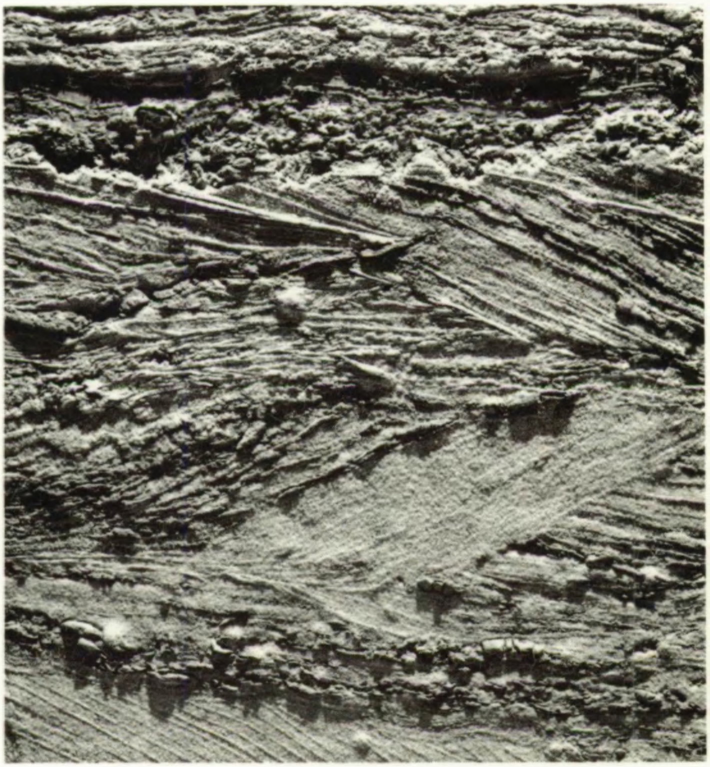
EARLY-PLEISTOCENE TIDAL AND FLUVIATILE ENVIRONMENTS IN THE SOUTHERN NETHERLANDS AND NORTHERN BELGIUM



Kees Kasse

STELLINGEN

1. Het feit dat de Formatie van Tegelen, die in westelijk Noord-Brabant en Noord België in een zoet tot brak, inshore getijden-milieu werd gevormd, vaak als een fluviatiele afzetting is geïnterpreteerd, is ten dele te verklaren door de afwezigheid van fauna in de sedimenten.
   Van Dorsser, H.J. (1956): Het landschap van westelijk Noordbrabant. Thesis, Utrecht, 133 pp.
   Geys, J.F. (1975): De sedimentologie en de morfogenetische betekenis van de Oudpleistocene afzettingen in de Antwerpse Noorderkempen. Thesis, Gent, 230 pp.
   Dit proefschrift.

2. De door Zagwijn aangetoonde verkoeling van het klimaat tijdens het Tiglien C4 blijkt de intensiteit van een glaciale periode te hebben.
   Zagwijn, W.H. (1963): Pollen-analytic investigations in the Tiglian of the Netherlands. Meded. Geol. Stichting, N.S. 16, p. 49-69.
   Dit proefschrift.

3. De hoge Chenopodiaceae-waarden in de Formatie van Tegelen in België duiden op de aanwezigheid van aanspoelselzones en wijzen niet op het voorkomen van zoute kwelders.
   Dricot, E.M. (1961): Microstratigraphie des Argiles de Campine. Bull. Soc. belge Géol., Paléont., Hydrol. 70, p. 113-141.
   De Ploey, J. (1961): Morfologie en Kwartair-stratigrafie van de Antwerpse Noorderkempen. Acta geographica Lovaniensia 1, 130 pp.
   Dit proefschrift.

4. De omslag van de magnetische polariteit in de top van de Formatie van Tegelen in België dient toegeschreven te worden aan cryoturbatie en liquefactie tijdens het Weichselien.
   Van Montfrans, H.M. (1971): Paleomagnetic dating in the North Sea basin. Thesis, Amsterdam, 113 pp.
   Dit proefschrift.

5. De belangrijke sedimentaanvoer uit Midden België tijdens het Vroeg-Pleistoceen is in paleogeografische reconstructies vaak onderbelicht.
   Zagwijn, W.H. (1974): The palaeogeographic evolution of the Netherlands during the Quaternary. Geol. Mijnbouw 53, p. 369-385.
   Dit proefschrift.

6. De Vroeg-Pleistocene glaciale perioden zijn wat betreft hun intensiteit en duur vergelijkbaar met de Laat-Pleistocene.
   Zagwijn, W.H. (1975): Variations in climate as shown by pollen analysis, especially in the Lower Pleistocene of Europe. In: Wright, A.E. and F. Moseley (eds.): Ice Ages: ancient and modern. Geol. Journ., Spec. Issue 6, p. 137-152.
   Vandenberghe J. and C. Kasse (1988): Periglacial environments during the Early Pleistocene in the southern Netherlands and northern Belgium. Palaeogeogr., Palaeoclimatol., Palaeoecol.
   Dit proefschrift

7. De genetische classificatie van het Nederlandse Kwartair geeft problemen bij de indeling van estuariene en andere kustnabije afzettingen.
   Zagwijn, W.H. en C.J. Van Staalduinen (1975): Toelichting bij geologische overzichtskaarten van Nederland. R.G.D., Haarlem, 134 pp.

8. De aanwezigheid van walsedimenten rond pingoruïnes in Nederland is twijfelachtig.
   De Gans, W. (1981): The Drentsche Aa valley system. Thesis, Amsterdam, 132 pp.
   De Gans, W., P. Cleveringa en G.P. Gonggrijp (1984): Een ontsluiting in de wal van een pingoruïne nabij Papenvoort (Dr.). RIN-rapport 84/6, 53 pp.

9. De afwisseling van organische en klastische (eolische) eenheden in periglaciale sequenties is niet noodzakelijkerwijs de weerspiegeling van respectievelijk warmere en koelere fasen tijdens de afzetting.
   Van der Hammen, T., G.C. Maarleveld, J.C. Vogel and W.H. Zagwijn (1967): Stratigraphy, climatic succession and radiocarbon dating of the last glacial in the Netherlands. Geol. Mijnbouw 46, p. 79-95.
   Bishop, W.W. and G.R. Coope (1977): Stratigraphical and faunal evidence for Lateglacial and Early Flandrian environments in South-West Scotland. In: Gray, J.M. and J.J. Lowe (eds.): Studies in the Scottish Lateglacial environment. Pergamon, Oxford, p. 61-88.
   Kocurek, G. and J. Nielson (1986): Conditions favourable for the formation of warm-climate aeolian sand sheets. Sedimentology 33, p. 795-816.

10. Helder taalgebruik in een wetenschappelijk artikel leidt tot meer kritiek op de inhoud.

11. Het indelen van landen in ontwikkelde en onderontwikkelde is een weerspiegeling van het huidige technische wereldbeeld.

12. Bij het oplossen van de verkeersproblemen zou men West-Nederland moeten benaderen als één grote stad, in plaats van afzonderlijke "grote" steden.

13. De uitbreiding van het Nederlandse autowegennet doet de marterachtigen de das om.

14. De Nederlandse boer heeft het landschap altijd vernield; leidde dat vroeger tot diversiteit, tegenwoordig tot uniformiteit.

15. De toename van het aantal plantensoorten langs de Nederlandse autowegen zou door de Minister van Verkeer en Waterstaat als argument gebruikt kunnen worden voor de aanleg van meer autowegen.

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ACADEMISCH PROEFSCHRIFT

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Cornelis Kasse

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CONTENTS

SUMMARY

The study area is situated in the Dutch-Belgian border region at the southern rim of the Quaternary North Sea basin. During the Early-Pleistocene perimarine deposition occurred at the mouths of the Rhine, Meuse and Scheldt. Because of the slight uplift since the Middle-Pleistocene, the Early-Pleistocene tidal and fluviatile deposits are found close to the surface. In this study the perimarine sedimentary environments and the paleogeomorphological evolution during the Early-Pleistocene have been reconstructed, in relation to glacial and inter-glacial climatic conditions.

In chapter 2 the lithostratigraphic sequence has been described in detail, in order to enable spatial reconstructions of the sedimentary environments. The lithostratigraphic units in The Netherlands and Belgium are correlated on lithological and sediment-petrographical grounds. The Rijkevorsel, Beerse and Turnhout Members of the Campine Clay and Sand Formation in Belgium are equivalent to the Tegelen Formation in The Netherlands.

In chapter 3 the depositional environments are reconstructed by the analysis of the sedimentary structures, texture and paleobotanical content of the sediments.

In chapter 4 the chronostratigraphic position of the units is derived from paleobotanical analysis and paleomagnetism.

In chapter 5 the provenance of the sediments is established by the heavy mineral and gravel composition.

In chapter 6 the paleogeographical evolution of the study area is given during glacial and interglacial Early-Pleistocene periods. Two major breaks are recorded in the sedimentary sequence. The first break occurred at the transition of the warm temperate Tiglian to the cold Eburonian, when tide dominated environments were replaced by fluvial environments. During the Tiglian C3 and Tiglian C5 interglacials inshore, micro- to mesotidal deposits were formed. Clayey sediments were deposited in the eastern part of the area (Rijkevorsel and Turnhout Members) in distal (landward), fresh to brackish tidal environments. Tidal litter zones dominated by Chenopodiaceae species, developed at the mean high water level along the southern fringe of the tidal environment. In the western part generally sandy sediments (Hoogerheide and Woensdrecht Members) were deposited in proximal (seaward), brackish tidal environments, in which wave processes besides ebb and flood currents were important in the upbuilding of the sedimentary sequence. The tidal range, concluded from the vertical distribution of certain sedimentary structures, amounted to approximately 1 to 2 meters. The sediments in the east were originally supplied by the Rhine and redistributed by tidal currents. In the west an important admixture was established with sediment from the Scheldt basin. Between the warm temperate Tiglian C3 and Tiglian C5 a major cooling of the climate took place during the Tiglian C4 (Beerse Glacial). A temporarily eustatic sea-level fall occurred and eolian sand-sheets and fluvial deposits (Beerse Member) from the Scheldt basin were formed on the distal tidal deposits of the Rijkevorsel Member. The type and intensity of the periglacial structures in the Beerse Member point to a mean annual temperature of maximal -5°C. They reveal a more or less similar temperature regime in comparison to the younger Early-, Middle- and Late-Pleistocene glacials.

During the warm temperate Tiglian C5 an important sea-level rise took place and the previously deposited Tiglian C4 glacial sediments were severely eroded. Therefore, the interglacial Tiglian C5 deposits

(Turnhout Member) are often found directly upon deposits of Tiglian C3 age (Rijkevorsel Member). Only at the most southern margins of the tidal sedimentary basin have the periglacial sediments been preserved against erosion. The lower part of the Turnhout Member was probably deposited during the Olduvai magnetozone (1.66-1.87 m.y.). The normal polarities at the top of the Turnhout Member were caused by remagnetization of the magnetic components in the clay in periglacial conditions (liquefaction) during the Weichselian. After the Tiglian the sea finally withdrew from the Dutch-Belgian border area.

During the Eburonian and Waalian sediments were laid down by rivers from the Scheldt basin (Gilze Member). These eroded the Tertiary deposits in Central Belgium and supplied the major part of the sediment. No definite proofs could be found for the presence of the Meuse in the study area during these periods. The Gilze Member (Kedichem Formation) is found predominantly in The Netherlands. In Belgium the Gilze Member is only locally present (Weelde, Ravels) because of more intense erosion during the Middle- and Late-Pleistocene. The succession of glacial (Eburonian) and interglacial (Waalian) climatic conditions is reflected by changes in the river system and associated sedimentary sequence. Widespread, fine-grained sediments (Appelenberg Sands) developed during the Eburonian. Rapidly alternating beds of fine sand, loam and peat were deposited in shallow fluvial channels, by overbank sedimentation and in shallow pools outside the active river courses. In contrast to the Eburonian period, deposition during the Waalian was more restricted to certain areas with concentrated river flow. In up to 10 m deep meandering channels deposition of fine sands occurred. Lateral migration of the channels resulted in fining-upward sequences, which are capped by clay- and peat-beds, deposited in backswamp environments (Gilze Clay).

The second major break in the paleogeographic evolution took place during the Menapian glacial, when low energy (fine-grained) river systems, characteristic of the Eburonian and Waalian, changed into high energy (coarse-grained) river systems (Alphen Sands). This break probably reflects an increased subsidence of the Central Graben, after a period of limited tectonic activity during the Eburonian and Waalian. The Meuse, which apparently was not present in the investigated area during the Tiglian, Eburonian and Waalian, followed a northwestern course during the Menapian over the Campine plateau in Belgium to Noord-Brabant. The so-called southern rivers from the Scheldt basin (Dender, Zenne, Dijle, Gete) flowed to the northeast and coarse-grained channel sediments (Alphen Sands) were deposited by sandy, braided river systems in the confluence area of the Meuse and the "Scheldt". The sediments are characterized by a stable heavy mineral association and a Scheldt gravel association with an admixture of typical Meuse components in the coarse gravel fractions.

At the end of the Menapian the Rhine re-entered the Central Graben from the southeast after its absence in the Eburonian, Waalian and Menapian. During the Bavelian the Rhine eroded a deep (10-20 m) valley at the western side of the Central Graben. It was filled with fine-grained sediments (Bavel Member), which were deposited in channels and meander cut-offs during the climatic optimum of the Bavelian Interglacial. The start of the Jaramillo magnetozone (0.97 m.y.) was established at the base of the infilling. During the Cromerian the tectonic activity of the western boundary faults of the Central Graben increased. Rhine and Meuse migrated laterally over large distances and coarse-grained deposition took place in and slightly westward of the Central Graben by sandy, braided rivers (Sterksel Formation; Woensel heavy mineral zone). The uplift of the study area west of the Central Graben during the

Cromerian period and later on caused the cessation of the fluvial deposition by the Rhine, Meuse and rivers from the Scheldt basin and was followed by severe erosion of the Early-Pleistocene deposits during the Middle- and Late-Pleistocene. The transition from sedimentation to erosion was already apparent during the Menapian and Bavelian. In the Tiglian, Eburonian and Waalian aggradation was still dominant and vertical stacking of the units prevailed. During the Menapian, Bavelian and Cromerian deep erosion and subsequent infilling occurred west of the Central Graben. Therefore, vertical stacking is less evident and sediments of different ages are found next to each other instead of overlying one another.

SAMENVATTING

Het studiegebied is gelegen in het Belgisch-Nederlands grensgebied. Tijdens het Vroeg-Pleistoceen werden hier perimariene sedimenten afgezet in het mondingsgebied van de Rijn, Maas en Schelde, aan de zuidrand van het Noordzeebekken. Ten gevolge van een geringe opheffing vanaf het Midden-Pleistoceen komen de Vroeg-Pleistocene afzettingen hier dicht onder het oppervlak voor. In deze studie is het sedimentaire milieu en de paleogeomorfologische ontwikkeling gereconstrueerd van het gebied tijdens het Vroeg-Pleistoceen, onder de toen heersende klimaatsomstandigheden met glaciale en interglaciale condities.

In hoofdstuk 2 is de lithostratigrafische sequentie gedetailleerd beschreven, ten behoeve van een correcte ruimtelijke opbouw en interpretatie van verschillende afzettingsmilieus. Nederlandse en Belgische eenheden zijn gecorreleerd op grond van hun lithologische en sedimentpetrografische eigenschappen. De Rijkevorsel, Beerse en Turnhout Members, die in België de Kempische Klei en Zand Formatie vormen, zijn gecorreleerd met de Formatie van Tegelen in Nederland.

In hoofdstuk 3 is het afzettingsmilieu van de verschillende lithostratigrafische eenheden bepaald, door middel van de analyse van de sedimentaire structuren, de textuur en de paleobotanische inhoud.

In hoofdstuk 4 is de chronostratigrafische positie van de eenheden vastgesteld, met behulp van paleobotanische analyses en het paleomagnetisme in de afzettingen.

In hoofdstuk 5 is de herkomst van het sediment vastgesteld met behulp van de analyse van zware mineralen en grind.

Hoofdstuk 6 behandelt de paleogeografische ontwikkeling van het gebied tijdens glaciale en interglaciale Vroeg-Pleistocene perioden. Er zijn twee belangrijke omslagen geregistreerd in de sedimentaire sequentie. De eerste vond plaats bij de overgang van het gematigd warme Tiglien naar het koude Eburonien, toen getijden gedomineerde afzettingsmilieus werden opgevolgd door fluviatiele milieus. Tijdens twee interglacialen van het Tiglien (Tiglien C3 en Tiglien C5) werden in het oosten kleiige sedimenten (Rijkevorsel en Turnhout Member) afgezet in distale (landwaarts gelegen), zoete tot brakke getijden milieus. Aanspoelselgordels, gekenmerkt door de aanwezigheid van Chenopodiaceae soorten, ontwikkelden zich omstreeks het gemiddeld hoog water niveau, aan de zuidelijke rand van het getijden milieu. In het westen werden meer zandige sedimenten (Hoogerheide en Woensdrecht Member) gevormd in proximale (zeewaarts gelegen), brakke getijden milieus, waarin naast de eb- en vloedbeweging ook golfwerking een belangrijke factor was in de verticale opbouw van de sedimentaire sequentie. Het getijdenverschil, afgeleid uit het vertikale bereik van kenmerkende sedimentaire structuren, bedroeg ongeveer 1 tot 2 meter. De sedimenten in het oosten, die oorspronkelijk door de Rijn werden aangevoerd, werden door de getijdenstromingen geredistribueerd. In het westen werd een belangrijke bijmenging geconstateerd met materiaal uit het Scheldebekken.

Tussen het gematigd warme Tiglien C3 en Tiglien C5 trad een sterke verkoeling van het klimaat op tijdens het Tiglien C4 (Beerse Glaciaal). Er vond een tijdelijke eustatische zeespiegeldaling plaats, waarbij wind- en rivierafzettingen (Beerse Member), afkomstig uit het Scheldebekken, werden gevormd op de distale getijdenafzettingen van de Rijkevorsel Member. De aard en de intensiteit van de periglaciale structuren in de Beerse Member wijzen op een gemiddelde jaartemperatuur van -5°C. Ze tonen aan dat de temperatuur van het Tiglien C4 glaciaal vergelijkbaar was met latere Vroeg-, Midden- en Laat-Pleistocene glacialen.

Tijdens het gematigd warme Tiglien C5 vond een sterke zeespiegelstijging plaats, waarbij de glaciale Tiglien C4 sedimenten grotendeels werden geërodeerd, zodat de afzettingen uit het interglaciale Tiglien C5 (Turnhout Member) direct worden aangetroffen op afzettingen uit het Tiglien C3 (Rijkevorsel Member). Alleen in de zuidelijke randzone van het sedimentatiebekken zijn de periglaciale afzettingen voor erosie gespaard gebleven. De Olduvai magnetozone (1.66-1.87 m.j.) werd mogelijk aangetroffen aan de basis van de Turnhout Member. De normale polariteit aan de top van de Turnhout Member is een gevolg van remagnetisatie tijdens periglaciale omstandigheden in het Weichselien.
Na het Tiglien trok de zee zich definitief terug uit het Belgisch-Nederlands grensgebied. Tijdens het Eburonien en Waalien werden sedimenten (Gilze Member) in het gebied neergelegd door rivieren uit Midden België (Scheldebekken), die de daar aanwezige Tertiaire afzettingen erodeerden. De aanwezigheid van de Maas in het gebied kon voor deze perioden niet worden vastgesteld. De Gilze Member, behorend tot de Formatie van Kedichem, kent zijn grootste verbreiding in Nederland en wordt in België nog slechts lokaal aangetroffen (omgeving Weelde, Ravels) ten gevolge van latere erosie. De opeenvolging van glaciale (Eburonien) en interglaciale (Waalien) klimaatsomstandigheden is weerspiegeld in de sedimentaire sequentie. Tijdens het Eburonien werden over een groot gebied fijnkorrelige sedimenten afgezet, bestaande uit dunne, elkaar snel afwisselende lagen van fijn zand, leem en veen (Appelenberg Zanden). Het afzettingsmilieu bestond uit een riviervlakte met ondiepe geulen, overslagsedimenten en ondiepe meren buiten het bereik van de actieve rivierlopen. In tegenstelling tot het Eburonien was de sedimentatie tijdens het Waalien tot enkele gebieden beperkt ten gevolge van een meer geconcentreerde waterafvoer. In meanderende tot 10 m diepe geulsystemen werden fijne zanden afgezet, die ten gevolge van laterale migratie van de geulen werden bedekt door klei- en veenlagen, afgezet in rivierkommen en moerasbossen (Gilze Klei).
De tweede belangrijke omslag in de paleogeografische ontwikkeling vond plaats gedurende het Menapien glaciaal toen energiearme (fijnkorrelige) riviersystemen, kenmerkend voor het Eburonien en Waalien, veranderden in energierijke (grofkorrelige) riviersystemen (Alphen Zanden). Deze omslag was mogelijk het gevolg van een toenemende tektonische activiteit in de Centrale Slenk, die volgde op een periode van geringe daling tijdens het Eburonien en Waalien. De Maas, die gedurende het Tiglien, Eburonien en Waalien niet in het gebied aangetoond kon worden, stroomde in het Menapien in noordwestelijke richting over het Kempisch plateau in België naar Noord-Brabant. De rivieren van het Scheldebekken (Dender, Zenne, Dijle, Gete) stroomden in noordoostelijke richting en in het confluentiegebied van de Maas en de "Schelde" werden grofkorrelige geulsedimenten (Alphen Zanden) afgezet door zandige, vlechtende rivieren. De sedimenten zijn gekenmerkt door een stabiele zware mineralen associatie en een Schelde grind associatie, met een karakteristieke bijmenging van Maascomponenten in het grovere grind. Tegen het einde van het Menapien drong ook de Rijn de Centrale Slenk weer binnen vanuit het zuidoosten (Spruitenstroom Klei), na een periode van afwezigheid in het Eburonien, Waalien en Menapien. Tijdens het Bavelien erodeerde de Rijn een diep (10-20 m) dal aan de westzijde van de Centrale Slenk. Het werd opgevuld tijdens het klimaatsoptimum van het Bavel Interglaciaal met fijnkorrelige sedimenten (Bavel Member), afgezet in geulen en afgesneden meanders. De aanvang van de Jaramillo magnetozone (0.97 m.j.) werd aangetoond in de basis van de opvulling.
Tijdens het Cromerien nam de tektonische activiteit toe langs de westelijke breuken van de Centrale Slenk. De Rijn en de Maas vertoonden grote laterale verplaatsingen en er werden grofkorrelige sedimenten

afgezet in en iets ten westen van de Centrale Slenk door zandige, vlechtende rivieren (Sterksel Formatie, zware mineraal zone van Woensel).
Ten gevolge van de opheffing van het gebied ten westen van de Centrale Slenk tijdens en na het Cromerien, stopte de sedimentatie door de Rijn, Maas en "Schelde", waarna in het Midden- en Laat-Pleistoceen de Vroeg-Pleistocene afzettingen sterk werden geërodeerd. De overgang van sedimentatie naar erosie manifesteerde zich al tijdens het Menapien en Bavelien. In het Tiglien, Eburonien en Waalien trad nog voornamelijk vertikale accumulatie op. Tijdens het Menapien, Bavelien en Cromerien werd het gebied ten westen van de Centrale Slenk gekenmerkt door diepe fluviatiele insnijdingen, gevolgd door opvulling, waardoor sedimenten van verschillende ouderdom naast elkaar voorkomen in plaats van boven elkaar.

1. INTRODUCTION

1.1 Aims and framework

The general aim of this study is to reconstruct the sedimentary environments of the so-called perimarine, Early-Pleistocene deposits in the southern Netherlands (province of Noord-Brabant) and northern Belgium. Perimarine sediments, which are formed in the transition zone between fluviatile and marine environments (Hageman, 1969), were previously studied in Holocene depositional areas in the western Netherlands (i.a. Van der Woude, 1981; Van de Plassche, 1982; Berendsen, 1982). Pleistocene perimarine environments, however, have been investigated less frequently, since they are normally found at greater depth in the North Sea basin.
In this respect western Noord-Brabant and adjacent Belgium form an exception. Early-Pleistocene deposits, characterized by a wide variety of lithofacies, occur close to the surface (1-2 m) in this area.
Former investigations of these deposits were mainly concerned with lithostratigraphical aspects (Lorié, 1907; Doppert and Zonneveld, 1955), but so far the sedimentary environments and paleogeographic evolution have hardly been investigated (Dricot, 1961; De Ploey, 1961; Paepe and Vanhoorne, 1970; Geys, 1975). No detailed bio-chronostratigraphical information was available for the area and the lithostratigraphical position of the units also remained far from clear.
The present study was started in 1984 with a view of contributing to the understanding of the sedimentary history and paleogeographic evolution of the area. Attention was focused on lithological and sediment-petrographical aspects and on the study of the sedimentary processes. Paleobotanical data were used additionally to provide a biochronostratigraphic framework.

The general aim stated above involved work on the following practical aspects:
- First a lithostratigraphic framework had to be developed. So far different lithostratigraphic units, of which the mutual relation was not clear, have been used in The Netherlands and Belgium, (Zagwijn and Van Staalduinen, 1975; Paepe and Vanhoorne, 1976). In this study a new local lithostratigraphic subdivision is proposed (chapter 2).
- The sedimentary environments of the newly established units had to be reinvestigated in detail because the Early-Pleistocene units had been interpreted in turn as estuarine, tidal flat and fluviatile deposits (Van Dorsser, 1956; Dricot, 1961; Paepe and Vanhoorne, 1970; Geys, 1975). Sedimentary environments have been reconstructed in this study by the analysis of sedimentary structures and on the basis of paleobotanical data (chapter 3).
- The age of the Early-Pleistocene deposits was still a matter of controversy. Tiglian up to Cromerian ages have been proposed for equivalent lithostratigraphic units (Van der Vlerk and Florschütz, 1953; De Ploey, 1961). However, the age has to be known more precisely, in order to enable reliable paleogeographic reconstructions. Systematic analysis of pollen, geomagnetic polarity and to a lesser extent, botanical macro remains provided a sufficiently precise chronology for the present study (chapter 4).
- The sediment sources of the various Early-Pleistocene units were incompletely known. Meuse (Van Dorsser, 1956), Rhine (Zagwijn and Van Staalduinen, 1975) and a Scandinavian supply (Dricot, 1961) have been proposed. The provenance of the sediments was investigated systemati-

cally, in order to reconstruct the transport directions (chapter 5). The final aim was to integrate the lithological, sediment-petrographical, palynological and sedimentological results into a paleogeographic and environmental reconstruction of the area during the Early-Pleistocene. The results obtained were fitted into a regional framework on the basis of previously published data (chapter 6).

1.2 Location

The study area is situated at the southern rim of the North Sea basin, in the border area of The Netherlands (province of Noord-Brabant) and Belgium (province of Antwerpen)(fig. 1.1).

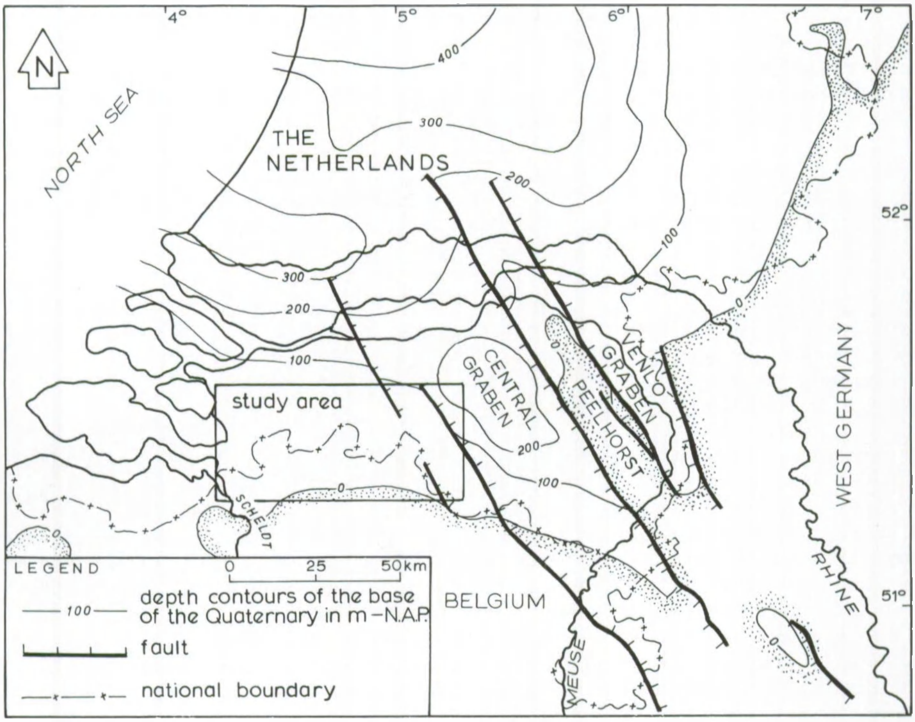


Fig. 1.1: Depth contours of the Quaternary and major tectonic units of the southern North Sea basin (after Zagwijn and Doppert, 1978).

The study area is bounded by the cities Bergen op Zoom, Breda, Tilburg, Turnhout and Antwerp. The southern boundary coincides with the so-called Campine microcuesta (De Ploey, 1961). The eastern limit is formed by the western faults of the Central Graben, where the Early-Pleistocene deposits occur at greater depth. The western boundary consists of the 20 to 25 m high escarpment at the right bank of the Scheldt.
Morphologically the area is characterized by a gently undulating cover-sand relief and south-north oriented brook valleys. The topography slopes to the north, descending from approximately 30 m on the Campine

microcuesta in Belgium to circa 15 m +N.A.P. in the neighbourhood of Gilze (0.7 °/oo). The lithological units in the subsurface show a dip to the north increasing from 1.1°/oo in the Tegelen Formation upto approximately 4 °/oo in the Boom Clay of Oligocene age (Van Rummelen, 1965). The gradient of the Early-Pleistocene units came into being during the Quaternary due to subsidence of the North Sea basin and uplift of the Tertiary deposits in Belgium (Zagwijn and Doppert, 1978). The Early-Pleistocene deposits therefore occur close to the surface (1-2 m) in the border area (fig. 1.2).

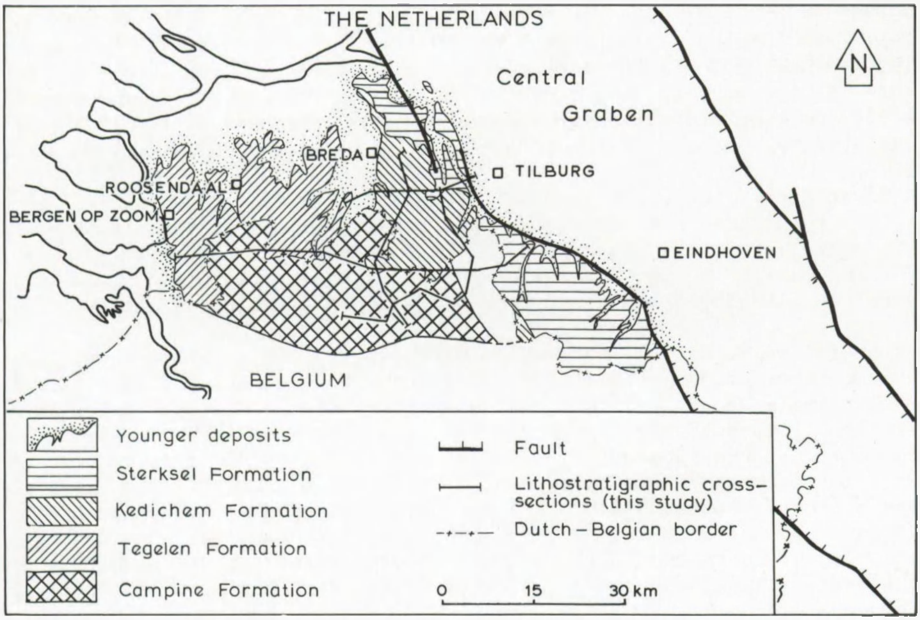


Fig. 1.2: Early- and Middle-Pleistocene formations in Noord-Brabant and northern Belgium (after Zagwijn and Van Staalduinen, 1975; Gulinck, 1962).

In Belgium they are part of the Campine Clay and Sand Formation (Paepe and Vanhoorne, 1976). In The Netherlands they have been mapped as the Tegelen and Kedichem Formations (Zagwijn and Van Staalduinen, 1975). Tertiary deposits crop out south of the Campine microcuesta, while Middle- and Late-Pleistocene deposits cover the Early-Pleistocene sediments north and east of the study area.

2. LITHOSTRATIGRAPHY

2.1 General introduction

Since World War II several studies have been published, which are concerned with parts of the Dutch-Belgian border area (Nelson and Van der Hammen, 1950; Doppert and Zonneveld, 1955; Van Dorsser, 1956; Van Voorthuysen, 1957; Dricot, 1961; De Ploey, 1961; Greguss and Vanhoorne, 1961; Van Oosten, 1967; Paepe and Vanhoorne, 1970; Geys, 1975; Haest, 1985). Recently, special attention has been given to the tectonic history of the western fault system of the Central Graben (Vandenberghe, 1982) and the Weichselian geomorphological development of the Mark and Reusel river basins (Vandenberghe et al., 1984; Van Huissteden et al., 1986).
A first aim of the present study has been:
1: to establish the relationships between the hitherto recognized lithostratigraphic units in The Netherlands and in Belgium.
2: to design a comprehensive regional lithostratigraphic framework, based on lithological and sediment-petrographical characteristics.

As an initial step all available data (literature, internal reports, boring descriptions) have been evaluated. On the basis of this information twelve borehole locations were selected in a S-N and E-W cross-section, respectively perpendicular and parallel to the structure contours of the Quaternary sedimentary basin. The results of the drilling campaign were complemented by detailed study of exposures, to get a better understanding of the three dimensional structure of the various layers encountered. The location of the borings and the exposures is given in fig. 2.1. The most important results are presented in the text, while the original data can be found in alphabetical order in the appendix.
The Geological Survey of The Netherlands and the Geological Survey of Belgium permitted the use of coring decriptions from their archives.

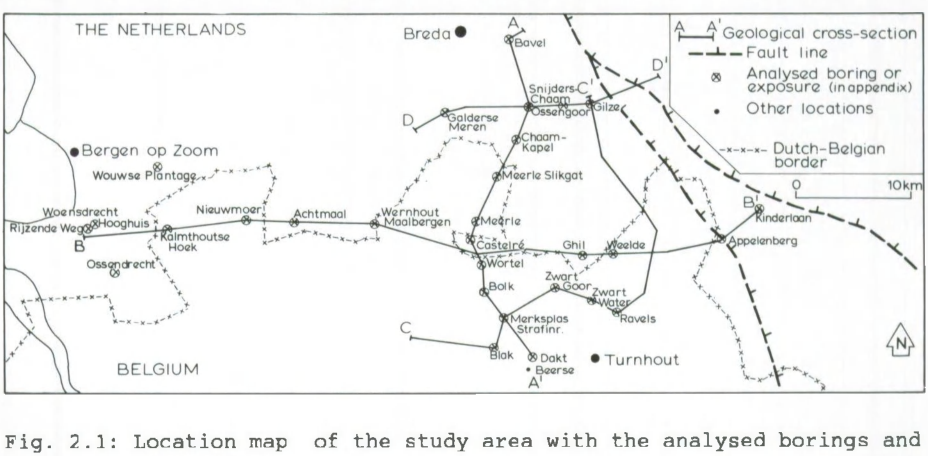


Fig. 2.1: Location map of the study area with the analysed borings and exposures.

2.2 Heavy mineral assemblages in the Upper-Tertiary and Quaternary of The Netherlands

Lithostratigraphic classification of the Upper-Cenozoic sediments in The Netherlands is based essentially on lithological and sediment-petrographical rock properties (Zagwijn, 1986). Edelman (1933) and Baak (1936) were the first in The Netherlands who used differences in heavy mineral composition to distinguish lithostratigraphic units. They introduced several sediment-petrographical provinces, which were denominated by capital letters and suffixes (fig. 2.2). Especially the terms A-, H- and B-Limburg associations are still often cited in the literature. Differences between these sediment-petrographical groups are not always clear and they can, to some extent, be due to grain-size effects (e.g. B-Lobith group and North Hinder group; A-group and H-group (Baak, 1936)). Therefore, it is preferred in this study, in accordance with Zagwijn and Van Staalduinen (1975), to differentiate between lithostratigraphic units on the basis of the relative frequency of stable and unstable heavy minerals. Characteristic and/or dominant heavy minerals are mentioned separately. The terms stable and unstable heavy minerals relate to the susceptibility of the heavy minerals to chemical weathering under temperate climatological conditions. The stable heavy mineral group comprises zircon, rutile, anatase, brookite, staurolite, tourmaline and the metamorphic minerals kyanite, andalusite and sillimanite (Boenigk, 1983). The unstable group includes epidote, alterite, hornblende, garnet and the volcanic heavy minerals hypersthene, augite and enstatite. Titanite, which is the most common mineral within the volcanic heavy mineral group in this study, is more stable however.

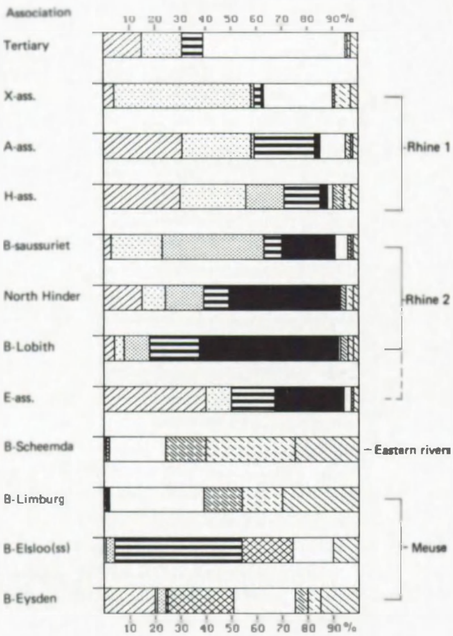
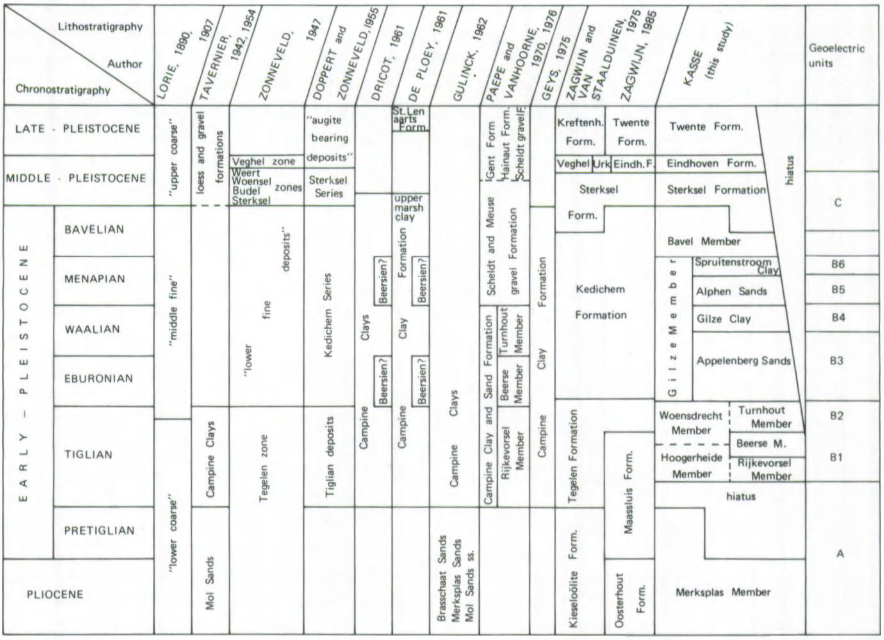


Fig. 2.2: Heavy mineral associations in The Netherlands (after Edelman, 1933; Baak, 1936) (legend in appendix).

2.3 Historical review

In this paragraph the evolution of the lithostratigraphic classification of the Early-Pleistocene deposits in Noord-Brabant and adjacent northern Belgium is reviewed. A general outline of the developments in the lithostratigraphic subdivision and terminology is given in table 2.1. More extensive reviews can be found in Tavernier (1942), Doppert and Zonneveld (1955), Geys (1975) and Paepe and Vanhoorne (1976).
In 1890 Lorié distinguished three major lithological groups in the Pleistocene of the western Netherlands. Later (1907), he labeled these

Table 2.1: Historical review of the lithostratigraphic subdivision of the Early-Pleistocene deposits in Noord-Brabant and northern Belgium.



units as "lower coarse", "middle fine" and "upper coarse" deposits. Until c. 1945 this tripartition was maintained, although the age of these deposits was a matter of debate (Doppert and Zonneveld, 1955).
Sediment-petrographical investigations by Zonneveld (1947) in the southeastern part of Noord-Brabant led to a new lithostratigraphical subdivision of the Quaternary of Noord-Brabant, which was extended to the western Netherlands.
In 1955 Doppert and Zonneveld presented a synthesis in which they integrated their own results with the previous tripartition of Lorié. They correlated the "lower coarse" unit with the Tiglian deposits, characterized by garnet, epidote, hornblende and saussurite-alterite heavy minerals (H-ass.) and by the presence of megasporangia of the waterfern *Azolla tegeliensis*. The "middle fine" unit was renamed as "Kedichem Series", dominated by an H-association; occasionally with a B-Limburg association in the upper part (metamorphic minerals and tourmaline). Doppert and Zonneveld (1955) divided Lorié's "upper coarse" unit into two sub-units. The lower part, with garnet, epidote, saussurite/alterite and hornblende (H-association) was called the "Sterksel Series". The upper part of the "upper coarse" unit, characterized by volcanic minerals, was indicated as the "augite-bearing deposits". According to Doppert and Zonneveld (1955, cross-section II) the Early-Pleistocene deposits in Noord-Brabant, occurring at or close to the surface, belong to the "Kedichem" (H-ass. with B-Limburg ass. at the top) and "Sterksel Series" (H-ass.).
Zagwijn and Van Staalduinen (1975) presented a synthesis of the lithostratigraphic subdivision and terminology of Late-Tertiary and Quaternary deposits in The Netherlands. In contrast to the ideas of

Doppert and Zonneveld (1955), they included large parts of the Early-Pleistocene deposits in western Noord-Brabant in the Tegelen Formation (fig. 1.2). Lorié's "lower coarse" and probably part of the "middle fine" unit, in their opinion, are part of this formation. In the type-locality of the Tegelen Formation in the province of Limburg (The Netherlands) this formation consists of a typical alternation of fluvial sand (gravel) and clay-layers with a H-association (Belfeld Gravel, Belfeld Clay, Tegelen Gravel, Tegelen Clay). According to Bisschops et al. (1985) this alternation cannot be recognized west of the Central Graben. In certain parts of Noord-Brabant the Tegelen Formation overlies the (marine) shell-bearing, sandy Early-Pleistocene Maassluis Formation or is laterally equivalent to it.
According to Zagwijn and Van Staalduinen (1975) the Kedichem Formation is equivalent to the "middle fine" unit of Lorié (1907) and the "Kedichem Series" of Doppert and Zonneveld (1955). The formation consists of fine sand-, clay- and peat-layers and occurs close to the surface in the vicinity of Baarle-Nassau (fig. 1.2). In the Central Graben area the Kedichem Formation is characterized by a B-Limburg heavy mineral association. This assemblage facilitates the distinction with the Tegelen and Sterksel Formations which both have an H-association. In the central Netherlands the Kedichem Formation is characterized by an H-association. A transition zone with a mixture of stable (B-Limburg) and unstable (H-ass.) heavy minerals was found in the northwestern part of the Central Graben (Zagwijn and Van Staalduinen, 1975). Part of Lorié's "upper coarse" unit and the "Sterksel Series" of Doppert and Zonneveld were incorporated in the Sterksel Formation, which crops out along the western fault system of the Central Graben (Zagwijn and Van Staalduinen, 1975)(fig. 1.2). Zonneveld (1947) distinguished four heavy mineral zones (Sterksel, Budel, Woensel, Weert) within the coarse, gravelly sands of the Sterksel Formation. The Budel zone contains more stable heavy minerals in comparison to the other zones, which are dominated by garnet, epidote, saussurite-alterite and hornblende in variable percentages. In the region directly east of the study area only the Sterksel and Woensel (and possibly Weert) zones are present (Bisschops et al., 1985).

The Early-Pleistocene deposits in Belgium (Campine region) have been studied since the end of the 19th century. A lithostratigraphic classification and terminology was established, which differs from the one applied in The Netherlands (table 2.1). Tavernier (1942), Geys (1975) and Paepe and Vanhoorne (1976) give a summary of the differences of opinion concerning the age and depositional environment of the deposits. Especially the Late-Tertiary Mol Sands and overlying Early-Pleistocene Campine Clays are important in the study area (Huyghebaert, 1961; Vanhoorne, 1962). Gulinck (1962) divided the Mol Sands s.l. into three units: the Brasschaat Sands in the west, the Merksplas Sands underneath the Campine Clays and Sands, and the Mol Sands s.s. around Mol. Paepe and Vanhoorne (1976) correlated the oölite bearing Merksplas Sands with the "kieseloölitic" terrace deposits along the Meuse in the Ardennes.
The Campine Clays (Tavernier, 1942, 1954), overlying the Merksplas Sands, consist of an alternation of sand- and clay-layers. Gulinck (1962) identified two clay-layers in exposures, while a third was found in some borings.
Dricot (1961) established a tripartition in the Campine Clays with an upper clay, middle so-called Beersien sand and lower clay-unit. This intercalated sand-unit is characterized by a B-Limburg association, whereas the clay-units are dominated by garnet, epidote and hornblende

(A-ass.).
De Ploey (1961) agreed with Dricot's tripartition of the Campine deposits. In addition he described a sand-layer on top of the upper Campine Clay at Ravels and St. Lenaarts, which is characterized by a B-Limburg heavy mineral association. This unit was called the St. Lenaarts Formation and interpreted as a Weichselian fluvial deposit (Early Glacial-Pleniglacial), which was later confirmed by Haest (1985).
Paepe and Vanhoorne (1970, 1976) renamed Dricot's units. The lower clay was called the Rijkevorsel Formation and later the Rijkevorsel Member; the upper clay the Turnhout Formation and Turnhout Member. The intervening sand-unit was called the Beerse Formation and later the Beerse Member, in spite of the fact that De Ploey (1961) had already defined a Beerse Formation of Weichselian age. They correlated the Rijkevorsel, Beerse and Turnhout Formations respectively with the Tiglian, Eburonian and Waalian periods in The Netherlands. The tripartition of the Early-Pleistocene deposits was also maintained by Haest (1985).
Geys (1975) studied several grain-size characteristics of Early-Pleistocene deposits in northern Belgium. In his study lithostratigraphical data and analysis of sedimentary structures are relatively scarce. Since Geys incorporated all Early-Pleistocene deposits in a single unit (Campine Clay Formation), regardless of their lithology, heavy mineral composition, stratigraphic position and age, his conclusions are at least inaccurate.

Summarizing, different lithostratigraphic systems are used in The Netherlands and Belgium. The lithostratigraphic nomenclature of the two countries is difficult to combine. Paepe and Vanhoorne (1970) for example placed Belgian lithostratigraphic units into the Dutch chronostratigraphic time scale, although the litho- and chronostratigraphic correlation is uncertain, while Geys (1975) based his environmental interpretations on doubtful lithostratigraphic correlations. In order to avoid erroneous correlations and to establish a coherent lithostratigraphic framework in the Dutch - Belgian border area, this study first concentrates on the lithology, heavy mineral composition and lithostratigraphic position of the various units.

2.4 Geoelectric cross-section in the Dutch-Belgian border area

J. Vandenberghe & C. Kasse

Introduction

In order to determine the general lithostratigraphic sequence of the Pleistocene layers in the study area and to get an impression of the spatial distribution and lateral variability of individual units a geoelectric survey has been performed between the Holocene Scheldt estuary and the Central Graben. Two important cross-sections are discussed in this paragraph: in the west a cross-section between Woensdrecht and Meerle and in the east a cross-section over the western faulted boundary zone of the Central Graben near Lage Mierde. The latter part is an extension and modification of the previously published "profile 3" (Vandenberghe, 1982). The geoelectrical cross-sections are more or less parallel to the lithostratigraphic cross-section BB' (fig. 2.1 and 2.6).
The geoelectric survey concerns the uppermost 20-30 m of the Quaternary strata. The profiles comprise Schlumberger soundings with a maximal half-electrode distance of 100-170 m and generally a 600 m interval in between. Power was provided by a lightweight generator and transformed

to direct electric current. In order to avoid manifest technical errors, the sounding curves were calculated in the field. The interpretation consisted of two steps: at first a geoelectric model was set up according to the method developed by Koefoed (1970) and Ghosh (1971), secondly the model was modified and upgraded by an interactive automatic method using the Marquardt procedure (Johansen, 1975); converted and adapted for handling on a PC by Drs. K. Hemker (Free University, Amsterdam) and by taking advantage of all the available geologic information.
Unfortunately, geoelectric interpretation is complicated by the rapidly changing character of the Quaternary deposits. Therefore, the data of a few deep boreholes combined with reference electric soundings and some "mini-electric" measurements (Vandenberghe, 1982) were very helpful in determining the specific resistivity values of the different lithologic units and thus in overcoming the problem of equivalence.

General results

According to Archie's law the specific resistivity of water-saturated materials is dependent on the specific resistivity of the water and the grains, the porosity and a cementation factor. The material properties are well approximated by clay content and sediment sorting. Consequently, the bulk layer resistivities may be expressed in terms of grain-size, provided that the pore-water resistivity is more or less constant in the region of study. Although the validity of the latter condition may be questioned, the following relationships between electric specific resistivities and lithologies are found (water-saturated conditions):

$< 50\ \Omega m$: clay, sandy clay
$50\text{-}73\ \Omega m$: (sandy) silt, alternations of sand (silt) and clay
$73\text{-}191\ \Omega m$: fine sand
$> 191\ \Omega m$: medium to coarse sand

In the westernmost part of the section extremely high specific resistivities occur due to the deep water table (WT) near the incised Scheldt valley (fig. 2.3). Non-saturated top sediments (1 to 2 m) also show high resistivities.
In southwestern and central N-Brabant three main lithologic units with characteristic specific resistivities are distinguished (Vandenberghe, 1982 and 1983). The lower unit (A) is only found in a few electric soundings with deep penetration in the western cross-section (fig. 2.3). Unit A shows relatively high specific resistivities indicative of medium to coarse sands. Corresponding borehole data allow correlation of this unit with the Maassluis Formation or Kieseloölite Formation (table 2.1). The overlying deposits have a considerable thickness and consist mainly of fine sands, clays and alternating fine sands and clays (unit B)(fig. 2.3, 2.4). Locally, gullies occur which are filled up with coarse (gravelly) sands. This unit comprises the Kedichem and Tegelen Formations and also the (clayey) top of the Maassluis Formation as defined by the Rijks Geologische Dienst (Zagwijn and Van Staalduinen, 1975). The upper unit (C) consists of gravels and sands of various specific resistivities (fig. 2.4). Generally, the sands are medium to coarse-sized, although layers of fine sand and thin, but widespread, sheets of loam are present too. These only occur in the eastern part of the area and correspond to the Sterksel Formation (table 2.1). Finally, the uppermost Late-Quaternary sediments have a thickness of max. a few meters. These will not be considered here.

Cross-section Woensdrecht-Meerle (fig. 2.3)

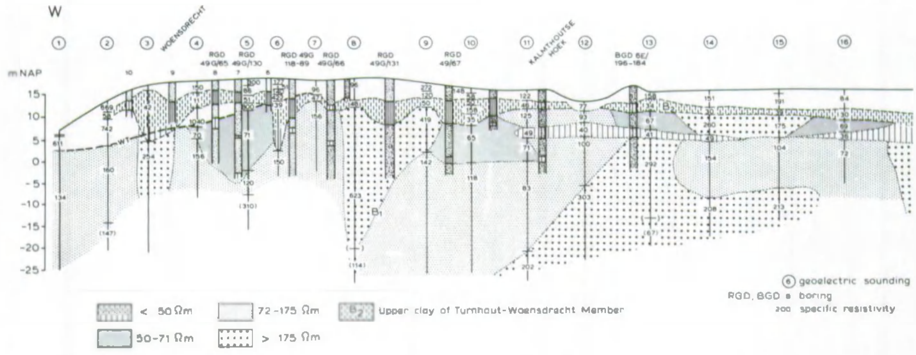
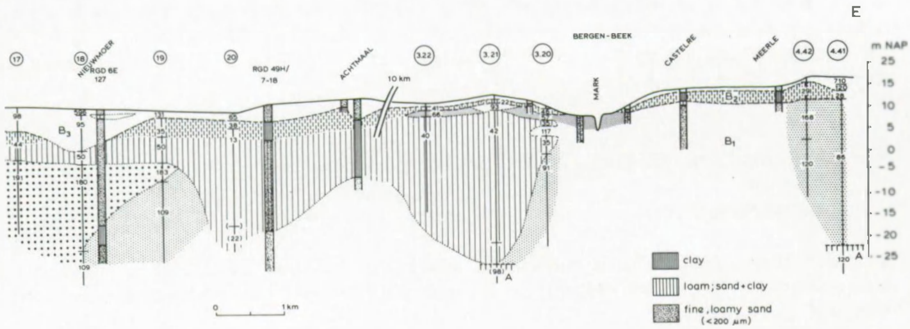


Fig. 2.3: Geoelectric cross-section in the Dutch-Belgian border area, between Woensdrecht and Meerle.

Unit B shows a general increase in grain-size in a western direction so that the distinction with unit A, even with the help of lithologic data from boreholes, is difficult to realize in the western part of the section. The lower part of unit B (B_1) is characterized by the occurrence of coarse-grained gully sediments (e.g. soundings 3, 8, 13, 17, 18, 19) surrounded by progressively finer deposits (fine sands and ultimately clays). These channel sediments may have been formed in different periods and their super-position was accompanied by erosion of the underlying deposits. Consequently, the occurrence of clay lenses may be rather irregular. Towards the top of sub-unit B_1, however, clay-layers of larger extent appear. A 0.5 to 2.5 m thick layer at c. 10 m depth in soundings 11 to 14 coincides further to the east with the more important upper clay-layer (B_2). The latter clay-unit (B_2) can be traced over the whole section, except at some points where it is eroded as a result of much younger river incision (e.g. soundings 1, 12). In the areas of a clayey developed (upper part of) sub-unit B_1, the individuality of the clay-unit B_2 is lost in the geoelectric results (e.g. soundings 20, 3.22 and 3.21). The top of sub-unit B_2 is marked by a clear erosional (gullying) contact with the sandy sub-unit B_3.

Cross-section Lage Mierde (fig. 2.4)

The clay-bed B_2 is most regularly developed and clearly expressed in the western part of this profile; B_1 and B_3 have strikingly higher specific resistivities. In contrast to cross-section Woensdrecht-Meerle the base of B_3 shows no gullying character. To the east (no. 19, 20, 23, 24), the lower boundary of B_2 is less pronounced, while the specific resistivities of B_1 and B_3 decrease somewhat so that the distinction between B_1 and B_3 is less obvious.
The most prominent feature of this profile is the presence of a few subvertical/vertical discontinuities, previously recognized as faults of the southwestern marginal zone of the Central Graben (Vandenberghe, 1982). Apart from local uplifting (soundings 21, 22), a general subsidence of the layers towards the east is obvious. Consequently, younger deposits start occurring in that direction. Sub-unit B_4 is a clay-layer showing similar specific resistivities to B_2. It is superposed by the sandy sub-unit B_5. In the easternmost part of the section the deposition within unit B is terminated by a thin (sandy) clay (sub-unit B_6).



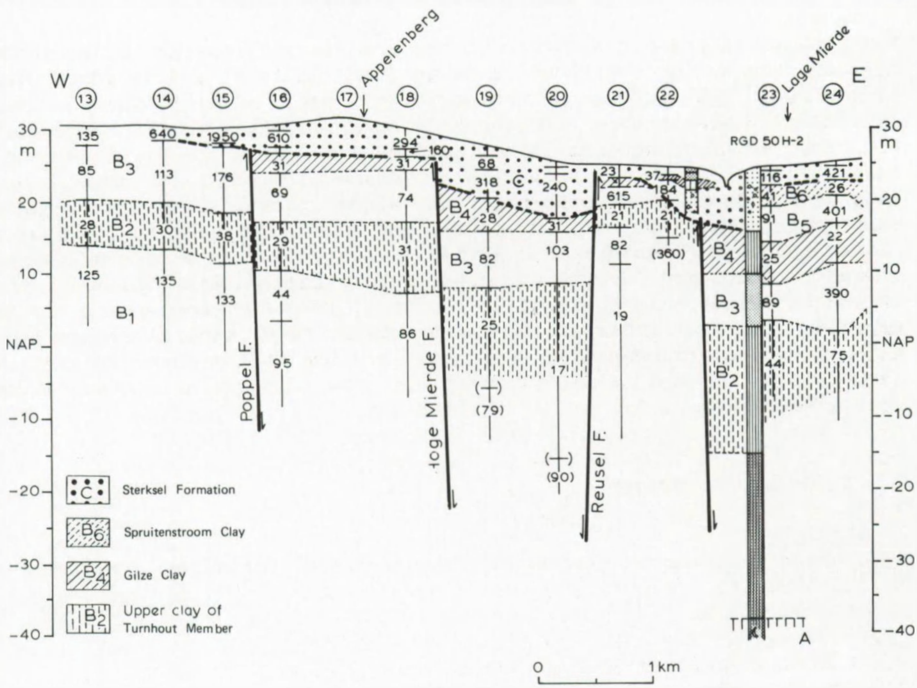


Fig. 2.4: Geoelectric cross-section in the Dutch-Belgian border area between Weelde and Lage Mierde.

The top of unit B is strongly affected by erosion at the base of unit C. The coarse deposits of unit C cross the fault zone to the west over a short distance. At the eastern side of the Hoge Mierde fault the altitude of the base of unit C is distinctly lower than at the west side while the thickness of unit C is only slightly increasing and thus the surface at the eastern side of the fault zone is somewhat lower than at the western side. This fact implies that the fault activity is contemporaneous and posterior to the deposition of the Sterksel Formation, more specifically the Woensel heavy mineral zone, which is dated as Cromerian-B by Bisschops et al., (1985). No evidence for such recent activity has been found along the Poppel Fault, the youngest activity of which - as revealed by the geoelectric data - did not exceed the Early-Pleistocene.

2.5 Description of the lithostratigraphic units

2.5.1 Introduction

In this paragraph a comprehensive picture of the lithostratigraphy of the Early-Pleistocene deposits in the Dutch-Belgian border area is presented, including:

1: definition of basic lithostratigraphic units.
2: subdivision of existing lithostratigraphic units (formations).
3: correlation of lithostratigraphic units on both sides of the Dutch-Belgian border.
4: an explanation of lateral east-west facies relationships in Noord-Brabant.

The lithostratigraphic sequence of the area is illustrated by a number of north-south and east-west cross-sections parallel and perpendicular to the structure contours of the North Sea basin (fig. 2.1, 2.5, 2.6, 2.7, 2.8). These enable the recognition of lateral facies changes. In the cross-sections several lithostratigraphic units are distinguished, based on lithological properties, sedimentological trends (fining-upward) and heavy mineral composition. Since formal lithostratigraphic units of formation rank have already been defined for this area (Zagwijn and Van Staalduinen, 1975), these units are designated as members. They are described according to the recommendations of the international stratigraphic guide (Hedberg, 1976). Stratotype, regional aspects, general lithologic characteristics, heavy mineral composition and references are presented systematically for each member. Grain-size classes follow the classification of the Geological Survey of The Netherlands (fig. 2.9).

2.5.2 Merksplas Member

<u>1. Stratotype</u>

- Location: Boring Merksplas Strafinrichting described by Delvaux, 1890-1891)(fig. 2.10).
  x = 4°49'55''EL
  y = 51°21'23''NL
  H (surface level) = 26.2 m +N.A.P.
  D (depth) = 28 to 46 m below the surface.
- Lithology: medium coarse, white quartz sands, with reworked organic material and some clay-beds. A gravel-bed occurs in the upper part of the member (26-28 m) and another at the base of the member (46.0-46.1 m).
- Heavy mineral composition: unknown.
- Gravel composition: 26-28 m: quartz, quartzite, flint, cristalline; 46.0-46.1 m: quartz, quartzite, flint, cristalline, psammite, schist, phyllite, arkose, grès (tertiary), silicic oölite.

<u>2. Regional aspects</u>

The Merksplas Member is confined to the southern part of the investigated area (fig. 2.5, 2.7). The top of the member dips approximately

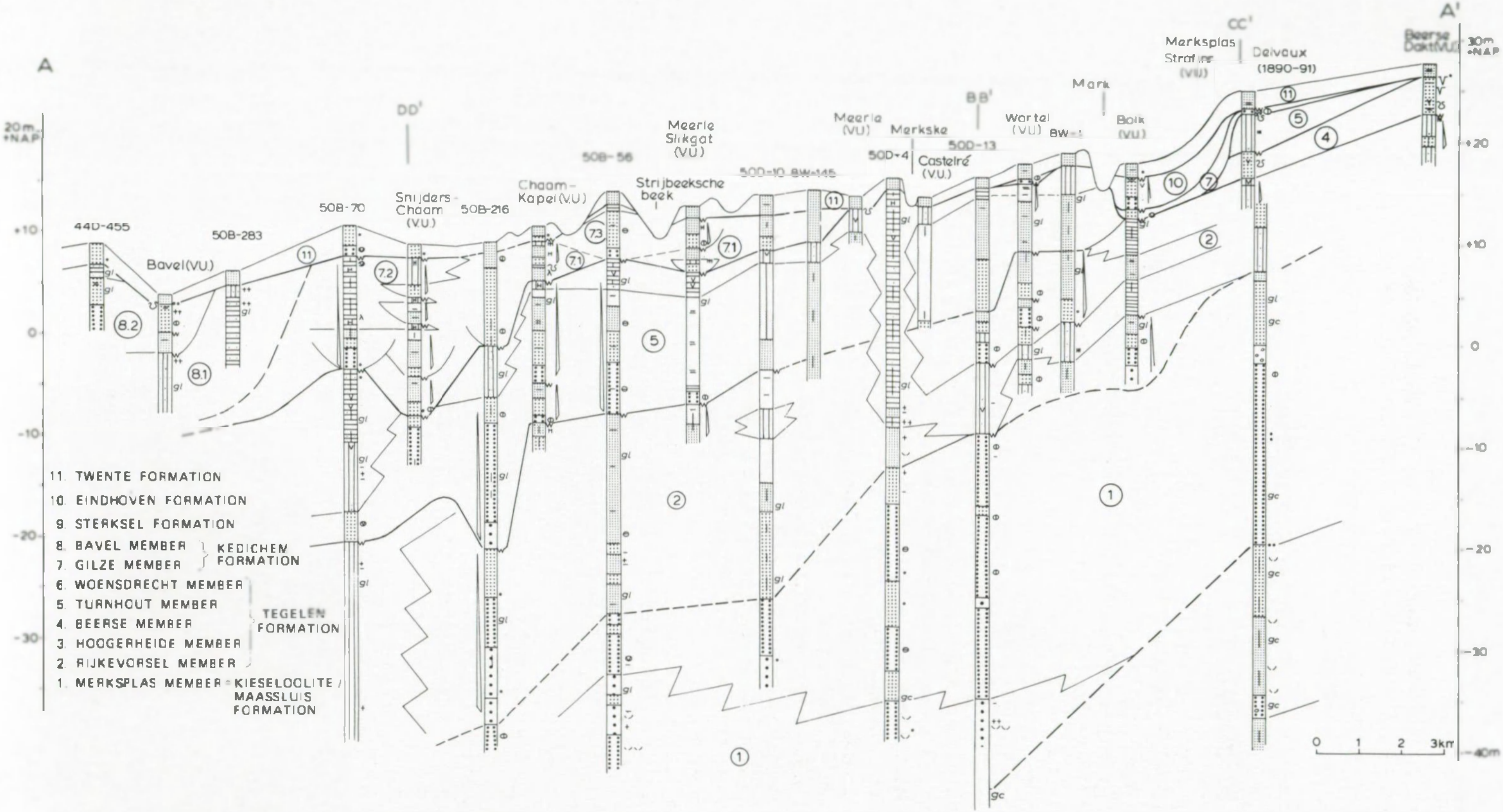


Fig. 2.5: Lithostratigraphic cross-section A-A' between Beerse and Bavel (legend in appendix).

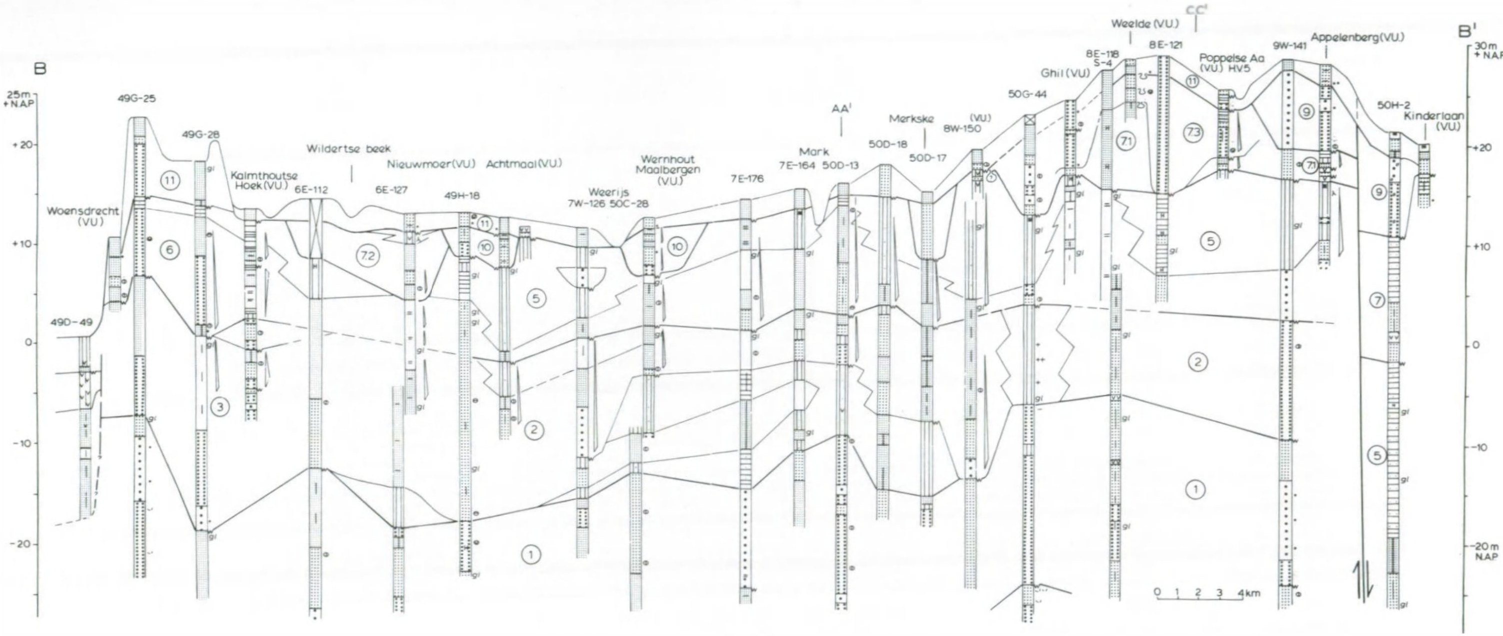


Fig. 2.6: Lithostratigraphic cross-section B-B' between Woensdrecht and Lage Mierde (more or less parallel to fig. 2.3) (legend in appendix).

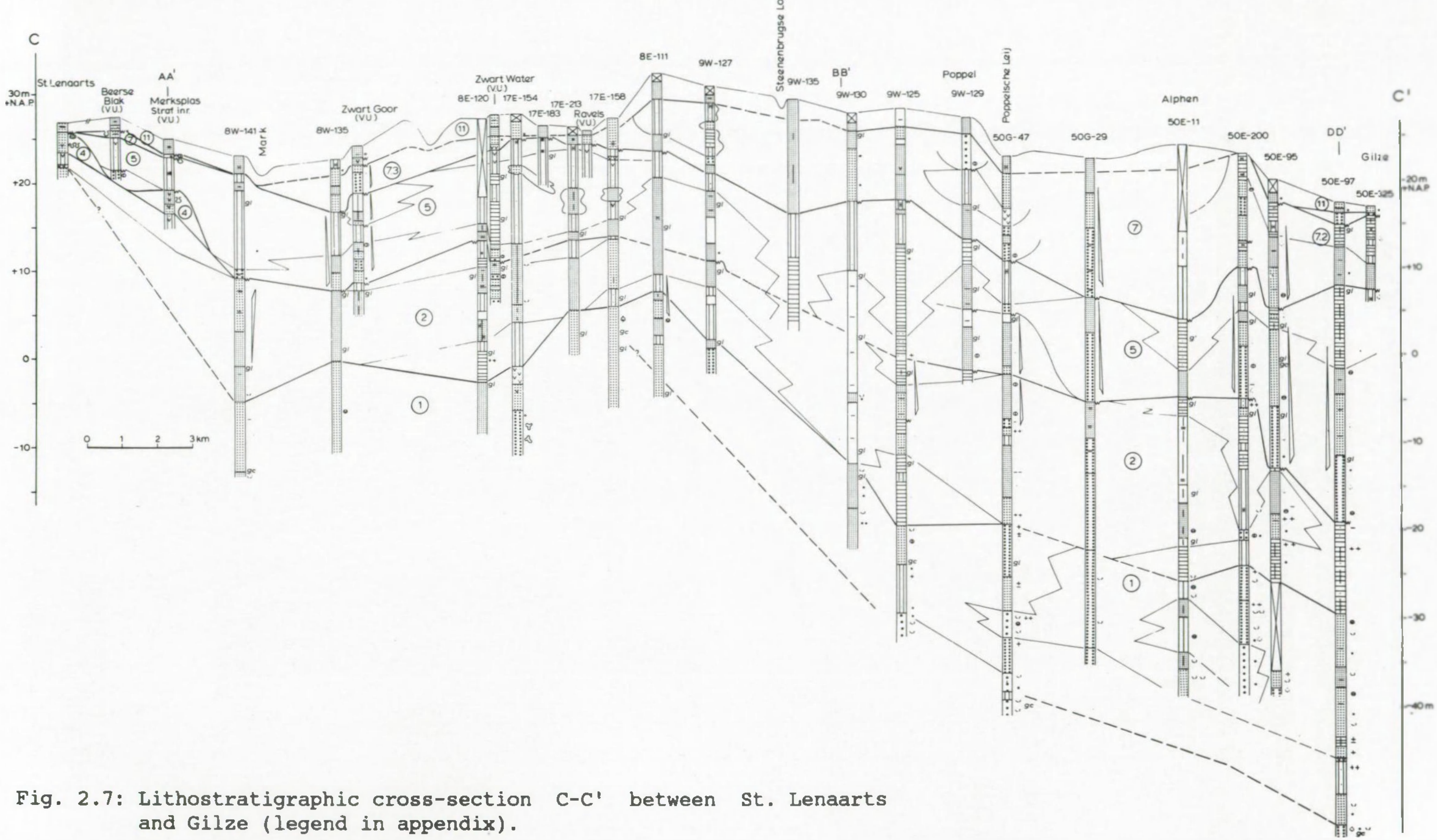


Fig. 2.7: Lithostratigraphic cross-section C-C' between St. Lenaarts and Gilze (legend in appendix).

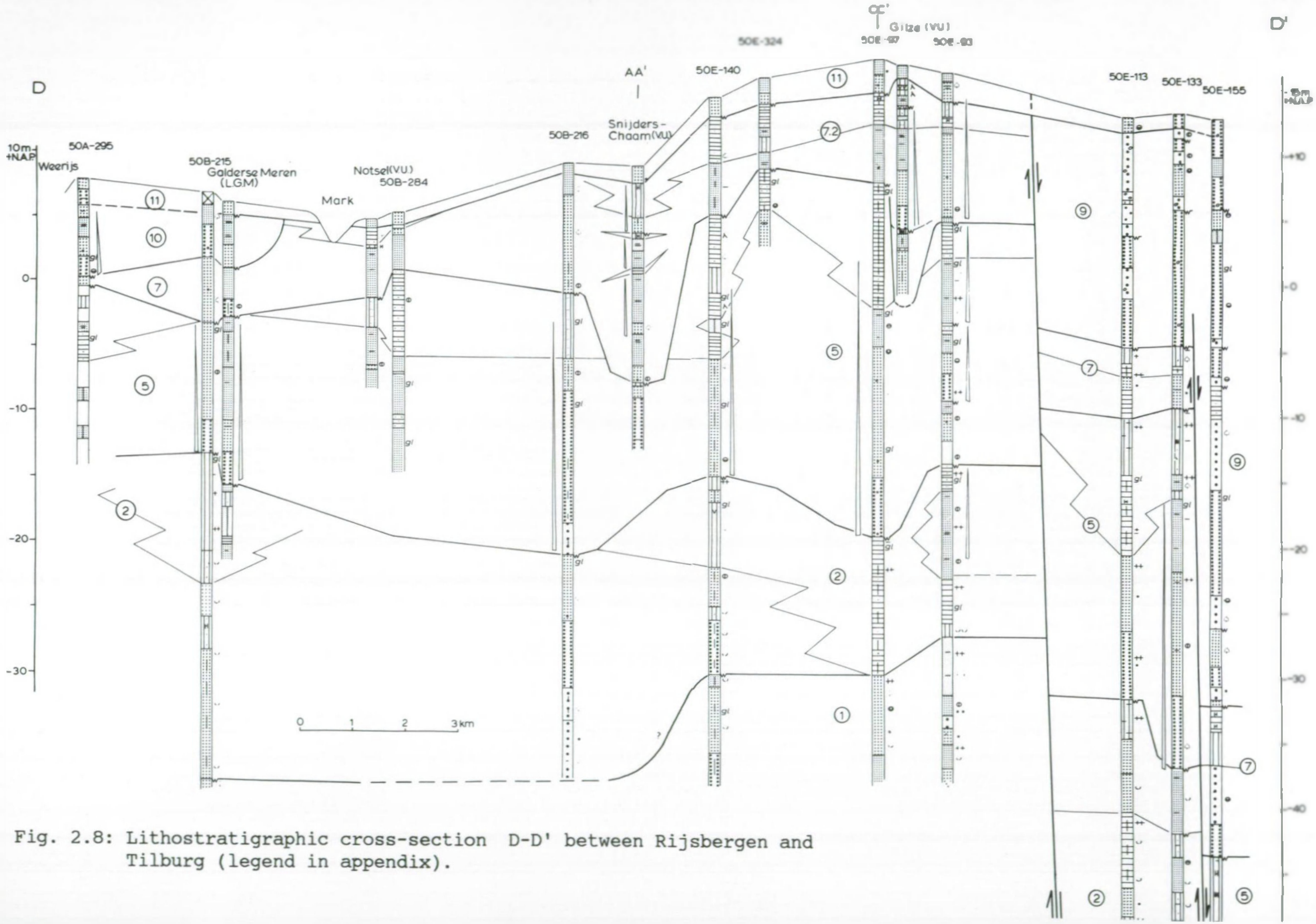


Fig. 2.8: Lithostratigraphic cross-section D-D' between Rijsbergen and Tilburg (legend in appendix).

2.3 °/oo to the north (fig. 2.5). This value is in close agreement with the mean dip of 2.2 °/oo of the top of the Merksplas Sands in 6 north-south cross-sections between Kalmthout and Poppel (pers. comm. P. Laga, Geological Survey of Belgium). If the same dip is extended to the south the Merksplas Member crops out south of the line Turnhout-Oostmalle.
The Merksplas Member is overlain by the Rijkevorsel Member. A sharp transition is found between sands of the Merksplas Member and clay of the Rijkevorsel Member in some borings (fig. 2.5: boring 50D-13; fig. 2.7: boring 17E-154). No gradual fining-upward sequence occurs in this respect, probably reflecting a hiatus in the sedimentation. The Merksplas Member overlies shell-bearing sands, locally with glauconite, belonging to the Maassluis and/or Oosterhout Formations (fig. 2.5: 50D-4, 50D-13).

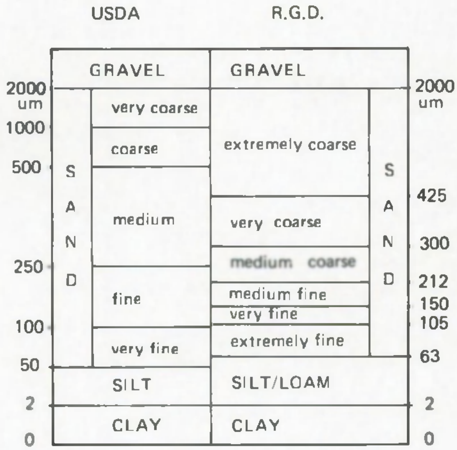


Fig. 2.9: Particle-size classes of the U.S. Department of Agriculture and the Geological Survey of The Netherlands.

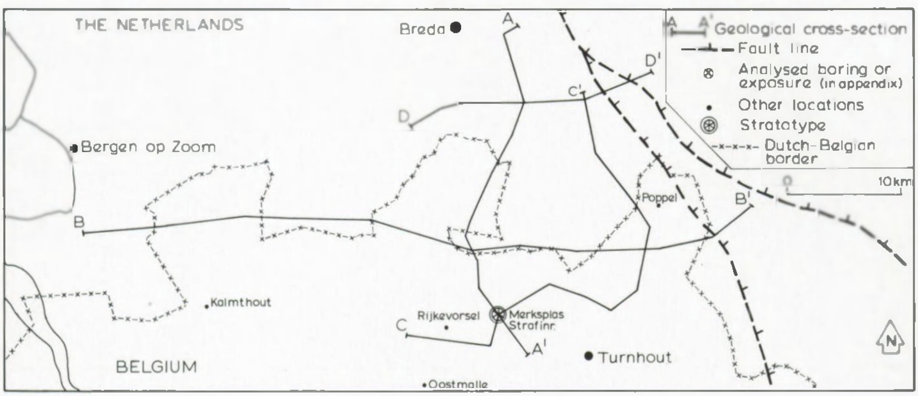


Fig. 2.10: Location map of the Merksplas Member.

3. General lithologic characteristics

The lithology of the Merksplas Member is dominated by medium fine to coarse, non-calcareous, gray sands, often with clay-pebbles and reworked organic material. Loam- and clay-beds are generally rare. A peat-layer is locally present at the top (fig. 2.6: boring S4; fig. 2.7: boring 17E-154 (Vanhoorne, 1962)).
Lateral facies changes have not been studied in detail. An important change occurs in a northern (western) direction. The non-calcareous Merksplas Sands grade laterally into calcareous, shell-bearing, medium to coarse sands of the Maassluis Formation (fig. 2.5, 2.7).

4. Heavy mineral and gravel composition

The Merksplas Member is characterized by predominantly stable heavy minerals. According to Huyghebaert (1961, boring Rijkevorsel: 27-60 m; boring Oostmalle: 1.5-15 m) 20-40% unstable heavy minerals are present. The gravel composition has been described above. White quartz, flint and silicic oölites are common constituents of the Merksplas Sands (Gulinck, 1962; Paepe and Vanhoorne, 1976).

5. References

Delvaux (1890-1891): Type locality boring Merksplas Strafinrichting.
Halet (1920): Qualitative description of gravel components.
Huyghebaert (1961): Sediment-petrographical composition.
Vanhoorne (1962): Mol Sands are overlain by Campine Clays.
Gulinck (1962): Merksplas Sands.
Paepe and Vanhoorne (1976): Merksplas Sands part of Mol Sands s.l.
Zagwijn and Van Staalduinen (1975): Kieseloölite Formation, Tegelen Formation, Maassluis Formation.
Haest (1985): Zanden van Berzegem, Zanden van de Konijnenberg.

2.5.3 Rijkevorsel Member

1. Stratotype

- Location: Clay-pit Beerse Dakt (fig. 2.11).
  x = 4°52'5''EL
  y = 51°19'55''NL
  H (surface level) = 28 m +N.A.P.
  D (depth) = 19-23 m +N.A.P.; below 19 m +N.A.P. not exposed.

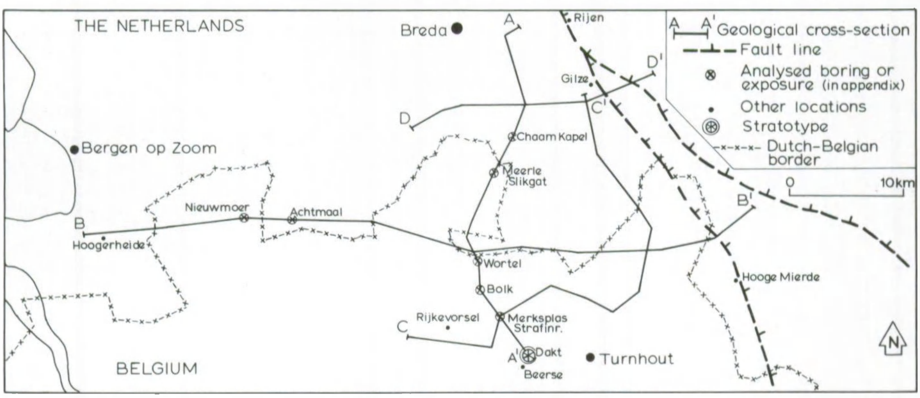


Fig. 2.11: Location map of the Rijkevorsel Member.

- Lithology: Bluish gray, non-calcareous, stiff clay, crumbly at the top, with intercalated beds (1 mm - 10 cm) of extremely fine to medium coarse, white sand in the central part of the unit (coarsening-up and fining-upward sequence)(App.). A peaty clay-layer is present at the top.
- Heavy mineral composition: Mixed heavy mineral association. Coarse-grained intervals dominated by stable heavy minerals (tourmaline, andalusite, staurolite, zircon, rutile). Finer grained beds are characterized by unstable heavy minerals (garnet, epidote, hornblende) (§5.3).
- Boundaries: Lower boundary unknown. Upper boundary is formed by a

sharp/erosive transition from clay of the Rijkevorsel Member to sand of the Beerse Member.

2. Regional aspects

The Rijkevorsel Member is well recognizable in the southeastern part of the investigated area in the neighbourhood of the villages Rijkevorsel, Merksplas and Beerse (fig. 2.11). In this region the Rijkevorsel Member is underlain by sands of the Merksplas Member and overlain by sands of the Beerse Member (fig. 2.5, 2.7). North of Merksplas the Beerse Member is absent, due to erosion, which occurred before the deposition of the Turnhout Member. The Turnhout Member directly overlies the Rijkevorsel Member in this situation and as both units have many characteristics in common they may be difficult to distinguish (fig. 2.5: boring Wortel). With the aid of erosional contacts, fining-upward sequences and a northward dip of approximately 1.4 °/oo (Beerse Dakt - Merksplas Strafinr.) it is possible to trace the top of the Rijkevorsel Member at least as far north as Breda (fig. 2.5: Meerle Slikgat, 50B-70).
The surficial occurrence of the Rijkevorsel Member is bounded in the east by the Central Graben. East of the Gilze-Rijen and Hooge Mierde fault the Rijkevorsel Member is situated at greater depth and cannot be reached by shallow boring techniques (fig. 2.4: B1).

3. General lithologic characteristics

The Rijkevorsel Member consists of a blue-gray, non-calcareous, stiff clay with a high lutum content in the southeastern part of the area. Peaty horizons occur locally in the uppermost part of the Rijkevorsel Member. In the middle part of the member an alternation of very fine to medium sand and clay-layers is present, forming one fining-upward sequence with the overlying clay at the top of the member. A clay-layer may be present at the base of the Rijkevorsel Member (fig. 2.5, 2.7: 8W-1, 50D-4, 50D-17, 7E-176, 17E-154, 9W-127). This clay overlies sand of the Merksplas Member and locally has a peaty or humic character (8E-120, 50D-13).
Towards the north the Rijkevorsel Member dips at approximately 1.4 °/oo and changes in lithology (fig. 2.5). The clayey lithofacies, which occurs in the neighbourhood of Rijkevorsel and Merksplas diminishes in thickness and sand-layers become more important. Clay-layers are locally important however (50D-4, 50B-70). Approximately 7 km south of Breda the Rijkevorsel Member becomes calcareous and shell-fragments of Cardium and Mytilus species occur locally, especially within or below thick (less permeable) clay-beds (fig. 2.5: 50B-70; fig. 2.8: 50E-140, 50B-215).
To the west the Rijkevorsel Member consists of non-calcareous, gray, moderately to well-sorted, extremely fine to medium fine, micaceous sands with clay-layers (fig. 2.6). In the neighbourhood of Nieuwmoer-Kalmthoutse Hoek the lithological and sediment-petrographical characteristics of the Rijkevorsel Member change and another lithostratigraphic unit is defined (Hoogerheide Member).

4. Heavy mineral composition

The Rijkevorsel Member in the southeastern part of the area is characterized by a mixed heavy mineral association, in which stable and unstable heavy minerals are equally represented (fig. 2.13). The amount of unstable heavy minerals increases rapidly to the north (fig. 2.5 and appendix: 40 % in Beerse Dakt, 55 % in Merksplas Straf., 70-80 % in Bolk and Wortel and up to 90 % in Meerle-Slikgat and Chaam-Kapel. As there is a tendency for grain-size to increase in a northern direction within the Rijkevorsel Member, this heavy mineral change could be caused by grain-size effects. However, the analysis of fractionated samples reveal that this change in heavy mineral composition is due to different sediment sources (see §5.3 and §5.4.2).

A change in the heavy mineral composition also occurs to the west (fig. 2.6 and appendix). Unstable heavy minerals percentages decline from 70-80% in Wortel to 60% in Achtmaal. This decline, which continues to the west in the Hoogerheide Member (20-40 % unstable heavy minerals), is also explained by changes in sediment provenance (§5.3 and §5.4.4).

5. References

Tavernier (1942, 1954): Campine Clays.
Dricot (1961): Tripartition in the Campine Clays.
Greguss and Vanhoorne (1961): Palynology of the Campine Clays.
Paepe and Vanhoorne (1970): Rijkevorsel Formation.
Geys (1975): Campine Clay Formation.
Zagwijn and Van Staalduinen (1975): Tegelen Formation.
Paepe and Vanhoorne (1976): Rijkevorsel Member. To prevent overburdening with new names this old lithostratigraphic name is maintained, although the definition of the Rijkevorsel Member in this study is not completely equivalent to the definition given by Paepe and Vanhoorne (1970, 1976). They studied the Rijkevorsel Member around Beerse, where a clay lithofacies dominates (Rijkevorsel Clay Formation: Paepe and Vanhoorne, 1970). To the north a sandy lithofacies is also present (see above), which in this study is considered to be part of the Rijkevorsel Member as well.

2.5.4 Beerse Member

1. Stratotype

- Location: clay-pit Merksplas Strafinrichting (fig. 2.12).
  x = 4°49'51''EL
  y = 51°21'45''NL
  H = 25 m +N.A.P.
  D = 16-19 m +N.A.P.

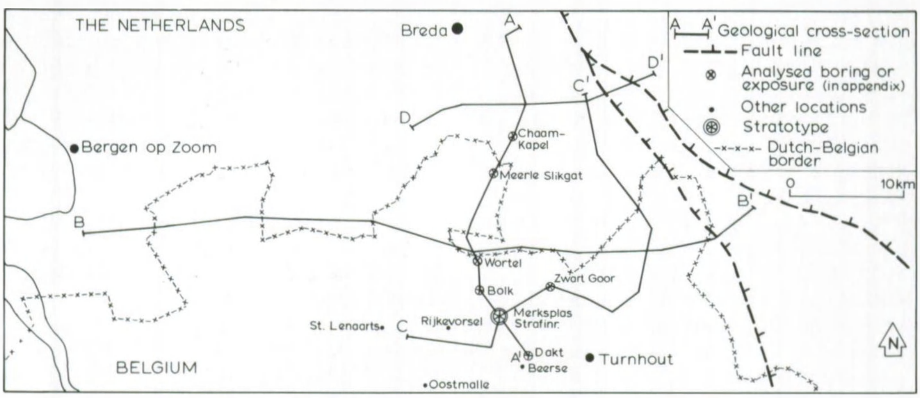


Fig. 2.12: Location map of the Beerse Member.

- Lithology: very fine to medium fine, yellowish-white, moderately to well-sorted sands, with 4 intercalated soil horizons. The soils are of the podzolic type on higher paleotopographic sites and peaty at lower lying places. The second soil from the top is strongly deformed (amplitude up to 70 cm)(see Appendix).
- Heavy minerals: stable heavy minerals (more than 90%: tourmaline, andalusite, staurolite, zircon, rutile)(fig. 2.13). Local variations

are due to grain-size effects (§5.3).

- Boundaries: the Beerse Member overlies the Rijkevorsel Member with a transitional boundary. The upper boundary to the Turnhout Member is clearly erosional (gullying).

2. Regional aspects

The Beerse Member is only found in the southeastern part of the investigated area in the neighbourhood of St. Lenaarts, Rijkevorsel, Beerse and Merksplas (fig. 2.5, 2.7). It is restricted to the highest parts of the Campine microcuesta (De Ploey, 1961) in a zone 5 km wide (north-south) and approximately 15 km long (east-west). The unit has not been found in The Netherlands (fig. 2.6, 2.8). The Beerse Member overlies the Rijkevorsel Member and is covered by the Turnhout Member. The lower boundary can be sharp/erosive (Beerse Dakt) or transitional (Merksplas Strafinr.). The upper boundary is always erosional. Prior to sedimentation of the Turnhout Member, the Beerse Member has been eroded by gullies (Stratotype Merksplas Strafinr.). North and east of Merksplas the Beerse Member has been removed completely. In boring Bolk (fig. 2.5) the Beerse Member may still be present although it cannot be excluded that the sand dominated by stable heavy minerals belongs to the Gilze Member as the stratigraphical position is ambiguous. North of boring Bolk the Turnhout Member rests directly onto the Rijkevorsel Member (fig. 2.5: Meerle Slikgat; fig. 2.7: Zwart Goor; see also: Paepe and Vanhoorne, 1970, fig. 5; Dricot, 1961, fig. 1; Haest, 1985, fig. II. 1b). Separation of the Turnhout and Rijkevorsel Members can be difficult then, especially if the erosional contact between the two members is not well expressed (e.g. boring Wortel), but further north a clear erosional boundary exists between the two members (fig. 2.5: Meerle Slikgat, Chaam Kapel).

3. General lithologic characteristics

The Beerse Member is characterized by two lithofacies. Lithofacies 1 (Stratotype Merksplas Strafinr., Beerse Dakt) consists of very fine to medium fine, moderately well-sorted sands, with maximal 4 intercalated humic to peaty soil horizons. The peaty soil horizons, which formed under moister conditions, locally reveal well-developed deformation structures with amplitudes up to 70 cm (appendix Merksplas Strafinr.: second soil; Beerse Dakt: fourth soil from the top). In the intervening sands small frost cracks (65 cm deep) and small ice-wedge casts (25 cm wide) were regularly found (Beerse Dakt). Lithofacies 1 of the Beerse Member resembles periglacial deposits of Weichselian age. For instance in clay-pit Beerse Dakt, where the Beerse Member occurs close to the surface, this member is found directly under the Weichselian deposits, only separated from the latter by a gravel-bed. However, sediment-petrographically the two units can be easily separated, as the Beerse Member is dominated by stable heavy minerals, while the Weichselian deposits have a mixed heavy mineral association (see appendix).
Lithofacies 2 (Beerse Dakt) consists of medium to coarse, moderately sorted sands with many clay-pebbles. Gravel occurs at the base. Lithofacies 1 has locally been eroded by lithofacies 2. Due to incomplete exposure the exact relation between the two facies could not be established. According to R. Mortier (pers. comm.) the two facies are laterally related and synchronous.

