen terrasvorming en klimaatwisseling, blijkt dat een snelheid van opheffing kan worden gereconstrueerd die een elegante extrapolatie laat zien van de periode na 1 miljoen jaar. Echter niet alle belangrijke klimaatswisselingen, in de periode voor 1 miljoen jaar, uiten zich in terrassen. Alleen tijdens de koudste perioden wordt voldoende sediment in de riviervlakte opgeslagen zodat de kans bestaat dat hier restanten van over blijven als de rivier gedurende de volgende perioden door laterale erosie zijn oevers weer verwijdt. In de specifieke situatie van Zuid Limburg werd de preservatie bevorderd door een zeer geringe kanteling van het gebied waardoor m.n. vooral de zuidoostoevers bewaard konden blijven.

In **Hoofdstuk 3** worden de aannamen getoetst die moesten worden gedaan in hoofdstuk 2 bij de inpassing in de tijd van de terrassen. Het gekombineerde effect van klimaat en tektoniek op de terrasvorming en grootschalige dalontwikkeling voor de laatste 2 miljoen jaar wordt gesimuleerd met een 3-dimensionaal model MATER. Het blijkt dat de gedane aannamen, samen met algemeen geldende hydrologische regels, voldoende consistent zijn om de terrassenserie als beschreven in hoofdstuk 2, te simuleren. De tektoniek speelt een beslissende rol in de preservatie.

Om echter een beter inzicht te krijgen in het verloop van de opbouw van een terras gedurende een vierde orde klimaatscyclus gaan we in **Hoofdstuk 4** nader in op het gedrag van de rivier gedurende een zo'n glaciaal-interglaciaal cyclus. Hiervoor gebruiken wij het tijdsinterval van de laatste 130.000 jaar omdat hierover, vanuit verschillende bronnen, een gedetailleerd beeld van de klimaatgeschiedenis bestaat. Daarnaast beschikken wij over een gedetailleerd ruimtelijk beeld van de interne opbouw van de riviersequentie die in deze tijd is gevormd. Dit beeld is gebaseerd op boringen. Stadiaal/interstadiaal klimaatwisselingen binnen een klimaatcyclus (vijfde orde cycli) vertonen een zelfde deterministische relatie met riviergedrag als wij in de vorige hoofdstukken vonden voor riviergedrag gedurende een reeks van glaciaal/interglaciaal cycli (vierde orde).

Er lijkt dus sprake te zijn van een hiërarchisch "genest" principe dat het rivier gedrag bepaalt. Ook in dit hoofdstuk maken wij gebruik van zowel de veld-benadering als de model-benadering. De laatste toont heel duidelijk hoe de rol van het regionale klimaat verweven is met het supra-regionale klimaat. Door de nabijheid van ijskappen is er periodiek veel bewegend dekzand dat een extra sedimentlast aan delen van de beneden Maas toevoegt. Hierdoor bouwt de rivier veel sneller en een dikker pakket op in deze perioden dan verwacht kan worden onder "normale" koude omstandigheden. Op de tijdschalen van enkele tienduizenden jaren overstemt de rol van het klimaat die van de tektoniek .
Gedurende het laatste stadiaal wordt veel dekzand vanuit de Roerdal slenk in de rivier geblazen. Dit gebeurt m.n. in een deel van de rivier waar zijn verhanglijn een opvallende knik vertoont. Deze twee factoren leiden ertoe dat over een zeer brede overstromingsvlakte een sedimentwaaier wordt opgebouwd gedurende dit stadiaal. Deze vlakte is plaatselijk zo'n tien meter dik en is opgebouwd uit een vrij uniform sediment.

In **Hoofdstuk 5** gebruiken wij ontsluitingen in deze sedimentwaaier om na te gaan of er binnen een stadiaal/ interstadiaal cyclus nog meer detail kan worden aangebracht in de riviersedimentatie. Doordat gedurende het stadiaal extreem veel sediment is opgehoopt, biedt dit de mogelijkheid een groot aantal gebeurtenissen te onderscheiden niet alleen die welke plaatsvonden gedurende de opbouwfase, maar ook die uit de daaropvolgende insnijdingsfase. In dit hoofdstuk worden sedimentologische en geomorfologische opeenvolgingen vergeleken met klimaatcycli die een duur hebben van duizend tot twee duizend jaar (zesde orde cycli). Hoewel het niet mogelijk is aan iedere onderscheiden eenheid een specifieke tijdspanne toe te kennen laat het aantal opeenvolgingen zien dat dergelijke kortdurende klimaatscycli ook door de rivier kunnen worden geregistreerd. Het eerder genoemde hiërarchische principe kan dus worden door getrokken tot op het niveau van deze kortdurende cycli. Eveneens blijkt dat de relatie niet altijd opgaat omdat de vegetatie een belangrijke intermediaire rol kan spelen tussen rivierdynamiek en klimaatsdynamiek. Tijdens de overgang van koud naar warm kan met name aan het begin enige vertraging optreden veroorzaakt door de migratietraagheid van de boomvegetatie. Tijdens de overgang van warm naar koud neemt allereerst de hoeveelheid afvoerwater toe waardoor het systeem wel insnijdt en nog niet direkt tot aggradatie overgaat omdat de bosvegetatie slechts geleidelijk degradeert.
In de vorige hoofdstukken is aangetoond dat het voorkomen van riviersequenties gerelateerd kan worden aan het optreden van klimaatwisselingen. Dit houdt in dat door nauwkeurige analyse van de riviersequenties het mogelijk is om hoge tijdresolutie in de stratigrafie te bereiken. Dit hebben wij in het navolgende toegepast.
Vergelijking tussen het breukenpatroon en het patroon van rivierbochten laat zien dat een uitermate gering verzet langs breuken sterk sturend werkt op het rivierpatroon. Door nu riviersequentieanalyse en of rivierpatroonanalyse te kombineren met bekende breukpatronen kunnen we een beter inzicht krijgen in het verloop van de factor tektoniek.