4. Heavy minerals and gravel

The Beerse Member can be discerned sediment-petrographically from the Rijkevorsel and Turnhout Members (fig. 2.13). In contrast to the mixed and unstable heavy mineral associations of the Rijkevorsel and Turnhout Members, the Beerse Member (lithofacies 1 and 2) is dominated completely by stable heavy minerals (more than 90 %).

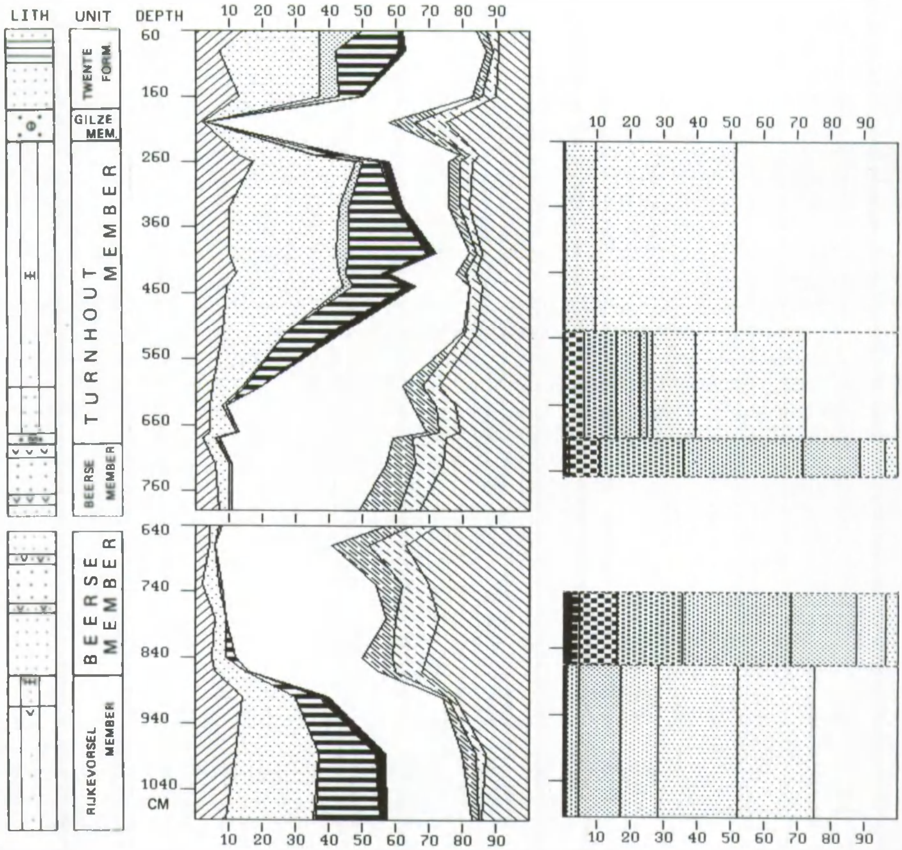


Fig. 2.13: Heavy mineral diagram of the Rijkevorsel Member, Beerse Member and Turnhout Member at Merksplas Strafinrichting (legend in appendix).

The stable heavy mineral association of lithofacies 2 distinguishes these medium coarse sands of the Beerse Member from medium sands at the base of the Turnhout Member at Wortel and Meerle Slikgat, which are characterized by an unstable heavy mineral association. Small variations within the stable heavy mineral association can be attributed to grain-size effects (§5.3). Tourmaline is concentrated within the coarser fractions; zircon and rutile in the finer fractions.
The stable heavy mineral association of the Beerse Member resembles the heavy mineral composition of the Merksplas and Gilze Members. In the south both Beerse and Merksplas Members occur close to the surface. The Gilze Member can directly overlie the Beerse and/or Merksplas Member in these situations and distinction may be difficult due to the absence of the Turnhout and/or the Rijkevorsel Member (see De Ploey, 1961: St. Lenaarts Formation at St. Lenaarts (Early/Middle Weichselian), later reinterpreted as Beerse Formation (Early-Pleistocene) by Paepe and Vanhoorne, 1970).
Gravel (3-5 mm) was found once at the base of lithofacies 2 in Beerse Dakt. The gravel composition is described in §5.4.3. The larger flint

particles often show well-rounded sides, which seem to reflect the outer parts of former, larger, ellipsoid particles, prior to disintegration.

5. References
Dricot (1961): Beersian deposits.
De Ploey (1961): Campine Clays; part of the St. Lenaarts Formation.
Paepe and Vanhoorne (1970): Beerse Formation.
Geys (1975): Campine Clay Formation.
Paepe and Vanhoorne (1976): Beerse Member.
Haest (1985): Old-Pleistocene deposits.

2.5.5 Turnhout Member

1. Stratotype

- Location: clay-pit Merksplas Strafinrichting (fig. 2.14).
  x = 4°49'51''EL
  y = 51°21'45''NL
  H = 25 m +N.A.P.
  D = c. 18.5-23 m +N.A.P.
- Lithology: bluish gray, stiff, non-calcareous clay. Fining-upward sequence: clay with medium coarse sand-beds at the base; crumbly clay with high lutum content and a very humic soil horizon in the upper part (App.).
- Heavy minerals: mixed association (garnet, epidote, hornblende, zircon, rutile, tourmaline)(fig. 2.13). More stable heavy minerals (zircon, rutile, staurolite, andalusite, tourmaline) at the base, due to reworking of underlying sands of the Beerse Member (see §5.3).
- Boundaries: lower boundary to the Beerse Member is erosive. The upper soil horizon of the Beerse Member has been eroded in the northwest corner of the exposure. Upper boundary to the Gilze Member and/or Twente Formation is sharp/erosive, with large (1 m) sand-filled involutions and ice-wedge casts formed in the clay.

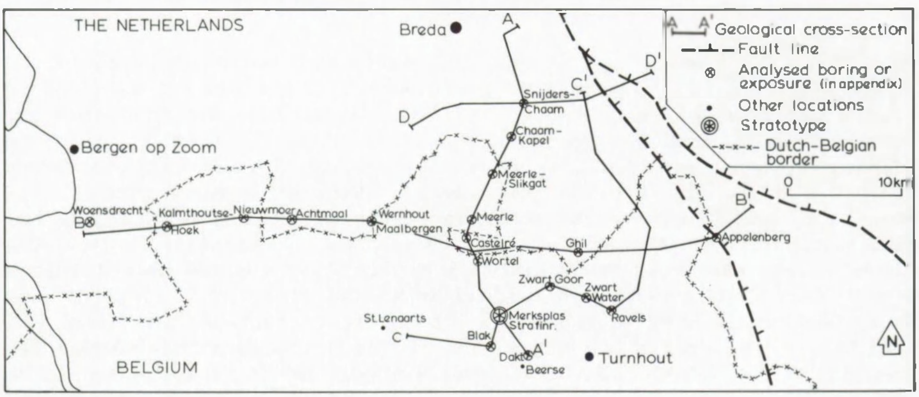


Fig. 2.14: Location map of the Turnhout Member.

2. Regional aspects

The areal distribution of the Turnhout Member is vast (fig. 2.5, 2.6, 2.7, 2.8). Only in the neighbourhood of Beerse, St. Lenaarts, is it absent (fig. 2.5, 2.7) due to non-deposition or by later erosion.
The Turnhout Member is underlain by the Beerse Member in the southeast-

ern part of the area (stratotype). In most cases however, the Turnhout Member rests directly on the Rijkevorsel Member, separated from the latter by an erosive boundary with accompanying changes in lithology (fig. 2.6, 2.8). The Turnhout Member is covered by the Gilze Member or Eindhoven and Twente Formation (fig. 2.5, 2.6, 2.7). The boundary is erosional in many cases. Locally no lithological break occurs (fig. 2.6: Appelenberg; fig. 2.7: Zwart Water), but sediment-petrographically a clear change is always present (fig. 2.15).
The thickness of the Turnhout Member increases to the north from less than 1 m at Beerse Dakt to 20 m in 50 B-216 (fig. 2.5).

3. General lithologic characteristics

In the south the Turnhout Member consists of a blue-gray, stiff, non-calcareous, micaceous clay, with high lutum content (fig. 2.5, 2.7: Merksplas Strafinr., Beerse Blak, Ravels). The clay-layer generally decreases in thickness to the north and changes laterally into extremely fine to very fine, loamy, bluish to greenish-gray, micaceous, well-sorted sand (fig. 2.5: Wortel). This sand is not the northward extension of the Beerse Member, because the heavy mineral composition and depositional environment are different. North of Wortel grain-size further increases and the sand becomes medium fine to medium coarse, non-calcareous, light gray, well-sorted and has low silt/clay content (fig. 2.5: Chaam Kapel, Snijders-Chaam). Local exceptions occur, consisting of thick (10-15 m) loam-, clay- or extremely fine sand-layers, which are lateral equivalent with the surrounding sands (fig. 2.5: Meerle Slikgat, 50D-4, 50B-70).
The Turnhout Member contains a clay-layer at the top (greenish-gray to bluish, non-calcareous, stiff, high lutum content, intercalated peaty soil horizons), which forms a marker-bed over the whole investigated area (fig. 2.3, 2.4, 2.5, 2.6). This clay-bed is separated from the underlying sands by alternating sand-clay laminations (fining-upward sequence; Kasse, 1986). Thickness variations of the clay are due to lateral facies changes (fig. 2.5: 50D-4, Meerle) and post-deposition erosion. The Turnhout Member has been eroded to a maximum of 10-15 m in the Mark-Weerijs river valleys, including Weichselian Late-Glacial erosion (Vandenberghe et al., 1984)(fig. 2.5, 2.6).

4. Heavy minerals

The Turnhout Member is in general dominated by garnet, epidote and hornblende (fig. 2.15; Appendix: Zwart Goor, Chaam Kapel). Glaucophane occurs in very low amounts (< 1 %). This assemblage distinguishes the Turnhout Member from the Beerse and Gilze Members (fig. 2.13). The Turnhout Member and the Rijkevorsel Member both have an unstable heavy mineral association (App. Wortel, Meerle Slikgat, Chaam Kapel). The amount of unstable heavy minerals decreases to the south (Chaam Kapel, Meerle slikgat: 80-90%, Wortel: 75%, Merksplas Strafinr.: 60-70%). The unstable heavy mineral association is replaced by a mixed heavy mineral association in the southeast around Merksplas (fig. 2.5, 2.7: Beerse Blak, Merksplas Strafinrichting). The southward decrease in garnet and alterite can be caused by grain-size effects, because grain-size decreases to the south. The increase in stable heavy minerals cannot be explained by grain-size effects. Probably a second sediment source with stable heavy minerals is involved, resulting in a mixed heavy mineral association (see §5.3). A comparable decrease in unstable heavy minerals was found in a western direction (fig. 2.6: Castelré: 85%, Wernhout Maalbergen: 75%, Achtmaal: 60-85%, Nieuwmoer: 70%, Kalmthoutse Hoek: 55-80%). West of Kalmthoutse Hoek (fig. 2.17) this tendency results in heavy mineral assemblages dominated by stable heavy minerals (fig. 2.18). On lithological and sediment-petrographical grounds a separate member is defined there, (Woensdrecht Member).

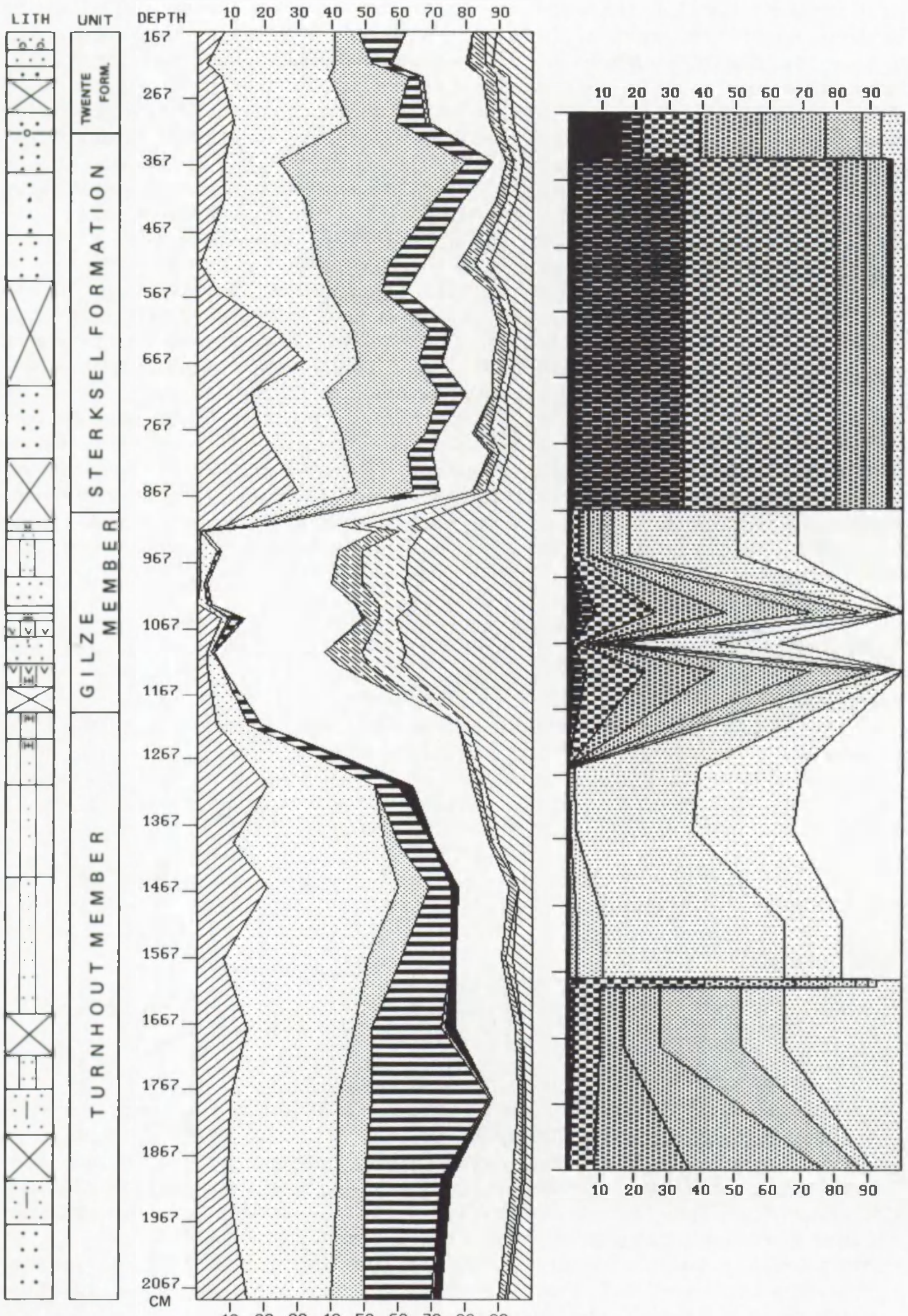


Fig. 2.15: Heavy mineral diagram of the Turnhout Member, Gilze Member and Sterksel Formation at Appelenberg (legend in appendix).

Besides lateral changes, vertical changes in the heavy mineral composition occur as well within the Turnhout Member. Hornblende and alterite content decrease upward, epidote and the stable heavy minerals increase. Garnet often shows a low peak in the upper spectra (fig. 2.15; Appendix: Zwart Water, Castelré, Meerle Slikgat). This vertical heavy mineral sequence is absent when the top of the Turnhout Member has been eroded before the deposition of the Gilze Member (Zwart Goor, Ghil, Wernhout Maalbergen, Achtmaal). Part of the vertical changes can be explained by grain-size effects, because they occur in a fining-upward sequence. Alterite is always connected to the coarser fractions. Epidote and zircon often dominate the finer grain-size fractions. However, the upward decrease of hornblende and increase of tourmaline are in contradiction to grain-size effects (§5.3). Boenigk (1983) explained comparable heavy mineral changes in the High Terrace deposits of the Rhine at Frechen by weathering and heavy mineral solution. According to Boenigk apatite will dissolve first, followed by pyroxenes, hornblende, garnet and epidote. Apatite and pyroxenes are indeed absent in most of our heavy mineral diagrams and a decrease of hornblende is accompanied by an increase of epidote towards the top. Locally, however, hornblende decreases towards the top, while apatite remains present (Appelenberg, Zwart Water, Zwart Goor). Although heavy mineral solution may have been important in certain borings it is more likely that part of the vertical heavy mineral changes were caused by a change in sediment source area.

5. References

Tavernier (1942, 1954): Campine Clays.
Doppert and Zonneveld (1955): part of the Kedichem Series in Noord-Brabant (cross-section II).
De Ploey (1961): upper marsh clay-layer.
Van Oosten (1967): Old-Pleistocene clay (estuarine deposits).
Paepe and Vanhoorne (1970): Turnhout Formation.
Geys (1975): Campine Clay Formation.
Zagwijn and Van Staalduinen (1975): Tegelen Formation.
Paepe and Vanhoorne (1976): Turnhout Member.

2.5.6 Hoogerheide Member

1. Stratotype

- Location: boring Kalmthoutse Hoek (fig. 2.16).
  x = 4°24'42''EL
  y = 51°26'02''NL
  H = 13.5 m +N.A.P.
  D = below 2.5 m +N.A.P. or below 0.5 m -N.A.P.
- Lithology: medium fine to coarse, non-calcareous, moderate to well-sorted sands, with many clay-pebbles.
- Heavy minerals: stable to mixed heavy mineral association (fig. 2.17), characterized by garnet, epidote, zircon, rutile, tourmaline. Coarser fractions dominated by staurolite, andalusite, tourmaline; finer fractions contain more garnet, epidote (40%)(§5.3).
- Boundaries: lower boundary unknown (below 7.5 m -N.A.P.). Upper boundary (0.5 m -N.A.P.) erosive to the overlying possibly fluviatile sands at the base of the Turnhout Member.

2. Regional aspects

The Hoogerheide Member is a newly introduced member in western Noord-Brabant (fig. 2.6). It is overlain by the Woensdrecht Member, separated from the latter by an erosive boundary and a change in grain-size. Below the Hoogerheide Member predominantly non-calcareous sands and locally clays are found, which could be part of the Merksplas Member. The Hoogerheide Member is laterally equivalent with the Rijkevorsel Member east of Kalmthoutse Hoek, since it overlies the Merksplas Member and is overlain by the Turnhout Member. Compared to the Rijkevorsel Member the Hoogerheide Member is coarser grained, contains less silt- and clay-layers and less unstable heavy minerals. The upper part of the Hoogerheide Member and the Woensdrecht Member are missing west of the south-north escarpment, which was formed by erosion of the Scheldt river (fig. 2.6). The Hoogerheide Member dips to the north and crops out in the neighbourhood of Putte.

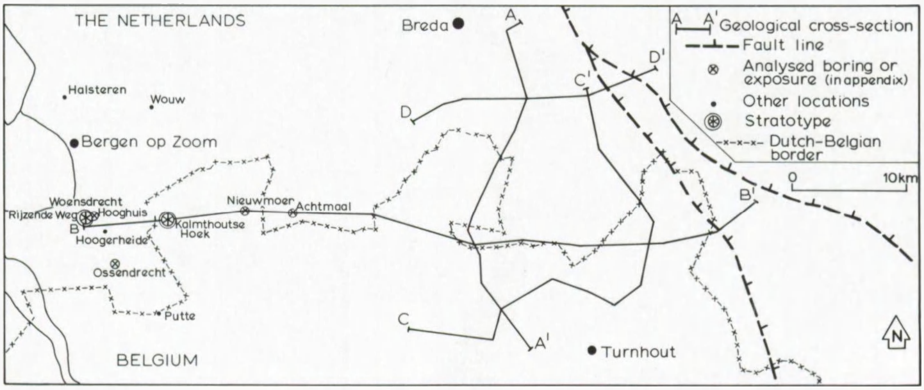


Fig. 2.16: Location map of the Woensdrecht and Hoogerheide Members.

3. <u>General lithologic characteristics</u>

Two lithofacies are recognized within the Hoogerheide Member. Lithofacies 1 (upper part Hoogerheide Member): very fine, very well-sorted, clean, white, non-calcareous sand, with flasers (clay, silt, organic material and mica)(App. Woensdrecht Hooghuis). Heavy mineral laminae are present locally (Woensdrecht Rijzende Weg).

Lithofacies 2 (fig. 2.6: Kalmthoutse Hoek, 49G-25, 49G-28): medium fine to medium coarse, non-calcareous, moderate to well-sorted, light brownish-gray sands, with occasional sharply-bounded clay-laminae and beds. Greenish-gray, compact clay-pebbles of all sizes and reworked organic material are common. Gravel is scarce however.

Lithofacies 1 and 2 often form a fining-upward sequence, at places capped by a clay-layer (borings 49G-66, 75, 90, 130, 164: not presented in fig. 2.6). In boring Kalmthoutse Hoek (fig. 2.6) the Hoogerheide Member consists of several relatively thin fining-upward sequences. Superposed upon these small-scale fining-upward sequences a general upward decrease in grain-size is present.

4. <u>Heavy minerals</u>

The Hoogerheide Member is characterized by a stable to mixed heavy mineral association, dominated by garnet, epidote, zircon, rutile and tourmaline (fig. 2.17). A comparable association was found by Nelson and Van der Hammen (1950: boring Bergen op Zoom and Wouw). Garnet, epidote and hornblende increase in the finer grained top of the member (fig. 2.17). The mixed heavy mineral association points to 2 sediment sources (§5.3 and §5.4.4). The fine-grained sediment contains a rather unstable heavy mineral association, while the coarser fractions have more stable heavy minerals.

The content of unstable heavy minerals increases to the west, north and

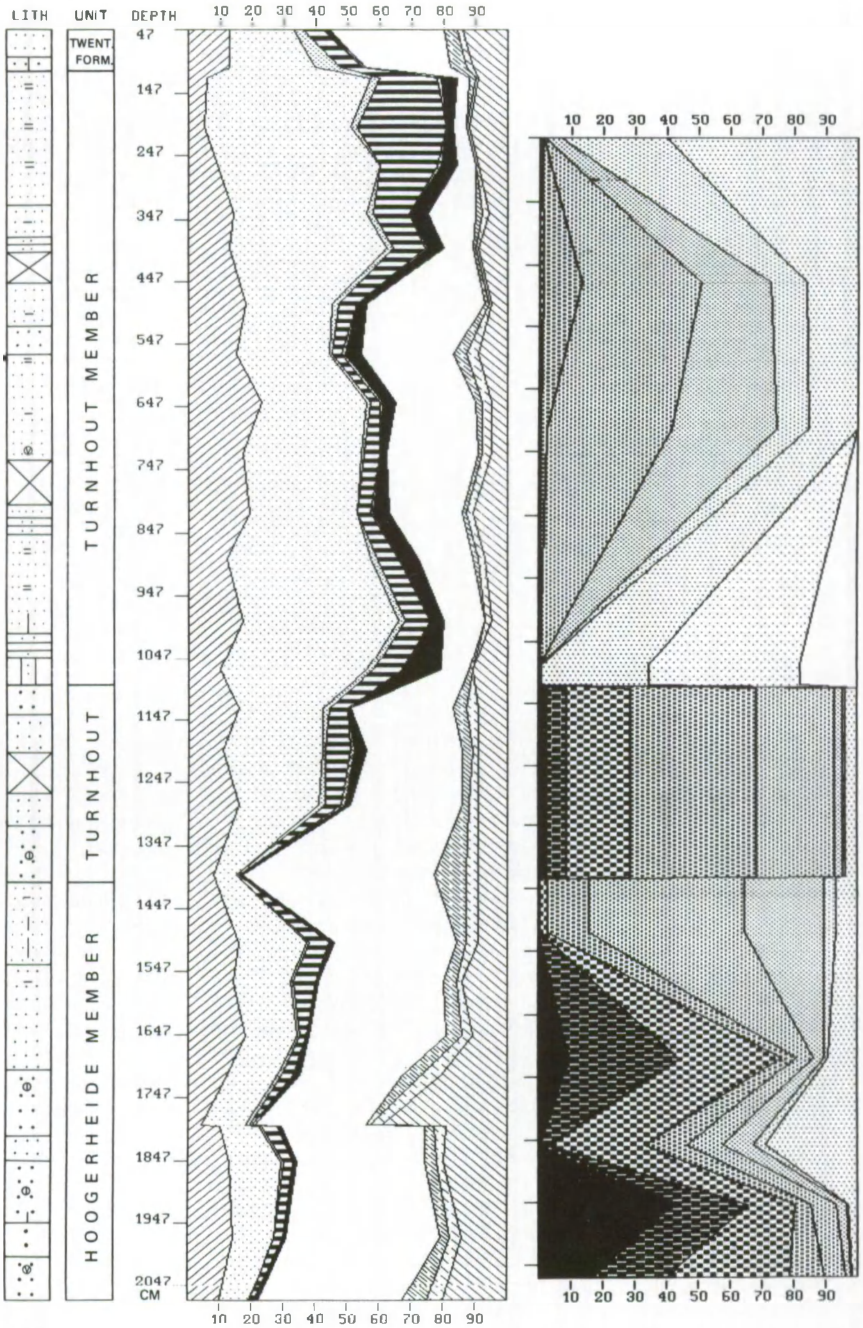


Fig. 2.17: Heavy mineral diagram of the Hoogerheide Member and Turnhout Member at Kalmthoutse Hoek (legend in appendix).

east (Rijkevorsel Member: Nieuwmoer, Achtmaal; Nelson and Van der Hammen, 1950: boring Wouw (22-26m?); Verbraeck and Bisschops, 1971: boring Klundert; Van Rummelen, 1970, 1972, 1978: Tegelen Formation; Van Staalduinen, 1979: boring Maassluis).

5. References

Nelson and Van der Hammen (1950): II-0 deposits ("Preglacial older than the High Terrace").
Van Dorsser (1956): II-0 sand- and loam-beds.
Van Voorthuysen (1957): Halsteren deposits.
Zagwijn and Van Staalduinen (1975): Tegelen Formation.

2.5.7 Woensdrecht Member

1. Stratotype

- Location: Woensdrecht Rijzende Weg (sand-pit)(fig. 2.16).
  x = 4°18'26''EL
  y = 51°26'03''NL
  H = 10.5 m +N.A.P.
  D = 3.5-9.5 m +N.A.P.
- Lithology: extremely fine to medium coarse, non-calcareous, well-sorted, white sands. Fining-upward sequences. Many clay-pebbles, clay-beds/laminae/flasers.
- Heavy minerals: mixed association (epidote, hornblende, zircon, rutile, tourmaline).
- Boundaries: erosional lower boundary to Hoogerheide Member; erosional upper boundary to Twente Formation (coversands).

2. Regional aspects

The Woensdrecht Member is a newly introduced lithostratigraphic unit in western Noord-Brabant. The Woensdrecht Member overlies the Hoogerheide Member erosively. It is covered with medium fine sand of the Twente Formation, often separated from the latter by a gravel-layer (Beuningen gravel-bed; fig. 2.6).
The Woensdrecht Member has been eroded by the Scheldt west of the escarpment (fig. 2.6). The clay-top of the Woensdrecht Member can be continued northwards to Bergen op Zoom and Halsteren (Van Voorthuysen, 1957: Halsteren deposits). The clay (lithofacies 3) forms a continuous layer to the east, connecting the top of the Woensdrecht Member with the top of the Turnhout Member (fig. 2.3, 2.6). In contrast to the Turnhout Member, the Woensdrecht Member is coarser grained and contains fewer clay-beds (fig. 2.6). Thick, isolated clay-lenses, which occur regularly within the Turnhout Member (Achtmaal) are absent in the Woensdrecht Member.

3. General lithologic characteristics

Three lithofacies are distinguished within the Woensdrecht Member.
Lithofacies 1 (App. Woensdrecht Hooghuis, Ossendrecht): medium fine to coarse, moderately well-sorted, non-calcareous, yellowish white sand with many clay-pebbles (up to 10 cm), widely spaced clay-pebble laminae or -beds, and some reworked plant material. Bioturbation is scarce (Woensdrecht Hooghuis: gully base).
Lithofacies 2 (App. Woensdrecht Rijzende Weg): extremely to very fine, well to very well-sorted, non-calcareous sands, with many clay-laminae or flasers (heterolithic facies) and few clay-pebbles and reworked organic remains. Bioturbation is locally present. Grain-size decreases upwards or remains constant, while locally the clay content and amount of clay-laminations diminish (Woensdrecht Rijzende Weg).
Lithofacies 3: massive, non-calcareous, compact, gray to blue-gray, micaceous clay. Humic or peaty horizons are present locally (fig. 3.21).

Lithofacies 3 caps the Woensdrecht Member in a large part of the area (fig. 2.6). The local absence of the clay-layer is probably due to later erosion along the escarpment.
The three lithofacies overlie each other in a fining-upward sequence. The lower boundary of lithofacies 3 can be transitional (Ossendrecht) or abrupt (fig. 3.21). In the latter case a humic or peaty clay directly overlies sand of lithofacies 2.

4. Heavy minerals

The Woensdrecht Member is characterized by a stable to mixed heavy mineral association (fig. 2.18).

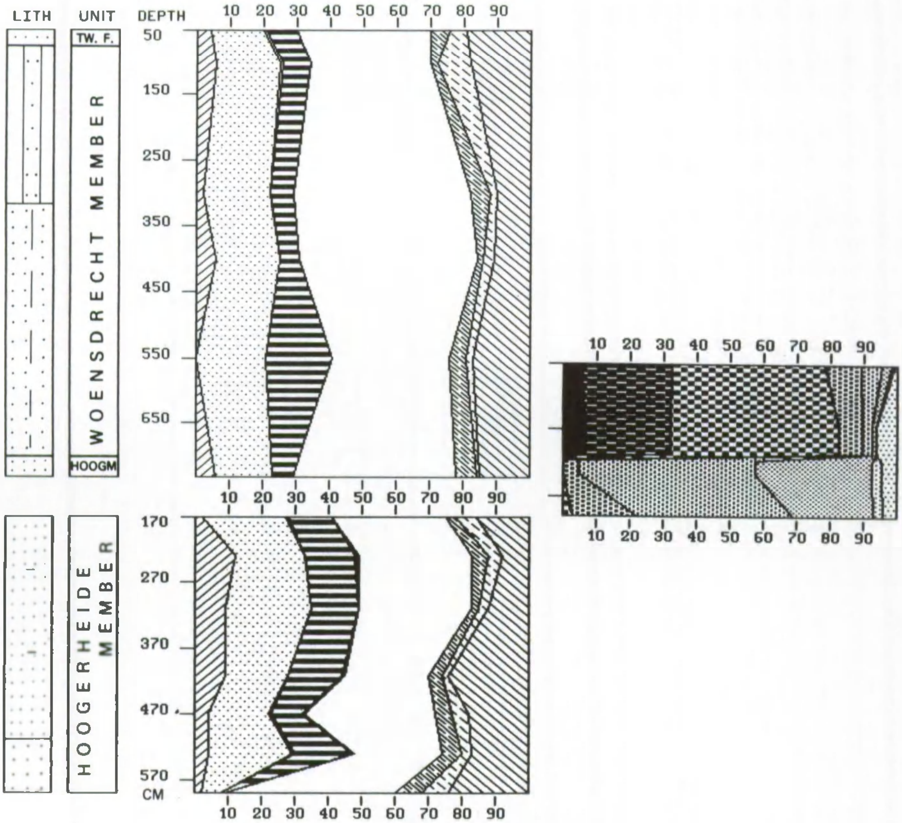


Fig. 2.18: Heavy mineral diagram of the Woensdrecht Member and Hoogerheide Member at Woensdrecht Hooghuis (legend in appendix).

Garnet (less than 10 %) and alterite content (0-2 %) are low, even in coarser sediments (which is rather unexpected, see §5.3). Staurolite and metamorphic minerals do not exceed 10-15 %. Glaucophane is present, but extremely scarce.
Differences in heavy mineral composition occur between lithofacies 1 (fig. 2.18: less than 40% unstable heavy minerals) and lithofacies 3 (up to 65% unstable heavy minerals)(Huyzer and Van Toor, 1986), which are due to different sediment provenances of the coarser and finer sediment particles (§5.3).

Nelson and Van der Hammen (1950) found 80-95 % stable heavy minerals (B-Limburg association) in the Woensdrecht Member. However, most of their analyses concern sands, which is probably the reason for the high content of stable heavy minerals mentioned by them.
The heavy mineral composition of the Woensdrecht Member changes to the north and east (Turnhout Member; Verbraeck and Bisschops, 1971: boring Klundert; Van Staalduinen, 1979: boring Maassluis). The heavy mineral assemblage of the overlying Twente Formation is comparable to the Woensdrecht Member, but rounding of the grains (especially epidote, hornblende) is much better and garnet (alterite) content is higher at the base of the Twente Formation (see App. Ossendrecht).

5. References

Nelson and Van der Hammen (1950): II-0 deposits.
Van Dorsser (1956): II-0 sand- and loam-beds.
Van Voorthuysen (1957): Halsteren deposits.
Zagwijn and Van Staalduinen (1975): Tegelen Formation.

2.5.8 Gilze Member

1. Stratotype

- Location: Gilze (clay-pit and boring)(fig. 2.19).
  x = 4°56'16''EL
  y = 51°31'45''NL
  H = 16 m +N.A.P.
  D = below 15 m +N.A.P. Base of the member was not reached, but is normally found at approximately 5 m +N.A.P. (fig. 2.7, 2.8).
- Lithology: very fine to medium coarse, brownish-gray, non-calcareous sand, fining-upwards into greenish-gray, stiff, clay-beds, with three intercalated humic to peaty soil horizons.
- Heavy minerals: predominantly stable with zircon, rutile, staurolite, metamorphic minerals and tourmaline.
- Boundaries: lower boundary strongly erosive (below 1 m -N.A.P.) to the Turnhout Member; upper boundary erosive to the Twente Formation.

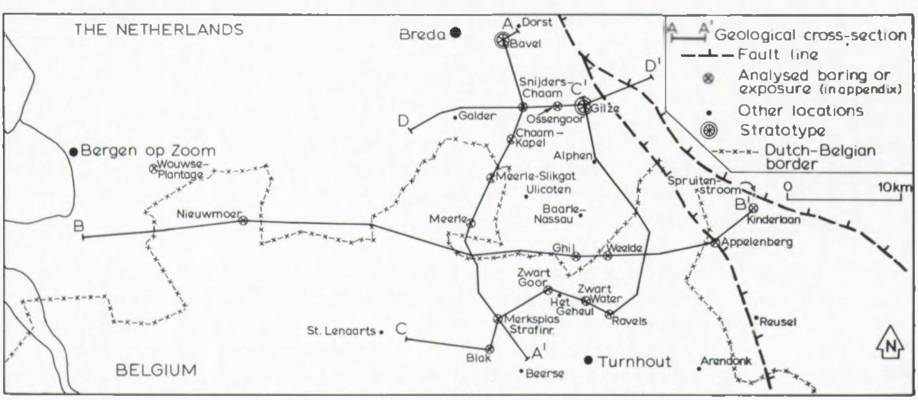


Fig. 2.19: Location map of the Gilze and Bavel Members.

2. Regional aspects

The member is important east of the Mark valley, where it occurs in a high topographical position (Baarle-Nassau plateau)(fig. 2.6, 2.8). The southwestern limit of continuous presence is formed by the villages

Turnhout, Het Geheul, Ulicoten, Meerle, Galder. The eastern limit of occurrences close to the surface coincides roughly with the western fault system of the Central Graben (Arendonk-Reusel, Poppel, Alphen, Gilze, Bavel-Dorst). Isolated patches of the Gilze Member are present west of the Mark-Weerijs basin in the neighbourhood of Nieuwmoer and the Wouwse Plantage (fig. 2.6) and on the Campine microcuesta (Beerse Blak, Merksplas Strafinrichting).

The Gilze Member overlies clays and sands of the Turnhout Member. The contact can be gradual (Appelenberg, Zwart Water) or erosive (Zwart Goor, Ghil). The Gilze Member is locally overlain by the Bavel Member (fig. 2.5). More commonly the Gilze Member is covered by eolian sands of the Twente Formation in the western part of the area or by the Sterksel Formation in the eastern part (fig. 2.6). The St. Lenaarts Formation in Belgium, which was found in the neighbourhood of St. Lenaarts and Ravels (De Ploey, 1961), can be correlated stratigraphically with the Gilze Member in The Netherlands (fig. 2.7). The latter member is overlain by the Sterksel Formation in boring Appelenberg (fig. 2.6) and therefore the St. Lenaarts Formation is part of the Early-Pleistocene deposits.

3. General lithologic characteristics

The Gilze Member is characterized by greenish or brownish-gray, very fine to medium coarse, poorly to well-sorted sands with isolated clay- and peat-beds (fig. 2.5, 2.6, 2.7, 2.8). Reworked organic material is common in the coarser grain-size fractions (Zwart Goor), but gravel is generally scarce. In a northward direction the amount and thickness of clay-layers seem to increase (fig. 2.5, 2.7).

Three sub-units could be distinguished in the Gilze Member. The lower sub-unit 1 (Appelenberg Sands) of the Gilze Member generally consists of very fine sand, with intercalated clay- and peat-layers (fig. 2.6: Appelenberg). The boundaries between the sand- and clay/peat-beds are often abrupt (fig. 2.5, 2.6: Appelenberg, Chaam Kapel, Meerle Slikgat). Fining-upward sequences are poorly developed in these situations.

Sub-unit 2 (Gilze Clay) consists of very fine to medium fine, moderately well-sorted sands, with clay- and peat-beds at the top (fig. 2.6: Nieuwmoer; fig. 2.7: Gilze). The clay-layer at the top of this fining-upward sequence can be continued over a large area (fig. 2.7, 2.8).

The upper sub-unit 3 of the Gilze Member is coarser grained (150-300 µm), with reworked plant remains, peat-clasts and locally fine gravel (fig. 2.6: Weelde, HV5; App. Ravels). This sub-unit is found in gullies, which eroded into the lower part of the Gilze Member. These channel sediments were described previously as "Alphen Sands" by Vandenberghe and Krook (1981). At the top of this sub-unit a clay-layer (Spruitenstroom Clay bed) is present locally in the east (fig. 2.6: Kinderlaan) (Van Huissteden and Van der Valk, unpubl.).

4. Heavy minerals and gravel

The Gilze Member is characterized by a stable heavy mineral association (B-Limburg association) with more than 90 % stable heavy minerals (fig. 2.15). Local variations in heavy mineral composition can be explained by grain-size density effects (§5.3). These grain-size heavy mineral relations are illustrated in fining-up and coarsening-upward sequences (App. Zwart Goor, Nieuwmoer). Oscillations in stable heavy mineral content not connected to grain-size effects occur as well (App. Gilze: 10.3 m). These variations possibly reflect changes in sediment source area, but it is yet impossible to separate the Gilze Member in different heavy mineral zones.

The stable heavy mineral association distinguishes the Gilze Member from under- and overlying units. The Turnhout and Bavel Members and the Sterksel Formation have an unstable heavy mineral assemblage, while the

Middle and Late-Pleistocene Eindhoven and Twente Formations are characterized by a mixed heavy mineral composition (Vandenberghe and Krook, 1985). The stable heavy mineral association of the St. Lenaarts Formation (De Ploey, 1961), is completely comparable to the one of the Gilze Member.
In the northern part of the investigation area the stable heavy mineral assemblage of the Gilze Member changes (fig. 2.5: Snijders Chaam; fig. 2.8: Gilze). Peaks of unstable heavy minerals occur, which are grain-size independent, because stable and unstable heavy mineral maxima are present in the same clay-bed (Snijders-Chaam: 1.68 m)(so-called abnormal variations: Edelman, 1933; Baak, 1936).
The heavy mineral composition also changes towards the top of the Gilze Member (Ossengoor, Weelde, Kinderlaan). Unstable heavy mineral content increases to 20-40% and even to 50-70% in the Spruitenstroom Clay bed, probably reflecting a "gradual" change in source area during deposition of the top strata of the Gilze Member. This Spruitenstroom Clay bed forms the transition between the Gilze Member dominated by stable heavy minerals and the Bavel Member/Sterksel Formation, which are characterized by unstable heavy minerals.
Gravel composition of the Gilze Member was described previously by Vandenberghe and Krook (1981) and Vandenberghe et al., (1986) (see §5.4.5).

5. References

Zonneveld (1947): "lower fine deposits".
Doppert and Zonneveld (1955): Kedichem Series (cross-section II).
De Ploey (1961): St. Lenaarts Formation.
Zagwijn and Van Staalduinen (1975): Kedichem Formation.
Geys (1975): part of the Campine Clay Formation (e.g. exposure Merksplas Pampa, Wortel Kolonie).
Vandenberghe and Krook (1981): Alphen Sands.
Haest (1985): Het Geheul Sands; Massief van Weelde Sands; Ravels Loam.

2.5.9 Bavel Member

1. Stratotype

- Location: Bavel (clay-pit)(fig. 2.19).
  x = 4°50'14''EL
  y = 51°35'00''NL
  H = 4 m +N.A.P.
  D = 3 m + N.A.P. to 6 m -N.A.P. (base not exposed).
- Lithology: Lithofacies 1: gray, plastic, very calcareous, fine sandy clay (15-50 % lutum) with sand-beds and -laminations at certain levels (see App.). Peaty (soil) horizons, reworked organic material and shells were not found.

Lithofacies 2: gray-brown, very fine to medium fine, well-sorted, calcareous sand with many coloured grains. Internal erosional surfaces (lag deposits) are marked by clay-pebbles, gravel and much organic material (wood fragments up to 1 m). The sand fills gully structures eroded in lithofacies 1. The top sediment is finer grained with loam-lenses (fining-upward).
- Heavy minerals (fig. 2.20): unstable heavy minerals, dominated by garnet, epidote, and hornblende (75-85%). Alterite is of less importance. Local fluctuations are explained by grain-size density effects (decreasing alterite content together with decreasing grain-size in the top of the clay: lithofacies 1).
- Gravel composition (see §5.4.6). Zandstra (1969) has chosen the Beuningen gravel-bed overlying the Bavel Member in a former pit as the

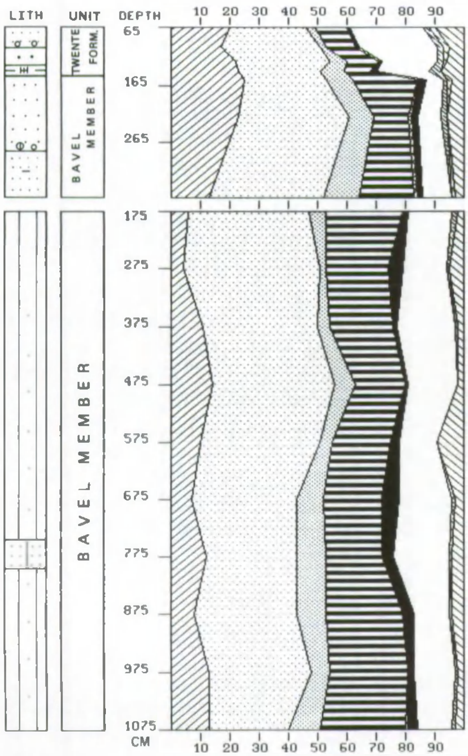


Fig. 2.20: Heavy mineral diagram of the Bavel Member at Bavel (legend in appendix).

"type-locality" for the Scheldt gravel association, because of its high flint content. Although flint content in his sample is higher (43.3%) than in ours (27%) it is lower than the flint percentages further south and west of Bavel (see §5.4.8: Meerle).

2. Regional aspects:

The Bavel Member was only found in the neighbourhood of Bavel (fig. 2.5). This location is part of the westernmost occurrences of the Bavel Member (Zagwijn and De Jong, 1984). The western boundary probably coincides roughly with the western faults of the Central Graben (Gilze-Rijen/Hooge Mierde fault). North of Gilze the Bavel Member also occurs west of the Gilze-Rijen fault.
The Bavel Member overlies fine sands and clays with a stable heavy mineral association belonging to the Gilze Member (fig. 2.5). The large lithological, sediment-petrographical and palynological differences between the members involve an erosional boundary. The Bavel Member is overlain by loamy sands of the Twente Formation.

3. References

Burck (1953): Boring Oosterhout (11.20-18.30 m).
Geys (1975): Campine Clay Formation (exposure Bavel).
Zagwijn and Van Staalduinen (1975): Kedichem Formation, Sterksel Formation.
Zagwijn en De Jong (1984): Bavelian Interglacial.

2.5.10 Middle- and Late-Pleistocene deposits

2.5.10.1 Sterksel Formation

The Sterksel Formation is found especially east of the Hooge Mierde and Gilze-Rijen fault (fig. 2.6, 2.8). It overlies the Gilze Member and is overlain by the Eindhoven and/or Twente Formations. The lithofacies consists of medium fine to coarse, non-calcareous, moderate to well-sorted sands, with some clay-pebbles and fine gravel (Appelenberg). Clay-beds are scarce.
The Sterksel Formation is dominated by unstable heavy minerals. Garnet, epidote and especially alterite are very important (fig. 2.15). Hornblende does not reach high values. Garnet content decreases sharply at 5.5 m below the surface. The gravel composition of the Sterksel Forma-

tion at Appelenberg is described in §5.4.7.

2.5.10.2 Eindhoven Formation

The Eindhoven Fomation was probably found in the drainage basin of the Mark and Weerijs rivers (fig. 2.6, 2.8). A close connection of the Eindhoven Formation to the (sub)recent river courses does not seem to exist. Boring Achtmaal (fig. 2.6) for instance, is situated on the interfluve of the Weerijs and Wildertse Beek.
The Eindhoven Formation overlies the Turnhout or Gilze Members (fig. 2.6: Achtmaal, 3 m erosion). The Eindhoven Formation is covered by the Twente Formation. Distinction between the two latter formations is not always easy as they resemble each other lithologically and sediment-petrographically (Achtmaal, Bolk; Zagwijn and Van Staalduinen, 1975: Nuenen Group). In other cases the boundary between the two formations is formed by a fine gravel-layer (fig. 2.5: Wortel). This gravel-bed is lying discordantly on the Eindhoven Formation, witnessing erosion prior to deposition of the Twente Formation.
The Eindhoven Formation consists predominantly of very fine to medium fine, moderately sorted, loamy, greenish-gray or greenish-yellow sands, with locally medium coarse sands, clay-pebbles and fine gravel at the base. Silt content generally increases upwards, forming a fining-upwards sequence with loam- and locally peat-layers at the top (fig. 2.5: Bolk).
The Eindhoven Formation is characterized sediment-petrographically by a mixed association with rather large variations (30-70% unstable heavy minerals). This distinguishes the Eindhoven Formation from the underlying Gilze Member (stable heavy minerals) and Turnhout Member (unstable heavy minerals). Furthermore, the unstable heavy minerals of the Turnhout Member are always angular, while the Eindhoven Formation contains more subrounded grains of epidote and hornblende. The overlying Twente Formation has a mixed heavy mineral assemblage too, but rounding and the percentage of well-rounded grains seems to be better and higher in the Twente Formation (no quantitative data). The coarser, fluvial sediment at the base of the Eindhoven Formation contains more stable heavy minerals (30-40%: Achtmaal, Bolk), because sediment was reworked from the underlying Gilze Member, which contains a stable heavy mineral association (Vandenberghe and Krook, 1985). The higher, unstable heavy mineral content in the finer grained upper part (fining-upward) (40-60%: Achtmaal, Wernhout Maalbergen, Bolk) is probably caused by the incorporation of an unstable eolian component, with rounded epidote and hornblende.
The gravel assemblage of the Eindhoven Formation is described in §5.4.8.

2.5.10.3 Twente Formation

The Twente Formation is found as a more or less continuous layer (1-2 m) (fig. 2.5, 2.6). The lower boundary of the formation is often formed by a gravel-layer (Beuningen gravel-bed), discordantly overlying older members.
On the interfluves two eolian lithofacies were frequently found above the Beuningen gravel-bed (Vandenberghe, 1983, 1985: Meerle Member). Lithofacies 1 consists of yellowish-brown, well-sorted, loamy, very fine sand with intercalated finer grained (loam-) laminae. On high topographical positions (fig. 2.6: Weelde; Witte Bergen) the loamy laminae are less expressed. Lithofacies 2 often overlies lithofacies 1 and is characterized by yellowish-brown, moderately well-sorted, medium

fine sands (Ossendrecht)(Huyzer and Van Toor, 1986). Lithofacies 2 is especially found along the eastern bank of the Scheldt valley (fig. 2.6) in southwest-northeast oriented, parabolic dunes (Meys, 1974).
Medium fine, fluvial sediments were found in a small river valley at Kalmthoutse Hoek. Fine-grained eolian sand- and loam-beds below the Beuningen gravel-bed occur at Merksplas Straf., Beerse Dakt and Beerse Blak (see app.). These fluvial and eolian sediments can be correlated respectively with the Blaak and Goirle Members (Vandenberghe, 1983, 1985).
The Twente Formation is characterized sediment-petrographically by a mixed heavy mineral association. This mixed composition is largely independent from underlying members (see App. Wortel, Zwart Water, Beerse Dakt), but minor regional variations exist, which are related to the subsoil. The Twente Formation overlying the Gilze Member in Ravels, Zwart Water and Zwart Goor contains 40-50% unstable minerals, while the Twente Formation overlying the Sterksel Formation and Bavel Member in Appelenberg and Bavel contains 60-70% unstable heavy minerals. Augite (diopside) is present sporadically, but was never found in older members below. Augite occurs frequently in the Kreftenheye and Urk Formations, approximately 10 km north of Breda (Zagwijn and Van Staalduinen, 1975). Towards the top of the formation (lithofacies 2 in Ossendrecht) the heavy mineral composition gradually becomes dominated by stable heavy minerals and is almost identical to the underlying Woensdrecht and Hoogerheide Members (App. Ossendrecht; Nelson and Van der Hammen, 1950). It is concluded that both local and regional sources were important in the heavy mineral composition of the Twente Formation (§5.4.8). The gravel assemblage of the Twente Formation (Beuningen gravel-bed) is described in §5.4.8.

2.6. Correlation with the regional lithostratigraphic units in The Netherlands and Belgium

In this paragraph the characteristics of the lithostratigraphic units are summarized and the members are integrated within the regional lithostratigraphic framework of The Netherlands (Zagwijn and Van Staalduinen, 1975) and Belgium (Paepe and Vanhoorne, 1976) (table 2.1).
The Merksplas Member consists of medium fine to coarse, non-calcareous sands. This member is part of the Mol Sands and Gravel Formation in Belgium and the Kieseloölite Formation in The Netherlands. The Merksplas Member dips to the north (2.3°/₀₀) and changes laterally into the shell-bearing Maassluis Formation around Breda (fig. 2.5, 2.7).
The Merksplas Member is overlain by the Rijkevorsel Member. The base of the Rijkevorsel Member is often formed by a clay-layer, resting directly on the Merksplas Member. Locally a peat or peaty layer occurs at this transition which reflects a hiatus in the clastic sedimentation (soil-formation). The Rijkevorsel Member and the overlying Beerse and Turnhout Members are grouped together into the Campine Clay and Sand Formation in Belgium and the Tegelen Formation in The Netherlands. The genetic base of the lithostratigraphic subdivision in The Netherlands presents some problems. The Tegelen Formation is defined as a fluviatile unit; the Maassluis Formation as a marine shell-bearing unit (Zagwijn and Van Staalduinen, 1975). Transitional (tidal flat, estuarine, lagoonal) deposits (like the Rijkevorsel and Turnhout Members: see chapter 3) can be incorporated either in the Tegelen Formation (non-marine deposits)(Zagwijn and Van Staalduinen, 1975) or in the Maassluis Formation (Van Staalduinen et al., 1979, fig. 24). In the present study the perimarine deposits of the Rijkevorsel and Turnhout

Members are included in the Tegelen Formation, since the lithological characteristics of the two members correspond more to the Tegelen than to the Maassluis Formation. It is shown that the top of the Turnhout Member in Belgium correlates lithologically, geometrically and palynologically (§4.3.1) with the top of the Tegelen Formation in The Netherlands (fig. 2.5). The Turnhout Member is not a part of the Kedichem Formation (as was suggested by Paepe and Vanhoorne, 1970); in its lithological and sediment-petrographical characteristics it strongly resembles the Tegelen Formation and is different from the Kedichem Formation (Gilze Member) in this area.

The Rijkevorsel and Turnhout Members are characterized by micaceous, extremely fine to medium fine, non-calcareous sands and blue-gray clays, dominated by unstable heavy minerals. The Beerse Member consists of very fine to medium fine sands with stable heavy minerals, which do not correspond to the characteristics of the Rijkevorsel and Turnhout Members. However, the Beerse Member is included in the Tegelen Formation because of its limited thickness and restricted areal distribution. It is only found in the southeastern part of the area and was elsewhere probably eroded prior to deposition of the Turnhout Member (fig. 2.5, 2.7).

The sediments of the Rijkevorsel and Turnhout Members become coarser to the north. The unstable heavy mineral content increases rapidly from south to north. In the region south of Breda the Rijkevorsel Member becomes calcareous and locally contains marine shell fragments (fig. 2.5, 2.8). The presence of marine shells is diagnostic for the Maassluis Formation (Zagwijn and Van Staalduinen, 1975). The contact between calcareous and non-calcareous sediments in fig. 2.5 intersects the lithostratigraphic boundaries of the members and probably illustrates the Pleistocene decalcification depth. Since it causes the disappearance of marine shells, this decalcification can involve the transition from the Maassluis Formation into the Tegelen Formation, according to the definition established by Zagwijn and Van Staalduinen.

The Rijkevorsel and Turnhout Members are laterally equivalent to respectively, the Hoogerheide and Woensdrecht Members in the west. Both Hoogerheide and Woensdrecht Members are regarded as part of the Tegelen Formation. The clay-layer at the top of the Turnhout and Woensdrecht Members, which is characterized by an unstable heavy mineral association and by the presence of *Azolla tegeliensis*, connects the two members (Tegelen clay)(fig. 2.3, 2.6). The Hoogerheide and Woensdrecht Members are coarser grained and are characterized by a stable to mixed, occasionally unstable heavy mineral association. The stable heavy mineral association (B-Limburg) is a mineralogical variation within the Tegelen Formation. The ideas of Doppert and Zonneveld (1955) and Van Dorsser (1956), who included part of the Early-Pleistocene deposits in western Noord-Brabant into the "Kedichem Series" because of their stable heavy mineral content, must be adjusted. The Kedichem Formation in western Noord-Brabant is present only in isolated erosion remnants (e.g. Wouwse Plantage) or channels (Nieuwmoer).

The Woensdrecht and Turnhout Members of the Tegelen Formation are covered by the Gilze Member. Sediments belonging to this member are found close to the surface on the so-called Baarle-Nassau plateau and west of the Mark river in isolated channels. The Gilze Member consists of very fine to medium coarse sand with clay- and peat-layers and is dominated by stable heavy minerals. North of Chaam and Gilze more unstable heavy minerals occur within the sediment, reflecting the interference of two sediment-petrological provinces. The lithological and sediment-petrographical characteristics of the Gilze Member are typical for the Kedichem Formation (Zagwijn and Van Staalduinen, 1975).

The Gilze Member extends southward into Belgium (Weelde-Ravels)(fig. 2.5, 2.7). The local "Het Geheul Sands", "Massief van Weelde Sands", "Ravels Loam" (Haest, 1985) and the St. Lenaarts Formation (De Ploey, 1961), which were formerly interpreted as periglacial deposits of Early/Middle-Weichselian age, are part of the Gilze Member (Kedichem Formation), because of their lithostratigraphical position and stable heavy mineral composition.
The Gilze Member of the Kedichem Formation is succeeded by the Bavel Member, which is only found in the northeastern part of the investigated area. In contrast to the Gilze Member the Bavel Member is dominated by unstable heavy minerals, which in this region is characteristic for the Sterksel Formation. However, fine-grained sand and claybeds like in the Bavel Member are more commonly found in the Kedichem Formation. In accordance with Zagwijn and De Jong (1984) the Bavel Member is included in the upper part of the Kedichem Formation on lithological grounds.
The occurrence of the Sterksel Formation is determined by the western boundary faults of the Central Graben (fig. 2.6, 2.8). The formation is characterized by medium fine to coarse sands, dominated by unstable heavy minerals. In Belgium, corresponding sediments of the Sterksel Formation are found east of Weelde-Arendonk ("Kruisberg Sands": Haest, 1985). They extend to the east in the Campine High Terrace, where they belong to the Meuse Gravel Formation (Paepe and Vanhoorne, 1976).
The Eindhoven Formation is found in the Mark-Weerijs river basin, eroding the Turnhout and/or Gilze Member (fig. 2.6). The formation is characterized by very fine to medium fine, loamy sands and a mixed heavy mineral association, with well-rounded unstable heavy minerals. The fluviatile deposits in the so-called "Breda Valley", which were found by Van Oosten (1967) in the Mark-Weerijs drainage basin, are probably partly equivalent to the Eindhoven Formation, but areal distribution, lithological and sediment-petrographical characteristics are insufficiently stated. It is possible that the deposits in the "Breda Valley" consist in fact of the Eindhoven-Twente Formations and/or the Gilze Member (fig. 2.6, 2.8: Galderse Meren).
The Twente Formation is formed by a continuous cover of loamy, very fine sands (predominantly Older Coversand II)(Van der Hammen et al., 1967). The mixed heavy mineral association represents local and regional sediment sources. East of the Scheldt valley escarpment, the upper part of the Twente Formation is coarser grained (medium fine sand) (Younger Coversand II). The high stable heavy mineral content reflects supply of local sediment, reworked from the Woensdrecht and Hoogerheide Members.

3. DEPOSITIONAL ENVIRONMENTS

3.1 Introduction

Until now facies studies of the Early-Pleistocene deposits in Noord-Brabant and northern Belgium have been scarce. Most investigations concentrated on local stratigraphical, palynological and sediment-petrographical aspects. The lack of paleoenvironmental information may, in part, be explained by the absence of fauna remains in the sediments. Only some deer antlers are known from the Campine Clay in northern Belgium (Germonpré, 1983). Molluscs and diatoms have not been found in the investigated area. Furthermore sedimentological analysis of exposures has been hampered by the predominantly clayey character of the deposits in Belgium. Therefore, the published ideas concerning the depositional environment of the Early-Pleistocene deposits range between a fluviatile and a marine origin. Belgian investigators more often concluded a coastal origin, while in The Netherlands a fluviatile genesis has in general been favoured.

The aim of chapter 3 is:
1: to summarize published ideas concerning the depositional environment and to evaluate different environmental interpretations (§3.2).
2: to describe and interpret the sedimentary structures and to reconstruct the depositional environments of the lithostratigraphic units defined in chapter 2 (§3.3).
3: to characterize the depositional environments by their paleobotanical content (§3.4).

The description and analysis of sedimentary structures is based on the study of lacquer peels from exposures and boring cores (fig. 3.1). Paleoecological information, based on the analysis of pollen and macroscopic plant remains, is additionally used.

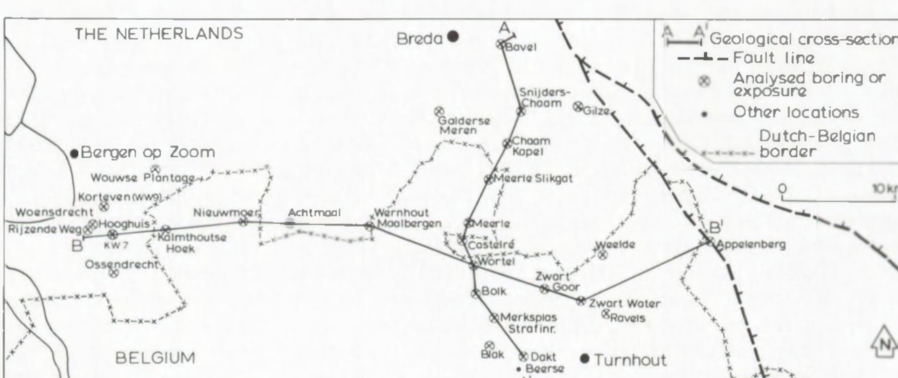


Fig. 3.1: Location map.

3.2 Historical review and discussion of former environmental interpretations

This paragraph presents a summary and discussion of articles published since 1950, in which an opinion is expressed concerning the sedimentary environments of the Early-Pleistocene deposits in Noord-Brabant and northern Belgium (see table 2.1). An extensive review of publications older than 1950 is given by Geys (1975).
In 1954 Tavernier interpreted the Mol Sands in Belgium ("Sables de Mol") as continental deposits. According to Tavernier, the overlying Campine Clays ("Argiles de la Campine") were deposited by westward flowing meandering rivers in a large alluvial plain. The clay-layers were formed in a backswamp environment, which was intersected by more sandy (gully) systems. Tavernier's interpretations were essentially based on geometrical similarities to recent alluvial plains.
One year later Doppert and Zonneveld (1955) correlated the Early-Pleistocene deposits in Noord-Brabant with the "Kedichem Series" in the central Netherlands on sediment-petrographical and lithological grounds. However, this correlation is incorrect since the deposits in Noord-Brabant belong to a large extent to the Tegelen Formation (our Turnhout-Rijkevorsel Members)(see §2.6). They stated already a priori (in the title of their publication) that the Pleistocene sediments in Noord-Brabant were deposited in a fresh water, fluviatile environment. They found that *Carya, Pterocarya* and *Tsuga* pollen was also present in deposits (so-called "Kedichem Series") younger than the Tiglian. They reinterpreted the pollen data from boring Oosterhout (Burck, 1953) and included them in the "Kedichem Series" instead of the Tiglian. This reinterpretation later proved to be only partly correct: Zagwijn and De Jong (1984) interpreted the lower pollen spectra again as Tiglian, while the upper spectra were regarded as representative of the Bavelian period.
Van Dorsser (1956) supported the ideas of Doppert and Zonneveld concerning the fluviatile origin of the Early-Pleistocene deposits in Noord-Brabant (our Woensdrecht Member), although she stated that the correlation of western and central Noord-Brabant is uncertain. Referring to earlier sediment-petrographical work by Nelson and Van der Hammen (1950), who in turn refer to Edelman (1933), she proposed a Meuse river system, which supplied stable heavy minerals to western Noord-Brabant during the Early-Pleistocene. Sands (locally homogenized) and overlying loams, in her opinion, had been deposited in braided and meandering river systems, during glacial and interglacial periods. However, the vertical and lateral relationships between the various deposits are not so clear in her work. The Early-Pleistocene sediments were regarded as one unit and they were used together in the reconstruction of the depositional environment. In our opinion it is more likely that different units are involved (see fig. 2.6). The sands with a stable heavy mineral association, which are described by Van Dorsser below a clay-layer at Hoogerheide (pit de Vijver), are part of our Woensdrecht Member. Sediments with a stable heavy mineral association above the clay-layer described at Langeschouw are probably part of our Gilze Member. Although these sediments are sediment-petrographically more or less identical they need not have been formed in the same depositional environment, nor be of the same age. The correlation from the central Netherlands to Noord-Brabant (Doppert and Zonneveld, 1955) and western Noord-Brabant (Van Dorsser, 1956) was essentially based on comparable heavy mineral sequences, which, however, are not isochronous by definition.
A fluviatile origin for the Early-Pleistocene deposits was also pro-

posed by Van Voorthuysen (1957), who studied the geology of western Noord-Brabant and the adjacent province of Zeeland. He introduced the fluviatile "Halsteren Deposits" (our Woensdrecht and Hoogerheide Members), which were found between "Icenian sands" and the Middle- to Late-Pleistocene "Vlissingen Deposits". The up to 80 m thick "Halsteren Deposits" filled a deep channel ("Zeeland Valley"), which was eroded into the underlying marine "Icenian sands" (modern Maassluis Formation). His idea of severe fluviatile erosion and deposition was later adopted by Van Rummelen (1965, 1970, 1972, 1978) in the geological survey of Zeeland.

In contrast with the ideas mentioned above, Dricot (1961) concluded that the Campine Clays around Beerse in Belgium (our Rijkevorsel and Turnhout Members) had been deposited in a salt-marsh environment. The fine sands underneath the clay were interpreted by Dricot as tidal flat deposits. The abundance of Chenopodiaceae pollen, the presence of Hystrichospheridae, Dinoflagellates, bioturbation by Polydora and heavy minerals of supposed northern origin (A-association) were used as arguments for a coastal origin of the deposits. However, his high Chenopodiaceae content of the deposits (up to 45%) was not confirmed by Greguss and Vanhoorne (1961) and Paepe and Vanhoorne (1970). Furthermore, Zonneveld (1948b) had already pointed out that an A-heavy mineral association is no proof for a marine Fenno-Scandian origin, as the Rhine supplied about the same heavy mineral association.

Intercalated between two clay-layers (our Rijkevorsel and Turnhout Members) Dricot found a sand-unit characterized by stable heavy minerals and a cold floral assemblage. He interpreted this so-called "Beersien" unit (our Beerse Member) as an eolian deposit formed in an arctic climate. De Ploey (1961), Paepe and Vanhoorne (1970) and Haest (1985) later confirmed Dricot's environmental interpretations of the Beersien sediments on palynological and granulometrical grounds.

In the same year De Ploey (1961) agreed with Dricot that the top of the Campine Clay (our Turnhout Member) is a salt-marsh deposit, underlain by tidal flat sediments with an unstable heavy mineral association. According to De Ploey, the clay-layer decreases in thickness to the north and a sandy tidal flat facies, deposited in an estuarine environment, becomes more important. De Ploey (1961) further described a sand-layer with a B-Limburg heavy mineral association (St. Lenaarts Formation), which occurred above the marsh-clay. He interpreted this sand-unit as a Weichselian, eolian deposit reworked by fluvial processes. Later Vandenberghe and Krook (1981) described a sand-layer ("Alphen Sands") with a stable heavy mineral association, overlying the Tegelen Formation in Noord-Brabant. In contrast to De Ploey they interpreted this unit as a braided river deposit of Early-Pleistocene age. However, in 1985 Haest again interpreted fluviatile sands with a stable heavy mineral association, resting on the Campine Clay at Beerse, as Early-Weichselian deposits. In the present study it is established on lithostratigraphical grounds that the St. Lenaarts Formation of De Ploey can be correlated with our Gilze Member ("Alphen Sands") of Early-Pleistocene age (fig. 2.5, 2.7).

Confusion is also possible between the "Beersien" deposits (our Beerse Member) and the St. Lenaarts Formation (our Gilze Member), since they have many sediment-petrographical and sedimentological characteristics in common, especially where the separating clay-layer (our Turnhout Member) is thin or absent. For instance, the Weichselian St. Lenaarts Formation at St. Lenaarts described by De Ploey (1961) was reinterpreted by Greguss and Vanhoorne (1961) and Paepe and Vanhoorne (1970) as the Early-Pleistocene Beerse Member.

During his pedological survey in western Noord-Brabant, Van Oosten

(1967) correlated a clay-layer (our Turnhout Member) with the salt-marsh clay-layer of De Ploey in adjacent Belgium. The occurrence of peat and wood fragments and cat-clay phenomena (acid sulphate soils) in the clay-layer and thinly laminated deposits below the clay were regarded by Van Oosten as indications of a fresh to brackish tidal environment. However, thin laminations are no proof of a tidal environment, but in combination with cat-clay phenomena his conclusions seem probable.

Paepe and Vanhoorne (1970, 1976) reinvestigated clay-pits around Beerse (Belgium) and confirmed the previous interpretations of Dricot (1961) and De Ploey (1961). The Early-Pleistocene deposits were subdivided into three lithological units. The lower clay-unit (Rijkevorsel Member) and the upper clay-unit (Turnhout Member) were interpreted as tidal flat deposits. However, the peat and vegetation remnants mentioned by Paepe and Vanhoorne (1970) are no proof of the tidal character of the Campine Clay. The regular alternations of clay and sand with opposed stream directions, which they found in gully-fill structures, are indeed a strong indication of a tidal environment. At the base of the Turnhout Member Paepe and Vanhoorne found a sand-layer with large-scale, low-angle, cross-bedding dipping to the northwest. They interpreted these structures as beach deposits, indicating the onset of the marine transgression of the Turnhout Member. They further described a sand-bed (Beerse Member) in between both clay-layers, which was characterized by soil horizons, frost wedges and cryoturbations. On sedimentological and palynological grounds they concluded that the Beerse Member was formed by eolian and fluviatile processes in a cold climate.

In 1975 and 1978 Geys returned to the previous environmental ideas of Tavernier (1954), Doppert and Zonneveld (1955) and Van Dorsser (1956). He concluded a fluviatile origin of the Early-Pleistocene deposits. His interpretation was especially based on granulometric and quartz morphoscopic methods. Geys stated that his Early-Pleistocene Campine Clay Formation was deposited by a meandering river flowing to the westnorthwest. Stream energy was low in general and limnic sedimentary environments occurred locally. He extended the fluvial sediments into the "Halsteren Deposits" filling the "Zeeland Valley" described by Van Voorthuysen (1957). According to Geys this Early-Pleistocene river system resembled the Holocene river district of the central Netherlands, which is characterized by clayey backswamp environments and sand-filled gullies. In his emphasis on granulometric data, however, he disregarded the lithostratigraphic position of the sediments. He included different lithostratigraphic units in the Campine Clay Formation. Fresh water molluscs at Bavel (our Bavel Member), granulometric data of large-scale, cross-bedded sands above the Campine Clay (in Merksplas-Pampa; Wortel-Kolonie) (our Gilze Member), together with granulometric data of the Campine Clay deposits (our Turnhout and Rijkevorsel Member) were all forced into his meandering river depositional model. It is questionable whether granulometric and morphoscopic analysis alone are able to reveal depositional environments, as most of the sediments concerned are polycyclic. Short distance reworking of older sediments need not be expressed granulometrically or morphoscopically. Besides, certain granulometric parameters and methods used by Geys differentiate only between e.g. river sands and beach deposits (Sk-sigma diagram from Friedman). They should not be applied to distinguish between other depositional environments. Estuarine sediments for instance will always belong to the river sediment group, although they are not fluviatile in a strict sense. The supposed geometrical similarity between the Early-Pleistocene deposits in Belgium (Geys: fig. 3.3.1.) and the sub-recent Holocene fluviatile deposits in the central Netherlands (Geys: fig.

3.9.1.) is misleading. The sand occurrences around Meerle (our Gilze Member) have no genetic relation to the clay deposits more to the south (our Turnhout Member; boring Wortel), since they belong to different lithostratigraphic units (fig. 2.5) and formed in different periods (Eburonian and Tiglian respectively; see chapter 4).
In the same year Zagwijn and Van Staalduinen (1975) presented a lithostratigraphic scheme of the Quaternary formations in The Netherlands. The Tegelen Formation (including our Rijkevorsel, Beerse, Turnhout, Hoogerheide and Woensdrecht Members) and the Kedichem Formation (including our Gilze and Bavel Members) were described as fluviatile deposits supplied by the Rhine, Meuse and smaller rivers from the south. The lower part of the Tegelen Formation was considered to be time-equivalent with the marine Maassluis Formation. The genetic element in this lithostratigraphic system can result in classification problems, which have been dealt with in §2.6.
Vandenberghe and Krook (1981) and Vandenberghe et al. (1986) investigated sedimentary sequences and structures in the coarse-grained "Alphen Sands" (our Gilze Member; Kedichem Formation). They concluded a braided river depositional environment. On sediment-petrographical and sedimentological grounds they suggested a southeast-northwest flowing Meuse, with smaller southwest-northeast oriented tributaries from Central Belgium.
Bisschops et al. (1985) interpreted the Tegelen Formation in eastern Noord-Brabant as a fluviatile, deltaic deposit of the Rhine and Meuse with near-coastal facies in certain intervals. Smaller rivers from the south were locally important during the deposition of the Kedichem Formation. Fluvio-periglacial and eolian periglacial environments are mentioned by them. The Sterksel Formation according to Bisschops et al. (1985) was deposited by the river Rhine, during glacial and interglacial periods.
Haest (1985) reinvestigated the Beerse Member (between our Rijkevorsel and Turnhout Members). He found five soil-horizons in the Beerse sands, of which the lower four were deformed and contained associated frost-wedges and some small ice-wedges. His granulometric analysis of the Beerse sands pointed to an eolian depositional environment, with local surficial runoff.

3.3 Sedimentary environments based on the description and interpretation of sedimentary structures

3.3.1 Introduction

In this paragraph sedimentary environments of the members defined in chapter 2 are reconstructed, based on the study of the sedimentary structures in lacquer peels and large-scale lateral and vertical facies changes in two cross-sections. Sedimentary structures are described and subsequently interpreted as bedform structures and correlated with specific sedimentary environments according to especially Reineck and Singh (1980) and Reading (1980).
The locations of the analysed borings and exposures are presented in fig. 3.1. The legend of the sedimentary structures is given in fig. 3.2.

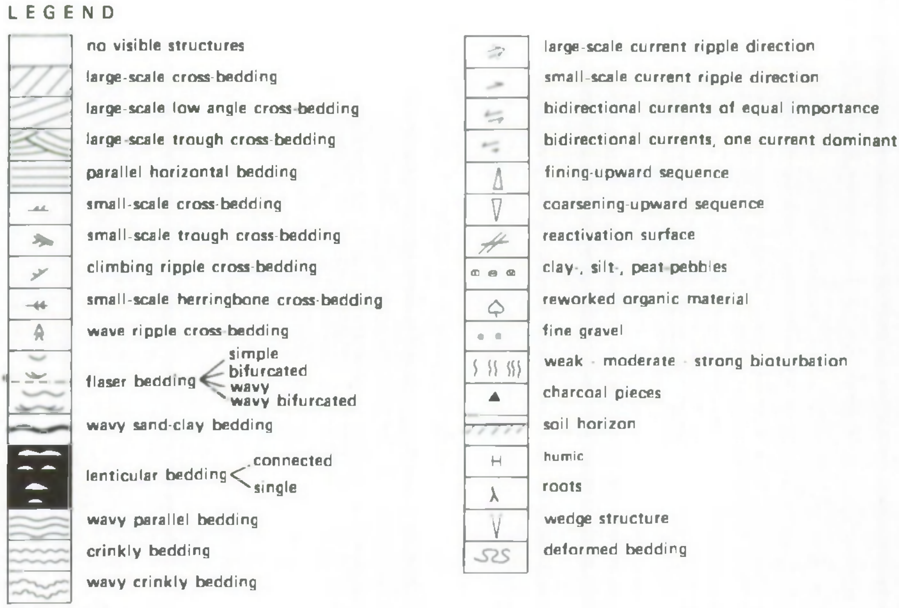


Fig. 3.2: Legend of sedimentary structures and additional symbols.

3.3.2 Rijkevorsel Member

Description of the sedimentary structures

Sedimentary structures in the Rijkevorsel Member are shown in fig. 3.3 and 3.4 in a north-south and east-west cross-section. Clay-pit Beerse Dakt is presented in more detail in fig. 3.5 and 3.6.
Four sedimentary units are distinguished within the Rijkevorsel Member.
Unit A (fig. 3.3: Bolk) is a fining-upward sequence with massive and large-scale cross-bedded, medium coarse sand, grading into flaser bedded, fine sand with some bioturbation.
Unit B (fig. 3.3: Bolk, Wortel; fig. 3.4: Wernhout Maalbergen) consists predominantly of lenticular bedding, with opposed cross-bedded sand-lenses covered by clay-drapes.
Unit C (fig. 3.3: Bolk, Wortel; fig. 3.4) is characterized by large-scale cross-bedding, flaser bedding and local lenticular bedding (fig. 3.4: Achtmaal). Around Breda marine shell-fragments (Cardium, Mytilus) are present (see fig. 2.8).
Unit D (fig. 3.3: Bolk, Meerle Slikgat, Chaam Kapel; fig. 3.5: Beerse Dakt) is formed by a fining-upward sequence. Flaser bedding grades into locally bioturbated lenticular bedding (fig. 3.3, 3.5: Beerse Dakt, Meerle Slikgat) and massive bedding with local peaty soil horizons (fig. 3.3: Merksplas Straf., Beerse Dakt).
A lacquer peel (fig. 3.5) of units (B), C, D at Beerse Dakt illustrates various sedimentary structures in the Rijkevorsel Member. Lenticular bedding (B) changes upward into flaser bedding and some low-angle cross-bedding (coarsening-up)(C). From the middle of the peel a fining-upward sequence is reflected by an increase of lenticular bedding (D).

Interpretation

The frequent sand-clay alternations with sharp transitions and the herringbone cross-bedding are indications for tidal processes. Sand was transported by ebb and flood currents, while mud was deposited during slack-water periods (Terwindt, 1981).
Unit A is interpreted as a tidal channel-fill. Medium coarse sand with clay-pebbles and reworked organic material was deposited under the form of megaripples. Stream velocity reduced by channel migration and fine sand was deposited by small current ripples. During slack-water clay-flasers covered the current ripples.
Unit B: The lenticular bedding points to quiet tidal deposition. Mud-draped ebb and flood ripples indicate a subtidal environment (lagoonal). These subtidal deposits locally overlay fluviatile sands and peat of the Merksplas Member (see fig. 2.7: boring 17E-154) indicating a sedimentary hiatus (see chapter 4: chronostratigraphy). Rapid drowning must have taken place.
Unit C: This sandy interval reflects more turbulent tidal conditions. The large-scale, low-angle cross-bedding and the flaser bedding were formed by megaripples and small current ripples in tidal channels. Coarse-grained channel sediment is rare however and clayey beds with lenticular bedding are present in unit C. This points to limited lateral channel migration, which is explained by the landward setting of the tidal depositional environment.
The absence of marine shells and the scarcity of bioturbation is an indication of a low salinity of the environment. A northward salinity increase is suggested by the presence of marine molluscs around Breda.
Unit D: The top of the Rijkevorsel Member is fining-upward. The flaser, wavy and lenticular bedding reflect the final stage silting of the previous tidal channels. Part of the channel infilling still occurred in subtidal situations (Meerle Slikgat), as both ebb and flood ripples are covered by slack-water mud.

The tidal sequence in the Rijkevorsel Member was well exposed in Beerse Dakt (fig. 3.5). The coarsening-upward sequence in the lower part of the lacquer peel reflects a gradual increase of tidal currents (transgression). The flaser and low-angle cross-bedding (in the middle) were formed in (pointbars of) small and shallow tidal channels (Reineck and Singh, 1980). The flaser and lenticular bedded couplets may have originated by seasonal cycles (Van den Berg, 1986), spring-neap cycles or variable weather and hydrological conditions.
Sub-horizontal coarser grained units in the middle of the lacquer peel, are interpreted as upper flow regime shallow water deposits. Current velocity diminished in the upper part of the lacquer peel and lenticular bedding was formed (fining-upward sequence). Ultimately, only clay was deposited, followed by plant growth.

Estimation of the tidal range

The tidal sequence at Beerse Dakt offers the opportunity to estimate the tidal range during the deposition of the Rijkevorsel Member (fig. 3.6). It is assumed that the gradual fining-upward sequence reflects the final silting of the area at a more or less constant sea-level stand. The mean low water (MLW) and mean high water (MHW) levels are fixed by the range of the sedimentary structures (fig. 3.6). MLW level is difficult to establish. The lower part of the lacquer peel was formed during a transgression (coarsening-up) with a rising sea-level.

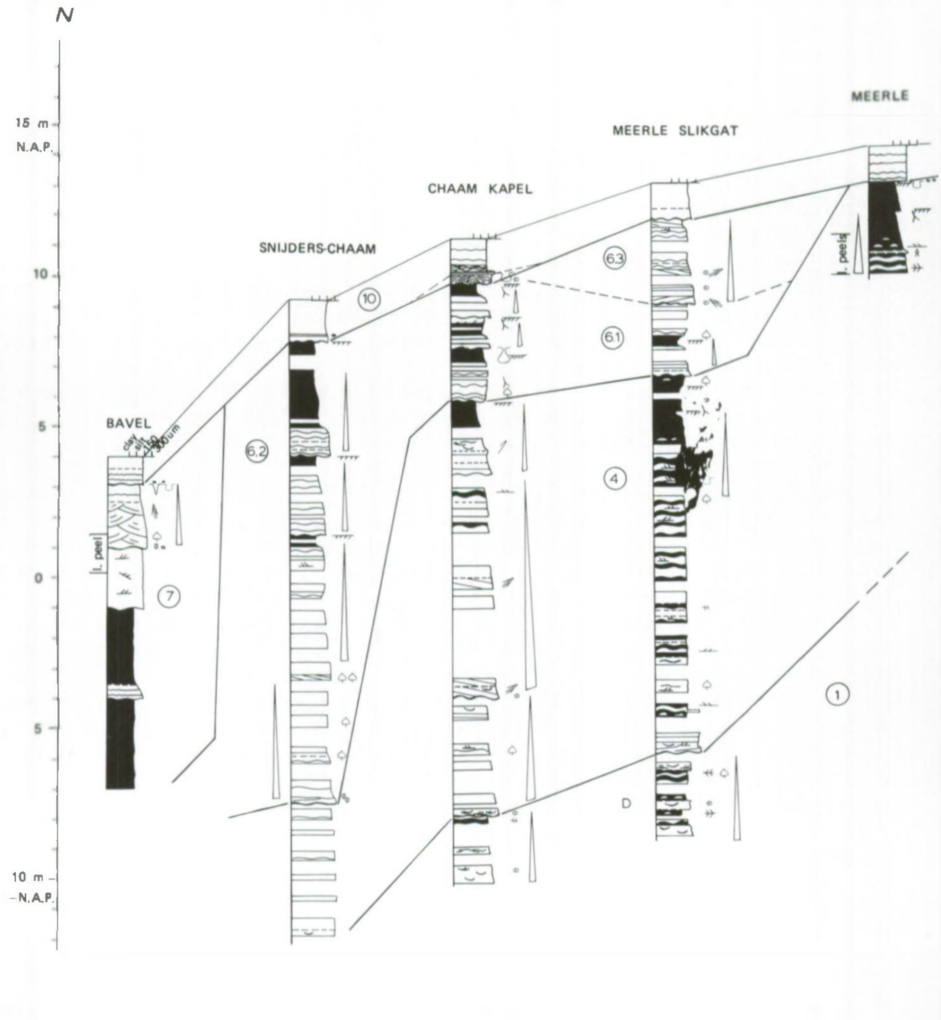
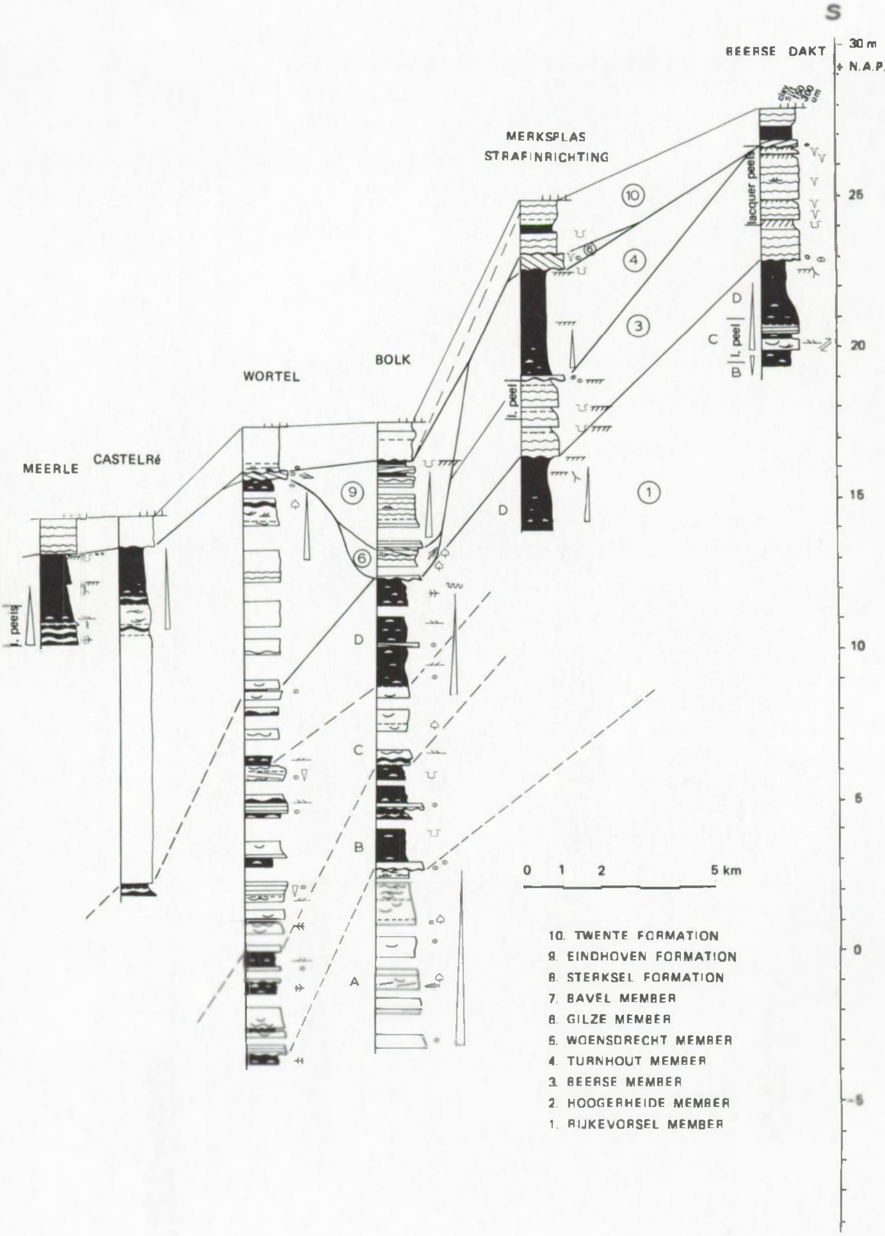


Fig. 3.3: Sedimentological cross-section between Beerse and Bavel, perpendicular to the depth contours of the North Sea basin.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BEERSE DAKT
30 m
+ N.A.P.
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WORTEL
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l. peel
0 1 2 5 km
10. TWENTE FORMATION
9. EINDHOVEN FORMATION
8. STERKSEL FORMATION
7. BAVEL MEMBER
6. GILZE MEMBER
5. WOENSDRECHT MEMBER
4. TURNHOUT MEMBER
3. BEERSE MEMBER
2. HOOGERHEIDE MEMBER
1. RIJKEVORSEL MEMBER

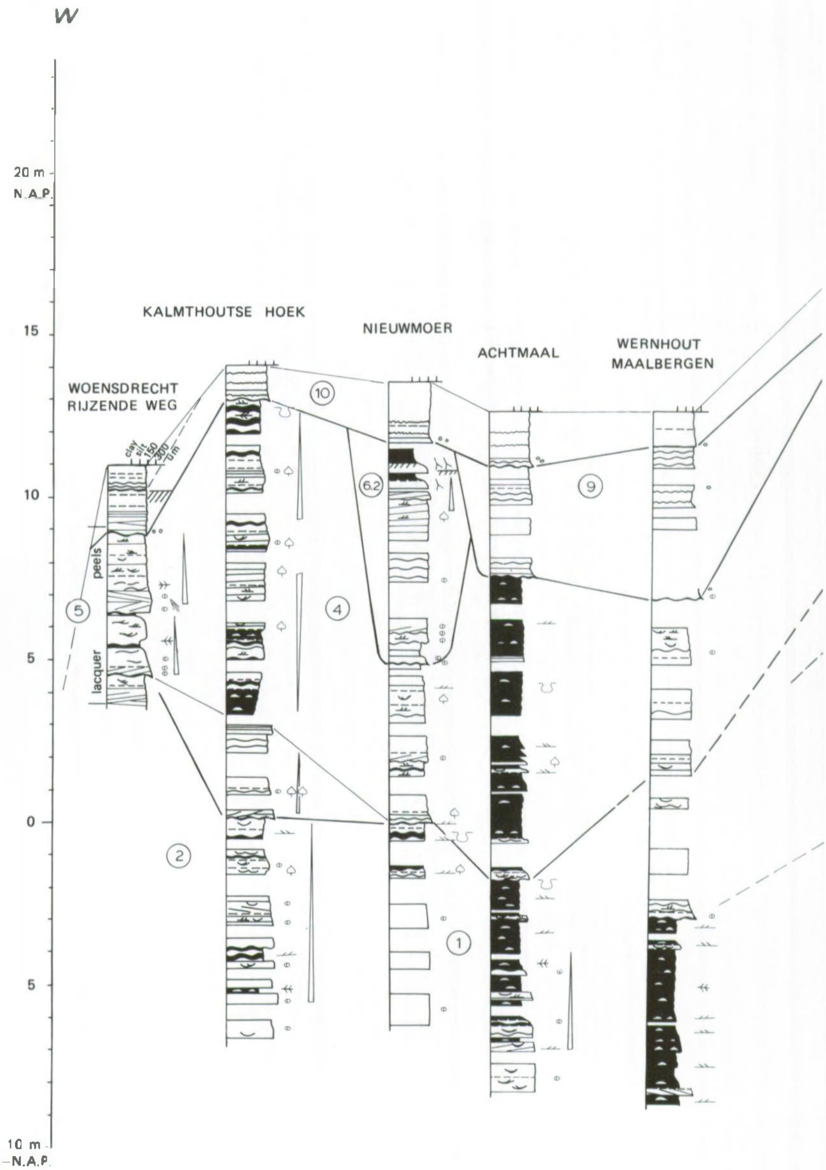
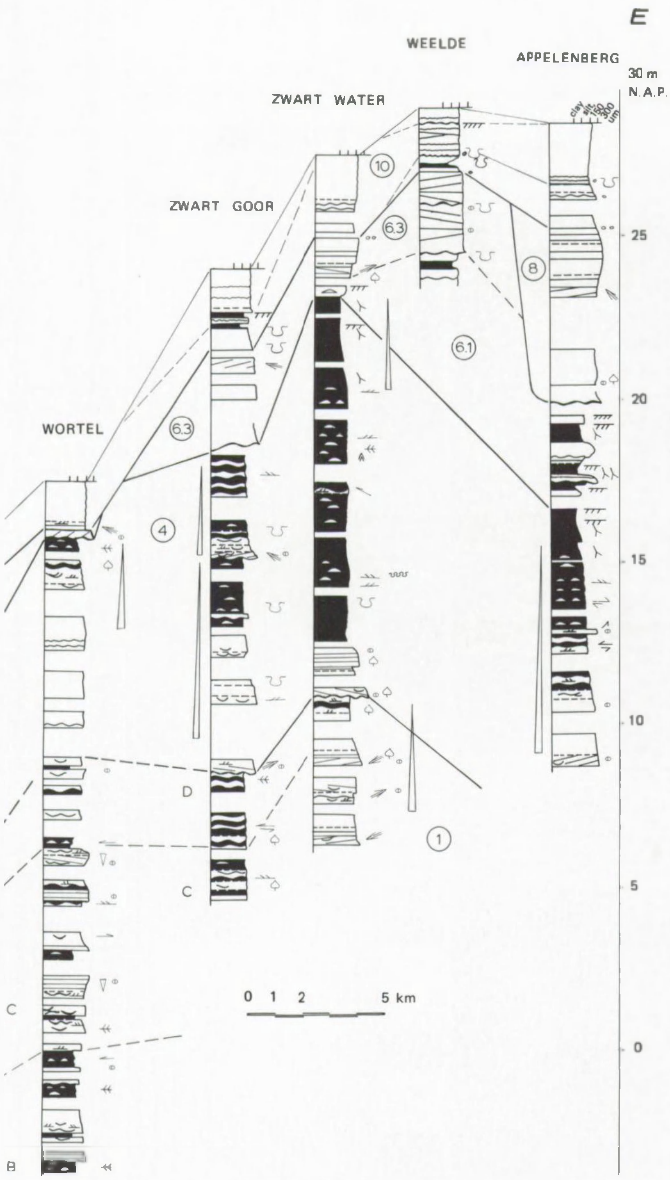


Fig. 3.4: Sedimentological cross-section between Woensdrecht and Appelenberg (Lage Mierde), parallel to the depth contours of the basin.


E
WEELDE
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30 m
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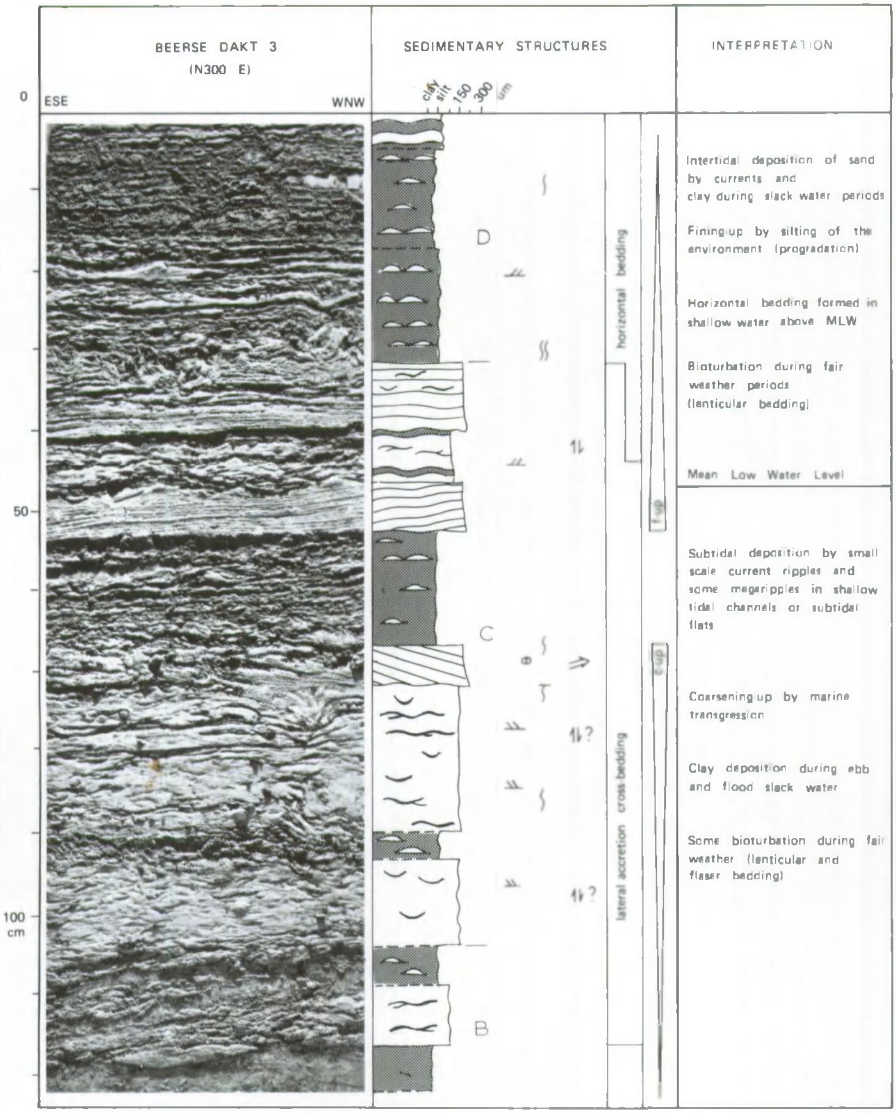


Fig. 3.5: Inshore tidal, landward facies of the Rijkevorsel Member at Beerse Dakt with weakly bioturbated, well developed lenticular and flaser bedding.

The largest sedimentological break occurs at the base of the low-angle, cross-bedded sand (fig. 3.5: at 70 cm). Large-scale cross-bedding in landward tidal environments is associated with megaripples below MLW (subtidal). Bidirectional ripples with mud-drapes (subtidal) were not found above the low-angle cross-bedding. The MLW level was probably not below the largest sedimentological break, as then the coarsening-up sequence (70-120 cm) would have been affected by erosion. A hypothetical MLW level just above the low-angle cross-bedding, is deduced from these facts.

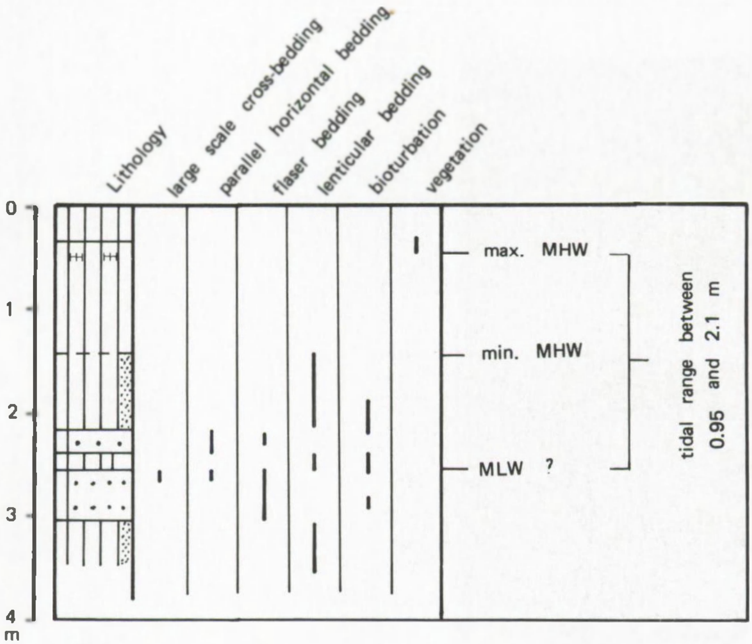


Fig. 3.6: Approximation of the tidal range during final silting of the Rijkevorsel Member at Beerse Dakt.

The MHW level can be estimated more accurately. The lenticular bedding and bioturbation in the fining-up sequence, are of subtidal or intertidal origin, below the MHW level (min. MHW). Wavy, crinkly, saltmarsh sediment, of which the base is formed 10-20 cm below MHW (Roep and Van Regteren Altena, 1988) was not found here, because of the landward depositional environment. The in situ humic soil horizon probably developed above MHW (max. MHW). When the MHW and MLW levels have been established, the tidal range at Beerse Dakt is valued between 0.95 and 2.1 m (fig. 3.6). This range is not corrected for later compaction, which according to Zonneveld (1960) can be as much as 50%.
As a conclusion, the tidal range estimation for the Rijkevorsel Member at Beerse Dakt is not very reliable, because the mean low water level is uncertain. A more detailed discussion and a more reliable tidal range estimation is given for the Turnhout Member at Meerle (§3.3.4).

3.3.3 Beerse Member

<u>Description of the sedimentary structures</u>

The Beerse Member is restricted in its occurrence to the southeastern part of the investigated area (fig. 3.3). Two lithofacies are distinguished (§2.5.4).

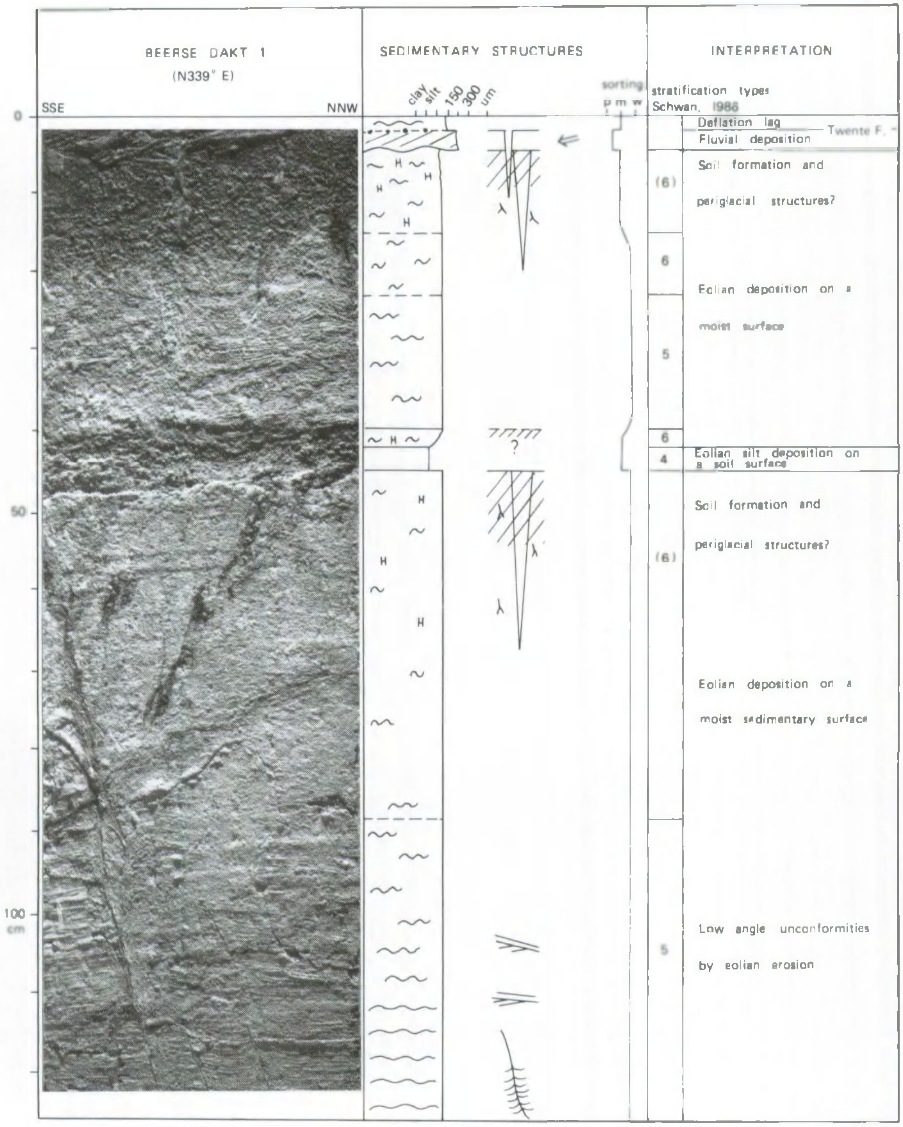


Fig. 3.7: Wet eolian deposition by adhesion ripples and two faint soil horizons in the upper part of the Beerse Member at Beerse Dakt.

Lithofacies 1 contains fine to medium sand with several humic or peaty soil horizons and periglacial structures (fig. 3.3: Beerse Dakt and Merksplas Strafinrichting). In Beerse Dakt (fig. 3.7, 3.8) the facies is characterized by continuous and discontinuous, crinkly bedding and four soil horizons. Massive bedding (on top of the paleosols), parallel horizontal bedding and small-scale cross-bedding occur infrequently. Deformations affect the lower paleosol (fig. 3.8). Frost cracks and

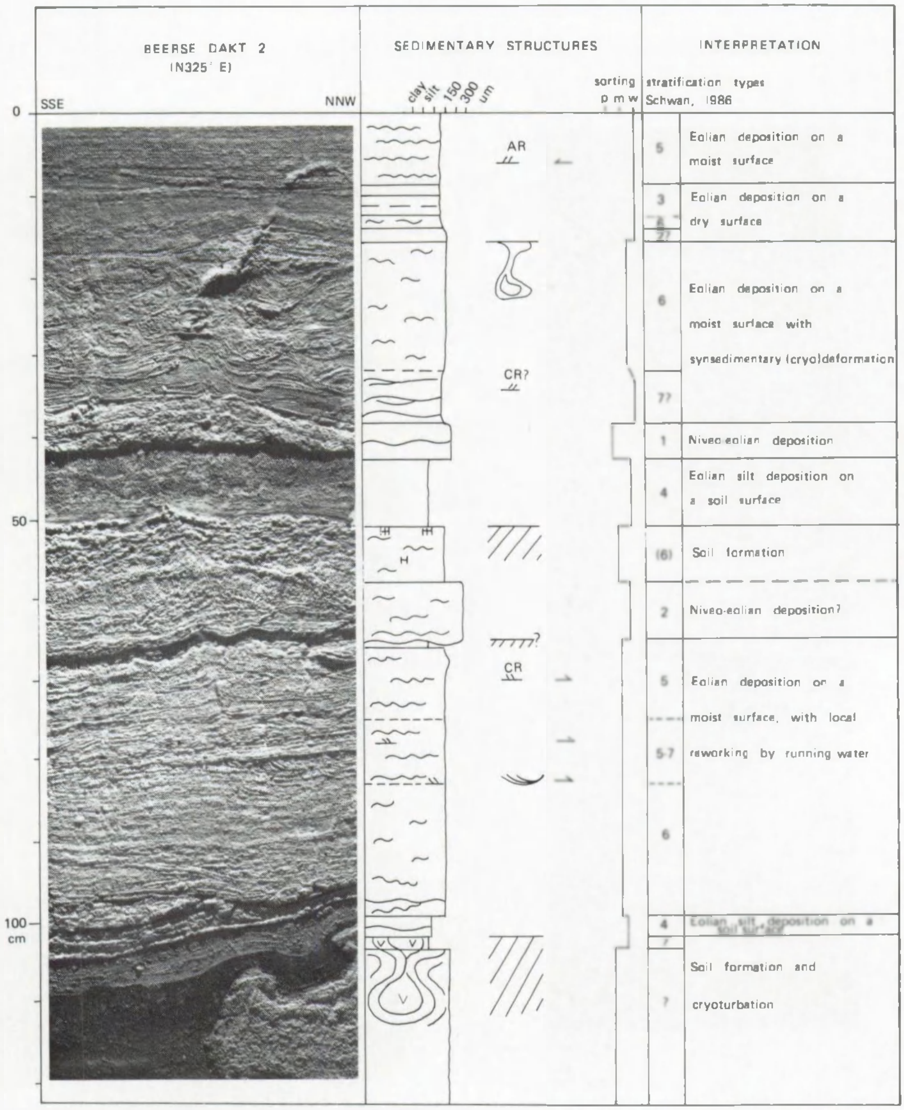


Fig. 3.8: Wet eolian deposition by adhesion ripples and local surficial runoff. The lower of the two soils has been cryoturbated. Lower part of the Beerse Member at Beerse Dakt.

small ice-wedge casts are present between the soils (see Appendix). In Merksplas Strafinrichting (fig. 3.9) lithofacies 1 shows a fining-upward sequence. Deformed, discontinuous, crinkly bedding changes upward into massive, wavy and parallel horizontal bedding, with coarse grained ripples (WR). Four paleosols are present in Merksplas Straf. (see App.), which are intensively deformed locally, with wave amplitudes up to 0.7 m. Lithofacies 2 (App. Beerse Dakt: bed 18) consists of medium to coarse sand with large-scale cross-bedding.

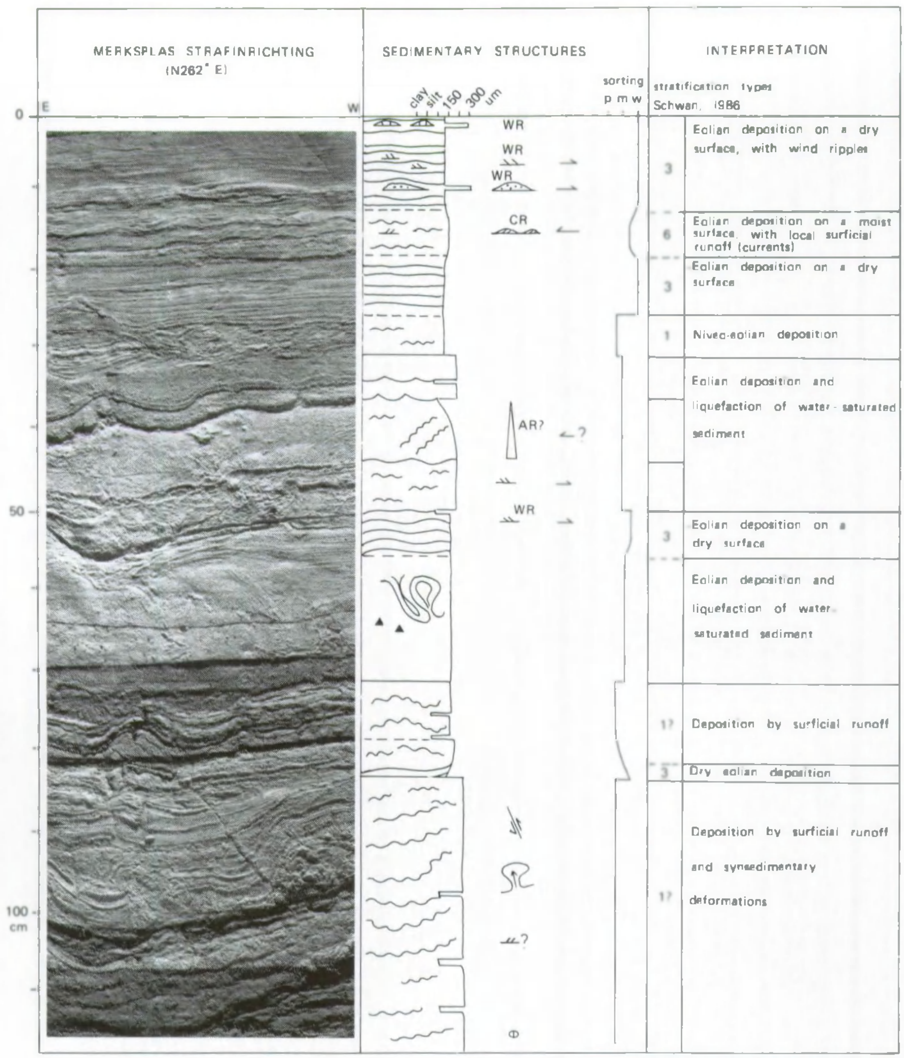


Fig. 3.9: Surficial runoff deposits changing upwards into dry eolian deposits with plane bedding. Beerse Member at Merksplas Strafinrichting.

Interpretation

The sedimentary structures in lithofacies 1 of the Beerse Member are interpreted in accordance with Schwan (1986). Several of his stratification types in Weichselian eolian coversands could also be distinguished in the Early-Pleistocene Beerse Member.
The discontinuous, crinkly bedding in Beerse Dakt (fig. 3.7, 3.8) is explained by the adhesion of windblown sand in plane beds or small adhesion ripples on a moist surface. Local reworking of these eolian

sediment by surficial runoff formed some small-scale current ripple cross-bedding (fig. 3.8 at ± 80 cm). The occurrence of parallel horizontal bedding at the top of fig. 3.8 points to eolian deposition in plane beds or by small wind ripples on a dry sedimentary surface. Periods of stagnation in the eolian sedimentation of sand are reflected by soil horizons and massive loam-beds (fig. 3.7, 3.8).
The fining-upward sequence at Merksplas Strafinrichting (fig. 3.9) reflects a synsedimentary decrease in moisture content and an increase of the eolian deposition. The deformed, discontinuous bedding in poorly to medium sorted sand is interpreted as a surficial runoff deposit. Deformations originated due to oversaturation of water in the pores between the sandgrains. Waterdepth must have been minimal, because intercalations of parallel horizontal bedding occur (interpreted as dry eolian deposits) in the surficial runoff sediments.
The massive beds in the middle of the peel (fig. 3.9) are explained by deposition of windblown sediment on a very wet surface. Periodic dryer conditions on the sedimentary surface are reflected by parallel horizontal bedding, which occur in between the liquefied beds. The parallel horizontal bedding was formed by eolian deposition of small wind ripples and plane beds on a dry surface. In the upper part of fig. 3.9 dry eolian, parallel horizontal bedding dominates. Coarse-grained (granule) wind ripples (WR) were locally preserved on the eolian surface.
The interpretations presented above confirm results of Haest (1985), who concluded to an eolian origin of the Beerse Member on granulometric grounds.

Geomorphological position

Geomorphological arguments support the eolian interpretation of the deposits. The stacked sand-units and paleosols, which are present in Merksplas Strafinrichting (App.), have all been formed in the same geomorphological situation. Humic, sandy soils are present on higher topographical positions in the eastern part of the exposure. Low lying, peaty soils occur in the west. The intervening sand-units occur as sand-sheets, with uniform thickness between the soils. Deposition by running water would have caused local erosion, flattening of the topography and infilling of low lying places. However, relief differences continued to exist during the successive sedimentation phases. Erosion hardly occurred in the receiving sites. This concordant upbuilding is explained by eolian deposition in sand-sheets (Ruegg, 1983; Schwan, 1986; Kocurek and Nielson, 1986).

Paleosols

The paleosols represent periods of non-sedimentation and surface stability, between the sand-sheet deposits. Peaty soils developed in low lying, wet places; humic, sandy soils formed on higher, dryer locations (App. Merksplas Straf.). The peaty wet soils were more subjected to cryoturbation than the dry soils, because of a higher soil moisture content.
The paleosols at Beerse Dakt (fig. 3.7, 3.8) are capped by a loam-layer. This loam is probably of eolian origin (no current ripples, no gullying). The deposition of eolian silt is explained by local surface stability (=soil formation), when sand transport was hampered by vegetation. Fall out of suspended silt formed a loam-layer on top of the soil horizon. Later the region was reached again by migrating sand-sheets and local sand transport dominated over the regional fall out of silt.

Periglacial phenomena

The Beerse Member is characterized by the presence of periglacial phenomena. Frost cracks, small ice-wedge casts (25 cm broad, 50 cm deep) and cryoturbations (50-70 cm amplitude) are present; associated with soil horizons and sand-units (fig. 3.7, 3.8, 3.9 and appendix).
The periglacial phenomena allow an estimation of the mean annual temperature. The frost cracks, cryoturbations and small ice-wedge casts point to a cool to cold climate with (local) permafrost conditions and a mean annual temperature around -5 °C (Romanovskij, 1985).
The pollen spectra from the paleosols (see chapter 4) indicate a cool climate as well. Thermophilous trees are absent and contrary to the opinion expressed by Haest (1985) there is no reason to suppose warmer conditions during soil formation.

Lithofacies 2:
The large-scale cross-bedding and coarse sand with clay-pebbles and fine gravel, point to high stream velocities and megaripple migration. Lithofacies 2 is interpreted as a fluviatile deposit.

3.3.4 Turnhout Member

The Turnhout Member is an important unit, which is present over the complete study area (fig. 3.3, 3.4). The member consists of fine to medium, micaceous sand and clay and is characterized by an unstable heavy mineral association (§2.5.5). Grain-size increases to the north and west, where the Turnhout Member is laterally connected to the Woensdrecht Member (fig. 3.3, 3.4). Vertical grain-size variations occur under the form of fining-upward sequences.

Description of the sedimentary structures (fig. 3.3, 3.4, 3.10, 3.11)

The base of the Turnhout Member was investigated at Beerse Blak (fig. 3.10; App.). Geys (1975, fig. 2.13.2) described a corresponding unit in a nearby pit, 5 to 8 m below the surface. Medium coarse sand with large-scale, tabular, unidirectional cross-bedding is laterally replaced, towards the NNW, by finer-grained sand with small-scale cross-bedding and possibly some herringbone cross-bedding. The cross-bedded sands are covered rather abruptly by clay.
Above the base of the Turnhout Member a sand- and clay-facies have been found. The sands are characterized by massive bedding, flaser bedding, wavy bedding and some large-scale cross-bedding (fig. 3.3 and 3.4). Fining-upward sequences are locally present in the sands (fig. 3.3: Chaam Kapel). The sands are laterally equivalent with thick clay/loam-beds, which are dominated by lenticular and wavy sand-clay bedding (fig. 3.3: Meerle Slikgat; fig. 3.4: Zwart Water, Achtmaal). Thick clay-beds are also found along the southern margins of the depositional environment (fig. 3.3: Merksplas Straf.; Appendix Beerse Blak). The massive or lenticular bedded clay at Merksplas Strafinrichting is found in gully structures eroded in the Beerse Member. Medium coarse sand and peat-lumps occur at the erosional contact.
A decrease in grain-size occurs towards the top of the member (fining-up sequence). Large-scale cross-bedding and flaser bedding give way to wavy sand-clay bedding, lenticular bedding and massive clay with peaty horizons (fig. 3.3: Merksplas Straf., Meerle, Meerle Slikgat; fig. 3.4: Appelenberg, Zwart Water). The top of the fining-upward sequence was studied in detail in Meerle (Kasse, 1986; fig. 3.3 and 3.11). Lenticu-

lar and flaser bedding with bidirectional, mud-draped cross-bedding are replaced upward by wave ripple cross-bedding (locally truncated)(fig. 3.11: ± at 60 cm). Lateral accretion bedding is replaced by horizontal bedding at the same level. Wavy sand-clay bedding, parallel horizontal bedding, lenticular bedding with unidirectional small-scale cross-bedding and massive clay occur in the upper part of the sequence (fig. 3.11). A peat-layer is present at 45 cm above the lacquer peels (fig. 3.3).

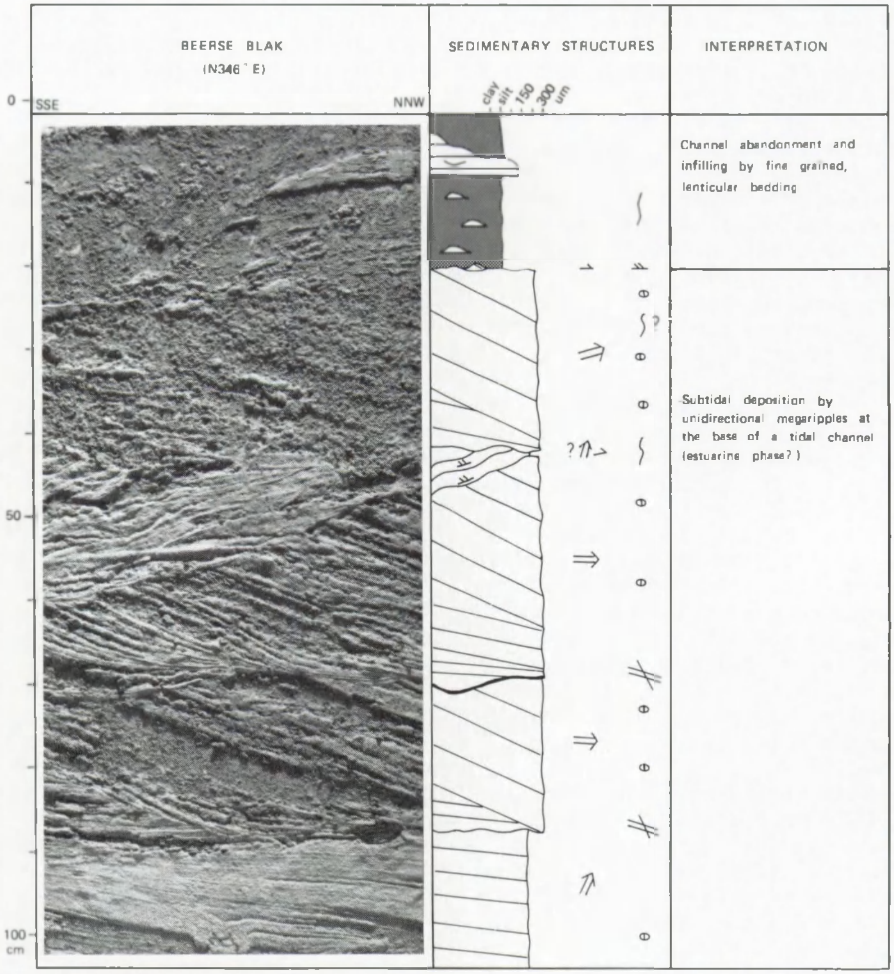


Fig. 3.10: (Sub)tidal channel-fill by unidirectional megaripples at the base of the Turnhout Member at Beerse Blak.

Interpretation

The lack of molluscs and diatoms hampers the environmental interpretation. The regular alternations of sand and clay-laminae, especially in the fining-up sequence at the top of the member, and the high-angle, opposed cross-bedding (fig. 3.3, 3.4, 3.11) are explained by tidal, ebb and flood sand transport and slack-water clay deposition (Kasse, 1986).

The medium coarse, large-scale cross-bedded sand with many clay-pebbles at the base of the Turnhout Member in Beerse Blak and Meerle Slikgat (fig. 3.3: 18-19 m) is interpreted as a megaripple deposit on the bottom of a tidal channel. It reflects an erosional phase, in which the underlying Beerse Member was partly (fig. 3.3: Merksplas Straf.) or completely eroded (fig. 3.3: Meerle Slikgat; App.: Beerse Blak). North-northwest flowing currents were dominant in Beerse Blak (fig. 3.10). Because of this erosion, the Turnhout Member only consists of a regressive, progradational, tidal sequence. A transgressive, vertical sequence (e.g. substratum, peat, fresh water lagoonal clay, salt water tidal deposits; Reineck, 1970) has not been found.
After this erosional phase the depositional environment changed rather abruptly. The tidal channels at Beerse Blak and Meerle Slikgat were abandonned and silted up with clay (fig. 3.3, 3.10). The environment became more quiet and sheltered (perhaps locally lagoonal). A progradational tidal sequence was formed (fining-upward) by seaward migration of the tidal environments (tidal channel sand, tidal flat sand and clay, marsh/peat)(fig. 3.3: Chaam Kapel, Meerle Slikgat; fig. 3.4: Appelenberg, Zwart Water). The flaser bedded, massive and large-scale cross-bedded sands in the lower part of the Turnhout Member were probably formed in tidal channels (fig. 3.3, 3.4). Stacking of channel sequences was found infrequently (fig. 3.3: Chaam Kapel; fig. 3.4: Zwart Goor). Only subtidal (and intertidal?) channel deposits, which have a higher preservation potential than supratidal deposits, have been found. Grain-size in the channels increases towards the north and west, which points to more turbulent tidal conditions with higher current velocities (fig. 3.3: Chaam Kapel, Snijders-Chaam) (Van Straaten, 1964). Beyond the tidal channels, thick (10 m) loam and clay-beds without major sedimentological breaks could develop in quiet conditions (fig. 3.3, 3.4: Meerle Slikgat, Zwart Water, Achtmaal). Deposition occurred predominantly in a subtidal setting, since ebb and flood sand ripples are both covered by slack-water mud. Supratidal deposits (marsh sediments, peat) were not formed in these loamy/clayey interchannel environments. The fine-grained interchannel areas point to little tidal channel migration. This channel stability is an indication for the landward (distal) setting of the tidal environment with respect to the sea.
At the southern margins of the depositional environment thick clay-beds were formed during the deposition of the channel sands and interchannel fines further north (fig. 3.3: Merksplas Straf.; App.: Beerse Blak). Part of these marginal clays are probably of supratidal origin: the absence of pollen in certain levels points to synsedimentary exposure and oxidation. The high amount of Chenopodiaceae pollen in this border zone (App.: Merksplas Straf., Beerse Blak, Ravels) is explained by local Chenopodiaceae stands in a (supra)tidal litter zone (§3.4).
The fining-upward sequence towards the top of the Turnhout Member reflects the final stages of deposition in the tidal environment. Current velocities decreased and flaser bedding was succeeded by wavy and lenticular bedding.

Tidal range at Meerle

The regular fining-upward sequence at Meerle, capped by a peat-layer, (fig. 3.11) is an indication for silting at a constant sea-level. In comparison to tidal range estimations for other members, the estimation

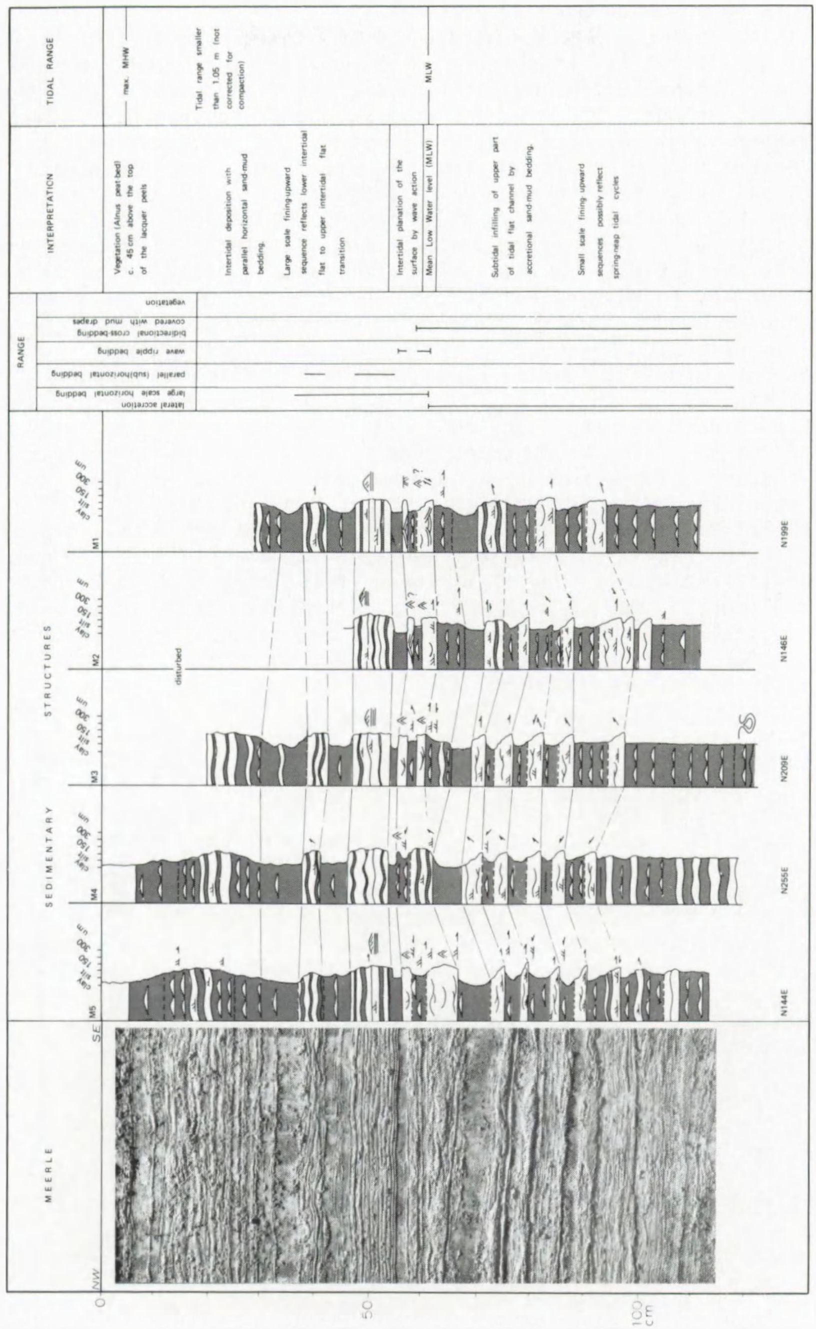


Fig. 3.11: Inshore tidal, landward facies of the Turnhout Member at Meerle. Well developed fining-upward sequence around the mean low water level and estimated tidal range.

presented here is more reliable, since mean low water (MLW) and mean high water (MHW) level could be fixed rather accurately by the range of certain sedimentary structures.
The large-scale, low-angle bedding in the lower part of the peels (fig. 3.11: 60-120 cm) is interpreted as a subtidal (mud-draped ebb and flood ripples) lateral accretion deposit. Current directions to the south are strongly dominant. The dominant southsoutheast current direction in the fining-up sequence, most likely represents the flood current.
The change from subtidal pointbar cross-bedding, into horizontal unidirectional cross-bedding with wave ripples, is interpreted as approximately the MLW level (fig. 3.11: at 60 cm). Wave action indicated by wave ripples, is most effective on the intertidal flats and tends to flatten depositional surfaces (Van Straaten, 1964). Above MLW silting continued by (unidirectional) flood currents, forming wavy bedded and lenticular bedded units. Current velocity sometimes dropped below the minimum velocity for current ripple formation and parallel horizontal sands were deposited during flood (Roep and Van Regteren Altena, 1988). Thickening up and thinning up of the parallel horizontal sets was perhaps caused by neap - spring - neap tide cycles. During spring tides flood current velocity exceeded the treshold stream velocity required for current ripple formation and current ripples developed on top of the parallel horizontal beds (fig. 3.11 at 50 cm).
In the final stages of silting, crumbly clay was deposited, capped by a peat-layer (fig. 3.3: Meerle, Meerle Slikgat; fig. 3.4: Appelenberg, Ravels, Zwart Water). The crumbliness of the clay and local absence of pollen (Meerle, Merksplas Straf.), is attributed to periodic intertidal or supratidal exposure. The fining-up sequence (from intertidal to supratidal deposits) is very gradual, which points to a landward tidal setting, for instance comparable to the recent Dollard basin in the northeastern Netherlands (Bakker, 1974). Salt-marsh sediments with crinkly, parallel bedding, of which the base is formed at 10-20 cm below the mean high water level (Roep and Van Regteren Altena, 1988), are not present.
The peat-bed at Meerle lies in situ on top of the clay (App. Meerle). The peat is dominated by *Alnus* pollen and must therefore have been formed above local mean high water (max. MHW). When both MLW and MHW levels are established, tidal range can be estimated at 1.05 m (fig. 3.11). If a 50% compaction rate is accepted (Zonneveld, 1960: Holocene, tidal fresh water deposits in the Biesbos area) for Early-Pleistocene deposits, then tidal range amounted to approximately 2 m. However, it is stressed that:
- this estimated tidal range is only valid for location Meerle.
- the compaction rate is not exactly known.
- increase and decrease of the tidal amplitude due to estuarine effects are unknown.
- a different tidal range may have occurred in earlier phases of deposition of the Turnhout Member.

Siderite concretions

Siderite nodules were found locally in a horizontal bed, 0.5 m below a peat-layer at Ravels. According to Wilson (1965) these nodules are commonly associated with fine, gray members; in particular below coal seams. He interpreted the nodules as precipitations from slightly reducing groundwater in a permanently saturated soil. The nodules in Ravels are still white and not oxidized. This indicates permanent reduction of the Turnhout Member after deposition. Nowadays, the tidal deposits are non-calcareous. The intertidal and supratidal sediments of the Turnhout Member could have been decalcified during the silting process (Van der Sluys, 1970). Subtidal deposits, however, are primary always calcareous (Van Straaten, 1964). The subtidal deposits of the Turnhout Member must, therefore, have been decalcified after deposition. As the siderite concretions are still unoxidized, it is concluded that the decalcification of the subtidal deposits occurred by post-depositional groundwater flow under reducing conditions.

Paleosalinity

Bioturbation is normally important in salt-water environments such as the intertidal flats of the recent Dutch Waddenzee (Van Straaten, 1964). The scarcity of bioturbations in the Turnhout Member is an indication for a low paleosalinity and a fresh or brackish water tidal environment. However, unlike the subrecent fresh water tidal environment of the Biesbos area southeast of Rotterdam (Zonneveld, 1960), the depositional environment was not completely fresh. Concentrations of pyrite and high percentages of Chenopodiaceae pollen occur locally in the top of the Turnhout Member (App. Ravels)(Dricot, 1961). Oxidation of the pyrite in the clay in clay-pit Ravels resulted in yellowish jarosite precipitation on exposed surfaces. Pyrite formation requires sulphate from seawater and a reducing environment. Therefore, regular influxes of salt or brackish water must have occurred in the depositional environment of the Turnhout Member.
The presence of peat with a high *Alnus* pollen content at the top of the member points to a temporarily sharp salinity decrease. The vegetation required regular flooding by fresh, eutrophic river water. Otherwise the peat would have developed into a more oligotrophic type (Roeleveld, 1974). These fresh water peats were later drowned by the sea and brackish water clay was again deposited (pyrite, Chenopodiaceae pollen) (App. Ravels, Beerse Blak).

Conclusion

The Turnhout Member was formed in an inshore, landward, fresh to brackish, micro- to mesotidal environment. Several depositional sub-environments are described: tidal channels, clayey interchannel areas, intertidal mixed flats and mudflats, tidal litter zones and fresh water swamp.

3.3.5 Hoogerheide Member

The Hoogerheide Member consists of fine to medium coarse sands with locally a clay-layer at the top. This member is in general more sandy than the laterally related Rijkevorsel Member, which contains more clay-lenses within the member (§2.5.6).

Description of the sedimentary structures (fig. 3.12, 3.13, 3.14)

The 7 m thick fining-upward sequence at Kalmthoutse Hoek is characterized at the base by large-scale, low-angle cross-bedding in medium coarse sand (fig. 3.12). The upper part is finer grained with wavy and flaser bedding and some bioturbation. Above this fining-upward sequence a 3 m thick sand-bed (11-14 m below the surface) is present dominated by low-angle cross-bedding, wavy parallel and parallel horizontal bedding. No sharp bounded clay-laminae or opposed, small-scale cross-bedding have been found. The stratigraphical position is uncertain.

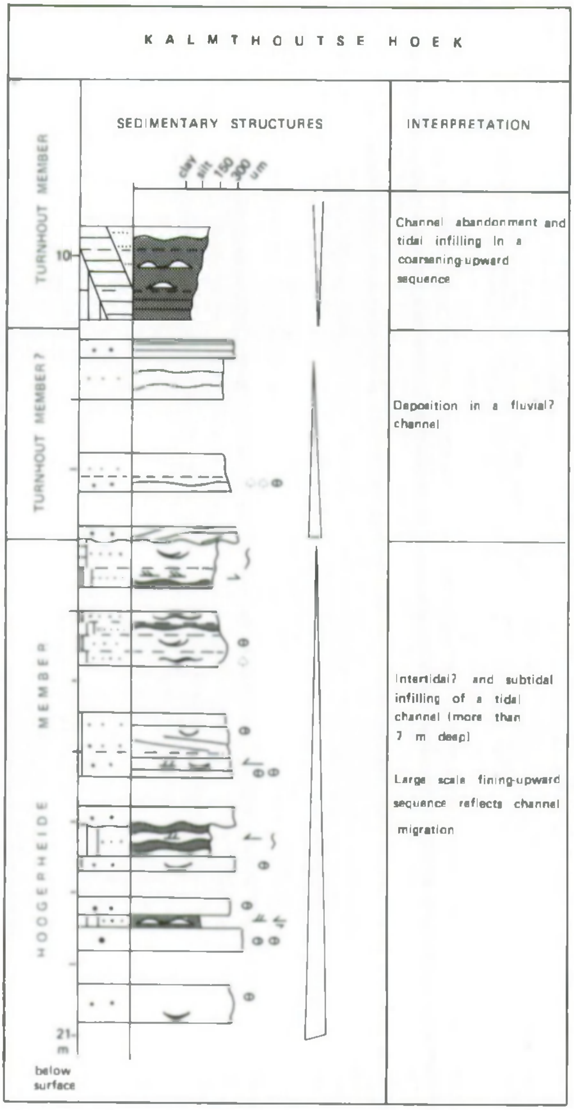


Fig. 3.12: Fining-upward sequence in a coarse-grained tidal channel-fill of the Hoogerheide Member at Kalmthoutse Hoek.

The upper part of the Hoogerheide Member was studied in several sand-pits west and south of Hoogerheide. The flaser bedding and herringbone cross-bedding at Woensdrecht Hooghuis (WH) (fig. 3.13) are the most characteristic structures in the very well-sorted sands. Towards the top a faint decrease of mud-flasers occurs (coarsening-upwards). Bidirectional, mud-draped current ripple cross-bedding in the lower part of the peel is succeeded by flaser bedding and small-scale cross-bedding. Climbing ripple bedding in drift (Reineck and Singh, 1980) grades into small-scale, tabular cross-bedding and parallel horizontal bedding.

Two types of bioturbation were found in Woensdrecht Hooghuis: Type 1 consists of round, 1 cm wide tubes with a concentric, layered sand-mud filling. Type 2 are very small (1 mm wide or less), elongate tubes, filled with mud. The bioturbation is present in small-scale current ripple bedding with ripple height of 1 cm formed by more or less unidirectional currents to the northwest. Non-bioturbated current ripple beds have larger ripple height (1-4 cm) and are formed by more bidirectional currents to the northwest and southeast; in the top of the lacquer peels the current direction is dominant to the east.

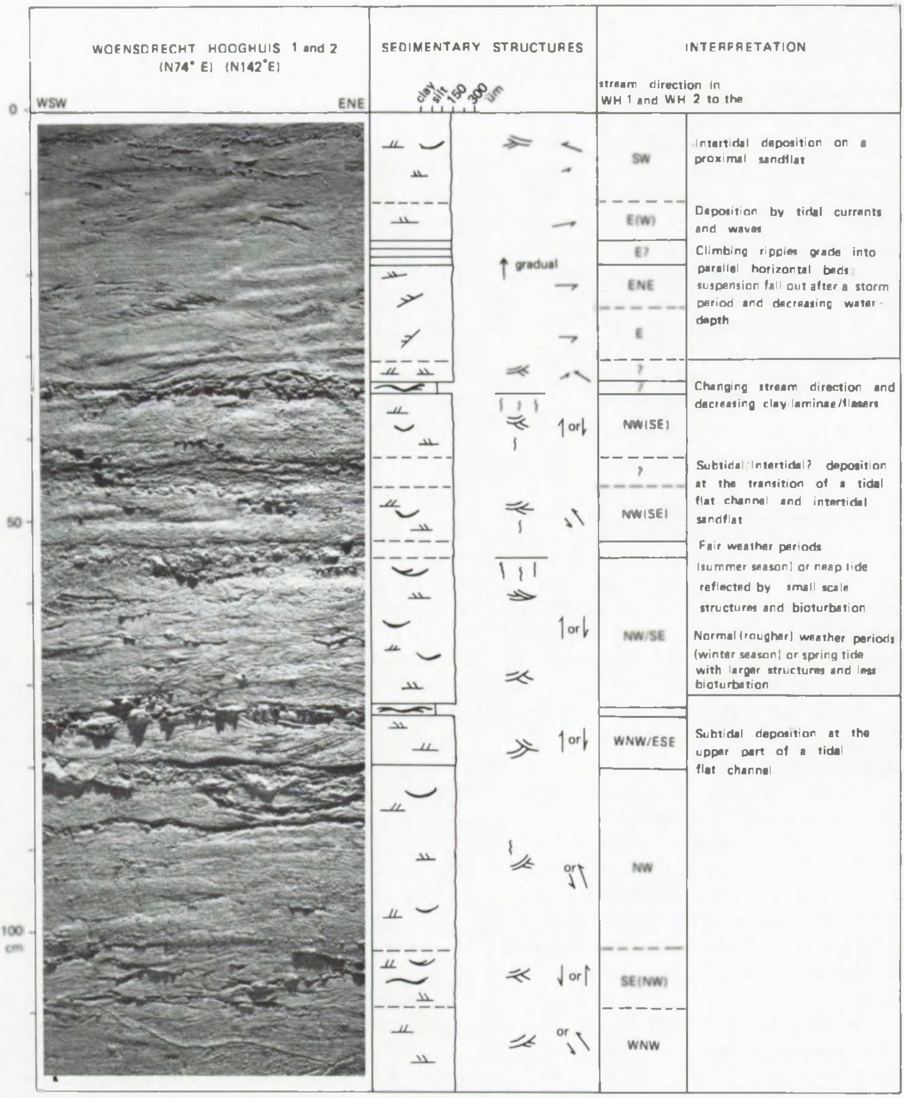


Fig. 3.13: Inshore subtidal and intertidal sandflat deposits characterized by flaser bedding and weakly bioturbated intervals. Hoogerheide Member at Woensdrecht Hooghuis.

The Hoogerheide Member in exposure Woensdrecht Rijzende Weg (WR) (fig. 3.14) is subdivided into four sub-units:
Unit 1 and 3 are dominated by small-scale, trough cross-bedding.
Unit 2 is formed by distinct sets with an undulating, erosional base. Each set consists of weakly developed relatively coarser grained large-scale, (low-angle) cross-bedded sand, which changes upward into finer grained horizontal and parallel bedded sand with an opposed dip.
Unit 4 is separated from unit 3 by a gradual boundary and it is dominated by climbing ripple cross-bedding. The angle of climb is variable (climbing ripple bedding in drift, type 1 and type 2: Reineck and Singh, 1980).

Interpretation

Kalmthoutse Hoek (fig. 3.12):
The regular occurrence of mud-laminae, the clay-pebbles and some bidirectional cross-bedding point to a tidal setting. The medium coarse, moderately sorted, large-scale cross-bedded sands with clay-pebbles are interpreted as megaripple deposits in a subtidal channel (Reineck and Singh, 1980; Van Straaten, 1964). Shells can be absent due to post-depositional decalcification. Lenticular and wavy sand-mud beds were formed on the channel bottom by sudden changes in hydrological conditions in the channel (Van Straaten, 1964).
The fining-upward sequence (7 m) is explained by the lateral migration of a more than 7 m deep channel. Silting of the channel is visualized by a change from large-scale cross-bedding into small-scale flaser and wavy sand-clay bedding. The bioturbation in the upper part of the channel-fill indicates a decrease in turbulence and subtidal or intertidal sedimentation.

Woensdrecht Hooghuis (fig. 3.13):
The mud-flasers and bidirectional herringbone cross-bedding illustrate the tidal character of the Hoogerheide Member. Bidirectional mud-draped ebb and flood current ripples point to subtidal sedimentation in the lower part of fig. 3.13. Northwestern directed ebb currents (see paleogeography) are slightly dominant.
The transition from climbing ripple bedding into current ripple bedding and parallel horizontal bedding (in sand of equal grain-size) indicates a shallowing of the water depth (fig. 3.13, top). The parallel horizontal bedding is interpreted as upper flow regime deposition on a nearly exposed intertidal sandflat. Although climbing ripple bedding is rare on tidal flats, it is locally abundant at the tidal flat - channel transition, when sand falls out from suspension after a storm period (Wunderlich, 1969). The slight decrease in mud-flasers towards the top of fig. 3.13 may point to the influence of wave action on the sandflat. According to Van Straaten (1964) wave action tends to flatten intertidal surfaces and to rework (remove) clay/silt from the intertidal sandflats. These considerations are in favor of a mean low water level in the middle of the lacquer peel (fig. 3.13).

The bioturbation types 1 and 2 in Woensdrecht Hooghuis resemble the burrows of Nereis (Reineck and Singh, 1980) and Heteromastus (Van Straaten, 1964). The bioturbated levels are related to smaller sedimentary structures; the non-bioturbated units to larger structures. These differences may be caused by seasonal or lunar variations in stream velocity (Reineck and Singh, 1980; Van den Berg, 1986). During fair weather (summer) or neap tide, tidal currents were weaker and sedimentary structures were smaller. As the sedimentation rate was relatively low, burrowing animals partly destroyed the fair weather structures. In winter time or during spring tide, the current velocities were larger and sedimentation was more rapid. Ebb and flood water produced herring-bone cross-bedding. The animals were buried or moved to higher levels rapidly and hardly destroyed the sedimentary structures.

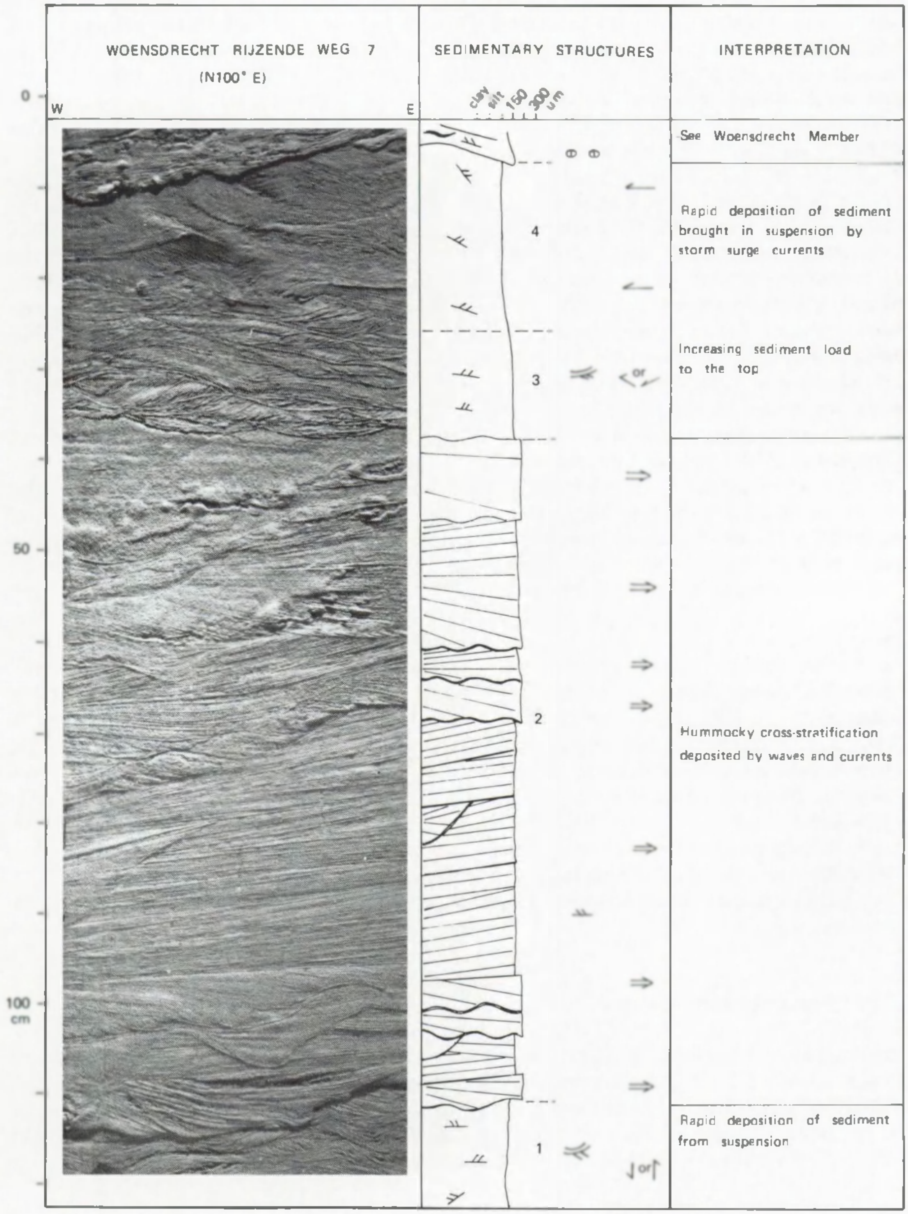


Fig. 3.14: Wave generated, low-angle (hummocky) cross-bedding and climbing ripple bedding. Hoogerheide Member at Woensdrecht Rijzende Weg.

Woensdrecht Rijzende Weg (fig. 3.14):
In this exposure there are no diagnostic indications of a tidal origin for the Hoogerheide Member. A tidal setting is proposed, however, because of the short distance between Woensdrecht Rijzende Weg and Woensdrecht Hooghuis.
Unit 4 was deposited in the form of climbing ripples in a short period of time by a westerly directed current. The trough cross-bedded units 1 and 3 are thought to have been formed by current ripples as well, at more or less right angles to the peel, because of the transitional boundary between unit 3 and 4 (environmental uniformity). Some of the sedimentary structures, however, can also be explained by wave action (offshooting and draping foreset laminae, bundled upbuilding and undulating lower set boundary (Reading, 1980).
Unit 2 is difficult to interpret. The individuality of the sets, the irregular base of the sets and the gradual change of large-scale cross-bedding into parallel bedding in one set, are not common in current dominated systems. The sedimentary structures in unit 2 show some resemblance to hummocky cross-stratification formed by the combined action of waves and currents in the nearshore surfzone of the Canadian Great Lakes (1.4-1.8 m water depth) (Greenwood and Sherman, 1986). Their hummocky cross-stratification is, however, more regular and wavy and does not possess a steep, irregular scoured base with low-angle cross-bedding as in unit 2.
Woensdrecht Rijzende Weg was probably situated at the border of an intertidal flat area. During storms the top of the intertidal flat was eroded and hummocky cross-bedding developed at the seaward side of the flats. Westward flowing ebb surge currents produced thick units of climbing ripple bedding (fig. 3.14). The hummocky cross-bedding suggests a nearshore setting for the Hoogerheide Member, but barrier island sediments have not been found so far.

Conclusion:
The sedimentary structures at Kalmthoutse Hoek and Woensdrecht suggest a proximal (=seaward), inshore, tidal depositional environment (Van Straaten, 1964). The environment is characterized by tidal channels (fig. 3.12: Kalmthoutse Hoek) and tidal sandflats. A MLW level could be established at Woensdrecht Hooghuis (fig. 3.13), but a MHW level is unknown. The proximality of the Hoogerheide Member with respect to the Rijkevorsel Member is indicated by:
- the relatively coarser-grained sediment
- the absence of intraformational clay-layers
- the presence of wave formed hummocky cross-bedding in the Hoogerheide Member.

3.3.6 Woensdrecht Member

This member consists of fine to coarse sands with a mixed heavy mineral association (§2.5.7). A continuous clay-layer at the top connects the Woensdrecht Member in the west with the Turnhout Member in the east. The former member is coarser-grained and no thick clay-layers occur within the member (fig. 2.3, 2.6).

Foraminifer *Ammonia beccarii*

Usually molluscs, foraminifers and diatoms are a common constituent in tidal deposits. In the Early-Pleistocene tidal units of Noord-Brabant, however, molluscs and diatoms are absent, possibly because of a low salinity during deposition or due to post-depositional leaching. The calcareous tests of foraminifers were also never found, but in several samples the chitinuous inner part of the foraminifers was preserved. In a total of 42 samples, prepared for the analysis of botanical macro remains, 9 samples contained 21 specimens all belonging to *Ammonia beccarii* (§3.4) (Huyzer and Van Toor, unpubl., bijlage 4). All the samples were taken from the clayey top of the Woensdrecht Member. Since other foraminifer spieces were not found it is assumed that *Ammonia beccarii* is present in situ. Reworking of forams from older (e.g. Tertiary) deposits would result in a mixed assemblage. According to Van Voorthuysen (1951) and Larsonneur (1975) *Ammonia beccarii* is found in marine, subtidal and intertidal flat environments. This ecological setting is in agreement with the results obtained from the sedimentary structures, from which a seaward, inshore (estuarine), tidal environment is proposed (see below).

<u>Description of the sedimentary structures (fig. 3.15, 3.16, 3.17, 3.18, 3.19, 3.20, 3.21)</u>

<u>Woensdrecht Rijzende Weg (WR)(fig. 3.15, 3.16):</u>
The section is divided into two sequences by two major sedimentological breaks (fig. 3.16: base WR 4; middle WR 3).
The lower sequence is characterized by large-scale, cross-bedded sands with many clay-pebbles at the base (fig. 3.16: WR 4), wavy flaser bedding, wavy sand-mud bedding (WR 4), small-scale trough cross-bedding with flasers and some lenticular bedding (WR 3).
The upper sequence starts somewhat above the middle of WR 3 (fig. 3.16). Large-scale cross-bedding with many clay-pebbles (WR 3) is covered by small-scale, trough cross-bedded sand with wavy, bifurcated and simple flasers and some parallel horizontal bedding (fig. 3.15: WR2; fig. 3.16: WR 1 and 2). The upper sequence has an upward decrease in grain-size and clay-flaser content (=decreasing heterolithicy). The sequences are summerized in fig. 3.16.
Bioturbation is restricted to one large burrow approximately at a level comparable to the base of WR 1.

<u>Woensdrecht Hooghuis (fig. 3.17):</u>
The major break at the base of fig. 3.17 reflects the erosive boundary of the Woensdrecht Member with the Hoogerheide Member. Large, curved bioturbation structures are present in the erosive base of the Woensdrecht Member. Large-scale, low-angle, accretional cross-bedding of sand and mud is the dominant bedding type. Clay-beds consist of clay-pebbles and in situ clay. The sand-beds reveal indistinct horizontal and small-scale cross-bedding.
The complete channel sequence in Woensdrecht Hooghuis is fining-upward with thinning-up of the sets, although some large-scale discordancies are present (see Appendix). The central part of the infilling seems to be more clayey than the slopes. The top of the sequence is horizontally bedded with thick, alternating beds of sand and clay. Sedimentary structures are locally destroyed by intense bioturbation.

<u>Ossendrecht(OSD) (fig. 3.18, 3.19, 3.20):</u>
Section Ossendrecht reveals 3 major breaks (middle of OSD 5, OSD 2 and OSD 1) separating 4 units (fig. 3.20):
Unit 1 is dominated by bidirectional cross-bedding with some continuous mud-laminae. Current direction is dominant to the west. Herringbone cross-bedding and climbing ripple bedding occur infrequently.
Unit 2 (fig. 3.18, 3.20) is characterized by large-scale, tabular and

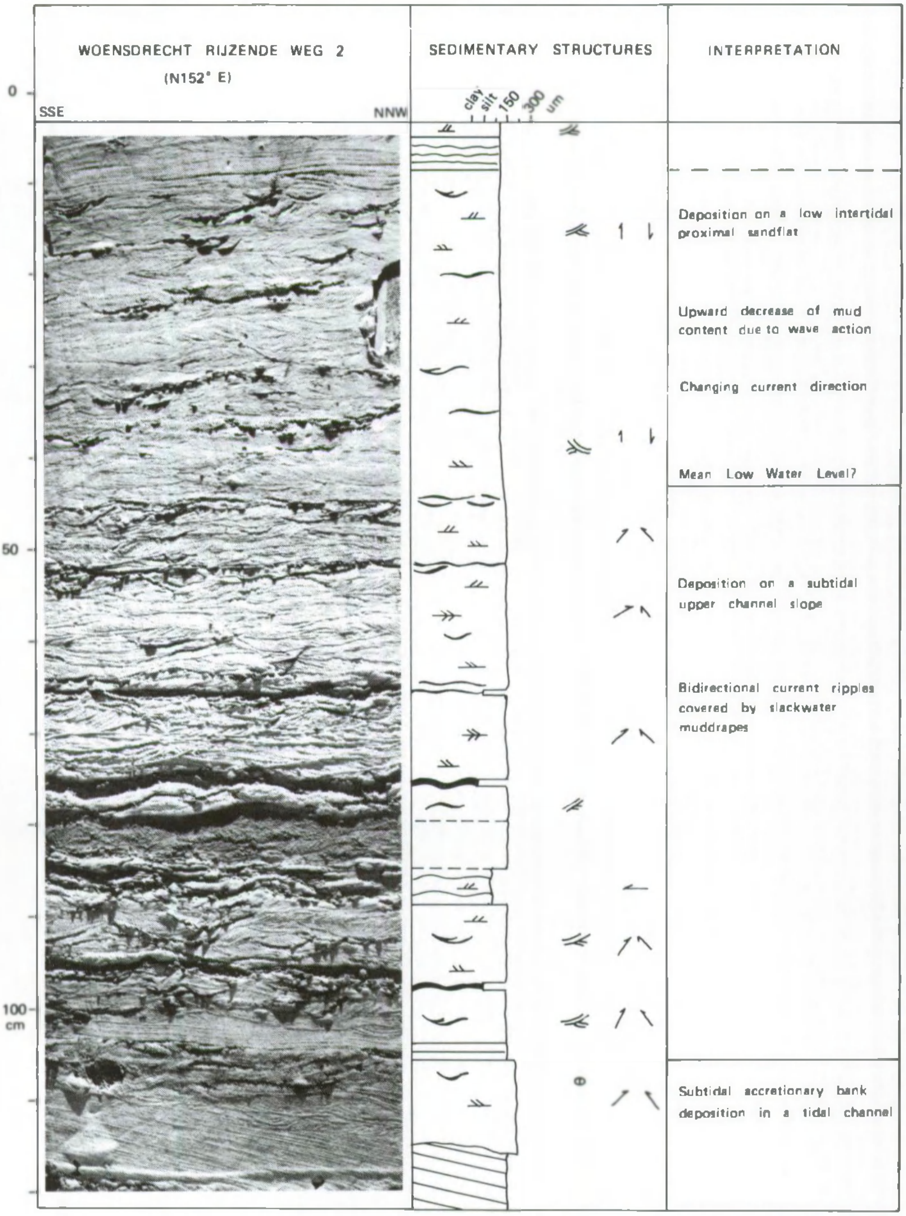


Fig. 3.15: Subtidal channel and (inter)tidal sandflat deposits with upward decreasing clay content. Woensdrecht Member at Woensdrecht Rijzende Weg.

tangential, bidirectional cross-bedding. The dominant current direction is to the westnorthwest. Clay-pebbles and reworked organic material are especially present above the erosional contact in OSD 4. The cross-bedded sets are maximal 20 cm high. The foresets are often covered with

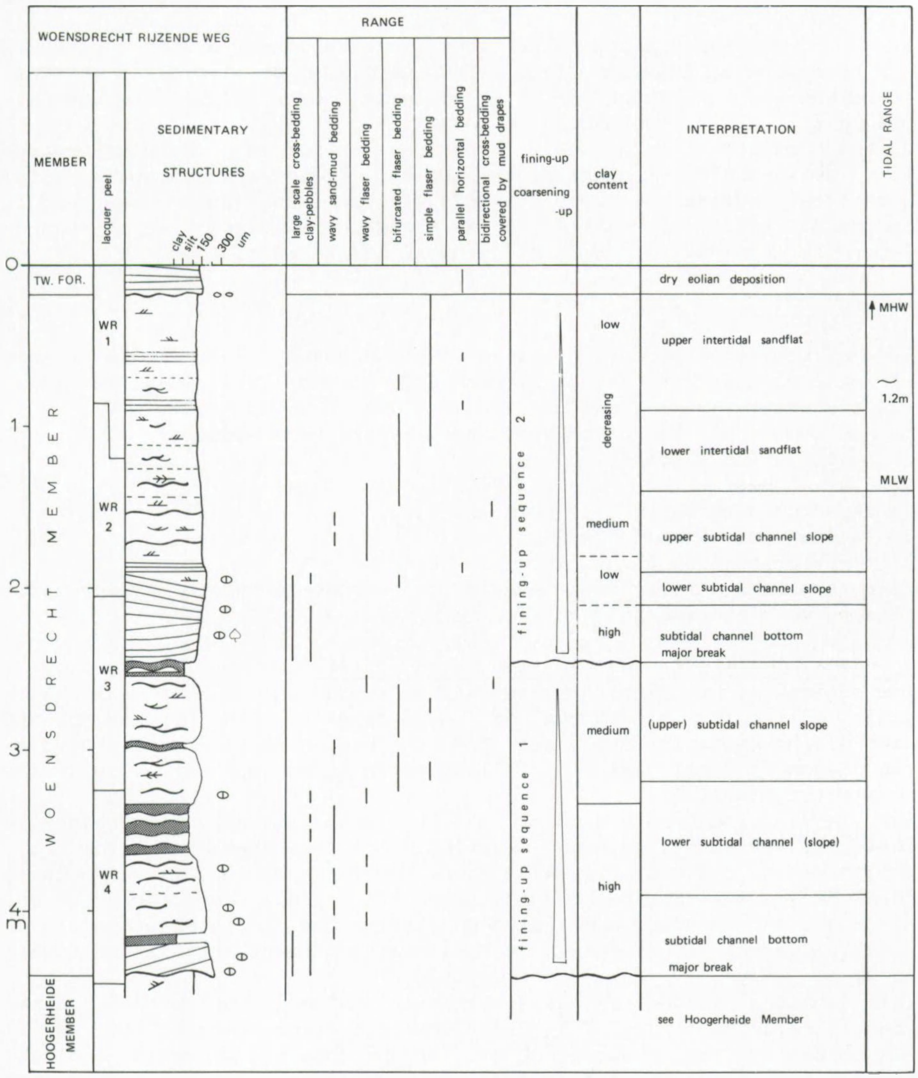


Fig. 3.16: Vertical sequence of channel and sandflat deposits and estimated tidal range in the Woensdrecht Member at Woensdrecht Rijzende Weg.

mud-drapes, perhaps locally arranged as clay-double layers (top OSD 4), and show reactivation surfaces (base OSD 2; top OSD 5). The individual cross-bedded sets have a length of a few meters at most. Thickening and thinning of the foreset-laminae in the set was not observed. However, steepening and flattening of the foreset-laminae and thickening and thinning of the set is present (not in the lacquer peels). Clay-laminae and small-scale bidirectional cross-bedding (herringbone cross-bedding) are common at the base of the large-scale cross-bedded sets.
Unit 3: The large-scale cross-bedded sets (unit 2) are suddenly replaced by wavy sand-mud bedding, lenticular bedding and small-scale

flaser cross-bedding. Clay-pebbles reflect the erosive character of the boundary (fig. 3.19: OSD 2). Wave ripple cross-bedding was found locally in 10 cm thick units which show a coarsening-up at the base and a fining-upward at the top (fig. 3.20: OSD 1). The imbricated internal structure and the stacking of the ripple crests illustrate the wave origin (fig. 3.19: top OSD 2).
Unit 4 consists of large-scale, trough cross-bedding, changing upwards into small-scale, trough cross-bedding, flaser bedding and parallel horizontal bedding (fig. 3.20: OSD 1). The large-scale cross-bedding and sandy nature of unit 4 is an isolated phenomenon as it changes laterally into wavy sand-mud bedding and laminated clay.

Interpretation

Several sedimentary structures in the Woensdrecht Member are characteristic of tidal processes (De Raaf and Boersma, 1971; De Vries Klein, 1971; Reading, 1980; Terwindt, 1981). These include:
1. bidirectional, large-scale and small-scale cross-bedding,
2. reactivation surfaces,
3. clay-drapes,
4. repeated alternations of sand and clay,
5. abundant clay-pebbles,
6. neap-spring-neap associated structures.

There are some differences between the exposures, which are probably related to differences in current velocity.

Woensdrecht Rijzende Weg (WR)(fig. 3.15, 3.16):
The sediments are finer grained, better sorted and contain more small-scale sedimentary structures in comparison to the Ossendrecht and Woensdrecht Hooghuis exposures. The two sequences are interpreted as two stages of tidal channel infilling with subsequent stacking of the sediments (fig. 3.16).
The large-scale cross-bedding at the base of the lower sequence is formed by westward migrating megaripples on the channel bottom (App. WR5). Due to channel migration wavy bedding and wavy flaser bedding developed on the subtidal channel slope (fining-up). Mud-draped ebb and flood current ripples are found up to the top of the sequence (=subtidal). Current direction was to the southsoutheast and northnorthwest in WR 3.
This subtidal deposition was interrupted by the formation of the upper channel sequence (fig. 3.16). First low-angle cross-bedding (top WR 3) was formed on the channel bottom. As the channel migrated, sediment accumulated on the channel slope in wavy sand-mud and wavy flaser bedding. The highest bidirectional mud-draped cross-bedding (=subtidal) is found approximately in the middle of WR 2 (fig. 3.15, 3.16). Above this level wavy flaser bedding changes into simple flaser bedding. The upward decrease in clay-flaser content is explained by the influence of wave action on higher, more exposed intertidal flats (Van Straaten, 1964). The parallel horizontal bedding (WR 1) is interpreted by upper flow regime deposition, formed during the emergence of the sandflat surface at low tide.
Current direction changes upward in the upper sequence. Northnorthwest-southsoutheast currents are replaced by more northeast-southwest currents, due to the change from subtidal channel into intertidal flat deposition.
Minimal tidal range can be approximated by the estimation of the mean low water level (MLW). The highest bidirectional mud-draped ripples (=subtidal) and overlying wave influenced decrease in flasers (inter-

tidal) allow designation of the MLW level in the middle of lacquer peel WR2 (fig. 3.15, 3.16). A mean high water level, based on marsh sediments or peat-beds, cannot be established, so a minimal tidal range of 1.2 m is assumed.

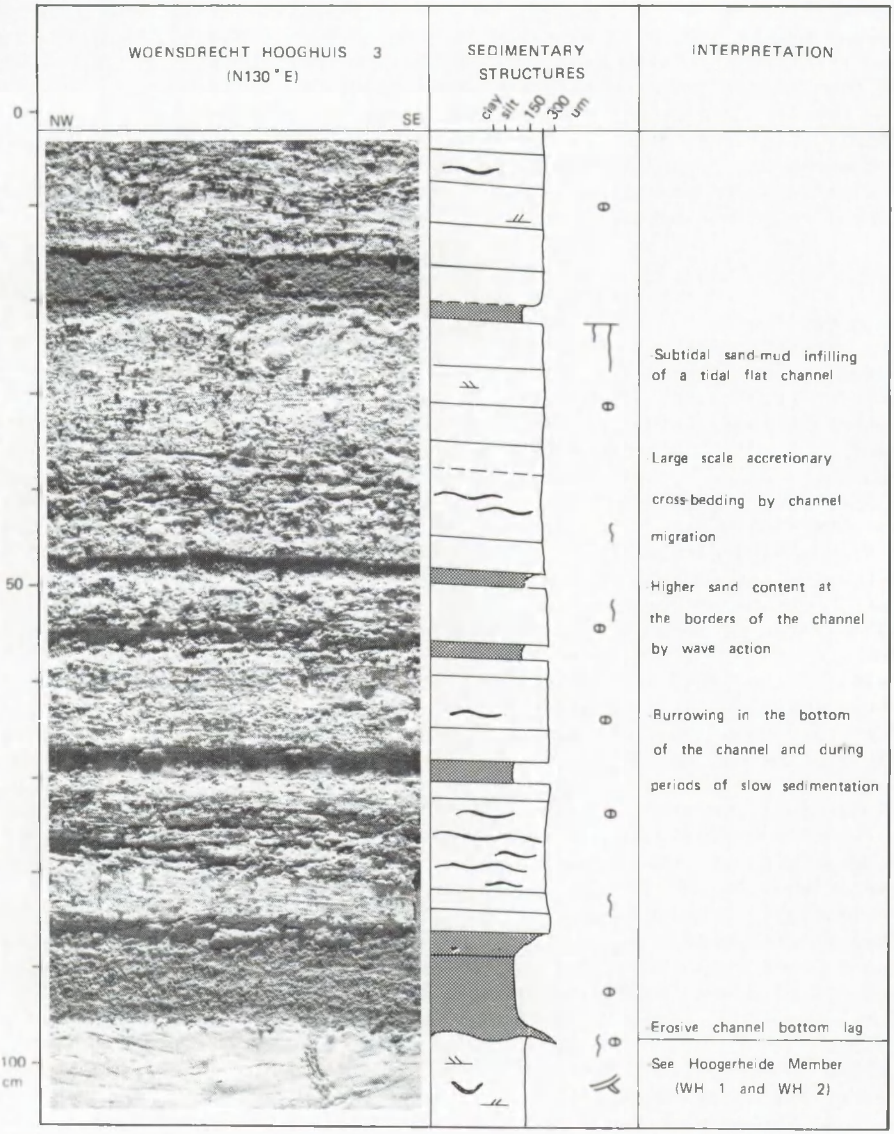


Fig. 3.17: Coarse-grained tidal channel base with low-angle cross-bedding and abundant clay-pebbles. Note the curved burrow in the bottom of the channel. Woensdrecht Member at Woensdrecht Hooghuis.

Woensdrecht Hooghuis (fig. 3.17):
The coarsely interlayered, low-angle sand-mud bedding is characteristic of relatively small tidal channels (Reineck and Singh, 1980). The clay-pebbles and coarse sand with one quartzitic sandstone pebble formed the lag deposit of the active channel. Animals at the channel bottom bioturbated the underlying Hoogerheide Member.
The large-scale unconformities in the sequence (see appendix) point to vertical stacking of channel sediments. The relatively sandy channel slopes may be explained by the winnowing action of waves. The fining-up and thinning-up of the beds towards the top reflect the final silting of the channel. The presence of well developed bioturbation levels at the top of the sequence points to a reduced sedimentation rate and a fairly high salinity. In combination with a change from low-angle cross-bedding to parallel horizontal bedding, the bioturbation might indicate an intertidal setting (Van Straaten, 1964), but MLW and MHW levels could not be designated.

Ossendrecht (fig. 3.18, 3.19, 3.20):
Unit 1 is interpreted as a subtidal deposit. The climbing ripple bedding may point to rapid infilling of a tidal channel, as was found in the upper part of the Hoogerheide Member in Woensdrecht Hooghuis and Woensdrecht Rijzende Weg.
Unit 2 (fig. 3.18, 3.20): The large-scale cross-bedding was formed by migrating megaripples in a 4 to 5 m deep tidal channel (Reineck and Singh, 1980). Current direction was dominant to the westnorthwest, probably caused by the ebb current. The subordinate flood current is reflected by reactivation surfaces in the ebb-dominated megaripples, by small-scale herringbone cross-bedding and some megaripple cross-bedding. The dominance of the ebb direction can be explained in an estuarine environment by additional river outflow or by the position in an ebb dominated channel.
Slack-water periods are registered as clay-drapes on foresets and bottomsets (fig. 3.18). The small-scale, herringbone cross-bedding at the base of the large-scale, ripple structures is interpreted as bottomsets, in which both ebb and flood currents are present.
The steepening and flattening of the megaripple foresets, together with the appearance (thickening) and disappearance (thinning) of megaripple sets, are attributed to neap-spring-neap tidal cycles in an estuarine environment (Terwindt, 1981; Visser, 1980). From neap to spring tide, currents accelerated and the set height and foreset angles increased. Deceleration of the current velocity from spring to neap tide caused the reverse.
The estuarine channel (unit 2) was filled in two phases (stacking of channel sediment), separated by an erosive contact with a clay-pebble lag deposit (base OSD 4). The sedimentary structures in unit 2 resemble the (sub)recent estuarine sediments of the Haringvliet (Oomkens and Terwindt, 1960; Terwindt, 1971; Terwindt, 1981). Channel depth (circ. 5 m) and set height (circ. 25 cm) at Ossendrecht are, however, less than in the Haringvliet. The sedimentary structures of the (sub)recent Oosterschelde are much larger (Terwindt, 1981; Van den Berg, 1986).
Unit 3: The major break between unit 2 and 3 may be regarded as a paleo mean low water level (fig. 3.19). The ebb directed, mud-draped (subtidal), large-scale cross-bedding below the break; the wave ripple bedding above the break, arranged in an overall fining-upward sequence, support this view. The transition from large-scale into small-scale structures is interpreted as the transition from tidal channel into channel slope or tidal flat.
Compared to the sediments below the MLW level the intertidal sediments

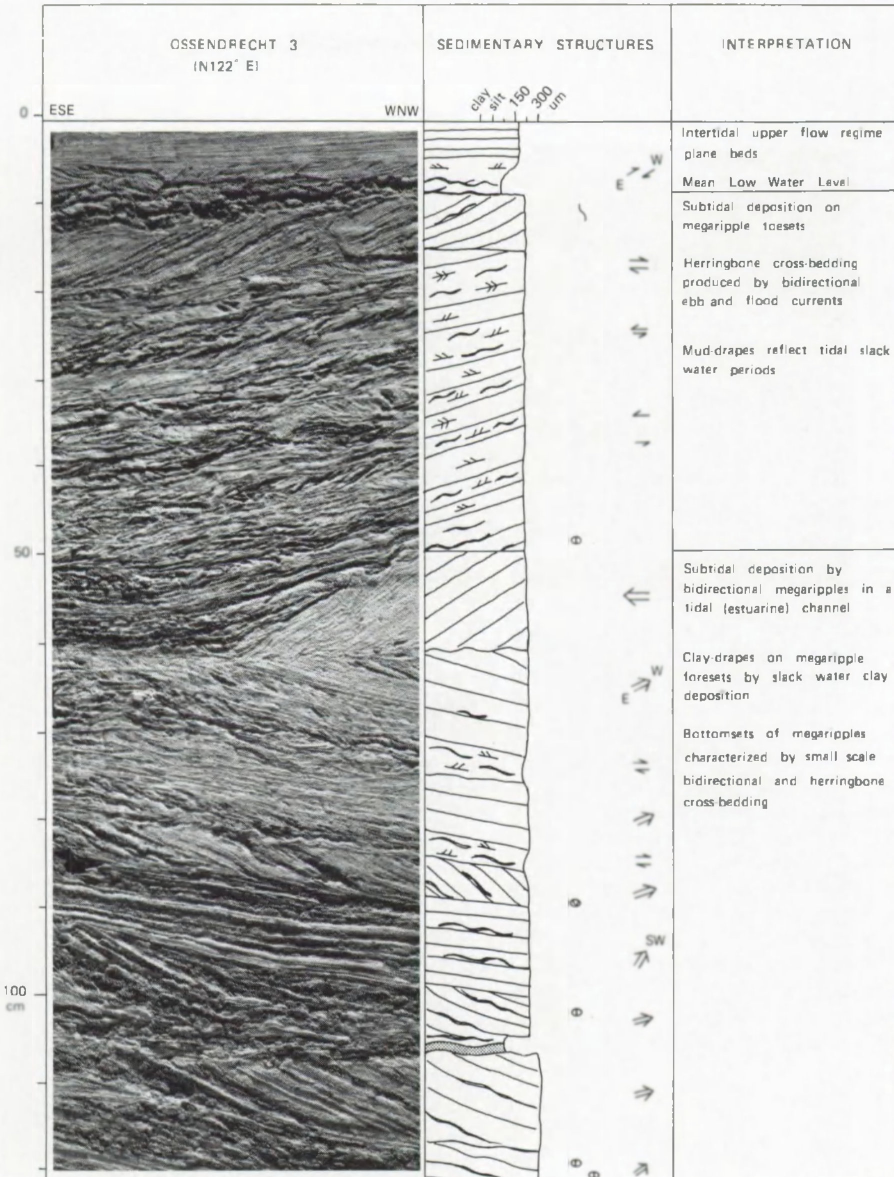


Fig. 3.18: Subtidal channel deposits with bidirectional large-scale and small-scale cross-bedding. Abundant mud-drapes formed during slack-water periods. Woensdrecht Member at Ossendrecht.

on channel slopes are often more clayey (Van Straaten, 1964). The wave ripple structures (fig. 3.19: top OSD 2) illustrate the intertidal conditions (Van Straaten, 1964). The small-scale coarsening-up/fining-up sequences, in which these wave ripples were found, may have been

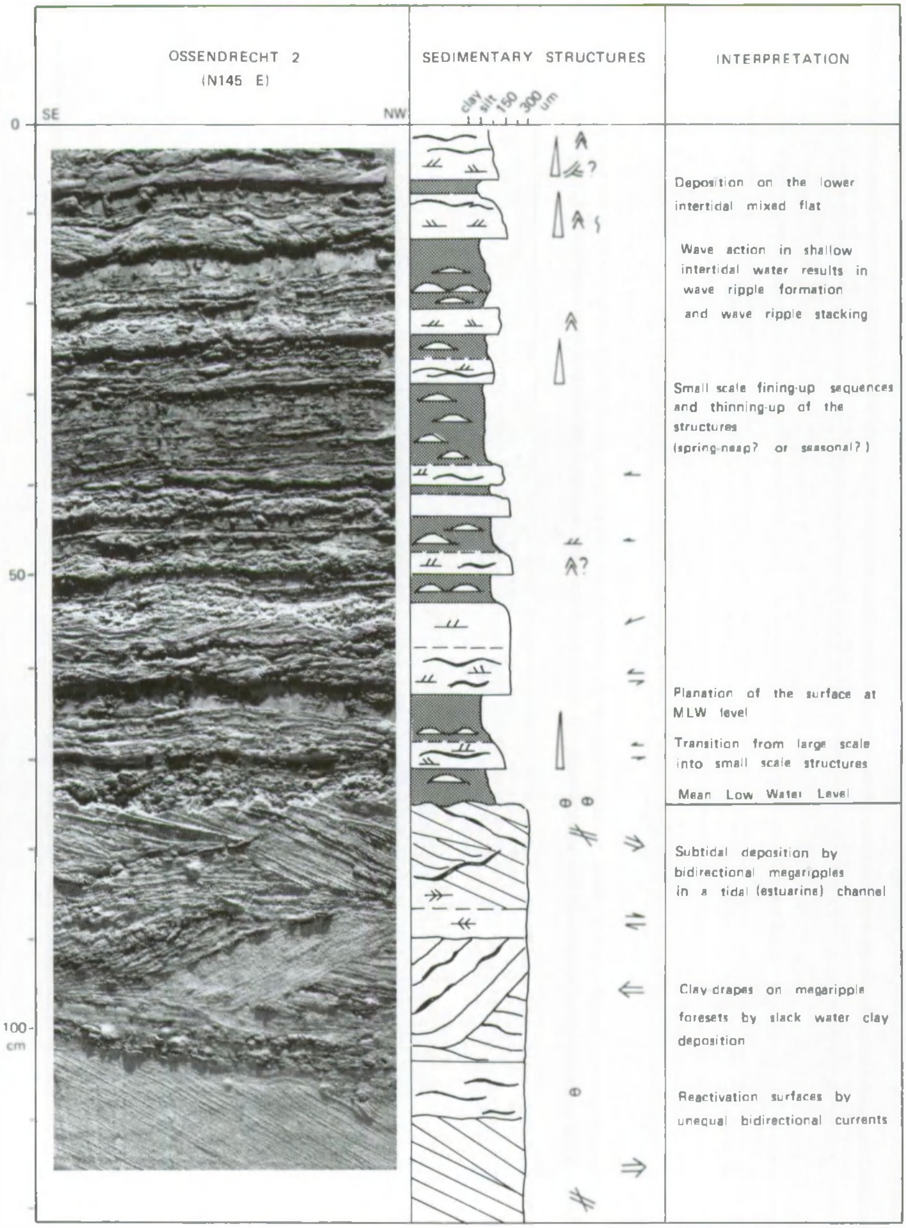


Fig. 3.19: Subtidal, bidirectional channel deposits with clay-drapes on the foresets and (inter)tidal mixed flat deposits with wave ripple bedding. The break in the deposition is interpreted as the mean low water level. Woensdrecht Member at Ossendrecht.

formed by neap-spring-neap tidal cycles or by seasonal, turbulent and fair weather periods.
Unit 4: The large-scale cross-bedded unit is a local phenomenon in an overall fining-up sequence (fig. 3.20). It is laterally equivalent to a laminated sand-clay bed, which forms a more or less continuous layer in the exposure (see App.). The large-scale, trough cross-bedding, was probably formed by megaripples in a small tidal channel. The restricted extent of the channel sediment points to limited lateral migration. Therefore, unit 4 is interpreted as a local tidal gully in an intertidal mudflat environment. Current direction in this intertidal channel was to the southwest, at right angles to the stream direction in the main channel below (unit 2). These intertidal gullies often drain the intertidal flat during ebb into the subtidal channel (Van Straaten, 1964). The fining-up sequence and thinning-up of the sets in the upper part of unit 4 indicate the final silting of the gully. Shallow water conditions resulted in upper flow regime conditions and caused the formation of parallel, horizontal bedding (top OSD 1).

Tidal range is based on MLW and MHW estimations (fig. 3.20). A constant sea-level is assumed, since a continuous sedimentary sequence (fining-upward) is present, which probably formed in a short period of time. The mean low water level is postulated from the presence of mud-draped ebb oriented ripples, wave ripples and a major break in the sedimentary structures and grain-size. The MHW level is unknown as crinkly bedded salt-marsh sediments or peat-beds are absent. The tidal range estimation is therefore less reliable than in Meerle (§3.3.4). As the top of the sequence is still of an intertidal character, tidal range is more than 1.7 m, not corrected for compaction.

Synthesis of the three exposures
The three exposures are characterized by fining-upward sequences with tidal channel sediments at the base and tidal flat deposits at the top. Sequential differences between the exposures seem to be related to differences in current velocity and wave influence. Channel depth is more or less constant (5-7 m).
The large-scale cross-bedding in relatively thick sets and coarse sand at Ossendrecht were formed by megaripples in estuarine channels with high current velocities (fig. 3.20). Channels with medium currents are characterized by thick interlayered bedding and lateral accretion cross-bedding (fig. 3.17: Woensdrecht Hooghuis). The low velocity tidal flat channels were filled with finer sand and small-scale sedimentary structures (fig. 3.16: Woensdrecht Rijzende Weg).
Lateral channel migration or decreasing currents in the channels led to a fining-upward sequence. Wave action was probably better registered in the tidal flat channels and adjacent tidal flats (WR), than in the larger, current dominated, estuarine channels (OSD). The top filling of the smaller channels and overlying tidal flat deposits are therefore characterized by a short, wave induced coarsening-up interval and decreasing heterolithicy (fig. 3.16: WR). The presence of wave-formed sequences points to a seaward (=proximal) setting of the tidal environment (Van Straaten, 1964).
The large-scale cross-bedded sets in the tidal channels are formed by bidirectional currents (fig. 3.20). The west to northwest current was dominant and probably reflected the ebb current (on paleogeographic grounds). Ebb dominated sediments are especially found in the estuarine channels, because of the connection to a fluvial system. In the overlying intertidal sandflats southwest-northeast currents were found (OSD, WR). This change in current direction was probably caused by the

difference between ebb channel flow and overbank flood water flow over intertidal flats (Reading, 1980).
The estuarine environment is characterized by mixing of water and sediment. The minor presence of bioturbation in the sediments was caused by variations in salinity and a high sedimentation rate, which hampered colonisation of the sediments by burrowing organisms (König, 1976).
Tidal range was estimated between 1.2 and over 1.7 m.

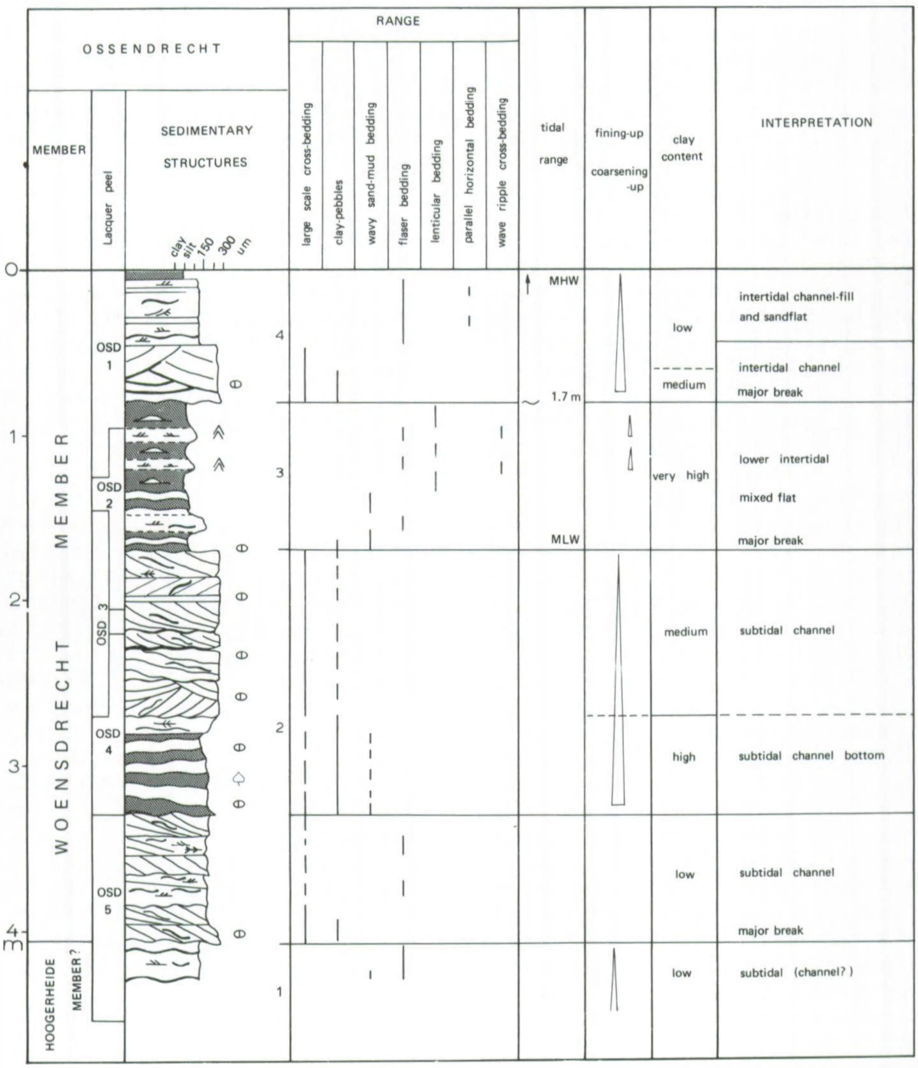


Fig. 3.20: Micro/mesotidal sequence with coarse-grained tidal channel and mixed tidal flat deposits. The intertidal channel is a local phenomenon. Woensdrecht Member at Ossendrecht.