In **Hoofdstuk 6** wordt de neotektoniek van het Roerdal riftsysteem m.b.v. hierboven geintroduceerde methoden in detail geanalyseerd. Allereerst worden sedimentaire cycli gebruikt om de opheffingsgeschiedenis van Zuid Limburg te reconstrueren. Gemeten in tijdstappen van 100.000 jaar laat de opheffing een beeld zien van vertraagde en versnelde bewegingsfasen. Deze verschillende episoden kunnen worden gekoppeld aan belangrijke geomorfologische gebeurtenissen in de rest van Europa. Deze lijken gestuurd door veranderingen in de bewegingsrichting van de platen in de lithosfeer. Vervolgens wordt aan de hand van de sedimentaire opvulling van de Roerdal slenk getoond dat horizontale bewegingen langs de begrenzende breuken een belangrijke rol spelen.

In **Hoofdstuk 7** wordt ten slotte de geomorfologie van het Roerdal rift systeem geïntegreerd met een lange tijdreeks van geodetische waarnemingen van vertikale bewegingen. Een netwerk van lineamenten (= lijnvormige elementen die door breuken zijn bepaald) op kilometerschaal geeft aan dat de bovenste korst kan worden opgedeeld in wigvormige blokken op 10 km schaal, die elk een eigen bewegingspatroon vertonen. Deze beweging wordt het best verklaard vanuit een rechtslaterale horizontale bewegingskomponent langs de hoofdbreuken. Dit is consistent met de resultaten uit het vorige hoofdstuk.

Dankwoord

Prof. Dr. Jan D. de Jong heeft, door zijn uitzonderlijk didactische vermogen, bij mij de belangstelling voor het vak gewekt. Hij heeft mij het belang duidelijk gemaakt om allerlei aspecten van de geologie te beschouwen en die tot een synthese brengen. In de persoonlijke sfeer hadden wij een hele goede band, daarom draag ik graag dit proefschrift ook aan hem op.

De resultaten van de geomorfologische kartering van de Maasterrassen in Zuid-Limburg waren de aanleiding tot het schrijven van dit proefschrift. Aan dat karteringsproject heb ik in team verband gewerkt met Karin Koelbloed, Bert de Lange, Dick Brus en Henk Wolfert; wij vormden een enthousiaste ploeg. Het resultaat was echter nooit bereikt als Werner Felder zich niet zo bijzonder coöperatief had opgesteld. Dit ondanks de terughoudende opstelling van Jan Bisschops, indertijd zijn formele chef. Het is een groot voorrecht geweest Werner, om met jou en jouw team: Peter Bosch, Paul Kisters en Erik Wijnen over een reeks van jaren met wisselende intensiteit te hebben kunnen samenwerken.

Mijn promotor Salle Kroonenberg komt de eer toe het vooralsnog unieke van de Maasterrassen te hebben ingezien. Je enthousiasme was zeer stimulerend en door jouw initiatief is het tot samenwerking met Tom Veldkamp gekomen waardoor de modelleringsaspecten in dit boekje konden worden opgenomen. Salle, ik waardeer het geduld dat jij, met jou stilistische gaven, hebt opgebracht bij het corrigeren van mijn teksten. Het is mij niet gelukt om er aantrekkelijk proza van te maken. Toen wij het promotieproject bespraken heb jij de belangrijke vraag gesteld of mijn huwelijk wel bestand zou zijn tegen de druk die het schrijven van een proefschrift met zich meebrengt. Ik heb dit criterium steeds als de belangrijkste toets onthouden. Eugénie en de kinderen hebben veel willen en kunnen opvangen.

Sierd Cloetingh heeft mij geweldig gestimuleerd om de morfo-tectonische aspecten verder uit te werken. Sierd, door jouw toedoen heeft mijn werk een extra dimensie gekregen.