Last stage clay deposition (fig. 3.21):

The top of the Woensdrecht Member is often formed by a continuous clay-layer, which extends laterally into the top of the Turnhout Member in the east (fig. 2.3, 2.6)(not present in exposure Woensdrecht Rijzende Weg). The lower boundary of the clay can be transitional or very sharp. The transitional boundary is found in tidal channel fillings with fining-up and thinning-up sequences (e.g. Woensdrecht Hooghuis), where the clay-layer represents the final phase in the tidal sequence.
At Korteven a very sharp boundary separates clean, white, tidal flat sand from overlying, stagnant water clay with high lutum content (fig. 3.21). Decreasing input of clastic sediment resulted in peat formation in a wet, fresh water environment (dominance of *Alnus* pollen, see §4.3.1). Later the peat drowned gradually and with increasing current velocity, well laminated, tidal, brackish water silts were formed (high Chenopodiaceae content). The sharp boundary between clay and underlying sands suggests a sudden break in the sedimentary sequence, which is illustrated by the presence of a peat-layer or soil (Geol. Survey of The Netherlands: borings 49G-66, 49G-75; Van Dorsser, 1956, p. 29 (soil), p. 33 (charcoal); Huyzer and Van Toor, 1986). The sudden decrease in transport capacity can be explained by a damming or silting of tidal channels in the estuarine environment. This may cause a decrease in tidal range and a local drop in water level. A regional sea-level change, however, cannot be excluded. The decline in water level is expressed on former high sedimentary surfaces (tidal sandflats) by soil formation (see Van Dorsser, 1956, p. 29: soil profile in the top of the sand, below overlying loam). In lower lying places conditions for peat formation were more favourable.

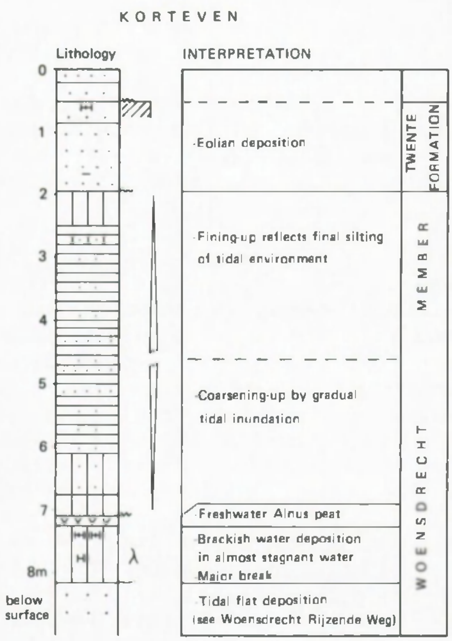


Fig. 3.21: Last stage deposition in the upper part of the Woensdrecht Member.

The water level drop was registered best in the most seaward part of the inshore tidal environment (Woensdrecht Member). High intertidal sandflats fell dry here and show a clear sedimentological break. In the landward estuarine environments to the east sediments are generally more clayey (Turnhout Member) (Reading, 1980, fig. 7.45). A temporary drop in water level was registered there as a peat-layer or soil horizon between clay-beds, or was not registered at all.

Conclusion:

The Woensdrecht Member was formed in an inshore, micro- to mesotidal (estuarine) environment. Wave action points to a seaward setting of the tidal flats and channels. A sudden drop in local water level caused peat formation and clay deposition in a more sheltered environment.

3.3.7 Gilze Member

The member consists of fine to medium, locally medium coarse sands with loam, clay and peat-beds (§2.5.8). Three units could be distinguished.

Description of the sedimentary structures (fig. 3.3, 3.4, 3.22)

Unit 1 (Appelenberg Sands) is the fine-grained lower part of the member with swift alternations of sand, peat and clay-beds (Appelenberg, Chaam Kapel, Meerle Slikgat base). Parallel horizontal bedding is the dominant bedding type.
Unit 2 (Gilze Clay) is also fine-grained, but characterized by thick fining-upward sequences, capped by a clay-layer (Nieuwmoer, Gilze, Snijders-Chaam). Parallel horizontal bedding, small-scale cross-bedding and some large-scale cross-bedding are the most characteristic bedding types.
Unit 3 (Alphen Sands) consists of medium coarse sand (Weelde, Ravels, Zwart Goor, Meerle Slikgat). Clear fining-upward sequences are absent. Massive bedding, parallel horizontal bedding and large-scale (trough) cross-bedding are regularly found. Clay-pebbles and peat-lumps are a common constituent (fig. 3.22).

Interpretation

The rapid lateral and vertical facies changes in the Gilze Member restrict the environmental interpretation. The fine to medium coarse sand, fining-upward sequences, reworked organic material and the intercalated humic to peaty soil horizons in the member may point to a fluviatile environment. Eolian deposits have not been found, although they seem to be present in equivalent deposits east of the investigated area in the Central Graben (Bisschops et al., 1985). However, distinction between eolian and fine-grained fluviatile deposits in borings can be difficult.
Unit 1: The thin (0.2-0.3 m) sand-units have an erosive lower and abrupt upper boundary and (wavy) parallel horizontal bedding. They are interpreted as floodplain overbank deposits (McKee et al., 1967) or shallow channel deposits, beyond the active river courses (Reading, 1980; Van Alphen, 1984 in: Bisschops et al., 1985). The clay- and loam-beds were formed by settling from suspension in a backswamp environment. During periods with limited supply of clastic sediment peat-beds could develop and gyttja-like beds were locally deposited in limnic environments (fig. 3.4: Appelenberg).
Unit 2: The large-scale fining-upward sequences stratigraphically overlie unit 1. The sands with associated small and large-scale cross-bedding are interpreted as fluvial channel deposits. The thick units point to deep channels (9.5 m at Gilze; 6 m at Nieuwmoer), which locally incised into the underlying Turnhout Member (fig. 3.4). The gradual transition of sand into clay and the large lateral extension of the clay-bed at Gilze probably reflect deposition by a meandering river. The clay at the top of the fining-up sequence is interpreted as a backswamp deposit outside the active river course. Swamp forests dominated by *Alnus* and other warm temperate species gave rise to local peat formation.
Unit 3: The large-scale, low-angle cross-bedding in medium coarse sand is interpreted as megaripple bedding formed in fluvial channels (fig. 3.3, 3.4: Weelde upper part, Zwart Goor, Meerle Slikgat upper part). The absence of fining-up sequences and clay-beds points to rapid migration of the channel systems, which is more characteristic of braided

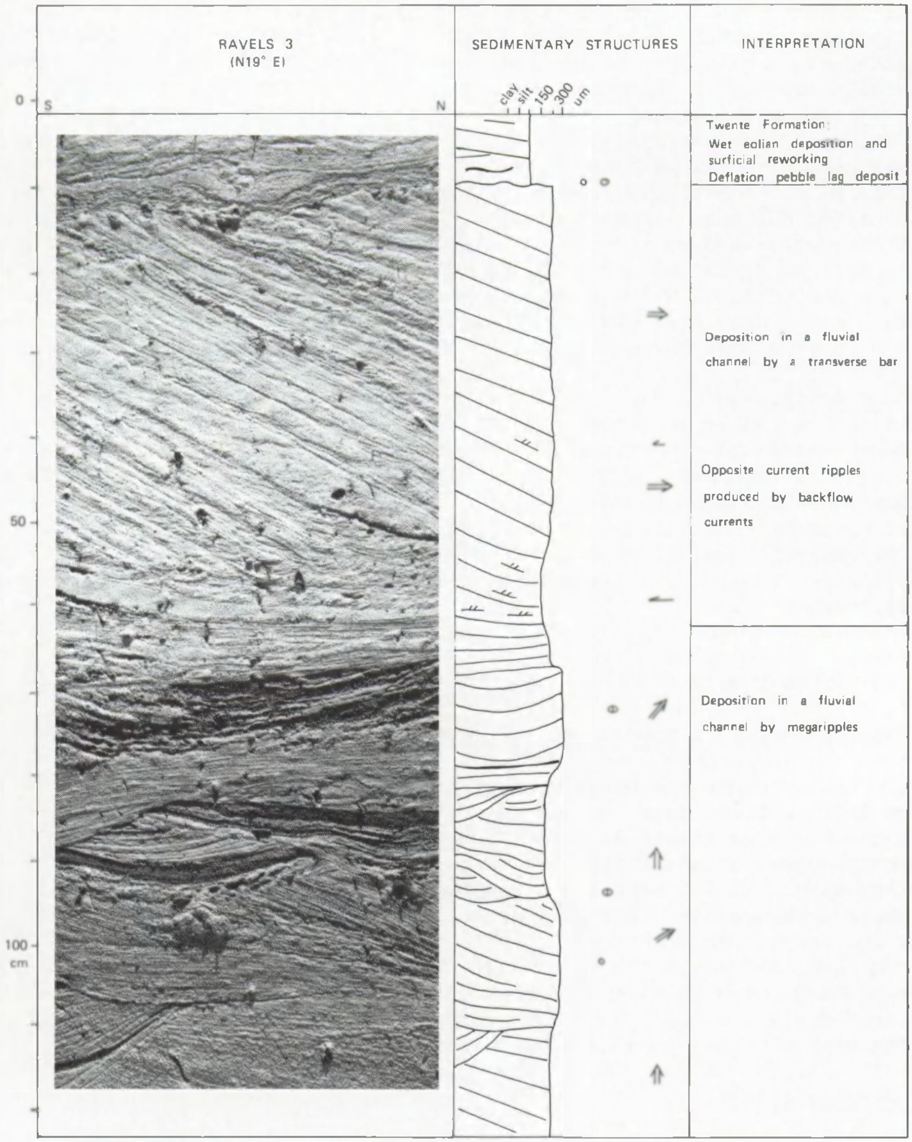


Fig. 3.22: Fluvial, braided deposition by megaripples and transverse bars in the Gilze Member at Ravels. Note the backflow ripples at the base of the large-scale foresets.

river systems (Vandenberghe et al., 1986). The large-scale involutions (50-100 cm)(cryoturbations?) at different levels within unit 3 in Weelde (not in the appendix) indicate periglacial sedimentary conditions (Vandenberghe and Krook, 1981).
Unit 3 was studied in detail in exposure Ravels (fig. 3.22). Large-scale, trough cross-bedding is dominant (see app. Ravels 1, bed 7). The

troughs are broad (approximately 6 m wide) and shallow (0.5 m deep). Large-scale cross-bedding, backflow ripple cross-bedding and climbing ripple cross-bedding are found in the trough structures. Reworked plant material and clay-pebbles are common (fig. 3.22). The trough cross-bedding at Ravels is interpreted as channel cut-and-fill structures in a braided river system (Picard and High, 1973; Singh, 1977; Reineck and Singh, 1980). The shallow troughs at Ravels generated during high discharges. A subsequent decrease of the current velocity caused the lateral filling of the channel with medium coarse sand (210-300 µm). Backflow currents produced backflow ripples at the base of the large-scale cross-bedding (fig. 3.22). The cut-and-fill structures at Ravels suggest a stream direction to the westnorthwest. During low discharges fine sand, silt and humic material was deposited on the trough slopes. This water-saturated sediment locally slumped towards the centre of the trough (see App. Ravels 1, left of the middle).

Conclusion:
The Gilze Member is a complex of three units which developed in different sedimentary environments. The lower unit 1 is deposited in a fluvial environment with small river channels, floodplain overbank deposition, backswamps and lakes. Unit 2 is interpreted as a meandering river deposit consisting of channel sand and backswamp clay and peat. The coarser grained upper part of the Gilze Member (unit 3) was formed by large, shallow (partly westnorthwest flowing), sandy braided rivers systems.

3.3.8 Bavel Member

Description of the sedimentary structures (fig. 3.3, 3.23)

The Bavel Member in clay-pit Bavel is subdivided into two lithological units. The lower unit 1 consists of gray, calcareous clay. Intercalated sand-beds show weakly developed, wavy parallel bedding (fig. 3.3).
Unit 2 consists of well-sorted sands, erosively overlying the clay. The fine sand (105-150 µm) at the base of unit 2 is characterized by small-scale cross-bedding and well-developed, climbing ripple cross-bedding (fig. 3.23). The upper part of unit 2 is coarser grained (150-210 µm) and separated from the fine sand by a lag deposit with clay-pebbles, wood fragments and fine gravel. Sedimentary structures are dominated by large-scale, trough cross-bedding. Sediment becomes finer grained to the top (fining-up) and set height decreases (thinning-up).

Interpretation

The sedimentary structures of the Bavel Member are associated with fluviatile depositional environments.
Unit 1: Thick clay-beds can be formed in backswamp environments or as clay-plugs in cut-off channels of a meandering river (Reineck and Singh, 1980). The clay is very calcareous, plastic and contains no soil horizons or bioturbation by roots. Therefore, deposition did not occur in a backswamp environment, but more probably in a channel cut-off of a meandering river in a warm temperate climate (chapter 4). The channel cut-off was quickly filled with suspended sediment (clay, silt, fine sand) introduced by overbank flows (no decalcification, no vegetation, limited physical ripening). The large thickness of the clay-plug points to a depth of the abandoned channel of more than 10 m. This deep channel eroded the Gilze Member between Snijders-Chaam and Bavel (fig.

3.3). In a newly excavated clay-pit the sand-beds in the thick clay-unit were better developed. They revealed abrupt alternations of sand and clay, some draping of clay-laminae on small-scale current ripple foresets and opposed current ripple cross-bedding approximately 50 cm apart. These features might reflect fresh water tidal influence during the filling of the meander cut-off.

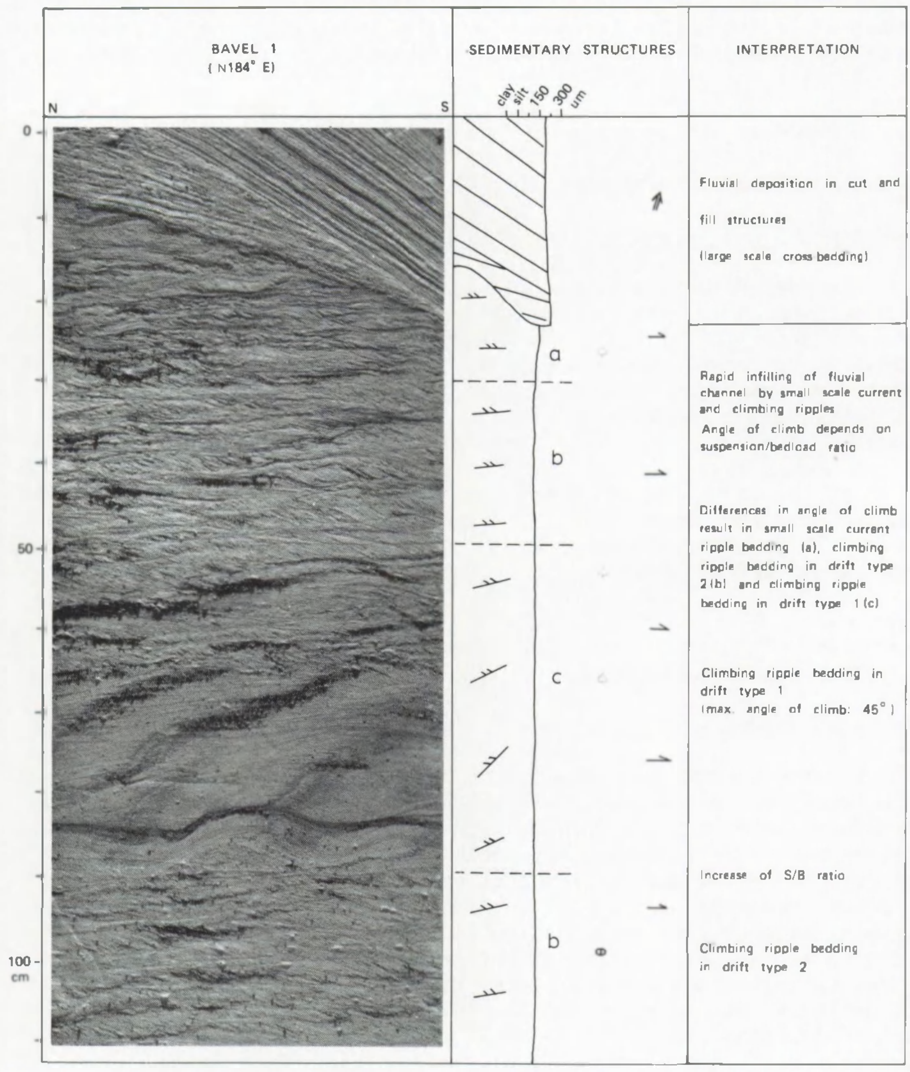


Fig. 3.23: Well developed climbing ripple bedding, truncated by large-scale cross-bedding in a fluvial channel of the Bavel Member at Bavel.

Unit 2: The sedimentary structures were formed by megaripples, small-scale current ripples and climbing ripples in a fluvial channel. After the underlying cut-off channel was filled with clay (unit 1), a new (meandering) river channel eroded part of the clay-plug. Filling first

occurred with fine-grained, well-sorted sands. The small-scale current ripple and climbing ripple beds are interpreted as pointbar deposits of a southsouthwest flowing river course (Reineck and Singh, 1980).
Later, medium fine sands were deposited in cut-and-fill structures. The large-scale, megaripple trough cross-bedding indicates a more western stream direction then. The Bavel Member was probably formed along the southwestern margin of a meander belt, since no sediments have been found southwest of Bavel. The southeast-northwest oriented floodplain occurred predominantly northeast of the Gilze-Rijen fault system and locally crossed this fault at Bavel (Bisschops et al., 1985, fig. 21).

3.3.9 Middle- and Late-Pleistocene formations

3.3.9.1 Sterksel Formation

Description and interpretation of the sedimentary structures (fig. 3.4)

The Sterksel Formation in boring Appelenberg is characterized by medium coarse, moderately well-sorted sands. The interpretation is hampered by the isolated position of the unit in the study area (fig. 3.4). The parallel horizontal and low-angle, large-scale cross-bedding are interpreted as megaripple foreset structures. The uniform vertical sequence is explained by stacking of fluvial channel sediments. The channel deposits fill a more than 5 m deep valley eroded into the underlying Gilze Member (fig. 3.4).
East of the investigated area extensive beds of equivalent channel sediments have been found in the Central Graben (Bisschops et al., 1985). This widespread occurrence can be explained by rapid channel migration, which is more characteristic of sandy, braided river systems (Allen, 1965).

3.3.9.2 Eindhoven Formation

Description and interpretation of the sedimentary structures

The Eindhoven Formation consists of loamy, fine to medium coarse sands. The small-scale cross-bedded units (fig. 3.4: Wernhout Maalbergen, Achtmaal) were probably formed by small current ripples in a fluvial environment. Since internal, structural evidence is weak, the fluvial origin is deduced mainly from the geomorphological and stratigraphical context (Reading, 1980). The scarcity of large-scale cross-bedding may be explained by fluvial deposition in relatively small river channels (fig. 3.4: Wortel: current direction northnorthwest). The dominance of parallel horizontal bedding can be related to ephemeral fluvial activity (Reineck and Singh, 1980), to floodplain overbank deposition along active shifting rivers (McKee et al., 1967) or to a combination of fluvial and eolian deposition (Vandenberghe and Krook, 1985). The fining-up sequences (Achtmaal, Wernhout Maalbergen) are locally capped by peat- and loam-layers (fig. 3.3: Bolk). The high ratio of channel sediment versus fine backswamp sediment suggests combing of the floodplain by low-sinuosity, sandy braided rivers (Allen, 1965). The characteristic silt component in the deposits is interpreted by reworking of eolian material (Van den Toorn, 1967; Bisschops et al., 1985).

3.3.9.3 Twente Formation

Description and interpretation of the sedimentary structures

The Twente Formation consists predominantly of fine to medium fine sand. Medium coarse sand, loam-beds and pebble-layers occur locally. The stratigraphical and environmental interpretations are based largely on Van der Hammen et al., (1967), Vandenberghe (1985) and Schwan (1986).
A gravel-bed, at the base of the formation (App. Beerse Dakt), is correlated with the Gilze gravel deflation lag of Weichselian Lower Pleniglacial age (Vandenberghe, 1985). The large-scale, low-angle cross-bedded medium fine sand above the Gilze gravel is interpreted as eolian sediment, deposited in low dunes or filling former topographical depressions. This sand is succeeded by a massive silt-layer and wavy, parallel bedded sand (fig. 3.3: Beerse Dakt, Merksplas Strafinrichting, Beerse Blak). The silt-bed is probably equivalent with the "Brabantse loam" of Weichselian age (Van den Toorn, 1967; Bisschops et al., 1985). The high topographical position of this silt-layer on the Campine microcuesta points to a primary eolian transport, but local concentration in standing water may have occurred as well (Bisschops et al., 1985; Haest et al., 1986).
The "Brabantse loam" is locally deformed and truncated by a second pebble-layer (fig. 3.4: Weelde), formed by surficial runoff and deflation, which is correlated with the Beuningen gravel-bed of Weichselian Upper Pleniglacial age. Loamy sands above the Beuningen gravel with a large regional extent are interpreted as eolian sand-sheet deposits and correlated with the Older Coversand II (Van der Hammen et al., 1967). The alternating bedding of sand and silt laminae is explained by seasonal variations in wind velocity (Schwan, 1986).
A humic soil or peat-layer on top of the loamy sands witnesses a period of surface stability (fig. 3.4: Weelde and Ossendrecht). The soil is equivalent to the Usselo-layer of Allerød age. After the Allerød period renewed local eolian sedimentation occurred at Ossendrecht and Weelde, which is correlated with the Younger Coversand II. The large-scale, low-angle, trough cross-bedding and parallel horizontal bedding, point to the deposition in low dunes at Ossendrecht. The parabolic dune forms indicate a southwest-northeast wind direction (Meys, 1974).

3.4 Paleoecological interpretation of the Early-Pleistocene deposits in western Noord-Brabant.

S.J.P. Bohncke and C. Kasse

Summary
Transformed pollendiagrams, in which each biozone is represented by a bar diagram consisting of ten widely defined vegetation units (physiognomical groups) together with macro remain analyses provided a generalized picture of the paleoenvironmental development for the Turnhout and Woensdrecht Members.
The lithologically and sedimentologically established tripartition (thick lower clastic unit, organic unit, thin upper clastic unit) could be recognized in the paleobotanical record.
Both clastic units show lateral changes in the pollen assemblage that can be interpreted by a northern increase of waterdepth, salinity or increased distance to the zone with upland forest, heath and floodplain forest.
The levels with increased Chenopodiaceae pollen (physiognomical group 2) have been interpreted as representing tidal litter zones. Tidal litter zones established due to tidal action along the southern rim of the basin, at places with temporarily emerging substratum in the understory of an intertidal herbaceous marsh.
The species composition of the intertidal marsh at the western side of the investigated area, preceding and overlying the organic level, indicate mesohaline conditions. Eastwards oligohaline to nearby fresh water conditions prevailed.
During the maximum of the peat formation the area must have been characterized by an extensive herbaceous marsh, dominated by *Typha spp.* and intermingled with shallow pools with stagnant fresh water. In some places soil formation took place in the emerging substratum. The local presence of alder flood plain forest could not be established. It is supposed that these were located south and eastwards of the study area.

3.4.1 Introduction and methods

Clayey and peaty intervals within the Turnhout and Woensdrecht Members have been subjected to pollen analyses. Besides biostratigraphical information, the pollen data from these Early-Pleistocene deposits yield information about changes in the paleoenvironment. To facilitate the evaluation of the data obtained, the following procedure has been developed.

The pollen diagrams have been subdivided into zones. For this subdivision major changes in the species composition as well as lithological changes have been taken into account. One zone in the pollen record is considered as a more or less stable phase in the environmental evolution, representing a single specific biotope. It is assumed in this study that climatic changes within the members are only marginal and that changes in the pollen record are also determined by changes in the depositional environment.

A second assumption is that within the Turnhout and Woensdrecht Members only one phase with increased *Alnus* values is present. Lithologically this level is characterized by an increase in organic content (humic clay or peat). This level forms a base for litho/biostratigraphical correlation.

In order to characterize the biotope represented by one pollen zone, the pollen types have been assigned to ten widely defined vegetational units (or physiognomical groups) in accordance with the occurrences of their present-day relatives (Van der Burgh, 1983). Throughout a pollen zone, the percental values of all pollen and spore types belonging to one specific physiognomical group have been added up. Subsequently the total of the percental values of the physiognomical groups has been fixed at 100% and the individual contribution of each group has been recomputed. The results have been plotted in bar diagrams. In this manner each biozone within a lithostratigraphic unit is simplified into a diagram consisting of ten bars (the physiognomical groups). The height of the bars is related to the percental contribution of the group to the total of 100%.

In this conception it should be possible to give a short overall characterization of that specific biotope and subsequently group these biotopes in a logical sequence, both laterally (space) and vertically (time), in order to compose a three-dimensional paleoenvironmental picture. The bar diagrams have been arranged in two north-south and one east-west cross-section, in accordance with the lithological (fig. 2.5, 2.6) and sedimentological cross-sections (fig. 3.3, 3.4). It is assumed that in this way gradients in the paleoenvironment would show up, which could subsequently be interpreted in spatially related more or less contemporaneous biotopes. Changes in these biotopes in time (vertical changes in the bore sections) can then give an indication of the processes that have exerted their influence on the study area during the period of deposition of one specific lithostratigraphic unit. It must be stressed that the sedimentological and paleobotanical record of the boreholes and exposures may contain hiatuses. The occurrence of non-deposition or erosion, however, can provide information that is of significance for the reconstruction of the paleoenvironment.

3.4.2 Grouping of the pollen and spores

Ten physiognomic groups have been designated in this study, representing most of the variation in the biotopes during deposition of the Turnhout and Woensdrecht Member. These groups are:

1. salt-marsh and mudflat vegetation: e.g. *Armeria*, Hystrichosphaeridae.
2. tidal litter zone vegetation: e.g. Chenopodiaceae, *Artemisia*, Compositae tubuliflorae.
3. fresh water vegetation, subaquatic and floating plants: e.g. Algae, *Myriophyllum*, *Nuphar*, Umbelliferae.
4. herbaceous shore vegetation: herbs from the telmatic zone in fresh water environments: e.g. *Typha*, Gramineae, Cyperaceae, *Chamaenerion*, Filicales.
5. carr vegetation: mostly trees and shrubs (ferns) occurring in stagnant shallow water or damp soils: e.g. *Pterocarya*, *Alnus*, *Betula*, *Salix*.
6. floodplain forest: trees and ferns from damp soils adjacent to streams, which are periodically flooded: e.g. *Alnus*, *Pterocarya*, *Fraxinus*, *Osmunda*.
7. upland forest: forest on relatively dry, well drained sites: e.g. *Quercus*, *Betula*.
8. coniferous forest: forest on dry, nutrient poor soils: *Pinus*, *Tsuga*, *Sciadopitys*.
9. heath: dwarf-shrub vegetation dominated by heather: Ericaceae, *Juniperus*.
10. oligotrophic peat bogs: *Sphagnum*, *Myrica*, Cyperaceae.

It is important for the paleogeographic reconstruction to correctly interprete the Chenopodiaceae in the paleobotanical record. As discussed in Bohncke (1984) a variety of habitats is suitable for species within the Chenopodiaceae family. The habitats that are associated with marine environments seem most applicable in this study: tidal mudflats of the eulittoral zone, salt-marshes of the supra-littoral zone and tidal litter zones (drift zones) near the mean high water level.
With respect to the salinity of the inundation water tidal litter zones embrace environments from polyhaline up to and including b-mesohaline, whilst salt-marshes and tidal mudflats occur in the poly- and eu-haline zone (salinity classification according to the Venice System, 1959).
The species associated with the Chenopodiaceae peaks can help to discriminate between the above mentioned possible environments. Salt-marshes and mudflats are characterized by obligate halophytes, while tidal litter zones are intermingled with a shore vegetation and bear species which are nitrophilous due to the decomposition of unwashed organic matter. Tidal litter zones are generally restricted to the contact zone between fresh water and brackish water and establish on damp soils or in temporarily inundated environments (Beeftink, 1965).
Pollen types present at or around the levels of increased Chenopodiaceae values (in Beerse Blak, Ravels, Merksplas Strafinrichting, being the southeastern part of the area) are *Artemisia*, *Armeria*, *Polygonum persicaria* type, Rubiaceae, *Rumex hydrolapatum*-type, Umbelliferae, Typhaceae, *Lythrum*, *Lysimachia* and sometimes Compositae tubelliflorae-type. Similar pollen assemblages have been found in a detailed study on Dutch Holocene coastal environments (Bohncke, 1984) and have been interpreted as derived from phytocoenoses occurring in mesohaline conditions. *Atriplex* species establish in the understory of a vegetation on nitrogen rich ruderal drift zones. Its fresh water equivalent is the Filipendulion alliance. A closely associated phytocoenose may have been present at Ravels were *Chamaenerion*-type pollen and the nitrophilous *Urtica* occur in the transition from ruderal mesohaline to fresh water with increased *Alnus* values.
Comparison with data from actuo environments can further elucidate the type of depositional environment represented by the fossil pollen

assemblage. Surface samples from salt-marshes in the Baye de Mont St. Michel proved to be poor in pollen (Morzadec - Kerfourn pers. comm., 1985). Long periods of aeration throughout the year in combination with the rather coarse-grained sediments appear to result in a poor preservation. Pollen in tidal mudflats from the East Frisian Islands (northern Germany) is generally well represented although preservation is sometimes poor (Chowdhury, 1982). In conclusion we may assume that the levels with high Chenopodiaceae content reflect the transitional zones with species from fresh water and slightly brackish environments, intermingled with ruderal elements and approach phytocoenoses that are most closely associated with tidal litter zones from lagoonal and estuarine environments. This is supported by sedimentological analyses of the vertical intervals (Kasse, 1986; §3.3.4, §3.3.6), in which the Chenopodiaceae maxima are formed in an intermittent position in relation to the underlying subtidal sediments and the overlying fresh water alder swamps. Hence the Chenopodiaceae are classified in this study as tidal litter zone species: physiognomical group 2.

3.4.3 Turnhout and Woensdrecht Members

North-south transect (fig. 3.24):

1. Lower clastic unit:

The thick, lower clastic unit of the Turnhout Member, is represented by pollen zones: Beerse Blak 4.1 and 4.2, Merksplas Strafinrichting 4.1, Wortel 4.1, Meerle Slikgat 4.1, Galderse Meren 4.1 and Chaam Kapel 4.1 (fig. 3.24).

The bar diagrams reveal the following trends in the physiognomical groups. In the southernmost locations of the cross-section (Merksplas, Beerse Blak) the tidal litter zones (group 2) are well established. Furthermore, the herbaceous shore vegetation (group 4), carr vegetation, floodplain forest (group 5 and 6) and heath (group 9) are important in the pollen assemblage. Northwards at Meerle Slikgat groups 4, 5 and 6 are still significant. Group 9 is declining in favour of group 8 (coniferous forest). At the northernmost end of the transect (Galderse Meren and Chaam Kapel) this trend is even more evident whereas groups 5 and 6 have reached a minimum.

Interpretation

At the southern end of the transect (Merksplas and Blak) a marshy shore vegetation was present. Under the influence of tidal action organic debris was trapped here and tidal litter zones consisting among others, of Chenopodiaceae species established. Water depth must have been limited to allow the establishment of tidal litter zones, which develop near the upper limit of the MHW level. These litter zones apparently did not form a small rim bordering a brackish water marsh, but formed quite an extensive zone.

The relatively high values for groups 5 and 6 indicate the close presence of fresh water with carr vegetation and floodplain forest. Locally at Merksplas at the termination of this phase the herbaceous shore vegetation is represented by *Alisma plantago aquatica* and *Typha spp.*, while in shallow water *Azolla tegeliensis* and *Salvinia natans* thrived (fig. 3.25: 5.2 m). The high values for group 9 possibly derive from a heath that grew on the nearby outcropping Tertiary sandy substratum.

At Meerle Slikgat open water (3) and shore line vegetation became relatively more important and it is assumed that the average water depth increased northwards. Consequently, in the absence of temporarily emerging substratum, tidal litter zones (2) did not develop, resulting

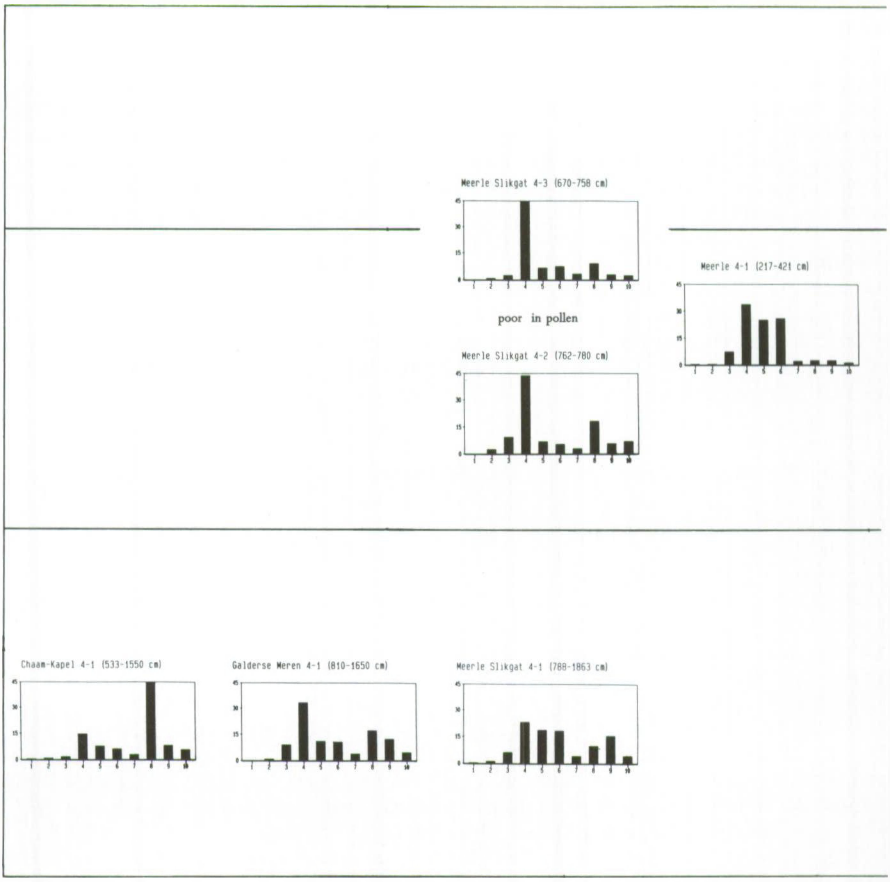
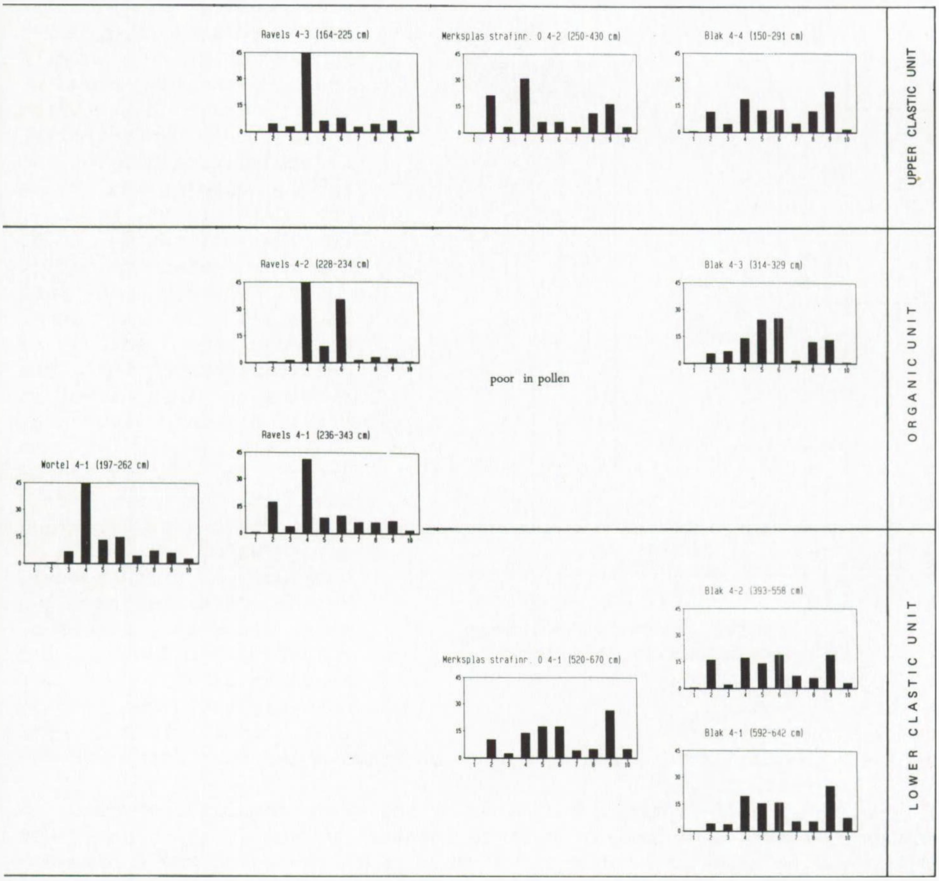


Fig. 3.24: North-south cross-section between Beerse Blak and Chaam Kapel. The bar diagrams with the physiognomical groups are arranged according to their lithostratigraphic position.

in the nearby absence of Chenopodiaceae. The northernmost sites (Chaam Kapel, Galderse Meren) appear to be further removed from the belt with floodplain forest (6) and carr vegetation (5). The increase in coniferous forest (8) seems to be an effect of selective enrichment with saccate pollen. Recent distribution patterns show an increase in *Pinus* pollen with increasing distance from the shore, especially in coarse-grained sediments (Chowdhury, 1982). Surface currents (Traverse and Ginsburg, 1966; Heusser and Balsam, 1977), high air current and resistance to destruction (Havinga, 1984) can all add to this phenomenon (cf. Chowdhury, 1982).

2. Organic unit

The organic unit in the Turnhout Member is registered at Beerse Blak (zone 4.3), Merksplas Strafinrichting (a sterile level), Wortel (zone 4.1), Meerle (zone 4.1) and Meerle Slikgat (zone 4.2, a sterile level and zone 4.3). The organic unit often consists of a humic clay or peatbed, which has been studied in closer intervals than the preceding and



overlying clastic sediments. Hence a more detailed picture of the environmental changes can be obtained.

Palynologically the organic unit shows up by an increase in groups 5 and 6, carr vegetation and floodplain forest. Fresh open water (group 3) and herbaceous shore vegetation (4) are firmly present. Sometimes a level, sterile in pollen, remains.

Interpretation:

The sequence at Meerle Slikgat, the most northerly investigated site with a clear peaty horizon, offers the opportunity to study the organic level in detail, since both the onset (zone 4.2) and the termination of the peat formation (zone 4.3) are present. Moreover macro remain analysis of the levels involved are available. Part of the peat-layer was poor in pollen.

Zone 4.2 (onset of the peat formation) shows low percentages for most of the groups, except for group 4 (shore vegetation) and 8 (coniferous forest). Group 3 is relatively high, indicating the presence of fresh open water. This is confirmed in the macro remain record (fig. 3.26) by the presence of *Potamogeton* cf. *alpinus*, *Scirpus* cf. *lacustris* , *Alisma plantago aquatica*, *Batrachium*, *Hippuris*, *Mentha aquatica*, *Salvinia*, *Typha* and undetermined Compositae species. The paleoenvironment can be characterized as shallow, fresh water, with weak currents.

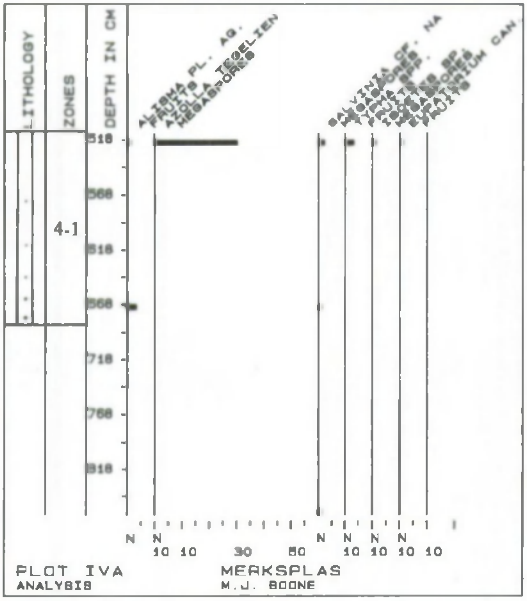
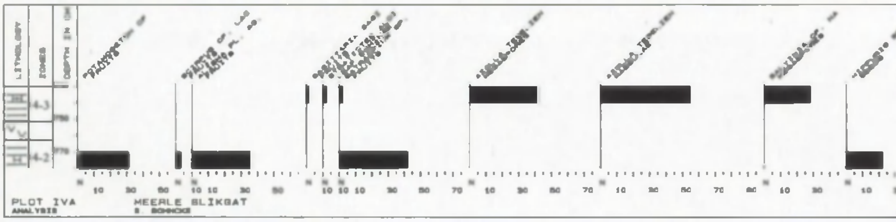


Fig. 3.25: Macro remain diagram of Merksplas Strafinrichting.

At a comparable level in site Meerle claypit the macro remains (fig. 3.27: 270-250 cm) indicate fresh, shallow standing water during this phase (*Sagittaria sagittifolia, Alisma plantago aquatica, Azolla tegeliensis, Typha sp., Salvinia cf. natans*). In the successive stage the open water area gradually declines and peat formation is initiated. Although the pollen is badly preserved the macro remains at this level in Meerle Slikgat (fig. 3.26: 7.58 m) reveal to some extent the species taking part in the peat formation. These are *Sparganium sp., Typha, Filipendula, Decodon* (Lythraceae), *Mentha, Carex rostrata* and other *Carex sp., Empetrum, Potentilla palustris* and *Menyanthes*. It is not unlikely that the peat at this level formed some sort of floating mat (quaking bog): an unsuitable substratum for the establishment of trees.

In view of the environment indicated by the macro remains, corrosion of the pollen must have been a post-sedimentary process, which has taken place during a period of minimum water depth preceding the subsequent transgression phase with deposition of the upper clastic unit. The peat-bed or humic clay extends southwards and is correlated with Ravels 4.2, Meerle clay-pit 4.1 and Blak 4.3. Here the organic unit appears as a zone dominated by floodplain forest (6) and carr vegetation (5). At Merksplas Strafinrichting soil forming processes have resulted in a sterile level as well. Macro remain analyses at Ravels did not confirm the local presence of *Alnus* (fig. 3.28: 2.32 m). Instead *Typha* dominates together with *Decodon, Juncus sp., Mentha aquatica, Hypericum cf. maculatum* and some megaspores of *Azolla tegeliensis*.

The overlying transitional zone 4.3 at Meerle Slikgat shows low values for all physiognomical groups, except for group 4 (shore vegetation), which is mainly due to the overwhelming amount of Monolete spores. In the macro remains (fig. 3.26: 7.35 m) this zone 4.3 is characterized by the abundant presence of water ferns e.g. *Salvinia cf. natans* (megaspores), *Azolla tegeliensis* (megaspores and massulae) and *Elatine hydropiper* (seeds), besides seeds of *Typha sp., Mentha aquatica, Carex rostrata* and *Carex sp., Batrachium* and leaf echinae of *Stratiotes*. In contrast to Meerle Slikgat zone 4.2, zone 4.3 reflects stagnant, fresh water conditions. The difference between zone 4.2 and zone 4.3 can be explained by differences in current velocity. At the beginning of the peat formation (zone 4.2) brackish water was replaced by slowly running fresh water. Zone 4.3 was formed during the beginning of a transgression (upper clastic unit), when a rising local water level inhibited



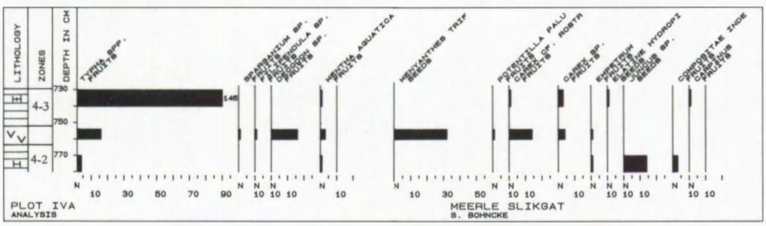


Fig. 3.26: Macro remain diagram of Meerle Slikgat.

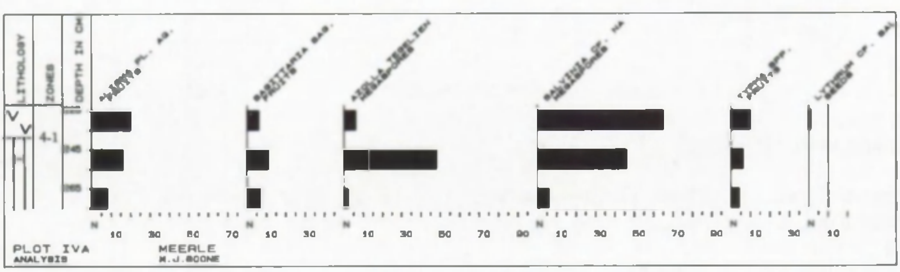


Fig. 3.27: Macro remain diagram of Meerle clay-pit.

the drainage and a stagnant, fresh water body built up, which subsequently became more brackish in the course of the transgression.

3. Upper clastic unit

This phase is present in bar diagram Blak 4.4, Ravels 4.3 and Merksplas Strafinrichting 4.2. The complexity of the lithological sequence more northwards does not allow a further correlation. Groups 2 and 4 (respectively tidal litter zones and herbaceous shore vegetation) are strongly represented. The amount of floodplain forest and carr vegetation (groups 5 and 6) is declining northwards (Merksplas 4.2) as does group 9 (heath).

Interpretation

At the time that the deposition of the upper clastic unit was at its maximum, tidal litter zones (group 2) reestablished at the southern shoreline of the intertidal basin (Blak 4.4, Merksplas 4.2 and to a lesser extent at Ravels 4.3). Tidal action resulted in a drift zone in the shore vegetation (group 4), which established in the high intertidal region at about mean high water. With an increased distance from the shore (Ravels 4.3) the influence of floodplain forest, carr vegetation (resp. groups 6 and 5) and heath (group 9) is declining.

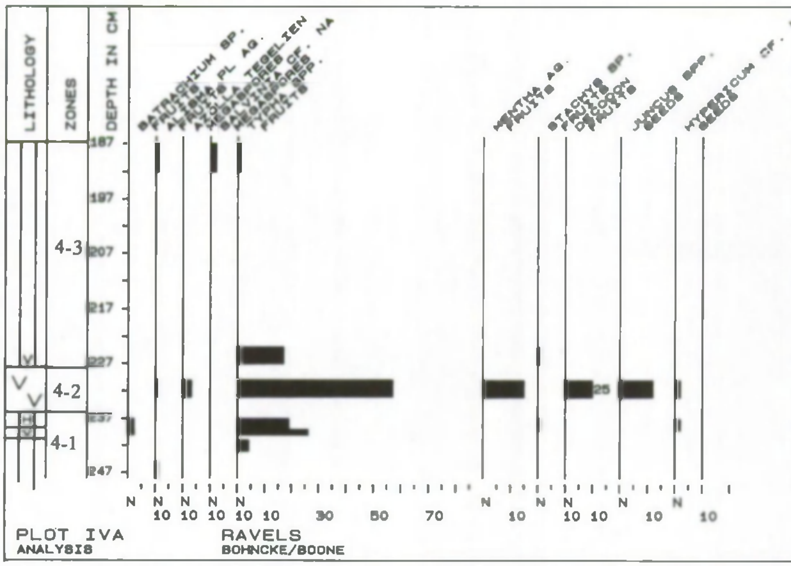


Fig. 3.28: Macro remain diagram of Ravels.

East-west transect (fig. 3.29)

Investigated sites: Appelenberg, Ravels, Zwart Water, Zwart Goor, Achtmaal, Wernhout Maalbergen and Kalmthoutse Hoek.

1. Lower clastic unit

The lower part of the Turnhout Member is represented by pollen zone Appelenberg 4.1, Zwart Water 4.1, Zwart Goor 4.1, Achtmaal 4.1 and 4.2, Wernhout Maalbergen 4.1 and Kalmthoutse Hoek 4.1 and 4.2.
The bar diagrams show the following trends: groups 4, 5 and 6 (resp. herbaceous shore vegetation, carr vegetation and floodplain forest) are well represented in all diagrams. From east to west the upland forest group 7 gradually declines. Open water species (group 3) and coniferous forest (group 8) show a concomitant increase.

Interpretation

The constant presence of the physiognomical groups 4, 5 and 6 possibly means that the east-west transect lies parallel to a paleo-vegetation zone consisting of a reed marsh associated type of vegetation (group 4) bordered by a carr and a belt with floodplain forest (groups 5 and 6). A westward decline in upland forest (group 7) presumably indicates that site Appelenberg in the southeast of the area lies most closely to this type of vegetation. More westward heath (group 9)is better represented (Kalmthoutse Hoek 4.2). The increase in coniferous forest (group 8) westward coincides with an increase in open water (group 3) and must be ascribed to the effects of water and air currents besides selective corrosion. The presence of open water is confirmed by the macro remain assemblage at Achtmaal (fig. 3.30).

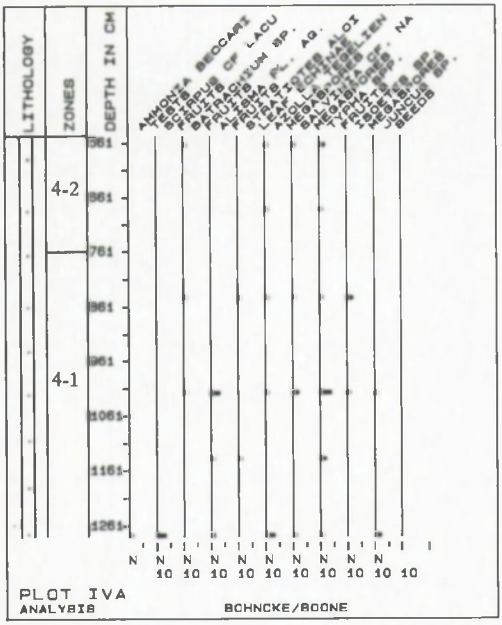


Fig. 3.30: Macro remain diagram of Achtmaal.

2. Organic unit

This phase is registered in Appelenberg 4.2, Ravels 4.1 (transitional) and 4.2 and Zwart Water 4.2.

The major characteristics of this zone are the dominating presence of groups 4, 5 and 6 (resp. herbaceous shore vegetation, carr and floodplain forest). All other physiognomical groups are only marginally present. The transitional phase at Ravels is characterized by groups 2 and 4 (tidal litter zones and herbaceous shore vegetation).

Interpretation

During the onset of the peat formation a belt with herbaceous shore vegetation intermingled with Chenopodiaceae (group 2) established at Ravels (zone 4.1). Possibly this belt moved northwards in the course of time and established at sites with temporarily emerging substratum. The fluvial environment enlarged its influence to the west. Macro remain analyses proved bad preservation conditions at Ravels.

Nevertheless the following species were found present (fig. 3.28: 2.38 m): *Batrachium sp.*, *Typha sp.*, *Hypericum cf. maculatum* and *Stachys cf. palustris*. The sediments from this period have been deposited under anaerobe conditions, as shown by the high pyrite content in the clay, which oxidized after the sampling (development of the yellow mineral jarosite). This may indicate that during the deposition of zone 4.1 at Ravels the tidal range was limited.

The absence of group 2 at the maximum of the peat formation indicates a freshening of the depositional environment. Floodplain forest and carr vegetation spread over the area (Appelenberg 4.2, Ravels 4.2, Zwart Water 4.2). Macro remain analyses of the peat-bed at Ravels could not prove the local presence of *Alnus*, but its nearby occurrence is clear from the pollen record. The local high content of group 4 pollen (reed marsh associated vegetation) is confirmed by the overwhelming amount of *Typha* seeds (fig. 3.28: 2.32 m).

3. Upper clastic unit

This phase is present in bar diagrams Appelenberg 4.3, Ravels 4.3 and Zwart Water 4.3.

Paleobotanical data are restricted to the southeastern part of the area. Group 4, the herbaceous shore vegetation, declines at Appelenberg but remains firmly present at Zwart Water 4.3 and dominates at Ravels 4.3. Tidal litter zones (group 2) are weakly present. Groups 8 (coniferous forest) and 9 (heath) increase somewhat.

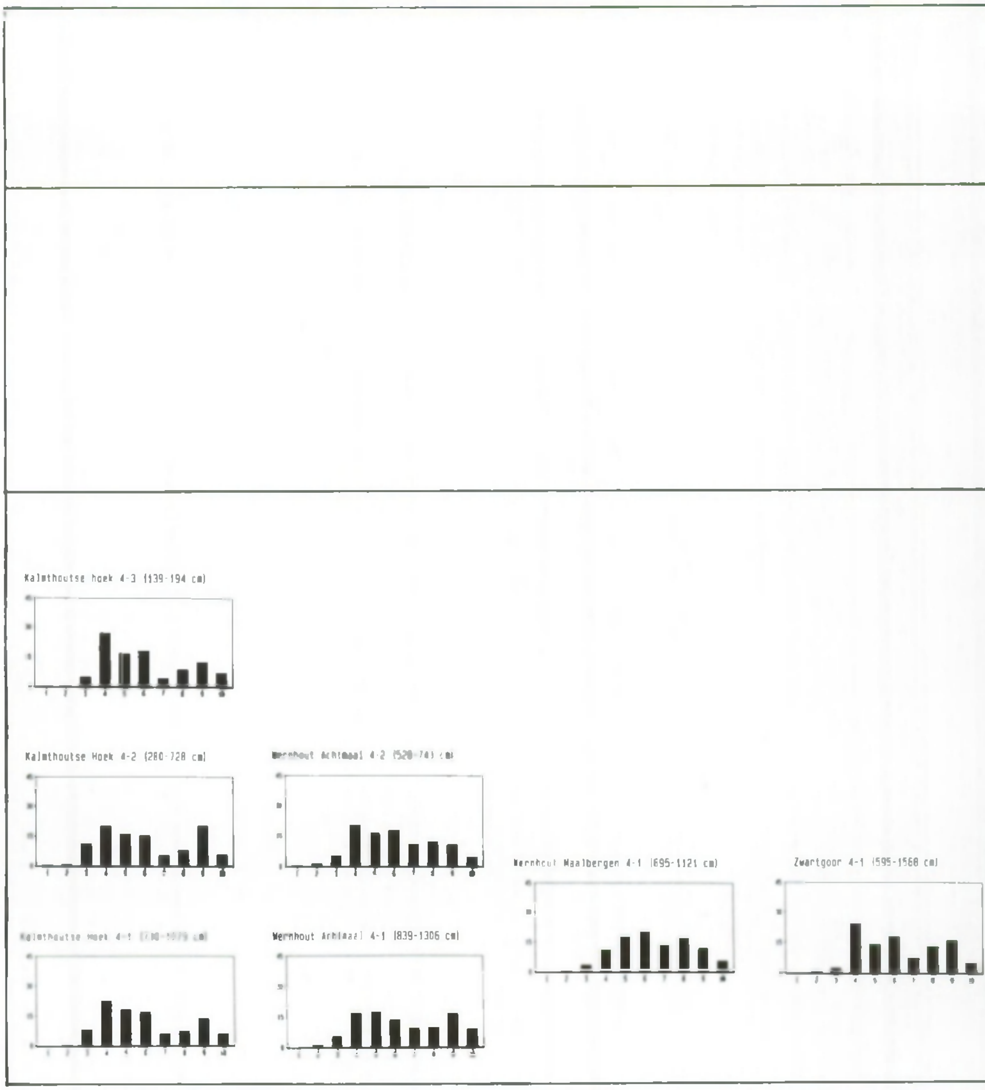
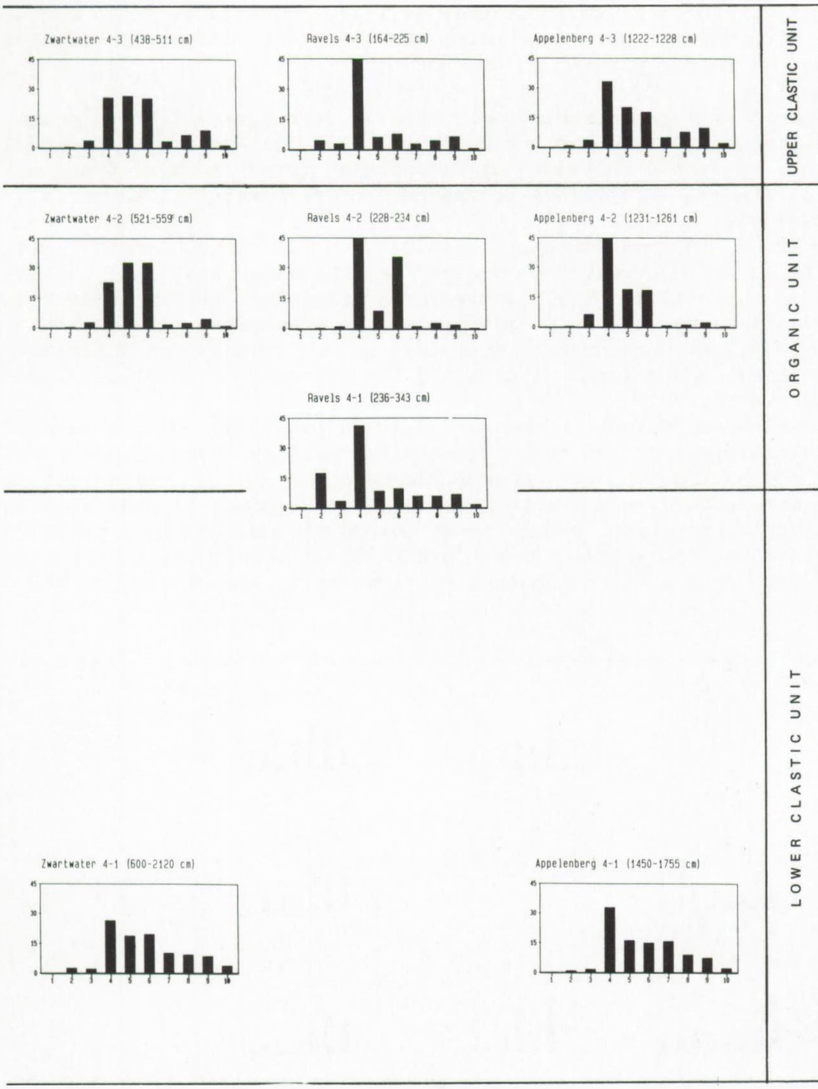


Fig. 3.29: East-west cross-section between Appelenberg and Kalmthoutse Hoek. The bar diagrams are arranged according to their lithostratigraphic position.

Interpretation

Only Ravels contains the transition from the organic unit to the over lying clastic unit. The transitional zone shows a drowning of the alder swamp forest, a short phase with strong increase of Gramineae (group 4) after which tidal litter zones reestablish (group 2: Chenopodiaceae). In the macro remains *Typha* seeds are present (fig. 3.28: 2.25 m). *Stachys cf. palustris* indicates, that the initial drowning occured under almost fresh water conditions. The pollen record even reveals the presence of *Myriophyllum sp.*
The influence of this transgression is hardly felt at Appelenberg.



Group 2 is found in very low values. At Zwart Water open water conditions increase somewhat (group 3) permitting a more regional registration of the pollen rain (increase of groups 7 and 8).

North - south transect in the western part of the area between Ossendrecht and Wouwse Plantage (fig. 3.31)

The following sites have been taken into consideration: Ossendrecht, KW7, WW9 (Korteven) and Wouwse Plantage.

1. lower clastic unit
Bar diagrams Ossendrecht 4.1, 4.2; Wouwse Plantage 4.1.
Main characteristics: The southern end of the transect at Ossendrecht is dominated by heath (group 9). The groups 5 and 6, carr vegetation

and floodplain forest, are relatively well represented at both sites. Group 4, the marshy shore vegetation, is important. Tidal litter zones (group 2) are mainly present at Ossendrecht.

<u>Interpretation</u>

At Ossendrecht the two bar diagrams show a development from a nearby marshy fringe vegetation (group 4) with tidal litter zones (group 2) towards an increased presence of floodplain forest (group 5 and 6). Apparently the tidal influence decreased and fresh water river environments developed.

The same trend, to some extent, is visible in the macro remains (fig. 3.32). The lower zone contains *Puccinellia* fruits, indicating the close presence of supra-tidal salt-marshes. It may not be excluded that *Puccinellia* invaded the reed marshes of the mesohaline zone (Gillham, 1957). Zone 4.2 at Ossendrecht contains macro remains of *Stratiotes* (leaf echinae) and *Alisma*, indicating the presence of shallow, standing, fresh water.

The high Ericaceae values in the bar diagrams (group 9) may be related to the close presence of the outcropping Tertiary substratum at the southern end of the transect. Wouwse Plantage, at the northern end of the transect, shows relatively more open water (group 3) and, associated with this, more coniferous forest (group 8) due to long distance transport. Also the litter zones (group 2) at this locality are less well represented due to the absence of temporarily emerging substratum.

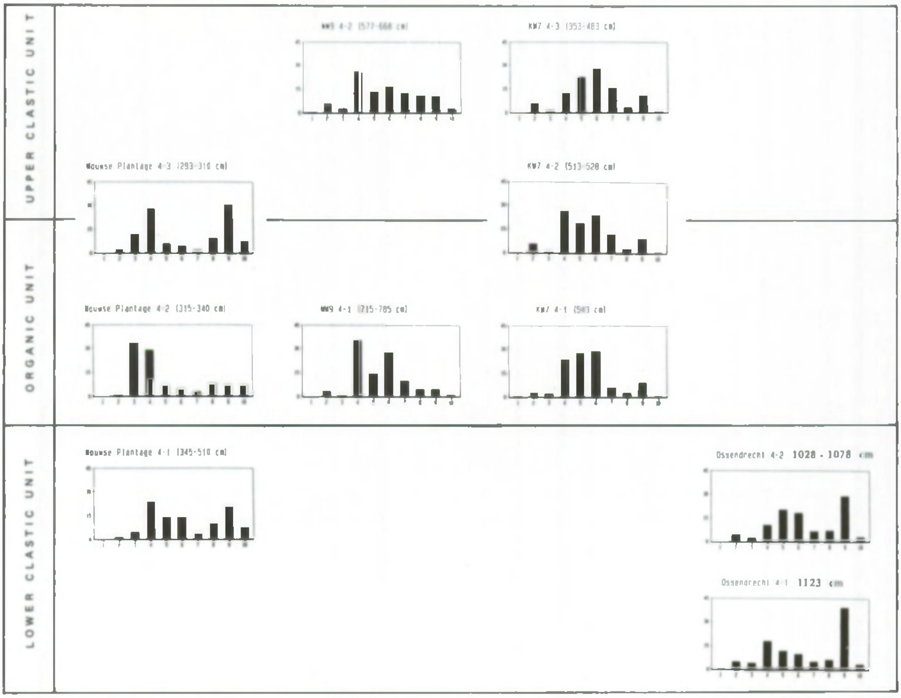


Fig. 3.31: North-south cross-section in the western part of the investigated area between Ossendrecht and Wouwse Plantage.

2. Organic unit

This unit is present in bar diagrams Wouwse Plantage 4.2, KW7-4.1 and WW9-4.1.
The main characteristics of this phase are the low presence of group 2 (tidal litter zones) and a northward shift from a belt dominated by groups 4 (herbaceous shore vegetation), 5 (carr vegetation) and 6 (floodplain forest) via an intermediate form to a belt with open, fresh water with herbaceous shore vegetation.

Interpretation

From south (KW7) to north (Wouwse Plantage) there is a transition from a floodplain forest (6) with carr vegetation (5) via a reed marsh dominated vegetation (4) with some floodplain forest elements (6) to an environment dominated by open water (3) bordered by herbaceous shore vegetation (4).
The open water environment contains macro remains of the following species (fig. 3.33, 3.34, 3.35): *Typha*, *Alisma plantago aquatica*, *Batrachium*, waterferns like *Callitriche sp.* and *Azolla tegeliensis*, *Juncus sp.* and *J. effusus*. In Korteven (WW9) badly preserved Chenopodiaceae seeds were found, which could not be identified to the species level.

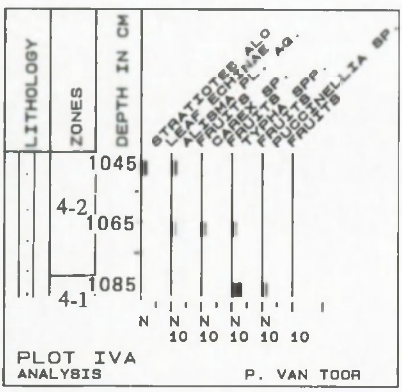


Fig. 3.32: Macro remain diagram of Ossendrecht.

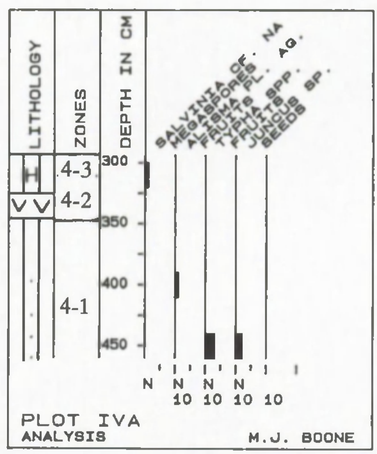


Fig. 3.33: Macro remain diagram of Wouwse Plantage.

3. The transition to the upper clastic unit

The transition to the overlying clastic phase is registered in KW7-4.2 and Wouwse Plantage 4.3, where the Chenopodiaceae content increases (group 2) and a herbaceous shore vegetation (group 4) with Gramineae and *Typha angustifolia* form the local elements. Alder swamp forest (group 5 and 6) is relatively low in the northern part of the cross-section but well represented in KW7-4.2, as is the upland forest (group 7). In Wouwse Plantage besides Ericaceae (group 9) the long distance transported pollen of the coniferous forest remain important.
Macro remains at KW7 (zone 4.2)(fig. 3.34) indicate the strong presence of open, standing water (*Potamogeton*, *Azolla tegeliensis*, *Typha*) and associated fringe vegetation (*Lycopus*, Cyperaceae, Gramineae). The presence of *Ammonia beccarii* points to periodical marine inundations.

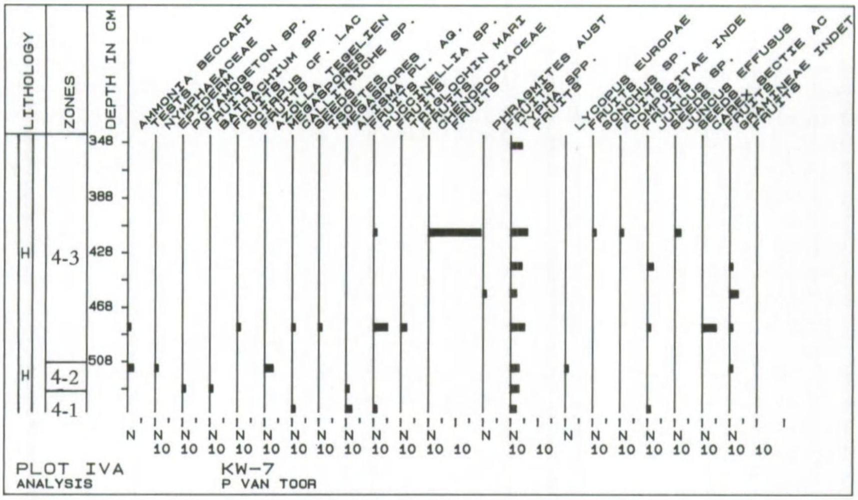


Fig. 3.34: Macro remain diagram of boring KW-7 (Huyzer and Van Toor, unpubl.).

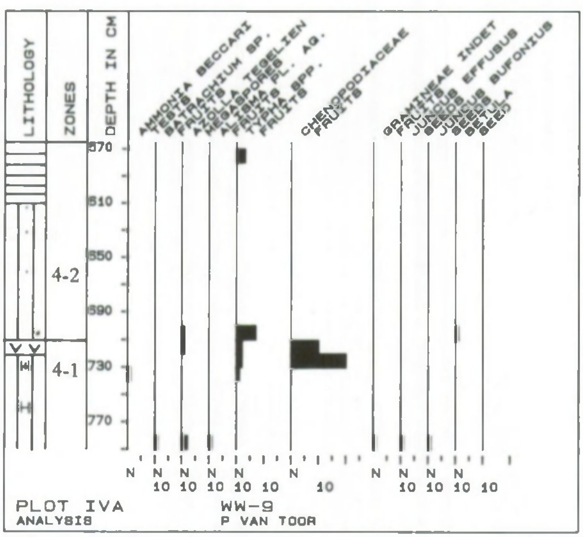


Fig. 3.35: Macro remain diagram of boring WW-9 (Korteven)(Huyzer and Van Toor, unpubl).

4. <u>Upper clastic unit</u>

This phase is registered in bar diagrams KW7-4.3 and WW9-4.2.
The tidal influence is shown by the presence of tidal litter zones (group 2) in both diagrams. Moreover, KW7-4.3 contains among others

fruits of *Puccinellia, Triglochin* and the foraminifer *Ammonia beccarii* (fig. 3.34). Also one fruit of *Phragmites australis* has been encountered, a species which was expected to be more frequently present in the macro remains. Other species present in the macro remains are *Scirpus sp., Callitriche, Isoëtes, Carex* sectie Acutae, *Typha, Juncus sp.* and *Sphagnum* leaf remains. WW9-4.2 contains *Typha* seeds (fig. 3.35). In conclusion the macro fossil assemblage represents a mixture from various environments. Moreover, the bar diagrams reveal that KW7 is located more closely to a belt with floodplain forest (group 6) and upland forest (group 7) in the south, than WW9.

3.4.4 Conclusions (fig. 3.36)

Towards the termination of the deposition of the lower clastic unit, tidal litter zones established in the understory of the upper zone of the intertidal marsh (fig. 3.36, phase a and b). These Chenopodiaceae rich tidal litter zones were located at the SE part of the investigated area. Northwards of the line Ravels-Merksplas Strafinrichting tidal action did not result in temporarily emerging substratum and tidal litter zones were absent. A belt with herbaceous marsh vegetation (predominantly *Typha*) bordering on pools was present.
In the western part of the area, at Ossendrecht, an intertidal marsh consisting of *Typha* intermingled with *Puccinellia* was present.
Within the onset of the peat formation, a freshening of the water took place allowing *Potamogeton, Hippuris, Batrachium, Salvinia* and *Alisma plantago aquatica* to establish in the shallow pools (Meerle Slikgat and Meerle) (fig. 3.36, phase c). Westwards *Stratiotes* and *Alisma* proved to be present (Ossendrecht).
At the maximum of the peat formation *Alnus* (groups 5 and 6) increased in the pollen record. Although macro remain analyses are very incomplete the analysed organic levels (Ravels, Meerle Slikgat, Meerle) never revealed macrobotanical remains of *Alnus*. Studies on the actuo pollen deposition in open systems with an inflow and an outflow (Peck, 1973; Bonny, 1976) showed a considerable supply of stream-borne pollen to the yearly pollen influx into lakes. For marine environments the highest pollen concentrations were found opposite to river mouths (Muller, 1959; Heusser and Balsam, 1977). It appears from the studies by Peck (1973) and Bonny (1976) that the water-borne component can attain values up to 90-97%, whereas the aerial deposition varies between 10 and 3% of the total. Moreover, amongst the pollen-types that show a differential input, *Alnus* occurs in significantly larger proportions in the stream-borne component (Bonny, 1976).
It is concluded that the increase of *Alnus* in the pollen diagrams results from an extensive zone with alder floodplain forest that established southeast of the investigated area. River systems running through this zone with alder floodplain forest are responsible for the large proportion of stream-borne *Alnus* pollen, deposited during the peat formation. Fruits and bud scales of *Alnus* show a different hydrodynamical behaviour and may have been trapped in the abundantly present intertidal marshes to the south and southeast. The organic level itself is characterized by an herbaceous vegetation intermingled with pools with fresh, standing water. These pools were vegetated by species like *Azolla tegeliensis, Salvinia cf natans, Alisma plantago aquatica, Sagittaria sagittifolia* and *Menyanthes* (fig. 3.36, phase d). The herbaceous marsh consisted of *Typha* spp. (dominant species), *Sparganium,*

Fig. 3.36: Tentative reconstruction of the paleobotanical environment in a north-south cross-section between Beerse Blak and Chaam Kapel. The following stages are depicted:
a. termination of the lower clastic unit.
b. transition to the organic unit.
c and d. maximal development of the organic unit.
e. transition to the overlying clastic unit.
f. development of the upper clastic unit.

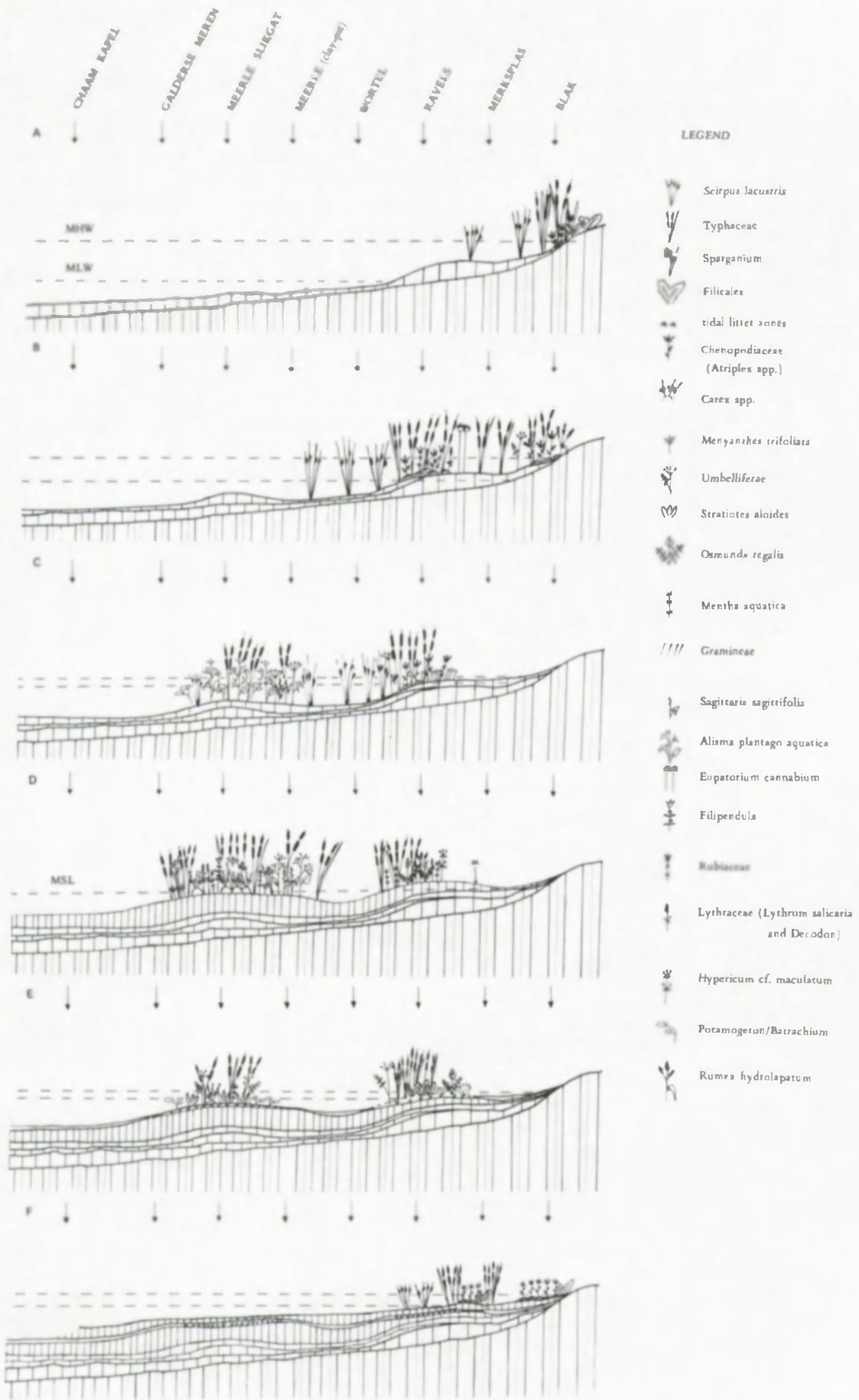
Mentha aquatica, Potentilla palustris, Carex rostrata, Carex sect. Acutae, *Carex* spp., *Filipendula, Decodon* and *Hypericum cf. maculatum*. At the western border of the area (Korteven (WW9), KW7) *Callitriche* and *Alisma plantago aquatica* are found besides *Typha* and *Juncus* seeds.
With the onset of the deposition of the upper clastic unit the area with fresh, open water increased (fig. 3.36, phase e). In the west (KW7) *Potamogeton* and *Azolla tegeliensis* established. At Meerle Slikgat (in the east) these species are accompanied by *Batrachium, Stratiotes, Salvinia cf natans* and *Elatine hydropiper*. The pools were bordered by *Typha* spp. (fig. 3.36, phase e). In the course of the transgression tidal litter zones reestablished in the SE part of the area (Ravels, Merksplas, Blak) probably under the influence of an increase in tidal range (fig. 3.36, phase f).
In the western part of the area (KW7) the intertidal marsh became invaded by *Triglochin, Puccinellia, Scirpus cf. lacustris* and *Phragmites australis*. Tests of *Ammonia beccarii* occur in the sediment. Apparently the intertidal marsh here lay within the then existing mesohaline zone.

3.5 Additional results concerning the depositional environments

3.5.1 Grain-size analysis

Grain-size characteristics have been used previously to define lithological units and sedimentary environments in the study area (Geys, 1975). The environmental conclusions based on grain-size parameters as presented by Geys are different from the results based on the interpretation of sedimentary structures in this study (§3.3). For this reason the sediments in the study area were first interpreted according to their lithostratigraphic position and sedimentary structures. Then the results of the grain-size analysis were plotted in diagrams in order to see whether the defined groups showed a certain coherence in grain-size parameters. The results are presented in the figures 3.37, 3.38 and 3.39.

<u>$<2\mu m/<16\mu m$ ratio and silt content:</u>
The $<2\mu m/<16\mu m$ ratio and silt content have been used successfully in the past to differentiate between fluviatile, marine and estuarine clay deposits (Zuur, 1936; Wiggers, 1955; Zonneveld, 1960; Van Straaten, 1964; Poelman, 1965).
Marine (tidal flat and salt-marsh) deposits were characterized by $<2\mu m/<16\mu m \times 100$ values between 65 and 70 (Zuur, 1936; Zonneveld, 1960). Fluviatile deposits showed somewhat lower $<2\mu m/<16\mu m$ values ranging between 55 and 65 (Zonneveld, 1960; Poelman, 1965). The fresh water tidal deposits of the Biesbos area revealed, however, much lower $<2\mu m/<16\mu m$ values, between 40 and 60 (Zonneveld, 1960; Poelman, 1965).


CHAAM KAPEL
GALDERSE MEREN
MEERLE SLIKGAT
MEERLE
WORTEL
RAVELS
MERKSPLAS
BLAK
A
MHW
MLW
B
C
D
MSL
E
F
LEGEND
Scirpus lacustris
Typhaceae
Sparganium
Filicales
tidal litter zones
Chenopodiaceae
(Atriplex spp.)
Carex spp.
Menyanthes trifoliata
Umbelliferae
Stratiotes aloides
Osmunda regalis
Mentha aquatica
Gramineae
Sagittaria sagittifolia
Alisma plantago aquatica
Eupatorium cannabium
Filipendula
Rubiaceae
Lythraceae (Lythrum salicaria
and Decodon)
Hypericum cf. maculatum
Potamogeton/Batrachium
Rumex hydrolapatum

Extreme, low <2µm/<16µm values (35-42) were found in fresh to slightly brackish lagoonal deposits (Wiggers, 1955: "sloef" deposits). Van Voorthuysen (1957, p. 51) pointed already to the similarity of the low <2µm/<16µm values in the "Halsteren Beds" (our Woensdrecht Member) and in the "sloef" deposits.

Poelman (1965) compared the (coarse) silt content (16-53 µm) of fluvial and fresh water estuarine deposits in the Land van Heusden en Altena (Central Netherlands). The fluviatile deposits commonly contain less than 35% coarse silt, while the estuarine deposits contain more than 35% coarse silt in 66% of the samples (N=219).

The low <2µm/<16µm values and high coarse silt content in fresh to brackish water tidal deposits was explained by peptisation of marine coagulated clay-flakes (Wiggers, 1955; Van Straaten, 1964). When a marine clay/silt/fine sand flake is transported from a salt into a brackish or fresh water environment, the change in salinity causes peptisation of the flake. The separate constituents of the flake are transported further and settle with current velocities corresponding to the particle size of the constituents. Clay-particles smaller than 2 µm only settle in the most quiet environments. Since silt has already settled previously, the <2µm/<16µm values increase in such sheltered, landward environments (Wiggers, 1955).

The fine-grained samples from clay- and silt-beds have been analysed with the pipette method and plotted according to their lutum (<2 µm) and lutum-fine silt (<16 µm) ratio (fig. 3.37) and to their coarse silt (16-53 µm) content (fig. 3.38).

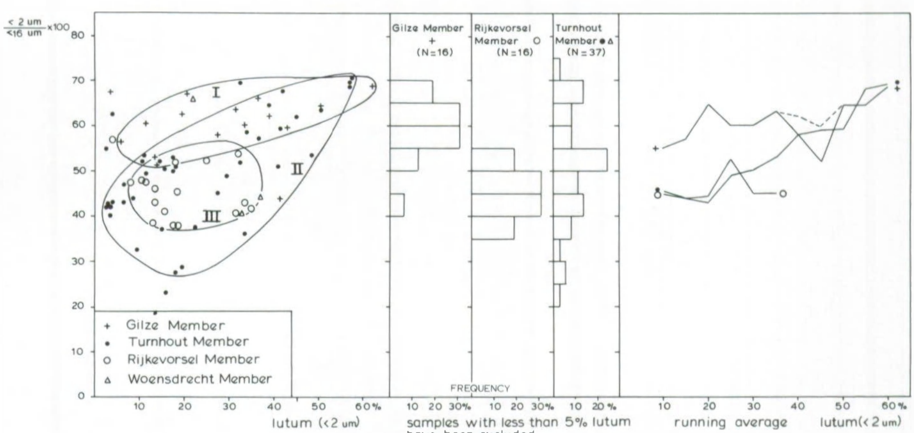


Fig. 3.37: Lutum/lutum-fine silt ratio of different lithostratigraphic units.

The <2µm/<16µm ratio varies between 23 and 70 (fig. 3.37). <2µm/<16µm values increase with higher lutum content. Three populations have been distinguished, corresponding to the Gilze, Turnhout and Rijkevorsel Members respectively.

Population I (Gilze Member) is characterized by <2µm/<16µm values between 55 and 65. Population II (Turnhout Member) reveals a very low <2µm/<16µm ratio (45) when the lutum content is low (10-25%) and is clearly different from population I. However, with lutum content increasing the <2µm/<16µm ratio of population II increases to values

which are comparable to the <2µm/<16µm values of population I (circ. 70). Population III (Rijkevorsel Member) is a subpopulation of group II, with <2µm/<16µm values between 40 and 50.
The (coarse) silt content (16-53 µm) of the samples (fig. 3.38) varies between 5 and 80%. The three populations show much overlap. Population I (Gilze Member) has a relatively low silt content (less than 40%). Population II (Turnhout Member) is characterized in general by a very high coarse silt content (40-75%) when lutum values are below 20%. Population III (Rijkevorsel Member) shows intermediate values between 40-60%, when lutum content is lower than 20%.

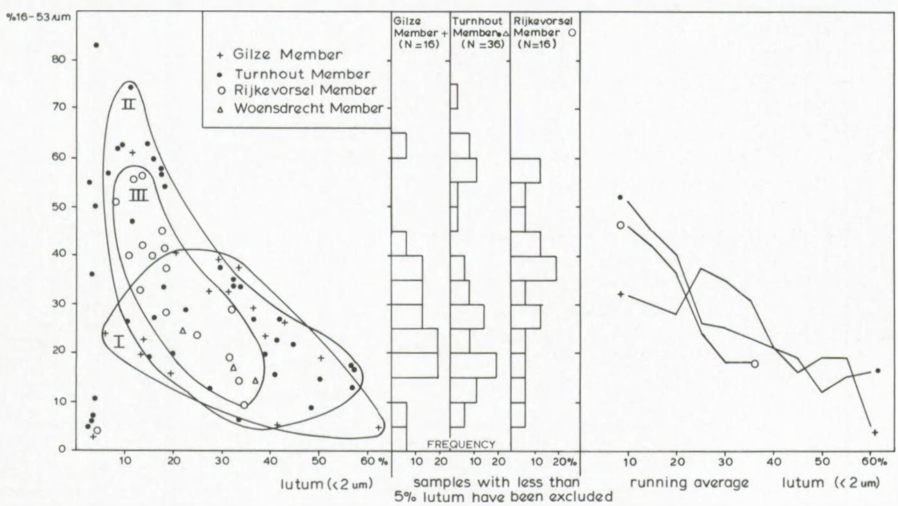


Fig. 3.38: Coarse silt content of different lithostratigraphic units.

Comparison of the <2µm/<16µm values and coarse silt content of the samples in the study area with former investigations leads to the following conclusions:

1. The high <2µm/<16µm ratio (55-65) and low coarse silt content (less than 40%) of the Gilze Member (population I) are comparable with subrecent fluviatile sediments (Poelman, 1965).
2. When lutum content is low, the low <2µm/<16µm ratio (40-55) and high coarse silt content (more than 40%) in both the Turnhout (population II) and Rijkevorsel Members (population III) are comparable to subrecent, fresh water, tidal deposits in the Biesbos area (Zonneveld, 1960). The high ratio (60-70) in samples with high lutum content (more than 45%) is not interpreted as an indication of salt water deposition. It reflects deposition of peptisized suspended clay in the most sheltered, landward, fresh to brackish water, tidal environments (e.g. tidal litter zone).
3. The <2µm/<16µm ratio (40-55) of the tidal deposits of the Turnhout and Rijkevorsel Members are higher than the <2µm/<16µm ratio (35-42) in fresh to brackish water, lagoonal deposits ("sloef")(Wiggers, 1955).
4. The low <2µm/<16µm values and the high coarse silt content in the brackish to fresh water, tidal deposits of the Turnhout and Rijkevorsel Members, is explained by peptisation of coagulated flakes during their landward transport from salt marine to brackish and fresh water tidal environments (Van Straaten, 1964).

<u>Mean, standard deviation and skewness:</u>
The coarser grained samples from sand-beds have been analysed by the dry sieving method and plotted according to their mean values, standard deviation (sorting) and skewness (fig. 3.39). The fractions coarser than 53 µm (4.25 phi) have been sieved in quarter phi intervals. The grain-size classes finer than 53 µm have not been investigated separately. The skewness and standard deviation have been calculated for a grain-size distribution above 2 µm (9 phi).

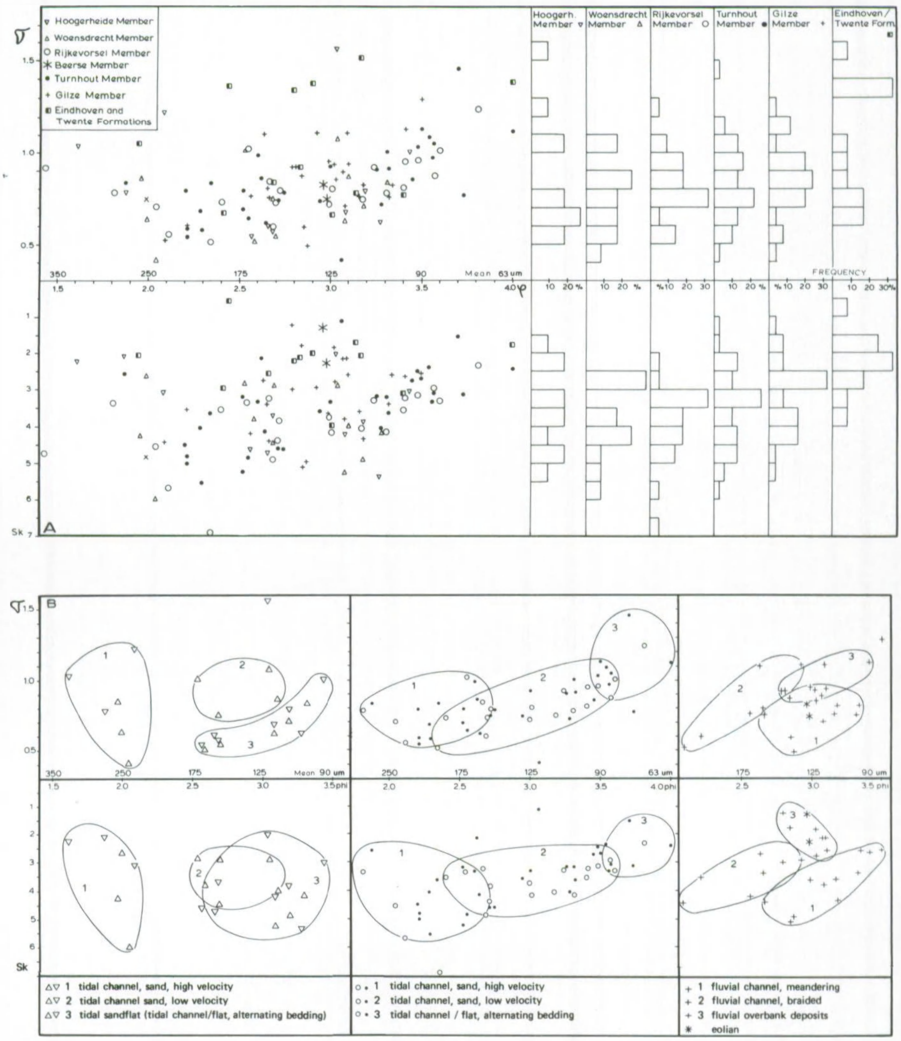


Fig. 3.39: Mean, skewness and standard deviation of coarse-grained samples in connection with the depositional environments.

Some general trends can be seen in the diagram (fig. 3.39A: left side):
1. The standard deviation increases (poorer sorting) with decreasing grain-size, which can be explained by a mixture of sediment in the fine sand samples (e.g. alternations of sand-silt-clay).

2. The generally positive skewness tends to zero with decreasing grain-size, because a fine-grained tail (positive skewness) is more important in coarser grained samples, than in finer grained samples.
3. The various members and formations do not form distinct populations, according to their grain-size parameters.

The frequency distributions of the skewness and the standard deviation of the individual members show that (fig. 3.39A: right side):
1. Sorting is good (standard deviation: 0.75) and skewness is highly positive (Sk = 3.25) for the tidal deposits of the Hoogerheide, Woensdrecht, Rijkevorsel and Turnhout Members.
2. Sorting is slightly poorer (sigma: 0.85) and the skewness is slightly lower (Sk = 2.75) in the fluviatile Gilze Member.
3. Sorting is poor (standard deviation: 1.35) and skewness is low (Sk = 2.25) in the Eindhoven and Twente Formations, which were formed by small, fluvial channels and surficial runoff.

In order to characterize the sedimentary sub-environments of the members the grain-size characteristics were plotted for each member or genetically related members (fig. 3.39B). The following conclusions can be drawn:
1. The sub-environments in most of the members are characterized more by their mean grain-size values, than by their standard deviation and skewness. However, some exceptions occur:
- The sandflat deposits of the Hoogerheide and Woensdrecht Members are generally better sorted and have a higher skewness than channel deposits of equal grain-size.
- The tidal channel/mixed flat deposits of the Rijkevorsel and Turnhout Members have a relatively poor sorting, because of their alternating sand-clay bedding.
- The fluvial overbank deposits of the Gilze Member have a poorer sorting and especially a lower skewness, than channel sediments with a comparable grain-size.
2. The sediments of sub-environments in the Hoogerheide and Woensdrecht Member are, in general, coarser grained than the sediments of corresponding sub-environments of the Turnhout and Rijkevorsel Members. This difference has been explained previously (chapter 3) by the more seaward location of the Hoogerheide and Woensdrecht Members in comparison with the more sheltered, landward position of the tidal environments of the Turnhout and Rijkevorsel Members.

Finally, some general conclusions can be formulated, based on the grain-size analysis of 185 samples:
1. The $<2\mu m/<16\mu m$ ratio and coarse silt content enable differentiation between clay-beds of a fluviatile and brackish to fresh water tidal origin, especially when the lutum content of the samples is low (between 5 and 45%).
2. The skewness and standard deviation are in general inappropriate to distinguish environments, although sandflat, fluvial overbank and local surficial runoff deposits seem to be characterized by specific skewness and/or sigma values.
3. The mean grain-size values of the samples is the most important factor to separate sub-environments.
4. The combination of sedimentary structures, the mean grain-size, the skewness and the standard deviation gives a good characteristic of the sub-environments.

3.5.2 Clay-mineralogy

Clay-mineralogical data from the Early-Pleistocene deposits in Noord-Brabant are scarce. Breeuwsma and Zwijnen (1984) investigated the clay-mineralogical composition of the Tegelen, Sterksel and Veghel Formations in Noord-Brabant. The Tegelen Formation is dominated by illite and smectite clay-minerals. The Sterksel and Veghel Formations are also dominated by illite, with chlorite and illite-chlorite content higher than in the Tegelen Formation. According to Breeuwsma and Zwijnen (1984), the differences in clay-mineralogical composition between the formations probably reflect different sediment sources: the Tegelen sediment was supplied by the Rhine, whereas the Sterksel and Veghel deposits were supplied by the Meuse. The supposed Meuse origin of the Sterksel Formation is only parly confirmed by the gravel and heavy mineral analysis (chapter 5).

3.5.3 Cation exchange capacity

According to Breeuwsma (1985) cation exchange capacity (CEC) can be related to the depositional environment. The CEC in Holocene river clay was found to be higher than in marine clay. In order to ascertain whether comparable differences are also present in Early-Pleistocene deposits in the study area, 8 samples from various members have been analysed. The results are presented in table 3.1 (values between brackets are percentages).

Table 3.1: Cation exchange capacity of Early-Pleistocene deposits in Noord-Brabant.

Location	Depth (in m)	Member	in meq/100 gr dry soil K	Na	Ca	Mg	Total	Ca in Mg extr.	CEC corr.	Environment based on sedimentary structures	
Bavel	10.1	Bavel	0.100	0.118	10.16	1.03	11.41	0.82	0.95	fluviatile	+++
			(0.9)	(1.0)	(89.0)	(9.0)					
Gilze	3.65	Gilze	0.178	0.199	3.97	3.78	8.13	1.32	7.56	fluviatile	-
			(2.2)	(2.4)	(48.8)	(46.5)					
Ravels	2.50	Turnhout	0.083	0.034	4.17	-	-	2.88	1.50	tidal litter zone	-
Meerle Slikgat	9.42	Turnhout	0.166	0.174	4.22	1.56	6.12	1.24	5.24	fresh-brackish tidal	-
			(2.7)	(2.8)	(69.0)	(25.5)					
Achtmaal	8.12	Turnhout	0.380	0.171	12.62	3.51	16.68	1.43	14.34	fresh-brackish tidal	±
			(2.3)	(1.0)	(75.7)	(21.0)					
Meerle Slikgat	21.3	Rijkevorsel	0.159	0.076	8.02	1.86	10.12	0.55	8.29	brackish tidal	+
			(1.6)	(0.8)	(79.2)	(18.4)					
Achtmaal	19.4	Rijkevorsel	0.135	0.118	3.15	2.82	6.22	1.20	4.55	brackish tidal	-
			(2.2)	(1.9)	(50.6)	(45.3)					
Zwart Goor	19.4	Rijkevorsel	0.132	0.075	6.64	0.74	7.59	0.81	6.89	brackish tidal	++
			(1.7)	(1.0)	(87.5)	(9.7)					

The CEC values for Bavel and Ravels are probably erroneous, since they are lower than the Ca content. The CEC of the rest of the samples is highly variable. The samples from fluviatile, fresh water tidal and brackish tidal environments do not form distinct CEC groups.
Calcareous (+), fluviatile sediment (Bavel) and calcareous, brackish tidal sediment (Zwart Goor) have a comparable high Ca and low Mg content. Non-calcareous (-), fluviatile (Gilze) and brackish tidal (Achtmaal: 19.4 m) deposits are characterized by a high Mg and relatively low Ca content. Samples with intermediate carbonate content (±) have intermediate Mg and Ca values. Na content is low in all samples.

The values are characteristic for a fresh ground water system (pers. comm. Dr. C.A.J. Appelo). This does not neccessarily imply a fresh water depositional environment, since the cation composition may have changed after the deposition. If the Na content had been high in some samples it would have indicated a salt water origin. Since the samples have low Na values it is impossible to differentiate on this basis between primary fresh water deposits and primary salt water deposits, in which the salt formation water was later replaced by fresh ground water. The deep decalcification in some members (Gilze, Turnhout, Rijkevorsel) points to intense percolation of ground water. It seems likely that the recent cation composition is different from the original one during deposition. Therefore, no conclusions can be drawn concerning the depositional environment based on the recent cation composition.

3.5.4 Deer antlers

During the last century several finds have been reported of deer antlers in the Campine Clay Formation (Rijkevorsel and Turnhout Members) in Belgium. They belong to the species Eucladoceros tegulensis, E. falconeri and Cervus s.l. incertae sedis (=Cervus rhenanus?)(Germonpré, 1983). Unfortunately the precise stratigraphic position of the finds is often unknown. One fragment of Eucladoceros tegulensis was found in the Turnhout Member. According to Germonpré (1983) the antlers are always well preserved. They do not show traces of alteration or abrasion. Therefore, the antlers probably occurred in situ in the Campine Clay Formation and they are not reworked from older deposits.
The Rijkevorsel and Turnhout Members are interpreted as landward, tidal flat and tidal litter zone deposits (§3.3.2 and §3.3.4). Since the antlers appear to occur in situ in the deposits, it is concluded that the deer probably lived or grazed in the tidal litter zones. They lost their antlers, which were subsequently buried by new sediment during spring tides. During later decalcification of the Campine Formation calcareous material disappeared, but the horny antlers (and an elephant tooth reported by Van Straelen, 1920) remained intact.

3.6 Discussion and summary of the sedimentary environments

In this paragraph, the sedimentological results and other evidence are summarized and evaluated, against the background of former ideas, concerning the depositional environments of the Early-Pleistocene deposits in The Netherlands and Belgium (see table 2.1).
The Rijkevorsel Member, which belongs to the Tegelen Formation and Campine Clay and Sand Formation, is characterized by regular and frequent alternations of sand and clay (lenticular and flaser bedding) in combination with bidirectional cross-bedding. These structures point to a tidal environment. The large-scale sedimentary sequence is often characterized by a coarsening-upward at the base of the member, followed by a fining-upward sequence. Since occasionally continental (peat) deposits are present below/at the base of the Rijkevorsel Member, the coarsening-up is interpreted as the drowning of the area. The most sandy part in the middle of the Rijkevorsel Member reflects the maximal tidal influence during the transgression. The fining-upward at the top of the Rijkevorsel Member indicates inshore silting of the tidal area, as offshore progradation would, on the contrary, be reflected by a coarsening-upward sequence. The general clayeyness of the

Rijkevorsel Member is explained by low energetic conditions and slow migration of the tidal channels, which is characteristic for landward (distal) parts of tidal environments. Salinity of the paleoenvironment is difficult to establish, since molluscs and diatoms are absent. However, bioturbation is scarce, which points to a limited biological activity, probably due to a low salinity. Tidal range was estimated between 0.95 and 2.1 m, uncorrected for compaction.
To the west grain-size increases in the Hoogerheide Member (Tegelen Formation), which is characterized by large-scale cross-bedding, flaser bedding and herringbone cross-bedding. The coarser grain-size and smaller thickness of the clay-layers are explained by active tidal channel migration. Hummocky cross-bedding at the top of the Hoogerheide Member points to wave activity. These characteristics indicate a more seaward (proximal), inshore, tidal environment with respect to the Rijkevorsel Member.
In the southeastern part of the investigation area around Beerse the Rijkevorsel Member is covered by the Beerse Member (Campine Clay and Sand Formation, Tegelen Formation). In The Netherlands the Beerse sands are always absent, due to erosion prior to deposition of the Turnhout Member. Sedimentary structures are dominated by dry eolian, parallel horizontal and low-angle cross-bedding, crinkly wet eolian adhesion bedding and contorted bedding. Superficial runoff was locally present. There is a fair resemblance to the bedding types distinguished by Ruegg (1983) and Schwan (1986) in Weichselian eolian sand-sheets. The periglacial structures (frost cracks, ice-wedge casts, cryoturbations) indicate cold climatic conditions with a mean annual temperature of approximately -5° C during the deposition of the eolian sand-sheet deposits of the Beerse Member.
The Beerse Member is covered by the Turnhout Member (Campine Clay and Sand Formation, Tegelen Formation). The Turnhout Member has many sedimentary characteristics in common with the Rijkevorsel Member. The regular alternations of sand and clay, the bidirectional cross-bedding, the fining-upward sequences, the general clayeyness of the sediment, the absence of bioturbations and the presence of pyrite point to landward (distal), inshore tidal, fresh to brackish environments. A tidal range of 1.05 m was estimated at Meerle. Tidal litter zones, characterized by Chenopodiaceae species, occurred at the southern fringe of the tidal environment. The Turnhout Member can easily be correlated with the coarser-grained Woensdrecht Member (Tegelen Formation) in the west. The larger grain-size, the abundance of clay-pebbles, the large-scale bidirectional cross-bedding with neap-spring-neap tidal cycles, the wave-induced coarsening-up at the top of tidal flat deposits and the presence of faint bioturbations are indications for a seaward (proximal), inshore (estuarine) tidal, brackish environment. Tidal range was more than 1.2 to 1.7 m.
The depositional environments of the Rijkevorsel and Turnhout Members are essentially in agreement with previous Belgian investigations. Dricot (1961), De Ploey (1961) and Paepe and Vanhoorne (1970) all stressed the tidal character and interpreted the sediments as tidal flat and salt-marsh deposits. They probably over estimated the salinity of the environment, since the sediments were described as salt-marsh ("schor") deposits. In our opinion, however, the upper parts of the Rijkevorsel and Turnhout Member in Belgium have been formed in tidal litter zones or brackish to fresh water marshes in which *Typha* was important. In The Netherlands the Rijkevorsel and Turnhout Members (Tegelen Formation) have been interpreted as fluviatile deposits of the Rhine and the Meuse (Doppert and Zonneveld, 1955; Van Dorsser, 1956; Van Voorthuysen, 1957; Zagwijn and Van Staalduinen, 1975), probably

because the sediments had to be studied in borings and did not contain marine shells and diatoms. Only when marine molluscs were found in the Rijkevorsel Member around Breda, was the tidal character recognised and the sediment incorporated into the marine, shell-bearing Maassluis Formation (Zagwijn and Van Staalduinen, 1975). The sandy deposits of the Woensdrecht Member in western Noord-Brabant have been misinterpreted as a braided or meandering river deposit (Van Dorsser, 1956; Van Voorthuysen, 1957: Halsteren Beds; Van Meurs, unpubl.; Damoiseaux, 1982), probably because of the dominant, westnorthwest directed stream direction, in the estuarine deposits. The stable to mixed heavy mineral association is not an indication of fluvial deposition by the Meuse (Van Dorsser, 1956), since it can be explained by estuarine processes (§5.4.4).

Up till now sedimentological information concerning the Kedichem Formation was scarce. A fluviatile, fluvio-periglacial and eolian origin had been proposed, but no details were given (Zonneveld, 1958; Zagwijn and Van Staalduinen, 1975; Bisschops et al., 1985). The recent investigations reveal that the Gilze Member, which is part of the Kedichem Formation, is a complex unit, which has been formed in several depositional environments. The rapid alternations of sand, loam and peat-layers in the lower unit (Appelenberg Sands) may be attributed to a fluvial environment with small river channels, overbank deposits, backswamps and lakes. The large-scale, fining-upward sequences with extensive clay-layers at the top of the second unit (Gilze Clay) are interpreted as floodplain deposits of a meandering river, consisting of channel and backswamp sediments. The coarser grained upper unit (Alphen Sands) with large-scale, trough shaped, cut-and-fill structures is interpreted as a sandy, braided river deposit (or deposits of ephemeral streams), which is in agreement with Vandenberghe and Krook (1981) and Vandenberghe et al. (1986). The Gilze Member occurs close to the surface west of the Gilze-Rijen and Hooge Mierde fault. It is correlated with the St. Lenaarts Formation at Ravels on the Campine microcuesta in Belgium (De Ploey, 1961). The large-scale, fluvial character of the sediments was under estimated by De Ploey, when he interpreted the St. Lenaarts Formation as an eolian deposit reworked by fluvial processes.

The Bavel Member (Kedichem Formation) was deposited by a large meandering river in a more than 10 m deep channel, incised into the Gilze Member. Thick, massive and laminated, calcareous clays were interpreted as the infilling of meander cut-offs. Large-scale, trough cross-bedded sand was deposited in fluvial channels by southwesterly directed currents (Zagwijn and De Jong, 1984: Bavel I, Ia).

The younger formations, belonging to the Middle- and Late-Pleistocene, are not discussed here (see §3.3.9).

4. CHRONOSTRATIGRAPHY

4.1 Introduction

The age of the Early-Pleistocene deposits in Noord-Brabant and northern Belgium has been investigated previously in several studies (Van der Vlerk and Florschütz, 1953; Tavernier, 1954; Van Dorsser, 1956; Dricot, 1961; De Ploey, 1961; Paepe and Vanhoorne, 1970; Van Montfrans, 1971; Geys, 1975; Hus et al., 1976). The conclusions, however, were not very coherent (see §4.2), since a Tiglian up to a Cromerian age has been proposed. Furthermore, equivalent deposits in The Netherlands and Belgium were correlated with different chronostratigraphic stages. Finally, biostratigraphical and paleomagnetic methods were used in combination only sporadically.

The present study therefore aimed at:

1. Accurately establishing the chronostratigraphic position of the various lithostratigraphic units with the highest possible resolution, in order to enable reliable paleogeographic reconstructions.
2. Correlating the chronostratigraphy in The Netherlands and Belgium.
3. Integrating magnetostratigraphic and bio-climatostratigraphic evidence.

In The Netherlands the Pliocene-Pleistocene boundary is commonly placed at the base of the Praetiglian glacial, 2.5 m.y. ago (Zagwijn, 1975b). The age of this boundary is still a matter for discussion, as it has also been dated at 1.6 m.y., at the top of the Olduvai magnetozone (Haq et al., 1977; Jenkins, 1987). However, in the study area cold periods, older than 1.6 m.y. and characteristic for the Quaternary are present. In accordance with Zagwijn these cold phases are included in the Pleistocene and the Pliocene-Pleistocene boundary is therefore maintained at 2.5 m.y.

As in other parts of The Netherlands and Belgium the Early-Pleistocene deposits in the area cannot be dated by absolute dating techniques. Therefore, the age of the deposits was established by pollen analysis, some guide fossils, the magnetic polarity and to a certain extent the lithostratigraphic position. The results have been correlated with the existing biochronostratigraphic time scale of The Netherlands (Zagwijn, 1986).

The pollen assemblages of Quaternary deposits in The Netherlands reveal an alternation in time of warm-temperate and cold climatic conditions, which define successive interglacials and glacials (Zagwijn, 1986). The pollen analytical investigations show an intricate climatic evolution. Several stages, which were formerly considered as interglacials, later appeared to be complexes of warmer and cooler periods (Zagwijn, 1957, 1960, 1971, 1974; Zagwijn and De Jong, 1984). The Early-Pleistocene warm-temperate phases are all characterized paleobotanically by low percentages of Early-Pleistocene species like *Pterocarya, Carya, Tsuga* and *Eucommia*. Tertiary floral elements such as *Nyssa* and *Sciadopytis* are absent or reworked from underlying deposits (Zagwijn, 1975b).

The geomagnetic method offers the opportunity to establish a magnetostratigraphic time scale. The latter is based on the recorded reversals of the earth magnetic field and associated magnetozones, which are correlated with the well established polarity time scale of the Quaternary (Mankinen and Dalrymple, 1979). During the Quaternary at least 8 reversals are known with certainty (Hus, 1988). The combination of the pollen record, the magnetostratigraphy and lithostratigraphy, results in a more complete chronological framework for the geological evolution in the investigated area.

4.2 Historical review

Since the start of the geological investigations in the Dutch-Belgian border area in the late 19th century, the age of the deposits has been a matter for discussion (see table 2.1 for the lithostratigraphic nomenclature). An extensive review has been given by Geys (1975).
A Pleistocene (Tiglian) age has been proposed for the Campine/Tegelen Formation by Van der Vlerk and Florschütz (1953) and Tavernier (1954), on the basis of paleobotanical similarities with the Tiglian deposits in the Tegelen type area in the Dutch province of Limburg. The underlying Mol Sands (and Merksplas Sands) were finally dated as of Tertiary age (Vanhoorne, 1962). A Tiglian or Taxandrian age ("Kedichem Series") has also been proposed for the Early-Pleistocene deposits close to the surface in Noord-Brabant (Nelson and Van der Hammen, 1950; Doppert and Zonneveld, 1955).
In 1957 Zagwijn introduced a more detailed subdivision of the Early-Pleistocene in The Netherlands, based on pollen analysis. In succession to the Tiglian, he defined the Eburonian, Waalian and Menapian periods, which correspond respectively to a glacial, interglacial and glacial climate.
Paepe and Vanhoorne (1970, 1976) later applied this Dutch chronostratigraphic subdivision in Belgium. The Rijkevorsel Member was connected with the Tiglian period, because of the presence of *Azolla tegeliensis* (Greguss and Vanhoorne, 1961). The overlying Beerse and Turnhout Members were correlated with the Eburonian and Waalian periods respectively, based on the observation of periglacial phenomena in the Beerse Member. A paleomagnetic reversal from reversed to normal in the upper part of the Turnhout Member was interpreted as corresponding to the base of the Jaramillo magnetozone (Van Montfrans, 1971; Paepe and Vanhoorne, 1970). Some discrepancies developed when Zagwijn (1975a, 1979) located the Waalian shoreline west of the recent Dutch coast, whereas Paepe and Vanhoorne (1970) described tidal flat and marsh deposits of Waalian age in Belgium, as far east as Turnhout (Turnhout Member).
On the top of the Campine/Tegelen Formation, sands with a stable heavy mineral composition are predominanty present (§2.5.8: Gilze Member). In Belgium these sands have been dated as Early- and Middle-Weichselian (De Ploey, 1961: St. Lenaarts Formation; Haest, 1985; Haest et al., 1986). The St. Lenaarts Formation has been continued in The Netherlands in the so-called Alphen Sands (§2.6), which have been interpreted as probably Menapian (Vandenberghe and Krook, 1981; Vandenberghe et al., 1986).
In 1984 Zagwijn and De Jong defined a new chronostratigraphic stage (Bavelian) in the Early-Pleistocene, which comprises two interglacials and two glacials. Deposits of Bavelian age were identified at Bavel, in the Central Graben, in terrace deposits of the Meuse and in the Central Netherlands.

Based on the review given above the following questions arise:
1. In which phase(s) of the Tiglian was the Rijkevorsel Member formed (Greguss and Vanhoorne, 1961; Paepe and Vanhoorne, 1970)?
2. Is the Beerse Member of Eburonian/Menapian age and the Turnhout Member of Waalian/Cromerian age (Dricot, 1961; De Ploey, 1961; Paepe and Vanhoorne, 1970, 1976)? How to explain in that case the occurrence of perimarine deposits in Belgium (Turnhout Member) and fluviatile deposits in The Netherlands (Kedichem Formation; Zagwijn and Van Staalduinen, 1975)?
3. Do both the Hoogerheide and Woensdrecht Members belong to the

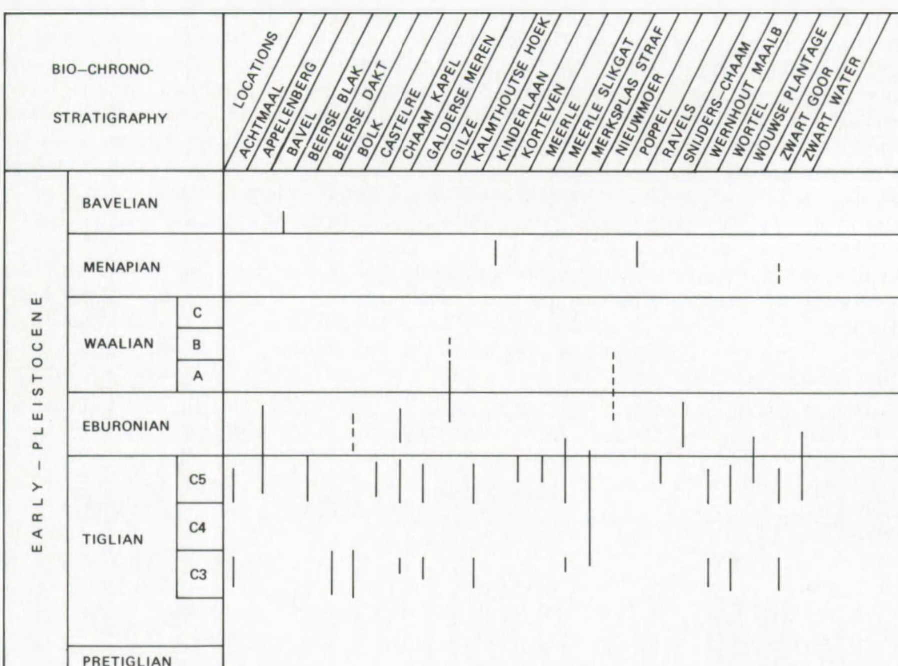
Tiglian (Nelson and Van der Hammen, 1950) or are they partly younger (Van Dorsser, 1956)? Is the top of the Woensdrecht Member in the west and the Turnhout Member in the east isochronous?
4. Has the Gilze Member been deposited in the Early-Pleistocene or Late-Pleistocene during one or several stages (Menapian) (De Ploey, 1961; Vandenberghe and Krook, 1981)?
5. Are deposits of Bavelian age present in western Noord-Brabant?

4.3 Results

4.3.1 Biostratigraphy

In this paragraph the paleobotanical results are presented in a chronological order. The chronostratigraphic interpretations have been summarized in table 4.1. Only the most relevant pollen diagrams are discussed in the text. The remainder is presented in the appendix.

Table 4.1: Bio(chrono)stratigraphic interpretation of the investigated pollen sections.



The pollen samples have been taken from borings and from exposures. In general only clay-, silt- and peat-layers were sampled. Sand-beds are often poor in pollen and the pollen may also be reworked. The vertical sampling distance varied widely, ranging from 1 cm to several meters. Short intervals were used in clay- and peat-beds, occurring in the upper part of several members. It is assumed that sediment accumulation was slow in these fine-grained beds and therefore a comparatively long period may be represented by thin lithological units. Large sampling intervals were applied in more silty sections, which probably reflect higher energy conditions and more rapid sedimentation. The pollen samples have been treated according to standard laboratory methods (Faegri and Iversen, 1975). Preparation includes: KOH treatment, sieving, acetolysis (Erdtman), removal of clastic material by heavy fluids separation and HF treatment. Wherever possible 300 pollen grains have been counted. All trees and herbs (except water plants) have been included in the pollen sum, which can lead to a dominance of local elements, for instance in peat-layers. The zonation in the pollen diagrams is a local one, based on the changes in tree pollen and the AP/NAP ratio.

The Early-Pleistocene age of the deposits is illustrated by the presence of *Pterocarya, Carya, Tsuga* and *Eucommia* pollen in low values, which are absent from the Middle-Pleistocene onwards (Zagwijn, 1975b). Tertiary pollen like *Nyssa* and *Sciadopytis* are very rare. Occasionally, they are found in fluviatile sediments (Gilze Member: e.g. boring Zwart Goor), where they are probably reworked from Tertiary deposits.

Pliocene - Tiglian A

According to Vanhoorne (1962) sands of Tertiary age are present below the Campine Clay (our Rijkevorsel and Turnhout Members). He described a pollen sample at Turnhout at 30 m below the surface (fig. 2.7: boring 17E-154), probably from an in situ peat-layer. It is characterized by *Nyssa* (1.9%), *Sciadopytis* (3.8%) and *Fagus* (3.3%)(sum of all trees is 100%) (Vanhoorne, 1962). According to Vanhoorne this pollen assemblage is similar to the pollen spectra of the Mol Sands at Mol, although the amount of Tertiary pollen is higher in the latter location. The sample was probably taken at the transition from the Merksplas Member to the Rijkevorsel Member. The low values of Tertiary pollen and the presence of *Fagus* in boring 17E-154 points to a Late-Pliocene to Early-Tiglian age for the Merksplas Member, since *Fagus* is present until the Tiglian A (Zagwijn, 1963a).

Fagus and Tertiary pollen have not been found in the lower part of the Rijkevorsel Member in the borings Bolk and Zwart Water (see appendix), although they are situated fairly close to boring 17E-154 of Vanhoorne (1962). The pollen spectra of silt/clay-laminae in these borings are dominated by *Alnus* and *Pinus*. It is therefore concluded that no depos-

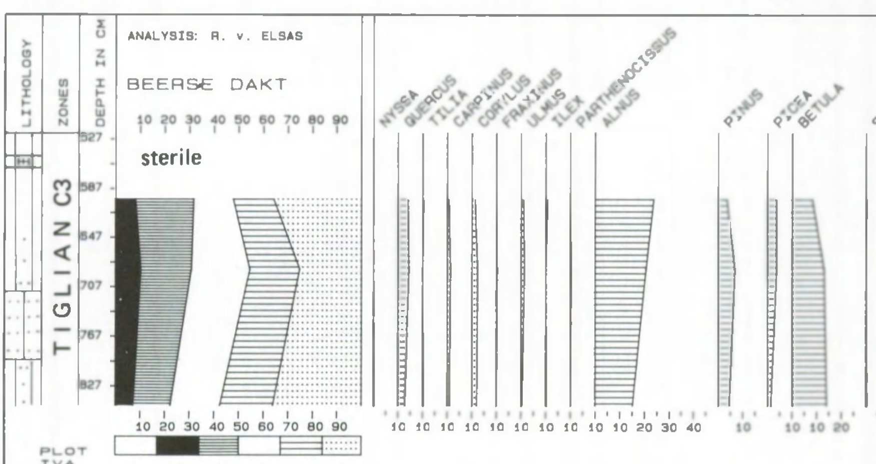


Fig. 4.1: Pollen diagram of the Rijkevorsel Member at Beerse Dakt (legend in appendix).

its of Late-Tertiary or Tiglian A age have been found in this study.

Tiglian

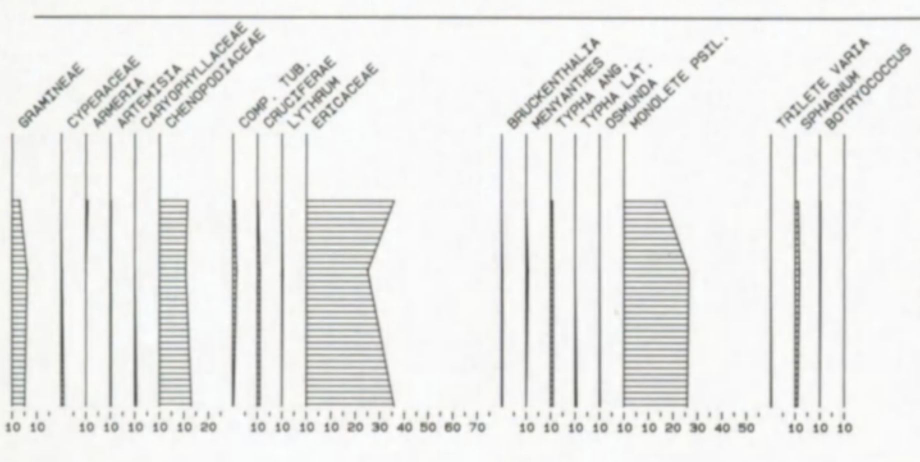
The Rijkevorsel Member (Tegelen Formation-Campine Clay and Sand Formation) overlies the Merksplas Member. It has been investigated in the stratotype at Beerse Dakt (fig. 4.1) and in several other borings and exposures (see app.).

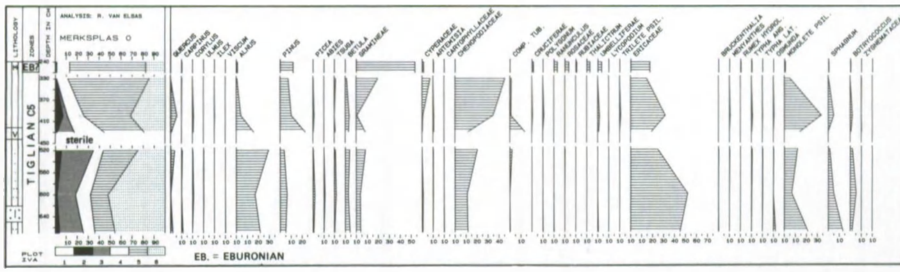
The member is characterized by fairly high values of thermophilous trees (30%) of dry (*Quercus*) and wet (*Alnus*) habitats. *Alnus, Betula, Pinus* and *Quercus* dominate the tree pollen. The high Chenopodiaceae content (12%) is characteristic for the distal tidal environment of the Rijkevorsel Member. This assemblage distinguishes the Rijkevorsel Member from the underlying Merksplas Member and the overlying Beerse Member, since practically no thermophilous species occur in the latter unit.

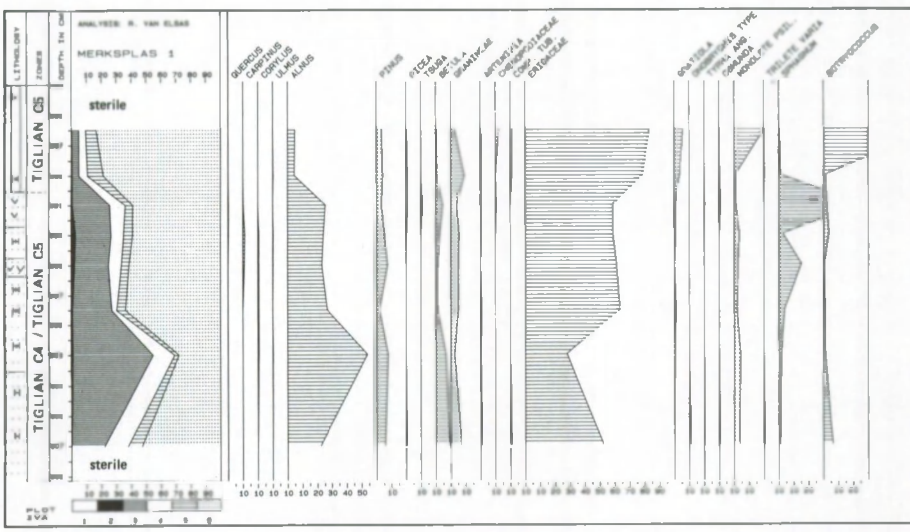
Several samples have been analysed on the presence of megasporangia of *Azolla*. This waterfern has not been found, but according to Greguss and Vanhoorne (1961) *Azolla tegeliensis* is a common constituent in the Rijkevorsel Member. Therefore the Rijkevorsel Member is regarded to be of Tiglian age, since *Azolla tegeliensis* is a guide fossil for the Tiglian period (Zagwijn, 1963a).

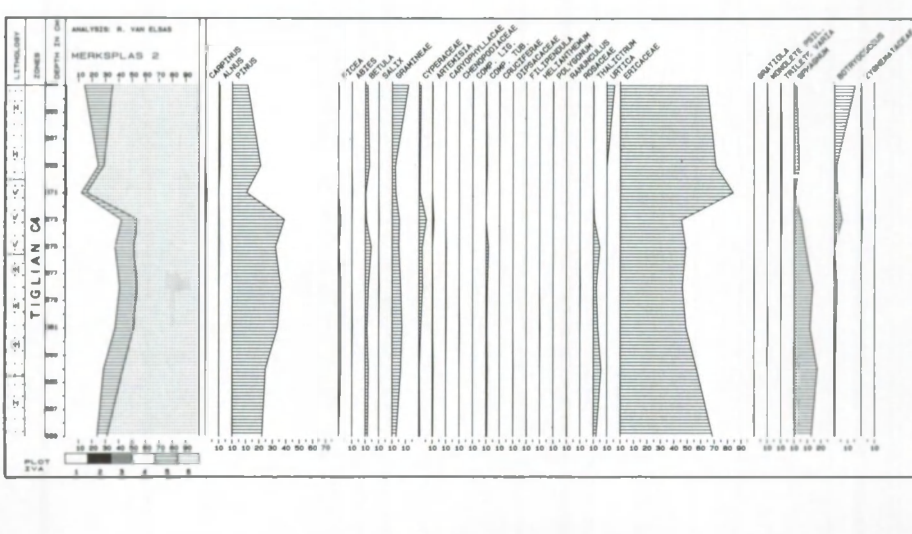
The distal tidal flat deposits of the Rijkevorsel Member are overlain by continental (eolian) deposits of the Beerse Member in the south-eastern part of the study area (fig. 4.2). The top of the Rijkevorsel Member at Merksplas Strafinrichting has not been eroded and therefore no hiatus occurs between the Rijkevorsel Member and the Beerse Member. Unfortunately the upper part of the Rijkevorsel Member appeared to be sterile in pollen probably because of oxidation and soil-ripening during silting above mean sea level in the tidal litter zone environment as revealed by the well developed crumbliness (§3.3.2). Therefore, the ecological and climatological transition from the Rijkevorsel Member into the Beerse Member could not be reconstructed.

Five soils were found at the base, in and on the top of the Beerse Member in Merksplas Strafinrichting (fig. 4.2: Merksplas(M) 1-5). In contrast to the underlying Rijkevorsel Member (M 5: 10.80 m) and over-




ANALYSIS: R. VAN ELSAS
MERKSPLAS 0
TIGLIAN C5
EB.
sterile
EB. = EBURONIAN
LITHOLOGY
ZONES
DEPTH IN CM
QUERCUS
CARPINUS
CORYLUS
ULMUS
ILEX
VISCUM
ALNUS
PINUS
PICEA
ABIES
TSUGA
BETULA
GRAMINEAE
CYPERACEAE
ARTEMISIA
CARYOPHYLLACEAE
CHENOPODIACEAE
COMP. TUB.
CRUCIFERAE
POLYGONUM
RANUNCULUS
ROSACEAE
RUBIACEAE
THALICTRUM
UMBELLIFERAE
LYCOPODIUM
TRILETE PSIL.
ERICACEAE
BRUCKENTHALIA
MENYANTHES
RUMEX HYDROL.
TYPHA ANG.
TYPHA LAT.
OSMUNDA
MONOLETE PSIL.
SPHAGNUM
BOTRYOCOCCUS
ZYGNEMATACEAE


ANALYSIS: R. VAN ELSAS
MERKSPLAS 1
TIGLIAN C4 / TIGLIAN C5
TIGLIAN C5
sterile
sterile
LITHOLOGY
ZONES
DEPTH IN CM
QUERCUS
CARPINUS
CORYLUS
ULMUS
ALNUS
PINUS
PICEA
TSUGA
GRAMINEAE
ARTEMISIA
CHENOPODIACEAE
COMP. TUB.
ERICACEAE
OSMUNDA
MONOLETE PSIL.
TRILETE VARIA
SPHAGNUM
BOTRYOCOCCUS


ANALYSIS: R. VAN ELSAS
MERKSPLAS 2
TIGLIAN C4
LITHOLOGY
ZONES
DEPTH IN CM
CARPINUS
ALNUS
PINUS
PICEA
ABIES
BETULA
SALIX
GRAMINEAE
CYPERACEAE
ARTEMISIA
CARYOPHYLLACEAE
CHENOPODIACEAE
COMP. LIG.
COMP. TUB.
CRUCIFERAE
DIPSACACEAE
FILIPENDULA
HELIANTHEMUM
POLYGONUM
RANUNCULUS
ROSACEAE
THALICTRUM
URTICA
ERICACEAE
GRATIOLA
MONOLETE PSIL.
TRILETE VARIA
SPHAGNUM
BOTRYOCOCCUS
ZYGNEMATACEAE

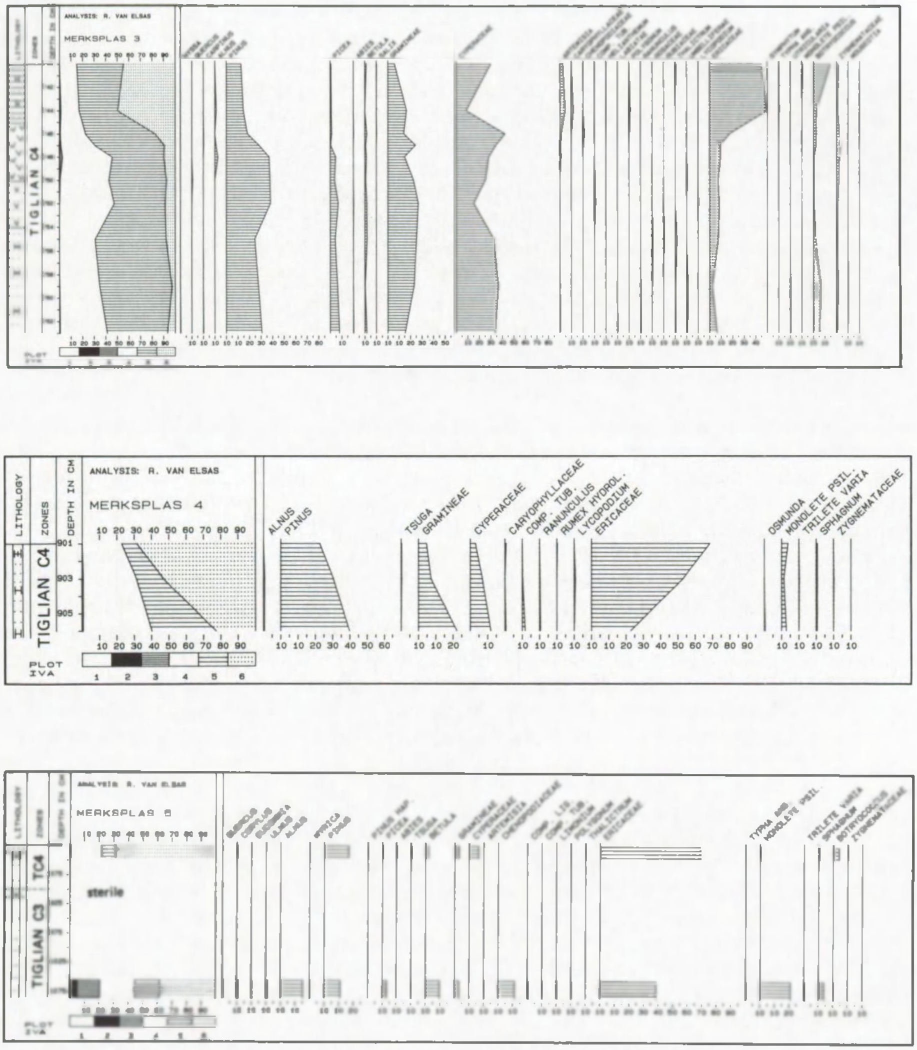


Fig. 4.2: Pollen diagram of the Rijkevorsel Member, Beerse Member and Turnhout Member at Merksplas Strafinrichting (Tiglian C3, C4, C5)(legend in appendix).

lying Turnhout Member (M 0) the thermophilous trees of dry and wet places are almost completely absent. The pollen diagrams (M 2, 3, 4) are dominated by so-called indifferent trees (*Pinus*), herbs (Gramineae, Cyperaceae) and Ericaceae. The pollen assemblages point to a taiga or tundra vegetation. The associated cryoturbations and small ice- wedge casts in the eolian sands between the soils, indicate a cold climate, probably with local permafrost conditions and a mean annual temperature around -5° C (§3.3.3). The upper soil (M 1) is different from the lower three. *Alnus* is more important, but the Ericaceae are still dominant. It is uncertain whether the *Alnus* pollen belong to *Alnus viridis* or *Alnus glutinosa*. The size of the pollen grains was measured at 23.7 μm

(N=50). Since the measurements were established in glycerine gel, a conversion factor of 0.8 must be applied to correct for expansion of the pollen grains (Faegri and Iversen, 1975). Pollen size then amounts to 18.9 µm. The size of *Alnus* pollen at Meerle, situated in the overlying warm temperate phase, amounts to 19.6 µm (N=50; conversion factor 0.8). According to Menke (1976) *Alnus glutinosa* and *Alnus viridis* have a diameter of 21.0 $\pm$ 1.0 µm and 17.9 $\pm$ 0.9 µm respectively. Both our values lie between those mentioned by Menke and therefore no conclusions can be drawn from the size measurements. However, the habitus of the *Alnus* pollen from soil M 1 resembles more the habitus of recent *Alnus glutinosa*. The increase of *Alnus* at the top of the Beerse Member in M 1 possibly represents the climatological change from the glacial phase to the next interglacial phase. Moreover, it cannot be excluded that soil M 1 was formed during the interglacial period itself. Wet and oligotrophic conditions may have been responsible for the presence of *Alnus*, Ericaceae and *Sphagnum* in diagram M 1.

The Turnhout Member overlies the Beerse Member in Merksplas Strafinrichting. The upper soil of the Beerse Member (M 1) was locally eroded. The Turnhout Member is dominated by *Alnus, Pinus, Quercus* and *Betula* (fig. 4.2: M 0). Thermophilous trees of wet and dry habitats are well represented (max. 30%). *Typha* and *Osmunda* point to a relatively high summer temperature. The Chenopodiaceae content is as high as or higher than in the Rijkevorsel Member (max. 40%). This again indicates a high sea-level and distal tidal flat and tidal litter zone environments (§3.3.4). The high content of Chenopodiaceae pollen suppresses the values of the thermophilous tree pollen. In spite of this occasional dominance of local pollen, the pollen assemblage as a whole is characteristic of a temperate climate, comparable to the Holocene. The upward decrease of Ericaceae in M 0 can be explained by the decreased reworking of pollen from the underlying soils (M 1) of the Beerse Member and by the increasing distance to the mainland in the south during the transgression of the Turnhout Member.
In the Turnhout Member at Merksplas Strafinrichting (fig. 4.2: M 0, 5.2 m below surface) 30 megasporangia of *Azolla tegeliensis* have been found. In Meerle the Turnhout Member contained up to 48 megasporangia of *Azolla tegeliensis* in one small sample. The presence of *Azolla tegeliensis* unmistakenly points to deposition during the Tiglian, since *Azolla tegeliensis* is regarded as a guide fossil for the Tiglian stage (Zagwijn, 1963a). *Azolla filiculoides* has never been found in the Turnhout Member, nor in the Rijkevorsel Member, which confirms the Tiglian date of the members. The Tiglian age of the Turnhout Member is in contradiction to previous ideas of Dricot (1961), De Ploey (1961) and Paepe and Vanhoorne (1970). The latter found *Azolla tegeliensis* in the Rijkevorsel Member, but not in the Turnhout Member (pers. comm. R. Vanhoorne). They therefore interpreted the Rijkevorsel Member as Tiglian and the Turnhout Member as Waalian age. The absence of *Azolla tegeliensis* in the Turnhout Member in previous studies can be explained by the fact that only few samples were investigated, since the Turnhout Member is presently thin or even absent (e.g. in Beerse Dakt) at the southern border of its distributional area. To the north the sedimentary sequence of the Turnhout Member is more complete and *Azolla tegeliensis* is a common constituent in the fining-upward sequences formed by the final silting of the fresh water, tidal environments (see Meerle). The Tiglian age of the Turnhout Member fits well with the finds of deer antlers of Eucladoceros tegulensis (Germonpré, 1983; §3.5.4).
Since both the Rijkevorsel and Turnhout Member are of Tiglian age, the

intercalated Beerse Member must be of Tiglian age as well. This result is new, as the cold Beerse Member has been interpreted formerly as Eburonian or even Menapian (see §4.2). Up to now two cool phases are known within the Tiglian (Zagwijn, 1963a: Tiglian B and Tiglian C4). If the Beerse Member is to be connected with the Tiglian B phase, then the underlying Rijkevorsel Member should have been formed in the warm-temperate Tiglian A. However, the Tiglian A is characterized by the presence of *Fagus* and low amounts of Tertiary pollen (*Sciadopytis, Sequoia, Taxodium*) (Zagwijn, 1963a: boring Eindhoven I, II). Since *Fagus* and Tertiary pollen have never been found (except for a few grains), the Rijkevorsel Member-Tiglian A correlation is rejected. The Beerse Member must be connected then with the cool Tiglian C4 phase (Zagwijn, 1963a). This implies that the Rijkevorsel Member and Turnhout Member were probably formed in the Tiglian C3 and Tiglian C5 respectively. The Tiglian C3 age of the Rijkevorsel Member implies a large hiatus between the Merksplas Member (Pliocene, Praetiglian) and the Rijkevorsel Member, comprising the Tiglian A, Tiglian B and part of the Tiglian C.

The Tiglian C3(b) zone has been described previously as the climatic optimum of the Tiglian. The Tiglian C5 phase reflects another period of warm temperate climatic conditions, almost as warm as the TC3(b) zone (Zagwijn, 1963a). The weak climatic differences between the Tiglian C3 and C5 phases could occasionally be distinguished in the pollen diagrams of respectively the Rijkevorsel and Turnhout Members. The Rijkevorsel Member often contains a somewhat higher content of thermophilous, dry trees than the Turnhout Member (compare fig. 4.1 and fig. 4.2; see also appendix Meerle Slikgat, Chaam Kapel, Galderse Meren). Especially *Eucommia, Carpinus, Ulmus, Tilia, Fraxinus* and *Ilex* are often better represented in the Rijkevorsel Member. It is possible to reconstruct the mean summer (and winter) temperatures from the pollen record (Zagwijn, 1963a, 1975b). The pollen of *Ilex, Hedera* and *Castanea* in the Rijkevorsel and Turnhout Member indicates mild winter conditions with mean temperatures above 0° C. The presence of (*Taxus*), *Hedera*, (*Vitis*) and *Eucommia* point to a mean summer temperature around 20° C. The lower values of the species mentioned above in the Turnhout Member might indicate somewhat lower temperatures in the Tiglian C5 phase.
The correlation of the Beerse Member with the Tiglian C4 implies the introduction of a glacial phase within the Tiglian. The estimated mean annual temperature of -5° C is much lower than described so far by Zagwijn (1963a: TC4c; fig. 6, fig. 15).
As has been stated above, the areal distribution of the Beerse Member is limited because of erosion by the Turnhout Member. Reworked pollen of the Beerse Member, deposited at the base of the Turnhout Member (fig. 4.2: M 0) can give the impression of a cooler phase. In Meerle Slikgat for instance (see app.) the interval between 7.55 and 19 m below the surface contains less thermophilous, dry trees and more herbs and Ericaceae between two pollen zones with a higher content of dry, thermophilous trees. The high values of dry thermophilous trees at the top of the Rijkevorsel Member (19-21 m) and at the top of the Turnhout Member (7-7.5 m) is possibly caused by the increase of pollen from the mainland during the final silting of the distal, tidal environments during the Tiglian C3 and C5 respectively. However, lithologically, sediment-petrographically and environmentally the unit between 7.55 and 19 m is part of the Turnhout Member. Furthermore, the *Alnus* content (30-35%) is much higher than in the in situ Beerse Member. Therefore, this interval most probably formed in the TC5 phase as well. The higher content of herbs and Ericaceae is probably caused by reworking of

pollen from the Beerse Member and by a basinwards decrease of *Alnus* (Chowdhury, 1982)(see §3.4). A comparable phenomenon is possibly present more to the north e.g. in boring Rotterdam E55 (Zagwijn, 1963a). The interval between 98 and 115 m below the surface is interpreted as the cool TC4c phase. The pollen assemblage, however, contains up to 10-15% thermophilous trees, which is much higher than in the Beerse Member (fig. 4.2). The pollen may have been reworked from sediments of Tiglian C4 age and deposited in marine sedimentary environments during the beginning of the TC5 phase.
In the upper part of the Turnhout Member very often a strong increase of thermophilous trees (especially *Alnus*) is found in humic to peaty layers (fig. 4.3 and App.: Meerle, Beerse Blak, Appelenberg, Zwart Water, Meerle Slikgat). This increase of thermophilous trees could reflect the climatic optimum of the interglacial Tiglian C5. However, in our opinion this is not the case. The high *Alnus* content always occurs in a peaty layer at the top of the Turnhout Member, which is interpreted as the final phase of silting of the distal, tidal flat-tidal litter zone environments. The increase of thermophilous trees is considered therefore not the result of more optimal climatic conditions, but of local edaphic factors.
The effects of progressive silting and subsequent drowning are well illustrated at Ravels (fig. 4.3). At the base of the diagram (3.41 m) sedimentation occurred in open water conditions (lenticular bedding) and regional pollen dominates the pollen assemblage. Approximately 15 cm below the peat-layer (2.53 m) the pollen spectrum is strongly dominated by local Chenopodiaceae pollen. At 2.39 m below the surface, peat formation started and Chenopodiaceae are replaced by Gramineae. This succession was interrupted at 2.37 m by renewed clay sedimentation with a temporarily increase of the Chenopodiaceae content. At 2.35 m renewed peat formation occurred and Gramineae content declined, followed by a decline of *Typha*, whereas at the same time *Alnus* increased (up to 55%). The *Alnus Osmunda* vegetation points to fresh and eutrophic water supply in an open swamp environment. The subsequent drowning of the peat at 2.28 m led to the disappearance of *Alnus* and first Gramineae reappeared, followed by Chenopodiaceae.

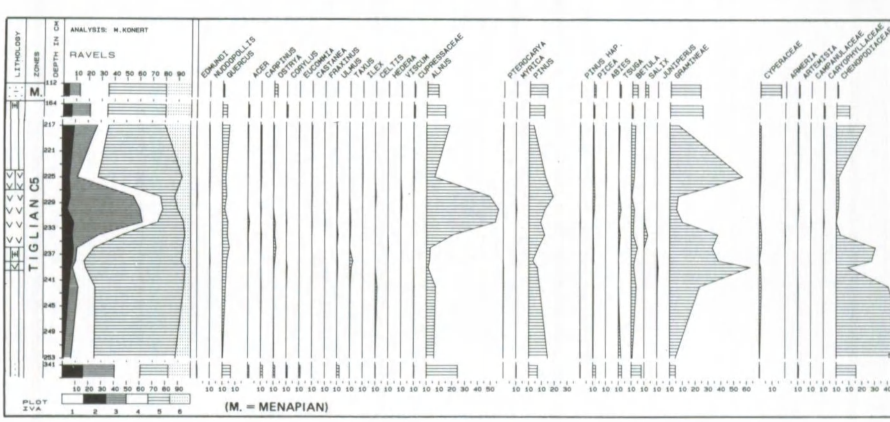


Fig. 4.3: Pollen diagram of the Turnhout Member at Ravels, illustrating the final silting at the top of the member (Tiglian C5).

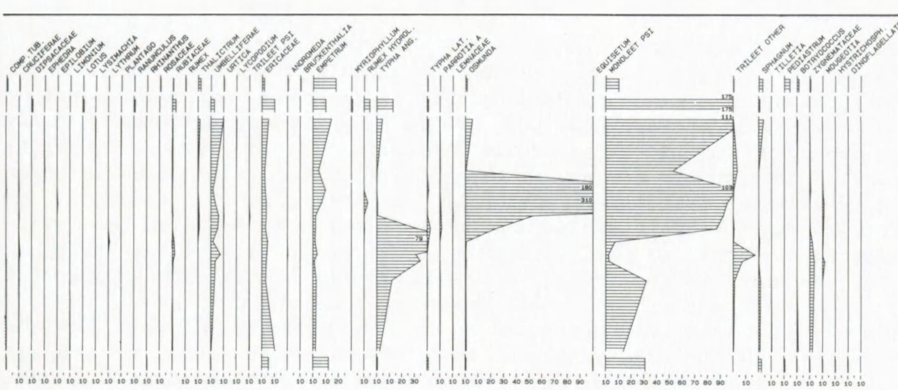
Chronostratigraphic position of the Woensdrecht and Hoogerheide Members.

The correlation of the Turnhout Member with the Woensdrecht Member is reliable, since the clayey top of these members continues through the whole area (fig. 2.3 and 2.6). The connection of the Rijkevorsel Member with the Hoogerheide Member is somewhat uncertain however. No evidence was found in the pollen record for a cool or cold interval either within the Hoogerheide and Woendrecht Members or at their transition.

The macro remains from the clay-bed in the upper part of the Woensdrecht Member yielded large amounts (37) of megasporangia of *Azolla tegeliensis* (Huyzer and Van Toor, 1986)(fig. 3.34, 3.35). The presence of this waterfern confirms the Tiglian age of the Woensdrecht Member (Nelson and Van der Hammen, 1950: II-0 deposits).

The Hoogerheide Member is characterized by the presence of *Alnus, Pinus, Betula* and Chenopodiaceae (*Armeria*)(App. Kalmthoutse Hoek). The Woensdrecht Member contains a comparable pollen assemblage, but the *Alnus* content can be higher (up to 45%), especially in peaty layers in the upper part of the member (fig. 4.4). The pollen assemblages in the upper part of the Turnhout and Woensdrecht Members resemble each other (fig. 4.3: Ravels and fig 4.4: Korteven; be aware of the large differences in vertical scale). Diagram Korteven (fig. 4.4) is dominated by *Quercus, Alnus, Pinus* and Chenopodiaceae. The peat-layer is characterized by *Alnus* and *Osmunda*. Below and above the peat-layer the content of Chenopodiaceae is more important, while the *Alnus* content is lower. The higher values of Gramineae in Ravels just below and above the peat are not found in Korteven, probably because of too wide sampling intervals.

As has been stated above, both the Turnhout and Woensdrecht Members are characterized by the presence of *Azolla tegeliensis*. Florschütz (1938) and Van der Vlerk and Florschütz (1953) described *Azolla tegeliensis* as well, at Hoogerheide at 3 m below the surface; this is in the Woensdrecht Member. At Wernhout they found this waterfern at 2 m below the surface; this is in the Turnhout Member (see fig. 2.6). Because of the comparable pollen associations and the presence of *Azolla tegeliensis*, the top of the Woensdrecht and Turnhout Members is interpreted as more



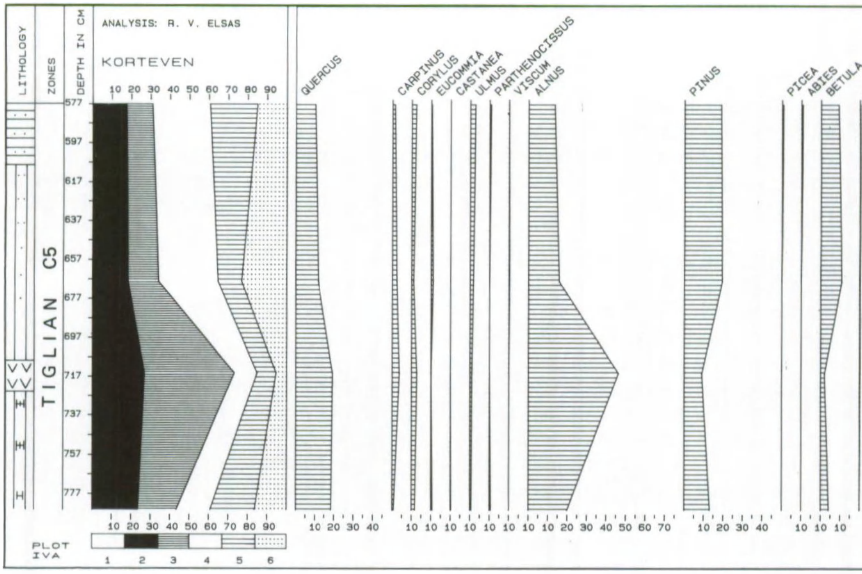
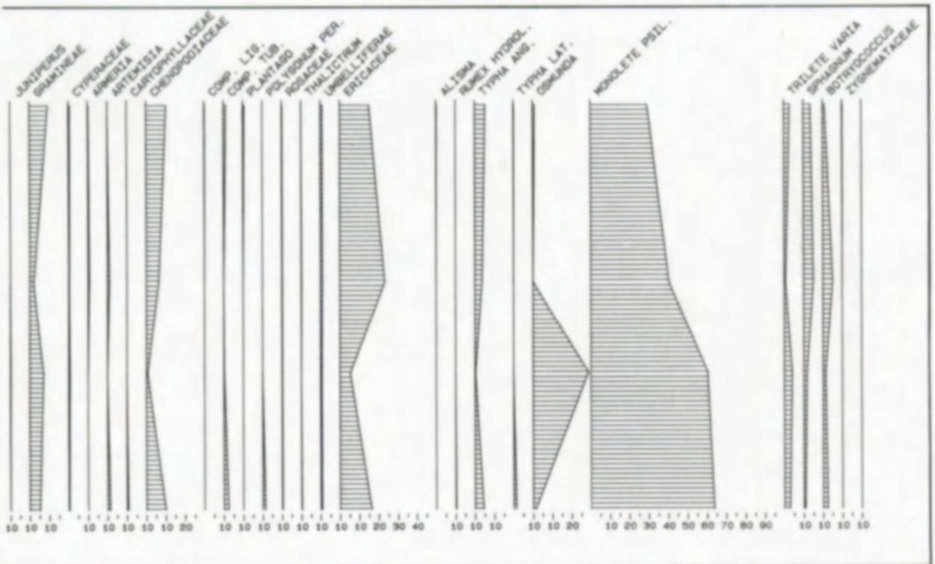


Fig. 4.4: Pollen diagram of the Woensdrecht Member in boring Korteven, showing the palynological resemblance to the Turnhout Member in Ravels (Tiglian C5).

or less isochronous (Tiglian C5). In the Hoogerheide Member Tertiary pollen and *Fagus* are absent, so it must be younger than the Tiglian A (Zagwijn, 1963a). Because of its pollen composition (App. Kalmthoutse Hoek: *Alnus, Quercus, Pinus, Betula*) and the close similarity to the pollen assemblage of the Woensdrecht Member a Tiglian C age seems the most likely. The Hoogerheide Member then might have been formed in the Tiglian C3.

The clayey top of the Turnhout and Woensdrecht Members has been eroded at many places (fig. 2.6: Kalmthoutse Hoek, Nieuwmoer, Ghil, Zwart Goor); especially in the Mark-Weerijs river basin (Achtmaal, Wernhout Maalbergen). In those situations a hiatus is present between the clay and overlying units (Gilze Member, Eindhoven Formation, Twente Formation). According to De Ploey (1961) and Van Oosten (1967) a humic horizon, which was dated as Eemian, is locally present in the top of the clay. However, deposits of Eemian age have not been found in the investigated sections probably because of erosion during the Early- or Middle-Weichselian. According to Vandenberghe (1985) Eemian deposits occur only locally on interfluvia and in subsiding areas (Central Graben) where they were protected against Weichselian erosion. In the study area the humic clay at the top of the Turnhout Member is commonly characterized by a cool or cold pollen assemblage (appendix Merksplas Strafinr., Appelenberg, Zwart Water). The cold pollen spectrum can date from any glacial between the Tiglian and the Holocene. The "Eemian" peat-layer at Meerle described by De Ploey (1961) was reinvestigated in this study (App. Meerle). The pollen composition and the presence of *Azolla tegeliensis* point to a Tiglian instead of an Eemian age.



Tiglian - Eburonian transition

In boring Appelenberg (fig. 4.5), Zwart Water and Meerle Slikgat (App.) a gradual lithological transition has been found between the Turnhout Member and the overlying Gilze Member. Here a more or less continuous sedimentary sequence is present, in which climatic changes have been registered.

The pollen assemblage of the Turnhout Member at Appelenberg is very characteristic of the Tiglian C5 period (fig. 4.5). *Alnus* dominates the pollen spectrum, especially in the humic soil horizon. In the upper 30 cm of the Turnhout Member *Alnus* decreases and *Pinus, Picea,* Gramineae and *Juniperus* increase. This change does not seem to be connected to a change in the sedimentary facies and it is therefore interpreted as a cooling of the climate. This deterioration of the climate is normally situated in the Tiglian C6 zone (Zagwijn, 1963a).

At 12 m below the surface the clay changes upwards into peat, gyttja, loam and very fine sand of the Gilze Member. The sediments between 11.23 and 11.55 m have been deposited in very wet, locally lacustrine environments. Gramineae and Cyperaceae dominate the pollen spectrum in this interval, but they may be over represented because of the local, wet sedimentary environment. Thermophilous trees are virtually absent, indicating cold climatic conditions. Because of the more or less continuous vertical sequence of the Turnhout Member into the Gilze Member, the pollen assemblages most probably represent the Eburonian cold stage.

Above 10.95 m fluviatile overbank deposits become more important and local dryer conditions are reflected by a higher *Pinus* content in the diagram (fig. 4.5). The increase of *Artemisia* and *Thalictrum* indicates a change to a more continental climate in the course of the Eburonian period. The increase of *Corylus, Quercus* and *Castanea* is explained by an increased reworking of pollen due to higher current velocities during the sedimentation.

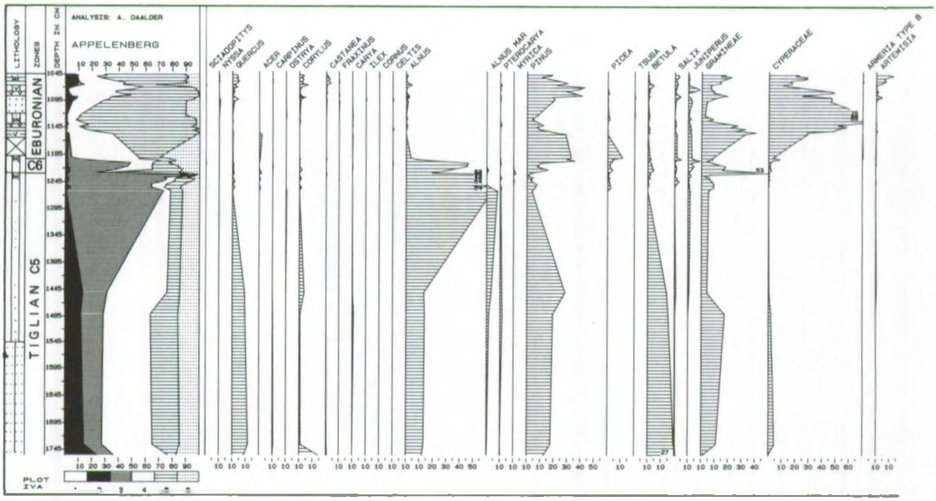


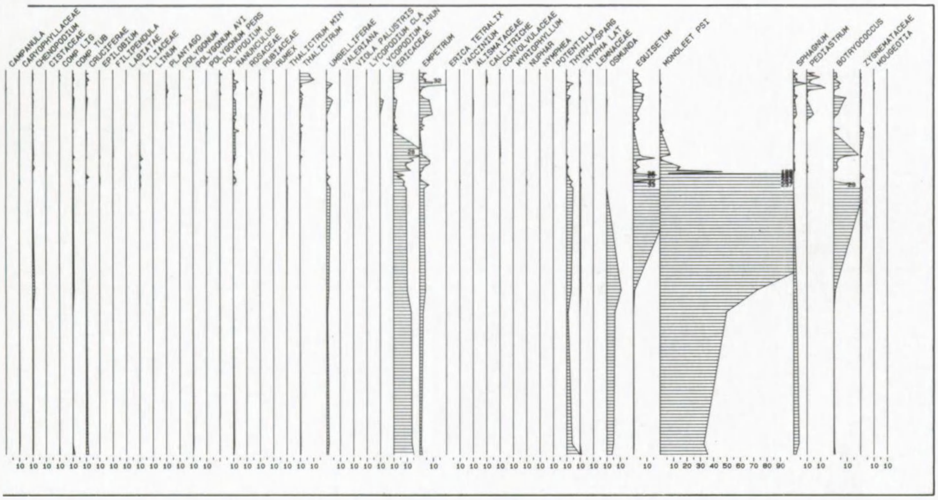
Fig. 4.5: Pollen diagram with the transition of the Turnhout Member into the Gilze Member in boring Appelenberg (Tiglian C5, C6, Eburonian).

In diagram Zwart Water (app.) the pollen spectra of the cold Eburonian stage directly overlie at 4.36 m below the surface the *Alnus* dominated spectra of the warm temperate Tiglian C5 phase, although no clear break is present in the sedimentary sequence. The Eburonian is characterized by fluctuations in the *Pinus*-Gramineae ratio, which probably represent stadial (low *Pinus* content) and interstadial phases (higher *Pinus* content) within the Eburonian period (Eburonian I,II,III: Zagwijn, 1963a: Pit Russel-Tiglia-Egypte, boring Eindhoven I).

Waalian

The Gilze Member consists predominantly of sand and continuous clay- or peat-beds are scarce. Therefore, the pollen diagrams from the Gilze Member are often short and difficult to correlate from one place to another.

Following the Eburonian stage, the next phase with a high content of thermophilous trees occurs in the Gilze Member (Gilze Clay) at Gilze (fig. 4.6). Unfortunately this pollen section does not form one continuous, vertical sequence with the Eburonian and Tiglian deposits, because of local erosion of the Turnhout Member at this spot (see fig. 2.8). The lower part of the diagram (fig. 4.6: 5.5 - 14 m below the surface) contains a high content of *Pinus, Betula* and Gramineae. It is situated stratigraphically above the upper part of diagram Appelenberg (fig. 4.5). Because of the high *Pinus* values a late Eburonian or early Waalian age is proposed for the base of the Gilze diagram. Above 5.48 m the pollen spectra become dominated by thermophilous dry (*Carpinus, Quercus, Ulmus*) and wet (*Alnus*, max. 68%) trees. The pollen assemblage is characteristic for a temperate-warm climate. Above 2.10 m below the surface the thermophilous trees decrease and the pollen spectrum becomes dominated again by *Pinus, Betula*, Gramineae and Ericaceae (fig. 4.6). The *Corylus* increase at 1.25 m is probably caused by a higher current velocity (clay changing into loam) and the connected reworking of pollen.



The chronostratigraphic position of this warm phase can be deduced in the following manner. The presence of *Tsuga* (4%) is an indication for the Early-Pleistocene age of this section (fig. 4.6). Analysis of the macro remains revealed large amounts of megasporangia and massulae of *Azolla filiculoides*, which were nicely attached to each other. This waterfern is not restricted to the Early-Pleistocene. According to Zagwijn (1963a) it occurs frequently after the Tiglian (and sporadically in Tiglian C4, TC5, TC6), in the Waalian, Bavelian, Cromerian and Holsteinian stages. The pollen spectrum at Gilze does not resemble the Bavel Interglacial pollen assemblages, since *Tsuga* and *Eucommia* content is much higher in the Bavel Interglacial (fig. 4.8) (Zagwijn and De Jong, 1984). To conclude, the pollen spectrum at Gilze dates from an Early-Pleistocene interglacial, younger than Tiglian and older than Bavelian, which is the Waalian.
The pollen spectra of Waalian age at Gilze resemble other spectra of Waalian age in the Central Netherlands (e.g. Zagwijn and De Jong, 1984: boring Leerdam). Zagwijn has been able to distinguish a tripartition in the Waalian (Waalian A, B, C) in some pollen diagrams (Zagwijn, 1957: boring Veldhuizen; Zagwijn, 1963a: boring Eindhoven I; Zagwijn and De Jong, 1984: boring Leerdam). The differences in the pollen assemblages of the warm temperate Waalian A and Waalian C are not always evident, especially when the intermediate, cooler Waalian B phase is absent. In Gilze only one warm temperate phase of the Waalian seems to be present (fig. 4.6). Because of the relatively high content of thermophilous dry trees (*Carpinus, Quercus*) it is probable that only the Waalian A phase has been registered. The strong decline of thermophilous trees above 2.1 m reflects a cooling of the climate, possibly representing the beginning of the Waalian B phase. The relatively high content of *Pinus* and *Betula* is more characteristic for the cool Waalian B, than for the cold Menapian phase.

Menapian
In the neighbourhood of Alphen, the clay-layers of Waalian age (exposed at Gilze) have been eroded by channels, which have subsequently been filled with medium to coarse so-called "Alphen Sands" (Vandenberghe and Krook, 1981). The latter unit is incorporated in the Gilze Member

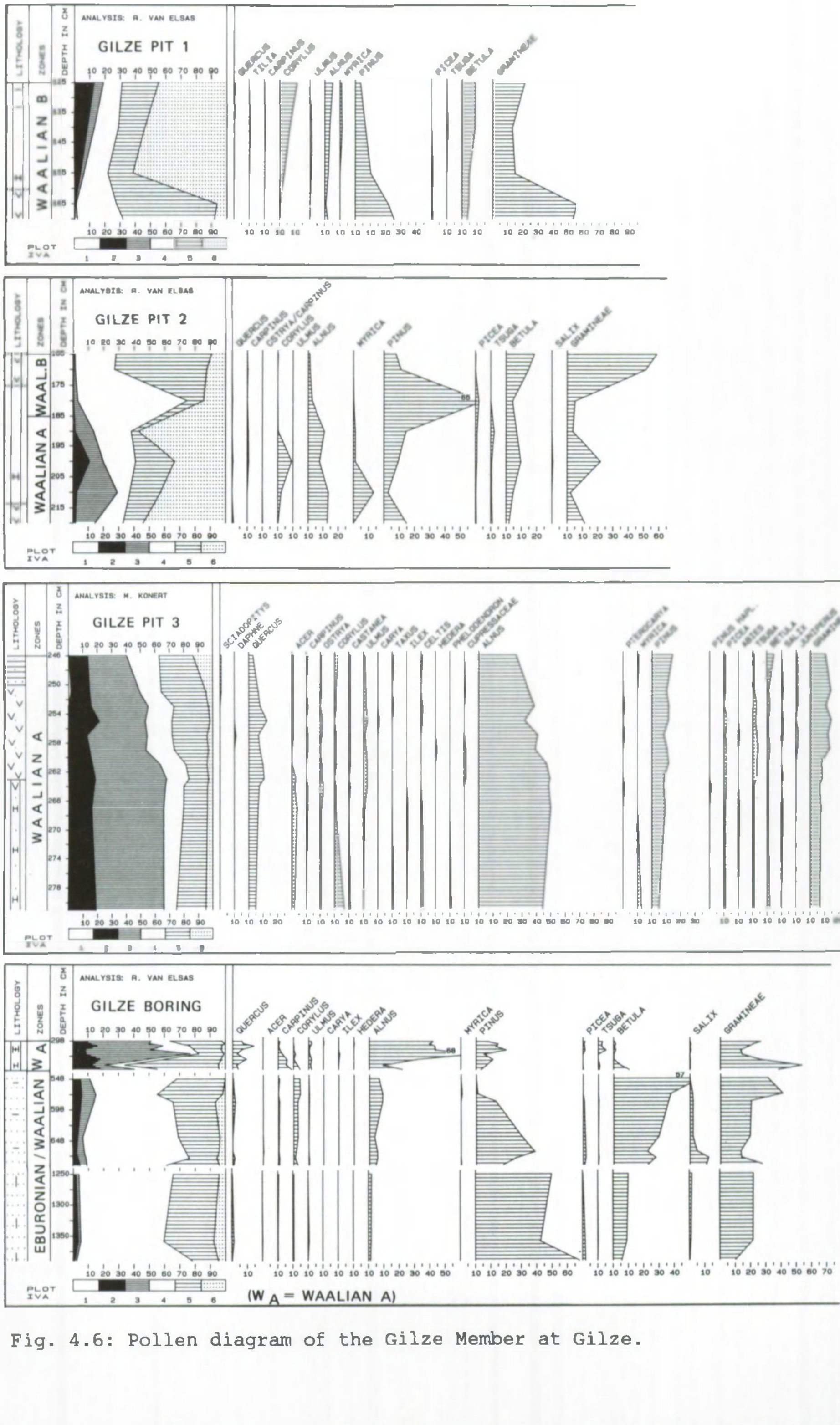
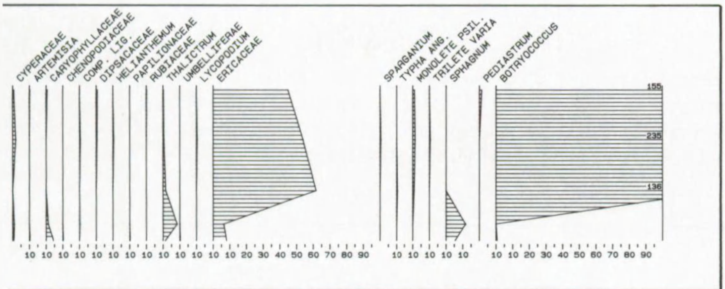


Fig. 4.6: Pollen diagram of the Gilze Member at Gilze.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DIPSACACEAE
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RUBIACEAE
THALICTRUM
UMBELLIFERAE
LYCOPODIUM
ERICACEAE
SPARGANIUM
TYPHA ANG.
MONOLETE PSIL.
TRILETE VARIA
SPHAGNUM
PEDIASTRUM
BOTRYOCOCCUS
155
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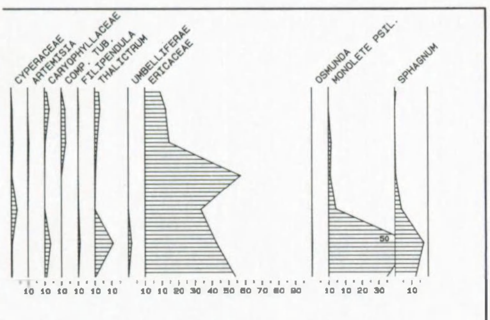
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MONOLETE PSIL.
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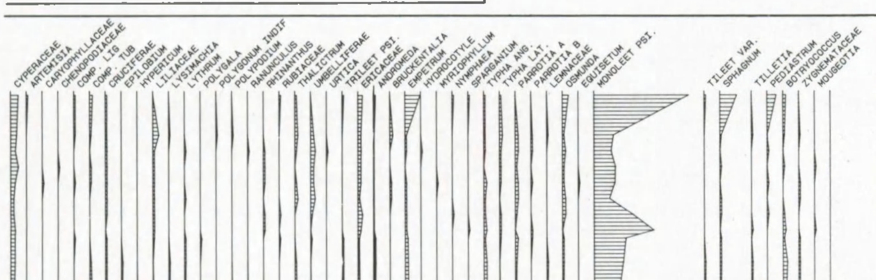
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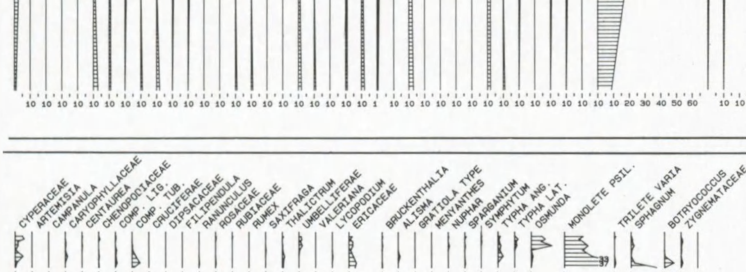
HYPERICUM
LILIACEAE
LYSIMACHIA
LYTHRUM
POLYGALA
POLYGONUM INDIF
POLYPODIUM
RANUNCULUS
RHINANTHUS
RUBIACEAE
THALICTRUM
UMBELLIFERAE
URTICA
TRILEET PSI.
ERICACEAE
ANDROMEDA
BRUCKENTALIA
EMPETRUM
HYDROCOTYLE
MYRIOPHYLLUM
NYMPHAEA
SPARGANIUM
TYPHA ANG.
TYPHA LAT.
PARROTIA A
PARROTIA B
LEMNACEAE
OSMUNDA
EQUISETUM
MONOLEET PSI.
TILEET VAR.
SPHAGNUM
TILLETIA
PEDIASTRUM
BOTRYOCOCCUS
ZYGNEMATACEAE
MOUGEOTIA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DIPSACACEAE
FILIPENDULA
RANUNCULUS
ROSACEAE
RUBIACEAE
RUMEX
SAXIFRAGA
THALICTRUM
UMBELLIFERAE
VALERIANA
LYCOPODIUM
ERICACEAE
BRUCKENTHALIA
ALISMA
GRATIOLA TYPE
MENYANTHES
NUPHAR
SPARGANIUM
SYMPHYTUM
TYPHA ANG.
TYPHA LAT.
OSMUNDA
MONOLETE PSIL.
TRILETE VARIA
SPHAGNUM
BOTRYOCOCCUS
ZYGNEMATACEAE
37

(coarse-grained lithofacies), because of its stable heavy mineral content (chapter 2). According to Vandenberghe and Krook (1981) the pollen diagram of the upper part of the "Alphen Sands" is dominated by *Pinus*, *Betula*, Gramineae and Cyperaceae. This association points to a park-tundra vegetation and a cool climate. The authors dated the pollen section at Alphen as "possibly Menapian". This age agrees well with the Waalian age of the underlying clay-bed at Gilze (fig. 4.6).

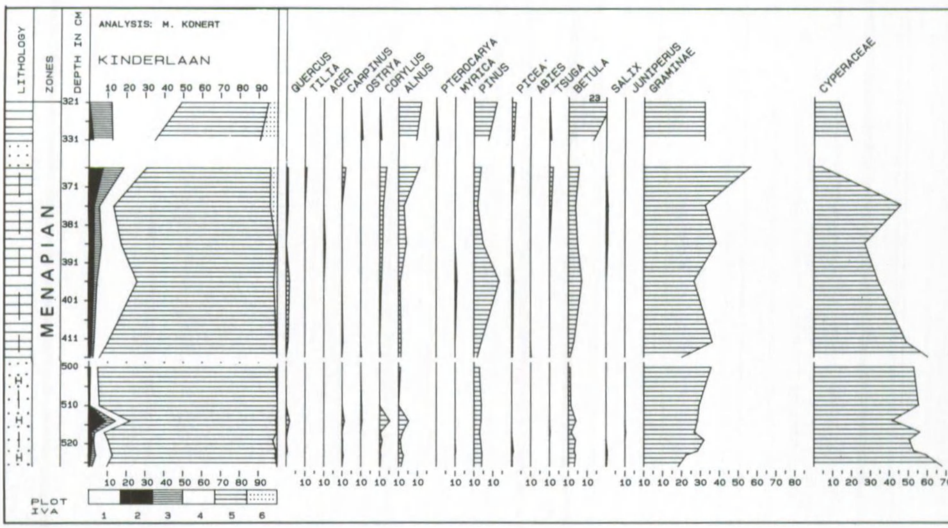


Fig. 4.7: Pollen diagram of the Gilze Member (Spruitenstroom Clay) at Kinderlaan (Menapian).

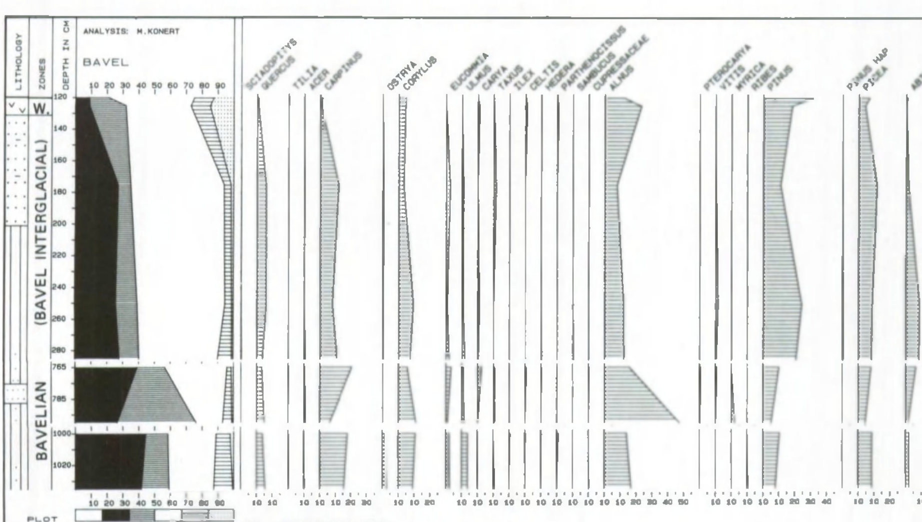
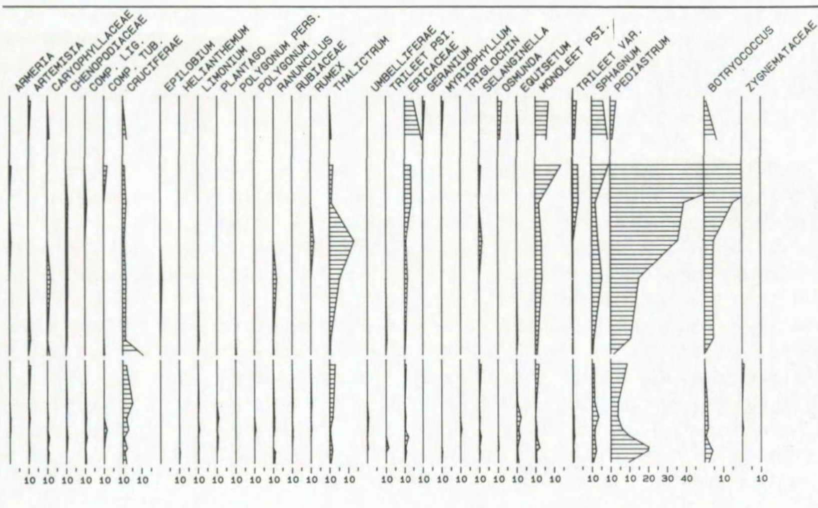
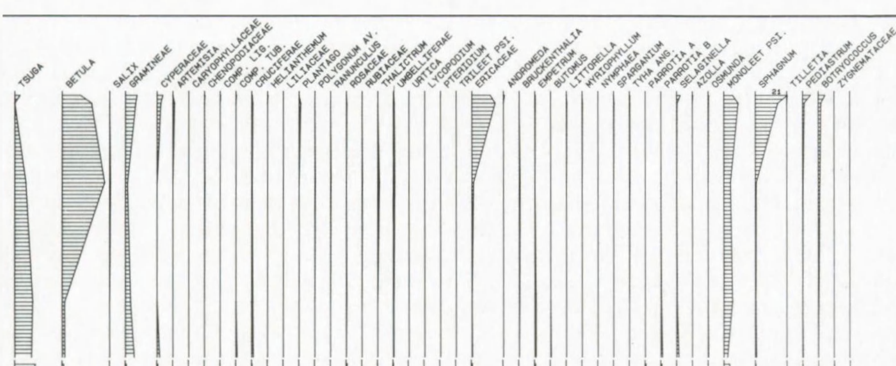


Fig. 4.8: Pollendiagram of the Bavel Member at Bavel (Bavelian Bv3b).

In the present study the dating of the coarse upper part of the Gilze Member (Alphen Sands) is generally impeded by the lack of peat- or clay-layers (App.: Ghil, Zwart Goor, Ravels). Reworking of pollen is locally very important. For instance, the presence of 10% pollen of *Sciadopytis, Nyssa* and *Sequoia* in the Gilze Member at Zwart Goor (App.) indicates the reworking of pollen from Tertiary deposits. The degree of reworking is in general well demonstrated by the higher content of





Corylus pollen (fig. 4.5: 10.65 m; fig. 4.6: 1.25 m).
In the east of the study area a clay-bed is found in the upper part of the Gilze Member (Spruitenstroom Clay). The pollen diagram from this unit at Kinderlaan is completely dominated by Gramineae and Cyperaceae (fig. 4.7). The continuous presence of *Thalictrum* (and Cruciferae) indicates cold climatic conditions. Because of the lithostratigraphic position in the upper part of the Gilze Member, above the Gilze Clay of Waalian age and below the Bavel clay of Bavelian age this pollen diagram is interpreted as Menapian. However, it must be remembered that the pollen sections at Kinderlaan (fig. 4.7) and Alphen (Vandenberghe and Krook, 1981) represent only parts of this cold stage. In any case it is concluded that the Gilze Member as a whole was formed during the Eburonian, Waalian and Menapian stages.

Previously the St. Lenaarts Formation in Belgium has been interpreted as Weichselian, because of the presence of cold pollen spectra and their position under Weichselian eolian sands (De Ploey, 1961; Haest, 1985; Haest et al., 1986). In our opinion these sands overlying the Turnhout Member can be correlated lithostratigraphically with the Gilze Member. Since the Gilze Member in its turn is overlain by the Sterksel Formation (Appelenberg) and contains Early-Pleistocene warm temperate pollen assemblages, it is evident that the St. Lenaarts Formation is of Early-Pleistocene age as well (§2.5.8). Furthermore, the Weichselian date of a peat-layer in the sand (55.300 $\pm$ 700 BP; Haest et al., 1986) is calculated on the basis of a very low C-14 concentration, which could as well result from a very slight contamination of dead carbon with younger material.

Bavelian

In the neighbourhood of Bavel clay- and sand-beds of the Bavel Member erosively overlay the Gilze Member (fig. 2.5). The clay (2 to 10 m below the surface) is characterized by very high values of dry, thermophilous trees (up to 40%), which point to a warm interglacial climate (fig. 4.8). Especially the high *Carpinus* (up to 20%), *Tsuga* (up to 11%) and *Eucommia* content differentiates the Bavel Member from all underlying Early-Pleistocene members in the area (compare fig. 4.1, 4.3, 4.6, 4.8). In accordance with Zagwijn and De Jong (1984: diagram Bavel Ia) the clay of the Bavel Member is correlated with the Bavel Interglacial (Bv3b) of the Bavelian stage.
The clay of the Bavel Member has been eroded by channels. The pollen assemblage at 1.75 m in the upper part of the channel-fill is comparable with the underlying clay, although *Betula* content is higher (26%). It is therefore incorporated in the Bavelian Interglacial. This observation contradicts the results of Zagwijn and De Jong (1984, diagram Bavel Ia: 3.55 m: Linge Glacial), who found a cold pollen spectrum in the upper part of the sand in a neighbouring pit.
The upper two spectra in fig. 4.8 have been taken from cryoturbated peat-layers just below a gravel-bed (probably Beuningen gravel-bed). These are interpreted as Weichselian. The high content of thermophilous trees (20-30%) is explained by the reworking of pollen from the underlying Bavel Member.
To conclude, deposits of the Bavel Interglacial are characterized by a high *Tsuga* and *Carpinus* content. This pollen assemblage is restricted to the northeastern part of the study area, close to the Central Graben and it has not been found in western Noord-Brabant and adjacent northern Belgium.

4.3.2 Magnetostratigraphy

Magnetostratigraphy is mainly based on changes of the magnetic polarities in rock strata. These so-called reversals of the earth magnetic field are quite well established (Mankinen and Dalrymple, 1979; Lowrie and Alvarez, 1981). They occur in a short time period (1000-10.000 years) and they therefore provide excellent time markers. Furthermore, polarity changes are registered globally in different lithologies. Marine and continental deposits can therefore be correlated by the presence of essentially isochronous polarity changes. With the aid of the well dated reversals, the biostratigraphical units (pollen zones) can be connected with the worldwide magnetostratigraphic time scale. However, it should be remembered that the magnetostratigraphic units are based on time independant physical properties. Magnetostratigraphic units are therefore not chronostratigraphic units, in spite of their worldwide applicability (Hus, 1988).
One of the major problems of geomagnetic dating in continental deposits is the lack of a continuous sedimentary sequence, in comparison to the deep sea record. Therefore, the geomagnetic record on the continent is often fragmentary. Geomagnetic reversals are found occasionally in different lithostratigraphic units of which the mutual relation is not always known. Furthermore, because of the repetitive nature of the reversals, it is difficult to establish which polarity zone is involved. Additional (i.e. biostratigraphic) information is required then to identify the reversals with the transition of specific magnetozones.
Several paleomagnetic investigations have been performed of Early-Pleistocene sediments in The Netherlands and Belgium (Van Montfrans, 1971; Zagwijn et al., 1971; Hus et al., 1976; Prakash Chandra Adhikary, unpubl.; Hus, 1988). Van Montfrans (1971) investigated several exposures in Noord-Brabant and northern Belgium, which are evaluated first, before the new results are presented.

4.3.2.1 Evaluation of former paleomagnetic investigations (Van Montfrans, 1971)

The geomagnetic results of Van Montfrans (1971) have been summarized in table 4.2. The upper part of the table contains data obtained by Van Montfrans from Noord-Brabant and northern Belgium. In the lower part of the table his results from exposures in the neighbourhood of Tegelen in the Dutch province of Limburg (stratotype of the Tegelen Formation) are added for comparison purposes. At the right side of the table litho- and biochronostratigraphic interpretations are given according to the present study.
The Dorst and Bavel locations were formerly interpreted as Waalian C, but were later reinterpreted as Bavelian (Zagwijn and De Jong, 1984). Because of this reinterpretation the Jaramillo magnetozone (normal polarity) shifted from the Waalian to the Bavelian, which illustrates the problems of bio- and magnetostratigraphy.
The clay-bed in location the Chaamse Bossen is correlated in the present study with the Gilze Member (Kedichem Formation) (see fig. 2.8), instead of the Tegelen Formation. The new Waalian B/Menapian date is inferred from the cold pollen spectra in the upper part of the clay-bed in Gilze (fig. 4.6: Gilze pit 1), which is in the neighbourhood of the Chaamse Bossen.
The sediments at the Wouwse Plantage probably date from the Tiglian and the Eburonian (see App.), instead of the Tiglian.
The paleomagnetic results in the upper part of the Tegelen Formation

Table 4.2: Former paleomagnetic investigations in Noord-Brabant and northern Belgium (after Van Montfrans, 1971).