Het gehele onderzoek is steeds nauw verweven geweest met aspecten van mijn dagelijks werk. Zowel bij de Rijks Geologische Dienst als bij het Staring Centrum-DLO heb ik steeds de vrijheid genoten om de balans te zoeken tussen dit onderzoek en ander werk. Erno Oele, Dirk Beets en Jan Klijn ben ik daar zeer erkentelijk voor. Ook de discussies met hen waren een stimulans om door te gaan. Later hebben Dick van Doorn en Ruud Schüttenhelm deze grondhouding voortgezet. Bij Ruud ging het aanvankelijk niet van harte. Ik vermoed dat de positieve instelling t.a.v. dit werk, van Aleid Bosch en Wim Westerhof zeker geholpen heeft om Ruud over de streep te trekken.

Voor de inhoudelijke kant van het werk heb ik veel profijt gehad van discussies met velen. Dat gebeurde zowel in het veld, bij het verzamelen van monsters en het interpreteren van de sedimentaire structuren, als achter de schrijf- of borreltafel als het over het conceptuele model ging. Zonder de anderen te kort te willen doen wil ik enkelen in het bijzonder noemen. Met Wim Groenewoud heb ik tijdens het schaven aan teksten heel veel gelachen; met Tom Veldkamp kon ik altijd een diepgaande discussie opzetten, zo ook met Waldo Zagwijn. Dirk Beets was steeds zeer stimulerend maar moeilijk te overtuigen, het was een uitdaging om jou op zijn minst aan het twijfelen te kunnen brengen. Met Jacques Schwan heb ik veel up's en enkele down's beleefd bij het schrijven van ons gezamenlijk stuk. Ik heb goede herinneringen aan de tijd dat ik samen met Wim Westerhof aan de interpretatie van boorbeschrijvingen van kaartblad 52 O werkte, het was voor mij de basis voor het denken over de sequentiestratigrafie.

Verder wil ik graag noemen: Ton van Hoof en Cor Langereis die het paleomagnetisch werk

deden; Dick Brus die de geo-statistiek deed: leuk dat onze wegen elkaar opnieuw kruisten; Gerard Ruegg, Richard Boersma, Poppe de Boer, Kees Kasse en Erik Wijnen voor hun kijk op de sedimentologie; Piet Cleveringa, Wim Hoek en Karin Koelbloed als het over pollen ging. Hilda Goslings-Nieuw Beerta, Carolien Slomp en Pieter Lubbers hebben als student in deelprojecten meegewerkt.
Tot slot zijn bij de afwerking van de artikelen die de ruggegraat van het boekje vormen, weer vele anderen betrokken geweest. Tekstfiguren en foto's vormen een wezenlijk onderdeel van het geheel. Ik heb steeds een beroep kunnen doen op de deskundigheid op dit gebied zowel bij de tekenkamer van de RGD onder leiding van Jan de Blaauw, als die van het Staring Centrum-DLO onder leiding van Ruud de Jonge en Martin Jansen. Roel van der Kraan, Henk van Ledden, Diana Moeliker, Dasja ten Cate, Rini Schuiling, John van Delft en Cor van der Schouw, ik heb jullie inzet en nauwkeurigheid zeer op prijs gesteld en ik ben trots op het resultaat van jullie vormgeving.

Curriculum Vitae

De schrijver dezes werd geboren op 10 maart 1949 te Rotterdam. In 1968 behaalde hij het diploma HBS-B aan het Libanon Lyceum aldaar. Waarna hij in Wageningen aan de Landbouw Hogeschool ging studeren. Na een onderbreking voor het vervullen van de militaire dienst plicht, behaalde hij in 1975 zijn kandidaats examen en vervolgens in 1977 het ingenieursdiploma in de Cultuurtechniek met de bijvakken Geologie en Natuurbeheer.
Na het afstuderen kreeg hij een baan als leraar in de Bodemkunde en Geologie aan de Hogere Bosbouw en Cultuurtechnische School in Velp. Na drie jaren werd deze baan gecombineerd met die van tijdelijk waarnemend districtsgeoloog bij de Rijks Geologische Dienst in district Oost. Na het aflopen van deze periode in 1983, werd hij namens de Rijks Geologische Dienst als geoloog/ geomorfoloog gedetacheerd bij de toenmalige Stichting voor Bodemkartering, thans onderdeel van het Staring Centrum -DLO te Wageningen. In deze functie werkte hij aan de kaartbladproductie van de Geomorfologisch kaart van Nederland, schaal 1:50.000. Toen deze serie, als gevolg van het zogenaamde "concentreren op kerntaken", door beide diensten voortijdig werd afgebroken in 1992, kwam er tijdruimte voor meer onderzoeksmatige projecten. Naast onderzoek, werkte hij mee aan de geologische kartering van enkele bladen in zuid-oost Nederland en was betrokken bij het Universitaire onderwijs. Dit laatse in de vorm gastcolleges in de Geologie van Nederland aan de Technische Universiteit van Delft en via het schrijven van leerstofeenheden voor de Open Universiteit in Heerlen.
Sinds 1990 is, naast aan bovengenoemde taken, gewerkt aan bouwstenen voor dit proefschrift.