VAN MONTFRANS, 1971					KASSE, this study	
	LOCATION EXPOSURE	CRYO-TURBATION	BIOCHRONO-STRATIGRAPHY	POLARITY + (normal) — (reversed)	MEMBER	BIOCHRONO-STRATIGRAPHY
NOORD – BRABANT	DORST	—	upper WAALIAN C	+	BAVEL	BAVELIAN
	BAVEL	—	upper WAALIAN C	+	BAVEL	BAVELIAN
	CHAAMSE BOSSEN	—	TIGLIAN ?	+	GILZE	WAALIAN B?
	WOUWSE PLANTAGE	—	TIGLIAN ?	rejected	GILZE / TURNHOUT	EBURONIAN / TIGLIAN
	OSSENDRECHT	—	upper TIGLIAN	rejected	WOENSDRECHT	TIGLIAN C6
	WERNHOUT	—	TIGLIAN	rejected	TURNHOUT	TIGLIAN C6
NORTHERN BELGIUM	DE TOEKOMST (BEERSE)	yes upper part	WAALIAN	+ upper part	TURNHOUT	TIGLIAN C6
	FRANCISCUS (BEERSE)	yes upper part	WAALIAN	— lower part	TURNHOUT	TIGLIAN C6
	DE TOEKOMST	—	EBURONIAN	+ upper part — lower part	BEERSE	TIGLIAN C4
	FRANCISCUS	yes upper part	EBURONIAN	rejected	BEERSE	TIGLIAN C4
	DE TOEKOMST	—	TIGLIAN	—	RIJKEVORSEL	TIGLIAN C3
	FRANCISCUS	yes upper part	TIGLIAN	— lower part	RIJKEVORSEL	TIGLIAN C3
LIMBURG	MAALBEEK	—	EBURONIAN	rejected		
	MAALBEEK	—	lower EBURONIAN	—		
	KURSTJENS	—	TIGLIAN / EBURONIAN	+		
	WAMBACH	—	TIGLIAN-EBURONIAN	+ upper part		
	LAUMANS	—	TIGLIAN-EBURONIAN	+ (upper) part — (lower)		
	OBEL	—	TIGLIAN C	+		
	EGYPTE	—	TIGLIAN C3-C4ab	+		
	VAN CLEEF	—	TIGLIAN A	+ (middle)		

(our Woensdrecht and Turnhout Members) at Ossendrecht, Wouwse Plantage and Wernhout were rejected by Van Montfrans.
The Rijkevorsel, Beerse and Turnhout Members in the neighbourhood of Beerse (pits De Toekomst and St. Franciscus) have been redated by us as respectively Tiglian C3, Tiglian C4 and Tiglian C5 (§4.3.1), instead of Tiglian, Eburonian and Waalian. The normal (+) polarity obtained by Van Montfrans in the upper part of the Turnhout Member in pit De Toekomst was found in a 0.5-1.0 m thick cryoturbated layer (Paepe and Vanhoorne, 1970, fig. 4). The base of the Turnhout Member in pit St. Franciscus was characterized by a reversed polarity. The Beerse Member showed a normal polarity for the upper part and a reversed polarity for the lower part of the member. The Rijkevorsel Member revealed a reversed polarity (De Toekomst, St. Franciscus).
The polarity of the Turnhout, Beerse and Rijkevorsel Members obtained by Van Montfrans (1971) in pits De Toekomst and St. Franciscus and by Hus et al. (1976) for the latter is different from the results in the Tegelen area. The Turnhout, Beerse and Rijkevorsel Members seem to be dominated by a reversed polarity (normal polarity in the cryoturbated upper part of the Turnhout Member), whereas time equivalent Tiglian deposits around Tegelen are characterized by normal (+) polarities.

4.3.2.2 Paleomagnetic results from the study area

The paleomagnetic samples have been taken from clay-beds in four exposures and nine borings. Although sand-layers possess a measurable magnetic signal, the results are less reliable due to remagnetization risks. The exposures at Bavel, Ravels and Gilze were sampled continuously and the inclination and azimuth (declination) were measured accurately with standard equipment developed by Prof. Dr. J. Hus at the Centre de Physique du Globe in Dourbes (Belgium). Exposure Meerle and the borings have been sampled discontinuously. The azimuth of the samples from borings is not very reliable, since the coring equipment may have rotated during lowering in the borehole.
At first, the natural remanent magnetization (NRM), which is the remanence "in situ", has been measured in all samples. It is assumed that the paleomagnetic directions are impressed in the sediment during the sedimentation or shortly afterwards, when the deposits are still unconsolidated. The remanence measurements were performed with a three-axis SCT superconducting magnetometer. Each sample was measured 4 times (prisms) or 6 times (cubes) and the mean natural remanent magnetization intensity and direction were calculated. After deposition or during storage of the samples in the laboratory the natural remanent magnetization can be influenced by the presence of the geomagnetic field and a viscous remanence may build up spontaneously. This viscous magnetization can be removed by alternating field demagnetization in order to find the original magnetization of the sample (so-called characteristic remanent magnetization). 18 samples were first selected from different members and beds for demagnetization tests. These pilot samples were demagnetized in 13 steps (from 0 to 700 Oersted) in order to establish the stability of the magnetization. The individual members required different demagnetization values ranging from 150 to 300 Oersted to remove the viscous component of the remanent magnetization.
After the demagnetization tests all samples were partially demagnetized according to the values obtained from the tests. The Rijkevorsel and Turnhout Members were demagnetized in alternating fields of 200 to 250 Oersted; the Gilze Member in 300 Oersted and the Bavel Member in 150 Oersted. Then the samples were measured again and the mean values for

the remanent magnetization after alternating field demagnetization were computed. The paleomagnetic measurements were plotted subsequently in a diagram, containing inclination, declination and magnetic intensity of the samples.
From each pair (inclination, declination) of each sample the geographic coordinates of the corresponding virtual geomagnetic north pole (VGP) were calculated. When the pole occurs in the northern hemisphere (or southern hemisphere), near to the geographic north pole (south pole), the magnetization is normal (reversed). Allowing for the secular variation, the geomagnetic field is considered as normal (reversed) when the VGP plots within a circle of 40° around true north (south). During a reversal intermediate directions result due to a shift of the geomagnetic pole from one hemisphere to the other.
Detailed information concerning the historic outline of paleomagnetic dating, sampling techniques, measurement errors and rejection criteria has previously been given by Van Montfrans (1971).

Exposure Bavel (fig. 4.9) (Bavel Member)
A thick (max. 8 m) clay-layer, pollen-analytically equivalent to the Bavel Interglacial of the Bavelian stage (§4.3.1), was investigated. A negative inclination and a southward declination (reversed polarity) was found in the lower part; a positive inclination and northward declination (normal polarity) in the upper part of the clay-layer. The transition zone (5-8 m below the surface) is characterized by an alternation of positive and negative inclinations. The tree roots in the upper part of the clay-layer do not affect the normal polarity. The transition zone is interpreted as a change in polarity from reversed to normal. The magnetic intensity is low with respect to the over- and underlying, normal and reversed units. According to several studies (see Hus, 1988) a reversal of the earth magnetic field lasts between 1000 and 10.000 years and probably near to the lower limit.
The magnetic results confirm the results of Van Montfrans (1971: Bavel) and Zagwijn and De Jong (1984: Bavel I and II), who found a reversed magnetozone at the base and a normal magnetozone in the middle and upper part of the Bavel Interglacial deposits.

Exposure Gilze (fig. 4.10) (Gilze Member)
The upper 3 m of the Gilze Member at this location have been analysed. The investigated section is lithologically heterogeneous (see App.). The top strata (1.3-2.2 m) are intensively cryoturbated, probably in the Weichselian. A very fine sand-layer separates the cryoturbated clay from the underlying compact, crumbly clay, which contains humic/peaty soil horizons and becomes coarser grained downwards.
The cryoturbated clay, the sand-layer and 5 samples from the crumbly clay reveal a positive inclination and an intermediate polarity. Below 2.8 m positive and negative inclinations both occur. Declination is to the south and a predominantly reversed polarity is inferred for the lower part of the section. The change in polarity occurs approximately 60 cm below the base of the cryoturbation level at the sampling site.

Exposure Meerle (fig. 4.11) (top Turnhout Member)
The investigated clay-layers (3 m thick) belong to the upper part of the Turnhout Member (see appendix). Cryoturbation of Weichselian age reaches to 1.1 m below the boundary of the Twente Formation and the Turnhout Member (Beuningen gravel-bed) at the sampling location. The samples have been taken from 11 boxes (4 samples from one box), approximately 20 cm apart. The section shows a normal polarity in the upper part, an interval with intermediate polarities and a reversed polarity

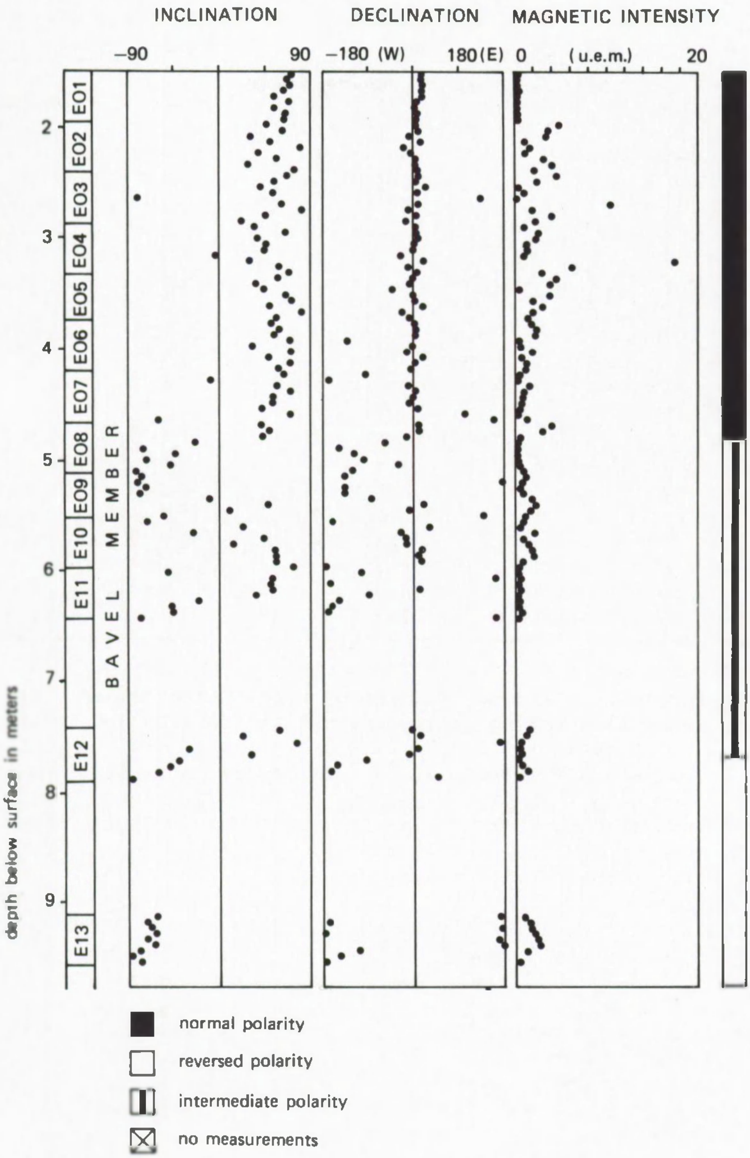


Fig. 4.9: Inclination and declination of the Bavel Member at Bavel after alternating field (AF) demagnetization (150 Oersted).

in the lower part of the clay-layers. The reversal occurs below the base of the cryoturbation level in the 50 cm thick interval with intermediate polarity.

Exposure Ravels (fig. 4.12) (top Turnhout Member)
The section has been sampled and analysed by Prof. Dr. J. Hus. He allowed us to present the results in this study. The samples have been taken from the upper part of the Turnhout Member (2 m). The "uncleaned"

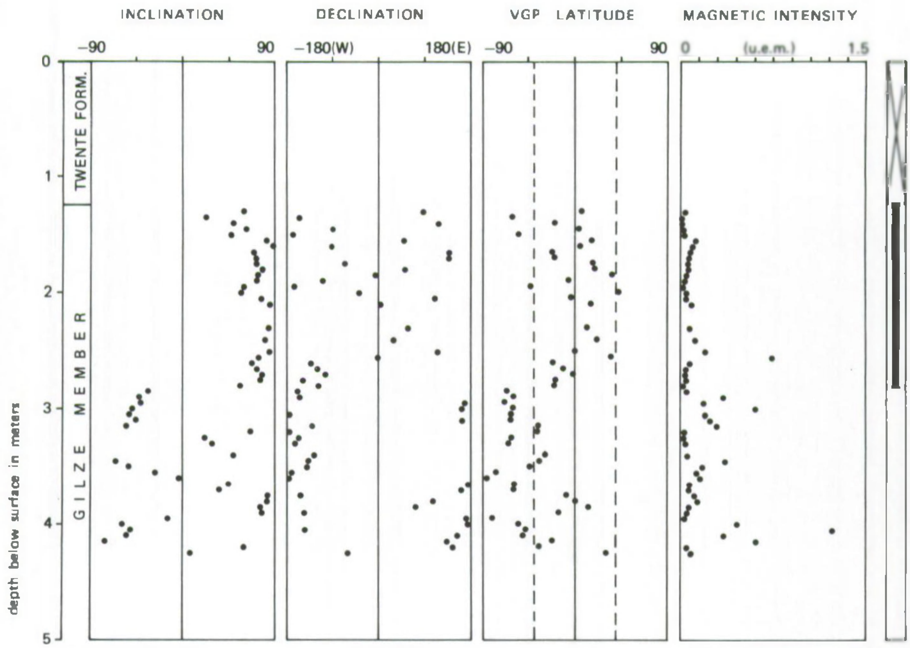


Fig. 4.10: Inclination and declination of the Gilze Member at Gilze after alternating field demagnetization (200-300 Oersted).

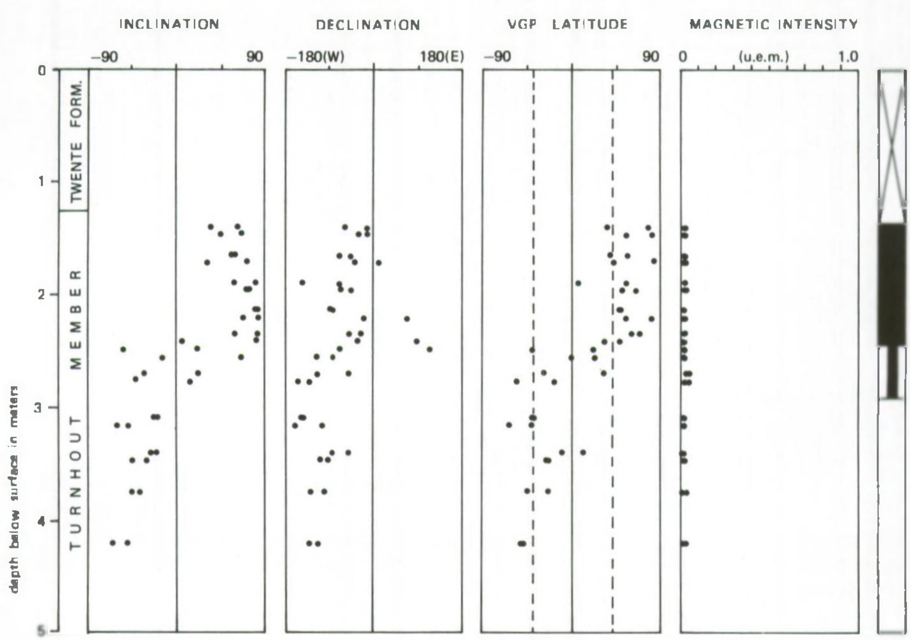


Fig. 4.11: Inclination and declination of the Turnhout Member at Meerle after alternating field demagnetization (300 Oersted).

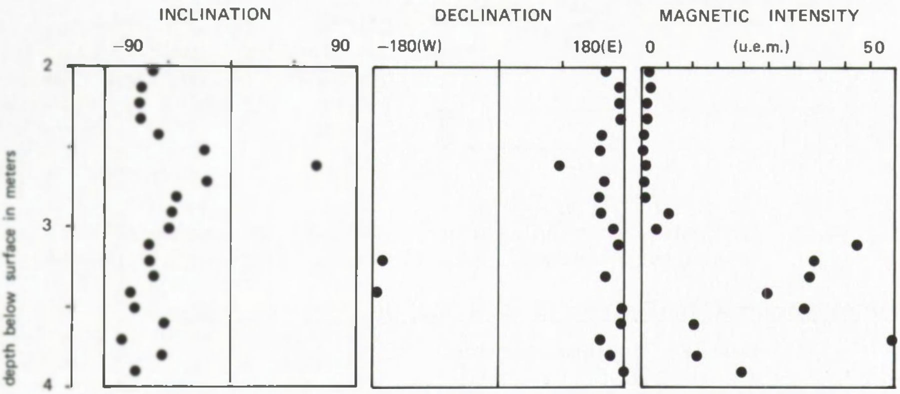
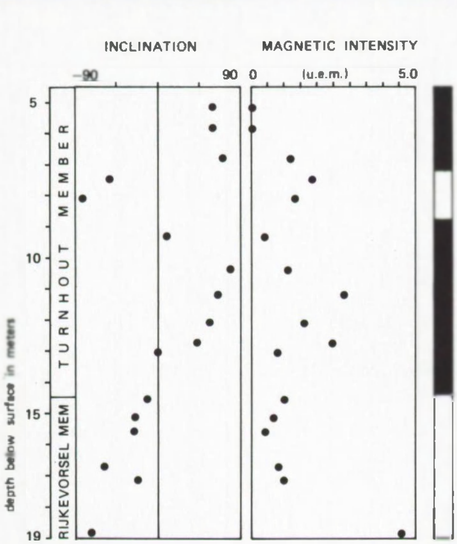


Fig. 4.12: NRM of the Turnhout Member at Ravels.

Natural Remanent Magnetization (NRM) of the samples indicate an overall negative inclination and a southward declination, which point to a reversed geomagnetic field. The one sample with a positive inclination might indicate a thin normal polarity zone, but the declination (85°) is still very large. The low magnetic intensity in the upper part and the high intensity in the lower part of the investigated section probably point to the presence of two clay-layers.
It is stressed that Weichselian cryoturbations did not affect the top of the Turnhout Member here, since a 1-1.5 m thick layer of fluviatile deposits of the Gilze Member is present between the Turnhout Member and the Weichselian cryoturbations.

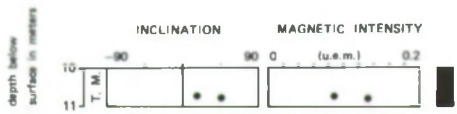
Boring Achtmaal (fig. 4.13) (Turnhout and Rijkevorsel Members)



17 samples have been analysed. The azimuth (declination) has not been established in the field. The Turnhout Member shows both positive and negative inclinations. The upper sample of the Turnhout Member is perhaps unreliable, due to remagnetization risks just below the erosional contact of the Eindhoven Formation and the Turnhout Member. The other samples of the Turnhout Member contain predominantly a positive inclination (normal polarity), especially in the lower part of the Turnhout Member (see also Meerle Slikgat).
The Rijkevorsel Member is characterized by negative inclinations and doubtlessly has a reversed polarity.

Fig. 4.13: Inclination of the Rijkevorsel and Turnhout Members in boring Achtmaal after alternating field demagnetization (300 Oersted).

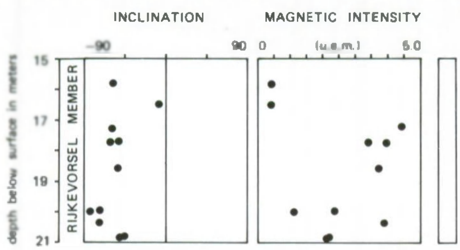
Boring Kalmthoutse Hoek (fig. 4.14) (base Turnhout Member)



The two samples at the base of the Turnhout Member display a positive inclination (declination is unknown), which is interpreted as having a normal polarity.

Fig. 4.14: Inclination of the Turnhout Member in boring Kalmthoutse Hoek after alternating field demagnetization (250 Oersted).

Boring Wernhout Maalbergen (fig. 4.15) (Rijkevorsel Member)



The samples were all derived from the Rijkevorsel Member. Only negative inclinations have been found, which indicate a reversed polarity.

Fig. 4.15: Inclination of the Rijkevorsel Member in boring Wernhout Maalbergen after alternating field demagnetization (200 Oersted).

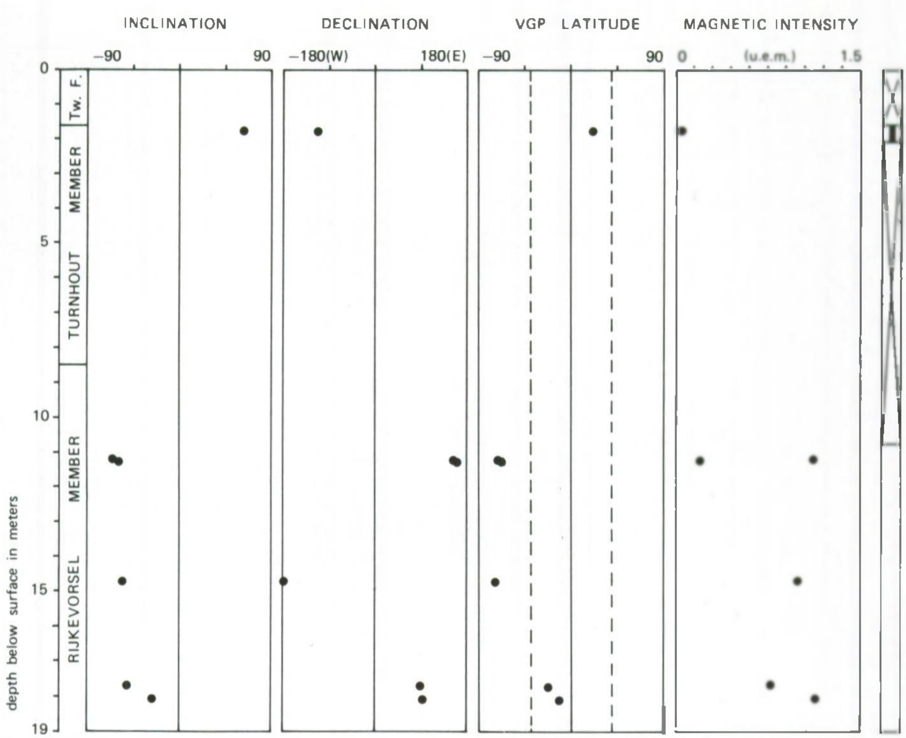


Fig. 4.16: Inclination and declination of the Rijkevorsel and Turnhout Members in Wortel after AF demagnetization (200 Oersted).

Boring Wortel (fig. 4.16) (Turnhout and Rijkevorsel Members)

Six samples have been investigated. The one sample of the Turnhout Member showed a positive inclination (intermediate polarity). Because of its position close to the Weichselian disconformity this result may be influenced by Weichselian remagnetization.

The 5 samples of the Rijkevorsel Member clearly have negative inclinations, i.e. a reversed polarity.

Boring Bolk (fig. 4.17) (Rijkevorsel Member)

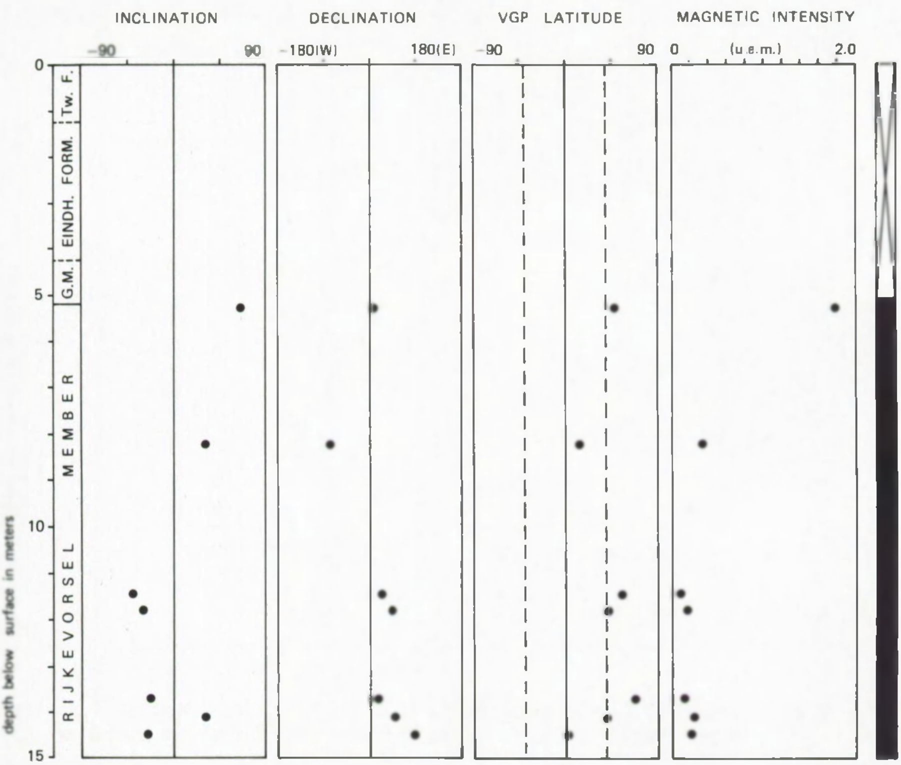


Fig. 4.17: Inclination and declination of the Rijkevorsel Member in boring Bolk after alternating field demagnetization (250 Oersted).

The samples are part of the Rijkevorsel Member. The Turnhout Member is missing here because of erosion in the Mark valley (see fig. 2.5). The inclinations are predominantly negative at the base and positive in the upper part of the member. The polarity seems to be intermediate or normal. The upper sample may be influenced by its position directly under the erosional boundary, which separates the Eindhoven Formation or Gilze Member from the Rijkevorsel Member.

Boring Zwart Water (fig. 4.18) (Turnhout Member)

The section has been taken from a thick, clayey facies of the Turnhout Member (comparable to Meerle Slikgat). The upper part of the member is probably equivalent to the Turnhout Member in the nearby clay-pit

Ravels. The negative inclinations and southward declinations in the upper part of the member point to a reversed polarity. The negative inclinations and northward declinations in the lower part result in an intermediate to normal polarity.

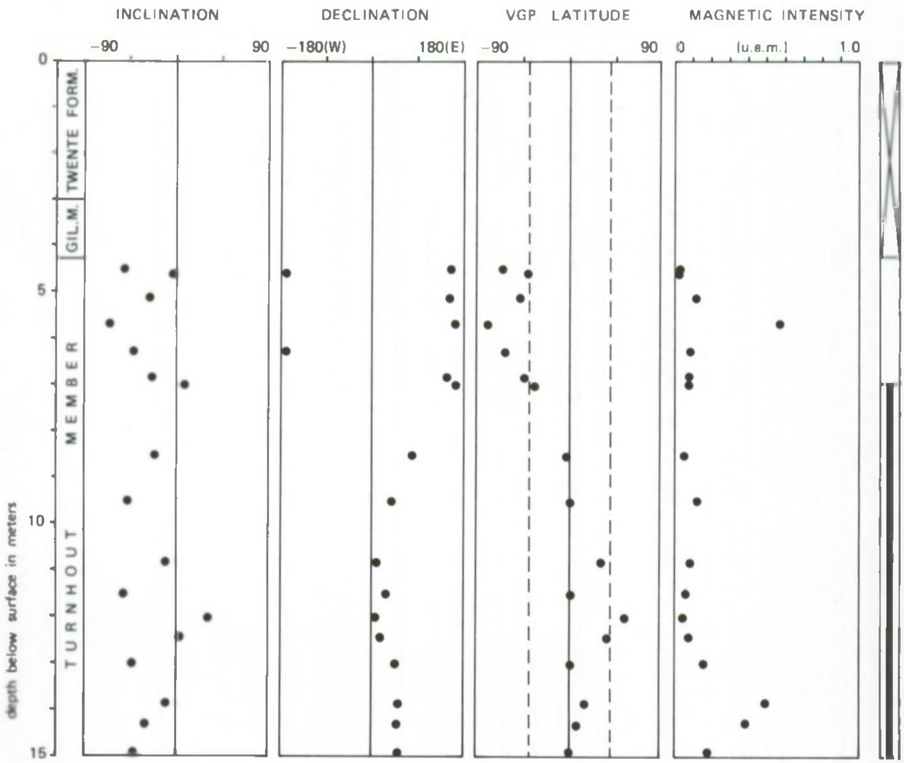


Fig. 4.18: Inclination and declination of the Turnhout Member in boring Zwart Water after alternating field demagnetization (250 Oersted).

Boring Meerle Slikgat (fig. 4.19)(Gilze, Turnhout, Rijkevorsel Members)

The Gilze Member (upper sample) reveals a negative inclination.

The Turnhout Member is almost completely fine-grained (clay-silt) at Meerle Slikgat (like in Achtmaal, Zwart Water). Therefore, geomagnetic changes within the Turnhout Member have been registered well. In other places only the clayey upper part of the Turnhout Member could be investigated, because the middle and lower part of the member consists of sand (fig. 2.5: Meerle, Wortel, Chaam Kapel). The upper samples of the Turnhout Member at Meerle Slikgat have a negative inclination (reversed and intermediate polarity). The lower samples show a positive inclination (intermediate polarity).

The Rijkevorsel Member (2 samples) is again characterized by a negative inclination (reversed polarity).

Boring Chaam Kapel (fig. 4.20) (Gilze, Turnhout, Rijkevorsel Members)

The Gilze Member has a positive inclination, perhaps because of its position 35 cm below the Weichselian unconformity with cryoturbation

phenomena.
The three samples from the upper part of the Turnhout Member clearly have a negative inclination, illustrating the reversed polarity. The sand in the middle and lower part of the Turnhout Member excludes reliable magnetic measurements.
The Rijkevorsel Member (2 samples) is characterized by negative inclinations and an intermediate to reversed polarity, as in Meerle Slikgat.

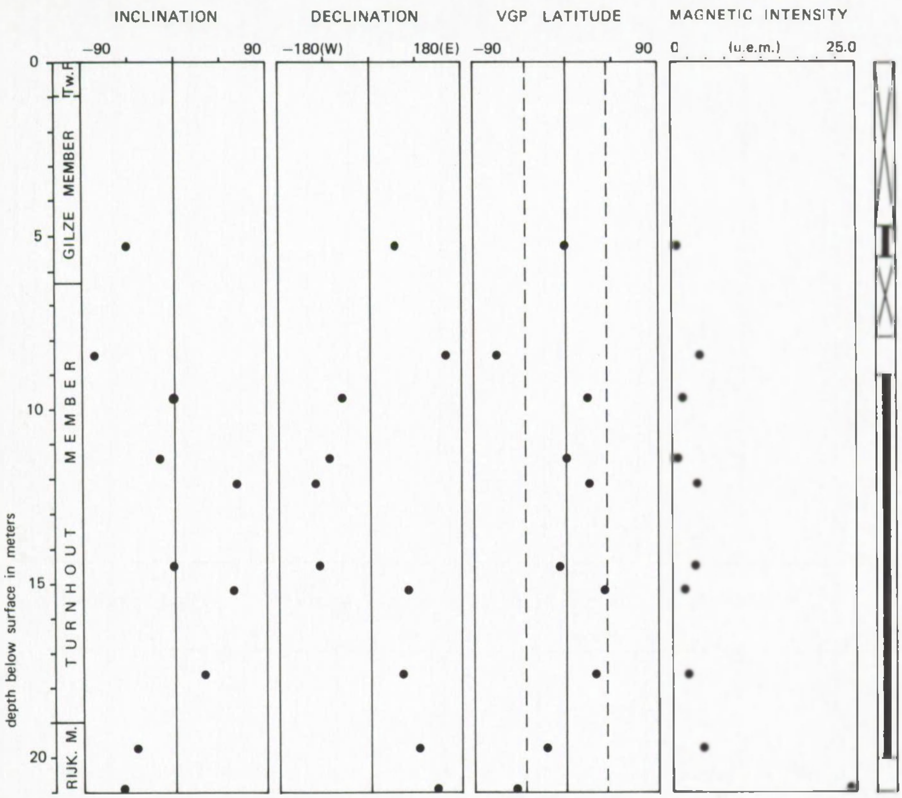


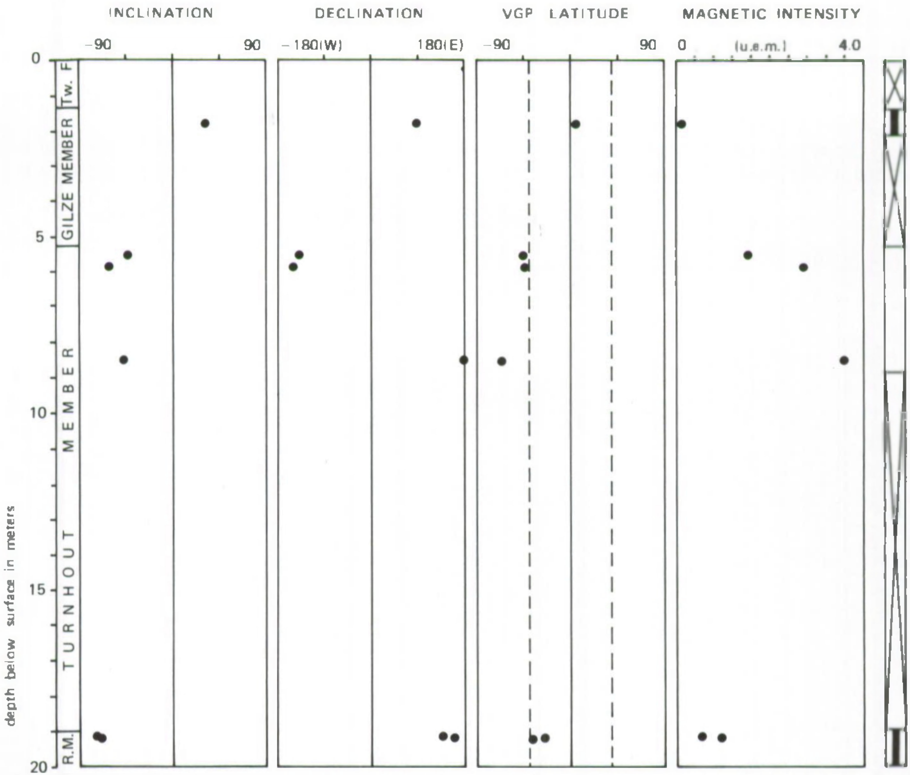
Fig. 4.19: Inclination and declination of the Rijkevorsel, Turnhout and Gilze Members in Meerle Slikgat after alternating field demagnetization (200 Oersted).

Boring Snijders-Chaam (fig. 4.21) (Gilze Member)

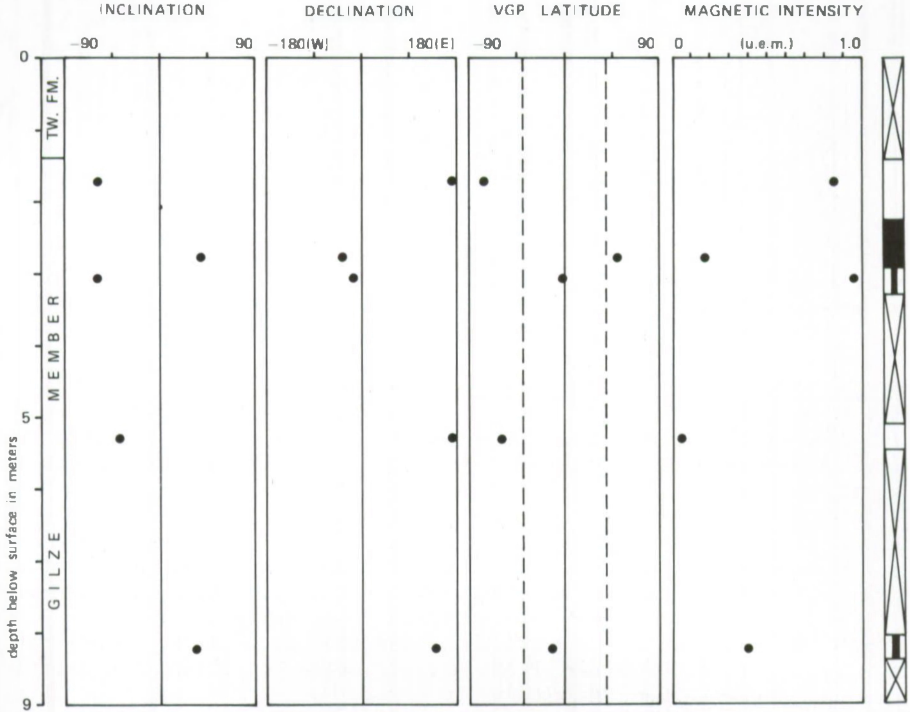
The 5 samples were taken from 3 different clay-layers, which are all included in the Gilze Member. The results are not very consistent. The upper three samples are from one clay-bed: two have a negative and one a positive inclination.

Fig. 4.20: Inclination and declination of the Rijkevorsel, Turnhout and Gilze Members in Chaam Kapel after alternating field demagnetization (250 Oersted)(next page, top).

Fig. 4.21: Inclination and declination of the Gilze Member in Snijders-Chaam after alternating field demagnetization (300 Oersted) (next page, base).



INCLINATION
DECLINATION
VGP LATITUDE
MAGNETIC INTENSITY
-90
90
-180(W)
180(E)
-90
90
0
(u.e.m.)
4.0
0
5
10
15
20
depth below surface in meters
Tw. F
GILZE MEMBER
TURNHOUT MEMBER
R.M.


INCLINATION
DECLINATION
VGP LATITUDE
MAGNETIC INTENSITY
-90
90
-180(W)
180(E)
-90
90
0
(u.e.m.)
1.0
0
5
9
depth below surface in meters
TW. FM.
GILZE MEMBER

4.3.2.3 Interpretation of the geomagnetic measurements

The results of the magnetic measurements are summarized in table 4.3. The relation between the Weichselian periglacial structures (frost wedges, ice-wedge casts, convolutions) and magnetic polarity is visualized in fig. 4.22.

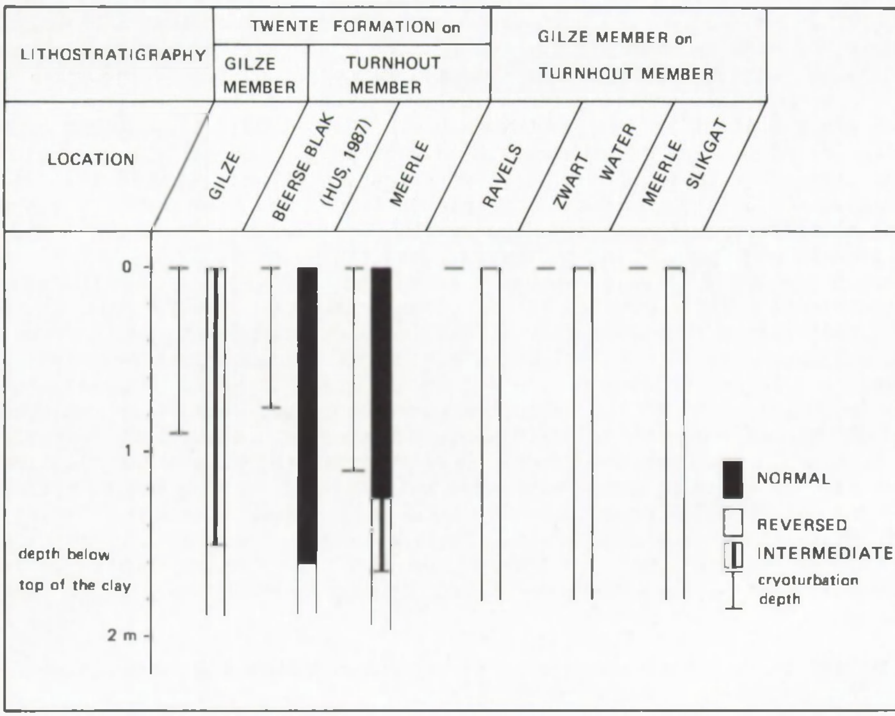


Fig. 4.22: Relation between the Weichselian periglacial structures and the magnetic polarity.

In Meerle and Gilze, where the clays of the Turnhout and Gilze Members are directly overlain by the Twente Formation, the magnetic change coincides with the presence of cryoturbated levels (fig. 4.22) (Van Montfrans, 1971: pit de Toekomst, pit Sint Fransiscus). It is true that the reversal occurs some distance below the base of the macroscopically visible cryoturbation at the sampling site. According to Hus (1988) in exposure Beerse Blak the change in polarity occurs at the top of a humic layer, well below the cryoturbated upper part of the Turnhout clay (see also app.). He warns against the influence of periglacial activity and proposes to apply a field test on deformed strata. Nevertheless, the positive inclinations above the humic bed are interpreted as a normal magnetosubchron in the Matuyama chron (Hus, 1988). However, it is not certain whether the thickness of the melting Weichselian permafrost was equal to the depth of the cryogene structures as seen nowadays in the exposures. Possibly remagnetization and load casting occurred during the melting of the ice-rich topzone of the permafrost, while melting of the underlying ice-poor permafrost zone resulted in (partial) remagnetization without load casting.

In those situations where the Turnhout Member is covered by the Gilze Member and Weichselian cryoturbation phenomena did not influence the clay-beds of the Turnhout Member (e.g. Ravels, Zwart Water, Meerle Slikgat, Chaam Kapel), no polarity changes are registered and the Turnhout Member shows a reversed polarity to the top. Furthermore, it is remarkable that sections influenced by later cryoturbation always reveal a change from a reversed to a normal polarity in different lithostratigraphic units (Gilze as well as Turnhout Members). New magnetozones would be necessary in the existing magnetostratigraphic scale, if each reversal in the top of the different lithostratigraphic units is regarded as a primary feature. Although other magnetosubzones, not given in the polarity time scale of Mankinen and Dalrymple, are certainly present in the Matuyama epoch (Hus, 1988), it does not seem wise to define them in problematic (cryoturbated) layers.
The discussion as to whether cryoturbation can effect magnetic polarity depends to a large extent on the plasticity of the clay-beds concerned during the cryoturbation. If the involutions are merely a rigid displacement and overturning of layers, then the original polarity will be partly preserved, giving however a large scatter in the magnetization directions. If cryoturbation involves complete liquefaction of the clay-bed then a reorientation of the magnetic components will occur, according to the then existing magnetic field (remagnetization). In the case of Weichselian remagnetization of a Tiglian clay-bed (Meerle), the reversed polarity of the Matuyama magnetozone will have been replaced by the normal polarity of the Brunhes magnetozone. Laboratory tests by Vandenberghe and Van den Broek (1982) indicate that oversaturation and excess pore water pressure must have been present during the formation of the Weichselian convolutions. Under such circumstances with a loss of intergranular contacts in the liquefied layer, a complete remagnetization of the sediment is likely to occur. The latter conclusion is adopted in the interpretation of the paleomagnetic results in this study.

The results of the paleomagnetic investigations have been summarized in the following conclusions (table 4.3):
1. The Rijkevorsel Member is almost always characterized by a reversed polarity.
2. The polarity of the Beerse Member could not be investigated because of its sandy nature. Therefore, the magnetostratigraphic position of the Beerse glacial is unknown.
3. The polarity of the Turnhout Member is variable. A normal polarity was found locally at the base of the member. The upper part of the Turnhout Member is characterized by a reversed polarity (Ravels). The normal polarity, which is found in the cryoturbated top of the Turnhout Member, can in our opinion be explained by the remagnetization of a liquefied layer, during melting of the permafrost in the Weichselian. However, it cannot be excluded completely, that an "in situ" magnetozone with normal polarity is present in the top of the Turnhout Member, since the reversal is located below the cryoturbated zone and locally coincides with a lithological boundary (Hus, 1988: Beerse Blak).
4. The Gilze Member is characterized by a reversed polarity, but exceptions are regularly found (Chaam Kapel, Snijders-Chaam). The upper part of the Gilze Member at Gilze reveals an intermediate to normal polarity, which is also explained by remagnetization during the Weichselian.
5. The Bavel Member is characterized by a reversed polarity at the base and a normal polarity in the upper part of the member, separated by a well developed transition zone.

4.4 Synthesis of the chronostratigraphy

The combination of the bio- and magnetostratigraphy leads to the following conclusions (table 4.3):

Table 4.3: Summary of the chronostratigraphical results based on paleobotanical evidence and paleomagnetism (+ is normal polarity, - is reversed polarity, ± is intermediate polarity; signs between brackets are derived from cryoturbated beds).

LITHOSTRATIGRAPHY	CHRONOSTRATIGRAPHY		MAGNETOSTRATIGRAPHY													POLARITY TIME-SCALE
			BAVEL	GILZE	MEERLE	RAVELS	ACHTMAAL	KALMTHOUTSE HOEK	WERNHOUT MAALBERGEN	WORTEL	BOLK	ZWART WATER	MEERLE SLIKGAT	CHAAM KAPEL	SNIJDERS - CHAAM	BRUNHES
STERKSEL FORMATION	M.PL	CROMERIAN														0.72
BAVEL MEMBER	EARLY - PLEISTOCENE	BAVELIAN	+ −													0.91 JARAMILLO 0.97
GILZE MEMBER	EARLY - PLEISTOCENE	MENAPIAN		(+) −												MATUYAMA
		WAALIAN		−												
		EBURONIAN											-	±	+/-	
TURNHOUT MEMBER	EARLY - PLEISTOCENE	TIGLIAN C5			(+) −	− −	+ − +	+		±		− ±	- ±	−		1.66
BEERSE MEMBER	EARLY - PLEISTOCENE	TIGLIAN C4														OLDUVAI 1.87 Ma
RIJKEVORSEL MEMBER	EARLY - PLEISTOCENE	TIGLIAN C3					−		− −	−	±		−	±		

1. The Rijkevorsel Member has been formed in the Tiglian C3 phase and is characterized by the reversed polarity of the Matuyama magnetochron.
2. The Beerse and Turnhout Members have been dated respectively as Tiglian C4 and Tiglian C5. The normal polarity at the base of the Turnhout Member and perhaps in the Beerse Member (Van Montfrans, 1971: pit De Toekomst) is correlated with the Olduvai subchron (1.66-1.87 m.y.; Lowrie and Alvarez, 1981) of the Matuyama chron.
3. The Hoogerheide and Woensdrecht Members have been interpreted respectively as possibly Tiglian C3 and probably Tiglian C5.
4. The top of the Turnhout Member dates from the Tiglian C5, Tiglian C6 and the beginning of the Eburonian period and is younger than the Olduvai subchron (reversed polarity).
5. The Gilze Member has been deposited during the Eburonian, Waalian and Menapian bio-chronostratigraphical stages. The predominantly reversed polarity is evidence of deposition during the Matuyama magnetochron.
6. The Bavel Member dates from the Bavel Interglacial of the Bavelian stage. The magnetic reversal from a reversed to a normal polarity is interpreted as the base of the Jaramillo subchron of the Matuyama magnetochron, dated at 0.97 milj. years (Zagwijn and De Jong, 1984).

5. PROVENANCE OF THE SEDIMENT

5.1 Introduction

Provenance studies in The Netherlands are mainly based on the analysis of the heavy minerals (specific density higher than 2.9) and gravel components in the sediments (Edelman, 1933; Baak, 1936; Van Straaten, 1946; Zonneveld, 1947, 1948a; Maarleveld, 1956; Zandstra, 1969, 1978) and to a lesser extent on light minerals (Van Baren, 1934). The investigations are not of a very recent date and they sometimes led to different conclusions. Edelman (1933, 1938) concluded a Scandinavian and Meuse provenance for certain deposits in Noord-Brabant. Later Zonneveld (1947) stressed the importance of the Rhine and proposed a Rhine and Meuse origin of the sediments in eastern Noord-Brabant. According to Van Dorsser (1956) especially the Meuse extended its influence during the Early-Pleistocene into western Noord-Brabant. Zandstra (1969) investigated the gravel composition at Bavel and concluded a Scheldt provenance.
These differences are partly due to the correlation and comparison of units of different age. Since a detailed lithostratigraphic framework has been provided (chapter 2) it is now possible to investigate the provenance of the sediments of the individual units (fig. 5.1).

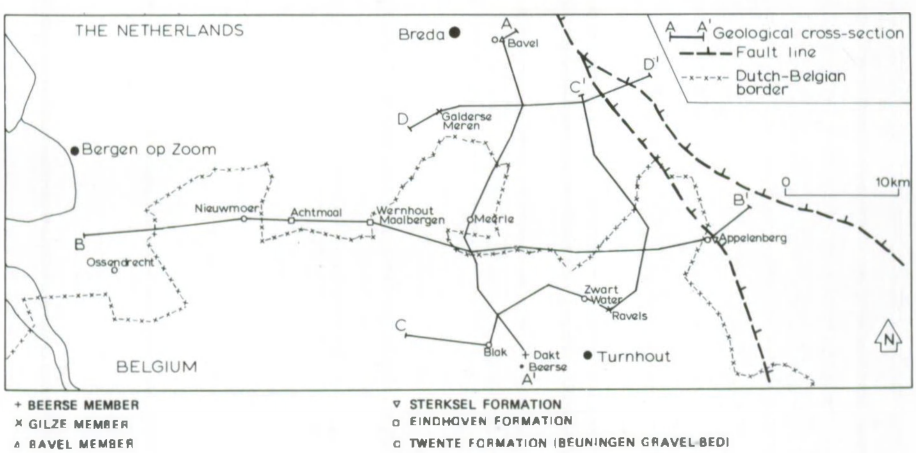


Fig. 5.1: Location map of the analysed gravel samples.

5.2 Methods

In this study heavy minerals (§2.5) and gravel from different members have been analysed. In general 200 transparent heavy mineral grains were analysed with grain-sizes ranging from 53 to 420 µm. In fine-grained deposits (clay-layers) the coarse silt fraction (32-53 µm) was counted as well, since few heavy mineral grains larger than 53 µm were available. The gravel has been analysed in 3 grain-size classes (3-5 mm; 5-8 mm; >8 mm). The >8 mm fraction is often of minor importance, indicating the fine-grained character of the gravel in western Noord-Brabant. 300 gravel components were determined, classified and compared with the gravel associations established by Zandstra (1978) (fig. 5.2).

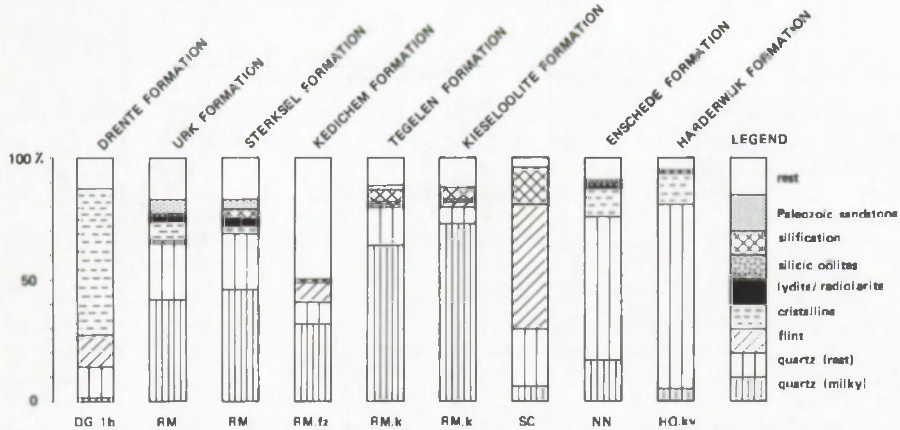


Fig. 5.2: Gravel associations in The Netherlands. The Paleozoic sandstone of the Kieseloölite and Kedichem Formation are included in the rest group (after Zandstra, 1978).

In the study area only the Rhine/Meuse (RM) and Scheldt (SC) gravel associations are to be considered. Glacial associations (DG) and gravel assemblages supplied by so-called eastern rivers from Germany (NN, HO) are added for comparison.

5.3 Grain-size fractionation and heavy mineral composition

Introduction

The provenance of the lithostratigraphic units in this study is essentially based on the analysis of unfractionated heavy mineral samples, for the following reasons:

1. It takes less time to analyse unfractionated samples.
2. If heavy mineral analysis is restricted to a single grain-size class (e.g. 105-150 µm), then it is impossible to analyse the very fine-grained samples.
3. Fractionation separates fine grains with high specific density from coarse grains with low specific density, which are hydrodynamically equivalent (Zonneveld, 1947).
4. Different grain-size classes may represent different sediment source areas.

However, Van Andel (1950) stressed the importance of the relation between the grain-size distribution and heavy mineral composition. For this reason a number of samples, characterized by a very different heavy mineral assemblage, were fractionated in five grain-size classes (53-77, 77-105, 105-150, 150-212, larger than 212 µm). The individual classes were analysed again.

Results (fig. 5.3, 5.4, 5.5)

In fractionated samples dominated by unstable heavy minerals alterite increases strongly in the coarser grain-sizes (fig. 5.3). Epidote and to a lesser extent hornblende dominate the finer fractions. Garnet occasionally reveals a peak value in the 77-105 µm class. Although the heavy mineral variations are large between the grain-size fractions, the heavy mineral composition as a whole is dominated by unstable heavy minerals (garnet, epidote, alterite, hornblende) in all fractions.

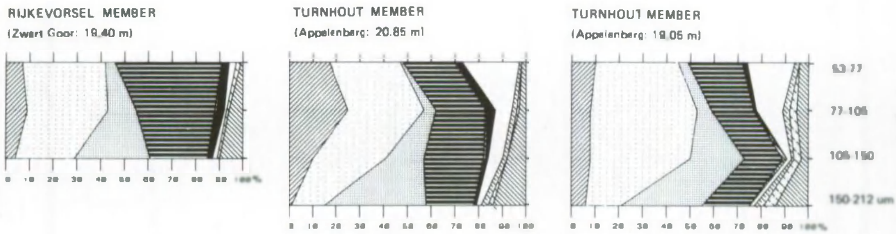


Fig. 5.3: Heavy mineral composition and grain-size fractionation of samples dominated by unstable heavy minerals (legend in appendix).

The fractionated samples dominated by stable heavy minerals show a strong increase of tourmaline, andalusite and staurolite in the coarser fractions (fig. 5.4). Staurolite appears to have its peak value around 150 µm. On the other hand zircon, rutile and anatase are strongly represented in the finest fractions. In spite of grain-size dependent variations the heavy mineral composition in all fractions is dominated by stable heavy minerals.

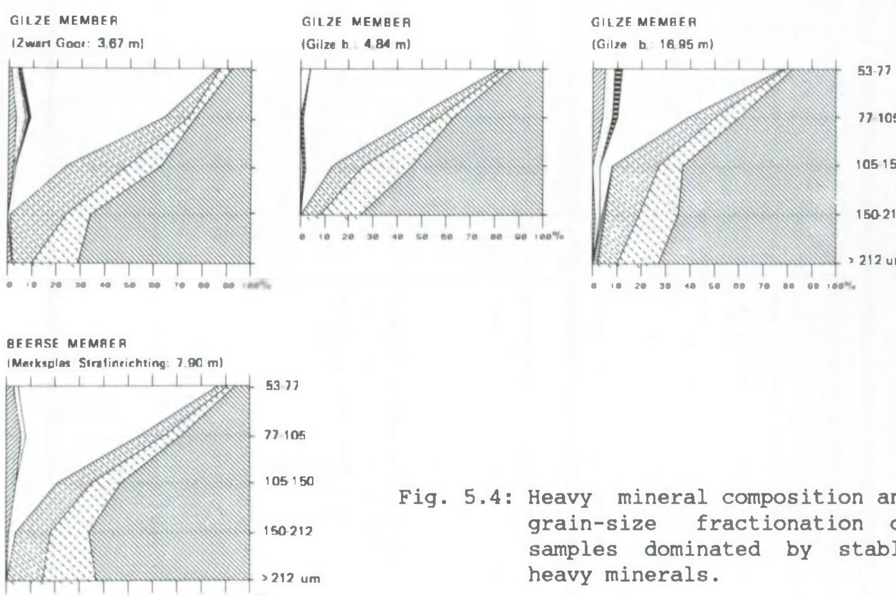


Fig. 5.4: Heavy mineral composition and grain-size fractionation of samples dominated by stable heavy minerals.

The third group of samples have a mixed heavy mineral assemblage (fig. 5.5). After fractionation the coarser fractions are dominated by stable heavy minerals (tourmaline, andalusite, staurolite), while the finer fractions are characterized by zircon, rutile and additional unstable heavy minerals, such as garnet, epidote and hornblende.

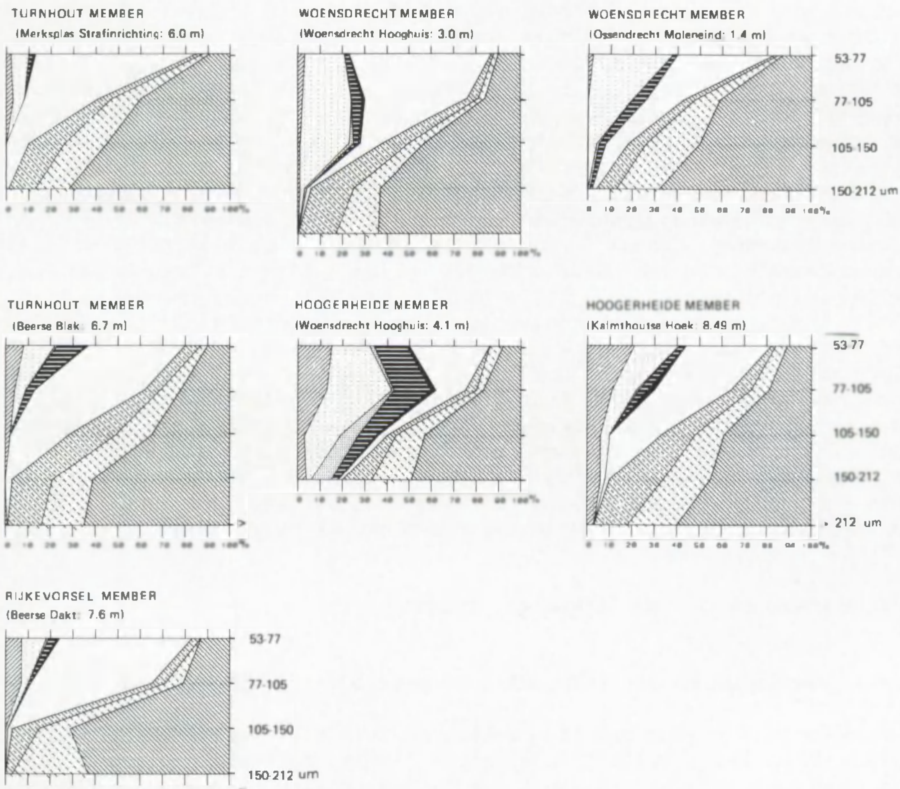


Fig. 5.5: Heavy mineral composition and grain-size fractionation of samples with a mixture of stable and unstable heavy minerals.

<u>Interpretation</u>

The results presented above reveal large grain-size dependent variations of individual heavy minerals. These variations are explained by differences in specific density (hydrodynamic equivalence), shape and original grain-size of the individual heavy mineral grains (Schuiling et al., 1985). Small grains with high specific density (e.g. zircon, rutile, garnet) are hydrodynamically equivalent to larger grains with lower specific density (e.g. tourmaline, alterite) and therefore occur in the same sample. Furthermore, certain minerals like zircon commonly occur as small mineral grains, while tourmaline normally has a larger original grain-size (Boenigk, 1983; Schuiling et al., 1985). The results further show that in the samples dominated by either stable or unstable heavy minerals (fig. 5.3, 5.4) the heavy mineral association as a whole is independent from grain-size. Unfractionated samples, dominated by stable or unstable heavy minerals, show the same dominance in different grain-size classes after fractionation. The heavy mineral assemblages of such unfractionated samples can therefore be used as a stratigraphic characteristic and they indicate a certain homogeneity in sediment provenance.

From the foregoing it appears that the differences in the heavy mineral composition of the third group, with mixed heavy mineral assemblages (fig. 5.5), can be explained only partly by differences in specific density of the minerals (e.g. zircon decrease and tourmaline increase in coarser fractions). The sharp decrease of unstable heavy minerals in coarser fractions cannot be explained however by the specific density of the minerals. It seems probable that in such cases two different sediment sources are involved; a finer grained one with unstable heavy minerals and a coarser grained one with stable heavy minerals. This situation is clearly present in the Turnhout Member at Merksplas Strafinrichting (fig. 5.5) in a sample from the base of a tidal channel of the Turnhout Member, where it overlies the Beerse Member with a clear erosive boundary. Medium sand with stable heavy minerals (see fig. 5.4) was reworked from the Beerse Member, while afterwards the channel was filled with fine sand and clay with more unstable heavy minerals (fig. 2.13). Furthermore, glaucophane is found infrequently in the samples with a mixed heavy mineral association. This mineral is known from the Alpes (pers. comm. Prof. Dr. J.L.R. Touret) and was supplied by the Rhine, which river transported predominantly unstable heavy minerals during the Pleistocene (Boenigk, 1983). It is therefore concluded that the unstable heavy minerals and glaucophane in the mixed heavy mineral association were delivered by the Rhine. The stable heavy minerals must have a different origin, which fact implies that two different sediment sources are involved in samples with a mixed heavy mineral composition.

5.4 Provenance of the lithostratigraphic units

5.4.1 Merksplas Member (according to previous investigations)

The heavy mineral and gravel composition have been described by Delvaux (1890-1891), Halet (1920), Huyghebaert (1961), Gullentops (1963), Paepe and Vanhoorne (1976) and Haest (1985). The Merksplas Member is characterized by a stable to mixed heavy mineral association. The gravel contains quartz, flint, silicic oölite, quartzite, psammite, schist, phyllite, arkose, grès (Tertiary) and cristalline.
According to Paepe and Vanhoorne the Merksplas Sands (as part of the Mol Sands s.l.) were deposited by an east-west flowing river system of the Rhine (and the Meuse). The presence of "grès tertiaire" and Eocene components, mentioned by Delvaux from the stratotype at Merksplas Strafinrichting, points to additional supply by rivers draining the Tertiary deposits in Central Belgium.

5.4.2 Rijkevorsel and Turnhout Members

Both members are characterized by an unstable heavy mineral association, dominated by garnet, epidote, alterite and hornblende (A- and H-association)(fig. 2.13 and 2.15). Edelman (1933) interpreted comparable assemblages as being of Fennoscandian origin. Later Zonneveld (1947) stressed that a similar heavy mineral composition was supplied by the Rhine. According to Boenigk (1983) the Rhine deposited this unstable heavy mineral association from the Late-Tertiary onwards (Reuverian), when a connection was established with the Alpine region. Before that period the Rhine transported stable heavy minerals such as zircon, rutile, staurolite and metamorphic minerals from the Rhine Graben area in Germany north of the Kaiserstuhl (Boenigk, 1983).
The presence of glaucophane (included in the hornblende group) in the Rijkevorsel and Turnhout Members supports a Rhine supply, since this

mineral is typical for the Alpine region, while it does not occur in Scandinavian rocks. Examination of samples from Tertiary deposits in Belgium, which were deposited before the Rhine was connected to the Alpes, did indeed not reveal any grain of glaucophane among the unstable heavy minerals (fig. 5.7).
At the base of the Rijkevorsel Member (App. Bolk) and at the southern margins of the Rijkevorsel and Turnhout Members (fig. 2.13) the stable heavy mineral content is higher. This phenomenon is explained by different sediment sources (see §5.3). The fine sediment is dominated by unstable heavy minerals of Rhine provenance (fig. 5.5). The coarser sediment contains more stable heavy minerals. The latter minerals were probably reworked from the Merksplas Member (channels at the base of the Rijkevorsel Member; e.g. in Bolk) or Beerse Member (channels at the base of the Turnhout Member; e.g. in Merksplas Strafinr.) or they were supplied by small, local rivers, which entered the brackish tidal and marsh environments from the south.
Towards the top of the Turnhout Member the content of unstable heavy minerals decreases (see e.g. Meerle). The increase of tourmaline in a fining-upward sequence is opposite to normal grain-size dependant heavy mineral changes (§5.3). During the final stage silting of the tidal environment, the influx of stable heavy minerals by rivers from the south probably became more important. It cannot be excluded, that some decrease in the unstable heavy mineral content in the upper part of the Turnhout Member has been caused by the postdepositional weathering of apatite and hornblende (Boenigk, 1983). The relative sea-level drop after the Tiglian period might be responsible for the leaching of easily soluable minerals.

5.4.3 Beerse Member

The Beerse Member is characterized by a high stable heavy mineral content (B-Limburg association)(fig. 2.13: Merksplas Strafinr.). The association is completely different from the underlying Rijkevorsel Member, which was supplied from the north and east and is found as a continuous deposit in the northern part of the area. The stable heavy minerals of the Beerse Member were most probably transported from the south by rivers and by the wind (§3.3.3). The Merksplas Sands and the Mol Sands, which crop out south of Beerse, are the closest sediment sources. Eocene and Oligocene deposits in Central Belgium may also be considered as a potential origin of the sediment (fig. 5.7).
The gravel of the Beerse Member is characterized by flint (29%), quartz (58%) and a small rest group (9%) (fig. 5.6). The association is comparable to the Scheldt(SC) gravel association (Zandstra, 1969) (fig. 5.2), although the quartz content is somewhat higher in the Beerse Member. Until now the SC-gravel was defined in the Beuningen gravel-bed at Bavel between Early-Pleistocene units and

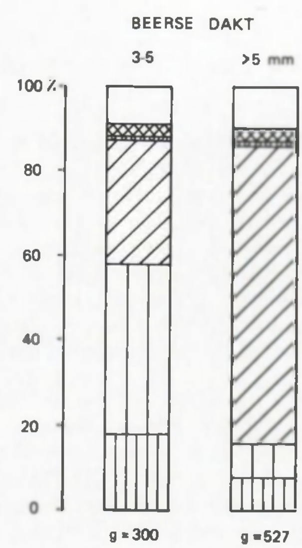


Fig. 5.6: Gravel composition of the Beerse Member at Beerse Dakt (legend in fig. 5.2; g=number of grains).

Weichselian coversands (Zandstra, 1969), which indicates a considerable hiatus. In such situations it is possible that the high flint and quartz content is due to chemical and mechanical weathering of other gravel components. The SC-gravel of the Beerse Member proves that the SC-association was already transported to the area during the Early-Pleistocene (Tiglian).

Typical tracer components of the Rhine or Meuse have not been found and, for that matter, a very high flint content is not found in the Early-Pleistocene terraces of the Meuse in Zuid-Limburg (Van Straaten, 1946). Generally a rather high content of ellipsoid, transparent quartz is present in the SC-association of the Beerse and Gilze Members and the Eindhoven and Twente Formations. Meuse gravel on the other hand contains a higher amount of irregularly shaped, milky quartz (see fig. 5.2).

The only source area that might be considered is the Central Belgian Tertiary region, which is drained by the Scheldt and smaller rivers such as the Dender, Zenne, Dijle and Gete. The marine Tertiary deposits in this region contain thin gravel-beds (transgression lag deposits), characterized by transparent quartz, flint and some quartzite (Gulinck, 1960; Tavernier and De Moor, 1974; Vandenberghe, 1977). The gravel components are well rounded and ellipsoid by coastal abrasion in the Tertiary near shore environment. The transparent quartz was probably not supplied from the Ardennes (more milky quartz), but originated from further south e.g. from the Vosges or Brittany (fig. 5.15). This supply would also explain the presence of staurolite and metamorphic minerals (andalusite, kyanite, sillimanite) in the Tertiary deposits (Demoulin, 1987), since the Paleozoic rocks of the Ardennes do not contain much of these minerals (Tavernier, 1947). In the Tertiary North Sea basin in Belgium this transparent quartz was mixed with flint abraded from the Cretaceous limestones in the southern Netherlands, Belgium and France. From Early-Pleistocene times onwards gravel and sand were eroded from the uplifted Tertiary strata, transported to the north, and deposited along the rim of the North Sea basin (fig. 5.15).

5.4.4 Hoogerheide and Woensdrecht Members

The members are characterized by a mixed to stable heavy mineral association (§2.5.6 and §2.5.7). The heavy mineral fractionation (§5.3: fig. 5.5) reveals that two sediment sources were involved. The finer fractions have a higher content of unstable heavy minerals. The presence of glaucophane points to an Alpine provenance of the fine-grained sediment. The coarser fractions show a higher amount of stable heavy minerals, which were formerly explained by a Meuse supply (Van Dorsser, 1956). This is unlikely, however, since sediments of the similar age (Tiglian) in the east (Rijkevorsel and Turnhout Members) are dominated by a Rhine heavy mineral association. Furthermore, the Meuse still flowed in a northeastern course through Zuid-Limburg during the Tiglian (Zagwijn and Van Staalduinen, 1975). The heavy mineral composition of the Woensdrecht and Hoogerheide Members strongly resembles the composition of certain Tertiary deposits in Belgium. Both members and the Tertiary Diest and Kesselberg Sands are characterized by a low garnet and alterite content in the unstable heavy mineral fraction (fig. 5.7) (Gullentops, 1963).

Hence it may be concluded that the stable to mixed heavy mineral association in western Noord-Brabant was supplied by the Rhine and from Tertiary deposits. In §3.3.5 and §3.3.6 the estuarine and tidal flat origin of the members was postulated. The fine, unstable, heavy miner-

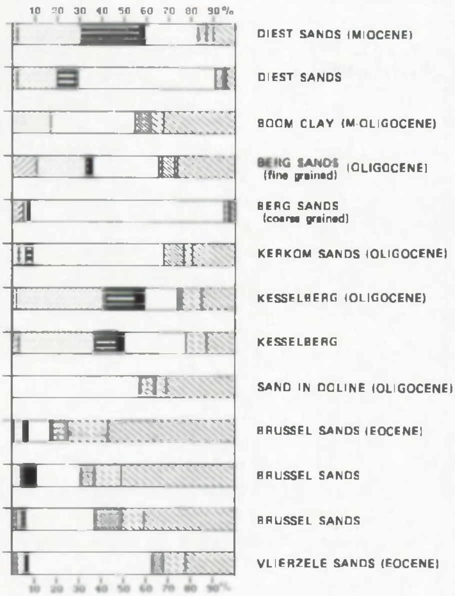


Fig. 5.7: Heavy mineral composition of Tertiary Formations in Belgium (legend in appendix).

als (Rhine provenance) were probably transported from the sea into the tidal environment by flood currents and settled during slack-water. The dominance of the ebb current was explained by a significant fluvial input. The Tertiary sediment was probably supplied from the hinterland by a river, which was connected with the estuaries of the Hoogerheide and Woensdrecht Members. Reworking from the underlying deposits by tidal channels may be considered as well.
Gravel is nearly always absent in both members, although current velocity was high enough to transport very large clay-pebbles (20 cm). Only one specimen was found in a lag deposit of a tidal channel, at the base of the Woensdrecht Member in Woensdrecht Hooghuis. The well rounded pebble consists of white, quartzitic sandstone. The provenance may be found in the Ardennes or the Brabant Massive, although the Paleozoic rocks are often not so light coloured.

5.4.5 Gilze Member

The data on the sediment provenance of the Gilze Member are somewhat confusing. The member is dominated by stable heavy minerals (§2.5.8). Turbid chloritoid and green-brown Vosges hornblende, indicative of a Meuse origin (Zonneveld, 1948a, 1955) have not been found in the study area. A Meuse provenance of the sand can therefore be excluded. On the other hand the influence of the Meuse in western Noord-Brabant was detected in the gravel components in the coarse-grained upper part of the Gilze Member: the Alphen Sands of Menapian age (Vandenberghe and Krook, 1981). The Meuse influence was established by the presence of Revinian quartzite, although typical Meuse minerals were not found among the heavy minerals.
The gravel composition of the Gilze Member at the Galderse Meren is presented in fig. 5.8 (after Vandenberghe et al., 1986). The fine gravel (3-5, 5-8 mm) is dominated by flint (70-80%). In the coarser fractions (8-23, >23 mm) the quartz content is higher (20-40%) and the flint content is lower (36-59%). This is remarkable, since normally quartz content decreases and flint increases with increasing grain-size (Maarleveld, 1956). Revinian quartzite and light coloured, possibly Westphalian quartzite, which occur in the Meuse basin, were found in the coarser gravel. However, the very high flint content (fig. 5.8), which is highly characteristic of the Gilze Member, has never been found in the Meuse terraces of Zuid-Limburg (Van Straaten, 1946).

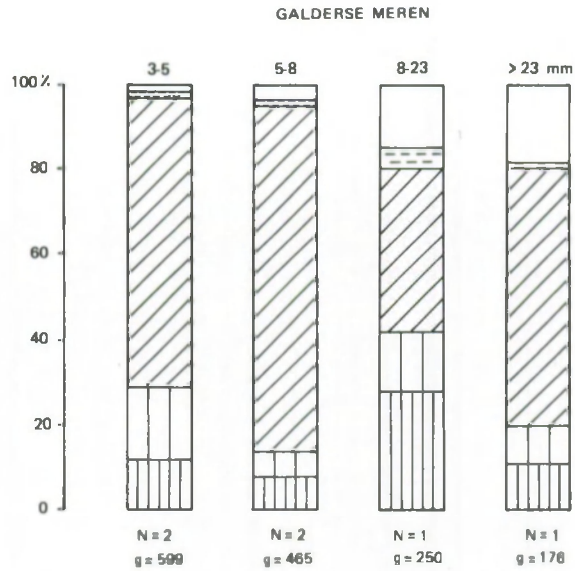


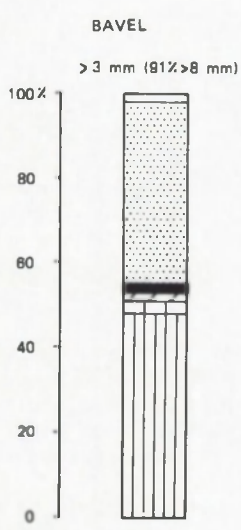
Fig. 5.8: Gravel composition of the Gilze Member at Galderse Meren (after Vandenberghe et al., 1986)(legend in fig. 5.2; N=number of samples, g=number of grains).

The gravel assemblage of the Gilze Member is explained by sediment supply from both the Scheldt and Meuse basin (Vandenberghe et al., 1986). The Scheldt yielded sand (stable heavy minerals) and fine gravel (flint and transparent quartz), which is identified as the Scheldt-association (fig. 5.8: 3-5, 5-8 mm). The Meuse contributed little to the finer fractions, but supplied relatively more coarse gravel with milky quartz and Revinian tracer components (fig. 5.8: 8-23, >23 mm). The Scheldt supply of sand and gravel in the Gilze Member exceeds the Meuse contribution where volume is concerned. Rivers from the Scheldt basin (Scheldt, Dender, Zenne, Dijle, Gete) drained and eroded the Tertiary of Central Belgium and transported sand with a stable heavy mineral association and gravel with a Scheldt-association to the north (fig. 5.15). The stable heavy mineral composition of the Gilze Member resembles the composition of the Eocene Brussel Sands, but it cannot be excluded that other weathered Tertiary formations were also eroded. The Tertiary weathering can be responsible for the large amount of gray flint in the gravel assemblage (Gulinck, 1960; Vandenberghe et al., 1986). Some thin sections revealed that this gray flint is an intermediate alteration product, which is found between the original brown to black Cretaceous flint in the centre and the white patina crust at the surface of flint particles. The gray flint was later frost shattered in the Quaternary and transported to the north to contribute to the Gilze Member (Tavernier and De Moor, 1974). Because of the large distribution of the Scheldt-association in Noord-Brabant and adjacent Belgium it is concluded that the rivers from the Scheldt basin are of greater significance in the depositional history at the southern rim of the North Sea basin than has been recognized up to now.

The northern limit of the Scheldt gravel and stable heavy mineral association in the Gilze Member (Kedichem Formation) could be fixed by the interfingering of stable and unstable heavy mineral zones, north of

Gilze and Snijders-Chaam (possibly also Wouwse Plantage)(see appendix). The unstable heavy mineral peaks are interpreted as the southernmost influence of the Rhine in the area during the Eburonian and Waalian periods. A comparable phenomenon was observed in the Central Graben (Zagwijn, 1960; Zagwijn and Van Staalduinen, 1975; Bisschops et al., 1985). North of the locations Breda, Gilze, Kampinasche Heide and Veghel, deposition by the Rhine was probably dominant. In the strato-type of the Kedichem Formation in the Central Netherlands unstable heavy minerals of Rhine provenance dominate the lower part of the formation, which dates from the Eburonian and Waalian (Zagwijn and Van Staalduinen, 1975). Furthermore, it should be realized that in a mix-ture of Rhine and Scheldt/Meuse sediments, Rhine characteristics will easily dominate due to the higher concentration of heavy minerals in Rhine sediment. On the average the content of heavy minerals in Rhine sand is about 4.5 times as high as in the sands supplied by the Meuse or Scheldt (boring Appelenberg: Turnhout Member: 0.87%; Gilze Member: 0.19% heavy minerals).
The northern extent of the Meuse and Scheldt depositional area was determined mainly by the position of the Rhine. A southern Rhine course (possibly caused by a high sea-level in interglacial periods: Tiglian C3, Tiglian C5, Waalian, Bavelian) restricted the influence of the Meuse and Scheldt to the southern Netherlands (Noord-Brabant), whereas a more northern situated Rhine course allowed a northern expansion of the Meuse and Scheldt systems. This process of extension of the south-ern depositional area dominated by the Meuse and Scheldt occurred during several periods of the Pleistocene (Tiglian: Beerse Member; Eburonian, Waalian, Menapian: Gilze Member; Bavelian: boring Leerdam, Zagwijn and De Jong, 1984; and Cromerian: boring Dordrecht, Zagwijn et al., 1971).

5.4.6 Bavel Member



The Bavel Member is characterized by garnet, epidote and hornblende, whereas the alterite content is relatively insignificant (fig. 2.20). This assemblage is comparable to the Sterksel heavy mineral zone (Zonneveld, 1947) deposited by the Rhine during the Bavel Inter-glacial (Zagwijn and De Jong, 1984; Bisschops et al., 1985). Since the sediments of the Sterksel heavy mineral zone at Bavel are fine-grained, they have been incorporated in the Kedichem Formation, instead of the Sterksel Formation.
The gravel in the Bavel Member is dominated by milky quartz (51%), Paleozoic quartzitic sand-stone (43%) and little flint (2%) (fig. 5.9). This assemblage is commonly found in both the Kedichem and Sterksel Formations (fig. 5.2) and points to a Rhine/Meuse provenance. In comparison to the Gilze Member the Scheldt has been replaced by the Rhine and Meuse. Although the Bavel Member shows an erosive contact with the Gilze Member (fig. 2.5), reworked sand and gravel of the Scheldt association was not

Fig. 5.9: Gravel composition of the Bavel Member at Bavel.
(legend in fig. 5.2).

found. The amount of sediments supplied by the Rhine far surpassed those of the Scheldt and the Meuse. The Rhine also dominated the other river systems in the sediment-petrographic record due to its elevated unstable heavy mineral content (1.5% at Bavel).

5.4.7 Sterksel Formation

The Sterksel Formation at Appelenberg is characterized by an unstable heavy mineral association, in which two heavy mineral zones can be distinguished (fig. 2.15). The lower zone, with a high alterite content, is correlated with the Woensel heavy mineral zone (Zonneveld, 1947). The upper zone, with low garnet content, resembles the Weert heavy mineral zone. Both heavy mineral sub-associations indicate a Rhine provenance.

Intercalations of stable heavy minerals (Budel heavy mineral zone), which indicate a local increase of Meuse deposition (Zonneveld, 1947), have not been found. However, the typical greenish brown Vosges hornblende (included in the hornblende group) does occur in the investigated section. This mineral points to an additional supply of heavy minerals by the Meuse (Zonneveld, 1953, 1955, 1956). Since the heavy mineral concentration is low (boring Appelenberg: 0.19%), it is possible that the major part of the total sediment was supplied by the Meuse, although the heavy mineral composition is dominated by Rhine minerals. The influence of the Meuse is also apparent from the gravel composition (fig. 5.10).

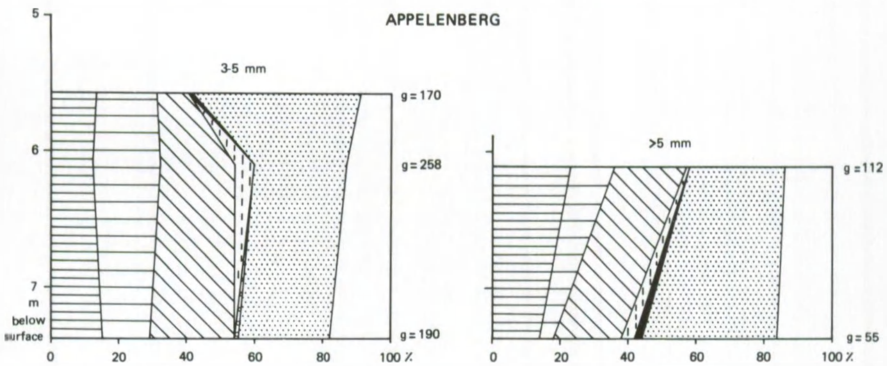


Fig. 5.10: Gravel composition of the Sterksel Formation in boring Appelenberg (legend in fig. 5.2).

The high quartz (30%) and Paleozoic sandstone content (27-50%) and the presence of Revinian quartzite probably indicate a Meuse provenance. The flint content is rather high (circ. 20%) for a pure Meuse association (Van Straaten, 1946). The flint was possibly reworked from underlying flint-rich gravel of the Gilze Member, or it was supplied by rivers from the south (Scheldt association).

The strong influence of the Meuse in the sediment probably points to a nearby confluence area of the Meuse and Rhine. According to Zonneveld (1974) the confluence of the Meuse (St. Pietersberg deposits) and the Rhine (Sterksel Formation) is located circ. 6 km south of the investigated boring Appelenberg.

5.4.8 Eindhoven and Twente Formations

Although the formations are of Middle- and Late-Pleistocene age, their sediment-petrographic composition is discussed here, since it provides important information concerning the distribution of the Scheldt gravel association.
The formations are characterized by a mixed heavy mineral association (§2.5.10). The occurrence of such an association can be explained by mixture of sediment with stable and sediment with unstable heavy minerals. The mixed assemblage in the formations is different from the sediment-petrographical composition of the underlying sediments (see appendix Achtmaal, Witte Bergen), which points to a regional sediment provenance. In comparison with the underlying Turnhout Member, Bavel Member and Sterksel Formation, the unstable heavy minerals in the Eindhoven and Twente Formations are well rounded. The roundness and mixed character of the heavy minerals was probably caused by eolian and fluvial deposition in periglacial environments (§3.3.9), after the Cromerian.
According to Schwan (1986) the eolian sediment of the Twente Formation was supplied from the emerged northern part of the North Sea basin during the Weichselian. Vandenberghe and Krook (1981) stated that the mixed heavy mineral association in the eolian sands was formed by the incorporation of heavy minerals from different formations. The southern transport direction is supported by the presence of colourless or light green augite (diopside) in the Twente Formation of the investigated area. The mineral is absent in the underlying Tegelen and Kedichem Formations and has most probably been supplied from the Kreftenheye and Urk Formations in the Central Netherlands.
In comparison to the eolian sediments, the fluvial deposits within the Eindhoven and Twente Formations contain more sediment from local underlying units (Vandenberghe and Krook, 1985). For instance, the high stable heavy mineral content at the base of the fluviatile Eindhoven Formation in boring Bolk can be explained by reworking from the underlying Gilze Member. On the other hand, eolian silt deposits (loess) (App.: Merksplas Strafinr., Beerse Blak, Beerse Dakt) have a higher content of unstable heavy minerals, which resembles the A-association of northern (Scandinavian?) origin in the northern North Sea basin (Baak, 1936).
Locally, at the top of the Twente Formation the sediment provenance of the eolian deposits changes from regional to local (app.: Ossendrecht). Just below and above the Allerød peat-layer in Ossendrecht the heavy mineral composition is still characterized by a mixed assemblage. The high garnet content (max. 21%) points to a regional supply, since the underlying Hoogerheide and Woensdrecht Members contain a low (< 10%) garnet content. Above the peat-layer unstable heavy mineral content decreases rapidly (garnet: 1%; alterite: 0%) and the association is almost similar to the underlying Woensdrecht Member (alterite: 1% ; garnet: 3%). The sediment was probably derived from outcrops of the Woensdrecht Member along the Scheldt escarpment, some kilometers west of this exposure.

The gravel assemblage of the fluvial Eindhoven Formation (Achtmaal, Wernhout Maalbergen)(fig. 5.11) and the Beuningen gravel-bed of the Twente Formation (fig. 5.12) is characterized by flint (circ. 50%), quartz (circ. 40%: especially transparent) and a small rest group (mainly quartzite and silification). Silicic oölites often occur (circ. 1%). The gravel composition of the Eindhoven and Twente Formations is comparable to the Scheldt association of the Beerse and Gilze Members.

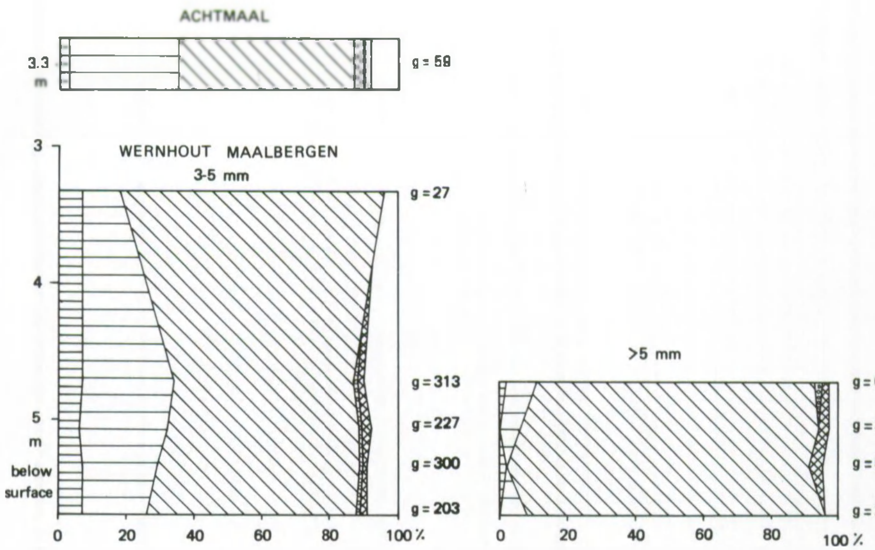


Fig. 5.11: Gravel composition of the Eindhoven Formation in borings Wernhout Maalbergen and Achtmaal (legend in fig. 5.2).

Since the formations were formed in a periglacial environment outside the range of the major rivers, it is likely that the gravel was reworked from the Gilze Member. The Scheldt gravel has originally been derived from the Tertiary deposits in Belgium (fig. 5.15).
In the western part of the area the flint content in the Twente Formation is lower, whereas the quartz and silification content is higher than in the east (fig. 5.12, 5.13: Ossendrecht). The silifications especially consist of calcedonic sandstone, occasionally with red dots (hematite?). The gravel composition at Ossendrecht is clearly part of the Scheldt gravel association, but the small differences in the SC-gravel association in the western (Ossendrecht) and the eastern (e.g. Meerle) part of the investigated area might reflect different source areas in Belgium.
The type locality of the Scheldt gravel association at Bavel (Zandstra, 1969) has been reinvestigated in a nearby pit (fig. 5.12: Bavel). The fine gravel fraction (3-5 mm) is characterized by a high flint and quartz content (80%) and is comparable to the Scheldt-association as described by Zandstra (1969). However, in the coarser fractions the lower flint and the higher Paleozoic sandstone content resembles more the underlying Bavel Member (fig. 5.9). Therefore, the gravel composition at Bavel is interpreted as a mixture of Scheldt and Rhine/Meuse gravel, reworked respectively from the nearby Gilze Member and from the Bavel Member. Further to the east (fig. 5.12: Appelenberg) the Rhine/Meuse gravel components, reworked from the Sterksel Formation (fig. 5.10) become dominant in the Twente Formation, although Scheldt gravel can still be detected by the high flint content.

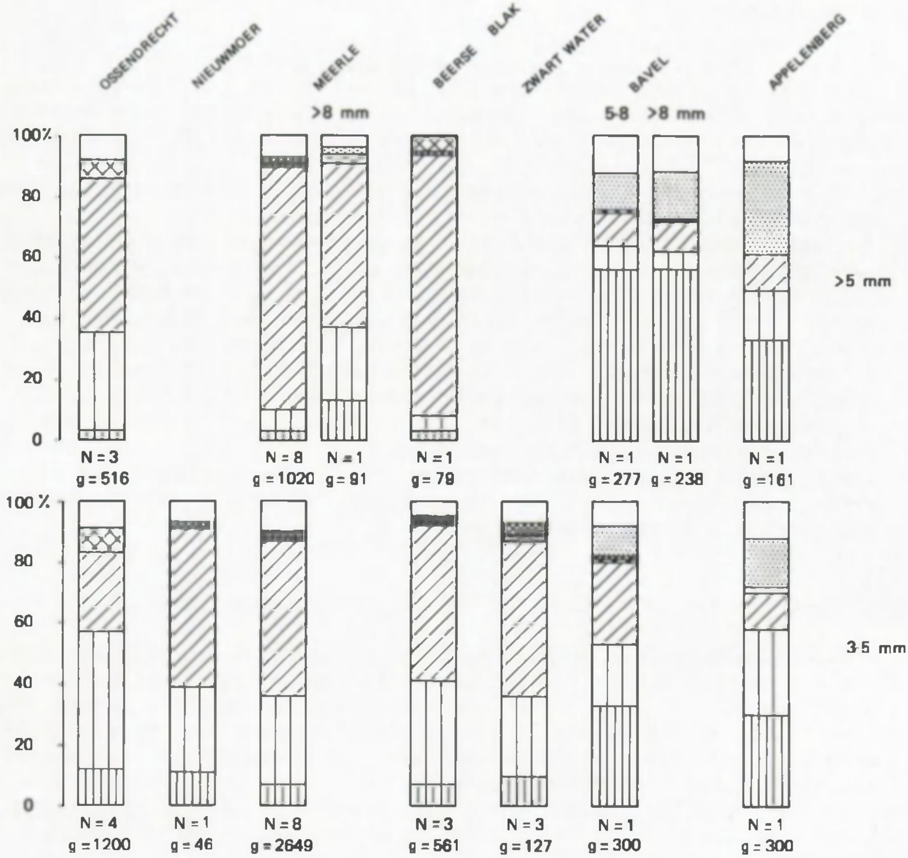


Fig. 5.12: Gravel composition of the Twente Formation (Beuningen gravel bed), arranged in an east-west section throughout the area.

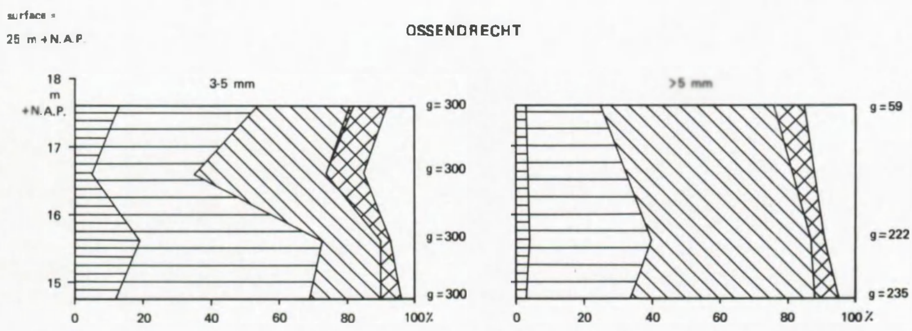


Fig. 5.13: Gravel composition of the Twente Formation in Ossendrecht. (legend in fig. 5.2).

5.5 Conclusions

1. The results of the gravel analysis presented in the previous paragraphs have been summarized in fig. 5.14. Unpublished data from the study area have been added to create more coherent groups. The diagrams reveal that:

- The Sterksel Formation (group 1) contains a high quartz and rest group content, indicative of a Rhine/Meuse provenance.
- The quartz and flint content of the Eindhoven and Twente Formation overlying the Sterksel Formation (group 2) is higher with respect to group 1. Reworking of Rhine/Meuse material is evident, although gravel with a Scheldt association has been supplied additionally.
- The Gilze Member (group 4) is dominated by flint and quartz in the 3-5 mm gravel fraction, which is characteristic for a Scheldt provenance. The Gilze Member (group 4) overlaps part of groups 1, 2 and 3 in the fraction larger than 5 mm, which indicates a combined Scheldt and Meuse provenance of the coarser gravel.
- The very high flint and quartz content in the Eindhoven and Twente Formations (group 3) points to reworking of gravel with a Scheldt association from the Gilze Member.

The heavy mineral composition of the sediments reveals that:

2. The Rhine supplied garnet, epidote, alterite, hornblende and glaucophane from the Alpine region into the fluviatile Bavel Member and Sterksel Formation and to the northern part of the Gilze Member. The sediments of the Rijkevorsel and Turnhout Members and partly of the Woensdrecht and Hoogerheide Members are of a Rhine provenance as well, but redistributed by tidal currents.
3. The rivers of the Scheldt basin supplied zircon, rutile, tourmaline, staurolite, kyanite and andalusite from the Belgian Tertiary deposits to the Beerse and Gilze Members and to a lesser extent to the Hoogerheide and Woensdrecht Members.
4. The Meuse supplied little sand to the Gilze Member, but some gravel was apparently transported from the Ardennes region, which can be deduced from the presence of Revinian quartzite. The Meuse contributed green-brown Vosges hornblende and a major part of the sand and gravel to the Sterksel Formation, although the heavy mineral assemblage is dominated by Rhine minerals.

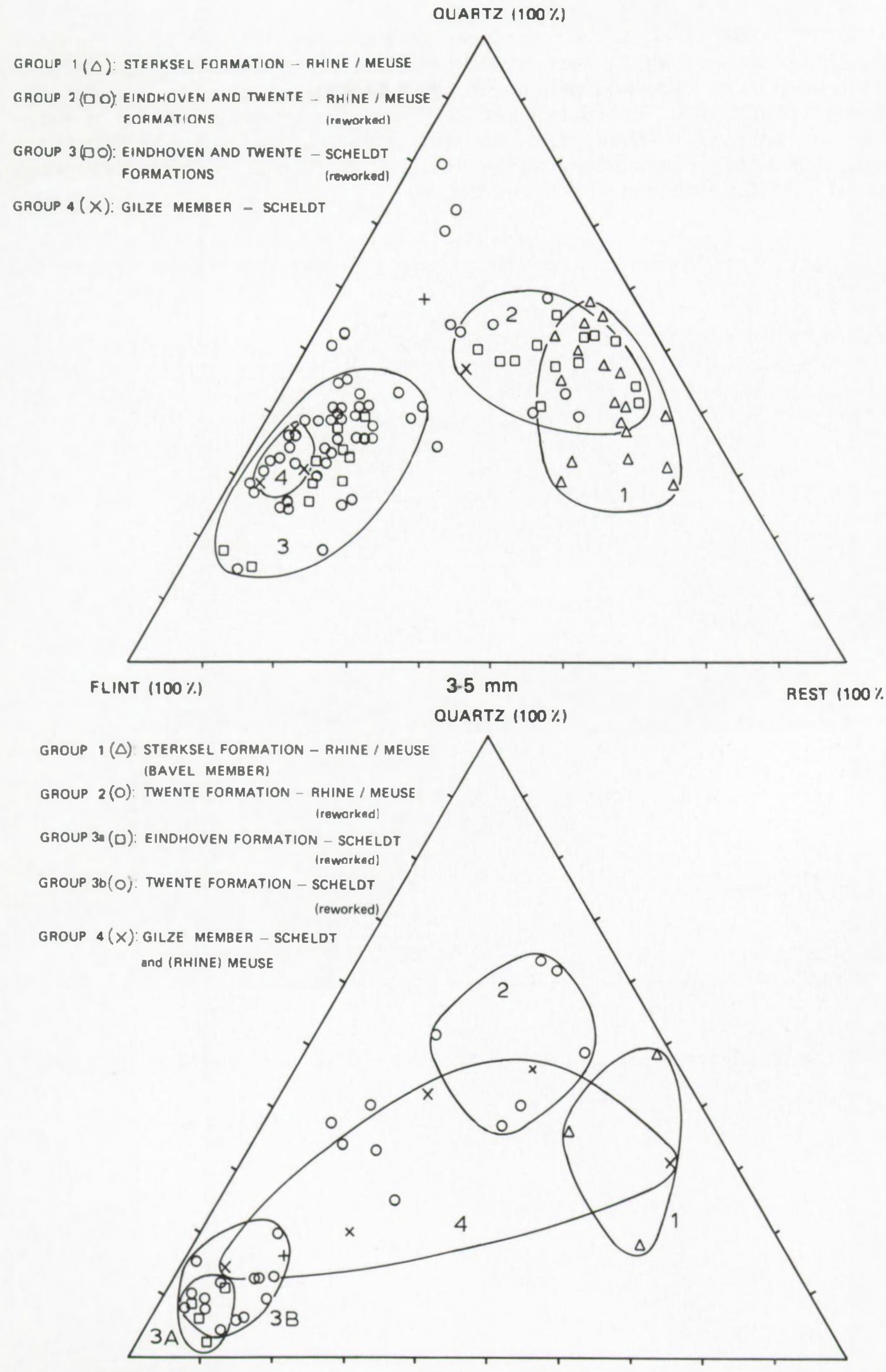


Fig. 5.14: Gravel assemblages of different lithostratigraphic units in Noord-Brabant and northern Belgium.

5. The Scheldt association in Noord-Brabant consists of flint, quartz and silifications, which were eroded from Tertiary deposits in Belgium and transported to Noord-Brabant at least since the Tiglian (Beerse Member) (fig. 5.15). Noord-Brabant is the northern extension of a large area in Belgium, where the Scheldt gravel association is dominant. During the Early-Pleistocene Noord-Brabant occurred in the confluence area of the Scheldt and the Rhine/Meuse.

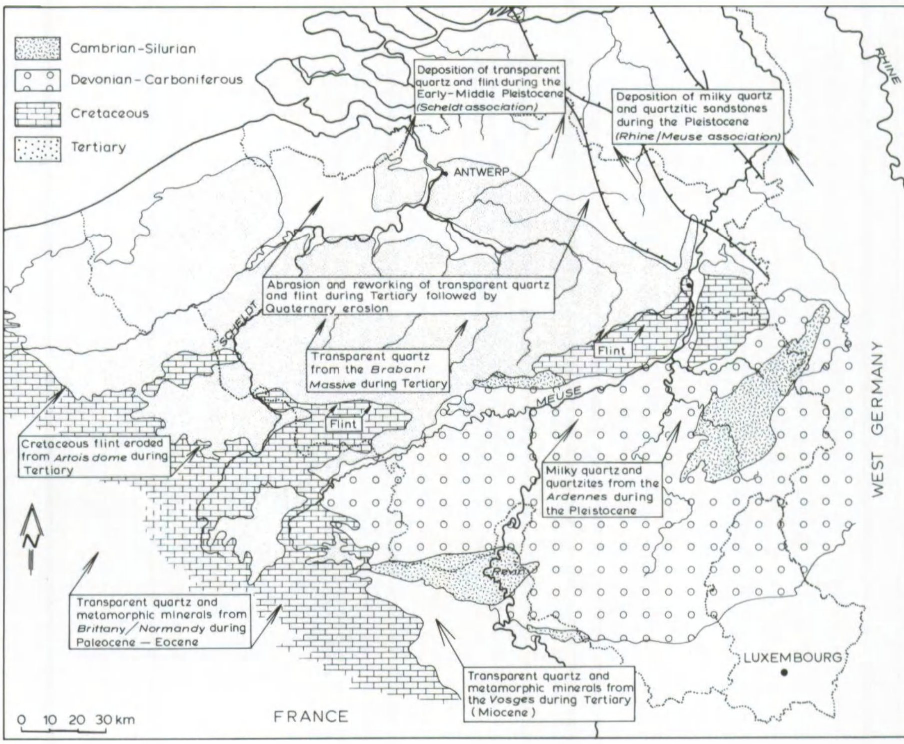


Fig. 5.15: Provenance of gravel in Noord-Brabant and northern Belgium.

6. PALEOGEOGRAPHIC EVOLUTION OF NOORD-BRABANT AND NORTHERN BELGIUM

6.1 Introduction

This chapter presents the paleogeographic history of Noord-Brabant and northern Belgium during the Early-Pleistocene. The Late-Tertiary and the Middle- and Late-Pleistocene evolution are discussed only briefly. Up to now the paleogeographical development of the area was described locally (De Ploey, 1961; Geys, 1975). Large-scale reconstructions were made for the complete Netherlands (Zagwijn, 1975a, 1979), while more detailed studies were performed southeast of the study area (Zonneveld, 1947, 1948a, 1955, 1974).
The Early-Pleistocene geomorphological evolution presented below is based on the results from the investigated area, presented in the previous chapters. Published information from neighbouring areas (Central Netherlands, Central Graben, Zuid-Limburg, Central Belgium) has been used as well. An attempt is made to give a picture for each glacial and interglacial Early-Pleistocene phase. Former reconstructions were mainly made for interglacials (Zagwijn, 1975a, 1979).
It is stressed, that the boundaries of the depositional environments are often hypothetical, since the number of observations is relatively small. Furthermore, it should be remembered, that the paleogeographical development is based primarily on lithostratigraphic units, which can to a certain extent be diachronous. Finally, the paleoenvironments need not evolve gradually into each other, as important changes and different paleogeographic situations could have occurred between two reconstructions.

6.2 **Pliocene-Pleistocene transition** (Praetiglian, Tiglian A, B): Fluviatile sedimentation followed by non-deposition (hiatus)

During the Late-Pliocene and Praetiglian the Meuse in Belgium (trainée mosane) (Pissart, 1974) continued its northeastern course downstream of Liège through Zuid-Limburg, where the Kieseloölite Formation was deposited on top of the Ubachsberg (Zonneveld, 1974) and in the Kosberg terrace (Zagwijn and Van Staalduinen, 1975) (fig. 6.1). The Meuse discharged into the Rhine, which followed the Roer Graben east of Heerlen-Sittard (Zuid-Limburg). From this point the Rhine (and Meuse) flowed in a westerly direction through northern Belgium (Paepe and Vanhoorne, 1976), where they deposited thick beds of the so-called Mol Sands and Merksplas Sands (fig. 6.1). Because of the southwest-northeast paleocoastline configuration (Zagwijn, 1975a), it is possible that the Merksplas Sands are the estuarine or nearshore equivalent of part of the fluviatile Mol Sands. To the north the Mol and Merksplas Sands change into marine, offshore (glauconite and) shell-bearing deposits of the Oosterhout and Maassluis Formations (pers. comm. Dr. P. Laga).
After the Late-Pliocene or Praetiglian the Rhine/Meuse sedimentation stopped in western Noord-Brabant and adjacent northern Belgium. This situation of non-deposition seems to have persisted until the Tiglian C3, at least in the southern part of the area (Turnhout, Ravels), where the Rijkevorsel Member of Tiglian C3 age overlies the Tertiary or Praetiglian Merksplas Sands (Vanhoorne, 1962). The sedimentary hiatus is probably smaller in the east (Central Graben) and north, where marine and fluviatile sediments of the Maassluis and Tegelen Formations were deposited during the Tiglian A, B and C periods (Zagwijn, 1963a:

borings Eindhoven; Bisschops et al., 1985).

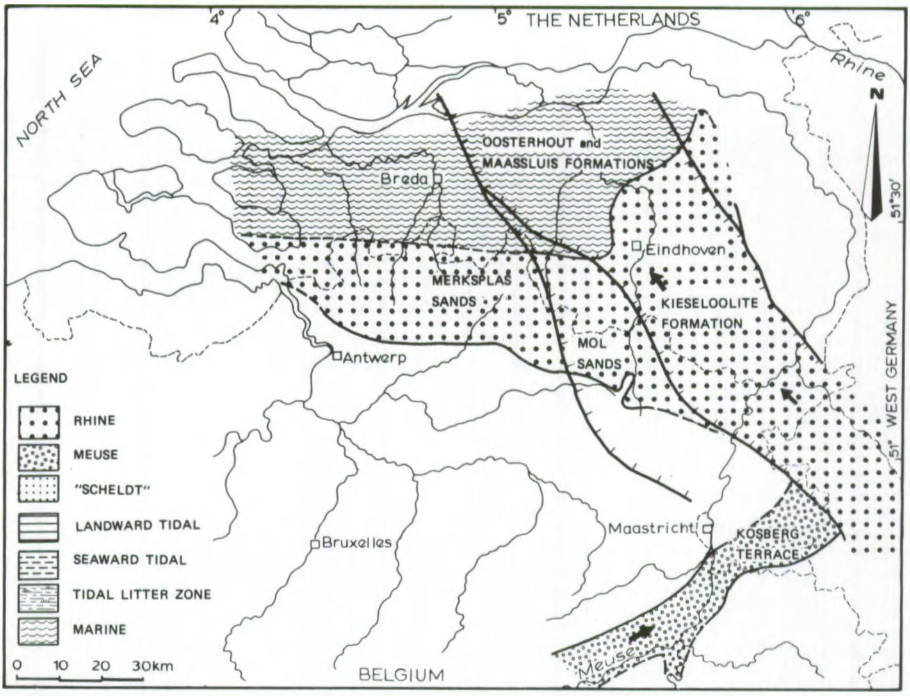


Fig. 6.1: Paleogeography of the Late-Tertiary and Praetiglian.

6.3 Tiglian: Noord-Brabant in the transition zone of marine and continental deposition

6.3.1 Tiglian C3: Marine transgression and deposition of distal (Rijkevorsel Member) and proximal (Hoogerheide Member), inshore, tidal sediments in a warm temperate climate.

After the cool Tiglian B, the climate ameliorated (Tiglian C)(Zagwijn, 1963a). The vegetation in Noord-Brabant and northern Belgium was characterized by a fairly high content of so-called warm temperate species. A rising sea-level resulted in a transgression in the investigated area (Rijkevorsel Member, Hoogerheide Member). The presence of the waterfern *Azolla tegeliensis* and the absence of *Fagus* pollen points to a Tiglian C age (Zagwijn, 1963a).

At the beginning of the Tiglian C3 transgression the Merksplas Member in northern Belgium was eroded by estuarine channels (fig. 6.2: stage 1). In the course of the transgression the estuarine zone shifted further landward and the former estuaries drowned. Outside the estuaries the flooding of the Tertiary plain is reflected by the deposition of a (lagoonal?) subtidal clay (lower clay of the Rijkevorsel Member) (fig. 6.2: stage 1 and 2). The preservation of this clay-facies might be explained by the deposition on the steeply dipping Tertiary surface. The Merksplas Sands show a dip nowadays of approximately 2.1 °/oo, while the top of the overlying Rijkevorsel Member reveals a value of

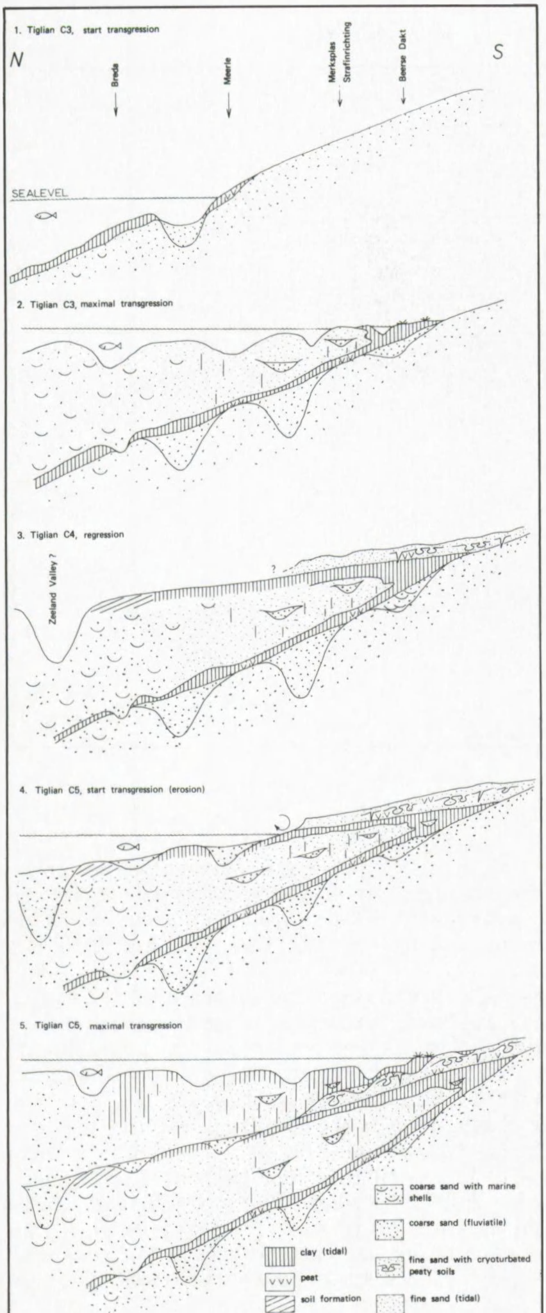


Fig. 6.2: Geological evolution of Noord-Brabant and northern Belgium during the Tiglian.

1.4 °/oo. Therefore, the surface dip amounted to 0.7 °/oo at the start of the Tiglian C3 transgression. Apparently this dip was caused by tectonic subsidence of the North Sea basin after the Late-Pliocene/Praetiglian and before the Tiglian C3. The transgression over the 0.7 °/oo dipping surface can have led to a rapid deepening of the tidal environments or to rapid sedimentation, by which the previous deposited clay-layer became protected against later tidal channel erosion.

During the transgression maximum large parts of the study area were affected by tidal processes (fig. 6.3). The tidal sediments are found north of the locations Putte, Westmalle, Turnhout, where they wedge out over the Merksplas Sands. The Rhine occupied a north-western course through the Central Graben area during this period (fig. 6.3) (Zagwijn, 1974, 1979). In the neighbourhood of Eindhoven the fluviatile environments of the Rhine probably merged into tidal environments (Bisschops et al., 1985), in which the Rhine sediments were redistributed. In the eastern part of the study area the Rhine derived sediments (Rijkevorsel Member) were deposited in an inshore, distal (landward) tidal environment, characterized by a fairly low salinity (fig. 6.3). The tidal range was estimated locally between 0.95 and 2.1 m. At the southern margins of the distal, tidal environments, in the neighbour-

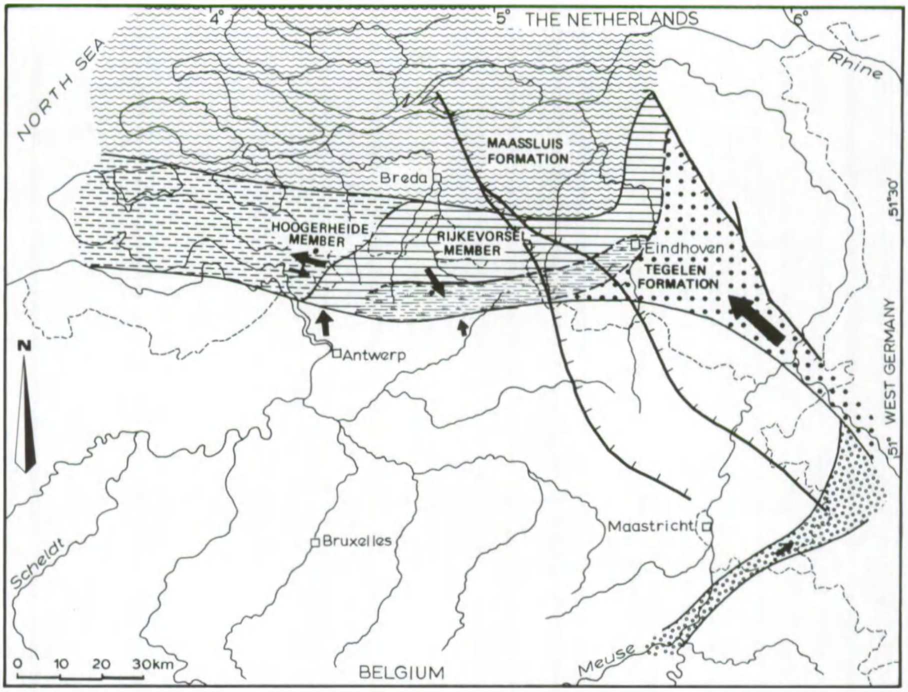


Fig. 6.3: Paleogeography of the Tiglian C3 period.

hood of Turnhout and Westmalle, tidal litter zones developed, characterized by a Chenopodiaceae vegetation. Sediment was transported predominantly from the north, but a minor amount of sediment was reworked from the Tertiary subsoil or supplied by small rivers systems from the south.

More to the west the simultaneously deposited Hoogerheide Member was formed in a proximal (seaward), inshore, tidal environment (fig. 6.3). Wave processes were more important here as reflected by the presence of hummocky cross-bedding. During storms sediment was brought into suspension above the sandflats and settled quickly afterwards on the sandflat or on channel slopes forming climbing ripple bedding. Salinity could have been somewhat higher here in the west. The paleocurrents were dominated by northwestern (ebb) and southeastern (flood) directions. With respect to the eastern part of Noord-Brabant, Rhine sediments were less important in the west during the Tiglian C3 transgression. The sediments in the proximal, tidal environments of the Hoogerheide Member were reworked from the Tertiary subsoil or were supplied by river systems from the southeast.

The inshore, non-calcareous, sparsely bioturbated, tidal sediments change in the neighbourhood of Breda into shell-bearing, littoral or offshore deposits of the Maassluis Formation (fig. 6.3) (Zagwijn and Van Staalduinen, 1975). This transition probably reflected a northward increase in salinity of the environment. However, post-depositional decalcification of the later uplifted sediments in the south can be responsible as well for the non-calcareous character of the tidal

deposits.
At the end of the Tiglian C3, silting of the tidal environments and a cooling of the climate resulted in a regression of the sea, which culminated during the Tiglian C4.

6.3.2 **Tiglian C4:** regression of the sea and deposition of eolian and fluviatile sediments (Beerse Member) in a cold climate

At the beginning of the Tiglian C4 the previous deposition in tidal environments was succeeded by continental sedimentation in a cold climate (Beerse Member) (fig. 6.4). Unfortunately, the transition of the Tiglian C3 with interglacial conditions to the Tiglian C4 with glacial characteristics was not preserved in the pollen record. It is assumed that the regression was an eustatic one, caused by a cooling of the climate. The tidal deposition of Rhine sediment in the study area stopped and the Rhine probably followed a northwesterly course through the Central Graben (fig. 6.4) (Bisschops et al., 1985). The Meuse was still occupying its northeastern course through Zuid-Limburg (Zagwijn and Van Staalduinen, 1975).

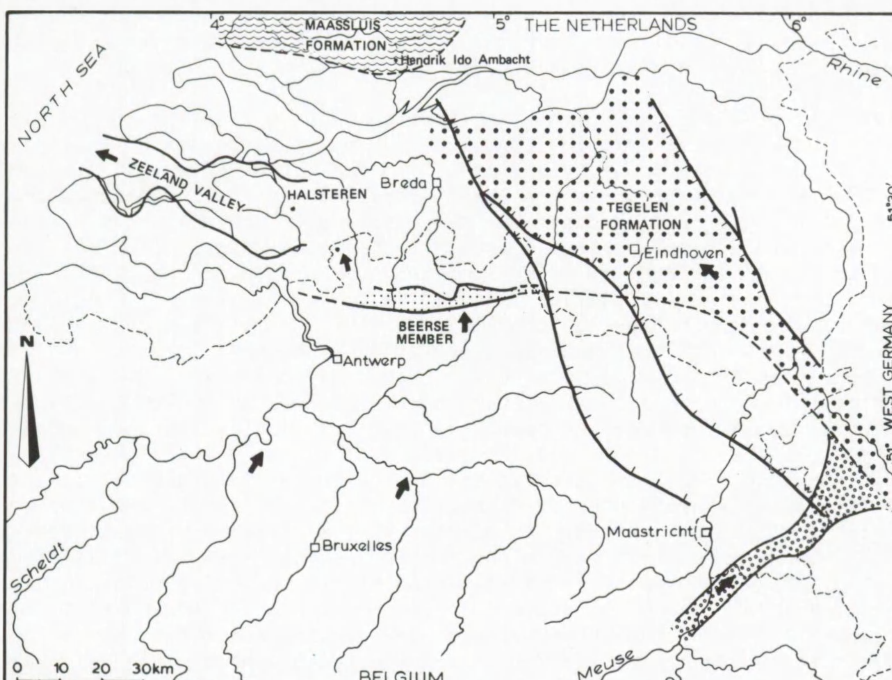


Fig. 6.4: Paleogeography of the Tiglian C4 period.

During the Tiglian C4 regression the Beerse Member was formed in northern Belgium. The sediments were probably supplied by rivers from the Scheldt basin (Central Belgium), where outcropping Tertiary formations are characterized by stable heavy minerals and flint and quartz gravel (Gullentops, 1963; Tavernier and De Moor, 1974). Deposition occurred in eolian and fluviatile, periglacial environments (fig. 6.2: stage 3). Dry eolian and wet eolian sands were deposited in sandsheets, which

resemble the Weichselian coversands (Schwan, 1986). Intervening periods of non-deposition are reflected by soils and peat formation. The small ice-wedge casts, frost cracks, cryoturbated soil horizons and the cold pollen spectra point to a glacial phase with (local) permafrost and a mean annual temperature around -5° C. The associated tundra or taiga vegetation was strongly dominated by herbs, pine and birch.
Besides eolian deposition in some areas, the lowering of the sea-level probably also led to deep erosion in other regions. The so-called Zeeland Valley, later filled by the Halsteren Beds (Van Voorthuysen, 1957), might have been formed during this period with a low sea-level stand.
The regression of the sea was registered at the southern rim of the North Sea basin by continental, periglacial sedimention (Beerse Member). More to the centre of the basin, however, marine deposition probably continued (fig. 6.4). The transition from the warm temperate Tiglian C3 into the cold Tiglian C4 might be reflected there by a change in the marine mollusc assemblage (pers. comm. T. Meijer: borings Brielle, Hendrik Ido Ambacht, Zuurland). Apparently the regression was less pronounced than during the Middle- and Late-Pleistocene glacials.

6.3.3 **Tiglian C5**: Transgression of the sea and deposition of inshore tidal, distal and proximal, fresh to brackish water sediments (Turnhout and Woensdrecht Members)

After the Tiglian C4 climate ameliorated during the Tiglian C5 with interglacial characteristics (Zagwijn, 1963a). The courses of the Rhine and the Meuse were probably comparable to the situation in the previous Tiglian phases (fig. 6.5). The Rhine occupied its northwestern course through the Central Graben. The Meuse flowed to the northeast through Zuid-Limburg and formed the Simpelveld terrace (Zagwijn, 1986). The rivers from northern Germany possibly prograded in the central and eastern Netherlands (Zagwijn, 1974, 1975a).
The climatic improvement at the beginning of the Tiglian C5 caused a transgression of the sea in Noord-Brabant, northern Belgium and the southwestern Netherlands (fig. 6.2: stages 4 and 5; fig. 6.5). In contrast to previous ideas (Paepe and Vanhoorne, 1970) the associated Turnhout and Woensdrecht Members date from this Tiglian period, because of the presence of *Azolla tegeliensis*.
The transition from the glacial Tiglian C4 to the interglacial Tiglian C5 is not registered in Noord-Brabant since in the first phase of the Tiglian C5 transgression, estuarine systems (Meerle Slikgat, Beerse Blak) almost completely eroded the Beerse Member of the Tiglian C4 glacial phase and part of the interglacial type deposits of the Rijkevorsel Member (fig. 6.2: stage 4). Paleocurrents at Beerse Blak were dominant to the northnorthwest (ebb). The erosion was more intense than during the start of the Tiglian C3 transgression, which might be related to the more gentle slope of the underlying deposits over which the sea transgreded (see §6.3.1). Nowadays the Tertiary deposits reveal a dip of approximately 2.1°/oo. The top of the Rijkevorsel (Tiglian C3) and the Turnhout Member (Tiglian C5) show a dip of respectively 1.4°/oo and 1.1°/oo (fig. 2.5). When the sea transgreded in the Tiglian C5, over the shortly before deposited Beerse Member, the surface dip of 0.3°/oo was less than in the Tiglian C3 period (0.7°/oo). Therefore, it might be possible that no rapid deepening of the environments occurred and the tidal estuarine channels were able to migrate and remove the underlying Tiglian C4 periglacial deposits. Only in the southernmost part of the inundated region have Tiglian C4 sediments (Beerse Member)

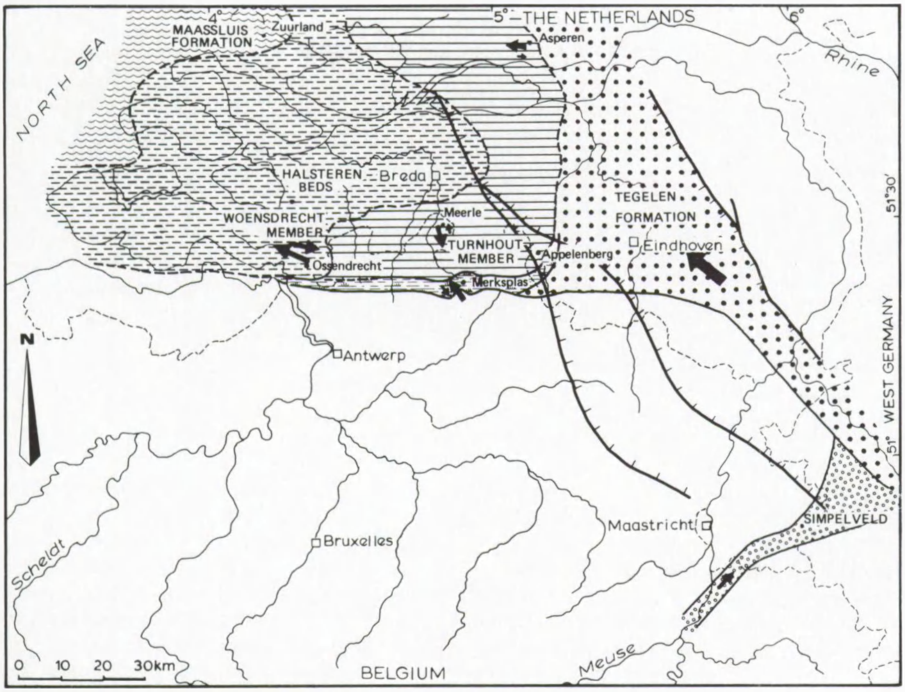


Fig. 6.5: Paleogeography of the Tiglian C5 period.

been preserved (fig. 6.6). More basinwards, in the western Netherlands, marine deposits with a cold mollusc assemblage (Tiglian C4?) have been protected from later erosion as well. During the Tiglian C5 transgression marine sedimentation probably continued and shell-bearing deposits with a warm temperate mollusc assemblage (possibly Tiglian C5) were stacked upon marine deposits of possibly Tiglian C4 and Tiglian C3 age (pers. comm. T. Meijer: boring Brielle, Zuurland) (fig. 6.6). The phenomena described above are represented schematically in fig. 6.6. Towards the centre of the sedimentary basin in the north, marine sediments were formed during glacials and interglacials of the Tiglian. In Noord-Brabant only tidal sediments of interglacial periods have been preserved, since glacial deposits were eroded during the subsequent interglacial transgression. At the southern rim of the sedimentary basin glacial and interglacial conditions are reflected by the alternation of respectively tidal and fluvial/eolian deposits.
After the erosional phase at the beginning of the Tiglian C5 the study area was completely covered by tidal deposits (fig. 6.2: stage 5). In the eastern and southern part of the area rather clayey sediments were laid down (Turnhout Member) in a distal (landward) micro- to mesotidal environment (fig. 6.5). More or less stationary tidal channels occurred adjacent to the clayey interchannel areas. The paleocurrent was dominated by south oriented flood currents (exposure Meerle). The salinity was low in these landward tidal environments (limited bioturbation), but periodic influxes of salt or brackish water locally occurred in a reducing environment (pyrite formation, exposure Ravels). The tidal range was estimated at Meerle at 1.05 m (not corrected for later com

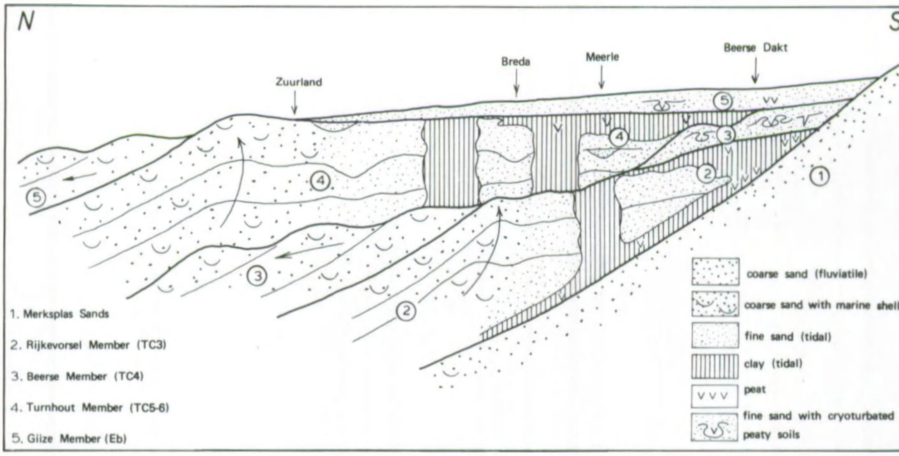


Fig. 6.6: Large-scale depositional model for Noord-Brabant and northern Belgium during the Tiglian.

paction), which is approximately equivalent to 2 m if a 50% compaction rate is assumed for fresh to brackish water tidal deposits (Zonneveld, 1960). To the south the tidal deposits of the Turnhout Member wedge out over the deposits of the Beerse Member and Rijkevorsel Member (fig. 6.6). In this east-west, land-sea border zone (Ravels, Merksplas and Beerse) a so-called tidal litter zone developed. Chenopodiaceae stands flourished at the high water level due to the accumulation and oxidation of organic material (fig. 6.5).

The distal tidal environments merge to the east in fluviatile environments of the lower reaches of the Rhine in the Central Graben (Zagwijn, 1974)(fig. 6.5). Tidal fresh water molluscs were found as far east as Hendrik Ido Ambacht and Asperen (pers. comm. T. Meijer).

In the western part of the investigated area a more sandy facies was formed (Woensdrecht Member) in a proximal (seaward) microtidal environment (fig. 6.5). Tidal range amounted to more than 1.2 to 1.7 m in Woensdrecht and Ossendrecht. The salinity was low in the Woensdrecht Member, but the local presence of bioturbation and *Ammonia beccarii* indicate that the environment was somewhat more brackish than in the east. The migrating channels removed the fine-grained material and only relatively coarse-grained channel sediments were preserved. Current directions were predominantly to the westnorthwest (ebb) and eastsoutheast (flood). During neap-spring-neap tidal cycles megaripples were formed and faded out again. Wave action over exposed intertidal sandflats removed mud and silt and levelled the sandflat surface.

The transition from the distal and proximal, tidal, inshore deposits (Tegelen Formation) into marine, shell-bearing deposits (Maassluis Formation) was, unlike during the Tiglian C3, not present in Noord-Brabant (fig. 6.5). Marine, shell-bearing sediments of possibly Tiglian C5 age have recently been found northwest of Noord-Brabant (pers. comm. T. Meijer; boring Zuurland).

The relation between the lithostratigraphic units and the period in which they were formed is illustrated schematically for the Noord-Brabant region in fig. 6.7. In the south the lithologic units belong to distinct periods and the lithostratigraphic boundaries coincide therefore, with chronostratigraphic boundaries. To the north marine deposits

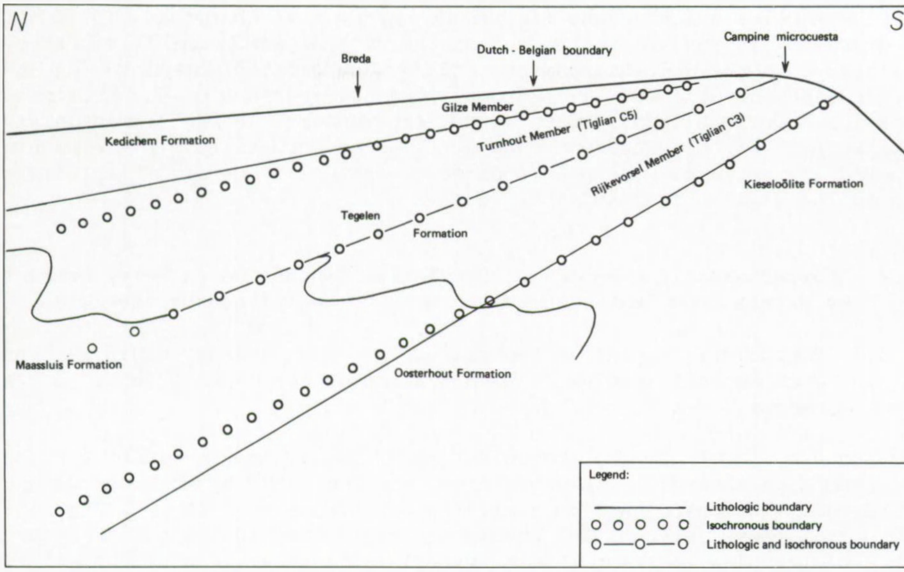


Fig. 6.7: Lithostratigraphic units and isochrons in the Late-Tertiary and Early-Pleistocene deposits of Noord-Brabant (modified after Van Montfrans, 1975, in: Zagwijn and Van Staalduinen, 1975).

were stacked during different periods and lithostratigraphic and chronostratigraphic boundaries do not coincide. Van Voorthuysen (1957) explained the presence of thick non-marine deposits in Noord-Brabant and Zeeland (see fig. 6.7: Tegelen Formation) by deep erosion (Zeeland Valley) followed by fluviatile deposition (Halsteren Beds). It is possible however, that the great thickness of the Halsteren Beds is due to a lateral facies change of marine shell-bearing deposits (Maassluis Formation) into the tidal, fresh to brackish water sediments (Tegelen Formation).

Towards the end of the Tiglian C5 the sea-level rise decreased and the depositional environments silted up completely and the area was covered by a continuous clay-layer. Brackish water environments were succeeded by tidal, fresh water, eutrophic environments, fed by large rivers (§3.4: Meerle). Peat formation commenced during a declined clastic input. In the western part of Noord-Brabant the silting of the proximal tidal environment was accompanied by a sudden drop in energy or tidal range, which caused an abrupt commencement of clay sedimentation. The abandonned channels of the proximal tidal flat environment were filled by clay in almost stagnant water. Fresh water, eutrophic peat, characterized by *Alnus* pollen, developed locally (§3.3.6: Korteven). It is unknown whether the regression of the sea, reflected by the peat-layer dominated by *Alnus* pollen, occurred at the same time over the complete area. Due to the fact that only one peat-layer is found with a comparable pollen assemblage in different locations, the peat-layer is supposed to be essentially isochronous.

The sea inundated Noord-Brabant once more after the regression period and a moderately thin clay-layer was deposited over the peat-layer. This transgression marks the last phase of tidal sedimentation during the Tiglian C5, before the sea finally withdrew from Noord-Brabant and northern Belgium during the Tiglian C6 and early Eburonian. The final regression of the sea at the end of the Tiglian was probably caused by silting of the tidal environments and by a climatic deterioration and concurrent eustatic sea-level drop, since the approach of the Eburonian glacial (Meerle Slikgat, Zwart Water, Appelenberg) is registered in the upper part of the Turnhout Member (Tegelen Formation). The temperate forest was replaced by a pine dominated vegetation, generally attributed to the Tiglian C6 (Zagwijn, 1963a).

6.4 Eburonian-Waalian-Menapian: Fluviatile deposition in Noord-Brabant by rivers from Central Belgium (Scheldt basin) and by the Meuse

6.4.1 **Eburonian**: fluviatile deposition by low energy river systems from Central Belgium in a cold climate (Appelenberg Sands: Gilze Member)

During the onset of the Eburonian the climate became colder (pollen diagram Appelenberg). The vegetation was dominated by pine, herbs and *Ericaceae*, indicating a mean summer temperature around 10° C (Zagwijn, 1963a). Milder interstadial phases are registered in the pollen record by a higher pine content (Zwart Water). Permafrost conditions with a mean annual temperature below -5° C existed during specific phases within the Eburonian and ice-wedges were formed in the top of the Tegelen Formation (Turnhout Member) at Merksplas Strafinrichting (probably) and in pit Laumans (Tegelen, province of Limburg).
The sediment transport directions changed in accordance with the climatic changes and the withdrawal of the sea (fig. 6.8). In the Tiglian C5 the Rhine sediments were redistributed by tidal currents over western Noord-Brabant and northern Belgium. During the Eburonian (Waalian, Menapian) the Rhine had left its southeast-northwest course in the Central Graben area and flowed more or less east-west through the Central Netherlands and Noord-Brabant, north of Veghel (Zagwijn, 1960; Bisschops et al., 1985) and Snijders-Chaam (fig. 6.8).
In the southern part of Noord-Brabant and northern Belgium sediments with a stable heavy mineral association were transported from the south by rivers from Central Belgium (so-called rivers from the Scheldt basin: Scheldt, Dender, Zenne, Dijle, Gete)(fig. 6.8). The sediment supply already started at the end of the Tiglian period, as documented by an increase of stable heavy minerals in the top of the Turnhout Member. These rivers from Belgium transported fine sand, without gravel to the north. Outside the river channels fine sand, loam-, clay- and peat-beds were formed in wet, floodplain environments (Gilze Member: Appelenberg Sands). Episodic inundation of the floodplain formed thin lobes of overbank sediments, which alternate abruptly with the backswamp clay- and peat-beds. Peat and gyttja-like sediments locally developed in wet places and small lakes on the river floodplain, when the clastic input of sediment was low.
In the confluence area of the Rhine and the rivers from the south, Rhine (unstable heavy minerals) and "Scheldt" (stable) sediments alternate (Snijders-Chaam)(abnormal variations: Edelman, 1933). The presence of the Meuse could not be detected in Noord-Brabant for the Eburonian. According to Zonneveld (1974) it is uncertain whether the Meuse still occupied its northeastern course through Zuid-Limburg. A new more northern branch had possibly developed. A faint Meuse influence was recorded at Herkenbosch, since Vosges hornblende was found in deposits older than the Waalian period (Zagwijn, 1960). The Margraten terrace

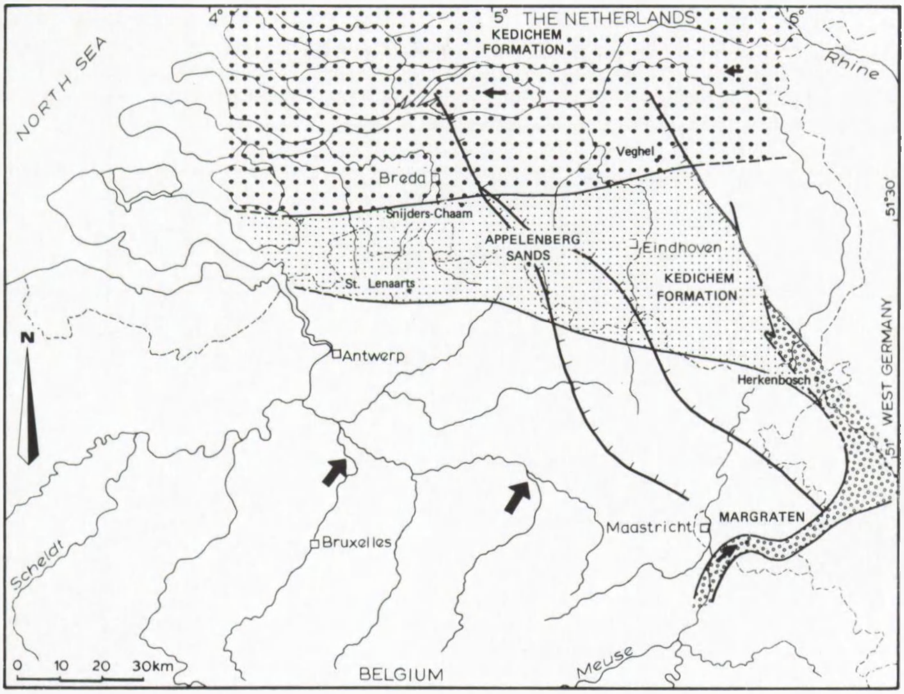


Fig. 6.8: Paleogeography of the Eburonian period.

(Zonneveld, 1948a), which is younger than the Simpelveld terrace of late-Tiglian age, was possibly formed during the Eburonian; the dating is uncertain however (fig. 6.8).

6.4.2 **Waalian**: localized fluvial deposition by meandering river systems (Gilze Clay: Gilze Member) in a warm temperate climate

The amelioration of the climate at the beginning of the Waalian is recorded by the reappearance of thermophilous trees (*Alnus, Quercus*) and the continuous occurrence of an Early-Pleistocene species like *Tsuga* (§4.3.1: Gilze). In contrast to the Eburonian, sedimentation in Noord-Brabant was restricted to local areas (Nieuwmoer, Weelde, Gilze). This difference was probably caused by a more fixed river pattern.
As in the Eburonian, the Rhine still occupied its east-west course north of Veghel (Zagwijn, 1960, 1963b; Bisschops et al., 1985) and Gilze (fig. 6.9). The Meuse had probably left its northeastern course through Zuid-Limburg and flowed west of the Ubachsberg in a northern direction. The Sibbe terrace level could have been formed in this period (Zonneveld, 1974).
Noord-Brabant south of Veghel-Gilze was dominated by rivers draining the Tertiary deposits in Central Belgium, from which they transported sediments with a stable heavy mineral association (fig. 6.9). The river flow was concentrated in specific areas. In the 5 to 10 m deep meandering river channels large-scale fining-upward sequences were formed (Nieuwmoer, Gilze). Thick backswamp deposits developed on the flood-

plain. *Alnus* was an important element in the river floodplain vegetation, while the presence of *Azolla filiculoïdes* points to temporarily open water backswamp environments (Gilze). The paleocurrent at Gilze was to the west. Outside the meanderbelts soils and peats were formed locally (Weelde, unpublished). The preservation potential of these thin interglacial deposits on interfluvia was low, because they were easily eroded in the subsequent cold Menapian stage.

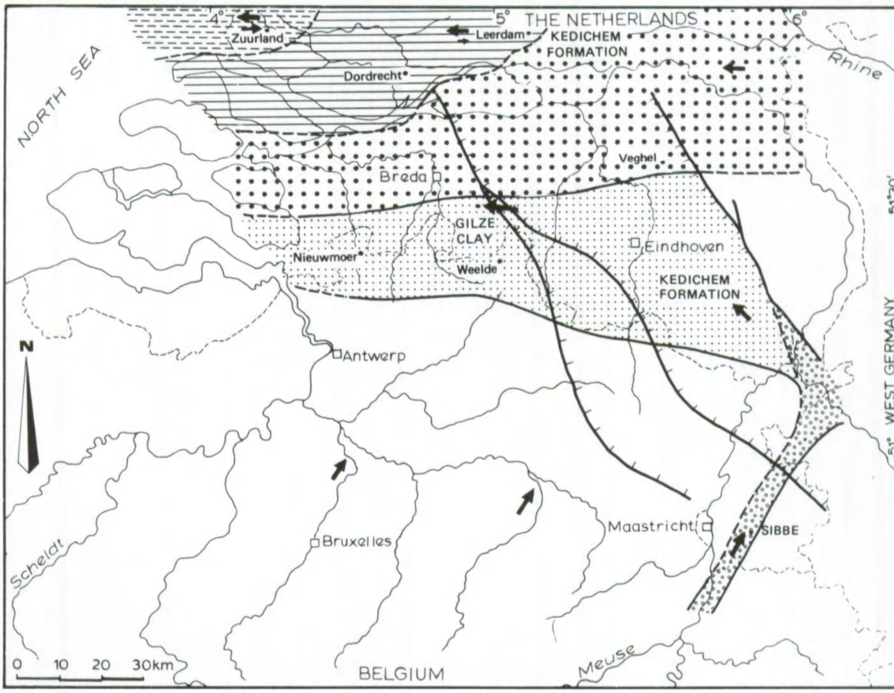


Fig. 6.9: Paleogeography of the Waalian period.

The temperate warm climate of the Waalian resulted in a high sea-level (Zagwijn, 1979), but in contrast to the previous Tiglian interglacial phases Noord-Brabant was not covered by tidal deposits (fig. 6.9). The marine transgression of the Waalian was recorded, however, in the western Netherlands. Sediments with a marine mollusc assemblage probably dating from the Waalian period have been found in the neighbourhood of Rotterdam (boring Hendrik Ido Ambacht and Zuurland; pers. comm. T. Meijer). East of Rotterdam the influence of the sea (tides) was felt as far as Dordrecht and Leerdam, which is indicated by the presence of Chenopodiaceae in the pollen record of the Waalian(A) (Zagwijn, 1974; Zagwijn and De Jong, 1984).

6.4.3 Menapian: The Meuse shifted to the northwest over the Campine plateau. Deposition of coarse-grained sediment in the confluence area of the Meuse and rivers from the Scheldt basin (Alphen Sands: Gilze Member)

After the Waalian the climate deteriorated during the Menapian. The thermophilous trees had disappeared and especially herbs and some pine

dominated the vegetation (Kinderlaan). Climate was cold, with a mean summer temperature of approximately 10° C. Permafrost conditions were occasionally present in the Menapian, indicating a mean annual temperature around -5° C (Vandenberghe and Kasse, 1988).
The Meuse followed a northwestern course (fig. 6.10). Leaving the Ardennes north of Liège the Meuse curved south of Maastricht to the northwest and possibly formed a terrace level at 140 m (pers. comm. Ing. W. Felder). Passing over the Campine plateau in Belgium the Meuse entered Noord-Brabant from the southeast. The sediments are coarse-grained and gravelly and have previously been described as Alphen Sands at Alphen and Galder (Vandenberghe and Krook, 1981; Vandenberghe et al., 1986). The presence of Revinian quartzite clearly indicates the Meuse provenance of at least the coarse gravel components.

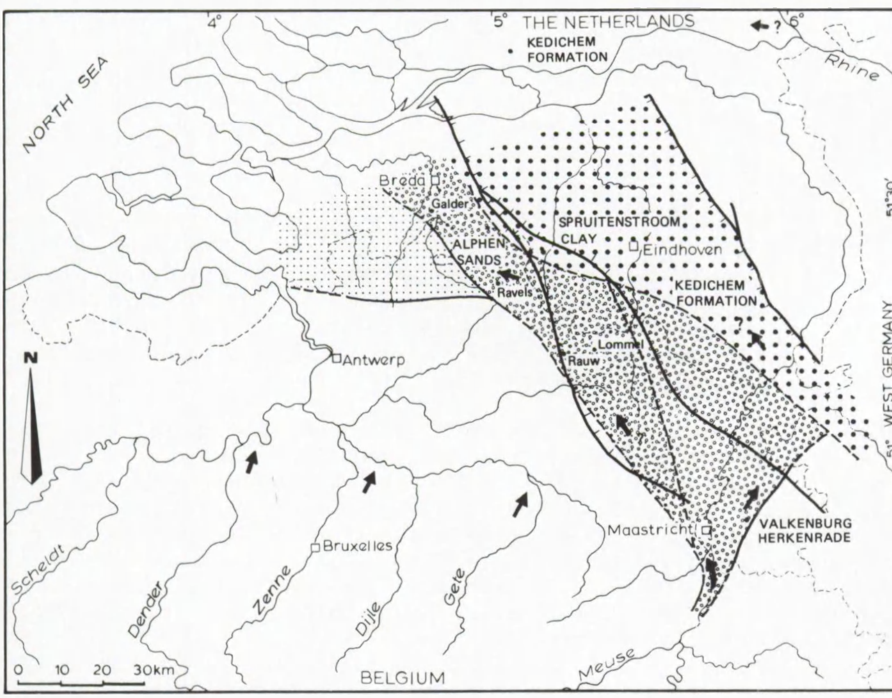


Fig. 6.10: Paleogeography of the late-Menapian period.

The Meuse sediments on the Campine plateau were probably later eroded during the Cromerian. In the section constructed in fig. 6.11 the top of the 140 m terrace on the westbank of the Meuse is connected with the top of the Gilze Member (Kedichem Formation) in Noord-Brabant. At the location of the Campine plateau the top of the Gilze Member constructed in this manner occurs above the present-day topography, which indicates erosion after deposition of the Gilze Member. The southernmost occurrences of the Gilze Member can be expected in the neighbourhood of Lommel in Belgium (Vandenberghe, 1982), since the intersection of the base of the Kedichem Formation and the present-day topography is found there (fig. 6.11).

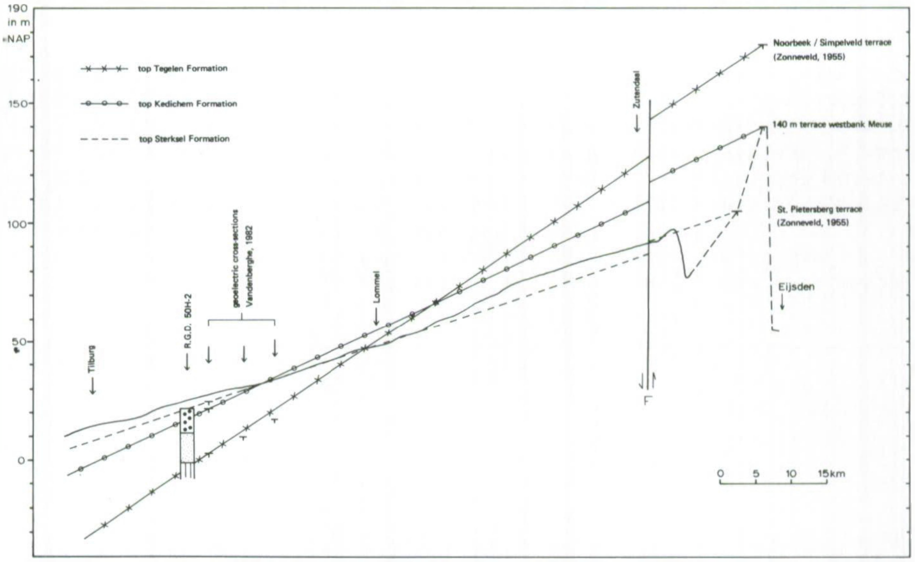


Fig. 6.11: Terrace levels, terrace intersections and vertical stacking of Early- and Middle-Pleistocene deposits in Zuid-Limburg, the Campine plateau and Noord-Brabant. The vertical displacement of the units along the fault is based on geoelectrical investigations (§2.4, fig. 2.4).

While the Meuse flowed to the northwest, the Central Belgian rivers such as the Scheldt, Dender, Zenne, Dijle and Gete continued to flow to the northeast, as in the previous Tiglian, Eburonian and Waalian (fig. 6.10). These so-called southern rivers from the Scheldt basin transported predominantly sand and fine gravel from the Tertiary source area in Central Belgium (Gullentops, 1963). These sediments, which are characterized by stable heavy minerals and flint and quartz gravel (Scheldt-association: Zandstra, 1969), dominate the sedimentation in Noord-Brabant during the Menapian. The Meuse, although clearly present, probably supplied only little sand and relatively more coarse gravel. The confluence area of the Meuse and the southern rivers may have been located on the Campine plateau along the southeast-northwest oriented Rauw fault.

In contrast to the sediments of previous periods, the sediments of Menapian age are generally coarser grained in Noord-Brabant (Alphen Sands). These were deposited by sandy braided river systems. The channels migrated rapidly and formed extensive sand bodies, consisting of stacked, trough-shaped, channel sediment (Ravels, Zwart Goor). The fine-grained overbank deposits of the floodplain (clay-, loam- and peat-beds) were seldomly preserved. The paleocurrent direction at Ravels was to the westnorthwest (fig. 6.10). As indicated by Zagwijn (1963b) the coarseness of the deposits in this period cannot be explained only by cold climatic conditions, since periglacial deposits can be fine-grained as well (such as the Appelenberg Sands of Eburonian age). Perhaps an increased tectonic activity in the hinterland area was also responsible for the change in the fluviatile regime. In the following periods (Bavelian and Cromerian) important fault activity is

recorded along the Gilze-Rijen, Hooge Mierde (§2.4) and Feldbiss faults (Zagwijn, 1963b).
The position of the Rhine in the late Menapian is somewhat uncertain. In the beginning of the Menapian the Rhine probably still followed its east-west course, as in the Eburonian and Waalian (Zagwijn, 1960; Bisschops et al., 1985). At the end of the Menapian the Rhine probably entered the Central Graben again, after its absence in the Eburonian, Waalian and early Menapian (fig. 6.10). The Rhine extended its influence westwards to approximately the Hooge Mierde fault, where sediments of a Rhine provenance (Spruitenstroom Clay bed) were laid down on the previously deposited sediments of the Meuse and the southern rivers. This westward shift of the Rhine culminated later in the Bavelian and Cromerian.

6.5 Bavelian-Cromerian: Increased faulting and fluviatile deposition by the Rhine and the Meuse

6.5.1 Bavelian: Deep incision by the Rhine in Noord-Brabant and infilling with fine fluviatile sediment of the Bavel Member (Sterksel heavy mineral zone)

The Rhine followed a southeast-northwest course through Germany, the Roer Graben and Central Graben (fig. 6.12). During the beginning of the Bavel Interglacial of the Bavelian stage the Rhine eroded a deep valley (10-20 m) in the Gilze Member, possibly because of increased faulting and connected uplift of the region west of the Central Graben.

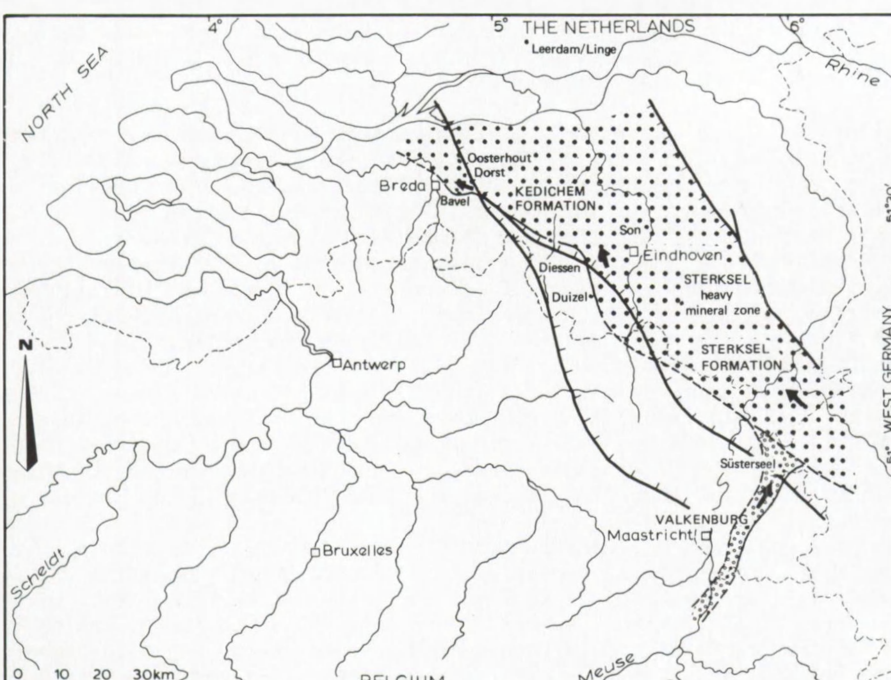


Fig. 6.12: Paleogeography of the Bavelian period.

The southwestern bluff of the valley is found at Duizel-Diessen (southwest of Eindhoven) (Bisschops et al., 1985) and Bavel. In the neighbourhood of Bavel the Rhine crossed the Gilze-Rijen/Feldbiss fault to the northwest. The course of the Rhine over the Campine plateau in Belgium is unknown.
The incised valley at Bavel was filled with fine sediments (Bavel Member) of Rhine provenance (Sterksel heavy mineral zone) during the Bavel Interglacial. The Rhine was probably characterized by a meandering river regime. In channel cut-offs more than 10 m thick clay-beds developed, which is an indication for the great channel depth and the dimension of the river. The clay-plugs were only partly eroded by river channels in a later phase. Their preservation is probably due to their position close to the southwestern margin of the southeast-northwest oriented floodplain.
After the Menapian the Meuse had shifted to the northeast and left its former depositional area in Noord-Brabant (fig. 6.10 and 6.12). The influence of the Meuse and southern rivers from Central Belgium was not detected in the sediment at Bavel. The Meuse probably took a northern course through Zuid-Limburg (fig. 6.12), since sediments dating from the Bavel Interglacial have been found at Süsterseel overlying the Meuse deposits of the Valkenburg terrace (Zagwijn and De Jong, 1984; Zagwijn, 1986).
At the end of the Bavelian (Leerdam Interglacial) the Rhine shifted to the northeast again. The Meuse and other southern rivers extended their influence to the north, which is indicated by the stable heavy mineral zone in the upper part of the Kedichem Formation at Leerdam (Zagwijn and De Jong, 1984).

6.5.2 **Cromerian**: increased tectonic activity and large-scale deposition of coarse fluvial sands by rapid lateral migration of the Rhine and Meuse (Sterksel Formation)

During the Cromerian tectonic activity became more important, expressed by increased faulting during deposition of the Sterksel Formation (fig. 6.13). The subsidence of the Central Graben led to a reinforcement of the Rhine sedimentation in the area. The increase in grain-size in the upper part of the Gilze Member (Alphen Sands) in the Menapian and the occurrence of the Rhine in the investigated area in the Bavelian possibly indicate the first signs of tectonic activity (in the hinterland). According to the vertical displacement and differences in thickness of the units along the Gilze-Rijen fault, the major tectonic activity occurred during deposition of the Sterksel Formation and afterwards (fig. 6.13). East of the fault the Sterksel Formation is up to 40 m thick, while this unit is nearly absent on the uplifted block west of the fault. The Gilze-Rijen fault seems to have been inactive in the previous (Tiglian?) Eburonian, Waalian and Menapian, since the Gilze Member (Kedichem Formation) shows the same thickness on both sides of the fault.
The paleogeographic reconstruction for the Cromerian is presented in fig. 6.14. During the Interglacial I (Waardenburg) and Glacial A of the Cromerian (Zonneveld, 1974; Zagwijn, 1985) the Rhine temporarily withdrew from the (western) Central Graben (fig. 6.14: phase 1). The Meuse formed the St. Geertruid terrace (and St. Pietersberg) in Zuid-Limburg (Zonneveld, 1974). The latter river extended its depositional area to the north and deposited sediments of the Budel heavy mineral zone as far as Eindhoven (Zonneveld, 1947). On the Campine plateau, the Meuse possibly began to deposit the Winterslag Sands (Gullentops, 1974).

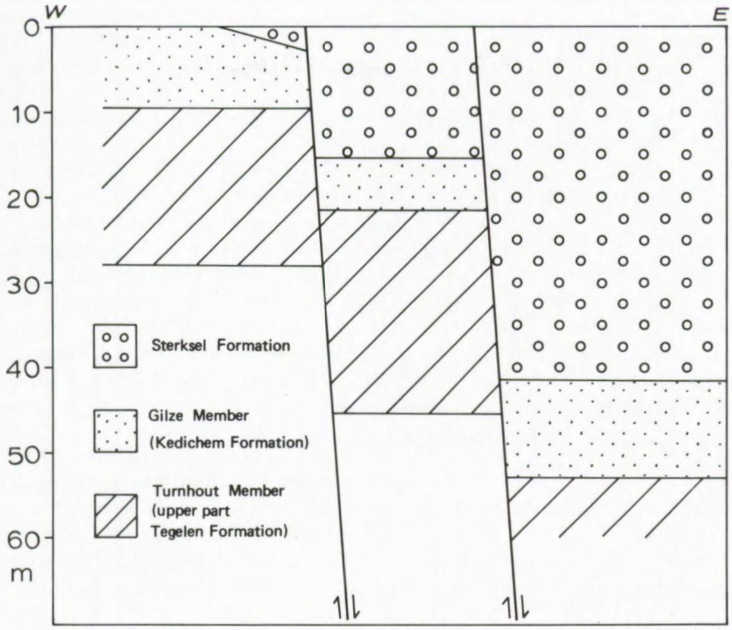


Fig. 6.13: Fault activity of the Gilze-Rijen fault at Gilze during the Early- and Middle-Pleistocene as established in figure 2.8.

Figure 6.14, phase 2 gives the paleogeographic reconstruction for the Cromerian Glacial A, Interglacial II (Westerhoven) and Glacial B phase (Zagwijn, 1985), when the Rhine reached its maximal western extension and deposited the Woensel heavy mineral zone (Zonneveld, 1947). The Rhine had shifted far to the southwest with respect to the previous phase 1. Entering The Netherlands from the southeast via the Roer Graben, the Rhine flowed to the west over the Campine plateau and deposited the Lommel Sands (Gullentops, 1974).

The Meuse followed, as in the late-Menapian, a northwestern course and deposited the Winterslag Sands on the Campine plateau (Gullentops, 1974) and the deposits of the St. Pietersberg terrace in Zuid-Limburg (Zonneveld, 1974). These Meuse sediments date from the Cromerian, because they merge northwards into Rhine deposits of Cromerian Interglacial II age (Westerhoven) (Zagwijn, 1985). In figure 6.11 the top of the St. Pietersberg terrace in Zuid-Limburg is connected with the top of the Sterksel Formation (Woensel heavy mineral zone) in the western part of the Central Graben. The top of the Sterksel Formation constructed in this manner, coincides fairly well with the present-day topography of the Campine plateau. Therefore, the sediments on the Campine plateau are probably equivalent to the Sterksel Formation in The Netherlands. The topography of the southern part of the Campine plateau occurs slightly above the constructed top of the Sterksel Formation. Part of the deposits on the southern Campine plateau are therefore probably somewhat older than the St. Pietersberg terrace and might be correlated with the St. Geertruid terrace (Zonneveld, 1974).

The confluence area of Rhine and Meuse occurred on the Campine plateau near Helchteren (fig. 6.14, phase 2). Typical Meuse minerals (Vosges hornblende) were found north of the confluence area as far as Hooge

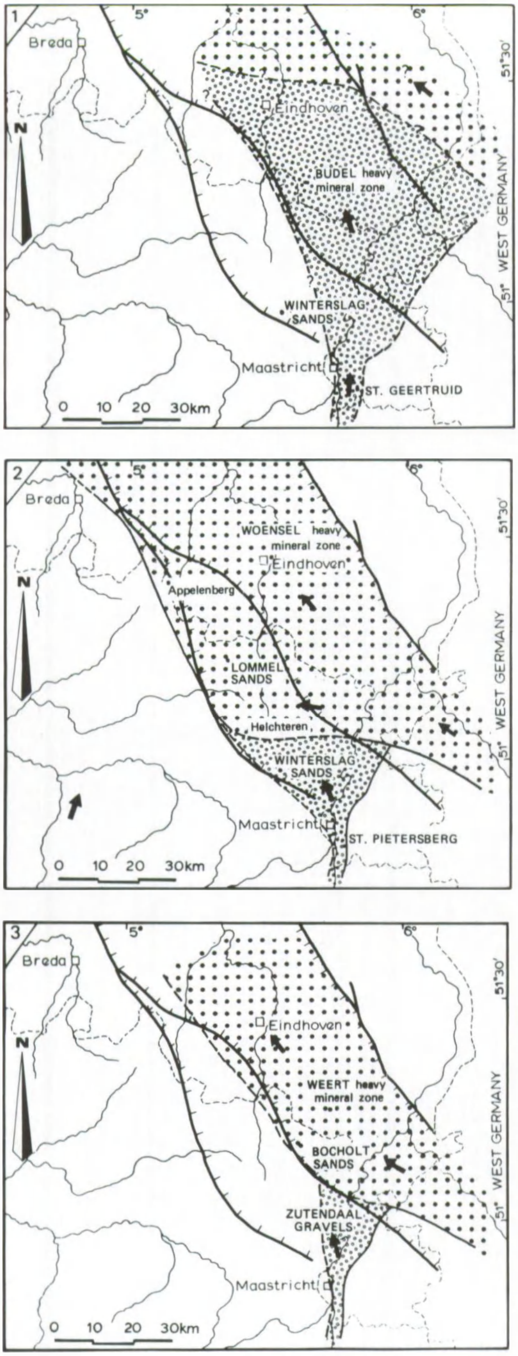


Fig. 6.14: Shifting of the Rhine and Meuse during the Cromerian.

Mierde (Appelenberg). West of Helchteren the Rhine and Meuse bounced against the uplifted block west of the Rauw fault and then followed the Rauw/Hoge Mierde/ Gilze fault to the northwest. The Rhine and Meuse locally crossed the Hooge Mierde fault and the uplifted Gilze Member was severely eroded. Predominantly coarse channel sediment (sand and gravel) was deposited (Appelenberg) in such situations (fig. 6.14, phase 2). Continuous subsidence of the Central Graben caused the stacking of channel sands in thick sequences (fig. 6.13). Clay-beds representing channel-fill or floodplain deposition rarely developed. The high channel/ floodplain sediment ratio was caused by the active migration of river channels, which is often associated with braided river systems. This does not necessarily involve cold climatic conditions during the deposition. As tectonic activity was important in the Cromerian, the river gradient could have been relatively steep. In combination with the coarse sediment, this steep gradient could have resulted in braided river systems during glacial and interglacial periods (Zagwijn, 1963b). Part of the channel deposits indeed originated in an interglacial period (Zagwijn and Zonneveld, 1956: Westerhoven Clay).

The southern rivers from Central Belgium supplied relatively less sediment in this period. The deposits in Noord-Brabant and on the Campine plateau are dominated by Rhine and Meuse components, although a higher flint content near

the southern boundary of the depositional area may point to an admixture with sediment from the Scheldt basin.
In the following Cromerian Glacial B period (Zonneveld, 1974; Zagwijn, 1985) the Rhine migrated to the northeast (fig. 6.14, phase 3). It deposited the Weert heavy mineral zone in The Netherlands (Zonneveld, 1947) and the Bocholt Sands in Belgium (Gullentops, 1974), predominantly east of the Feldbiss fault. The Meuse extended its influence to the northeast and the Zutendaal Gravels were laid down over the previously deposited Lommel Sands of Rhine provenance (Gullentops, 1974).
In the paleogeographic reconstructions presented above, the Winterslag Sands are regarded to a certain extent as a time transgressive unit (fig. 6.14: phase 1, 2). During the deposition of the Woensel and Weert heavy mineral zones by the Rhine, the Meuse rotated clockwise to the east. The depositional area of the Meuse was primarily determined by the location of the Rhine. The high concentration of heavy minerals in Rhine sediments dominated in the sediments north of the confluence area.

6.6 **Middle- and Late-Pleistocene:** western Noord-Brabant and northern Belgium outside the depositional area of the Rhine, Meuse and "Scheldt"; important erosion in the study area

This paleogeographic reconstruction of western Noord-Brabant and northern Belgium is concluded with an overview of the Middle- and Late-Pleistocene evolution of the area (fig. 6.15). In contrast to Early-Pleistocene stages, which are characterized by deposition and vertical stacking, the Middle- and Late-Pleistocene development is dominated by erosion. The Rhine and Meuse had shifted further to the northeast (Urk and Veghel Formations), because of tectonic activity of the Central Graben and Peel Horst (Zonneveld, 1947).
The upthrown block west of the Central Graben was eroded and Middle- and Late-Pleistocene sediments are therefore scarce in western Noord-Brabant and northern Belgium. Because of this uplift the southwest-northeast oriented rivers from Central Belgium probably began to erode their lower courses. The downcutting in the underlying sands of the Gilze Member continued until the massive clay-beds of the Rijkevorsel and Turnhout Members were reached. These resistant east-west oriented clay-beds hampered further erosion and perhaps by the concurrent opening of the English Channel (Sommé, 1979; Zagwijn, 1979: Cromerian or Holsteinian) the rivers from Central Belgium diverged to the northwest. Subsequent erosion in the Nete basin, south of the occurrences of the resistant Tiglian clay-beds, led to the genesis of the Campine microcuesta between Turnhout and Ossendrecht (De Ploey, 1961) (fig. 6.15). The westward flowing rivers induced increased erosion in northwestern Belgium (Flemish Valley)(Tavernier and De Moor, 1974; Vandenberghe and De Smedt, 1979) and in the Dutch province of Zeeland. This erosion could have been responsible for the genesis of the Scheldt escarpment between Antwerp and Halsteren (fig. 6.15), where at least 30 m of the Hoogerheide and Woensdrecht Members were eroded during the Middle- and Late-Pleistocene.
After the upper courses of the Central Belgian rivers had been captured and a new west-oriented drainage system had come into being, erosion continued in the south-north oriented lower reaches which are nowadays known as the Molenbeek, Weerijs and Mark brooks. The Gilze Member, originally present as a continuous deposit in Noord-Brabant, was almost completely eroded in these lower reaches. For instance in the Mark valley approximately 15 m were removed from the Gilze Member and the

Turnhout Member. Only a pebble lag with a Scheldt gravel association testifies the original presence of the Gilze Member. Between the brook valleys erosion was less severe and the sediments of the Gilze Member were preserved in plateau like structures (Wouwse Plantage, Rucphense Bossen, Baarle Nassau). The Baarle Nassau plateau (25-30 m +N.A.P.) was formed during the Middle- and Late-Pleistocene between the subsiding Central Graben in the east (20 m +N.A.P.) and the incision of the Mark in the west (10-15 m +N.A.P.).

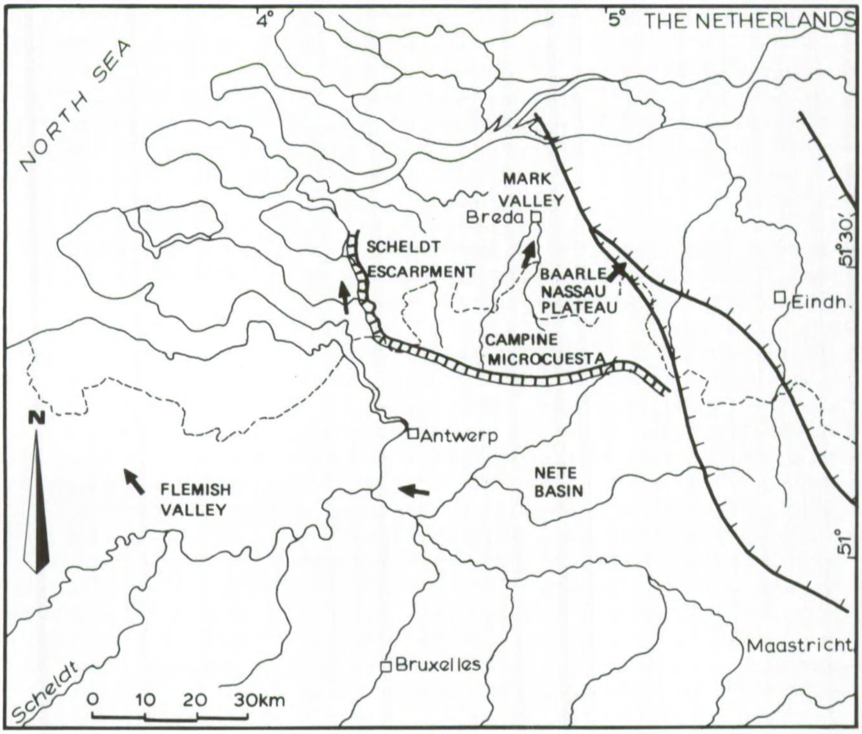


Fig. 6.15: Erosion of Noord-Brabant and northern Belgium during the Middle- and Late-Pleistocene.

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My uncle Wim stimulated unconsciously my interest in physical geography by his permission to dig pits in his land.
Finally, my friend Els, who remained my dearest companion in stressful conditions. Her mental and physical support enabled me to finish this thesis.

APPENDIX

LEGEND OF THE LITHOLOGY IN BORINGS AND EXPOSURES

SAND

very coarse (vc)(> 300 µm)

medium coarse (mc)(210-300 µm)

medium fine (mf)(150-210 µm)

very fine (vf)(105-150 µm)

extremely fine (ef)(< 105 µm)

clay

silt

peat

gyttja

sandy

very sandy

clayey

very clayey

humose

very humose

silty

very silty

peaty

SAMPLES:

heavy minerals

grain-size

gravel

pollen

macro remains

paleomagnetism

cation exchange capacity

reworked organic material

fine gravel

clay-pebbles

silt-pebbles

peat-pebbles

glimmers (muscovite)

glauconite

rooted

marine shell-fragments

very calcareous

calcareous

non-calcareous

erosive boundary

fining-up sequence

coarsening-up sequence

involutions

wedge structures

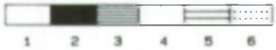
LEGEND OF HEAVY MINERALS

- garnet
- epidote
- saussurite
- alterite
- hornblende
- chloritoid
- volcanic minerals
- other minerals
- topaz
- staurolite
- metamorphic minerals
- tourmaline

LEGEND OF GRAIN-SIZE

- > 420 μm
- 300-420 μm
- 212-300 μm
- 150-212 μm
- 105-150 μm
- 75-105 μm
- 53-75 μm
- 16-53 μm
- 2-16 μm
- 0-2 μm

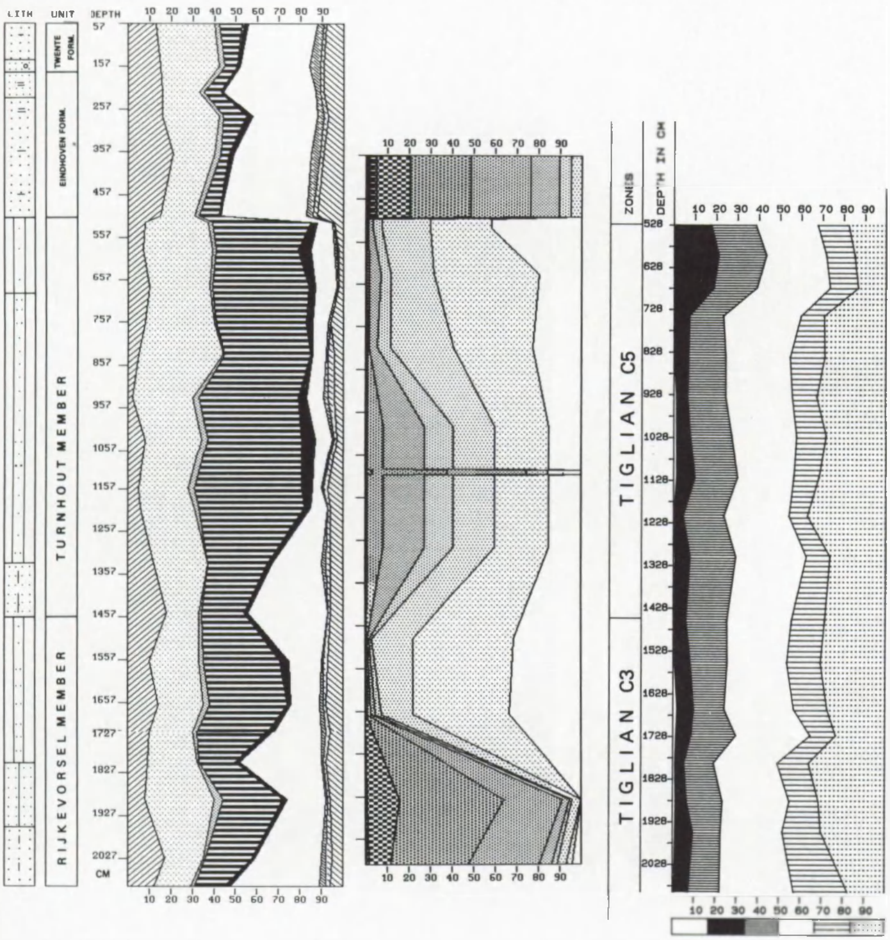
LEGEND OF POLLEN DIAGRAMS



MAIN DIAGRAM:
1. Tertiary trees
2. Warmth-loving trees of relatively dry soil
3. Warmth-loving trees of relatively wet soil
4. "Indifferent trees"
5. Terrestrial herbs, excl. Ericales
6. Ericales

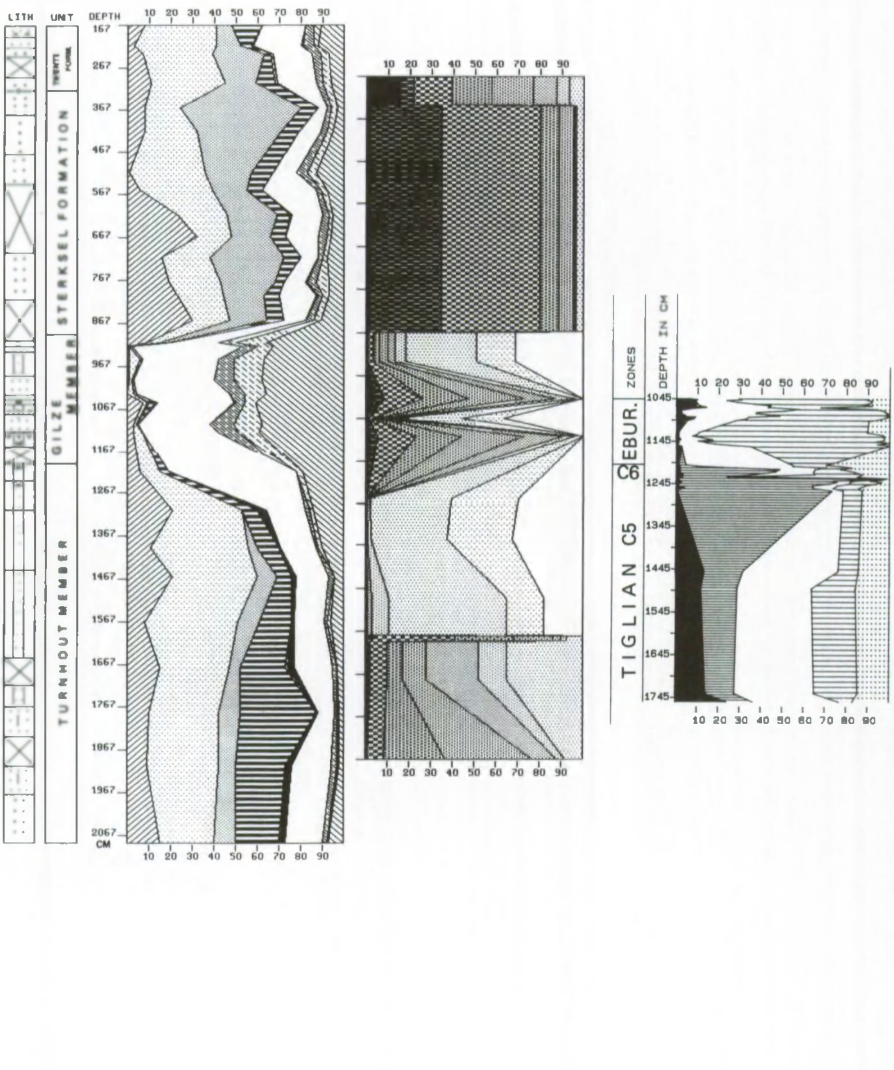
ACHTMAAL

X = 4°33'59''OL
Y = 51°26'19''NB
H = 12.5 m +N.A.P.



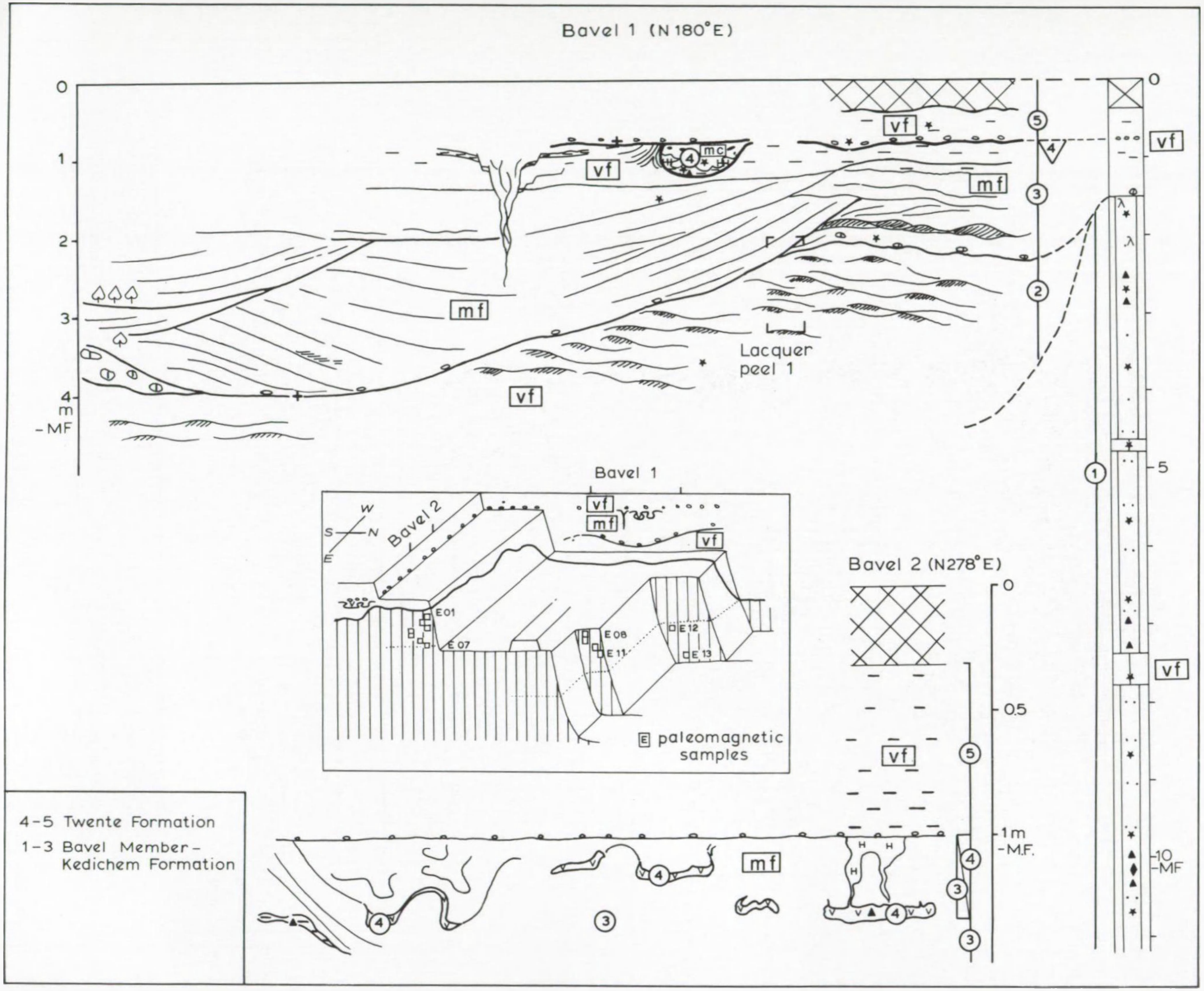
APPELENBERG

X = 5°06'26''OL
Y = 51°25'21''NB
H = 29 m +N.A.P.



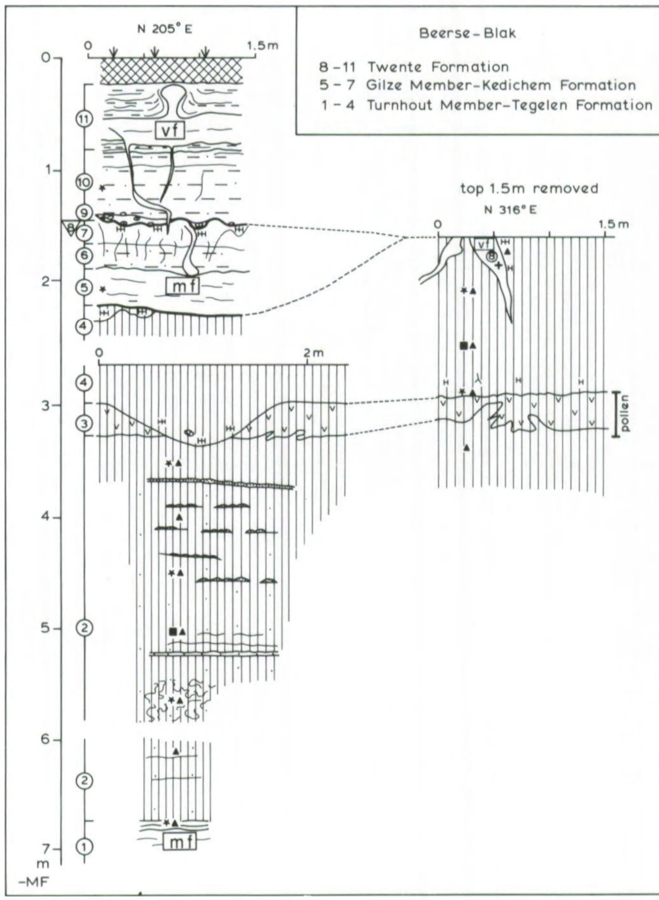
BAVEL

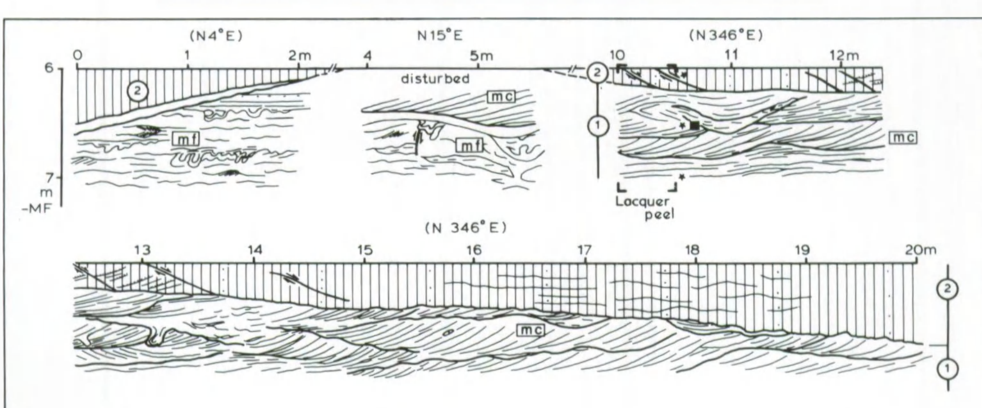
X = 4°50'14''OL
Y = 51°35'00''NB
H = 4 m +N.A.P.

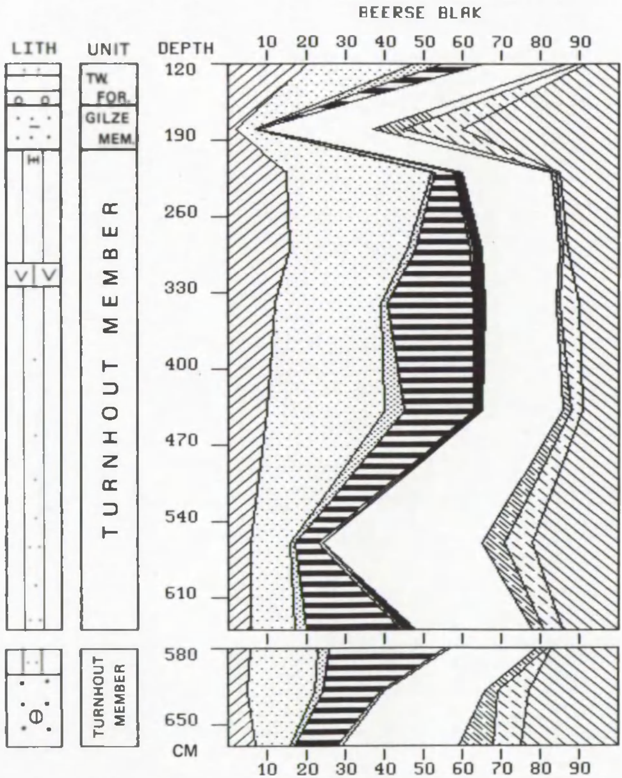


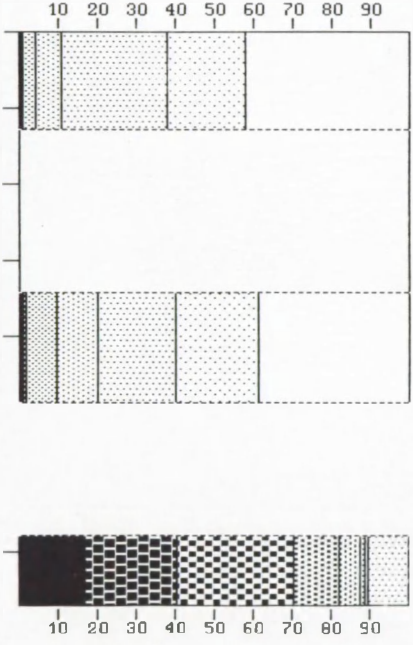
BEERSE BLAK

X = 4°49'03''OL
Y = 51°20'20''NB
H = 27.5 m +N.A.P.

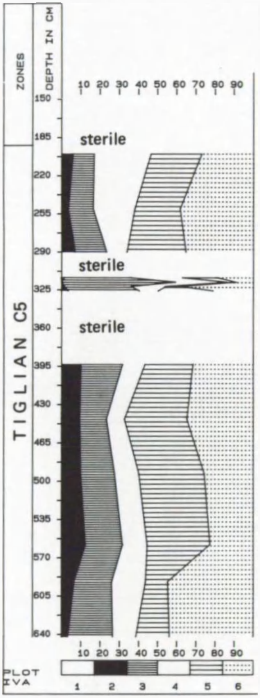





BEERSE BLAK
LITH
UNIT
DEPTH
10 20 30 40 50 60 70 80 90
TW. FOR.
GILZE MEM.
TURNHOUT MEMBER
120
190
260
330
400
470
540
610
580
650
CM
TURNHOUT MEMBER



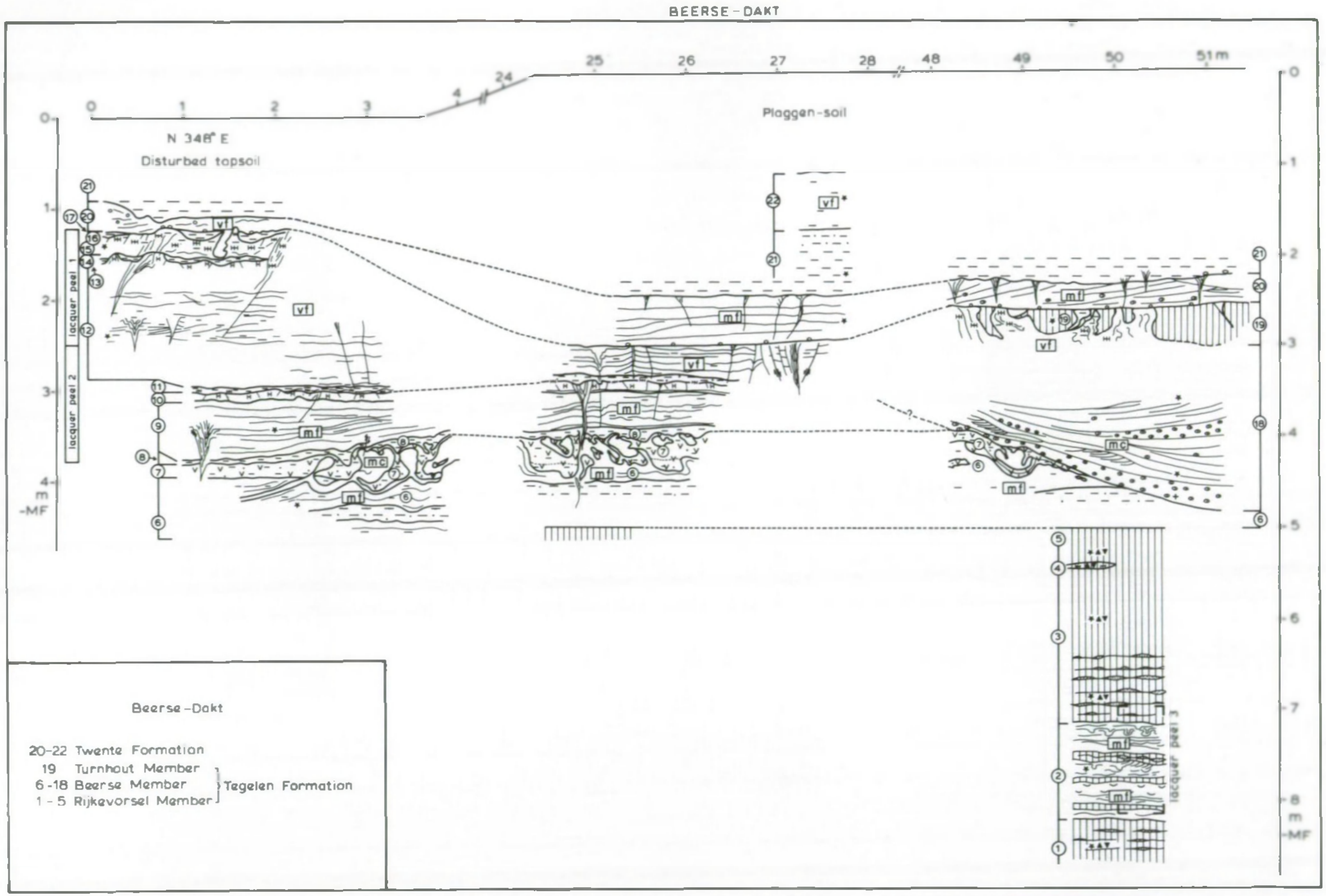
10 20 30 40 50 60 70 80 90
10 20 30 40 50 60 70 80 90

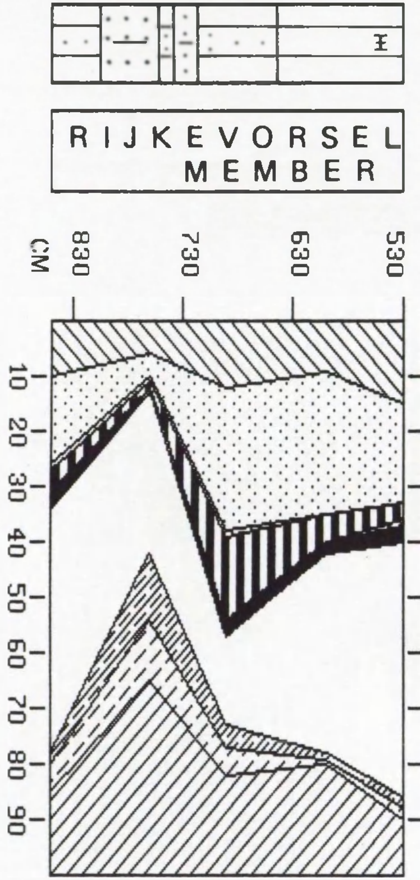


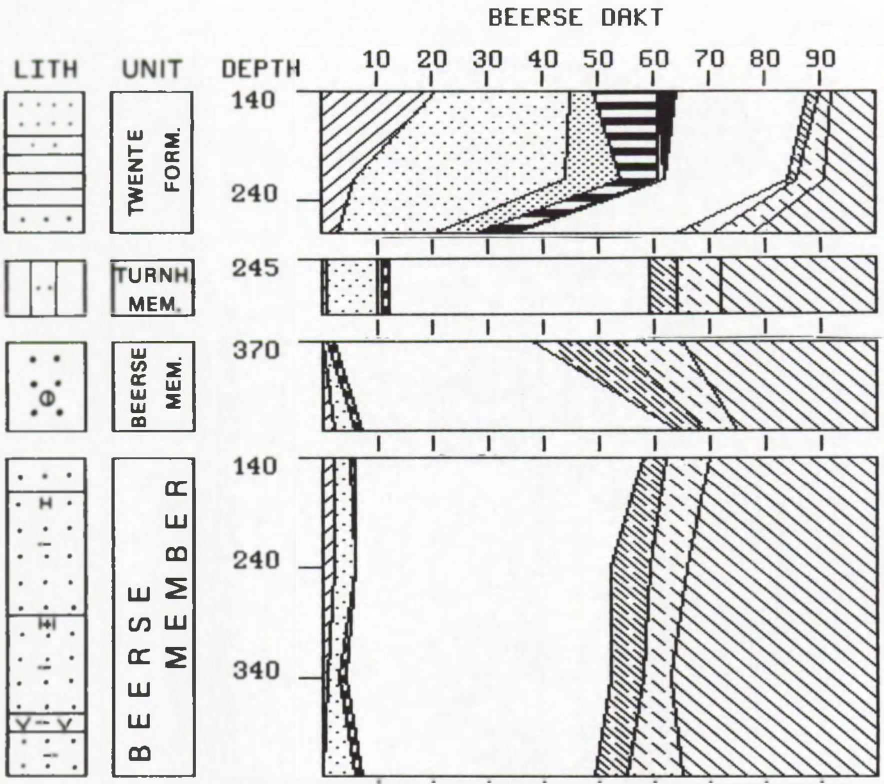
ZONES
DEPTH IN CM
10 20 30 40 50 60 70 80 90
150
185
sterile
220
255
290
sterile
325
360
sterile
395
430
465
500
535
570
605
640
TIGLIAN C5
10 20 30 40 50 60 70 80 90
PLOT IVA
1 2 3 4 5 6

BEERSE DAKT

X = 4°52'05''OL
Y = 51°19'55''NB
H = 28 m +N.A.P.

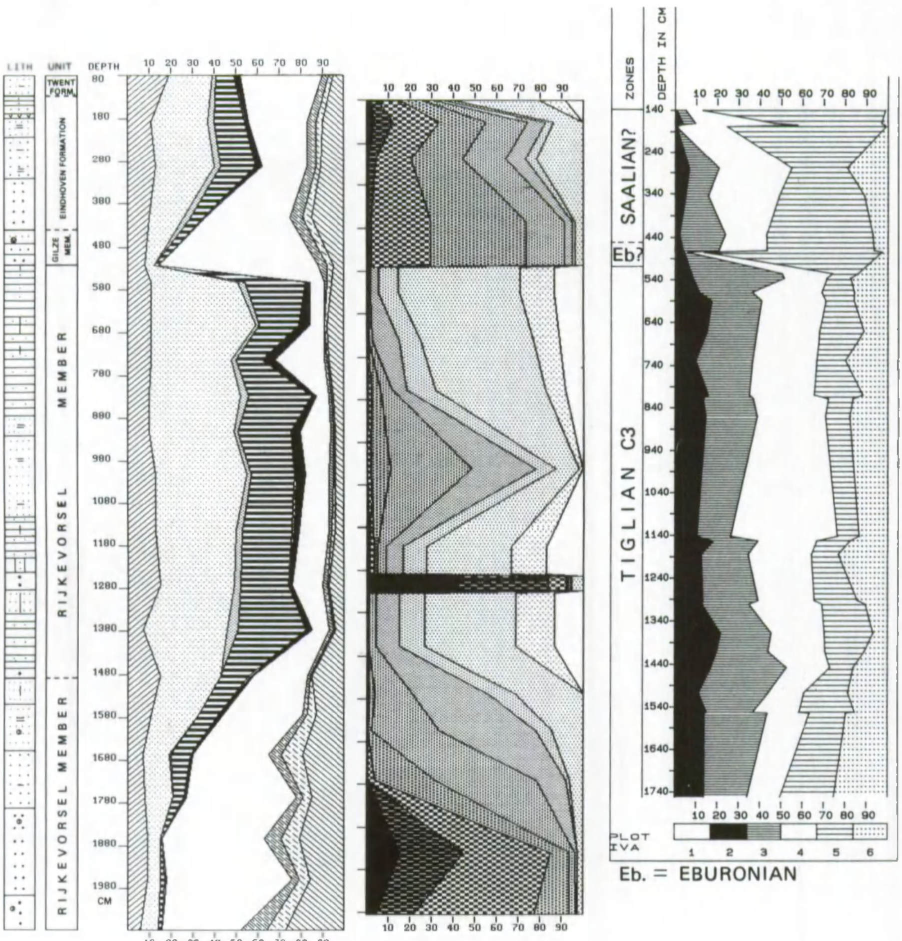



RIJKEVORSEL MEMBER
CM
830
730
630
530
10
20
30
40
50
60
70
80
90


BEERSE DAKT
LITH
UNIT
DEPTH
10
20
30
40
50
60
70
80
90
TWENTE FORM.
140
240
TURNH MEM.
245
BEERSE MEM.
370
BEERSE MEMBER
140
240
340

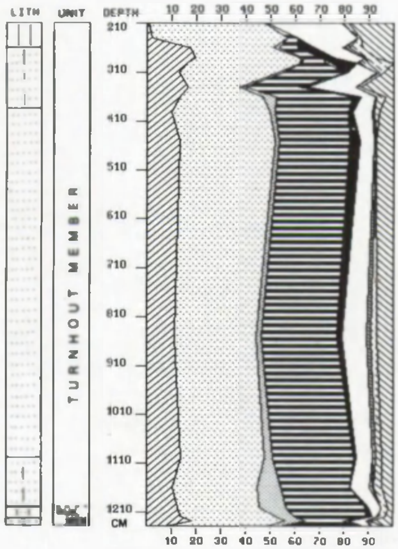
BOLK

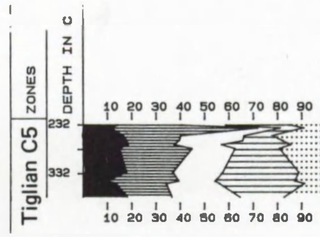
X = 4°48'16''OL
Y = 51°22'55''NB
H = 18 m +N.A.P.



CASTELRE

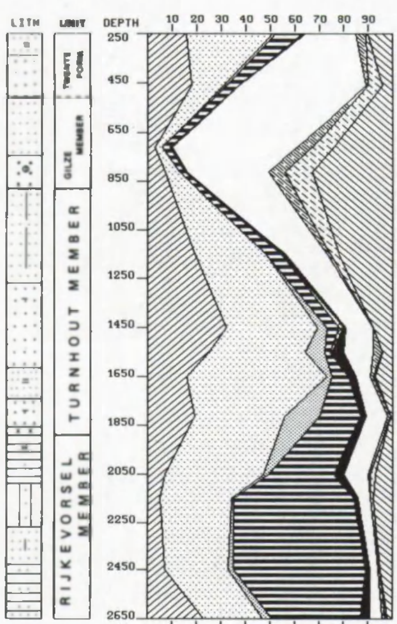
X = 4°46'50''OL
Y = 51°25'40''NB
H = 14.5 m +N.A.P.

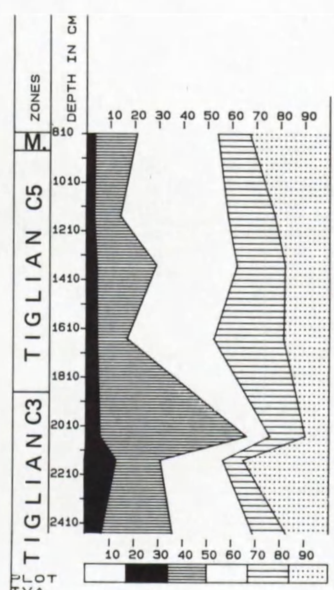




GALDERSE MEREN

X = 4°45'20''OL
Y = 51°31'33''NB
H = 6.2 m +N.A.P.

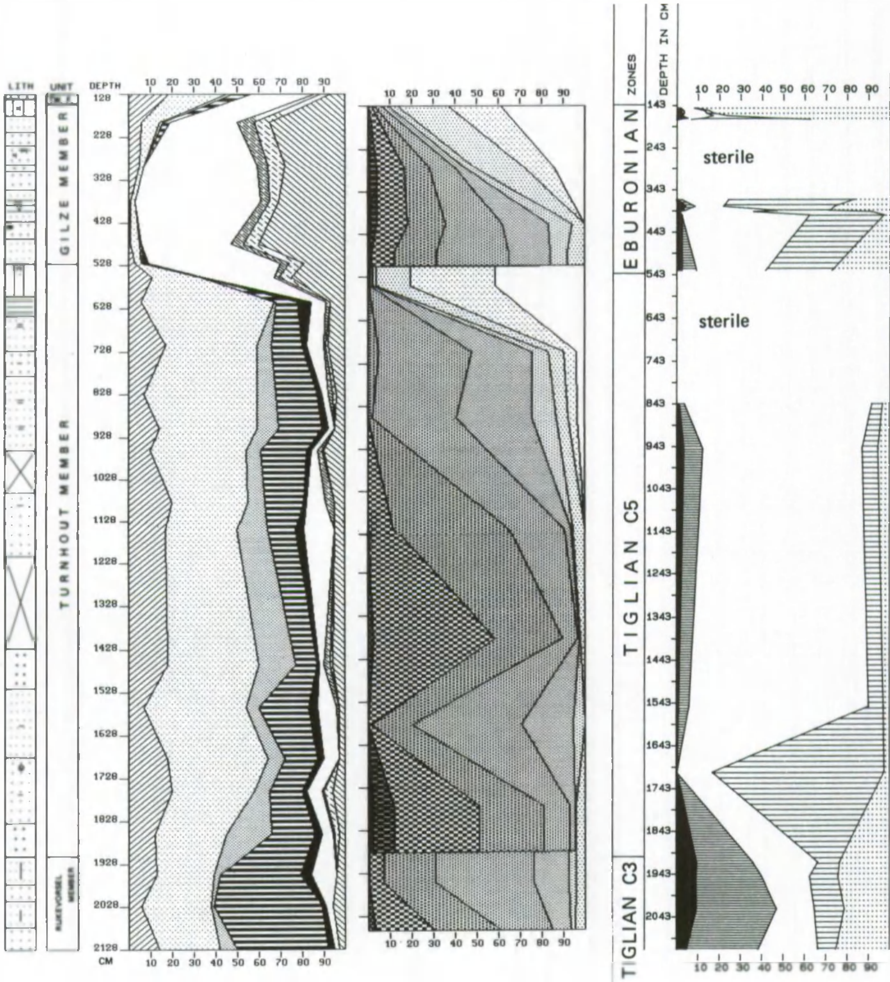




M. = MENAPIAN

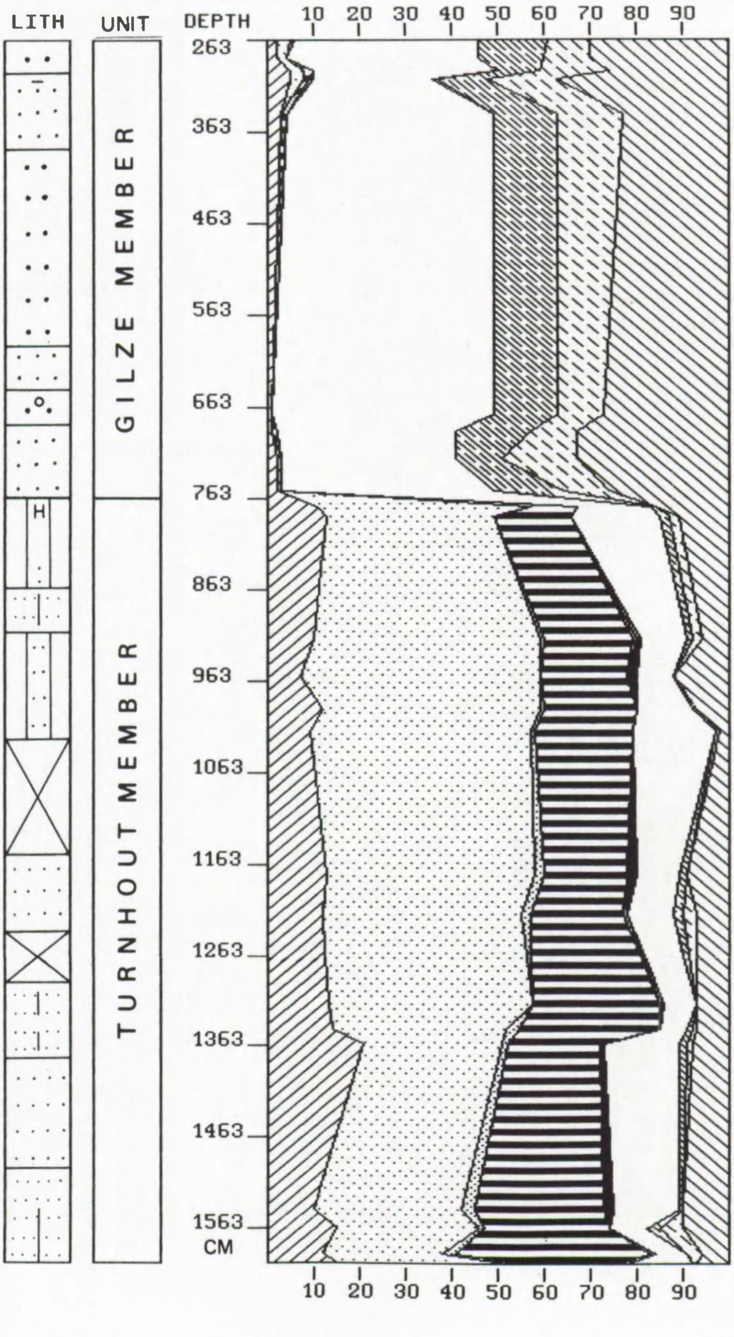
CHAAM KAPEL

X = 4°50'56''OL
Y = 51°30'15''NB
H = 11.5 m +N.A.P.

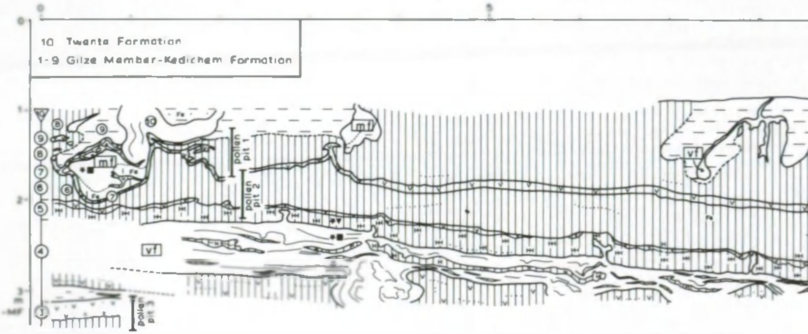


GHIL

X = 4°55'53''OL
Y = 51°24'29''NB
H = 25 m +N.A.P.



Gilze (N 6° E)

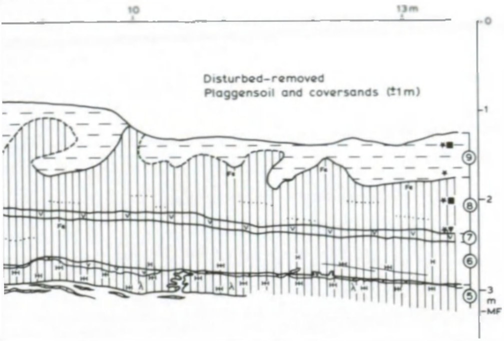

10 Twente Formation
1-9 Gilze Member-Kedichem Formation
pollen pit 1
pollen pit 2
pollen pit 3
mf
vf
Fe
0
1
2
3
m
-MF
5

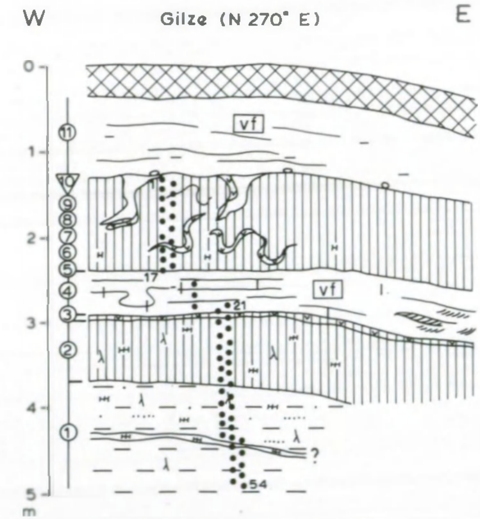
GILZE

X = 4°56'16''OL

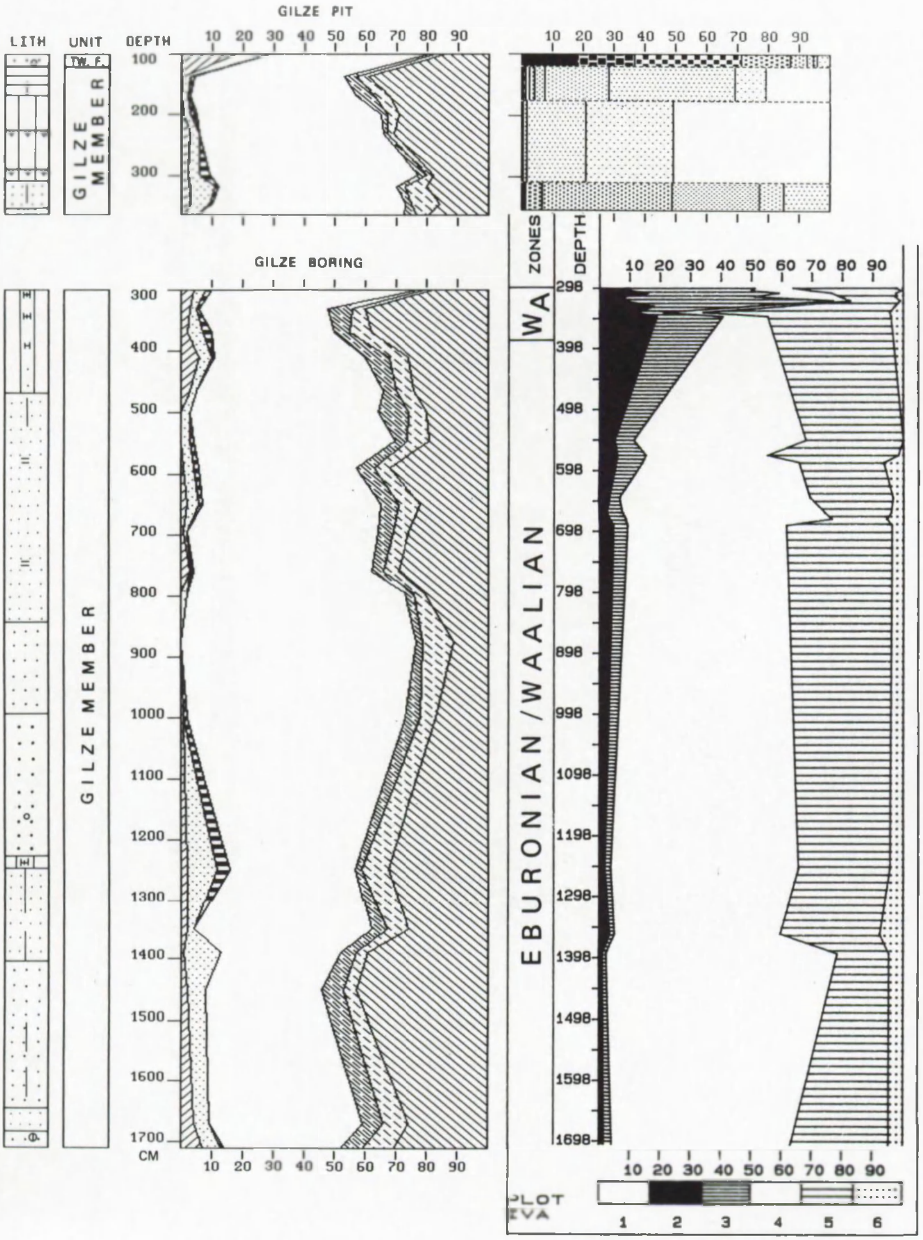
Y = 51°31'45''NB

H = 16 m +N.A.P.





• palaeomagnetic measuraments

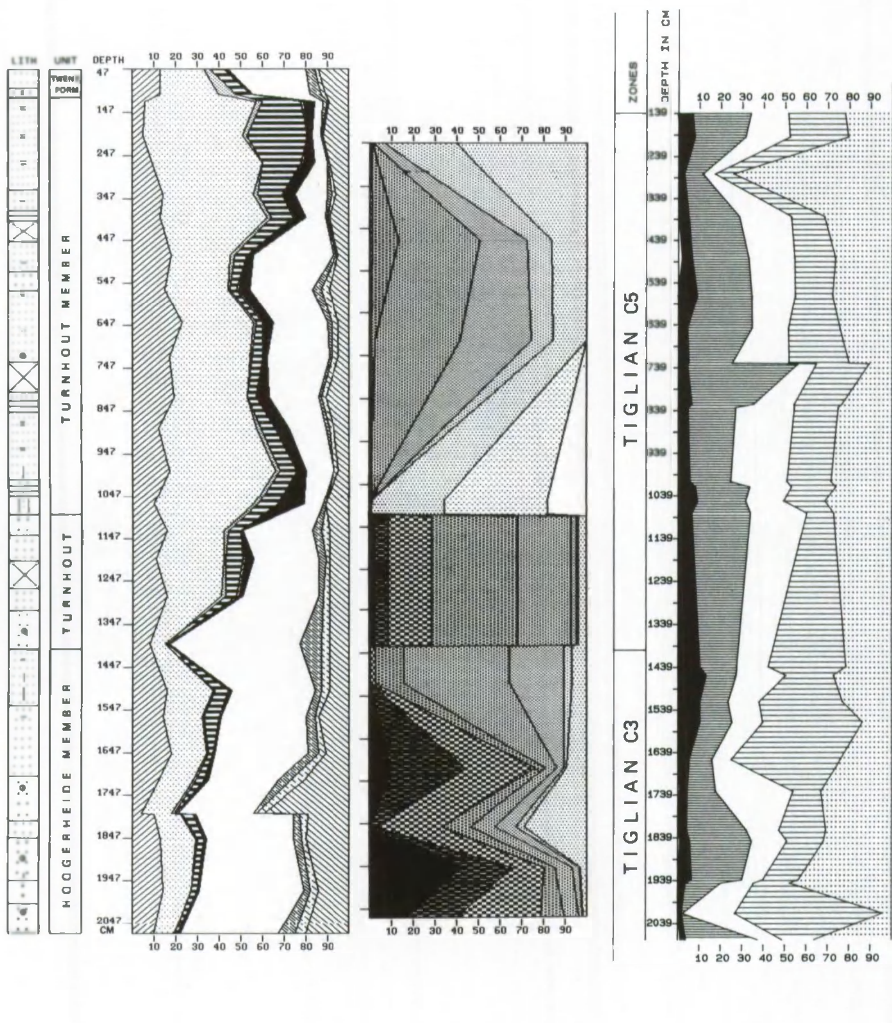


GILZE PIT
LITH
UNIT
DEPTH
TW. F.
GILZE MEMBER
10 20 30 40 50 60 70 80 90
100
200
300
GILZE BORING
300
400
500
600
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
CM
ZONES
DEPTH
WA
EBURONIAN / WAALIAN
298
398
498
598
698
798
898
998
1098
1198
1298
1398
1498
1598
1698
1 2 3 4 5 6

W_A = WAALIAN A

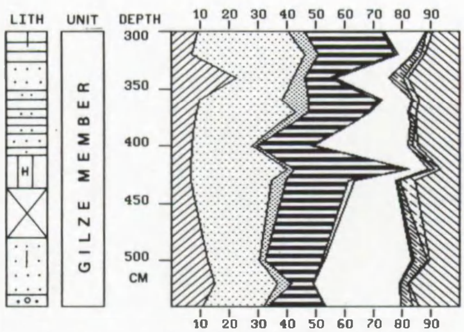
KALMTHOUTSE HOEK

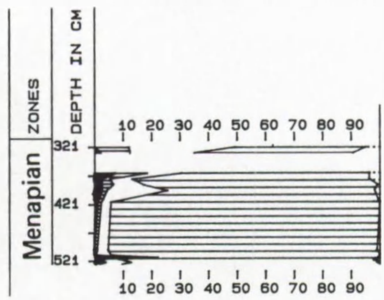
X = 4°24'42''OL
Y = 51°26'02''NB
H = 14 m +N.A.P.



KINDERLAAN

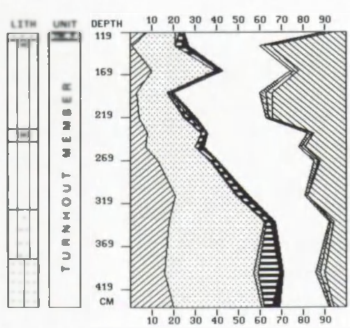
X = 5°09'20''OL
Y = 51°26'47''NB
H = 20 m +N.A.P.

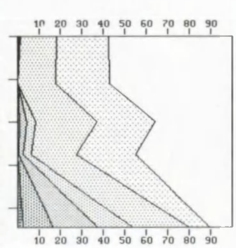


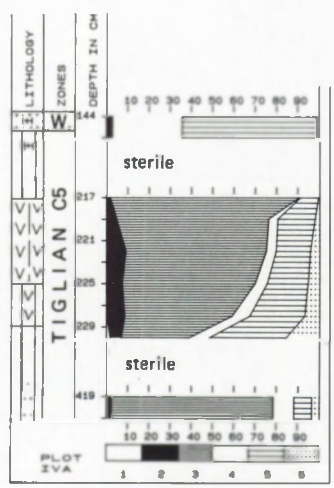


MEERLE

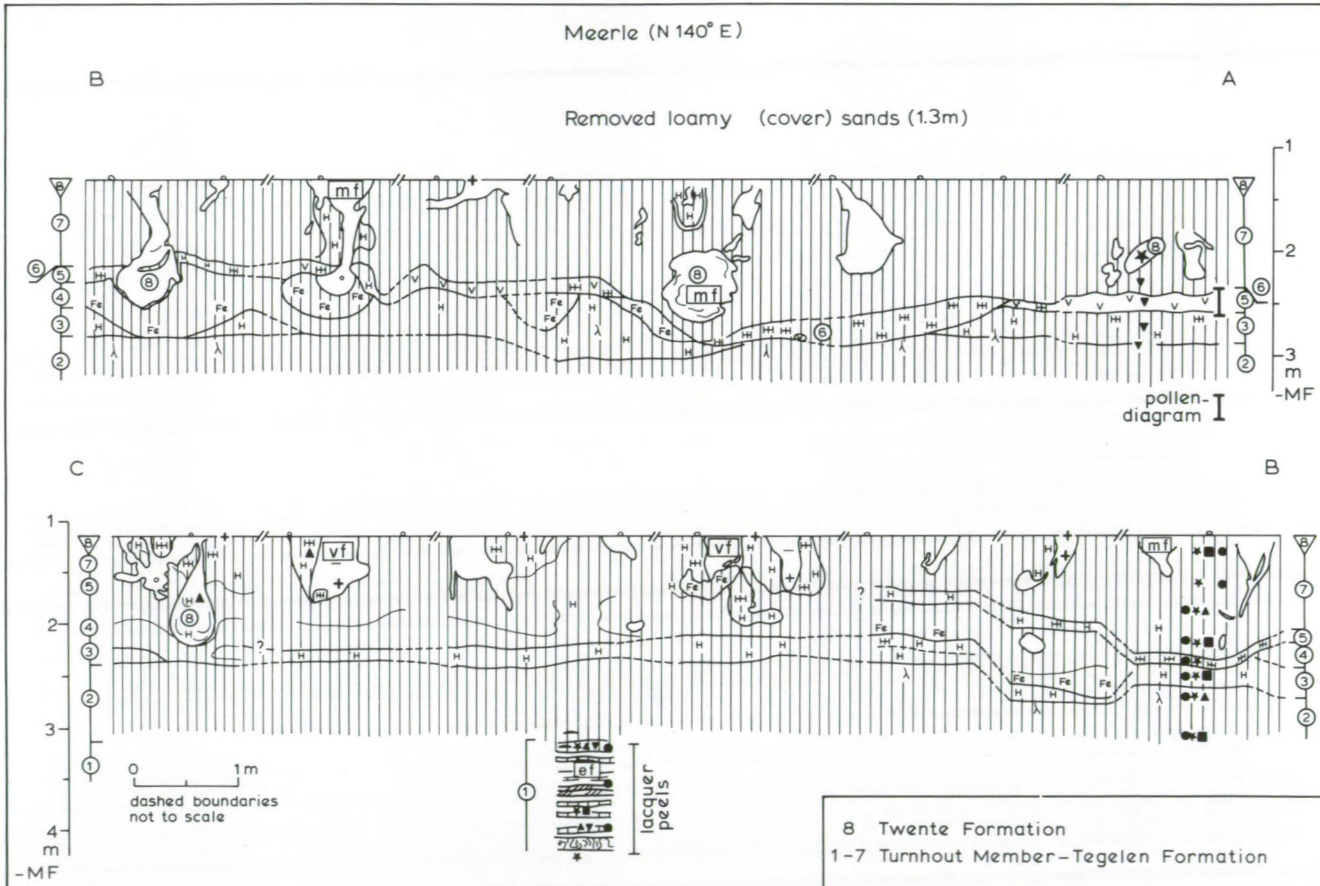
X = 4°47'57''OL
Y = 51°26'26''NB
H = 15 m +N.A.P.





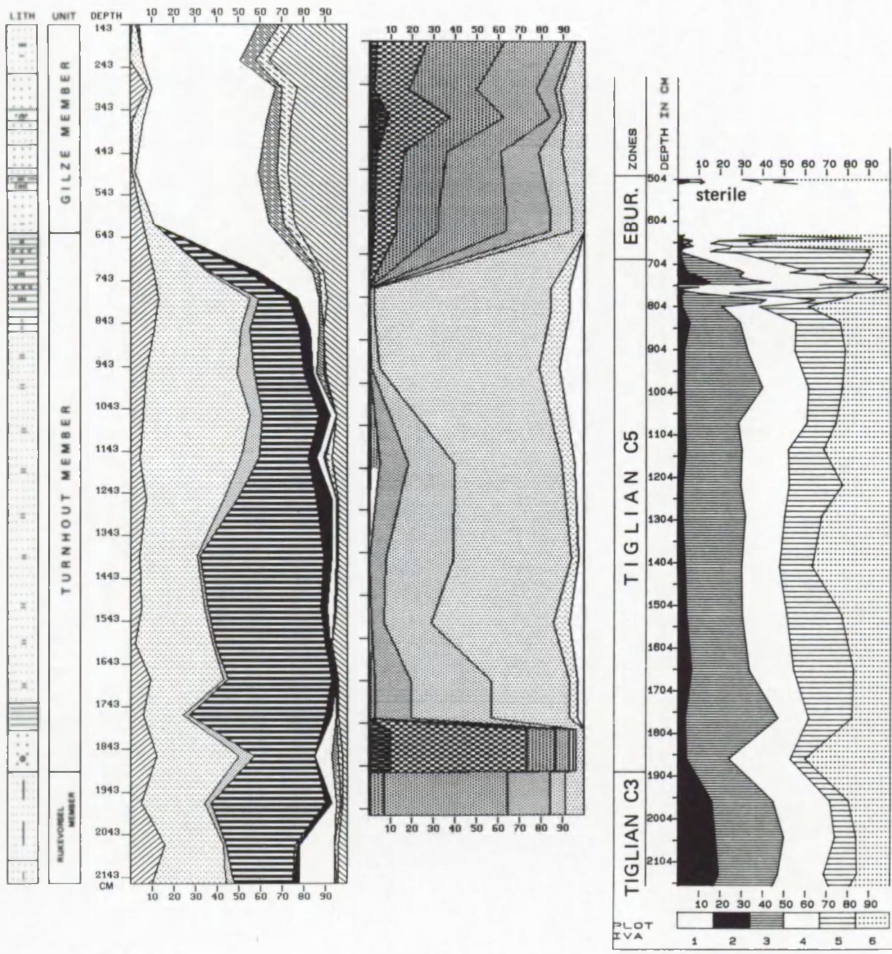


W. = WEICHSELIAN


Meerle (N 140° E)
B
A
Removed loamy (cover) sands (1.3m)
pollen-diagram
m
-MF
C
B
dashed boundaries not to scale
0
1m
lacquer peels
8 Twente Formation
1-7 Turnhout Member-Tegelen Formation

MEERLE SLIKGAT

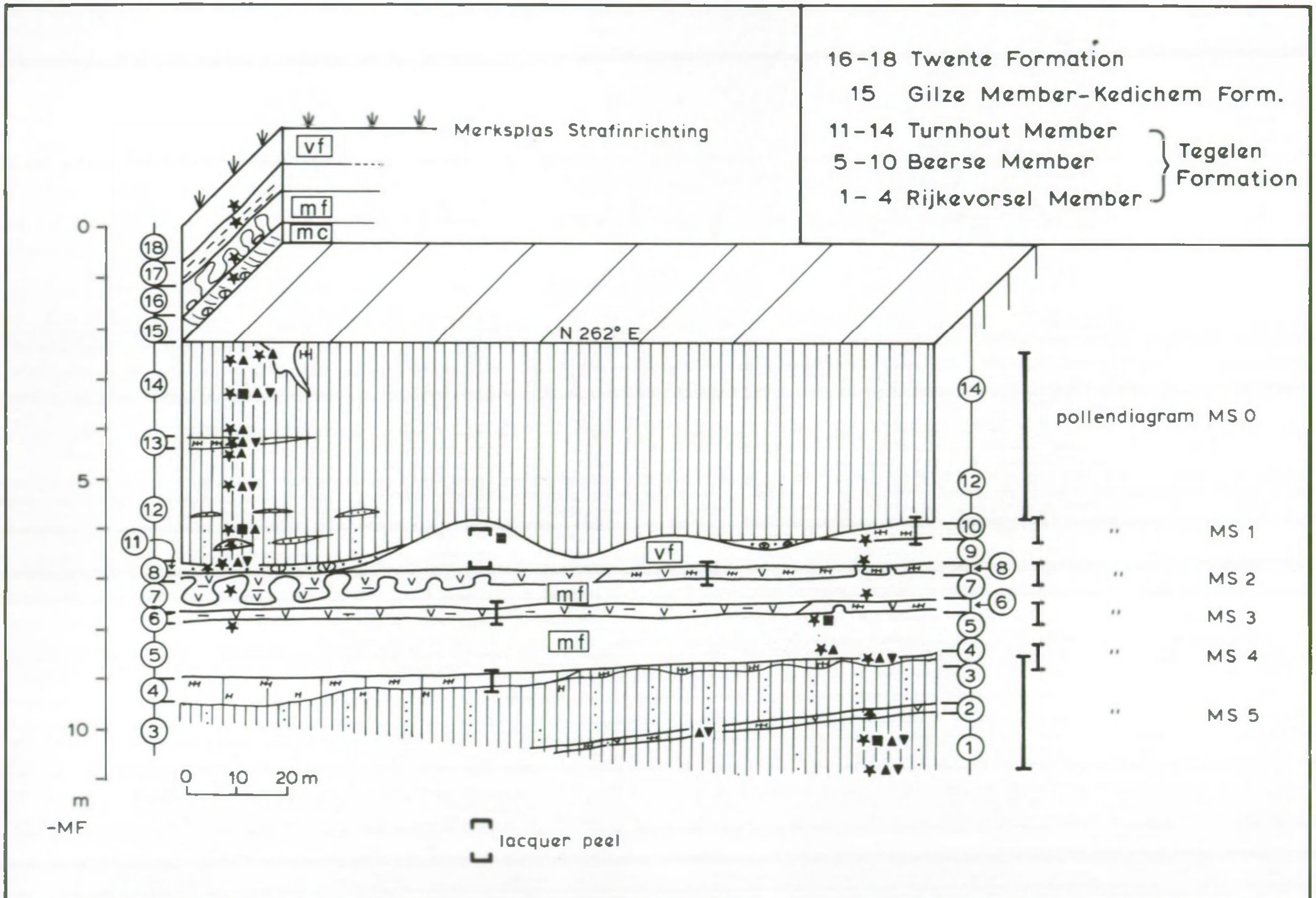
X = 4°49'18''OL
Y = 51°28'35''NB
H = 13 m +N.A.P.



EBUR. = EBURONIAN

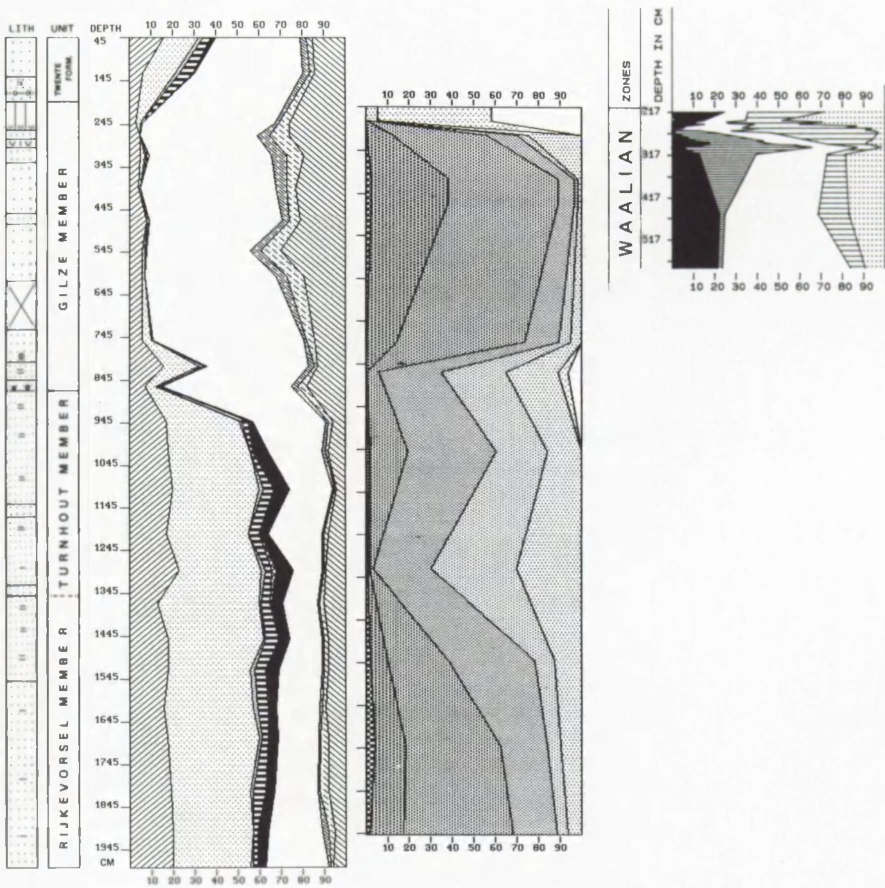
MERKSPLAS STRAFINRICHTING

X = 4°49'51''OL
Y = 51°21'45''NB
H = 25 m +N.A.P.



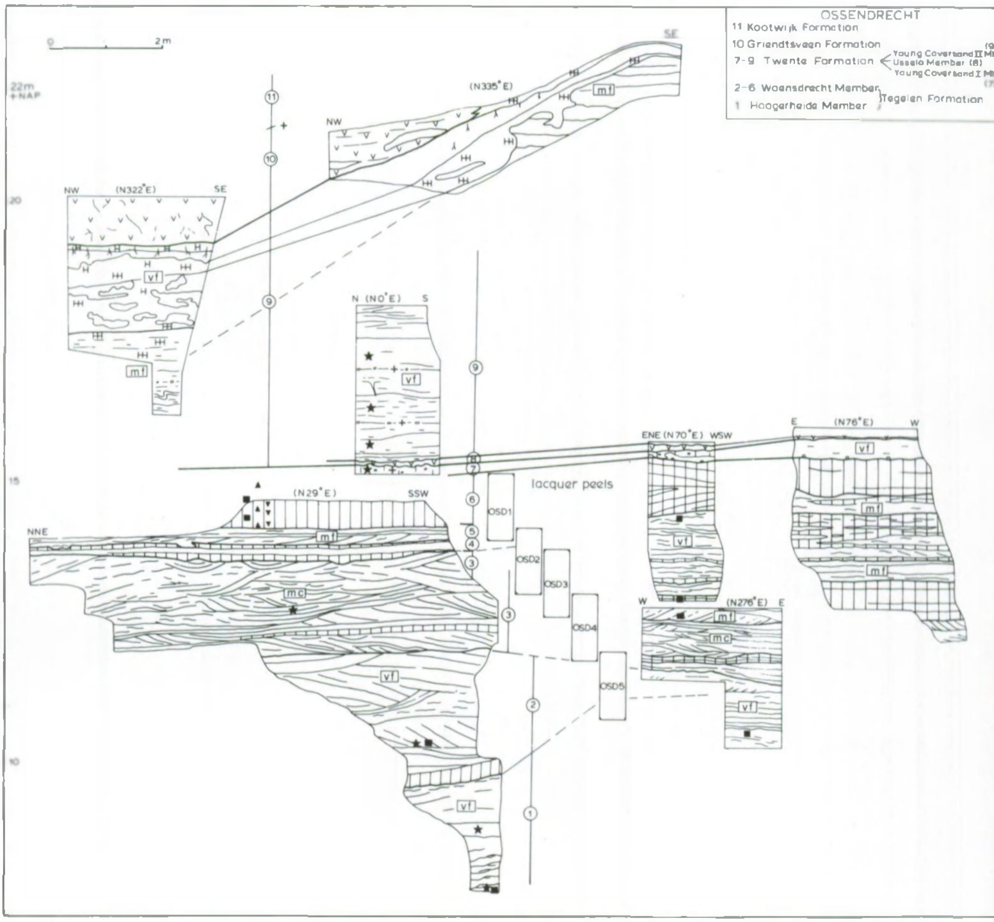
NIEUWMOER

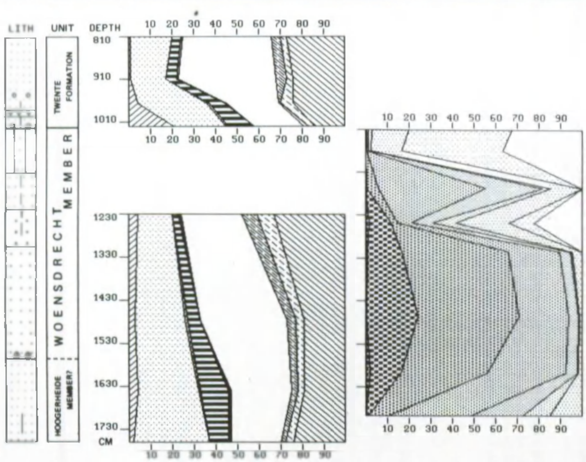
X = 4°30'22''OL
Y = 51°26'32''NB
H = 14 m +N.A.P.

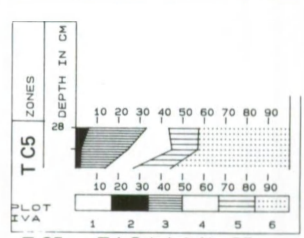


OSSENDRECHT

X = 4°20'44''OL
Y = 51°23'51''NB
H = 24 m +N.A.P.



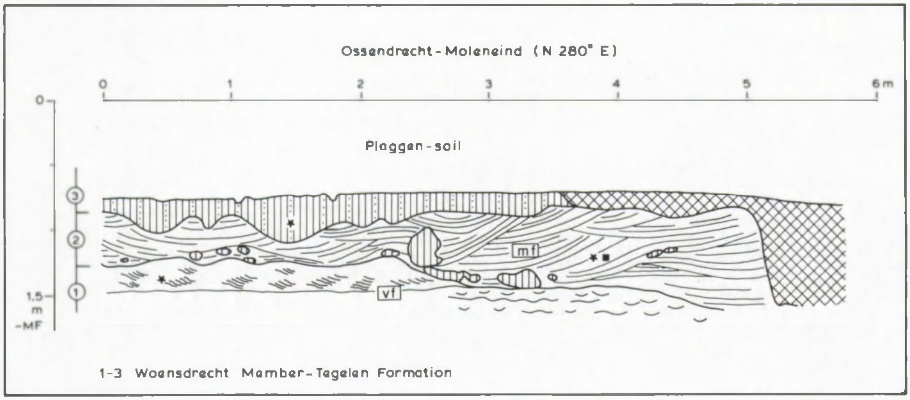


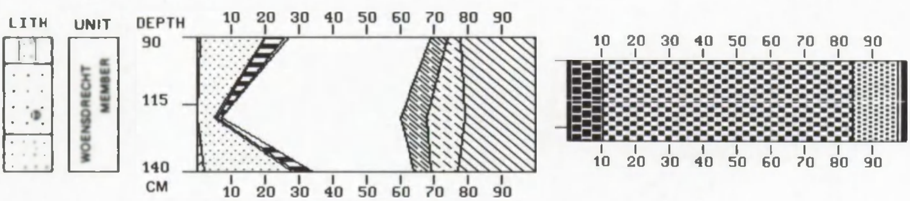


TC5 = TIGLIAN C5

OSSENDRECHT MOLENEIND

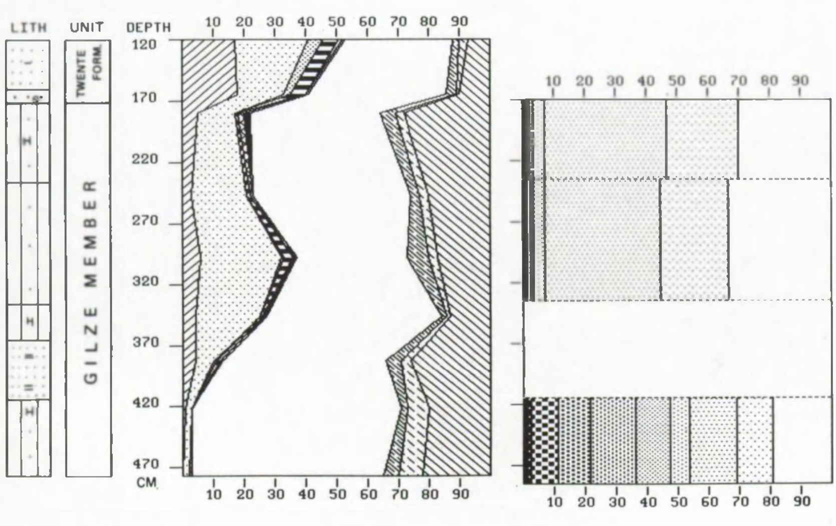
X = 4°20'29''OL
Y = 51°23'30''NB
H = 13 m +N.A.P.





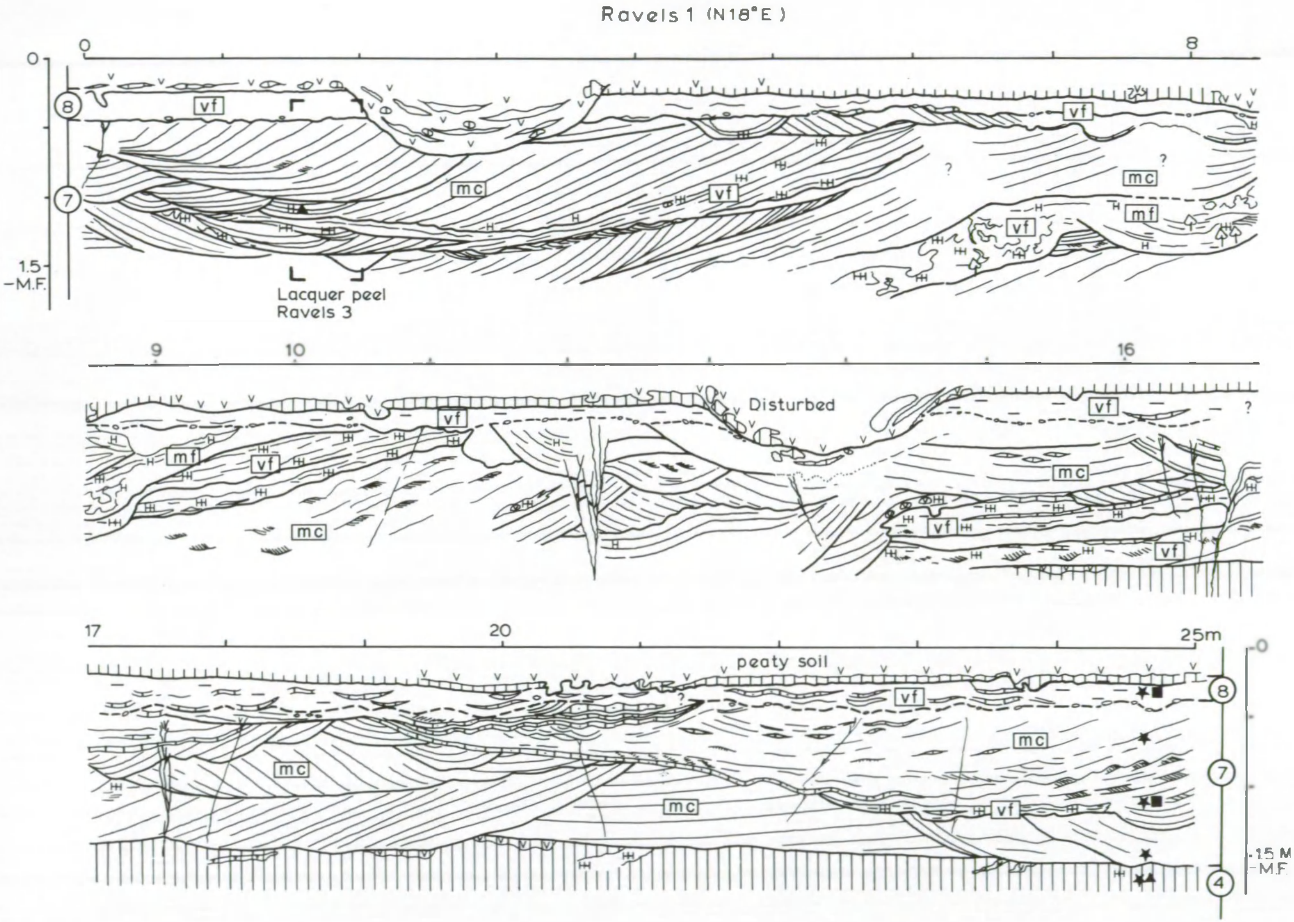
OSSENGOOR

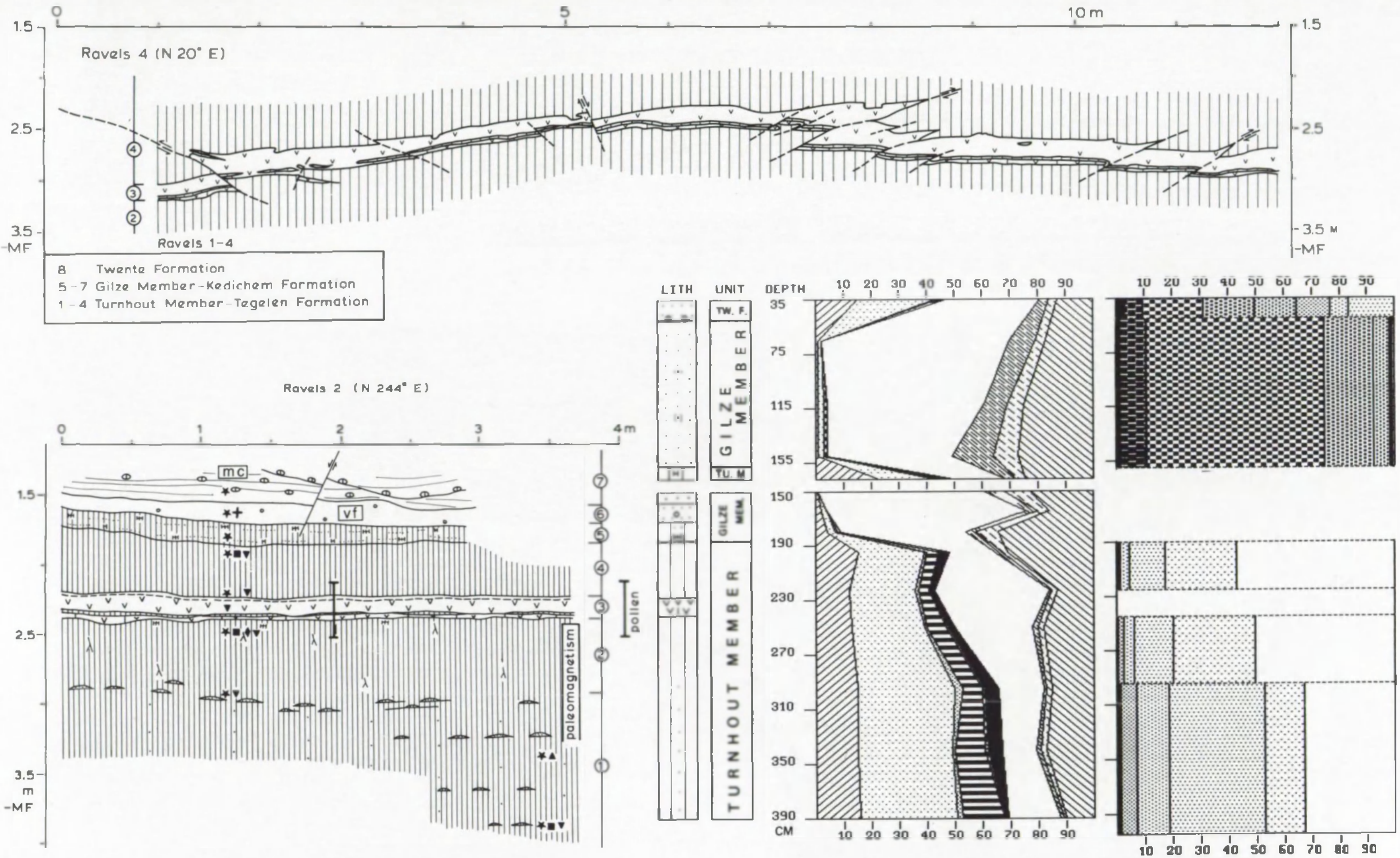
X = 4°54'02''OL
Y = 51°31'19''NB
H = 16 m +N.A.P.



RAVELS

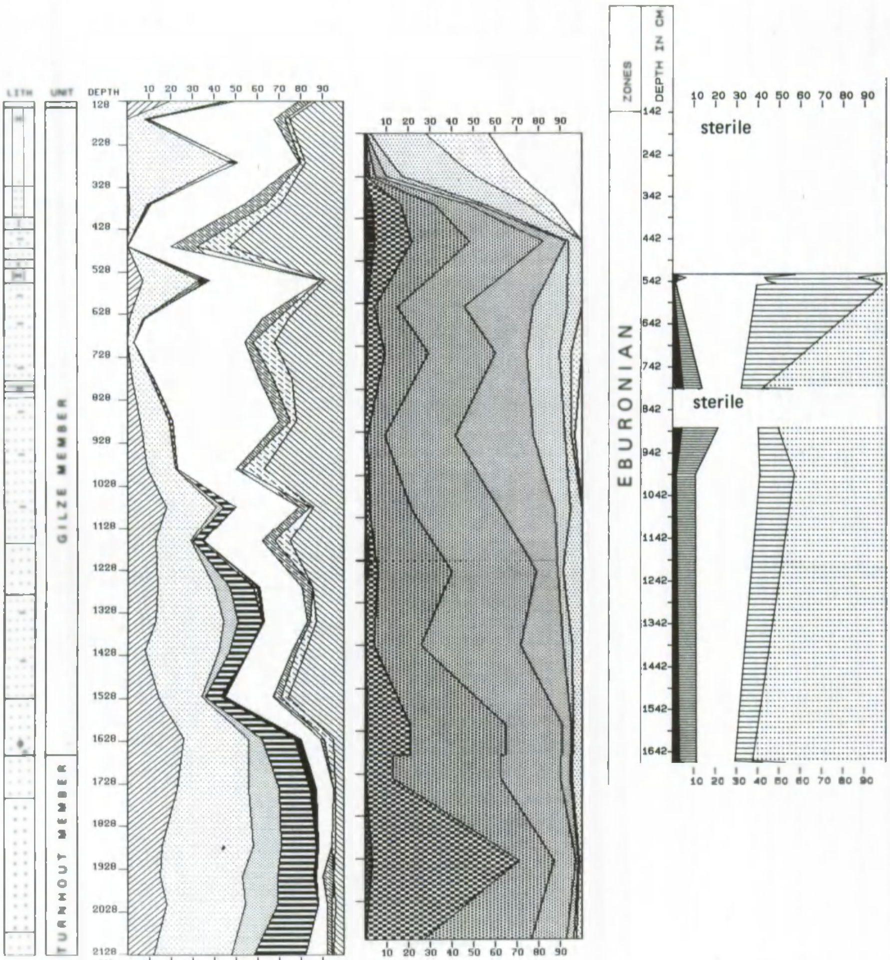
X = 4°58'26''OL
Y = 51°21'47''NB
H = 26.5 m +N.A.P.




Ravels 4 (N 20° E)
0
5
10 m
1.5
2.5
3.5
-MF
3.5 M
-MF
Ravels 1-4
8 Twente Formation
5 - 7 Gilze Member-Kedichem Formation
1 - 4 Turnhout Member-Tegelen Formation
Ravels 2 (N 244° E)
0
1
2
3
4 m
1.5
2.5
3.5
m
-MF
mc
vf
pollen
paleomagnetism
LITH
UNIT
DEPTH
TW. F.
GILZE MEMBER
TU. M
GILZE MEM.
TURNHOUT MEMBER
35
75
115
155
150
190
230
270
310
350
390
CM
10 20 30 40 50 60 70 80 90

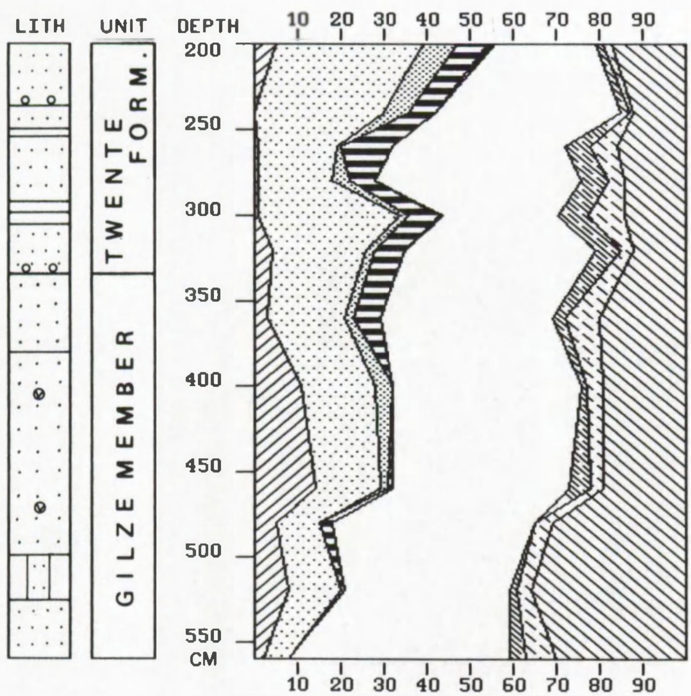
SNIJDERS-CHAAM

X = 4°51'51''OL
Y = 51°31'51''NB
H = 9 m +N.A.P.



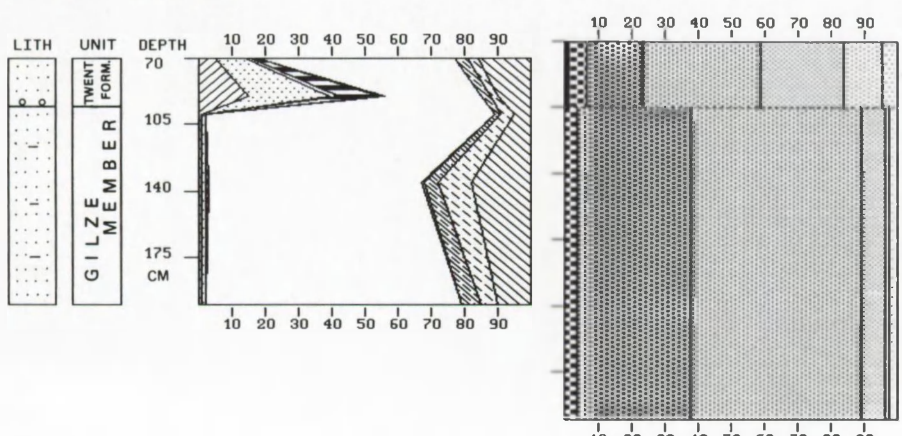
WEELDE

X = 4°57'58''OL
Y = 51°24'45''NB
H = 30 m +N.A.P.



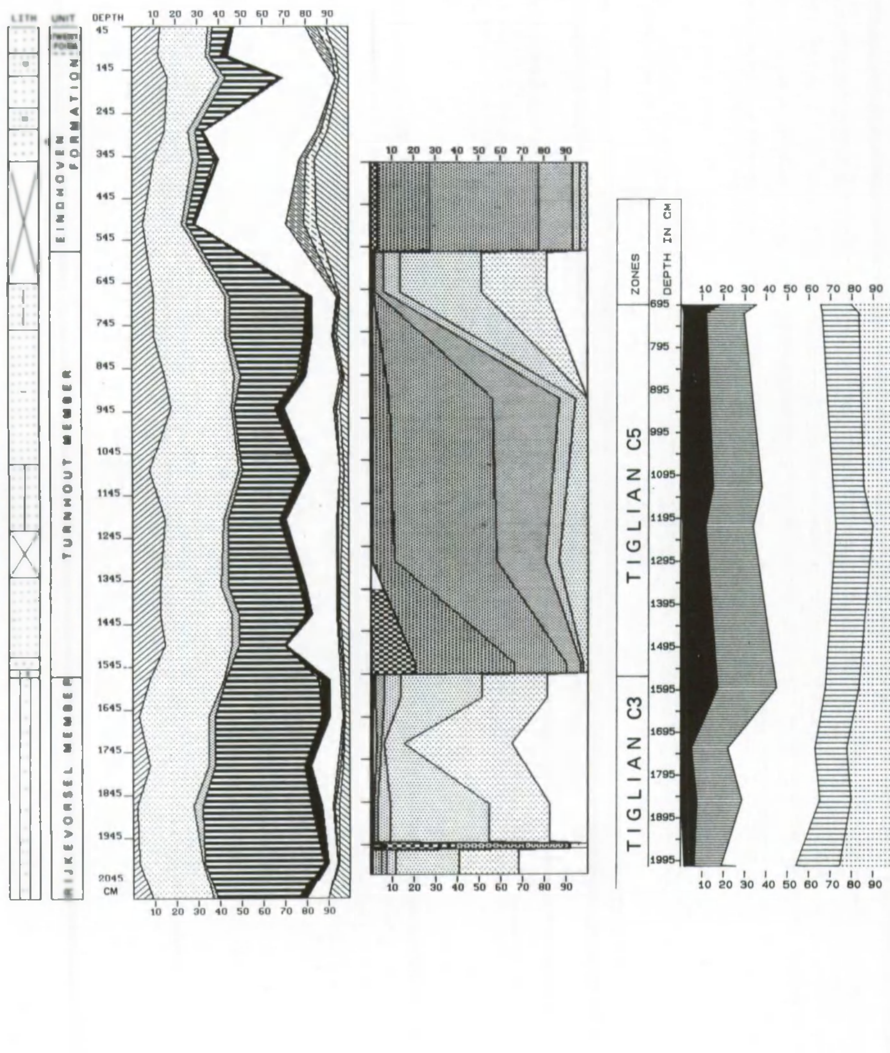
WITTE BERGEN

X = 4°53'09''OL
Y = 51°26'12''NB
H = 21.5 m +N.A.P.



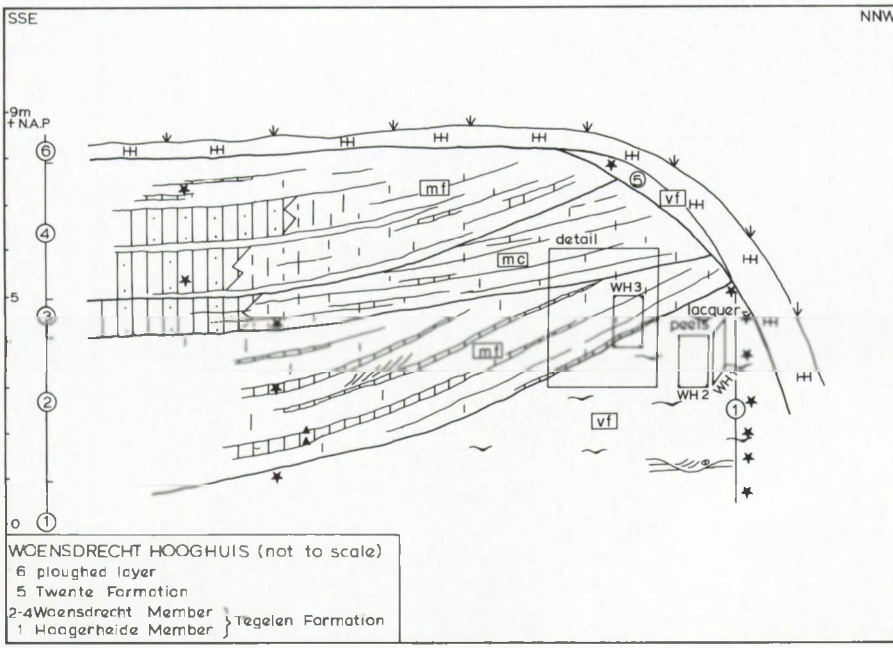
WERNHOUT MAALBERGEN

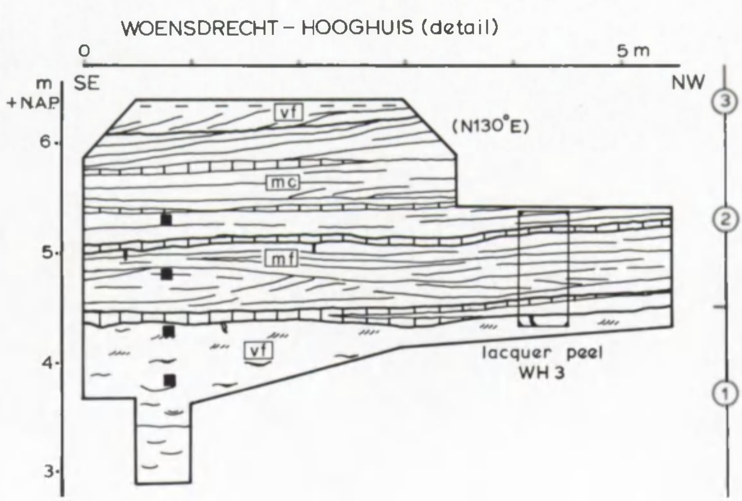
X = 4°40'00''OL
Y = 51°26'16''NB
H = 13 m +N.A.P.



WOENSDRECHT HOOGHUIS

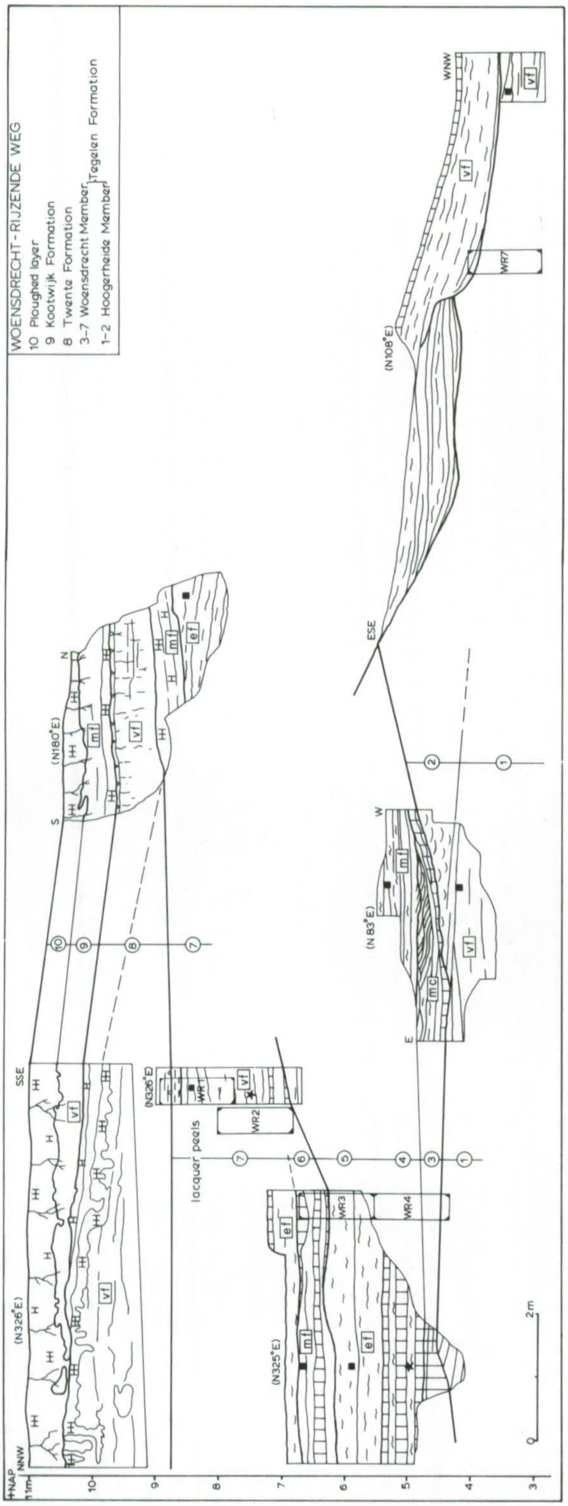
X = 4°18'33''OL
Y = 51°26'20''NB
H = 8 m +N.A.P.

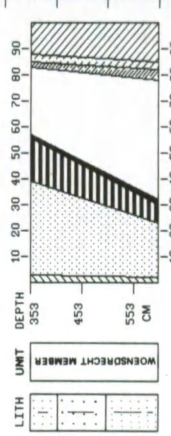


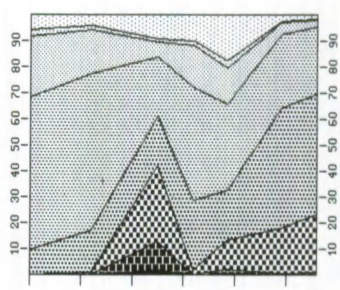


WOENSDRECHT RIJZENDE WEG

X = 4°18'26''OL
Y = 51°26'03''NB
H = 10.5 m +N.A.P.

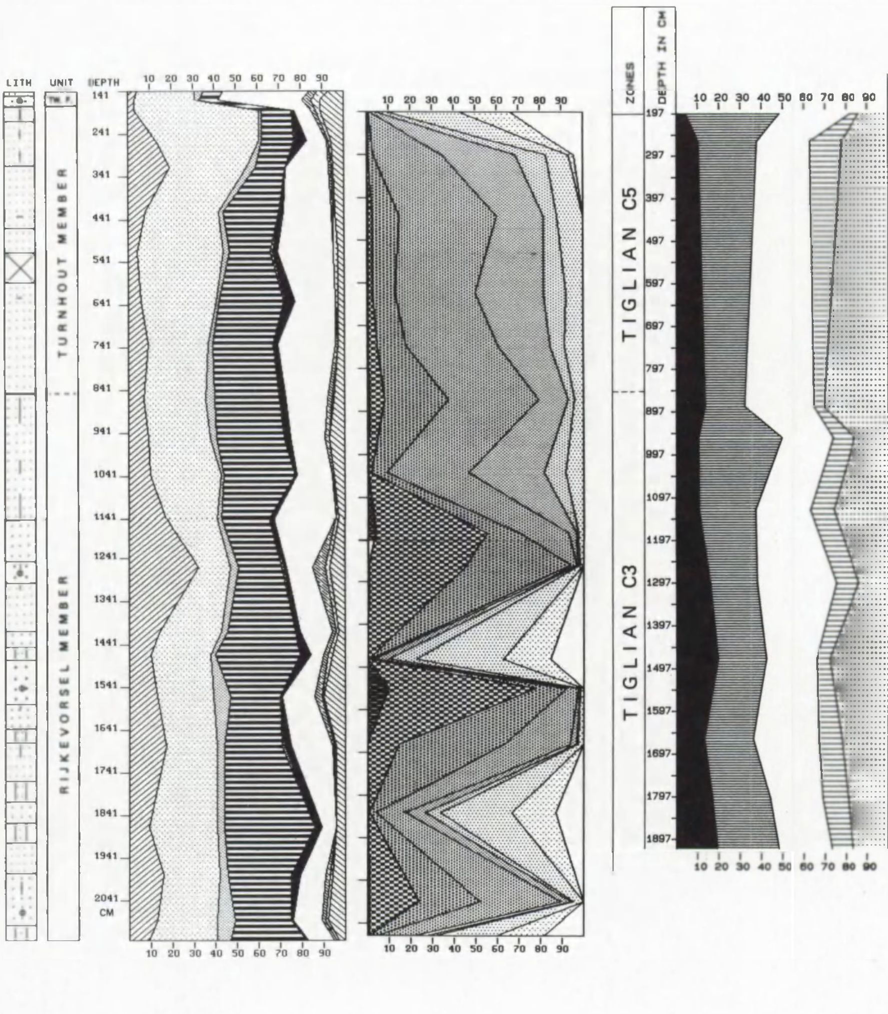






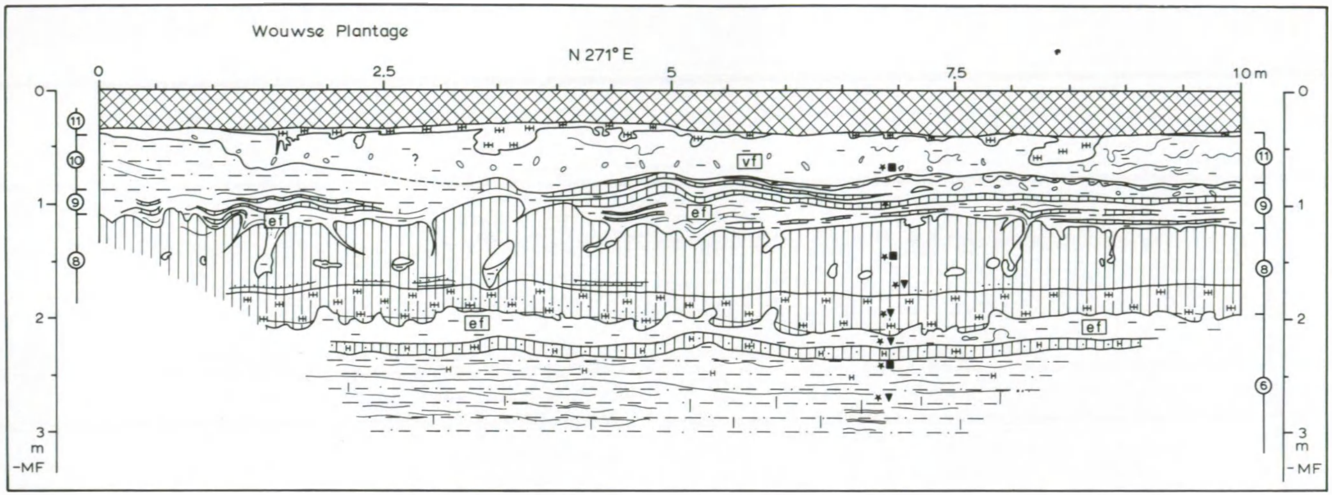
WORTEL

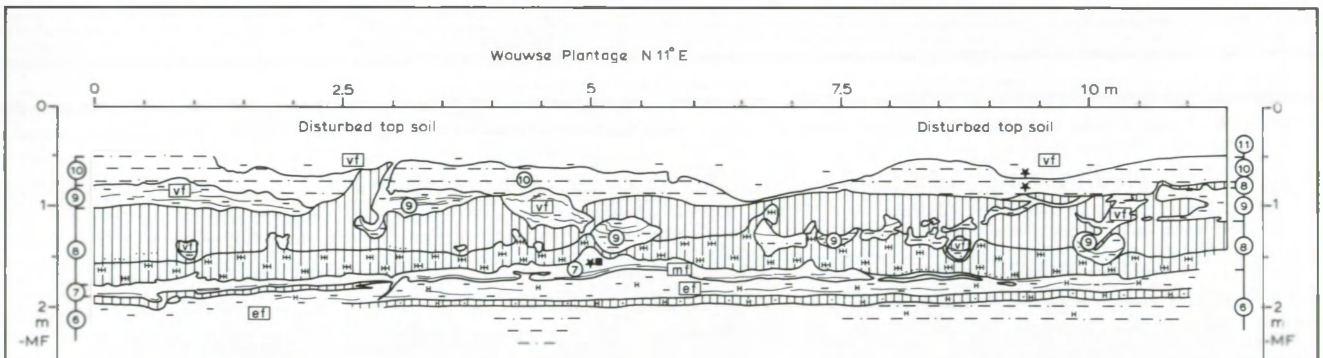
X = 4°47'59''OL
Y = 51°24'20''NB
H = 17 m +N.A.P.



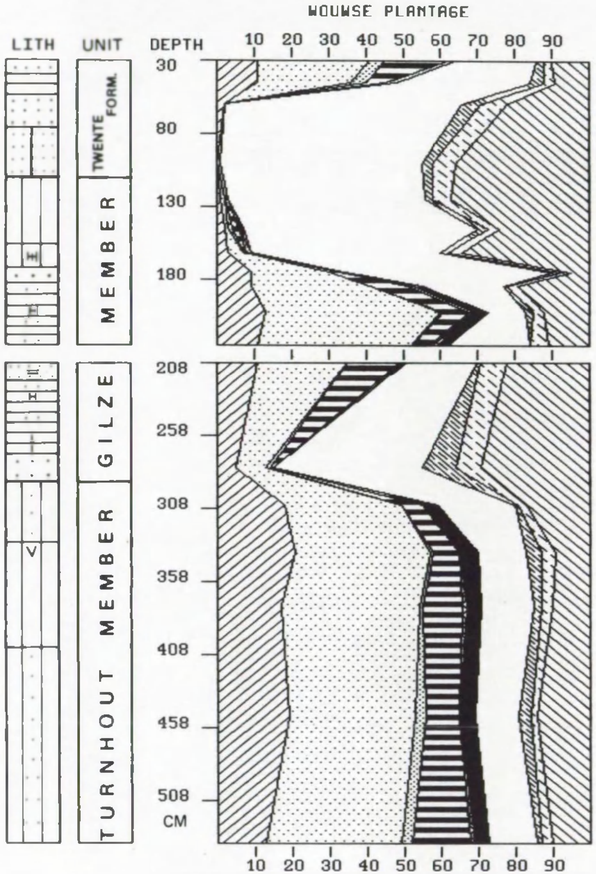
WOUWSE PLANTAGE

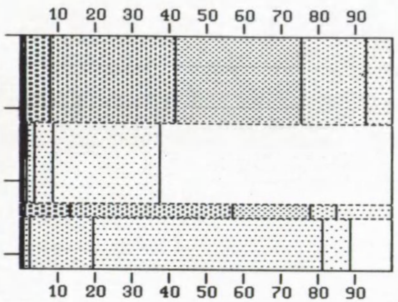
X = 4°23'30''OL
Y = 51°28'51''NB
H = 10 m +N.A.P.

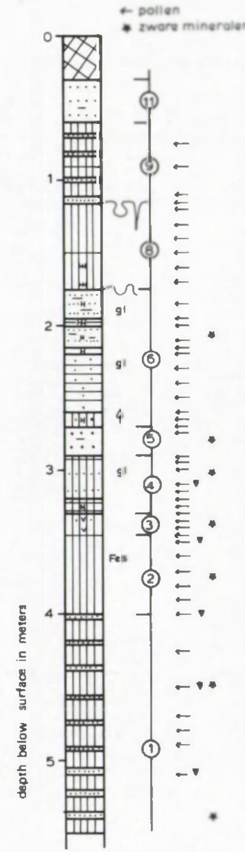


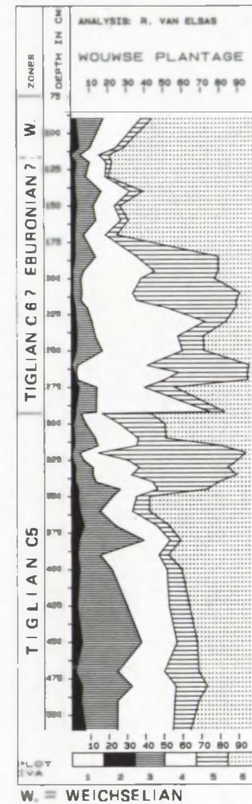


9 -11 Twente Formation
5 - 8 (Turnhout M.) Gilze Member-(Tegelen) Kedichem Formation
1 - 4 Turnhout Member-Tegelen Formation


WOUWSE PLANTAGE
LITH
UNIT
DEPTH
TWENTE FORM.
MEMBER
GILZE
TURNHOUT MEMBER
30
80
130
180
208
258
308
358
408
458
508
CM
10 20 30 40 50 60 70 80 90

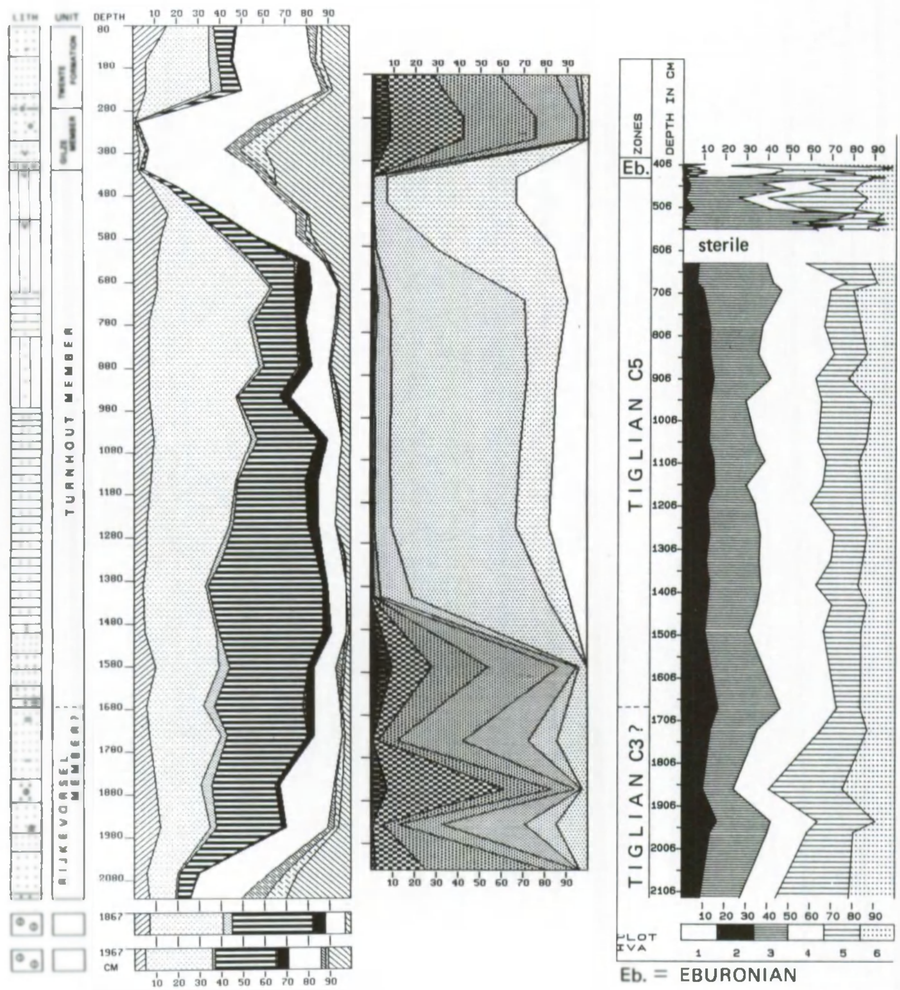

10 20 30 40 50 60 70 80 90


← pollen
* zware mineralen
depth below surface in meters
0
1
2
3
4
5
gl
Fe
1
2
3
4
5
6
8
9
11


ANALYSIS: R. VAN ELSAS
WOUWSE PLANTAGE
ZONES
DEPTH IN CM
10 20 30 40 50 60 70 80 90
W.
EBURONIAN ?
TIGLIAN C 6 ?
TIGLIAN C5
1 2 3 4 5 6
W. = WEICHSELIAN

ZWART WATER

X = 4°56'27''OL
Y = 51°22'29''NB
H = 27.5 m +N.A.P.



ZWART GOOR

X = 4°53'38''OL
Y = 51°23'06''NB
H = 24 m +N.A.P.